Geochronology and geochemistry of the Chuanwulu complex in the South Tianshan, western Xinjiang, NW China: Implications for petrogenesis and Phanerozoic continental growth

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Abstract

The South Tianshan Collisional Belt (STCB), formed by the collision between the Tarim and Central Tianshan blocks, is pivotal for understanding the complex and prolonged tectonic evolution of the Central Asian Orogenic Belt (CAOB). The Chuanwulu complex is located in the eastern part of the STCB, Xinjiang province, NW China. It consists of a gabbro-diorite unit (unit I) and a monzonite-syenite unit (unit II), which are both cut by some granite dykes (unit III). Based on LA-ICP-MS U-Pb zircon dating, the units I and II were emplaced at 287.8 ± 4.3 Ma and 286.4 ± 2.5 Ma, respectively. The SiO₂ contents of the samples from the complex display a wide range from 50.52 wt.% to 70.64 wt.%, and most samples are of alkaline affinities. The magma mixing process during the formation of the complex is suggested by considerable petrographic and geochemical evidence such as mafic microgranular enclaves (MEEs) that occur within unit II, disequilibrium textures, linear negative correlations between SiO₂ and some major elements, and a wide range of (87Sr/86Sr)t ratios and εNd(t) values. Unit I is characterized by relatively low SiO₂ contents (50.52–55.05 wt.%), high MgO contents (3.29–5.43 wt.% with Mg# = 0.50–0.56), high Sr contents (1646–3101 ppm), pronounced negative Nb-Ta anomalies and light rare earth element (LREE) enrichment. In combination with their isotopic compositions [(87Sr/86Sr)t = 0.70543–0.70751, εNd(t) = −2.3 to −1.8, δ18O SMOW = 5.7–9.4‰, εNd(t) = 0.2–4.9], these features indicate that they are derived from an incompatible element-enriched lithospheric mantle. The SiO₂ contents of units II and III range from 58.88 wt.% to 65.91 wt.% and 67.10 wt.% to 70.64 wt.%, respectively. They exhibit pronounced positive Zr-Hf anomalies and negative P and Ti anomalies, and relatively elevated Sr and O isotopic compositions [(87Sr/86Sr)t = 0.70543–0.70751, δ18O SMOW = 8.9–9.1‰]. In addition, samples of unit III have high Sr (600 ppm to 1201 ppm) contents and Sr/Yb (373 to 1905) and low Y/Yb (8 to 12) ratios. The felsic rocks (units II and III) have a “C-type” adakite-like geochemical signature and are comparable with those of rocks derived from an ancient garnet-bearing amphibolite facies lower crust. Our study, compiled with other geological evidence, indicates that the collision between the Central Tianshan and Tarim blocks and the final amalgamation of the CAOB should have occurred during Late Carboniferous. On the regional scale, we propose that the recycling of ancient lithosphere is the predominant mechanism which is able to account for Phanerozoic continental evolution in STCB, differing from other tectonic units in CAOB.

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collision between them. Almost all paleogeographic studies put the Tarim block as the last block to dock to the huge accretionary system of the CAOB, which terminated the Altai accretionary orogeny (Xiao et al., 2010, and references therein). Thus, the study on the STCB is pivotal to understanding the tectonic evolution of the CAOB. In recent years, considerable efforts have been made to unravel the enigmatic tectonic history and metallogenesis of the STCB (Gao et al., 2011, and references therein). However, the nature and tectonic evolution of the STCB is still debated, and some contrasting tectonic models have been proposed. The controversy chiefly focuses on the time of the collision. The collision of the Central Tianshan and Tarim blocks is traditionally thought to have occurred during the Late Paleozoic (e.g. Gao et al., 1998, 2011, and references therein). However, a Triassic collision, mainly based on recent SHRIMP U–Pb ages of 233–226 Ma obtained for zircon rims separated from eclogites in NW China (Zhang et al., 2007a, 2007b), was also proposed. Moreover, since most of the South Tianshan igneous rocks are characterized by "crust-like" signatures, such as higher initial 87Sr/86Sr ratios and negative εNd (t) values, whether the STCB was affected by the injections and/or recycling of large amount of the juvenile and/or mantle materials during Phanerozoic, which were common in other parts of the CAOB (Jahn et al., 2000a, 2000b), is still an open question (Gao et al., 2011; Jiang et al., 1999, 2004).

The Chuanwulu complex is located in the Chinese section of the STCB, southern Xinjiang (Northwest China). The complex, consisting of mafic and felsic rocks, intrudes a fragment of a Paleozoic ophiolitic mélangé. Thus, the special scenario provides a rare opportunity to constrain the time of the collision and the mechanism of continental growth during the late Paleozoic. In this paper, we report the ICP-MS zircon U–Pb ages of 233±4 Ma and some younger U–Pb ages (e.g., 233±4 Ma, 233±3 Ma, 233±4.6 Ma) have been inter-related high-pressure/low-temperature metamorphism is also preserved (Gao et al., 1998; Zheng et al., 2006). Subduction- and/or collision-related high-pressure/low-temperature metamorphism is also preserved (Gao et al., 1998; Zheng et al., 2006). It is mainly composed of blueschist-, eclogite- and greenschist-facies meta-sedimentary rocks and some mafic metavolcanic rocks with N-MORB, E-MORB, OIB and arc basalt affinities (Gao et al., 1998, 2011; Zhang et al., 2007a, b). Most metamorphic zircon ages of those rocks were reported to be Carboniferous (Gao et al., 2011, and references therein). However, some younger U–Pb ages (e.g., 233 ± 4–226 ± 4.6 Ma) have been interpreted to represent the collision time (Zhang et al., 2007a, 2007b).

The basement of the Chinese part of STCB is represented by strongly deformed and metamorphosed rocks of the Paleo-Proterozoic Xingditagh Formation (Dong et al., 2011). It is covered

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**Fig. 1.** (a) Tectonic sketch map of the Central Asia Orogenic Belt showing the location of Tianshan Mountains, and (b) the geological map of the South Tianshan Collisional Belt (modified from Gao et al., 2011).
by the metamorphic Mid-Proterozoic Akesu Formation which is composed predominantly of gneisses and amphibolites. Lower to Middle Paleozoic chert, limestone, and flysch mainly outcrop along the SCTS. Upper Paleozoic limestone, sandstone, and shale with minor volcanic rocks, are widespread in the belt. It is noticeable that the Permian terrestrial bimodal volcanic strata (Xiaokantilike Formation), which crops out in the central part of the belt, unconformably overlies the upper Carboniferous marine carbonate rocks (Kangkelin Formation).

