An Early Permian (ca. 280 Ma) silicic igneous province in the Alxa Block, NW China: A magmatic flare-up triggered by a mantle-plume?

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1. Introduction

Silicic large igneous provinces (SLIP), composed of predominantly silicic extrusive and intrusive rocks with areal extents >0.1 Mkm² and volumes >0.25 Mkm³ (Bryan, 2007; Bryan and Ernst, 2008; Bryan et al., 2002), are a recently recognized type of large igneous province (LIP) defined by Bryan et al. (2002). They have potentially important implications for geodynamic processes, economic resources (Bryan, 2007; Bryan and Ferrari, 2013; Pankhurst et al., 2011a), and the environment (Bryan, 2007).

During the last two decades, a significant number of SLIPs have been recognized including the Tertiary Sierra Madre Occidental of Mexico (Bryan et al., 2002), the Early Cretaceous Whitsunday igneous province (Bryan, 2007; Bryan et al., 1997, 2000), the Jurassic Chon Aike Province (Pankhurst et al., 1998, 2000), the Paleozoic Kennedy–Conners–Auburn province, northeast Australia (Bryan et al., 2003) and the Mesoproterozoic Gawler SLIP, southern Australia (Allen et al., 2008; Pankhurst et al., 2011b). These SLIPs are characterized by large volumes (>10⁵ km³), of predominantly rhyodacite–rhyolite compositions, long periods of magmatic evolution (<40 myr) and were located along paleo- or active continental margins (Bryan, 2007; Bryan et al., 2002). Unlike the mafic LIPs commonly related to mantle plumes (e.g., Bryan and Ernst, 2008; Chung et al., 1998; Coffin and Eldholm, 1994; Ernst and Buchan, 2001, 2003), the mechanisms for the formation of the SLIPs are not as clear and the generation of SLIPs may be related to continental rifting, mantle plumes or back arc extension (Betts et al., 2005; Bryan, 2007; Bryan and Ferrari, 2013; Bryan et al., 2002; Pankhurst et al., 2000, 2011b).

Early Permian (ca. 280 Ma) igneous rocks with predominantly ultramafic–mafic rocks are widespread in the Tarim Craton and the Central Asian Orogenic Belt (CAOB) (e.g., Li et al., 2011; Wei et al., 2014; Xia et al., 2012; Xu et al., 2013; Yang et al., 2007, 2013; Zhang et al., 2010; Zhou et al., 2004, 2009), and correspond to the Tarim LIP (e.g., Qin et al., 2011; Su et al., 2011, 2012). Recent dating of rocks on the Alxa Block, adjacent to the Tarim Craton and the CAOB, reveals that a large area of Early Permian (ca. 280 Ma) granitoids occur in the Block (e.g., Geng and Zhou, 2012; Li, 2006; Shi et al., 2012). Thus, although they may represent a silicic igneous province, its area (0.05 Mkm²) is smaller than a SLIP as defined by Bryan (2007) (>0.1 Mkm²). However, the petrogenesis and the tectonic setting of these granitoids, and their relationship to the broadly contemporaneous Tarim LIP are unclear. In this contribution, we present in situ zircon U–Pb age and Hf–O isotope data and whole-rock element and Nd–Sr isotope data for two newly
identified rock associations from the Alxa Block: Early Permian I-type and A-type granites plus high Sr/Y granodiorites in the Bayannuoergong area, and gabbros, diorites and quartz diorites in Bijierai Complex. Our work shows that a silicic igneous province corresponding to a ca. 280 Ma magmatic flare-up can be found in the Alxa Block, and suggest that the magmatic event was probably triggered by the contemporaneous Tarim mantle plume.

2. Geological setting

The Alxa Block, located in westernmost North China, connects the North China Craton, the Tarim Craton, the CAOB and the Qilian Block (Fig. 1a). It has recently been interpreted as an independent block by Dan et al. (2012, 2013, 2014), although it was previously thought to be a part of the North China Craton (NCC) (e.g. NMBGMR (Nei Mongol Bureau of Geology and Mineral Resources), 1991; Zhao et al., 2005). The Alxa Block is bounded by the Late Paleozoic Enger Us Ophiolite Belt in the north, the western margin fault of Bayanwulashan in the east and the southern margin fault of Longshoushan in the south (Fig. 1b). It is largely covered by deserts, and outcrops of pre-Neoproterozoic crystalline basement rocks are only seen in the southwestern parts of the block. The oldest rocks are the newly discovered ~2.5 Ga tonalite-trondhjemite-granodiorite (TTG) exposed in the Beidashan Complex (Gong et al., 2012; Zhang et al., 2013a) and Paleoproterozoic rocks with ages of ~2.3–1.9 Ga exposed in the Longshoushan Complex that were metamorphosed at ~1.9 Ga and ~1.8 Ga (Gong et al., 2011; Tung et al., 2007). The Mesoproterozoic medium to low-metamorphic grade Alxa Group (Geng et al., 2007) was unconformably covered by Neoproterozoic un-metamorphosed to very low-grade metamorphosed sedimentary sequences (NMBGMR (Nei Mongol Bureau of Geology and Mineral Resources), 1991; Chen et al., 2004). In the Neoproterozoic, a few S-type granites with ages of 930–910 Ma were intruded in the central Alxa Block (Dan et al., 2014; Geng et al., 2002).

