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Geodynamic Evolution of the Proto-Tethys Ocean in the West Kunlun Orogenic Belt, NW Tibet: Implications from the Sub-Arc Crust and Lithospheric Mantle Modification

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ABSTRACT

Radiogenic isotopes serve as a crucial tool for investigating crustal evolution, playing a pivotal role in revealing magma sources and petrogenesis. However, they can be ineffective in distinguishing distinct magmatic sources with similar radiogenic isotopic compositions, a common occurrence in nature. This paper addresses this challenge in the Ordovician igneous rocks from the West Kunlun orogenic belt (WKOB), to distinguish between two potential magmatic sources (i.e., the Tarim Craton and the Tianshuihai terrane) with similar isotopic

compositions, by using appropriate thermodynamic and geochemical modeling based on mineral and whole-rock geochemistry. Zircon U–Pb dating yields ages of 483 ± 3 Ma, 469 ± 2 Ma and 461 ± 2 Ma for the Pushou gabbros and the Datong monzogranites and syenites, respectively. The Pushou gabbros exhibit low SiO_2 (47.4–49.1 wt.%), high MgO (5.5–6.9 wt.%), high large ion lithophile elements (LILEs, e.g., Rb, Ba, Th, and K) and low high field-strength elements (HFSEs, e.g., Nb, Ta, Zr, Hf, P, and Ti), which suggest an origin in subduction-modified mantle. They display high whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios (0.7156 to 0.7192), negative whole-rock $\epsilon\text{Nd}(t)$ values (-7.1 to -7.8), as well as high zircon $\delta^{18}\text{O}$ values (7.6–7.9 ‰) and enriched zircon Hf isotopic compositions ($\epsilon\text{Hf}(t) = -5.3$ to -7.7), consistent with 1–5% subducted sediments in an enriched mantle source. Trace element models further confirm that the gabbros are most likely derived from low-degree (~15 %) partial melting of subduction-modified Tarim mantle in the spinel-garnet facies, rather than from the Tianshuihai mantle. The Datong syenite samples belong to the shoshonitic series and are characterized by medium SiO_2 (59.5–61.4 wt.%), relatively low MgO (0.9–1.2 wt.%) and $\text{Mg}^\#$ (37–42), enrichment in LILEs and depletion in HFSEs. They have high whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios (0.7103 to 0.7105) and negative whole-rock $\epsilon\text{Nd}(t)$ values (-3.8 to -4.3), along with negative to slightly positive zircon $\epsilon\text{Hf}(t)$ values (-3.8 to $+2.6$), similar to coeval mafic rocks. Thermodynamic and geochemical modeling suggests that the Datong shoshonitic rocks likely originated via crystal fractionation of shoshonitic basaltic magmas in the SW Tarim Craton. The Datong monzogranites have high SiO_2 (69.7–72.6 wt.%), low MgO (0.6–0.7 wt.%) and demonstrate a typical enrichment in alkalis, Zr, and Nb with depletion in Sr, P, and Ti, consistent with A-type granites. They are characterized by high whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios (0.7321 to 0.7323), negative whole-rock $\epsilon\text{Nd}(t)$ (-11.3 to -11.8), negative zircon $\epsilon\text{Hf}(t)$ (-11.0 to -16.5) and high zircon $\delta^{18}\text{O}$ (7.2–8.0 ‰), indicating derivation from the remelting of an ancient crustal source. Thermodynamic, major, and trace element modeling indicate that their parent magma may have been generated by water-deficient (~2 wt.%) partial melting of ancient crustal material beneath the SW Tarim Craton rather than that of the Tianshuihai terrane, under high-temperature ($T > \sim 950$ °C), low-pressure ($P = 5$ –8 kbar) conditions. Based on the tectonic framework of the WKOB, we propose

that the original mantle and crust beneath the southern Kunlun terrane may have been modified or partially replaced by that beneath the SW Tarim Craton during the Ordovician. Therefore, this evidence for Tarim-derived magmatism, when combined with regional sedimentary and structural records, indicates that Ordovician magmatism in the southern Kunlun terrane is most consistent with episodic northward subduction of the Proto-Tethys Ocean, commencing at ~485 Ma. Middle Ordovician slab break-off can explain the formation of the A-type granites, but re-instated northward subduction is required for the generation of late Ordovician Datong syenites.

Key words: Thermodynamic modeling; Magmatic source; Northward subduction; Proto-Tethys Ocean; Slab break-off

INTRODUCTION

Understanding the causes of geochemical and isotopic variations of igneous rocks in terms of magma sources and evolution processes is arguably the fundamental goal of igneous petrology. Many models and hypotheses have been proposed to explain their diversity (few of which are mutually exclusive), including direct mantle partial melts (e.g., Kushiro, 1974); partial melting of deep crustal lithologies (e.g., Rapp & Watson, 1995; Drummond & Defant, 1990); fractional crystallization of mantle-derived parental magmas (e.g., Grove et al., 2003; Lee & Bachmann, 2014) and magma mixing/assimilation (e.g., Barbarin, 2005; Collins, 1996; Kemp et al., 2007; Yang et al., 2007). Among these, partial melting and closed-system fractional crystallization can produce magmatic diversity from a single source. Since radiogenic isotope systems remain unaffected by melting or fractional crystallization, many studies have utilized whole-rock Sr-Nd-Pb isotopes and zircon Lu-Hf isotopes to constrain source contributions (e.g., Kemp et al., 2007; Yang et al., 2007; Béguelin et al., 2019; Deng et al., 2020). Unfortunately, different sources with similar isotope compositions are challenging to distinguish using this approach (e.g., Turnbull et al., 2010; Zeng et al., 2011). However, careful elemental comparison between igneous rocks and thermodynamically-modeled melts based on likely source compositions can potentially differentiate magmatic sources with similar isotope

compositions (Liu et al., 2022; Hu et al., 2022) and thus further elucidate petrogenetic processes (Heinonen et al., 2020). An example of this approach is presented here, from the West Kunlun orogenic belt (WKOB).

The WKOB in the northwestern Tibet Plateau (Figs. 1a-b) was formed by accretion-to-collision orogenesis during Paleozoic subduction and closure of the Proto-Tethys Ocean (Xiao et al., 2003, 2005; Liu et al., 2019; Zhang et al., 2018a, 2018b; Yin et al., 2020). The southern Kunlun terrane has the most complex tectonic assemblages in the WKOB, including ophiolites, arc-forearc rocks, accretionary complexes, magmatic and metamorphic rocks (Xiao et al., 2003; Zhang et al., 2018a, 2021; Tao et al., 2022), all of which provide valuable insights into geological processes at ancient convergent plate margins. However, the tectonic setting and petrogenesis of its Ordovician magmatism in the southern Kunlun terrane remain controversial. Contrasting models have different interpretations of subduction polarity—northward subduction of the southern Proto-Tethys Ocean beneath the southern Kunlun terrane and Tarim Craton (Yuan et al., 2003; Xiao et al., 2003, 2005; Zha et al., 2022; Tao et al., 2022), or southward subduction of the northern Proto-Tethys Ocean beneath the southern Kunlun terrane and Tianshuhai terrane (Liao et al., 2010; Zhang et al., 2018a, 2018b, 2019a). Therefore, clarification of the geodynamic mechanism that generated Ordovician magmatic rocks in the southern Kunlun terrane is essential to understanding the evolution of the Proto-Tethys Ocean.

Given that the southern Kunlun terrane was located between the Tarim Craton and Tianshuihai terrane (Fig. 1b; Xiao et al., 2003; Zhang et al., 2010), source characterisation of its arc magmatism (i.e., within the Tarim Craton or the Tianshuihai terrane) has the potential to provide critically test between northward and southward subduction models. Previous studies have attempted to use radiogenic isotope compositions of the granitic rocks in the southern Kunlun terrane to trace their sources from either the Tarim craton (corresponding to northward subduction; Yuan et al., 2003) or Tianshuihai terrane (corresponding to southward subduction; Xu et al., 2021). However, the location of the source is still controversial because the lower crust of the Tarim craton has similar isotopic compositions to that of the Tianshuihai terrane (see Supplementary Table S1 for details; Zhang et al., 2016, 2018c, 2019b). Therefore,

accurately resolving the magmatic source from different end-members with similar isotope compositions has profound implications for studying the geodynamic evolution of the Proto-Tethys Ocean.