The outcrops of late Paleozoic intrusive rocks make up ~5% of the total area of the STOB. Five pulses of magmatic activity in the STOB have been identified by previous studies (Zhang et al., 2009, and references therein): (1) Neoproterozoic, (2) Early Paleozoic, (3) Middle Paleozoic, (4) Late Carboniferous to Early Permian and (5) Cenozoic (mafic dykes). However, most of the igneous rocks were formed in the fourth pulse, i.e., late Carboniferous to early Permian.

3. Geology and petrography of the Chuanwulu complex

The Chuanwulu complex is situated in the western part of the Chinese STOB, about 15 km from the border between China and Kyrgyzstan. The late Devonian to early Carboniferous Tuyoukeaxiu Formation (TYA Fm.), a part of Kuokesayan ophiolitic mélangé, occupies most of the area of the study region (Fig. 2a). According to the rock type, the TYA Fm. is subdivided into three members from base to top.

Member 1 consists of beige-sage green basic volcanic and clastic rocks, and is conformably overlain by member 2, which is intruded by the Chuanwulu complex and is composed of interbedded silty slates, feldspathic-quartz sandstones and detrital feldspathic sandstones. Above member 2, member 3 is composed of a suite of cherts interlayered with silicified slate. The Chuanwulu complex occupies an area of ~7 km² and is oblong in a WNW–ESE direction (Fig. 2b). From the margin to center, the complex consists of s gabbro–diorite unit (unit I) and a monzonite–syenite unit (unit II). Granite dykes (unit III) are abundant and locally cut the rocks of unit I and unit II, and country rocks.

Unit I has sharp contacts with unit II. The outcrops of unit I comprise <20% of the complex, but drilling data implies that some amount of gabbro is probably buried by the Cenozoic sediments. This unit predominantly consists of fine- to coarse-grained gabbro, biotite gabbro, gabbroic diorite and biotite diorite. The different types of rocks gradually vary mineralogically, and they are composed of plagioclase (60–70 vol.%), K-feldspar (5–10 vol.%), clinopyroxene (0–15 vol.%), hornblende (2–10 vol.%), biotite (2–5 vol.%), and quartz (0–5 vol.%). Accessory minerals, including magnetite, ilmenite and apatite, are common in this unit. Some apatite with acicular shapes reflects the fast quenching process of the magma (Fig. 3c).

Unit II crops out in the inner zone of the complex and makes up about >70% of the complex. Xenoliths of unit I and country rock are common in the margin of this unit, suggesting this unit was emplaced

Fig. 2. (a) Sketch map showing the geology of the study area. (b) Geological map of the Chuanwulu complex.
after unit I (Fig. 3a). Unit II consists of coarse-grained monzonite, biotite monzonite, biotite adamellite and syenite. The different types of rocks gradually vary in mineral assemblages and chemical compositions, suggesting they formed nearly simultaneously. This unit predominantly consists of plagioclase (30–60 vol.%), K-feldspar (10–40 vol.%), hornblende (5–10 vol.%), biotite (5–10 vol.%), quartz (3–10 vol.%) and accessory minerals including magnetite, ilmenite, titanite, zircon and apatite. Some K-feldspar is partly altered to kaolinite and is replaced by secondary muscovite. Poikilitic texture characterized by several anhedral quartz grains enclosed by subhedral K-feldspar grains is also observed in some thin sections. Mafic microgranular enclaves (MMEs) are widespread within unit II, mainly concentrated near the boundary with unit I (Fig. 3b). Compared to the host rocks, the MMEs are more fine-grained and dioritic gabbro, diorite and quartz diorite in composition. Some MMEs contain large, rounded phenocrysts, which are chemically similar to those in the host rocks (Fig. 3d).

Unit III comprises <10% of the complex and consists of granite with porphyric (or porphyritic?) texture. The phenocryst phases include K-feldspar, plagioclase, biotite and quartz, occupying 20 vol.% of the granite in thin sections. The groundmass is composed of K-feldspar, biotite, quartz and accessory minerals such as magnetite, ilmenite, titanite, zircon and apatite.

4. Analytical methods

4.1. Zircon U–Pb and Lu–Hf isotopic data

One biotite diorite sample (DYCWL-4) and one biotite monzonite sample (DWCWL-3) are chosen for age determination by LA-ICP-MS U–Pb and Lu–Hf zircon methods. The two samples are actually the same with samples CWL-30 and CWL-7, respectively. Both of them were carefully examined through polarizing microscope to ensure the freshness and the quantity of zircons. Zircons were separated using standard heavy mineral techniques, and were mounted in epoxy blocks for analysis. Internal structures of the zircon grains were examined using transmitted electron, backscattered electron (BSE) and cathode luminescence (CL) prior to U–Pb isotopic analyses. These images have been used to identify different stages of zircon growth, and to select the positions for LA-ICP-MS analyses (Fig. 4). The U–Pb isotopic analyses for samples were obtained with an Elan 6100 DRC ICP-MS equipped with 193 nm Excimer lasers, housed at the Department of Geology, Northwest University, Xi’an, China. Zircon 91500 was used as a standard and NIST 610 was used to optimize the machine. A mean age of 1062 ± 6 Ma was obtained for the 91500 zircon standard. The spot diameter was 30 μm. Corrections for common-Pb were made using the method of Andersen et al. (2002). Data were processed using the GLITTER and ISOPLOT (Ludwig, 2003) programs. Errors on individual analyses by LA-ICP-MS are quoted at the 95% (1σ) confidence level. The details of the analytical procedures have been described by Yuan et al. (2004).

Zircon Lu–Hf isotopic analysis was carried out in-situ using a New-wave UP213 laser-ablation microprobe, attached to a Neptune multicollector ICP-MS at Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing. Instrumental conditions and data acquisition were comprehensively described by Hou et al. (2007) and Wu et al. (2006). A stationary spot was used for the present analyses, with a beam diameter of either 40 μm or 55 μm depending on the size of ablated domains. Helium was used as carrier gas to transport the ablated sample from the laser-ablation cell to the ICP-MS torch via a mixing chamber mixed with Argon. In order to correct the isobaric interferences of 176Lu and 176Yb on 176Hf, 176Lu/177Lu = 0.02658 and 176Yb/177Yb=0.796218 ratios were determined (Chu et al., 2002). For instrumental mass bias correction Yb isotope ratios were normalized to 172Yb/171Yb of 1.35274 (Chu et al., 2002) and Hf isotope ratios to 176Hf/177Hf of 0.7325 using an exponential law. The mass bias behavior of Lu was assumed to follow that of Yb. Zircon CJ1 was used as the reference standard, with a weighted mean 176Hf/177Hf ratio of 0.282013 ± 0.00008 (2σ; n = 10) or 0.282013 ± 0.000024 (2σ; n = 10) during our routine analyses. It is not distinguishable from a
weighted mean $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282013 ± 19 (2σ) from in-situ analysis by Elhlou et al. (2006).