A large volume of Phanerozoic granitoids are exposed in the Alxa Block (NMBGMR (Nei Mongol Bureau of Geology and Mineral Resources), 1991). Granitoids with ages >320 Ma are rare, however, and only a few are exposed near the eastern margin of the Alxa Block (Dan et al., 2013; Li, 2006). Recent studies suggest that they were mainly formed at 300–260 Ma (Fig. 1b) (Geng and Zhou, 2012; Lai et al., 2007; Li, 2006, 2012; Li et al., 2010a, 2010b; Pi et al., 2010; Shi et al., 2012; Wu, 2011). However, the corresponding geochemical data are relatively scarce, and the petrogenesis and tectonic setting remain unclear.

Late Carboniferous–Early Permian volcanic rocks were discovered in the Upper Carboniferous–Lower Permian Amushan Formation. The formation is the oldest stratum in the Alxa Block and its age was established by LA–ICP-MS zircon dates of 320–302 Ma from the lower part of the formation (Lu et al., 2012). The lower and middle sections of the Amushan Formation consist of volcanic (mainly acidic rocks and subordinate basic rocks), clastic, and carbonate rocks; the upper section is a molasse composed of silty shale, sandstone, gravel-bearing
sandstone and conglomerate (Lu et al., 2012). The basic rocks show affinities with intraplate magmatism, indicating that this formation was formed in a continental rifting setting (Dang et al., 2011; Jiang et al., 2011).

3. Regional geology

3.1. Bayannuoergong batholith

The Bayannuoergong batholith is one of the largest in the Alxa Block, and outcrops over 2000 km². It intruded into Precambrian strata and contains a few Precambrian xenoliths (Fig. 2a). The batholith consists mainly of coarse monzogranites and granodiorites, along with a few diorites and A-type granites (Fig. 2a), and some later felsic dykes (Fig. 2a). Mafic microgranular enclaves are commonly observed within the monzogranites, especially at pluton margins, and also occur within granodiorites. Recent LA–ICP-MS zircon U–Pb dating indicates that the emplacement ages of the monzogranites and granodiorites are 285–273 Ma (Li, 2012; Wu, 2011) and 265 Ma (Li, 2012), respectively.

The monzogranites and A-type granites consist of quartz (20–30 vol.%), alkali feldspar (20–30 vol.%), plagioclase (20–30 vol.%), biotite (0–10 vol.%), rare hornblende and accessory minerals (1–5 vol.%), including zircon and apatite. The granodiorites consist of quartz (20–25 vol.%), alkali feldspar (10–20 vol.%), plagioclase (50–60 vol.%), biotite (5–15 vol.%), hornblende (3–5 vol.%) and accessory minerals, including zircon and apatite.

3.2. Bijiertai Complex

The Bijiertai Complex, about 35 km to the north of Bayannuoergong batholith, is composed mainly of Early Paleozoic gabbros, diorites, quartz diorites, granodiorites, granites and two maﬁc-ultramaﬁc lenses within these rocks (Fig. 2b) (Feng et al., 2013; Geng and Zhou, 2012). Abundant gneissic sedimentary xenoliths with unknown-ages occur in the complex (Geng and Zhou, 2012). Recent LA–ICP-MS zircon U–Pb dating indicates that the diorites and granodiorites were formed at 288–276 Ma (Geng and Zhou, 2012), and the gabbro at 274 Ma (Feng et al., 2013).

The gabbros, diorites, and quartz diorites are studied in this paper. The gabbros are composed mainly of clinopyroxene (60–65 vol.%), plagioclase (30–40 vol.%) and minor biotite (2 vol.%), the diorites consist mainly of hornblende (30–40 vol.%), plagioclase (45–55 vol.%) and quartz (<5 vol.%), and the quartz diorites consist of hornblende (10–20 vol.%), plagioclase (60–75 vol.%) and quartz (5–10 vol.%). In contrast to the medium-grained gabbros and diorites, the quartz diorites are coarse-grained, and the plagioclases show cumulative texture.

4. Analytical procedures

4.1. Zircon U–Pb dating

Measurements of U, Th and Pb for samples 09AL22-1, 09AL27, 09AL33 and 09AL70 were conducted using the Cameca IMS-1280 SIMS at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG-CAS), Beijing. U–Th–Pb ratios and their absolute abundances were determined relative to the standard zircon Plešovice (Sláma et al., 2008) and 91500 (Wiedenbeck et al., 1995), respectively, using operating and data processing procedures similar to those described by Li et al. (2009, 2010c). Uncertainties on individual analyses in the data tables are reported at a 1σ level. Mean ages for pooled U–Pb and Pb/Pb analyses are quoted with 2σ and/or 95% conﬁdence intervals. The weighted mean U–Pb ages and Concordia plots were processed using an Isoplot/Ex v.3.0 program (Ludwig, 2003). SIMS zircon U–Pb isotopic data are presented in Supplementary Table A.1.

The LA–ICP-MS U–Pb dating for samples 09AL12-3, 09AL37 and 09AL73 was conducted using an Agilent 7500a ICP-MS with an attached 193 nm excimer ArF laser-ablation system (GeoLas Plus) at IGG-CAS. The analytical procedures are similar to those described by Xie et al. (2008). 207Pb/206Pb and 206Pb/238U ratios were calculated using the ICPMSDataCal software (Liu et al., 2010a, 2010b), using the zircon standard 91500 as an external standard. Common Pb was corrected according to the method proposed by Andersen (2002). The weighted mean U–Pb ages and Concordia plots were processed using the Isoplot/Ex v.3.0 program (Ludwig, 2003). Analyses of the zircon standard CJ1 as an unknown yielded a weighted mean 206Pb/238U age of 606 ± 4 Ma (2σ, n = 25), which is in good agreement with the recommended value (Jackson et al., 2004). LA–ICP-MS zircon U–Pb isotopic data are presented in Supplementary Table A.1.

