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Late Cretaceous granitoids in Karakorum, northwest Tibet: petrogenesis and tectonic implications

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ABSTRACT

New zircon LA-ICP-MS U–Pb age, zircon Hf isotope, and whole-rock major and trace elemental data of the Late Cretaceous Ageledaban complex in the Karakorum Terrane (KKT), northwest Tibet, provide new constraints on the tectonic processes of the collision and thickening of the terrane between the Lhasa and Qiangtang terranes. The granitoids from the Ageledaban complex have a variable SiO₂ content, from 62.83 to 73.35 wt.% and A/CNK<1.1 (except for YM61-2). They have rare earth element and trace element patterns that are enriched in light rare earth elements, Rb, Pb, Th, and U, and are depleted in Ba, P, Sr, Ti, and Nb, indicative of weakly peraluminous-metaluminous I-type affinity. Zircon U–Pb dating reveals that the Ageledaban complex was emplaced at ca. 80 Ma. Zircons from the monzogranite and monzonite samples with concordant 206Pb/238U ages about 80 Ma have a zircon εHf(t) of -6.6 to -1.1, corresponding to the Mesoproterozoic Hf crustal model ages (TDM1 = 1.2–1.6 Ga); the remaining inherited zircons from the monzonite with concordant 206Pb/238U ages of about 108.1 Ma have εHf(t) values that range from -8.3 to -5.0, corresponding to the Mesoproterozoic Hf crustal model ages with an average of 1.6 Ga. These signatures indicate that the Ageledaban granitoids may have been derived from the partial melting of a mixed mantle-crust source. Together with the age and geochemical data in the literature, we propose that the collisional event in the KKT in northwestern Tibet would postdate the northern Lhasa–southern Qiangtang collision, which occurred first in the Amdo in the east and later in the Shiquanhe in central Tibet. Our results support the previous view that the collision of the Bangong–Nujiang suture zone (BNSZ) may be diachronous.

1. Introduction

The formation and evolution of the Tibetan Plateau was closely related to the accretion of terrains and landmasses to the southern edge of the Asian continent since the Palaeozoic (Chang and Zheng 1973; Allègre et al. 1984; Dewey et al. 1988; Yin and Harrison 2000; Pan et al. 2012; Jiang et al. 2014). However, the specific collision and thickening events that occurred in the various collision zones have escaped serious scrutiny. For example, in the case of the collision zone between the Lhasa–Qiangtang terranes in the central Tibetan Plateau, most studies suggest that the collision first occurred in the Amdo area in the eastern part during the Late Jurassic before propagating to the west, where the collision is inferred to occur during the late Early Cretaceous or early Late Cretaceous in the Shiquanhe area (Dewey et al. 1988; Matte et al. 1996; Yin and Harrison 2000; Fan et al. 2014; Xu et al. 2014; Chen et al. 2015; Wang et al. 2016; Zhu et al. 2016). Farther to the west, around the Karakorum Terrane (KKT), the timing of the collision remains poorly constrained due to the absence of available data.

The KKT is located on the northwestern edge of the Tibetan Plateau (Figure 1(a)). Extensive Palaeozoic and Mesozoic igneous rocks in this terrane record the complex histories of the opening and closure of both the Palaeo-Asian Ocean and the Tethyan Ocean and the resultant crustal growth (Bi et al. 1999, 1999; Ren 1999; Jiang et al. 2000; Xue et al. 2005; Fan et al. 2014; Xu et al. 2014; Zhang et al. 2016). However, few previous studies considered the Cretaceous history in the region, such that the timing of the collision and accretionary events remains unclear. Wang et al. (2009) reported the early Late Cretaceous granitoids with I-type/S-type affinity and the late Late Cretaceous granitoids with S-type affinity within northwest Tibet, which underwent crustal re-melting and
magma mixing in different degrees. Recently, the Late Cretaceous magmation (76–95 Ma) along the Lhasa–Qiangtang collision zone was also successively found at Rutog (Zhao et al. 2008; Ma 2013), Gaerqiong (Yao et al. 2013), Balazha (Yu et al. 2011; Wang et al. 2013), Minqianri (Ma and Yue 2010), Zhuogapu (Wang et al. 2014), Baingoin (Haider et al. 2013), Amdo (Bai et al. 2009), and Guogencuo (Li et al. 2013). Currently, the existing interpretations on the petrogenesis and tectonomagmatic processes of the Late Cretaceous magmation are mainly in two aspects: (1) northward subduction of the Yalung Zangbo oceanic crust with a low angle induces partial melting of the thickening continental crust (Zhao et al. 2008; Zhang et al. 2012) and (2) partial melting of the lower crust following the delamination (Yu et al. 2011; Li et al. 2013; Wang et al. 2013).