In this work, we present an integrated study of petrology, mineralogy, whole-rock geochemistry, whole-rock Sr–Nd isotope and zircon U–Pb–Hf–O isotope characteristics of the igneous rocks in the southern Kunlun terrane. Based on these results combined with a compilation of existing data, we have separately conducted phase equilibrium, major and trace-element modeling of potential mantle and crustal sources in the Tianshuihai terrane and the SW Tarim Craton. Our new data and modeling results, in combination with regional sedimentary and structural records, allow us to (1) identify the mantle source of the Ordovician intermediate-mafic rocks in the southern Kunlun terrane, (2) determine the basement of Ordovician arc crust in the southern Kunlun terrane and (3) reconstruct the Ordovician subduction polarity and geodynamic evolution of the Proto-Tethys Ocean.

GEOLOGICAL BACKGROUND

The WKOB extends northwest-southeast along the northwest margin of the Tibetan Plateau and represents the northernmost orogenic collage within the Tethys domain. It is subdivided from north to south into the Northern Kunlun terrane (SW Tarim Craton), Southern Kunlun terrane and Tianshuihai terrane, separated by the Oyttag-Kudi and Mazar-Kangxiwa-Subashi suture zones, respectively (Fig. 1B; Xiao et al., 2003, 2005; Zhang et al., 2019a, 2021; Zha et al., 2022).

The Northern Kunlun terrane is an uplifted part of the SW Tarim Craton (Zhang et al., 2014), which features a Precambrian basement comprising the Paleoproterozoic Heluositan complex, Mesoproterozoic greenschist- to amphibolite-facies metamorphosed and intensely folded sedimentary sequences (the Kalakashi Group), mid-Neoproterozoic greenschist-facies metamorphosed and folded volcanic-sedimentary sequences and late Neoproterozoic unmetamorphosed carbonate-clastic-tillite sequences (Zhang et al., 2019b, 2019c). Devonian molasse, a Carboniferous–Permian shallow marine platform and Late Permian terrestrial facies

sedimentary rocks overlie the Precambrian basement, while granitic rocks with ages of 477–430 Ma are observed along several northwest-southeast-trending faults and record early Paleozoic magmatic processes (Fig. 1C; Zhang et al., 2019d, 2019e; Wang et al., 2022a, 2022b).

In the Oyttag-Kudi suture zone, ophiolitic mélanges are sporadically distributed from west to east in the Oyttag, Kegang and Kudi areas (Fig. 1B; Jiang et al., 2008; Li & Zhang, 2014). Notably, in the Kudi area, these ophiolitic rocks are mainly composed of tectonic slices of peridotite, dunite, cumulate pyroxenite and gabbro, pillowed and massive basalts, and have been interpreted as relics of the northern Proto-Tethys Ocean (Xiao et al., 2003; Yuan et al., 2005). Zircon U-Pb ages of ca. 525–494 Ma have been determined for these rocks (Li & Zhang, 2014), and their basaltic and gabbroic components exhibit geochemical characteristics of normal or enriched mid-ocean ridge basalt (N-MORB or E-MORB) and island-arc basalt, suggesting a formation in a supra-subduction zone (SSZ) setting (Wang et al., 2002; Yuan et al., 2005). Structural and geological cross sections across the Kudi ophiolite complex indicate that the structures of the southern Kunlun terrane are dominated by NW-SE trending thrust faults, tight folds and foliations (Xiao et al., 2002, 2003). In the northern part of the complex, the Yixieke Cambrian volcanic and sedimentary rocks mostly dip to the south and have been thrust northwards towards the accretionary prism in the southern side of Akaz Daban (Xiao et al., 2002, 2003). In the southern part of this area, the structural deformation is characterized by compressive deformation recorded by SE verging thrust faults and folds, resulting in southward thrusting of Cambrian Buziwan ultramafic rocks onto the Kudi gneiss-schist (i.e., early Paleozoic accretionary wedge; Xiao et al., 2003, 2005; Zhu et al., 2021a).

The southern Kunlun terrane comprises various tectonic assemblages, including Precambrian strata, Yixieke arc, Kudi gneiss, Xiananqiao arc, the Pushou-Menggubao ophiolite suites, seamount and continental fragments (incorporated into the Kudi accretion-subduction complex) (Xiao et al., 2003, 2005; Yuan et al., 2005; Cui et al., 2007; Wu et al., 2021a; Yang et al., 2022). The Precambrian strata comprise the Pushou Group and Sangzhutage Group in the northern part and the Saitula Group in the southern part, separated by the Pushou-Menggubao ophiolitic mélanges (Cui et al., 2007; Wu et al., 2021a). Early workers considered

the Precambrian strata in the southern Kunlun terrane to be an integral part of the Tarim Block or an exotic block accreted during the early Paleozoic (Xiao et al., 2003; Zhang et al., 2007), but recent models propose that the Saitula Group represents an early Paleozoic accretionary wedge (Zhang et al., 2007; Zhang et al., 2018a). The Pushou-Menggubao ophiolite comprises ultramafic rocks, gabbros, basalts, diabase and siliceous rocks. Some of the basalts share similar geochemical features with oceanic-island basalts (Wu et al., 2021a). The age of the Pushou-Menggubao ophiolite is suggested to be Early or Middle Ordovician (Cui et al., 2007). Widespread igneous rocks with age groups of ca. 530–400 Ma and 270–200 Ma (Zhang et al., 2007, 2018b; Liu et al., 2014a; Yin et al., 2020; Xu et al., 2024) indicate two distinct phases of orogeny in the southern Kunlun terrane (Xiao et al., 2003, 2005).

The Mazha-Kangxiwa-Subashi suture zone extends about 340 km from the Gongger in the west, via Mazha and Kangxiwa, to Subashi in the east (Fig. 1B). Typical ophiolitic rocks are mainly preserved in the southeastern part of the WKOB (Zhang et al., 2021; Zha et al., 2022). Geological mapping has revealed that these ophiolitic mélanges can be further divided into the Qimanyute ophiolite in the north and the Subashi ophiolite in the south (Fig. 1B; Zhang et al., 2021; Zha et al., 2022). The ophiolitic mélanges in the Qimanyute area are mainly composed of basalt, dolerite sheets, plagiogranite and gabbro with only minor serpentinite (altered peridotite) and pyroxenite (Han et al., 2002; Zhang et al., 2021). The sedimentary rocks are mainly gray siliceous rocks and limestone. The contacts between structural units are ductile shear zones. In addition, the basalt, gabbro and dolerite in the ophiolitic mélange were metamorphosed at greenschist facies. Recent work has revealed that the gabbros in the Qimanyute ophiolitic rocks show a compositional transition from oceanic plagiogranites to boninitic rocks, with subsequent Nb-enriched gabbros (ca. 494–485 Ma; Zhang et al., 2021). These magmatic sequences are likely to have been generated in response to the initiation of subduction of the Proto-Tethys Ocean (Zhang et al., 2021). To the south, the Subashi ophiolite is composed of lenticular serpentinite, peridotite, gabbro, dolerite, basalt and chert (Zha et al., 2022). The basalt, dolerite and gabbro from the Subashi ophiolite display U-Pb zircon ages of ca. 455 ± 4 Ma, 451 ± 6 Ma and 446 ± 6 Ma, respectively, revealing that the ophiolite was

formed in the Late Ordovician. The basaltic rocks in the Subashi ophiolite exhibit geochemical characteristics similar to those of N-MORB, and are interpreted as remnants of the southern Proto-Tethys Ocean (Zha et al., 2022).