4.2. 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. (2010d). The measured oxygen isotopic data were corrected for instrumental mass fractionation (IMF) using the Penglai zircon standard (δ18O = 5.3‰) (Li et al., 2010e).

![Fig. 2.](image-url) (a) Geological sketch-map of the Bayannuoergong batholith (NMBCMR, Nei Mongol Bureau of Geology and Mineral Resources, 1991) and (b) Bijiertai Complex, with locations of dating samples.
The internal precision of a single analysis generally was better than 0.2‰ (1σ standard error) for the δ18O/δ16O ratio. The external precision, measured by the reproducibility of repeated analyses of the Penglai standard, is 0.41‰ (2σ, n = 120). Ten measurements of the 91500 zircon standard during the course of this study yielded a weighted mean of δ18O = 10.2 ± 0.5‰ (2σD), 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 Supplementary Table A.2.

4.3. 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 GeoLas 2005 excimer ArF laser ablation system at the State Key Laboratory of Geochemical Processes and Mineral Resources, China University of Geosciences in Wuhan. Lu–Hf isotopic analyses were conducted on the same zircon grains that were previously analyzed for U–Pb and O isotopes. Detailed analytical procedures were similar to those described by Hu et al. (2012). Measured 176Hf/177Hf ratios were normalized to 176Hf/177Hf = 0.7325. Further external adjustment was not applied for the unknowns because our determined 176Hf/177Hf ratios for zircon standards 91500 (0.282308 ± 0.000004) and CJ-1 (0.282021 ± 0.000011) were in good agreement within errors with the reported values (Griffin et al., 2006; Wu et al., 2006; Zeh et al., 2007). Zircon Hf isotopic data are listed in Supplementary Table A.2.

4.4. Major and trace elements

Sixteen powdered rock samples of ~200-mesh size were used for geochemical analyses. Major element oxides were analyzed on fused 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 Supplementary Table A.3.

4.5. Sr and Nd isotopic compositions

Sixteen samples were selected for whole rock Rb–Sr and Sm–Nd isotopic analyses (Supplementary Table A.4). Sr–Nd isotopic compositions were determined using a Micromass isotope multi-collector ICP-MS at SKLIG GIG CAS, and analytical procedures described by Li et al. (2004). Sr and Nd were separated using cation columns, and Nd fractions were further separated by HDEHP-coated Kef columns. The measured 87Sr/86Sr ratio of the NBS 987 standard and 143Nd/144Nd ratio of the JNd1-1 standard were 0.710274 ± 18 (n = 11, 2σ) and 0.512093 ± 11 (n = 11, 2σ), respectively. All measured Nd and Sr isotope ratios were normalized to 146Nd/144Nd = 0.7219 and 88Sr/86Sr = 0.1194, respectively. The Sr–Nd isotopic results are listed in Supplementary Table A.4.

5. Results

5.1. Zircon U–Pb dating results

Zircon grains separated for dating are mostly euhedral to subhedral, with lengths of ~100–250 μm, and length to width ratios of 2:1 to 4:1. The oscillatory zoning cathodoluminescence images in most grains and high Th/U ratios (0.12–1.04) (Supplementary Table A.1), suggest magmatic origins (Belousova et al., 2002). SIMS was used to date all samples except for samples 09AL12-3, 09AL37 and 09AL73, which were dated by LA-ICP-MS. U–Pb concordia diagrams of analyzed zircons are shown in Fig. 3, and the U–Pb age data are given in Supplementary Table A.1.

Seven samples were selected for dating. Samples 09AL22-1 and 09AL27 are monzogranites and give weighted mean 206Pb/238U ages of 278 ± 2 and 284 ± 3 Ma, respectively (Fig. 3a, b). Sample 09AL12-3 is an A-type granite and produces a weighted mean 206Pb/238U age of 279 ± 2 Ma (Fig. 3c). Samples 09AL33 and 09AL37 are granodiorites and have weighted mean 206Pb/238U ages of 281 ± 2 Ma and 283 ± 2 Ma, respectively (Fig. 3d, e). Samples 09AL70 and 09AL73 are diorites and quartz diorites, respectively. They give weighted mean 206Pb/238U ages of 280 ± 2 Ma (Fig. 3f) and 282 ± 2 Ma (Fig. 3 g), respectively.

In summary, the zircon U–Pb age data indicate that the granitoids in the Bayanuervoergong batholith and the diorites and quartz diorites in the Bijiertai Complex were all formed at ca. 280 Ma (284–278 Ma).