4.2. Major and trace elements

After petrographic examination, 20 fresh samples from the complex were crushed and powdered in an agate mill for geochemical analysis. Nine of the twenty samples are from unit I, seven samples are from unit II and the other four samples are from unit III. Major elements were determined by X-ray fluorescence (XRF) in the Key Laboratory of Orogenic Belts and Crustal Evolution, Ministry of Education, School of Earth and Space Sciences, Peking University, with an analytical uncertainties ranging from 1 to 3%. Trace (including rare earth) element analyses were conducted in the Institute of Geology and Geophysics Chinese Academy of Sciences in Beijing, China, determined by inductively coupled plasma mass spectrometry (ICP-MS). For trace element determination, about 50 mg of powder was dissolved for about 7 days at ca. 100 °C using HF–HNO$_3$ (10:1) mixtures in screw-top Teflon beakers, followed by evaporation to dryness. The material was dissolved in 7 N HNO$_3$ and taken to incipient dryness again, and then was re-dissolved in 2% HNO$_3$ to a sample/solution weight ratio of 1:1000. The analytical errors vary from 5 to 10% depending on the concentration of any given element. An internal standard was used for monitoring drift during analysis; further details have been given by Gao et al. (2008).

4.3. Sr, Nd and O isotope analysis

The isotope ratios of Nd and Sr and associated isotope–dilution concentrations were measured at the Central of Modern Analysis, Nanjing University, China. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were determined on a VG354 mass spectrometer, and isotopic ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd}$ = 0.7219 and $^{86}\text{Sr}/^{88}\text{Sr}$ = 0.1194. Repeated analyses of standards yielded averages of 0.710236 ± 0.000007 (2σ, n = 6) for Sr standard NISTSRM987, and 0.51184 ± 0.000003 (2σ, n = 6) for the La Jolla Nd standard. Total chemical blanks were <200 pg for Sr and <100 pg for Nd.

Oxygen isotope analyses on whole-rock samples were analyzed at the Institute of Mineral Resources, Chinese Academy of Geological Sciences, using the BrF5 procedure according to Clayton and Mayeda (1963) and a MAT251 EM Mass Spectrometer. The precision and accuracy are better than ±0.2/million.

5. Results

5.1. Geochronology

Most of the zircon grains from the Chuanwulu complex are colorless and transparent prismatic crystals containing some micro inclusions and tiny fractures. As a result of the physical deterioration in the zircon separation process, most zircons do not exhibit prismatic crystals and magmatic growth zonings are not clear. However, those zircon crystals have Th/U ratios ranging from 0.6 to 3.0 (Table 1), indicating they crystallized from magma (Yuan et al., 2004). Thirteen zircons from one biotite diorite sample (DYCWL-4) were analyzed. They are characterized by relatively low U (145–571 ppm) and high Th (62–761 ppm) contents and Th/U ratios of 0.6–1.6. The 206Pb/238U ages range from 282.4 to 299.0 Ma, forming a coherent group with a weighted mean 206Pb/238U age of 286.4 ± 2.5 Ma (MSWD = 0.56, 2σ) (Fig. 5b). The two weighted mean 206Pb/238U ages are interpreted as the crystallization ages of unit I and unit II, indicating they were emplaced nearly synchronously.

5.2. Zircon Lu–Hf isotope

Fifty-five Lu–Hf analyses were obtained from 55 representative zircon grains from the Chuanwulu complex. The results are given in Table 2 and illustrated in Fig. 6. The $^{176}\text{Lu}/^{177}\text{Hf}$ ratios range from 0.000372 to 0.004535 with a mean of 0.001095, indicating that
these zircons are very weak in radiogenic Hf. Twenty-five analyses were performed on 25 zircon grains from sample DYCWL-4. The analyses yield variable $^{176}$Hf/$^{177}$Hf ratios between 0.282576 and 0.282737, corresponding $\varepsilon_{\text{Hf}}$ (t) values of 0.2 to 4.9 (calculated at t = 287 Ma). Corresponding $T_{\text{DM1}}$ is calculated at ca. 0.74 Ga–0.95 Ga and $T_{\text{DM2}}$ at ca. 1.00 Ga–1.30 Ga. Thirty igneous zircons of DYCWL-3 gave $^{176}$Hf/$^{177}$Hf ratios ranging from 0.282602 to 0.282738. Computations based on crystallization ages (286 Ma) of the magmas yielded $\varepsilon_{\text{Hf}}$ (t) between −1.5 and 4.9. Corresponding $T_{\text{DM1}}$ is calculated at ca. 0.73 Ga–1.06 Ga and $T_{\text{DM2}}$ at ca. 1.00 Ga–1.40 Ga.

5.3. Major and trace elements

5.3.1. Major elements

Major element compositions of the representative samples of the Chuanwulu complex are given in Table 3. The loss on ignition (LOI) for all samples was ~1 wt.%. The complex has a wide range of chemical compositions, e.g., 50.52–70.64 wt.% for SiO$_2$, 0.48–5.43 wt.% for MgO and 0.86–7.88 wt.% for CaO. Particularly, the TiO$_2$, Fe$_2$O$_3$, MnO, MgO, CaO and P$_2$O$_5$ contents linearly decrease with increasing SiO$_2$ content (Fig. 7), which is supported by high values of the regression coefficient (R$^2$) that are predominantly higher than 0.9. Besides, the Al$_2$O$_3$ content is nearly constant, whereas K$_2$O, Na$_2$O and K$_2$O+Na$_2$O show positive correlation with SiO$_2$ content at ~61.69 wt.%, whereas they show a negative correlation at >61.69 wt.% SiO$_2$.

The gabbro-diorite unit (unit I) has the lowest SiO$_2$ (50.52–55.05 wt.%) contents among the three units, and displays moderate TiO$_2$ (0.84–1.47 wt.%), Fe$_2$O$_3$ (7.03–11.22 wt.%), MnO (0.10–0.15 wt.%), MgO (3.29–5.43 wt.%), CaO (5.17–7.88 wt.%) and P$_2$O$_5$ (0.70–1.24 wt.% contents but considerably high K$_2$O (3.50–6.16 wt.%) and Na$_2$O (3.44–5.95 wt.) contents. Except for sample CWL-46 with peralkaline affinities, most samples of this unit fall into the alkaline field on the AR vs. SiO$_2$ diagram and in the metaluminous field on the A/NK vs. A/CNK diagram (Fig. 8a and b).