Figure 1. Tectonic framework of the Tibetan Plateau (Zhu et al. 2013) and the distribution of magmatic rocks dated to the Late Cretaceous (marks by ovals with numerals) (a); distribution of the Cretaceous granitoid rocks in northwest Tibet (Wang et al. 2009) (b); sketch geological map of the Ageledaban complex and the sampling locations and Zircon U–Pb ages are also labelled (c). Abbreviations: WKT, West Kunlun Terrane; EKT, East Kunlun Terrane; TST, Tianshuihai Terrane; SP-GZT, Songpan–Ganzi Terrane; KKT, Karakorum Terrane; NQT, Northern Qiangtang Terrane; QT, Southern Qiangtang Terrane; SL, Southern Lhasa subterrane; CL, Central Lhasa subterrane; NL, Northern Lhasa subterrane; KKF, Karakorum Fault; ATF, Altyn Tagh Fault; LSSZ, Longmu Tso–Shuanghu Suture Zone; BNSZ, Bangong–Nujiang Suture Zone; SNMZ, Shiquan River–Nam Tso Mélange Zone; and LMF, Luobadui–Milashan Fault. Literature data from Bai et al. (2009), Cowgill et al. (2003), Kapp et al. (2000, 2005), Li et al. (2007), Li et al. (2013), Li et al. (2014), Ma and Yue (2010), Miller et al. (2000), Schwab et al. (2004), Wang et al. (2007, 2008), Wang et al. (2013), Wang et al. (2014), Yu et al. (2011), Zhao et al. (2008) Ma (2013), Haider et al. (2013), Yao et al. (2013), Zhai et al. (2007), Zhu et al. (2013), 1:250000 geologic reports of Sikadu, Tatulugou, Mazha, and Shenxianwan.
This article reports the zircon U-Pb ages, Hf isotopic data, and whole-rock major and trace elemental data for the Ageledaban complex (Figure 1(a)) in the KKT. This new data set provides insight to a region of the Tibetan Plateau that is insightful in terms of tectonic location and timing. These data and analyses further an evolving understanding in the origins of granitoids within collisional settings.

2. Geological setting and sampling

2.1. Tectonic framework of the Tibetan Plateau

The Tibetan Plateau is bordered by the Tarim Basin to the north and the Himalaya to the south and consists of a series of E-W-trending crustal terranes. From the north to the south, the crustal terranes include the Western Kunlun Terrane (WKT), the Eastern Kunlun Terrane (EKT), the Tianshuihai Terrane (TST), the Songpan–Ganzi Terrane (SPGZT), the KKT, the Qiangtang Terrane, the Kohistan Terrane, and the Lhasa Terrane. These terranes are bounded by five tectonic suture zones and are disrupted by the Altyn Tagh Fault (ATF) with sinistral strike-slip and the Karakorum Fault (KKF) with dextral strike-slip (Jiang et al. 2012, Pan et al. 1994, 1996). The Lhasa Terrane is divided into the Northern Lhasa subterrane (NL), the Central Lhasa subterrane (CL), and the Southern Lhasa subterrane (SL), which are each separated by the Shiquan River–Nam Tso Mélange Zone (SNMZ), and the Luobadui–Milashan Fault (LMF) (Zhu et al. 2011, 2013). The Qiangtang Terrane is divided into the Northern Qiangtang Terrane (NQT) and the Southern Qiangtang Terrane (SQT), which are separated by the Longmu Tso–Shuanghu Suture Zone (LSSZ) (Zhu et al. 2013) (Figure 1(a)).