5.2. Whole-rock major and trace element compositions

5.2.1. The granitoids from Bayanuervoergong batholith

The monzogranites from Bayanuervoergong batholith are high in silica and alkalis, with SiO2 ranging from 69.5 to 78.6 wt.% (volatile-free) and total K2O + Na2O contents varying from 7.0 to 8.7 wt.% (Supplementary Table A.3). The granodiorites have slightly lower silica and alkalies contents with SiO2 ranging from 63.6 to 66.6 wt.% and total K2O + Na2O varying from 5.3 to 6.0 wt.% (Supplementary Table A.3). The monzogranites and granodiorites are subalkaline on the alkali versus silica diagram (Fig. 4a). They have lower TiO2, Fe2O3, MnO, MgO, CaO, and P2O5 contents (Supplementary Table A.3), but higher Th and (Na2O + K2O)/CaO than the gabbro, diorite and quartz diorite from the Bijiertai Complex (Fig. 5). The A-type granites (09AL12-1, 09AL12-2 and 09AL12-3) have high SiO2 (78.5–78.6 wt.%), alkalis (Na2O + K2O = 8.2–8.5 wt.%) and FeOt/MgO ratios, and low CaO, MgO contents (Fig. 5; Supplementary Table A.3).

Chondrite-normalized RIE patterns of the monzogranites and granodiorites (Fig. 4b) show relative enrichment of light rare earth elements (LREEs), with variable (La/Yb)N ratios (3–58). The monzogranites are characterized by significant negative Eu anomalies, whereas the granodiorites display negligible Eu anomalies. Five monzogranites (09AL18-1, 09AL18-2, 09AL19-1, 09AL19-2, 09AL21) show heavy rare earth element (HREE) depletion, with high (La/Yb)N ratios (16–58). Three samples (09AL12-1, 09AL12-2, 09AL12-3) show negative anomalies in Ba, Nb, Ta, Sr, P, Eu and Ti (Fig. 4c), characteristics of A-type granites (e.g., Eby, 1990, 1992; Yang et al., 2006). The HREE abundances of the granodiorites are lower than those of the monzogranites. Furthermore, the granodiorites are characterized by relatively high Sr but low Yb and Y contents with high Sr/Y ratios of 22–71, similar to those of modern adakites and Archean TTG suites (Fig. 4d) (e.g., Defant and Drummond, 1990; Martin et al., 2005).

5.2.2. Bijiertai Complex: gabbro, diorite and quartz diorite

The gabbros, diorites and quartz diorites have SiO2 contents of 43.0–45.3 wt.%, 50.3–56.5 wt.% and 57.6–62.2 wt.%, respectively. The gabbros have high Mg# values of 46.5–66.4, but the diorites and quartz diorites have lower Mg# values of 47.1–52.1 and 38.1–40.1, respectively. All three rock types plot in the subalkaline field on the alkali versus silica diagram (Fig. 4a).

All of these rocks are enriched in LREE. The gabbros and diorites have negligible Eu anomalies, whereas quartz diorites have significant positive Eu anomalies. Gabbros show nearly flat LREE and fractionated HREE patterns with significant negative Ta and Nb anomalies. The quartz diorites are characterized by high Sr but low Yb and Y contents with high Sr/Y ratios of 53–155 (Supplementary Table A.3).
Fig. 3. In situ U–Pb dating results for the Early Permian rocks in Alxa Block. (a, b) Monzogranites (09AL22-1 and 09AL27), (c) A-type granite (09AL12-3), (d, e) granodiorites (09AL33 and 09AL37), (f) diorite (09AL70) and (g) quartz diorite (09AL73). Data-point error ellipses are 2σ.
5.3. Whole rock Sr–Nd isotopic compositions

The monzogranites exhibit the lowest and most variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios and highly variable $\varepsilon_{\text{Nd}}(t)$ values (Fig. 4f), corresponding to two-stage Nd model ages ($T_{\text{DM}}$) of 2.27–1.72 Ga. The granodiorites have higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios (0.7086–0.7098) and $\varepsilon_{\text{Nd}}(t)$ values from −12.0 to −9.5. The gabbros, diorites and quartz diorites from the Bijiertai Complex have initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios ranging from 0.7086 to 0.7098, and $\varepsilon_{\text{Nd}}(t)$ values ranging from −12.0 to −9.5. The monzogranites exhibit a significant negative correlation between $\varepsilon_{\text{Nd}}(t)$ values and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Fig. 4f).

Fig. 4. (a) $\text{SiO}_2$ vs. $\text{K}_2\text{O} + \text{Na}_2\text{O}$ diagram for intrusive rocks (Middlemost, 1994); (b) chondrite-normalized REE diagrams and (c) primitive mantle-normalized incompatible trace element diagrams for the Early Permian rocks, and the normalization values are from Sun and McDonough (1989); (d) Sr/Y vs. Y diagram (after Defant et al., 2002); (e) $\text{SiO}_2$ vs. $\text{MgO}$ diagram. The field of metabasaltic and eclogite experimental melts (1–4.0 GPa) is from the following references: Sen and Dunn (1994); Rapp and Watson (1995); Springer and Seck (1997); Rapp et al. (1999); Sajevic and Patiño Douce (2002), and references therein. Fields of adakites inferred to be derived from subducting oceanic crust is after Wang et al. (2006). Melts formed by partial melting of the lower mantle crust are from the following references: Atherton and Petford (1993); Chung et al. (2003); Johnson et al. (1997); Muir et al. (1995); Petford and Atherton (1996); Wang et al. (2005). (f) Nd–Sr isotope composition for the Early Permian igneous rocks. EMI, I-type enriched mantle, EMII, II-type enriched mantle (Zindler and Hart, 1986).
Fig. 5. Selected major oxides (wt.%) and trace elements (ppm) vs. SiO$_2$ (wt.%) for the Early Permian rocks.
5.4. Zircon Hf–O isotopic compositions