The monzodiorite–syenite unit (unit II) ranges from 58.88 to 65.91 wt.% for SiO$_2$, 0.47–0.77 wt.% for TiO$_2$, 3.47–5.67 wt.% for Fe$_2$O$_3$, 1.26–2.37 wt.% for MgO, 2.63–3.96 wt.% for CaO and 0.30–0.58 wt.% for P$_2$O$_5$ and exhibits high K$_2$O (5.55–6.35 wt.%), Na$_2$O (4.86–5.35 wt.%) and Na$_2$O + K$_2$O (10.50–11.70 wt.) contents. All samples in this unit plot into the alkaline field on the AR-SiO$_2$ diagram (Fig. 8a). Their Al$_2$O$_3$ contents are between 14.50 and 15.84 wt.%, exhibiting metaluminous and peralkaline affinities on the plot of A/CNK versus A/NK (Fig. 8b).

Granite dykes (unit III) range from 67.10 to 70.64 wt.% for SiO$_2$, 0.20–0.40 wt.% for TiO$_2$, 1.13–2.74 wt.% for Fe$_2$O$_3$, 0.03–0.05 wt.% for MgO, 0.86–2.27 wt.% for CaO and 0.07–0.23 wt.% for P$_2$O$_5$, and have moderate K$_2$O (5.18–7.17 wt.%), Na$_2$O (4.78–5.99 wt.%) and K$_2$O + Na$_2$O (10.19–11.95 wt.) contents. They plot into the alkaline and peralkaline fields on the AR-SiO$_2$ diagram (Fig. 8a). Their A/CNK and A/NK values ranging from 0.77 to 0.84 (Fig. 8b) suggest that the rocks of this unit are metaluminous and peralkaline.

In spite of the significant chemical differences, all the samples of the Chuanwulu complex display a K-rich characteristic (K$_2$O+ 2 wt.% > Na$_2$O) and are plotted into the shoshonitic field on a K$_2$O versus SiO$_2$ diagram (Fig. 8c). Compared with other granitoids in the Chinese part of South Tian Shan Collisional Belt, the Chuanwulu complex is characterized by relatively low A/CNK and A/NK values but relatively high K$_2$O contents (Fig. 8b and c).
5.3.2. Trace elements

The gabbro–diorite unit (unit I) contains high Rb (72–135 ppm) and Ba (1399–2280 ppm) contents and moderate Nb (11.87–33.96 ppm), Ta (0.71–1.71 ppm) and Nd (30.78–53.09 ppm) contents. This unit has the highest Sr (1646–3101 ppm) contents among the three units, leading to the low Rb/Sr (0.03–0.07) and high Sr/Nd (37–103) ratios. Notably, most samples of unit I have moderate Cr and Ni concentrations ranging from 38.6 ppm to 165.6 ppm and 14.4 ppm to 42.5 ppm, respectively, with the exception of CWL-9 that has high Cr (562.1 ppm) and Ni (311.01 ppm) contents. As expected for their alkaline affinity, samples of this unit have relatively high total rare earth element (REE) contents ranging from 163 ppm to 267 ppm, and they show light rare element (LREE)-enriched patterns [(La/Yb)N = 16.3–24.1] with no to slightly negative Eu anomalies (Eu/Eu* = 0.77–1.02) (Fig. 10).

The monzonite–syenite unit (unit II) also has high Rb (82.46–145.43 ppm) and Ba (1257–1884 ppm) contents, and moderate to high Nb (17.24–27.34 ppm) and Ta (1.15–1.96 ppm) contents. Samples of this unit are enriched in Sr (1219–1752 ppm), with Rb/Sr and Sr/Nd ratios ranging from 0.06 to 0.12, and from 43 to 63, respectively. The Cr and Ni contents of this unit vary from 11.49 ppm to 25.27 ppm, and from 4.77 ppm to 10.97 ppm, respectively, much lower than those of unit I. This unit has total REE contents ranging from 141.18 ppm to 174.40 ppm. Chondrite-normalized REE patterns (Fig. 10) of the unit II shows moderate enrichment of LREEs relative to HREEs [(La/Yb)N = 16.89–19.69] with moderately to slightly negative Eu anomalies (Eu/Eu* = 0.77–0.86).

Granite dykes (unit III) have high Rb (111.65–151.86 ppm) and Ba (875–1581 ppm), and moderate to high Sr (600–1201 ppm), Nb (6.16–31.21 ppm) and Ta (0.57–3.10 ppm) concentrations, with Rb/ Sr and Sr/Nd ratios varying from 0.10 to 0.25, and from 27 to 92, respectively. They have total REE contents varying from 151.38 ppm to 187.51 ppm, and moderate enrichment of LREEs relative to HREEs [(La/Yb)N = 14.13–25.09] with slightly to moderately negative Eu anomalies (Eu/Eu* = 0.74–0.88) (Fig. 10).

Collectively, all studied samples show strong enrichment of large ion lithophile elements (LILE), such as Rb, Ba, Th U and Sr, and negative Nb, Ta, P and Ti anomalies on normalized incompatible element patterns (Fig. 9). Besides, the unit I to unit III shows a trend of decreasing total REE but a restricted variation in (La/Yb)N ratios.

5.4. Sr–Nd–O isotopes

Sr, Nd and O isotopic data for the Chuanwulu samples are listed in Table 4, and the various initial Sr and Nd isotopic ratios are calculated to the ages of 287 Ma for gabbro–diorite samples and 286 Ma for other samples. The data are shown in the plot of (87Sr/86Sr)t versus εNd (t) in Fig. 11, and the published compositional fields for Late Paleozoic granitoids from the CAOB are shown for comparison. Samples of the Chuanwulu complex have low 87Sr/86Sr ratios (0.70835–0.70349), and demonstrate a wide range of age-corrected (87Sr/86Sr)t ratios. Units I, II and III have (87Sr/86Sr)t ratios from 0.70455 to 0.70606, 0.70543–0.70749 and 0.70751, respectively. The complex exhibits 143Nd/144Nd ratios between 0.50956 and 0.50102, and ratios of 143Nd/144Nd between 0.512156 and 0.512193. The age-corrected εNd values for samples define a small range from −2.3 to −1.4. In addition, samples yield relatively consistent and old Nd modal ages (TDM1Nd = 1.0–1.1 Ga, TDM2Nd = 1.2–1.3 Ga). The three units of the complex show variable whole-rock oxygen isotope values, ranging from 5.7 to 9.4‰, 8.5–9.1‰, and 9.1, respectively.