The nature of the Tianshuihai terrane is controversial. Xiao et al. (2002, 2003, 2005) have proposed it is a massive accretionary prism containing thick flysch sequences recording the Paleozoic–Mesozoic subduction-related orogenic process between the southern Kunlun and Karakorum terrane. Yuan et al. (2003) have argued it is part of the southern Kunlun terrane, while others propose it to be independent (Qiao et al., 2015; Hu et al., 2016). High-grade, metamorphosed sedimentary rocks in the Tianshuihai terrane, previously thought to be from the Paleoproterozoic Bulunkuole Group and Saitula Group, have a controversial deposition age ranging from Paleoproterozoic to Middle Cambrian (Zhang et al., 2007; Gao et al., 2013). Some studies report Early–Late Cambrian (ca.530–485 Ma) plutons and volcanic rocks in the Tianshuihai terrane, indicating Early Paleozoic magmatism (Gao et al., 2013; Zhu et al., 2016; Hu et al., 2017).

FIELD OBSERVATIONS AND SAMPLE DESCRIPTION

We systematically investigated and collected 12 samples from two distinct plutons (Pushou pluton and Datong pluton) in the southern Kunlun terrane, including 3 gabbros, 4 syenites and 5 monzogranites.

The Pushou gabbros (Fig. 2b), located in the northern part of the Pushou area, exhibit a massive and fine- to medium-grained texture. The mineral assemblage of the massive gabbros is dominated by plagioclase (45 vol.%), clinopyroxene (35 vol.%), orthopyroxene (8 vol.%), biotite (10 vol.%), amphibole (~2 vol.%), and accessory minerals such as magnetite, ilmenite, titanite and apatite (Fig. 3a-b). The Datong monzogranites, the predominant lithology in the Datong complex (Fig. 2a), have massive texture, are medium-grained and grey in colour. Their mineral assemblage comprises perthitic K-feldspar (35 vol.%), plagioclase (30 vol.%), quartz (25 vol.%), interstitial biotite (5 vol.%), and muscovite (1–5 vol.%). Accessory minerals include zircon, apatite, monazite, allanite, fluorite and Fe–Ti oxides (Fig. 3c-d). The Datong

syenites were sampled at the northern margin of the Datong complex batholith (Fig. 2a). They are typically fine- to medium-grained, primarily composed of plagioclase (60–65 vol.%), K-feldspar (~30 vol.%) and biotite (<5 vol.%), with minor quartz. Additionally, they contain accessory minerals including titanite, apatite and zircon (Fig. 3e-f).

ANALYTICAL METHODS

Electron Microprobe analysis

Mineral compositions were determined at the Institute of Geology, China Academy of Geological Sciences (CAGS), Beijing, using a JXA-8100 electron microprobe. Analyses were carried out at a 15 kV accelerating voltage, a 20 nA beam current and a focused (nominal 1–5 μm) beam. Natural minerals and synthetic pure oxides from SPI Supplies Inc. (West Chester, PA, USA) were used as standards. For pyroxene, the calibration standards used were hornblende (for Si, Ti, Al, Fe, Ca, Mg, Na and K), fayalite (for Mn), and Cr_2O_3 (for Cr). For plagioclase, the standards used were hornblende (for Si, Ti, Al, Fe, Ca and Mg), albite (for Na), orthoclase (for K), and fayalite (for Mn). The analytical errors are generally less than 2%.

Zircon U–Pb dating and trace element analyses

Zircon grains were separated by conventional heavy liquid and magnetic techniques, followed by hand picking. Representative grains, together with the zircon standards, were cast into an epoxy mount and then polished to expose the grain centers. All zircon grains were photographed in transmitted and reflected light, as well as cathodoluminescence (CL), and imaged to study their internal structures.

The LA–ICP–MS zircon U–Pb isotopic compositions and trace elements were performed using an Agilent 7500a ICP–MS with an attached 193 nm excimer ArF laser–ablation system (GeoLas Plus) at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG–CAS), following standard procedures described by Xie et al. (2008). Zircon standards 91500 and GJ–1 were used as external and internal standards, respectively. NIST SRM610 glass was used to optimize the instrument, and common Pb was corrected according to the

method proposed by Andersen (2002). Trace elements were calibrated using NIST SRM 610 as an external standard and using $\text{ZrO}_2 = 66.1 \text{ wt\%}$ as an internal standard. Isoplot (version 4.5) was used to draw U–Pb age Concordia diagrams and weighted mean calculations (Ludwig, 2003).

Zircon Lu–Hf isotope analysis

In situ zircon Lu–Hf isotope analyses were undertaken using a Nu Plasma HR MC–ICP–MS (Nu Instruments), coupled to a 193 nm excimer laser ablation system (Resolution M–50, Resonetics LLC), installed in the IGG–CAS. Lu–Hf isotopic analyses were made on the same spots as the U–Pb isotopic analyses. Depending on the zircon size, a spot size of 60 μm or 40 μm was used for analysis, with a laser repetition rate of 6 Hz. The detailed analytical procedures for the laboratories were provided by Xie et al. (2008). The measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratios were normalized to $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$, using exponential correction for mass bias. During analysis, the $^{176}\text{Hf}/^{177}\text{Hf}$ and $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of the standard zircon (91500) were 0.282294 ± 15 (2σ , $n = 20$) and 0.00031, which is in good agreement with the low peaks of $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282284 ± 22 measured by Griffin et al. (2006).

Zircon O isotope analysis

Zircon oxygen isotope analysis was measured using a Cameca IMS–1280 HR SIMS at the State Key Laboratory of Isotope Geochemistry, the Guangzhou Institute of Geochemistry, Chinese Academy of Sciences (SKLaBIG, GIG–CAS). The Cs^+ primary ion beam with an intensity of $\sim 2 \text{ nA}$ was accelerated at 10 kV, and the size of analytical spots was about 20 μm in diameter. Oxygen isotopes were measured in multi-collector mode using two off-axis Faraday cups. Total analytical time per spot was ~ 3.5 mins, including 30 s of pre-sputtering and 120 s of automatic tuning of the secondary beam, as well as 64 s of analysis. The measured oxygen isotopic values were corrected for instrumental mass fractionation (IMF) using the standard Penglai zircon ($\delta^{18}\text{O}_{\text{VSMOW}} = 5.3 \pm 0.10 \text{ ‰}$, (2σ); Li et al., 2010). A weighted mean of $\delta^{18}\text{O} = 5.49 \pm 0.09 \text{ ‰}$ (2σ) was obtained from 16 measurements of the Qinghu zircon standard,

within analytical error of the recommended value ($\delta^{18}\text{O}_{\text{VSMOW}} = 5.39 \pm 0.22 \text{ ‰}$, 2σ ; Li et al., 2013). Detailed analytical procedures are provided by Yang et al. (2018).

Whole-rock major and trace element analysis

Whole-rock samples were washed and trimmed in order to remove weathered surfaces. The samples were then pulverized and powdered in an agate mill to about 200 mesh for major and trace elements analysis. Major-element oxides were analyzed with fused glass beads using a Rigaku RIX 2000 X-ray fluorescence spectrometer at the SKLaBIG GIG-CAS. The detailed analytical procedures are described by Yuan et al. (2010). The analytical uncertainties are generally better than 5% for all elements. Trace-element compositions were determined using an ELAN DRC-e ICP-MS at the State Key Laboratory of Ore Deposit Geochemistry, Institute of Geochemistry, Chinese Academy of Sciences, following procedures described by Liang et al. (2000). About 40 mg of each powdered sample was dissolved in a high-pressure Teflon bomb for 2 days at 190 °C using a mixture of HF + HNO₃ + HClO₄ (1:1:0.2). Analytical precision is 2–5%.