In situ LA–MC–ICP–MS Lu–Hf isotopic analyses were conducted on the zircon grains that were previously analyzed for U–Pb and/or O isotopes. Their results are presented in Supplementary Table A.2 and Fig. 6. The monzogranites (samples 09AL22-1 and 09AL27) and granodiorites (09AL33 and 09AL37) have a large range of εHf(t) values of −17.8 to −3.3 (one point is −20.0), corresponding to two-stage zircon Hf model ages (TDM) of 2.43–1.51 Ga (one point is 2.56 Ga). The diorite (09AL70), quartz diorite (09AL73) and A-type granite (09AL12-3) have a smaller range of age-corrected εHf(t) values (Fig. 6b, Supplementary Table A.2), corresponding to TDM of 2.08–1.90 Ga, 1.95–1.67 Ga and 1.55–1.37 Ga, respectively. Most of the two-stage zircon Hf model ages are 2.2–1.8 Ga, but samples 09AL12-3 (A-type granite) and 09AL22-1 (monzogranite) give ages between 1.6 and 1.4 Ga. It is noted that the A-type granites have the highest εHf(t) values among these rocks.

Zircon O isotopes were analyzed on the diorite, granodiorite and monzogranite. The measured δ18O values for zircons from the diorite (09AL70) and granodiorite (09AL33) show similar and limited ranges forming normal Gaussian distributions, with averaged values of 6.80 ± 0.21‰ (1SD) and 6.61 ± 0.21‰ (1SD), respectively (Fig. 7). Measured zircon δ18O values for the monzogranites show a few anomalous spots. For sample 09AL22-1, spots show a limited range of 5.5–6.4‰, with a peak of 6.2‰, except for two anomalous spots at 4.9 and 5.0‰. For sample 09AL27, most spots show a slightly large range of 5.0–6.4‰, with a peak of 6.2‰, apart from two spots with anomalous values of 4.0‰ and 7.4‰ (Fig. 7).

6. Discussion

6.1. Ca. 280 Ma magmatic flare-up in the Aba Block

To better understand the petrogenesis and tectonic significance of the Permian magmatic rocks in the Aba Block, we carried out a regional data compilation (Supplementary Table A.5). Fig. 1b shows the representative Permian igneous rocks in the Aba Block, marked with representative ages. Fig. 8 shows the frequency distribution of the known Paleozoic igneous ages from the Aba Block and adjacent areas using Isoplot (Ludwig, 2003), highlighting peak activities at ~278 Ma. The Paleozoic magmatism can be broadly divided into three main episodes encompassing the Silurian to Early Carboniferous (424–337 Ma), the Late Carboniferous (320–300 Ma) and the Permian (290–250 Ma). The Silurian to Early Carboniferous igneous rocks are distributed in the eastern margin of the Aba Block and adjacent areas (Fig. 1b) and are probably caused by Paleozoic orogenesis in the eastern Aba Block (Dan et al., 2013). The Late Carboniferous magmatic rocks consist mainly of volcanic rocks and are distributed in the northern Aba Block and adjacent areas (Fig. 1b) and were possibly generated in an intraplate setting (Dang et al., 2011; Jiang et al., 2011).

The Permian igneous rocks are largely distributed in the Aba Block and adjacent areas. Most of them were formed at ca. 290–270 Ma, with a peak age of ~278 Ma (Fig. 8). This period of igneous activity is characterized by voluminous granites rather than mafic rocks (Supplementary Table A.5). Although the rocks are distributed over an area of 0.15 Mkm², the total area of outcrop (0.05 Mkm²) does not meet the 0.10 Mkm² threshold required for classification as a SLIP (Bryan, 2007; Bryan et al., 2002). Therefore, they cannot be classified as a SLIP at present. However, it is significant that the adjacent and areally extensive Upper Carboniferous–Lower Permian Amushan Formation in the northern Aba Block contains large exposures of andesitic to rhyolitic volcanic rocks (Dang et al., 2011). Establishing whether or not they are a part of the Aba silicic igneous province requires more high-precision geochronological data.

6.2. Petrogenesis of the ca. 280 Ma magmatic rocks

6.2.1. Bayannuoergong batholith

6.2.1.1. The monzogranites. Two end-member mixing may have played an important role in the generation of the monzogranites, based on the following lines of evidence. Their Nd isotopes negatively correlate with 87Sr/86Sr, suggesting a common mantle source of the monzogranites or contaminated their magmas. This is evident from the two-stage zircon Hf model ages (TDM = 1.90 Ga, 1.95 ± 0.15 Ma), the total area of outcrop (0.05 Mkm²) does not meet the 0.10 Mkm² threshold required for classification as a SLIP (Bryan, 2007; Bryan et al., 2002). Therefore, they cannot be classified as a SLIP at present. However, it is significant that the adjacent and areally extensive Upper Carboniferous–Lower Permian Amushan Formation in the northern Aba Block contains large exposures of andesitic to rhyolitic volcanic rocks (Dang et al., 2011). Establishing whether or not they are a part of the Aba silicic igneous province requires more high-precision geochronological data.
material is the most plausibly basalt that had undergone high-
temperature alteration in an oceanic setting. A high zircon O value 
(7.4‰) indicates that some other material was also added to the 
I-type monzogranite. These materials were probably Precambrian 
sedimentary rocks, as these are found as xenoliths in the monzogranites. 
This suggestion is consistent with the high Hf mantle model age (2.56 Ga) 
of this zircon grain (Supplementary Table A.2).