The Bangong–Nujiang Suture Zone (BNSZ) is thought to have formed by the Mesozoic closure of the Tethyan Ocean between the Lhasa and Qiangtang Terranes (Yin and Harrison 2000, Kapp et al. 2003; Qu et al. 2012; Chen et al. 2004; Wang et al. 2005, 2015; Zhu et al. 2006, 2016; Bao et al. 2007; Chen and Jiang 2002; Zhang et al. 2012; Fan et al. 2014). The BNSZ extends east from Kashmir for over 2000 km towards Bangguonuo, Gerzi, Dongqiao, and Jiayuqiao and continues south towards Burma, Thailand, and Malaysia. Recently, coeval magmatic rocks have also been reported along the BNSZ (Table 1), including volcanic rocks from the Minqianri, Amdo, Guogencuo, and Zhuogapu (ca. 91.2 Ma, Bai et al. 2009; ca. 75.9–79.9 Ma, Ma and Yue 2010; ca. 79.9 Ma, Li et al. 2013; ca. 91 Ma, Wang et al. 2014); porphyries from Balazha (ca. 88 Ma, Yu et al. 2011; ca. 93 Ma, Wang et al. 2013); granitoids from Ageledaban (ca. 94.5 Ma; Bi et al. 1999); and andesitic
porphyrites and rhyolites from Baingoin (ca. 94 Ma; Haider et al. 2013). In addition, abundant zircon xenocrysts of ca. 87 Ma and quartz diorites from the Gaerqiong Cu–Au deposit (Yao et al. 2013) indicate that magmatic activity occurred approximately 87 Ma along the southern margin of the NL (Figure 1(a)). These geochronological data suggest a widespread pulse of magmatism along the Lhasa–Qiangtang collision zone.

2.2. Lithology of the Ageledaban complex

The Cretaceous granitoid rocks including the Ageledaban complex in Karakorum, northwest Tibet, are disrupted by the ATF with sinistral strike–slip and KKF with dextral strike–slip. The Ageledaban complex is located north of Mountain Qiaogeli and east of the conjunction area of Mountain Kongka in the northwestern part of the BNSZ in the KKT (Fig. 1b, 2a). The complex intrudes into the late Palaeozoic Qiater Group. It has an irregular elliptic and strip shape that follows a NW–SE trend, and its outcrop area is approximately 167.1 km² (Figure 1(c)). The direction of the long axis of this complex is generally in agreement with the direction of regional tectonic lineaments (Figure 1(a)).

The Ageledaban complex and its surrounding rock distribute quartz veins, which intruded into along the joint and stratification. In this complex, it is composed of monzogranite, monzonite, tonalite, and granodiorite, and the sampling locations are presented in Figure 1(c). All of the igneous rocks in the study area have a medium-coarse inequigranular granitic texture (Figure 2(b)), with variable degrees of carbonation, sericitization, and chloritization. The monzogranite, tonalite, and...
granodiorite contain plagioclase, K-feldspar, biotite, quartz, and hornblende, with accessory zircon, apatite, magnetite, titanite, and so on (Figures 2(c, e, and f)). The plagioclase grains are characterized by variable degrees of oscillatory zoning. The monzonite contains plagioclase and K-feldspar, with accessory opaque mineral, such as magnetite (Figure 2(d)). In addition, the monzonite is filled with quartz veins occurring as an interstitial phase between mineral grains.

3. Analytical methods

3.1. Whole-rock geochemical analysis

Whole-rock major element concentrations were analysed using an X-ray fluorescence wavelength-dispersive spectrometry (PANalytical Axios X) at the Southwest Institute of Metallurgical Geological Measurement (Sichuan). The analytical precision of the major element concentrations was better than 5%. For FeO, H2O^+ and CO2 were determined by standard wet chemical techniques.