Whole-rock Sr–Nd isotope analysis

Whole-rock Sr and Nd isotopic compositions of selected samples were determined using an MC-ICP-MS at SKLaBIG, GIG-CAS. Analytical procedures are similar to those described by Wei et al. (2002) and Liang et al. (2002). All isotopic ratios were corrected for fractionation using a $^{86}\text{Sr}/^{88}\text{Sr}$ ratio of 0.1194 and a $^{146}\text{Nd}/^{144}\text{Nd}$ ratio of 0.7219. Reported $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were further adjusted relative to the NBS SRM 987 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.71025$) and the Shin Etsu JNdi-1 standard ($^{143}\text{Nd}/^{144}\text{Nd} = 0.512115$), respectively. The basaltic standard BHVO-2 was used to monitor ion-exchange chromatographic purification, and yielded an $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.703483 ± 0.000014 (2 standard error, SE) and a $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.512978 ± 0.000008 (2SE), consistent with the certified values of $^{87}\text{Sr}/^{86}\text{Sr} = 0.703476 \pm 0.000012$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512972 \pm 0.000008$, respectively (Raczek et al., 2003).

Thermodynamic and trace element modeling

Thermodynamic modeling was performed using a Gibbs free energy minimization approach with the software Perple_X (version 6.9.1) to constrain the melting temperature and pressure. Based on the mineral assemblages and bulk rock compositions, we used a system of $\text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{O}_2$ (NCKFMASHTO) and the hp633ver data set, assuming $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ value as 0.1 (Pourteau et al., 2020). The P–T phase diagram was calculated with 2 wt.% and 4 wt.% water, corresponding to water-undersaturated and water-saturated melting at 5–8 kbar. The following activity-composition relations were used for modeling: silicate melt, garnet, muscovite, biotite, orthopyroxene (White et al., 2014), amphibole (Green et al., 2016), plagioclase and K-feldspar (Fuhrman & Lindsley, 1988), clinopyroxene (Holland & Powell, 1996), spinel (White et al., 2002) and ilmenite (White et al., 2000). Kyanite, sillimanite, rutile, quartz and H_2O were considered as pure phases. The equilibrium mineral assemblages and melt compositions were calculated at discrete P–T points for every 10 °C and 0.1 GPa using a Perple_X-based program Rcrust (Mayne et al., 2016). The P–T estimation uncertainty is about ± 0.1 GPa and ± 50 °C at the 2-sigma level (Palin et al., 2016).

Based on the weight percentages of melt and residue minerals obtained from thermodynamic modeling, we further modeled trace element compositions of felsic melts through batch melting equation (Shaw, 1970):

$$C_{\text{melt}}/C_{\text{source}} = 1/[D + F*(1 - D)]$$

Where C_{melt} and C_{source} represent the concentration of a trace element in the melt and source rock, respectively; D is the bulk partition coefficient, and F is the degree of melting (i.e., the weight percentage of melt). The partition coefficients of minerals used in the batch melting model are given in Supplementary Data Table S16.

RESULTS

Mineral chemistry

Representative major element compositions of clinopyroxene, feldspar, apatite and biotite

from the Pushou gabbro and Datong monzogranite are presented in Fig. 4 and Supplementary Data Tables S2–5.

Clinopyroxene

Clinopyroxene phenocrysts from Pushou gabbro sample 21DW04 exhibit low MgO (12.4–13.1 wt.%), CaO (20.6–22.2 wt.%) and Mg[#] (67–71), together with high Al₂O₃ (1.9–2.6 wt.%) and FeO^T (9.9–11.3 wt.%). The clinopyroxenes plot as diopside–augite (Fig. 4a), with a chemical composition of Wo = 42.7–45.7 mol%, En = 35.8–37.4 mol%, and Fs = 15.7–18.8 mol%. Crystallization temperatures and pressures were estimated using thermobarometers for hydrous systems from Putirka (2008), with results listed in Supplementary Data Table S2. Thermobarometry calculations using the clinopyroxene–melt compositions yield estimates of the potential temperature (T_{Cpx}) of the mantle source ranging from 1175 to 1185 °C (average 1179 °C) and equilibration pressures of 7.1 to 8.0 kbar (average 7.6 kbar), employing the equation of Putirka (2008).

Biotite

Biotite in the Pushou gabbro is mainly Mg-biotite (Fig. 4b), exhibiting relatively high concentrations of SiO₂ (37.1–38.0 wt.%), MgO (14.5–16.0 wt.%), Mg[#] (66–70) and TiO₂ (4.9–6.4 wt.%). In the Datong syenites, biotites are predominantly Mg-biotite, with MgO (15.5–16.4 wt.%) and Mg[#] (68–69) and low FeO^T (12.8–13.3 wt.%). Biotite in the Datong monzogranite ranges from Fe²⁺-biotite to Mg-biotite (Fig. 4b), characterized by low SiO₂ (36.6–37.4 wt.%), MgO (10.7–11.4 wt.%) and Mg[#] (49–52), and high Al₂O₃ (15.4–16.0 wt.%), FeO^T (18.9–19.6 wt.%) and TiO₂ (2.1–2.5 wt.%). Overall, these biotites have negligible CaO (0–0.05 wt.%) and consistent FeO^T/(FeO^T + MgO) ratios (Pushou gabbro of 0.4–0.45; syenites of 0.60–0.64; monzogranite of 0.4–0.5). These characteristics suggest that all analyzed biotites are primary and have not undergone significant chloritisation due to post-magmatic deuteric alteration or meteoric fluid circulation (Kumar & Pathak, 2010).

The crystallization temperature and pressure of biotite in granitic magma were estimated

using the machine learning geothermometer and geobarometer developed by Li & Zhang (2022). The results are listed in Supplementary Data Table S3. The calculated temperature for Datong syenites ranges from 933 to 948 °C (average 941 °C) and the calculated pressure ranges from 6.5 to 7.7 kbar (average 7.1 kbar). Datong monzogranites give crystallization temperatures ranging from 867 to 874 °C (average 871 °C) and pressures between 5.7 and 7.0 kbar (average 6.4 kbar).

Feldspar

Plagioclases from the Pushou gabbro are andesine ($An_{48-51}Ab_{48-51}Or_{1-2}$; Supplementary Data Table S4; Fig. 4a), with low SiO_2 and Na_2O of 55.8–56.4 wt.% and 5.4–5.8 wt.%, respectively, and high Al_2O_3 and CaO of 27.1–27.4 wt.% and 9.85–10.37 wt.%, respectively. Plagioclase phenocrysts in the Datong syenites are predominantly albite ($An_{0.6-1.1}Ab_{98-99}Or_{0.3-1.3}$; Fig. 4c), showing higher SiO_2 and Na_2O (67.8–68.8 wt.% and 11.6–11.7 wt.%, respectively), and lower Al_2O_3 and CaO (19.0–19.3 wt.% and 0.14–0.24 wt.%, respectively), compared to those in the Pushou gabbro. Additionally, K-feldspars from both the Datong syenites and monzogranites are sanidine ($An_{0-0.2}Ab_{3-8}Or_{91-97}$; Fig. 4c), featuring SiO_2 (64.4–65.3 wt.%), Al_2O_3 (17.8–18.1 wt.%), K_2O (15.6–16.7 wt.%) and Na_2O (0.3–4.0 wt.%). No compositional zonation was observed in any feldspar phenocrysts.

The plagioclase–liquid hygrometer allows for the quantitative estimation of magmatic H_2O contents by consideration of plagioclase–liquid exchange between the anorthite and albite components (An_{17-95} ; Zellmer et al., 2013; Waters & Lange, 2015). By applying the method of Waters & Lange (2015) with a temperature input of 1180 °C and pressure of 7.6 kbar for the Pushou gabbro (derived above), we obtained values of 3.1–3.4 wt.% H_2O , with a mean of 3.2 wt.% H_2O , supporting the inference of a high- H_2O parent magma.