A few monzogranite samples display middle REE depletion (Fig. 4b). 
The concave REE patterns indicate that amphibolites may have been 
retained in their sources or were removed by crystal fractionation 
during magma evolution. These monzogranites have Sr–Nd isotopic 
compositions similar to the other monzogranites (Fig. 4f), suggesting 
that they have similar sources.

6.2.1.2. Occurrence of A-type granite in the Bayannuoergong batholith. 
Although many ~280 Ma granitoids have been reported in the Alxa 
Block, no A-type granitoids have been identified in previous studies. In 
general, A-type granites are comparatively enriched in high field 
strength elements (HFSEs), such as Zr, Nb, Y, REE and Ga (e.g., Collins 
et al., 1982; Eby, 1992; King et al., 1997, 2001; Sylvester, 1989; 
Whalen et al., 1987; Yang et al., 2006). In this study, a few granites in 
the Bayannuoergong batholith have the characteristics of A-type gran-
ites. Samples 09AL12-1, 09AL12-2 and 09AL12-3 have high K2O + 
Na2O, Zr, FeO/TiO2/MgO and Ga/Al ratios. On discrimination diagrams 
(Fig. 10a, b), they plot in the A-type granite field. Moreover, they 
show zircon saturation temperatures (849–854 °C) higher than those 
(683–833 °C) of other monzogranites (Fig. 10c). We therefore conclude 
that the three samples described in this study are A-type granites.

The origins of A-type granites have been attributed to: (1) direct frac-
tionation of mantle-derived alkaline basalts (e.g., Mushkin et al., 2003; 
Turner and Rushmer, 2009; Turner et al., 1992); (2) partial melting of 
crustal materials at high temperature (e.g., Collins et al., 1982; King 
et al., 1997; Wang et al., 2010a) and (3) hybridization between anatectic 
crust-derived felsic and mantle-derived mafic magmas (e.g., Kerr and 
Fryer, 1993; Mingram et al., 2000; Yang et al., 2006). The absence of 
any mafic microgranular enclaves in the A-type pluton does not support 
the generation of these A-type granites by mixing of mantle-derived 
magmas and crustal-derived silicic melts. The A-type granite has a high 
εNd(t) value, similar to the end-member component of the 
monzogranites with the highest εNd(t) values (Fig. 4f). Thus, it is difficult 
to distinguish a potential mantle-derived origin from the alternative 
origin via partial melting of crustal materials. The lack of coeval, proximal 
mufic rocks or rocks with intermediate compositions, their high SiO2 
contents relative to other ~280 Ma magmatic rocks, and their relatively 
low zircon saturation temperatures (~850 °C) compared to the 900– 
1100 °C temperatures calculated for mantle-derived A-type granites 
(e.g., Pankhurst et al., 2011a; Turner and Rushmer, 2009; Turner et al.,
6.2.1.3. High Sr/Y granodiorites. Although the granodiorites show some geochemical affinities with the “classical” adakites defined by Defant and Drummond (1990) (Fig. 4d), their K₂O/Na₂O ratios (0.58–0.80) are higher than those of typical adakites (K₂O/Na₂O ratios = 0.42) (Richards and Kerrich, 2007). Thus, they should be classed as high Sr/Y granitoids (Moyen, 2009). Several models have been proposed for the generation of high Sr/Y granitoids, including (a) melting of subducted young and hot oceanic crust (Defant and Drummond, 1990; Kay et al., 1993; Martin et al., 2005; Stern and Kilian, 1996; Tang et al., 2010a; Wang et al., 2007a, 2008); (b) assimilation and fractional crystallization (AFC) or fractional crystallization (FC) from parental basaltic magmas (Castillo et al., 1999; Macpherson et al., 2006; Richards and Kerrich, 2007); (c) magma mixing between felsic and basaltic magmas (Guo et al., 2007; Streck et al., 2007); and (d) partial melting of thickened lower crust (Atherton and Petford, 1993; Chung et al., 2003; Petford and Atherton, 1996; Wang et al., 2005, 2007b).

In the case of the Alxa Block, the first three models cannot account for the petrogenesis of the granodiorites. Oceanic slab melting is unlikely to have generated these high Sr/Y granitoids because of the apparent lack of contemporaneous subduction in the eastern Alxa Block and a slab source is inconsistent with their evolved Sr–Nd–Hf isotopic compositions (Figs. 3f and 5b). In addition, they have lower Mg# (42–47) values than those (47–56) of metasaltic rock-derived experimental melts contaminated by mantle peridotites (e.g., Rapp et al., 1999), indicating that the interaction between felsic magmas and mantle peridotites was unlikely. Additionally, their high δ¹⁸O values of zircon (6.61 ± 0.21‰ (1SD)) (corresponding to a calculated whole-rock magmatic δ¹⁸O value of 8.2‰) are not consistent with the mantle-like δ¹⁸O values of the best global examples of slab-derived adakites (Bindeman et al., 2005). Mantle-derived magmatic suites generated by FC processes generally exhibit a continuous compositional trend from basaltic rocks to felsic rocks derived from residual magmas (e.g., Castillo et al., 1999; Macpherson et al., 2006; Richards and Kerrich, 2007). The absence of contemporaneous basaltic rocks associated with these granitoids rules out this mechanism. Rocks formed by magma mixing are usually intermediate in composition and have high Mg (e.g., adakitic high-Mg andesites) (Guo et al., 2007; Streck et al., 2007) rather than acidic adakitic compositions in this study. Additionally, the homogeneous O and Hf isotope characteristics of these high Sr/Y samples are also inconsistent with a magma mixing model. Consequently, the remaining scenario of partial melting of a thickened lower crust is discussed below.