Trace element analyses were conducted using an ICP-MS (Finnigan Element 2) at the State Key Laboratory for Mineral Deposits Research, Nanjing University. Approximately 50 mg of powdered sample was dissolved in high-pressure Teflon bombs using a HF+HNO3 mixture. Rh was used as an internal standard to monitor signal drift during counting in ICP-MS analysis. The USGS rock standards GSP-1 and AGV-2 were chosen for calibrating element concentrations of the measured samples. The relative standard deviation was lower than 5%. The detailed analytical conditions and procedures for the elements are similar to those described by Gao et al. (2003). The whole-rock compositional data are listed in Supp. Table 1.

3.2. LA-ICP-MS Zircon U–Pb dating

One monzogranite sample (D3219) and one monzonite sample (YM61-1) were selected for zircon U–Pb and Hf isotope analyses. The zircon was separated by heavy-liquid and magnetic methods. Cathodoluminescence (CL) images were taken at the Institute of Mineral Resources, Chinese Academy of Geological Sciences (Beijing) to analyse the internal structures of individual zircon samples and to select the appropriate spots for the zircon isotope analyses.

Zircon U–Pb dating analyses were conducted using an LA-MC-ICP-MS at the Institute of Mineral Resources, Chinese Academy of Geological Sciences (Beijing). The detailed operating conditions of the laser ablation system and the MC-ICP-MS instrument and data reduction are the same as described by Hou et al. (2009). Laser sampling was performed using a Newwave UP 213 laser-ablation system; mass populations were measured with a Thermo Finnigan Neptune MC-ICP-MS instrument. Single-point denudation ablation used a laser spot with a diameter of 25 μm, an energy density of 2.5 J/cm^2, and a repetition rate of 10 Hz, applying helium as a carrier gas. We used reference zircon GJ1 as the external U–Pb dating standard, reference zircon M127 (U: 923 ppm, Th: 439 ppm, Th/U: 0.475) as the external U and Th concentration standard (Nasdala et al. 2008), and the Plešovice reference zircon as an internal standard. During the analysis, after measuring every 8–10 specimens on multiple grains, two GJ1 zircons were measured in succession, followed by one Plešovice zircon to ensure the accuracy of the measurement. The ICPMSDataCal software was applied to process the collected data (Liu et al. 2010), applying the Andersen (2002) method of common-Pb correction. The concordia diagrams of the zircon data were obtained using the Isoplot 3.0 software (Ludwig 2003). The zircon U–Pb isotopic data are summarized in Supp. Table 2.

3.3. Zircon Hf isotopic analysis

Hf isotopic measurements were conducted on the same U–Pb age spots or as close as possible within the same CL zonation. The zircons were analysed on a Neptune multiple collector MC-ICP-MS combined with an ArF excimer laser ablation system (New Wave Research) at the Tianjin Institute of Geology and Mineral Resources using techniques and analytical procedures described by Cheong et al. (2013). During sample analysis, after measuring every five to seven specimens on multiple grains, two GJ1 zircons were measured in succession. GJ1 standard zircons yielded an average ^176Hf/^177Hf ratio of 0.282006 ± 24 (n = 159, 2 SD). The initial epsilon Hf value was calculated using a ^176Lu decay constant of 1.865 × 10^-11 y^-1 (Scherer et al. 2001) and chondritic values suggested by Blichert-Toft and Albarède (1997). The Hf model age (TDM^Hf) was calculated based on the depleted mantle model described by Griffin et al. (2000). The two-stage model age (TDM^SF^Hf) was calculated assuming an average continental crust (^176Lu/^177Hf = 0.0116; Rudnick and Gao 2003) that was originally derived from the depleted mantle. All of the errors in this study were quoted at one sigma standard deviation, unless otherwise started. The Lu–Hf isotopic data of the zircons are listed in Supp. Table 3.