Apatite

Apatite grains from the monzogranite (sample WK1610) display relatively scattered CaO (46.4–54.7 wt.%) and P_2O_5 (36.0–39.5 wt.%), high F (2.8–3.3 wt.%) and low Cl (0.01–0.02 wt.%) contents (Supplementary Data Table S5), classifying them as fluorapatite (Fig. 4d; Zhang

et al., 2020).

Zircon U–Pb geochronology

Representative cathodoluminescence (CL) images of zircons from specimens of the Pushou gabbro (21DW04), Datong syenites (WK1612) and Datong monzogranite (WK1610) are presented as insets in Fig. 5. The U–Pb isotopic data are listed in Supplementary Data Table S6.

Zircons from the Pushou gabbro (21DW04) range from 150 to 300 μm in length (Fig. 5a). They are euhedral with weak, broad zoning and lack complex internal structures in CL images (Fig. 5a), which are typical features of zircon grains formed in mafic magmas. Their high Th (58–1766 ppm) and U (170–1572 ppm) contents with elevated Th/U ratios (0.3–1.1; Fig. 5d) are indicative of a magmatic origin (Hoskin & Schaltegger, 2003). Twenty-three analyses of 25 grains reveal $^{206}\text{Pb}/^{238}\text{U}$ ages from 474 to 494 Ma with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 483 ± 3 Ma (MSWD = 1.2), interpreted as the igneous crystallization age (Fig. 5a). The remaining two analyses yielded ^{207}Pb corrected $^{206}\text{Pb}/^{238}\text{U}$ ages of 528 and 529 Ma and are interpreted as xenocrystic.

Zircon crystals from the Datong monzogranite samples (WK1610) are also euhedral with numerous narrow oscillatory zones, and from ~ 100 to 150 μm in length (Fig. 5b). They exhibit similar Th (210–790 ppm) and U (513–1506 ppm) and Th/U ratios between 0.2 and 0.8 (Fig. 5d), consistent with a magmatic origin. Nineteen analyses yield $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 466 to 473 Ma, with a weighted mean of 469 ± 2 Ma (MSWD = 0.36) (Fig. 5b) consistent with the zircon U–Pb age (470 ± 1 Ma) reported in a previous study (Gao et al., 2013). This age is interpreted as the igneous crystallization age.

Zircon crystals from the Datong syenites (WK1612) are long and prismatic (~ 100 to 300 μm in length) with most showing oscillatory zoning (Fig. 5c). They possess elevated Th (235–1000 ppm) and U (342–1353 ppm) with Th/U ratios exceeding 0.1 (Fig. 5d), indicating a magmatic origin. Twenty-three analyses yield concordant $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 456 to 465 Ma, also consistent with that reported in previous studies (479–443 Ma, Jiang et al., 2002;

Liao et al., 2010; Wang et al., 2017; Zhu et al., 2018). The weighted mean age of 461 ± 2 Ma (MSWD = 0.39) (Fig. 5c) is considered to represent the crystallization of the Datong syenites.

Magmatic zircon trace element characteristics

Magmatic zircon trace element data are listed in Supplementary Data Table S7. Decreases in Th/U and increases in Hf and Yb/Gd are typically interpreted to reflect the cooling and fractionation crystallization of silicic magma; the Eu/Eu* variation may be associated with either feldspar in the residue or the fractionation of the magma (Wooden et al., 2007; Claiborne et al., 2010). Zircons from the Pushou gabbro, Datong syenite and monzogranites all exhibit consistent trends wherein Yb/Gd increases while Th/U, Eu/Eu* and Ti decrease with increasing Hf contents (Fig. 6a–d).

Recent research has suggested that zircon Eu/Eu* (mainly serving as a proxy for water content) is the most reliable oxygen fugacity (fO_2) indicator for magmatic rocks in Phanerozoic terranes. Loucks et al. (2020) have introduced a novel magmatic oxybarometer to estimate fO_2 of magma, employing the trace-elements Ce, Eu and Ti in zircon. This method mitigates the significant temperature influence on these ratios by incorporating the Ti content of the zircons. Zircons from Pushou gabbro (Sample 21DW04) show high fO_2 values ($-0.9 < \text{FMQ (fayalite-magnetite-quartz buffer)} < +1.3$; an average of +0.6; see Table S6 for results), indicating a relatively oxidized parent magma. The Datong syenites and monzogranites also have high fO_2 values ($+1.0 < \text{FMQ} < +1.8$ (average +1.4) and $+0.7 < \text{FMQ} < +2.7$ (average +1.5), respectively).

Whole-rock geochemistry

Pushou gabbros

The Pushou gabbros exhibit low SiO₂ (47.4–49.1 wt.%), high K₂O (2.0–3.0 wt.%) and high alkali contents (K₂O+Na₂O = 4.6–5.0 wt.%; Supplementary Data Table S8), plotting in alkali fields on the SiO₂ vs. (K₂O+Na₂O) diagram (TAS; Fig. 7a) and the shoshonite field on the K₂O vs. SiO₂ diagram (Fig. 7b). These samples exhibit MgO contents from 5.5 to 6.9 wt.%

with variable $Mg^{\#}$ values of 49–59, and V of 205–251 ppm, Cr of 18–46 ppm and Ni of 6–23 ppm. In contrast, they exhibit relatively high and constant Al_2O_3 (17.2–17.8 wt.%), CaO (10.0–10.1 wt.%) and TiO_2 (1.1–1.4 wt.%) contents, but low $FeO^T/(FeO^T+MgO)$ ratios (0.47–0.50). Chondrite-normalized rare earth element (REE) diagrams illustrate moderate REE fractionation ($(La/Yb)_N = 9.3–10.2$) and moderate Eu anomalies ($Eu/Eu^* = 0.67–0.75$) (Fig. 8a). The primitive mantle-normalized patterns reveal enrichment in large ion lithophile elements (LILE, e.g., Rb, Ba, Th, U and K) and relative depletion in high field strength elements (HFSE, e.g., Nb, Ta and Ti) (Fig. 8b).

Datong syenites

Datong syenite samples have moderate SiO_2 (59.5–61.4 wt.%; Supplementary Data Table S8) and high alkalis ($K_2O + Na_2O = 8.1–10.2$ wt.%), plotting mostly in the syenite field on the TAS diagram (Fig. 7a). High K_2O (3.3–4.2 wt.%) places them in the shoshonite series (Fig. 7b). The rocks are characterized by high Al_2O_3 (18.8–19.7 wt.%) and CaO (3.7–5.1 wt.%) with a metaluminous nature ($A/CNK = 0.87–0.93$) and magnesian characteristics ($FeO^T/(FeO^T + MgO) = 0.58–0.61$). Chondrite-normalized REE patterns demonstrate enrichment in LREE [$(La/Yb)_N = 19.3–30.1$] and moderate Eu anomalies ($Eu/Eu^* = 0.75–0.79$) (Fig. 8c). These rocks are also enriched in LILE (e.g., Rb, Ba, Th and U) and show negative anomalies in HFSE (Nb, Ta and Ti) on primitive mantle-normalized trace element patterns (Fig. 8d).

Datong monzogranites

The Datong monzogranites are characterized by high SiO_2 (69.7–72.6 wt.%; Supplementary Data Table S8), high K_2O (5.5–7.0 wt.%), total alkali contents (8.4–10.0 wt.%) and K_2O/Na_2O (1.5–2.2), categorizing them as alkali (Fig. 7a) and shoshonitic (Fig. 7b). These rocks are metaluminous to peraluminous (Fig. 7c), with A/CNK ratios of 0.97–1.15 (mostly < 1.1; Fig. 7c). They display low Al_2O_3 (13.8–14.7 wt.%), CaO (0.7–1.2 wt.%), MgO (0.6–0.7 wt.%), P_2O_5 (0.12–0.15 wt.%) and $FeO^T/(FeO^T + MgO)$ ratios (0.62–0.66). They have high zircon saturation temperature ($T_{Zr} = 816–848$ °C; Supplementary Data Table S8), high

Zr+Nb+Ce+Y values (360–428 ppm) and plot in the field of the A-type granites (Fig. 7d). Chondrite-normalized REE diagrams reveal obvious enrichment in LREE $[(La/Yb)_N = 26.5–30.5]$ and significant negative Eu anomalies ($Eu/Eu^* = 0.54–0.57$; Fig. 8e). These rocks exhibit enrichment in LILE but with large negative Ba and Sr anomalies and moderate to pronounced negative anomalies in HFSE (Fig. 8f).