Partial melting of a thickened garnet-bearing mafic lower crust due to heat flux from the mantle (e.g., Atherton and Petford, 1993; Wang et al., 2005, 2007b) is a plausible mechanism for the generation of the high Sr/Y granodiorites. If the high Sr/Y magmas are actually derived from the mafic lower crust, then they should have relatively low MgO contents and be compositionally similar to 1–4.0 GPa experimental melts of metasalts. In Fig. 4e, the samples display linear trends that are parallel to the fields of melts formed by partial melting of the thickened lower crust. This process is also consistent with the relatively low zircon δ¹⁸O values (6.3–7.0‰) (Fig. 7a), which are within and slightly higher than the upper limit value of previously reported mantle-derived magma (6.5‰) (e.g., Valley et al., 2005). Additionally, the current crustal thickness of the Alxa Block is >45 km based on seismic data (Li, 2010).

6.2.2. Bijiertai Complex

The Bijiertai Complex contains many types of rocks, such as peridotites, gabbros, diorites, quartz diorites and newly discovered norites (Wang, 2012). However, the gabbros are cumulate rocks with REE patterns similar to other cumulate rocks (e.g., Ma et al., 2013a), thus, whole rock geochemical data cannot be used to constrain their sources. The low εNd(t) values (−9.9 to −9.5) of the gabbros, however, suggests that they were probably generated by partial melting of the lithospheric mantle.

The quartz diorites from the Bijiertai Complex have high Sr/Y ratios and the positive correlation between Sr/Y and Eu/Eu* (Fig. 9b) suggests that the plagioclase accumulation elevated the Sr/Y ratios (e.g., Ma et al., 2013b). The quartz diorites have εNd(t) values similar to the gabbros, indicating that they were probably generated by fractional crystallization from mafic magmas. Given that the quartz diorites have higher δ¹⁸O, it is likely that they also assimilated small amounts of supraarc crustal materials.

The diorite εNd(t) value of −12.0 is lower than those of the gabbros and quartz diorites but similar to Bayannuoergong batholith granodiorites, indicating that they may have been generated by the partial melting of the crustal materials, although the involvement of mantle-derived materials cannot be excluded. The diorites have low Sr/Y ratios, suggesting that they were produced at mid-crustal depths. The diorites have zircon δ¹⁸O values (6.80 ± 0.21‰) that are distinctly higher than those of Bayannuoergong batholith rocks derived by magma mixing (6.2–6.3‰) and slightly higher than those (6.61 ± 0.21‰) of the high Sr/Y granodiorites, consistent with a predominantly mid-crustal source.

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**Fig. 9.** Plots of (a) Ba/La vs. εNd(t) values, (b) Eu/Eu* vs. Sr/Y from the Early Permian igneous rocks.
6.3. Geodynamics of the ca. 280 Ma magmatic flare-up

6.3.1. An extensional setting for the ca. 280 Ma magmatic rocks

The ~280 Ma granitoids were thought to have formed at an active continental margin based on geochemical discrimination diagrams (Li, 2006; Shi et al., 2012). However, these diagrams could instead reflect the tectonic settings in which the protoliths were formed, or reflect the melting of mixed sources (e.g., Li et al., 2003). The tectonic setting of the ~280 Ma granitoids can be constrained, however, by the timing of ocean closure between the CAOB and Alxa Block (i.e., Paleo-Asian Ocean), although this timing is itself highly disputed (e.g., Wang et al., 2010b; Wilhem et al., 2012; Xiao et al., 2009, 2010). Numerous Early Permian (ca. 280 Ma) A-type granites have been identified in the western CAOB (Tianshan, Junggar, and Altai) (Tang et al., 2010b; Zhang and Zou, 2013a and references therein) and are commonly attributed to the Permian Tarim mantle plume (e.g., Zhang and Zou, 2013a). This scenario implies that the Paleo-Asian Ocean was probably closed by this time. Recently, the Alxa Block has been invoked as a critical area to constrain the timing of Paleo-Asian Ocean closure, and subduction was proposed to extend into the Early Permian in the northern Alxa Block (Feng et al., 2013; Zhang et al., 2013b). However, this suggestion is not consistent with the spatial distribution and tectonic setting for the Upper Carboniferous–Lower Permian Amushan Formation. The Amushan Formation contains the only sedimentary stratum deposited in the Alxa Block during the Paleozoic. Moreover, the Amushan Formation occurs on both sides of the suture as adopted in this paper (Fig. 1b) but also to the north and south of the two sutures proposed by Feng et al. (2013). This distribution indicates that the tectonic setting of the northern Alxa Block had changed before the formation of the Amushan Formation. Indeed, the mafic rocks from the Amushan Formation were formed in a continental rifting setting (Dang et al., 2011; Jiang et al., 2011). Thus, the ocean represented by the Enger Us Ophiolite must have been closed before the deposition of the Amushan Formation (Late Carboniferous–Early Permian).