4. Results

4.1. Whole-rock geochemistry

The Ageledaban complex consists of monzogranite, monzonite, tonalite, and granodiorite. These granitoids
have a variable SiO₂ content from 62.83 to 73.35 wt.% (Supp. Table 1), and these values were plotted in the field of monzogranite, quartz monzonite, tonalite, and granodiorite on a total alkalis (K₂O+Na₂O) versus SiO₂ (TAS) diagram. However, there is an inconsistency between the TAS classification and rock slice identification that the monzonites named by rock slice identification fall into the area of quartz monzonite, which is likely due to the existence of quartz veins formed from unrelated fluids that can be seen from the rock slices of the monzonites (Figure 3(a)). The Ageledaban intrusives are weakly peraluminous- metaluminous containing A/CNK (= molar Al₂O₃/(CaO+Na₂O+K₂O)) values of <1.1 with the exception of sample YM61-2, which has a value of ~1.2 (Figure 3(b) and Supp. Table 1). These granitoids have high K₂O abundances with data mainly plotting in the field of the calc-alkalic to alkali-calcic series in a (Na₂O+K₂O-CaO) versus SiO₂ diagram (Figure 3(d)). On SiO₂-variation diagrams, the samples from the Ageledaban pluton define curve trends for Al₂O₃, TFeO, CaO, MgO, P₂O₅, and TiO₂ that are consistent with the fractionation of feric minerals, Ti-Fe oxide, apatite, and others.

Figure 4 illustrates all of the samples are enriched in light rare earth elements (LREEs), nearly flat in heavy rare earth elements (HREEs), inconspicuous depleted in middle rare earth elements (MREEs), and negative anomalies in Eu (except for D80 and D81). The distribution patterns of the rare earth elements (REEs) have a similar pattern in the samples, which are seen in the aspect of the fractionation in LREEs and HREEs (La₉/ Yb₉ = 5.60–76.2), and the REE pattern deviates to the right (Figure 4). The REEs have the same fractionation trend as the major elements, which indicate fractionation of the minerals that are enriched in LREEs (e.g. apatite). In a primitive mantle-normalized trace element diagram (Figure 4), the Ageledaban complex is clearly enriched in Rb, Pb, Th, and U, and depleted in Ba, P, Sr, Ti, and Nb. Compared with Rb and Th, Ba is depleted. Moreover, they exhibit variable Sr/Y ratios (2.47–56.1) and Yb content (1.44–9.84 ppm), and have low abundances of compatible trace elements, e.g. Cr = 9.21–26.6 ppm; Ni = 3.54–17.6 ppm (Supp. Table 1).

### 4.2. Zircon U–Pb dating

The zircon grains from the monzogranite (D3219) are mostly euhedral or subidiomorphic, with long or short columnar forms (60–200 μm long, 70–110 μm wide) and an aspect ratio of 1.1:1 to 4.2:1 (Figure 5(a)).
Zircons in the monzonite (YM61-1) exhibit euhedral or subidiomorphic (80–150 μm long, 50–130 μm wide) morphology with an aspect ratio of 1.2:1 to 3.5:1, but some of the grains have inherited cores (Figure 5(c)). All of the zircon grains exhibit oscillatory zoning.

The analysed zircons from the monzogranite and monzonite have variable U (129–3,615 ppm) and Th (136–985.6 ppm) concentrations, with a Th/U ratio ranging from 0.12 to 1.19 (Supp. Table 2), consistent with a Th/U ratio of magmatic origin (Hoskin and Schaltegger 2003; Griffin et al. 2004). Seventeen...
individual spots from the monzogranite (D3219) yield concordant U–Pb ages ranging from 77.3 to 78.6 Ma, with a weighted mean of 78.0 ± 0.3 Ma (MSWD = 0.4) (Figure 5(b)). Figure 5(d) displays the U–Pb concordia diagram of the monzonite (YM61-1); the zircon grains from this monzonite yield concordant U–Pb ages from 79.4 to 81.2 Ma (12 spots) with a weighted mean of 80.4 ± 0.4 Ma (MSWD = 0.86), which excludes seven spots with inherited ages (108.1 ± 0.5 Ma). The results suggest the monzogranite intruded at 78.0 ± 0.3 Ma and the monzonite intruded at 80.4 ± 0.4 Ma.