Zircon Hf–O isotope compositions

The results of in-situ analyses of Hf isotopes in zircons are listed in Supplementary Data Table S9, and the corresponding data are illustrated in Fig 9a. The Hf isotopic compositions of the Pushou gabbro are homogeneous with negative $\epsilon_{Hf}(t)$ ranging from -5.3 to -7.7 . These values are comparable to those of the Cambrian–Ordovician mafic rocks in the Tianshuihai terrane and SW Tarim Craton (Fig. 9a; Zhang et al., 2019e; Wang et al. 2021a; Sui et al., 2021). By comparison, the Datong syenites exhibit relatively heterogeneous Hf isotopic compositions, with negative to slight positive $\epsilon_{Hf}(t)$ values (-3.8 to $+2.6$), overlapping the field of Early Paleozoic mafic rocks in the Tianshuihai terrane and SW Tarim Craton (Fig. 9a). Zircons from the Datong monzogranite demonstrate enriched Hf isotopic compositions with $\epsilon_{Hf}(t)$ from -11.0 to -16.5 , similar to the Precambrian basement beneath the Tianshuihai terrane and Tarim Craton (Fig. 9a; Zhang et al., 2018c, 2019b).

Measured zircon $\delta^{18}O$ values for Pushou gabbro (Supplementary Data Table S10) exhibit a limited range from $+7.6$ ‰ to $+7.9$ ‰, averaging $+7.7$ ‰ (2SD). In contrast, Datong monzogranites have (relatively) variable zircon $\delta^{18}O$ values from $+7.2$ ‰ to $+8.0$ ‰, with a mean of $+7.6$ ‰(2SD) (Fig. 9b).

Whole-rock Sr–Nd isotope compositions

The Sr–Nd isotope compositions of the Pushou gabbros, Datong syenites and monzogranites are presented in Supplementary Data Table S11. The Pushou gabbros display relatively high $(^{87}Sr/^{86}Sr)_i$ of $0.7156–0.7192$ and negative $\epsilon_{Nd}(t)$ values of -7.1 to -7.8 , and plot in the field of the Tianshuihai terrane and SW Tarim Craton (Fig. 9c). The Datong syenites

are enriched and homogeneous, with $(^{87}\text{Sr}/^{86}\text{Sr})_i$ of 0.7103 to 0.7105 and negative $\epsilon\text{Nd}(t)$ values of -3.8 to -4.3 , similar to the coeval mafic rock in the WKOB (Fig. 9c; Zhang et al., 2019e; Sui et al., 2022; Wang et al., 2021a). Datong monzogranites have higher $(^{87}\text{Sr}/^{86}\text{Sr})_i$ of 0.7321–0.7323 and more negative $\epsilon\text{Nd}(t)$ of -11.3 to -11.8 , comparable with the Nd isotopic compositions of the Precambrian basement beneath the Tianshuihai terrane and the Tarim Craton (Fig. 9c; Zhang et al., 2018c, 2019b).

DISCUSSION

Crustal Contamination and Fractional Crystallization

The initial compositions of arc magmas are determined by the mantle source compositions, slab-derived components (including fluids and melts) and the P-T conditions of partial melting. Several factors, such as shallow-level crustal assimilation and fractional crystallization (AFC), can modify these initial compositions during ascent. Crustal contamination can lead to variations in elemental and isotopic compositions of arc magmas (Zheng & Hermann, 2014; Bezard et al., 2015). In contrast, closed-system fractional crystallization produces elementally-evolved rock types but does not significantly affect radiogenic isotope compositions. In general, crustal components exhibit distinctly low $\epsilon\text{Nd}(t)$ and high $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios (Rudnick & Fountain, 1995). Any crustal assimilation during magma ascent would likely increase $(^{87}\text{Sr}/^{86}\text{Sr})_i$ and decrease $\epsilon\text{Nd}(t)$ values in the magma suites. However, the Pushou gabbro and Datong syenites show limited Sr-Nd isotopic variation (Fig. 9c) and slight decrease of $(^{87}\text{Sr}/^{86}\text{Sr})_i$ at relatively consistent $\epsilon\text{Nd}(t)$ with increasing SiO_2 (Figs. 10c–d), providing robust evidence against significant pervasive crustal contamination.

On a diagram of $(\text{La}/\text{Sm})_N$ vs. La and SiO_2 content (Fig. 10a and b), Pushou gabbro samples show relatively constant $(\text{La}/\text{Sm})_N$ (4.3–4.8) but variable La content (33.5–44.0 ppm), suggesting that some crystal fractionation may have occurred during magma ascent. In general, mantle-derived primary melts have $\text{Ni} > 300$ ppm, $\text{Cr} > 750$ ppm and $\text{Mg}^\# > 68$ (Herzberg & O'Hara, 2002). Compared with these primary magma compositions, the Pushou gabbro rocks

have low $Mg^\#$ (49–59), Ni (<23 ppm) and Cr (<46 ppm) contents, indicating that olivine and clinopyroxene fractionation might indeed have occurred. The relatively constant TiO_2 with decreasing MgO (Fig. 11b) indicates negligible involvement of Fe-Ti oxides. Also, the limited range of CaO (10.01–10.10) with variable SiO_2 and moderate Eu anomalies, suggest that minor fractionation of plagioclase also occurred. Collectively, moderate fractionation of olivine and clinopyroxene with a little plagioclase did not result in marked fractionation of incompatible trace elements in the Pushou gabbros, as shown by their subparallel trace-element distribution patterns (Fig. 7b). Thus, the ratios of incompatible elements can be used to help identify the nature of the mantle source.

The low MgO contents and $Mg^\#$ values of the Datong shoshonitic rocks relative to those of primitive mantle-derived magmas suggest different degrees of crystal fractionation. Fractionation of Fe–Ti oxides (e.g., ilmenite or magnetite) in the shoshonitic rocks possibly occurred, as suggested by negative correlations of FeO^T and TiO_2 with SiO_2 (Fig. 11c, and 11e). The negative Eu anomalies ($Eu/Eu^* = 0.63–0.92$; Supplementary Table S8), together with the negative correlations of SiO_2 with Eu/Eu^* (Fig. 11g), indicate significant fractionation of plagioclase. Additionally, the decrease in $(Dy/Yb)_N$ ratios with increasing SiO_2 (Fig. 11h) suggests that fractionation of amphibole played an important role.

In summary, although affected by fractional crystallization, Pushou gabbro and Datong shoshonitic rocks underwent relatively little crustal contamination. Thus, the geochemical and isotopic characteristics (e.g., Sr–Nd–Hf–O isotopes) of these rocks can be used to constrain the nature of the mantle source.

Subduction polarity — Tarim Craton or Tianshuihai protoliths?

Previous studies have shown that Ordovician magmatic rocks in the southern Kunlun terrane have enriched Sr–Nd–Hf isotopic compositions compared with those of the Cambrian magmatic rocks (Figs. S1a–c; Liu et al., 2014a; Wang et al., 2021a), suggesting that the original depleted upper mantle/lower crust beneath the southern Kunlun terrane may have been partially modified or replaced by enriched mantle and/or crustal sources during the Ordovician (Tao et

al., 2022, 2024; see details in the Supplemental Material). Considering the interpretations of an Ordovician arc setting in the WKOB, the magmatic rocks in the southern Kunlun terrane have been correlated with either the southward subduction of the Proto-Tethys Ocean slab beneath the southern Kunlun terrane and Tianshuihai terrane (Liao et al., 2010; Zhang et al., 2018a, 2018b), or the northward subduction of Proto-Tethys Ocean slab beneath the southern Kunlun terrane and the SW Tarim Craton (Yuan et al., 2003; Xiao et al., 2003, 2005; Zha et al., 2022; Yang et al., 2022; Tao et al., 2022). According to these authors, there are two contrasting blocks that could provide potential isotopically enriched mantle or crustal sources, the Tianshuihai terrane and the Tarim craton.