6. Discussion

6.1. Mixing or fractional crystallization

The Chuanwulu complex has a wide SiO2 range, from 50.52 to 70.64 wt.%, which could be produced by several possible magmatic processes, i.e. fractional crystallization or crustal assimilation accompanying fractional crystallization (AFC) of mantle-derived melts, and simple magma mixing between the mafic and felsic melts (e.g. Clemens and Vieleuf, 1987; Creaser et al., 1991; DePaolo, 1981; Eby, 1992; Hibbard, 1991; King et al., 1997; Waight et al., 2000; Whalen et al., 1987; Yang et al., 2006). If felsic rocks were derived by fractional crystallization or AFC of mantle-derived melts, then the volume of mafic magma intruding into the crust should be an order of magnitude greater than that of granitoids (Frost et al., 2002; Turner et al., 1992). Thus, it appears unlikely that the felsic magmas of the Chuanwulu complex were derived from a basaltic parent magma by fractional crystallization or AFC process since the volume of felsic rocks is significantly larger than mafic rocks, even though some amount of gabbro has been buried by the Cenozoic sediments. In addition, the sharp boundary between units I and II could not be interpreted as a result of a continuous fractional crystallization process. Considering the wide ranges of (87Sr/86Sr)t ratios and δ18OSMOW values of the complex cannot be readily explained by a single mafic source, we propose a magma mixing model as the controlling mechanism for generating the Chuanwulu complex. Such a model is supported by the evidence discussed below.

The shape of MMEs which have sharp boundaries with their host rocks in unit II (monzonite–syenite) is commonly round to oblong. They are distinctive from their host rocks both by mineral assemblages and textures, and some of them contain large and rounded biotite phenocrysts that are chemically similar to those in the host rocks (Fig. 3d). Therefore, the MMEs were probably formed by a magma interaction...
process. The magma mixing model is also supported by the disequilibrium textures commonly observed, such as acicular apatite that in some cases occurs in mafic rocks (Fig. 3c), and poikilitic texture in the felsic rocks.

Geochemically, the mixing process is evidenced by the remarkable overlap of the data from mafic and felsic samples in the $f_{\text{DME}}$ (t) versus $\text{Nd} (t)$ versus $\text{Sr} (t)$ (Fig. 9). The linear correlations between some major elements and SiO$_2$, as stated above, can also act as evidence for a magma mixing model (McBirney, 1980; Perugini and Poil, 2004). The defined hyperbolic arrays on Rb/Sr versus Ti/Zr and the linear trend on Sr/Zr versus Ti/Zr, MgO versus FeO$_{\text{t}}$, and Ni versus MgO plots, all of which are expected for the mixing between the two distinct geochemical end-members, are evidenced by a simple interaction process between at least two magmas in the genesis (Holland et al., 2003; Karsli et al., 2007) (Fig. 12).

As stated above, it seems that the magma mixing processes played a dominant role in the formation of the Chuanwulu complex. However, fractional crystallization is also capable to impart influence and modify the geochemical features of the complex. Sharp contacts between unit I and unit II, acicular apatite and fine grained texture of the MMEs point out a process of fast quenching of mafic magma in a dynamic magma chamber (Kaygusuz and Aydinca, 2009; Yuan et al., 2010). Therefore, it is reasonable to infer that two stages of the crystal fractionation occurred: (1) formation of the variable
Besides, felsic samples have variable Ti contents and Ti* \[\text{Ti}^* = 2 \times \text{Ti}/(\text{Nd} + \text{Hf})\] prior to the mixing process, and (2) fractionation of some accessory minerals from the high-viscosity felsic melts. The felsic units (units II and III) are composed of variable rock types including monzonite, biotite monzonite, biotite adamellite, syenite and granite dykes, and have wide SiO2 contents ranging from 50.75 wt.% and 50.52 wt.% respectively) as well as low \[\delta^{18}O\text{-SMOW}\] values (6.8 and 5.7 respectively), all of which suggest a mantle source (Blattner et al., 1989, 2002), ruling out the possibility that the mafic magma was derived from the lower crust.

It is notable that the pronounced Nb and Ta troughs, together with the relative enrichment of LILEs and LREEs of the mafic samples, display ‘crust-like’ signatures. Furthermore, unit I has relatively low initial \[({}^8\text{Sr}/{}^{46}\text{Sr})_i\] ratios (0.70455–0.70606) but similar Nd and Hf isotopic features \[\epsilon_{\text{Nd}}(t) = -2.1 \text{ to } -1.4, \epsilon_{\text{Hf}}(t) = 0.2 \text{ to } 4.9\] relative to the felsic units \[({}^8\text{Sr}/{}^{46}\text{Sr})_i = 0.70543–0.70751, \epsilon_{\text{Nd}}(t) = -2.3 \text{ to } -1.8, \epsilon_{\text{Hf}}(t) = -1.5 \text{ to } 4.9\]. Experiments show that the isotopic equilibrium advances faster than the elemental equilibrium, and moreover, the Sr isotopic equilibrium is faster than that of Nd and Hf (Griffin et al., 2002; Pin et al., 1990; Scherer et al., 2000). Since remarkable differences in Sr isotopes between mafic and felsic rocks have been observed, the Nd and Hf isotopic similarities could not be attributed to the simple magma mixing but primarily resulted from the nature of their mantle source. Furthermore, all mafic samples have high Sr contents (1646–3101 ppm) and high Th/U ratios (2.91–4.36), combined with geochemically ‘crust-like’ signatures, indicating that the mafic magmas were derived from an incompatible elements-enriched mantle source, as proposed by Hawkesworth et al. (1993) and Rötting et al. (1998). Thus it is likely that the enriched sub-continental lithospheric mantle (SCLM) is a plausible source region for the mafic rocks of Chuanwulu complex. The two-stage Nd and Hf model ages of unit I relative to depleted mantle are older than 1.0 Ga, implying an ancient lithospheric mantle source. As shown in Fig. 11 and 15, the Sr–Nd isotope compositions and Nd model ages are comparable with those of the coeval Keping basalts in the Tarim block (Jiang et al., 2004), which suggests that they were derived from similar mantle sources.

Recent experiments and many cases reveal that the presence of water, which generally exists in some hydrous mineral phases such as phlogopite and amphibole, plays a crucial role in significantly lowering the partial melting temperature of such refractory mantle sources (Foley, 1991; Mengel and Green, 1989). Moreover, the large quantity of hydrous minerals such as biotite and hornblende in unit I is consistent with a water-bearing source region. Samples CWL-48 and CWL-49, both of which have the lowest \[\delta^{18}O\text{-SMOW}\] values and SiO2 contents, can geochemically represent the mafic magma derived from the mantle and then experienced minor fractional crystallization and little mixing with the felsic melts. The two samples are
Table 3
Major (wt.%) and trace (rare earth included) element data of the Chuanwulu complex.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Gabбро–diorite unit (unit I)</th>
<th>Monzonite–syenite unit (unit II)</th>
<th>Granite dykes (unit III)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lithology</td>
<td>CWL-04</td>
<td>CWL-05</td>
<td>CWL-09</td>
</tr>
<tr>
<td>LiO₂</td>
<td>52.92</td>
<td>54.60</td>
<td>54.69</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.84</td>
<td>1.06</td>
<td>1.07</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>8.16</td>
<td>7.83</td>
<td>8.23</td>
</tr>
<tr>
<td>MnO</td>
<td>0.11</td>
<td>0.10</td>
<td>0.13</td>
</tr>
<tr>
<td>MgO</td>
<td>4.14</td>
<td>3.42</td>
<td>3.93</td>
</tr>
<tr>
<td>CaO</td>
<td>7.00</td>
<td>5.31</td>
<td>5.83</td>
</tr>
<tr>
<td>Na₂O</td>
<td>4.25</td>
<td>4.61</td>
<td>4.15</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.76</td>
<td>5.08</td>
<td>5.42</td>
</tr>
<tr>
<td>Loss</td>
<td>0.84</td>
<td>0.90</td>
<td>0.95</td>
</tr>
<tr>
<td>A/CNK</td>
<td>0.68</td>
<td>0.71</td>
<td>0.63</td>
</tr>
</tbody>
</table>