4.3. Zircon Hf isotopes

Seventeen spots on the zircons from the monzogranite (D3219) have negative εHf(t) values (from −6.6 to −2.8), corresponding to Mesoproterozoic Hf crustal model ages (TDMC = 1.3–1.6 Ga, Supp. Table 3). These zircons from the monzonite (YM61-1) with concordant U–Pb ages from 79.4 to 81.2 Ma (12 spots) have εHf(t) values that range from −5.7 to −1.1 (Supp. Table 3), and the TDMHf values are 1.2–1.5 Ga; the remaining inherited zircons (seven spots) from the monzonite (YM61-1) with concordant U–Pb ages about 108.1 Ma have εHf(t) values that range from −8.3 to −5.0 (Supp. Table 3), and the average TDMHf values are 1.6 Ga (Supp. Table 3). These results suggest that monzonite (YM61-1) is derived from a Mesoproterozoic crustal source. The limits of variation of the εHf(t) value in monzogranite (D3219) and monzonite (YM61-1) are 3.8 and 7.2, respectively, which suggests that the samples should exhibit an inhomogeneous zircon Hf isotopic composition. Accordingly, the samples have a wider range of Hf isotope-crust model ages (Figure 6).

5. Discussion

5.1. Magmatism in the Late Cretaceous along the Lhasa–Qiangtang collision zone

The LA-ICP-MS U–Pb zircon geochronological data suggest that the Ageledaban complex formed ca. 80.0 Ma, which is clearly younger than 94.5 Ma by a biotite K–Ar age for the Ageledaban granodiorite (Figure 1(b); Bi et al. 1999). The biotite in the granodiorite shows chloritization along its joint and edge (Figure 2(f)). In addition, the Ageledaban complex is a complex pluton and experiences magmatism in two times, the later magmatism may affect the biotite that formed at an earlier stage. Hence, the chloritization of the biotite and/or the second magmatism may destroy part of the crystal lattice of the biotite and affect the reliability of age. However, zircons have rich U and Th, low common Pb, and very high stability in the mineral, making the Zircon U–Pb dating become one of the most frequently used and most effective methods in isotope geochronological research (Wu and Zheng 2004).

Geochemically, the ca. 88 Ma Balazha porphyries, the ca. 88 Ma Gaerqiong quartz diorites, and the ca. 94 Ma Baingoin andesitic porphyrites and rhyolites to the east of the ca. ~80.0 Ma Ageledaban granitoids have a geochemical affinity with adakitic rocks (Yu et al. 2011; Haider et al. 2013; Yao et al. 2013) that is characterized by a high Sr/Y ratio (26.6 in average), low abundance of Yb (1.0 × 10⁻⁶ in average), and a high, variable abundance of compatible trace elements (Cr: 107 × 10⁻⁶ in average; Ni: 13 × 10⁻⁶ in average). The abundance of ca. 83.1 Ma Rutog granitoids in the NL have a positive zircon εHf(t) value from 6.3 to 11.9 (Zhao et al. 2008; Ma 2013). However, the abundance of ca. 75.9–79.9 Ma Guogencuo volcanic rocks in the southern Qiangtang subterrane has a negative zircon εHf(t) ranging from −5.8 to −2.1 (Li et al. 2013). In summary, the currently

Figure 6. Histograms of the εHf(t) and TDMC of zircons for monzonite and monzogranite samples from the Ageledaban complex in the Karakorum Terrane.
available geochronological data reveal that Late Cretaceous magmatism occurred along the Lhasa–Qiangtang collision zone, resulting in the formation of a diverse array of rocks and these rocks include Mg-rich andesites and dacites, basaltic rocks, adakite-like porphyries, and adakitic andesitic porphyrites (Table 1).