Therefore, the petrogenesis of the Ordovician magmatic rocks in the southern Kunlun terrane becomes instructive for evaluating the subduction polarity of the Proto-Tethys Ocean. In one scenario, these Ordovician magmatic rocks are linked to mantle or crustal sources of the Tianshuihai terrane, which would require the southward subduction. Alternatively, they could be derived from magma sources beneath the SW Tarim Craton, and thus be attributable to northward subduction. In the latter case, the parental magmas are derived from the mantle wedge of the SW Tarim Craton and the felsic magmas would have been derived from the partial melting of crustal material within the SW Tarim Craton. Therefore, clear identification of the protolith affinities (Tarim Craton vs. Tianshuihai terrane) may provide direct evidence for the subduction polarity of the Proto-Tethys Ocean. Unfortunately, the two alternative protoliths exhibit similar, enriched Sr–Nd–Hf isotopic compositions (Zhang et al., 2018c; Zhang et al., 2010, 2014; See Supplementary Table S1 for details), which are appropriate as the source of the Ordovician igneous rocks in the southern Kunlun terrane (Fig. 9a and c). Therefore, Sr–Nd–Hf isotopes cannot be used to accurately identify the magmatic source and so determine the subduction polarity of the Proto-Tethys Ocean.

With the continued development and improvement of thermodynamic databases (Holland & Powell, 2011; Green et al., 2016), thermodynamic modeling is now becoming an effective approach to investigating the evolution of magmatic rocks, including phase equilibrium constraints, melt fractions and compositions, as well as proportions and compositions of solid

residues under a wide range of P–T conditions (e.g., Mayne et al., 2020; Pavan et al., 2021). Furthermore, based on mass balance and trace element partition coefficients, trace element compositions of melts and solid residues (or fractional minerals) can also be modeled (e.g., Pourteau et al., 2020; Johnson et al., 2021). Therefore, comparison between natural samples and modeled melt compositions might allow the distinction between isotopically-similar source end-members. Below, we explore thermodynamic and trace element modeling of protolith compositions from the two potential source regions, the Tianshuihai terrane (Zhang et al., 2016; Sui et al., 2021) and the SW Tarim Craton (Wang et al., 2022b; Zhang et al., 2010, 2019f).

Source of the Pushou gabbros

The gabbros in the Pushou area exhibit low SiO₂ (47.4–49.1 wt.%) and high MgO (5.5–6.9 wt.%) relative to continental crust (Rudnick & Gao, 2003) or crust-derived melts (Patiño Douce & Beard, 1995), indicating that they were derived from a mantle source. These gabbros also display pronounced negative Nb-Ta-Ti anomalies on primitive mantle-normalized trace element diagrams (Fig. 8b), which suggests metasomatism by subduction-related fluids or melts from an altered oceanic slab (Guo et al. 2005; Huang et al., 2010; Zhu et al., 2017) and/or subducted sediments (Gao et al., 2007; Conticelli et al., 2008; Liu et al., 2015).

In general, fluids possess a high capacity for transporting water-soluble elements but a low capacity for transporting water-insoluble ones, while hydrous melts exhibit a high capacity for transporting both (Zheng, 2012). As shown in Figs. 12a–b, the relatively uniform Rb/Nb and Rb/Y ratios with variable Th/Yb ratios of the Pushou gabbros indicate that their mantle source experienced metasomatism by subduction-related melts rather than fluids (Kepezhinskis et al., 1997). Because Hf is also generally regarded as immobile in slab-derived fluids, a selective enrichment of Nd relative to Hf sometimes occurs, leading to decoupling of Hf–Nd compositions toward less radiogenic $\epsilon\text{Nd}(t)$ (Pearce et al., 1999) with positive $\Delta\epsilon\text{Hf}(t)$ (Vervoort et al., 2011). The positive $\Delta\epsilon\text{Hf}(t)$ value of the Pushou gabbros (+4.33) might, therefore, indicate source contamination by zircon-free pelagic sediments (Chauvel et al., 2008). An alternative - selective melting of a mantle source containing high Lu/Hf minerals (e.g., Choi

& Mukasa, 2012) - cannot lead to the conspicuous elevation of O isotope compositions (Wang et al., 2014a) to +7.6 to +7.9 ‰, considerably higher than those of the primary mantle ($\delta^{18}\text{O} = 5.3 \pm 0.3$ ‰; Valley, 2003) (Fig. 9b). A surface-derived component is therefore indicated, likely to have been oceanic sediments ($\delta^{18}\text{O}$ ca. 20–40 ‰; Bindeman et al., 2005). Pushou gabbros may have been derived from a mantle source similar to those of the local ophiolitic basalts but metasomatized by sediments. However, binary mixing curves (Fig. 9b) indicate that the Hf isotopes of the Pushou gabbros are too enriched (negative) to have been generated by partial melting of a depleted upper-mantle source metasomatized by any reasonable proportion of sediments (mostly < 5 ‰; Chauvel et al., 2008). In contrast, a mixing model of already-enriched mantle and oceanic sediments could explain the Hf–O isotope compositions of the Pushou gabbro samples, with a realistic amount of sediment-derived melt (1–5 ‰, Fig. 9b). We, therefore, propose that prior to modification by subducted sediment-derived melts, the mantle wedge was comparable with enriched rather than depleted mantle beneath the southern Kunlun terrane.

A lack of significant Eu anomalies and low $(\text{La/Yb})_{\text{N}}$ (9.26–10.21) and $(\text{Dy/Yb})_{\text{N}}$ (1.4–1.5) ratios in the Pushou gabbro indicate a low abundance or even absence of plagioclase and garnet in the source region. Thermobarometry calculations using clinopyroxene compositions result in estimates of a potential temperature (T_{Cpx}) of the mantle source ranging from 1175 to 1185 °C (average 1179 °C) and equilibration pressures of 7.1 to 8.0 kbar (average 7.6 kbar) (Putirka, 2008), suggesting their derivation from the source region with spinel as a significant residual phase. According to the standard equation of Shaw (1970) for non-modal melting, the Pushou gabbros were likely produced by 15–17% partial melting of a spinel-garnet lherzolite mantle source (Fig. 12c; see Supplementary Table S12). The presence of amphibole and biotite (Fig. 3b) in the Pushou gabbro indicates a relatively high H_2O content (>3 wt.%; Frost & Frost, 2008; Scaillet et al., 2016), strongly suggesting water-saturated melting. The high zircon $f\text{O}_2$ values (mean $\Delta\text{FMQ} = +0.6$) also indicate a relatively oxidized parent magma. Thus, we propose that this rock series likely originated from moderate degree (~15–17 ‰; Fig. 12c) partial melting of a heterogeneous, hydrous and oxidized lherzolite-mantle source.