Based on this evidence, the ca. 280 Ma magmatic rocks in the Alxa Block are likely to have been formed in a post-orogenic extensional setting. The proposed tectonic setting is also supported by the characteristics of the ca. 280 Ma magmatic rocks themselves. Although the I- and A-type granites and high Sr/Y rocks are very widespread globally, these three types of granitoids rarely occur synchronously together, except in a ridge subduction-related slab window (Tang et al., 2010a, 2012). However, the ca. 280 granitoids have evolved Sr–Nd isotopic compositions that are distinct from the depleted compositions found in their ridge subduction counterparts (Tang et al., 2010a, 2012). Moreover, no spatial or temporal zonation is apparent in the distribution of Alxa Block examples (Fig. 1b). Additionally, none of the magmatic rocks are significantly metamorphosed and no contemporaneous metamorphism occurred in the region. Thus, the ~280 Ma magmatic rocks must have formed in a post-collisional extension or intraplate setting rather than in a ridge subduction environment.

6.3.2. An Early Permian mantle plume trigged the ca. 280 Ma magmatic flare-up?

The wide distribution and lack of zonation exhibited by the ca. 280 Ma Alxa Block magmatic rocks also argue against a slab break-off model (e.g., von Blanckenburg and Davis, 1995; Whalen et al., 2006, 2010). This model predicts a relatively narrow, linear zone of magmatism and uplift that is located along a suture zone. Two models could be proposed to account for intense post-collisional magmatism and the tectonic evolution, including (a) the detachment of an orogenic root zone or lithospheric delamination (e.g., Aydin et al., 2008; Whalen et al., 2006, 2010; Zhang et al., 2007), and (b) a mantle plume...
of the thick (~200 km) lithosphere and thick (45 km) crust also reduces the plausibility of lithospheric delamination. The thick (~200 km) lithosphere and thick (>45 km) crust also reduces the plausibility of lithospheric extension as a mechanism for generating the ca. 280 Ma granitoids.

A mantle plume model does not share the weaknesses of the previously described scenarios (e.g., Bryan et al., 2002; Li et al., 2003). Early Permian igneous rocks are widely distributed in NW China (e.g., Qin et al., 2011; Su et al., 2011; Xia et al., 2012; Xu et al., 2013; Zhang and Zou, 2013a, 2013b; Zhang et al., 2010), including the Tarim, Tianshan and CAOB, and constitute a large igneous province referred to as the Tarim or the Tianshan. A mantle plume has been invoked to explain the characteristics of these rocks (e.g., Qin et al., 2011; Su et al., 2011; Xu et al., 2013; Zhang et al., 2010; Zhou et al., 2004), e.g., widely distributed basaltic rocks (e.g., Zhang et al., 2010), extremely high magma temperatures estimated to be in the range of 1100–1600 °C (Qin et al., 2011), and widely distributed mafic–ultramafic complexes (e.g., Qin et al., 2011). The Early Permian mantle plume was recently proposed to be situated closer to the Beishan Rift than the Eastern Tianshan and the western Tarim (Su et al., 2012). The ca. 280 Ma granitoids in the Alxa Block are coeval with the ca. 280 Ma Tarim mantle plume. Thus, the heat required to generate these crust-derived granitoids was likely supplied by the Tarim mantle plume given that the Alxa Block was probably near its margin. A similar example is the Chon Aike SLIP (188–153 Ma) (Pankhurst et al., 1998, 2000), which is adjacent to the Karoo–Ferrar LP (Pankhurst et al., 2000).

Generating these large volume granitoids probably required the involvement of underplated mafic rocks in the lower crust, similar to other Phanerozoic SLIPs (e.g., Bryan et al., 2002). The rare mafic rocks in the Alxa Block support the possibility of mafic underplating. Moreover, the cumulus characteristics of the gabbro and quartz diorite imply the presence of magma chambers in the lower and upper crust. The main requirement for the generation of these large volume felsic melts is that their protoliths are hydrous (e.g., Bryan et al., 2002), which is consistent with the fact that reported Phanerozoic SLIPs are all restricted to continental margins that acquire fertile, hydrous lower-crustal materials by long-lived subduction (e.g., Bryan and Ferrari, 2013; Bryan et al., 2002). Although there is no contemporaneous subduction around the Alxa silicic igneous province, the hydrous lower-crust would have been created by Paleo-subduction in the southern (e.g., Song et al., 2013), eastern (Dan et al., 2013) and northern (Zheng et al., 2013) part of the Alxa Block.

7. Conclusions

Most of the Phanerozoic igneous rocks in the Alxa Block were formed during a short period (290–250 Ma) with a magmatic flare-up at ca. 280 Ma and their large areal extent is similar to that of other silicic large igneous provinces. The A-type granite, I-type monozonigranites and high Sr/Y granodiorites from the Bayannuoergong batholith were probably generated by partial melting of underplated basalts, by the mixing of lithospheric mantle-derived and lower crust-derived melts and by the remelting of the thickened lower crust, respectively. The gabbros, quartz diorites and diorites from the Bijiertai Complex were likely generated by partial melting of the lithospheric mantle, differentiation from basaltic magmas and partial melting of the middle crust. The ca. 280 Ma magmatic event was probably formed in a post-orogenic extensional setting, and triggered by the adjacent ca. 280 Ma Tarim mantle plume.

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