5.2. Petrogenesis of the Ageledaban complex

Granitic rocks display great diversity, the source of which, even today, is still one of the most debated subjects in granite petrogenesis. The consensus view is that granites are typically generated during periods of heat and/or mass transfer from mantle to crust, by partial melting of crustal rocks or by fractional crystallization or assimilation combined with fractional crystallization of mantle-derived basaltic magmas (DePaolo 1981; Holden et al. 1987; Collins 1998; Snyder and Tait 1998).

The Ageledaban complex is weakly peraluminous-metaluminous with the exception of sample YM61-2, having A/CNK values of <1.1, which is characteristic of I-type granitoids (Chappell and White 1992, 2001; Chappell 1999). Furthermore, all of the samples contain abundant amphibole and do not contain typical peraluminous minerals such as muscovite, garnet, or cordierite. It is also shown that P₂O₅ decreases with increasing SiO₂ content and Th increases with increasing Rb. This feature is considered to be an important criterion to distinguish I-type from S-type granite by Chappell and White (1992). In addition, the zircon saturation temperatures of the granitic magma are estimated between 73°C and 77°C with an average of 75°C (Supp. Table 1, Watson and Harrison 1983), which is lower than that of A-type granite (King et al. 1997, 2001).

In general, A-type granites contain alkaline dark minerals such as arfvedsonite and riebeckite, are comparatively rich in HFSEs, such as Zr, Nb, and Y, and strongly depleted in Ba, Sr, P, Ti, and Eu. The Ageledaban granitoid is free of mafic alkaline minerals, not rich in Nb content, and strongly depleted in Eu and HREEs, which suggests that it is not an A-type granite. The lack of high-temperature anhydrous phases (e.g. pyroxene and fayalite) and late-crystallizing biotite and amphibole is also inconsistent with A-type granite characteristics (Huang et al. 2008, 2013). All of these characteristics suggest the Ageledaban complex is I-type granite (Figures 3(c,d)).

The Ageledaban complex has negative εHf(t) values ranging from −8.3 to −1.1, with TDMHf values of 1.2–1.6 Ga (Supp. Table 3). In addition, the average model age of the seven inherited zircon crustal units (1.6 Ga) was older than that of the magmatic zircons (Figure 7).

Figure 7. Zircon εHf(t) versus t plot of monzonite and monzogranite samples from the Ageledaban complex in the Karakorum Terrane.

These zircon Hf isotopic data indicate that the granitoids from the Ageledaban complex were derived from the partial melting of the Mesoproterozoic lower crust. However, the limit of εHf(t) value variations is heterogeneous (Figure 6). The inhomogeneous phenomenon of the zircon Hf isotopic data is similar to data from other regions around the world (Griffin et al. Griffin et al. 2002; Kemp et al. 2007; Bolhar et al. 2008), which implies that an open system existed and led to the variable ¹⁷⁶Hf/¹⁷⁷Hf value (Kemp et al. 2007). The zircon ¹⁷⁶Hf/¹⁷⁷Hf value cannot be changed by partial melting or fractional crystallization. Thus, the inhomogeneous zircon Hf isotopic data likely signifies the interaction of mantle radiogenic and crust with less radioactive factors (Bolhar et al. 2008). On the MgO–TFeO diagram shown in Figure 8, the data from the Ageledaban samples are distributed towards the region of magma mixing (Zorpi et al. 1991). Wang et al. (2009) discovered the emplacement of abundant dioritic

Figure 8. Magma mixing trend diagram by MgO versus TFeO for the Ageledaban complex samples (Zorpi et al. 1991).
enclaves in the Ageledaban complex, and the appearance of dark enclaves is direct evidence of magma mixing (Kemp 2004; Janoušek et al. 2004a; Chen et al. 2007; Ma et al. 2012). Therefore, these results hint that the Ageledaban granitoid melts were derived from a mixed mantle-crust source.