(continued on next page)
and greater than 80 km (Olafsson and Eggler, 1983), respectively, we postulate that amphibole in the mantle would not be stable at depth less than 70 km. Garnet could account for the observed REE data. Since garnet and amphibole is stable between the two lines between partial melts in spinel- and garnet-facies peridotite, the mafic rocks of the Chuanwulu complex were produced by partial melting of an ancient sub-continental lithospheric mantle that is chiefly composed of garnet-, phlogopite- and amphibole-bearing spinel lherzolite.

6.3. Source of felsic rocks

As stated above, the diversity of felsic rocks possibly resulted from variable contributions of mafic magma. Therefore, the granite dykes (unit III), which have high SiO2 contents (67.10–70.64 wt.%), imply a residual garnet-bearing and plagioclase-free (eclogite) nature. As shown in Fig. 13, variable degrees of partial melting of an enriched mantle source either in the spinel stability field or the garnet stability field cannot generate the La/Yb-Yb and La/Yb-Dy/Yb systematics of the parental magma of mafic rocks (unit I). However, most samples of unit I plot between the two lines between partial melts in spinel- and garnet-facies peridotite. Therefore, the granite dykes (unit III) plot in the adakite field with one exception that has relatively high Y contents (Y = 16.3 ppm). Besides, the mafic samples have higher Y contents relative to felsic rocks and thus the process of magma mixing may have elevated the Y contents of felsic magma to variable degrees. Therefore, we propose that such geochemical signatures of units II and III are comparable to those of adakites and experimentally stimulated slab melts (Defant and Drummond, 1990; Peacock et al., 1994; Qin et al., 2010), implying a residual garnet-bearing and plagioclase-free (eclogite) source.

Given the high K contents, the felsic rocks of the Chuanwulu complex have similar geochemical characteristics to "c (continental)-type adakite" granite in some Chinese literature (Xiao and Clemens, 2007; Zhang et al., 2001). Recent studies indicate that these rocks can be formed by several ways: (1) fractional crystallization of hydrous basaltic magma at high pressure (Castillo, 2006; Chiaratia, 2009; Macpherson et al., 2006; Rodríguez et al., 2007); (2) partial melting of ancient lower crust at high pressure (Chung et al., 2003; Petford et al., 2007).
and Atherton, 1996); or (3) partial melts of delaminated lower continental crust that experienced subsequent reaction with surrounding asthenosphere (Gao et al., 2004; Xu et al., 2002). As discussed above, the felsic rocks were not generated by fractionation of basaltic magmas or AFC processes, therefore model (1) can be easily excluded. The felsic samples of the Chuanwulu complex have low Mg# values ranging from 0.44 to 0.54 with an average of 0.47, with (87Sr/86Sr)t ratios ranging from 0.70543 to 0.70751 and whole rock two-stage Nd model ages of ca. 1.2 Ga. Zircon grains from DYCWL-3 that represent the felsic units show εHf (t) values ranging from −1.5 to 4.9 and two-stage Hf model ages of 1.0 Ga to 1.4 Ga. Such geochemical features are inconsistent with the melts being derived from delaminated lower continental crust that experienced subsequent reaction with surrounding asthenosphere (model 3, Qin et al., 2010) but suggests

Fig. 7. Harker diagrams for the Chuanwulu complex.
that the felsic rocks could have been derived from reworked ancient lower crustal rocks, similar to some Miocene Tibetan adakites and Dabie adakites (Wang et al., 2007b).

Experimental data suggests that adakitic felsic rocks can be generated by partial melting of variable source rocks at given P–T and H2O conditions, including melting in the system tonalite–H2O, high-pressure (1.0–3.2 GPa) dehydration melting of tonalities and high-pressure (1.0–4.0 GPa) dehydration melting of metabasaltic rocks and eclogites (Wang et al., 2007b). However, melting in the system tonalite–H2O generally produces tonalitic and trondhjemitic magmas that have higher A/NK (Al2O3/(Na2O+K2O)) values (Fig. 8b) and lower K2O contents (Fig. 8c) than Chuanwulu felsic rocks (e.g. Bryant et al., 2004). High-pressure (1.0–3.2 GPa) dehydration melting tonalites produces granitic melts with slightly to strongly peraluminous compositions (Fig. 8b). However, Chuanwulu felsic samples are metaluminous and peralkaline, arguing against high-pressure dehydration melting of tonalite gneisses.

In contrast, high-pressure (1.0–4.0 GPa) dehydration melting of metabasaltic rocks and eclogites can readily generate high-K granitic melts chemically similar to Chuanwulu felsic rocks. According to experimental data (e.g. Beard and Lofgren, 1991; Rapp and Watson, 1995), the K2O content of melts is mainly controlled by the K2O contents of source rocks and pressure. Furthermore, dehydration melting, involving the breakdown of hydrous minerals such as hornblende or zoisite, more readily gives rise to K-rich melts (Fig. 8c). Thus, K-rich metabasaltic rocks and eclogites are suitable sources for the Chuanwulu felsic rocks.

The granite dykes have low Yb/Lu (6.0 to 6.4), Dy/Yb (1.4 to 1.9) and (Ho/Yb)N (0.9 to 1.1) ratios, suggesting that amphibole was the predominant phase in their source region (Moyen, 2009). Additionally, on the (La/Yb)N vs. YbN diagram (Fig. 14b), most felsic samples follow the melting line of amphibolite, with two exceptions of samples CWL-1 and CWL-71 which follow the melting line of 10%-garnet amphibolite. Since all of the mafic samples (unit I) are characterized by higher Yb contents ranging from 1.12 ppm to 2.18 ppm relative to felsic rocks, we propose that the magma mixing process should have elevated the Yb contents of felsic magma. We note that CWL-1 has the highest SiO2 content among all samples and CWL-71 is characterized by the highest δ18OV-SMOW (9.1‰) value and (87Sr/86Sr)t ratio (0.70751) amongst our samples. Thus, the two samples with the lowest Yb contents may best represent the primary crustal partial melts. Therefore, we propose that felsic rocks of the Chuanwulu complex may have formed by partial melting of garnet-bearing amphibolite facies lower crust rocks. Comparably, the samples of Mangqisu pluton in the east part of STCB formed at ca. 300 Ma (Zhu et al., 2008), exhibit an adakitic signature as well.