The data from the Ageledaban samples in Figures 8 and 9 do not support an obvious trend of fractional crystallization. However, the granitoids in SiO₂-oxide variation affinity and chondrite-normalized REE pattern (Figure 4(a)) follow very nice trends from tonalite/granodiorite to monzogranite that look like the tail end of fractionation processes. The formation processes of the Ageledaban complex could be interpreted to be the result of a MASH process generally observed above the subduction zones (Hildreth and Moorbath 1988).

In summary, this MASH process could be described as follows (Figure 10(b)). (1) The partial melting of

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**Figure 9.** Plots of Sr versus Rb and Ba, Eu* versus Sr and Ba for the Ageledaban complex samples according to Janoušek et al. (2004b). Pl, plagioclase; Kf, K-feldspar; Bt, biotite; Ms, muscovite; Grt, garnet; Amp, amphibole.

**Figure 10.** Schematic illustrations of the evolution of the BNZS, modified from Wang et al. (2016). L, Lhasa terrane; KKT, Karakorum Terrane; SQT, South Qiangtang subterrane; NQT, North Qiangtang subterrane; B, Baoshan terrane; YZ, Yangtze platform.
mantle wedge peridotite that was modified by slab-derived melts and fluids to generate basaltic magmas (Kepezhinskas et al. 1996), and then the magmas stagnate at the magma chamber at or near the Moho (Richards 2011) or hot zone (Annen et al. 2006) and experience fractional crystallization (Crawford et al. 1987). (2) The mantle-derived magmas stagnating at the magma chamber afford the heat and material for the partial melting of the Mesoproterozoic lower mafic crust (Annen and Sparks 2002), which resulted in the development of calc-alkaline I-type granitoids with isotopic homogenization. These granitoid melts move upward and finally pond at shallow magma chambers and experience slight or significant fractional crystallization of plagioclase and K-feldspar to generate normal calc-alkaline granitoids. (3) Isotopically homogenized doric magmas are developed owing to magma mixing between mantle-derived magmas and Mesoproterozoic lower mafic crust-derived granitic melts within the magma chamber at the hot zone. The compositional variations observed in these rocks were mainly controlled by their source compositions and the degree of partial melting.

5.3. Tectonic interpretation


Together with abundant Late Cretaceous geochronological and geochemical data, the ca. 88 Ma Balazha porphyry, the ca. 88 Ma Gaerqiong quartz diorite, and the ca. 94 Ma Baingoin andesite porphyrite and rhyolite have the geochemical signatures of adakite (Table 1), and suggest that crustal thickening in the east and middle of the BNSZ occurred before magmatism. In contrast, the Ageledaban complex in the KKT does not have the characteristics of adakite. These results suggest that the crust had thickened in the eastern and middle of the Lhasa–Qiangtang collision zone in the Late Cretaceous, but the study region in the KKT did not thicken (Figure 10(a)). Hence, the collisional event that occurred in the KKT in northwestern Tibet would postdate the northern Lhasa–southern Qiangtang collision in central Tibet. Our results support the previous view (Yin and Harrison 2000; Pan et al. 2006; Kapp et al. 2007; Zhu et al. 2013) that the BNSZ may have diachronously closed, and provided insight to the KKT in northwestern Tibet, which is insightful in terms of tectonic location and timing.

6. Conclusions

(1) The LA-ICP-MS zircon U–Pb dating reveals that the Ageledaban complex formed ca. 80.0 Ma in the KKT, northwest Tibet. Compared with the magmatic rocks along the Lhasa–Qiangtang collision zone, the emplacement of the Ageledaban complex occurred coevally.

(2) The Ageledaban granitoids are weakly peraluminous-metaluminous l-type granitoids that could be interpreted to be a result of a MASH process generally observed above subduction zones.

(3) The collisional event in the KKT in northwestern Tibet postdates the northern Lhasa–southern Qiangtang collision, which occurred first in the Amdo in the east and later in the Shiquanhe in central Tibet.

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