In summary, the most felsic rocks of the Chuanwulu complex were possibly derived from a thickened basaltic lower crust source, as indicated by their geochemical characteristics (Petford and Atherton, 1996; Petford and Gallagher, 2001).

6.4. Implications for geodynamic settings and continental growth

6.4.1. Constraint on the time of collision

Proposing a general geodynamic model concerning every aspect of tectonic evolution in South Tianshan Collisional Belt is beyond the scope of this paper. Here, we only discuss the collisional time between the Central Tianshan and Tarim blocks based on the nature of Chuanwulu complex together with other geological evidence.

Based on the concept of a ‘duality of orogens’ (Sylwester, 1998; Zwart, 1967), the eastern part of the STCB has been widely accepted as a high-pressure collisional belt, similar to Alps and Himalayas (Gao et al., 2011; Solomovich, 2007). In such a scenario, the time of collision between the two blocks is most likely to be inferred by the dating of these high pressure/ultra-high pressure (HP/UHP) metamorphic rocks. However the various U–Pb ages obtained for the HP and UHP rocks (Gao et al., 2011, and references therein) give rise to diverging opinions concerning the timing of the collision of the Yili and Tarim blocks and the final amalgamation of the CAOB. As
mentioned above, in spite that most metamorphic zircon ages of HP–UHP were reported to be Carboniferous (Gao et al., 2011, and references therein), some younger U–Pb ages (e.g., 233 ± 4 Ma–226 ± 4.6 Ma) were also reported and interpreted to represent the collision age (e.g. Zhang et al., 2007a, 2007b). However, these younger zircon ages can also result from fluid-mediated recrystallization of the zircon grains (e.g., Jong et al., 2009; Su et al., 2010) or the rejuvenation of old mountain belts by intra-continental deformation (e.g., Gilotti and McClelland, 2007), both of which were common in the STCB during the prolonged geological evolution.

Although the complex has high Na contents (3.44 wt.%–6.59 wt.%), both mafic and felsic rocks of the Chuanwulu complex display K-rich (K₂O + 2 wt.% > Na₂O) and alkaline characteristics, and no Early Permian Na-rich (K₂O + 2 wt.% > Na₂O) igneous rocks have been identified in STCB and the south margin of the Central Tianshan Block (Huang et al., 2011; Jiang et al., 1999, 2004; Konopelko et al., 2007; Long et al., 2011; Ma et al., 2010; Solomovich, 2007; Wang et al., 2007a; Zhang et al., 2009; Zhu et al., 2008, and references therein). Obviously, these features cannot coincide with subduction-related granites. The Chuanwulu mafic rocks were produced by the partial melting of an enriched lithospheric mantle source at depth between 70 km and 80 km, implying the STCB was already in an extensional setting and underwent lithospheric thinning before ~285 Ma (Hou et al., 2010; Solomovich, 2007). In addition, no contractional structures are present at the contact zones between the Chuanwulu complex and the country rocks, further suggesting that the complex was emplaced in an extensional regime rather than a contractional regime. However, adakitic signatures of the felsic rocks indicate that a thickened lower crust beneath the STCB still remained. Thus, the Chuanwulu complex is likely to have formed in the early stage of the post-collisional setting, when changes in the tectonic regime, e.g. from dominantly contraction and thickening to subsequent

![Fig. 9. Primitive mantle-normalized spidergrams for samples of Chuanwulu complex. Normalized values are from Sun and McDonough (1989).](image-url)
extension and thinning, had occurred but the crustal thickness was still high enough to produce adakitic magmas. Such a conclusion is compatible with the contemporary (285 Ma) leucogranite dikes crosscutting the HP–LT metamorphic belt in the Chinese part of the STCB (Gao et al., 2011).

Hence, we propose that the collision between Tarim and Central Tianshan blocks should have occurred prior to Early Permian. This conclusion is also supported by the regional geology. As mentioned above, Permian terrestrial volcanic strata unconformably overlie late Carboniferous marine carbonate rocks. Such unconformity is always interpreted to represent a collisional event between Tarim and Central Tianshan blocks (Huang et al., 2011). In addition, the extensive diabase dike swarms formed at ca. 275 Ma in the CTB, STCB (Gu et al., 2001; Jiang et al., 2005) and north margin of Tarim block (Sun et al., 2007) indicate an intraplate extensional regime. Additionally, arc-related granites and volcanic rocks with ages ranging from middle Silurian to early Carboniferous are widely exposed along the southern margin of the Central Tianshan block (Yang and Zhou, 2009; Yang et

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**Fig. 10.** Chondrite-normalized rare earth element patterns of Chuanwulu samples. Normalized values are from Sun and McDonough, 1989).
Table 4
Sr–Nd–O isotopic data for Chuanwulu samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>CWL-04</th>
<th>CWL-10</th>
<th>CWL-46</th>
<th>CWL-48</th>
<th>CWL-49</th>
<th>CWL-23</th>
<th>CWL-38</th>
<th>CWL-40</th>
<th>CWL-72</th>
<th>CWL-71</th>
</tr>
</thead>
<tbody>
<tr>
<td>(87Sr/86Sr)i</td>
<td>0.70543</td>
<td>0.70538</td>
<td>0.70562</td>
<td>0.70649</td>
<td>0.70639</td>
<td>0.70794</td>
<td>0.70791</td>
<td>0.70768</td>
<td>0.70819</td>
<td>0.70864</td>
</tr>
<tr>
<td>(87Sr/86Sr)m</td>
<td>0.70521</td>
<td>0.70455</td>
<td>0.70571</td>
<td>0.70606</td>
<td>0.70602</td>
<td>0.70682</td>
<td>0.70543</td>
<td>0.70585</td>
<td>0.70749</td>
<td>0.70751</td>
</tr>
<tr>
<td>εNd (t)</td>
<td>1.4</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
<td>5.6</td>
</tr>
<tr>
<td>εNd (t)</td>
<td>1.4</td>
<td>1.4</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
<td>2.1</td>
</tr>
<tr>
<td>141Nd/144Nd</td>
<td>0.512385</td>
<td>0.512379</td>
<td>0.512349</td>
<td>0.512381</td>
<td>0.512378</td>
<td>0.512354</td>
<td>0.512362</td>
<td>0.512357</td>
<td>0.512342</td>
<td>0.512336</td>
</tr>
<tr>
<td>143Nd/144Nd</td>
<td>0.512191</td>
<td>0.512192</td>
<td>0.512156</td>
<td>0.512185</td>
<td>0.512178</td>
<td>0.512175</td>
<td>0.512178</td>
<td>0.512175</td>
<td>0.512173</td>
<td>0.512152</td>
</tr>
</tbody>
</table>

al., 2005; Zhu et al., 2005; Zhu et al., 2006a; our unpublished data), suggesting that the collision occurred plausibly after the Early Carboniferous.

Consequently, we conclude that the collision between the Central Tianshan and Tarim blocks most likely took place during Late Carboniferous, and the Chuanwulu complex was emplaced in a post-collisional setting. This conclusion is comparable to the palaeomagnetic data that indicates that the North Tianshan belt, Central Tianshan block, Siberian and Tarim blocks were amalgamated in Late Carboniferous (at ca. 300 Ma, Bazhenov et al., 2003; Gao et al., 2011).

6.4.2. Implications for Phanerozoic crustal growth and evolution of CAOB
The problem of the growth and evolution of the continental crust has always been an important subject of research in earth sciences. On a regional scale, numerous previous studies of Phanerozoic granitoids of the CAOB have been published (e.g. Xu et al., 2002; Zhu et al., 2002; our unpublished data), suggesting that the collision occurred plausibly after the Early Carboniferous.

On a regional scale, numerous previous studies of Phanerozoic granitoids in the Chinese South Tianshan have been published (e.g. Xu et al., 2002; Zhu et al., 2002; our unpublished data), suggesting that the collision occurred plausibly after the Early Carboniferous.

The proportions of the juvenile or mantle component for most granitoids of the northern parts of CAOB were estimated to vary from 70 to 100% by Jahn et al. (2000a, 2000b), suggesting a significant juvenile component in tectonic units comprising the Altai, Junggar and North Tianshan and Central Tianshan. In the STCB, however, Permian igneous rocks, including granitoids, mafic intrusions and basalts, are uniformly characterized by negative εNd (t) values and ancient Nd model ages of > 1 Ga. These data are compatible with εNd (t) values and U–Pb ages previously published and obtained in this study for the Precambrian basement in the Tarim block that is distinguished from the other tectonic units as shown in Fig. 12.

The isotopic diversity of magmatic and basement rocks in various tectonic units of the CAOB may reflect the different compositions of the lithosphere. Since most collision-related igneous rocks in STCB originated from Precambrian basement sources, we propose that the lower lithosphere beneath the STCB should be dominated by Precambrian crystalline rocks, as well as very limited juvenile materials, if any. Thus, we propose two alternative mechanisms to account for the Phanerozoic continental evolution of the CAOB: (1) in the Northern tectonic units of the CAOB, such as Altai, Junggar and NTOB and CTB, the Phanerozoic continental growth mainly involved the re-melting of juvenile materials (i.e. juvenile lower crust, oceanic crustal and arc materials) and/or input of asthenospheric magmas, while (2) the Phanerozoic continental evolution of the South Tianshan Collisional Belt is characterized by the reworking of ancient lithosphere (i.e. ancient lithospheric mantle and ancient crust).

7. Conclusions
(1) The Chuanwulu complex consists dominantly of a gabbro–diorite unit (unit I), a monzonite–syenite unit (unit II) and granite dykes (unit III). LA-ICP-MS zircon U–Pb dating shows that unit I and unit II were emplaced at 287.8 ± 4.3 Ma and 286.4 ± 2.5 Ma, respectively. This time span fits the known ages of post-collisional magmatism elsewhere in the STCB.
(2) The complex probably formed by magma mixing between mafic and felsic magmas. The mafic rocks (unit I) were generated by partial melting of incompatible element-enriched spinel-facies lithospheric mantle, whereas the felsic rocks (units II and III) resulted from partial melting of the lower crust.
(3) The nature of the Chuanwulu complex, together with other geological evidence, suggests that the complex formed in the early stage of the post-collisional extension and the collision between the Central Tianshan and Tarim blocks occurred in the Late Carboniferous.
(4) The Sr–Nd isotopic characteristics of magmatic and basement rocks from STCB suggest that the reworking of ancient lithosphere is the dominant mechanism that is responsible for Phanerozoic continental evolution of the STCB, which significantly differs from other tectonic units of CAOB.
Fig. 12. (a) Plot of MgO versus FeOt (after Zorzi et al., 1991); (b) Plot of Ni versus MgO (after Dong et al., 2006); (c) Plot of Sr/Zr versus Ti/Zr (after Karsli et al., 2007); (d) Plot of Rb/Sr versus Ti/Zr (after Karsli et al., 2007). Symbols are as in Fig. 7.

Fig. 13. La/Yb versus Yb and Dy/Yb diagrams showing that the rocks need an enriched mantle source for garnet-bearing spinel mantle partial melting. All the mantle components are from Miller et al. (1999) and therein. Symbols are as in Fig. 7.

Fig. 14. Plots of (a) Sr/Y vs. Y and (b) La/Yb(n) vs. Yb(n) for felsic rocks of the Chuanwulu complex and Mangqisu pluton (modified after Defant and Drummond, 1990). Adakite fields are from Martin et al. (2005). Symbols are as in Fig. 7.
from Jiang et al. (1999, 2004), Chen and Jahn (2002, 2004), Xu et al. (2005) and Zhu et al. (2006b). Fields for basement rocks (amphibolites and gneisses) of the Junggar, Altay, Tien Shan (undivided) and Tarim are from Hu et al. (2000). Fields of evolution of the Early Proterozoic and Archean crust are after Jahn et al. (2000a). Nd isotope evolution of the depleted mantle is shown according to DePaolo (1981) (DM1) and as linear evolution $\varepsilon_{DM} = 0$ at 4.56 Ga to +10 at present after Jahn et al. (2000a) (DM2).

Fig. 15. $T_{124}$ (Nd) or Age versus $\varepsilon_{DM}$ (t) diagram showing initial Nd isotopic composition of the Chuanwulu complex (this study) and magmatic rocks of the adjacent units of CAOB. Aboisclia means 1-stage Nd model age for magmatic rocks and primary age for the fields of basement rocks of the CAOB, respectively. Data of other magmatic rocks from CAOB are from Jiang et al. (1999, 2004), Chen and Jahn (2002, 2004), Xu et al. (2005) and Zhu et al. (2006b). Fields for basement rocks (amphibolites and gneisses) of the Junggar, Altay, Tien Shan (undivided) and Tarim are from Hu et al. (2000). Fields of evolution of the Early- to Middle Proterozoic and Archean crust are after Jahn et al. (2000a).

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References


Chiaradia, M., 2009. Adakite-like magmas from fractional crystallization and melting assimilation of mafic lower crust (Eocene Macuchi arc, Western Cordillera, Ecuador).
Chemical Geology 265, 468–487.