To further explore possible mantle sources and differentiate between Tarim and Tianshuihai protoliths, we performed trace element calculations employing a non-modal batch melting model ($C_L/C_i = 1/(F + (1 - F)*D)$; Shaw, 1970), a common process in the generation of mantle-derived magmas (e.g., Kinzler & Grove, 1992; Walter, 1998). Partition coefficients are from Pilet et al. (2011) (Supplementary Data Table S12), and other modeling parameters and results are described in Supplementary Data Table S12. Considering that only 1–5% of sediment melt was added to the source of the Pushou gabbro, the Nd isotopic composition of the potential mantle source should be broadly similar to or slightly more depleted than that of Pushou gabbro, which has low $\epsilon Nd(t)$ (–7.1 to –7.8; Supplementary Data Table S11). A mafic sample from the Tianshuihai terrane (15ZK11; $\epsilon Nd_{(480\text{ Ma})} = -5.6$; Sui et al., 2022) and a mafic dyke from the SW Tarim Craton (2018KL112H6; $\epsilon Nd_{(480\text{ Ma})} = -5.8$; Zhang et al., 2019f) were used to back-calculate trace element compositions of potential primary magmas from each (See Supplementary Table S12 for details). Model results (Fig. 12d) suggest that mantle magmas equivalent to the mafic rocks in the Tianshuihai terrane are not a potential mantle source because the calculated non-modal batch melting lines do not closely fit the range of trace element compositions of Pushou gabbro, particularly with respect to heavy REE and Y abundances. In contrast, the modeling results of the mafic dyke sample from the Tarim Craton are closely comparable with one of the trace element compositions of the Pushou gabbro (Fig. 12d). These results, therefore, provide geochemical support for the involvement of Tarim mantle in the sources of the Pushou gabbros. These gabbros are most likely derived from a low degree (~15 %) of partial melting of subduction-modified enriched Tarim mantle sources in the spinel-garnet facies (with added 1–5 % sediment melts) (Fig. 12d; McKenzie & O’Nions, 1991).

The Mesoproterozoic Hf–Nd model ages of the Pushou gabbro samples suggest that their mantle sources were affected either by an old metasomatic event beneath the SW Tarim Craton during the Mesoproterozoic and Neoproterozoic period, or by young metasomatism with “older” materials. Although these two processes cannot be easily differentiated, the following geological observations are consistent with the second explanation: (1) Mesoproterozoic mafic dykes (Zhang et al., 2019b) and Neoproterozoic metabasaltic rocks (Zhang et al., 2010) in the

SW Tarim Craton show OIB- or MORB-like trace-element patterns without signs of subduction-related metasomatism; (2) the paucity of Early Cambrian–Early Ordovician (530–485 Ma) subduction-related records (Fig. 1c) in the SW Tarim Craton indicates that the Proto-Tethys Ocean did not subduct northward before the Ordovician period (Tao et al., 2024). These observations imply that the enriched lithospheric mantle was not metasomatized by subduction-related components before the Early Ordovician. Therefore, the subduction-modified trace element signals of lithospheric mantle sources for the Pushou gabbro were likely imparted by recycled oceanic sediments shortly before its generation.

Source of the Datong shoshonitic rocks

The Datong shoshonitic rocks (including the Datong syenite) exhibit moderate silica ($\text{SiO}_2 = 54.4\text{--}65.0$ wt.%), high potassium ($\text{K}_2\text{O} = 2.8\text{--}6.6$ wt.%) with variable MgO, Cr and Ni (Jiang et al., 1999, 2002; Liao et al., 2010; Wang et al., 2017; Zhu et al., 2018), and therefore are not directly derived from typical mantle rocks. Three proposed genetic models include (1) mixing between mantle-derived basaltic and lower crust or/and subducted sediments-derived felsic magmas (Liao et al., 2010; Chen et al., 2012; Hao et al., 2019), (2) partial melting of potassic lower crust (Sen & Dunn, 1994; Ferreira et al., 2015; Wang et al., 2017), and (3) fractional crystallization of subducted fluids/melts metasomatic mantle-derived high-K and shoshonitic magmas (Jiang et al., 1999, 2002; Liu et al. 2014b; Soder et al. 2018; Zhu et al., 2018; Zhu et al., 2021b). The magma mixing model is considered unlikely due to the lack of mafic microgranular enclaves (Fig. 3c) and consistent Sr–Nd isotopic compositions (Fig. 10d). Experimental studies have shown that dehydration melting of meta-basaltic rocks can produce granitoid magmas (Rapp et al., 1999). However, the relatively high $\text{Mg}^\#$ (35–56) and low SiO_2 are dissimilar to the experimental melts ($\text{Mg}^\# < 40$; Patiño Douce & Johnston, 1991; Rapp & Watson, 1995; Sisson et al., 2005), and they also exhibit distinct Sr–Nd–Hf isotopic compositions from coeval lower crust-derived shoshonitic rocks in the Datong area (Datong monzogranite—see below).

Conversely, fractional crystallization of high-Mg basalt to basaltic andesite at 2–10 kbar

and 700–1230 °C is known to produce basaltic-andesitic-rhyolitic magmas (Fig. 12a; Sisson & Grove, 1993; Grove et al., 2003; Nandedkar et al., 2014; Müntener & Ulmer, 2018; Ulmer et al., 2018; Lee & Bachmann, 2014). In the Harker diagrams (Fig. 11a–f), the data follow a differentiation trend similar to those of wet liquid lines of descent controlled by the fractional crystallization of plagioclase, amphibole and Fe-Ti oxides from primitive basaltic magmas (Fig. 11b; Deng et al., 2019; Liu et al., 2022). Importantly, the Datong shoshonitic rock geochemical signature (e.g., shoshonitic and magnesian) and Sr–Nd isotopic compositions align with those of coeval gabbro in the southern Kunlun terrane (Fig. 9c), interpreted to be derived from partial melting of metasomatized lithospheric mantle (Jiang et al., 2002; Zhang et al., 2019e, 2021; Wang et al., 2021a, 2022a). Nd–Hf isotopic decoupling (Fig. 9d) and relatively constant Rb/Y ratios with variable Th/Yb ratios suggest that their mantle source was metasomatized by subduction-related melts (Figs. 9d and 12b). Therefore, the Datong shoshonitic rocks likely result from the fractional crystallization of high-K and shoshonitic basaltic melts (i.e., Ordovician mafic rocks), derived from the partial melting of lithospheric mantle metasomatized by subduction-related melts (Fig. 9d).

To validate the crystal fractionation hypothesis of mantle-derived basaltic melts, we employed the thermodynamically calibrated Rhyolite-MELTS program (Gualda et al. 2012). The crystallization conditions were established at a pressure below 7.0 kbar, as estimated from biotite barometers (Table S3; Li & Zhang, 2022) and oxygen fugacity based on zircon oxygen barometer ($\text{zircon } f\text{O}_2 = \Delta\text{QFM} + 1.4$; Table S7). Additionally, the initial H₂O content was set at 3.0 to 5.0 wt.%, representing water-undersaturated and water-saturated differentiation, respectively. Gabbro sample 17WP48 from the southern Kunlun terrane with high MgO (5.0 wt.%) and Mg[#] of 55 was selected as the parental melt (Fig. 12c–f; Wang et al., 2021a). Under these conditions, the Datong shoshonitic rocks can be effectively reproduced by 10–47 % crystal fractionation of amphibole (10–35 %) + plagioclase (0–8 %) + magnetite (0.3–4.0 %) (Figs. 11c–f and 13a).

In addition to the southern Kunlun terrane, high-K and shoshonitic mafic rocks also occur in the Tianshuihai terrane and the SW Tarim Craton. They share similar Sr–Nd isotopes with

the Datong shoshonitic rocks (Fig. 9c; Ye et al., 2008; Sui et al., 2021; Wang et al., 2022a, 2022b). Utilizing the Rhyolite-MELTS modeled mineral assemblage and corresponding partition coefficients, plus the simple fractional crystallization equation ($C_L/C_i = F^{(D-1)}$; Shaw, 1970) the evolving trace-element abundances were calculated from appropriate parents of each terrane - mafic rock samples with highest MgO and similar $\epsilon\text{Nd}(t)$ values - 15ZK10 (Tianshuihai terrane; $\epsilon\text{Nd}(t) = -4.1$; Sui et al., 2022) and wp17-86 (SW Tarim Craton; $\epsilon\text{Nd}(t) = -3.99$; Wang et al., 2022b). As shown in Fig. 13b, parental magma from the Tianshuihai terrane would lead to a greater enrichment in HREEs (Dy, Y, Yb and Lu) and depletion in HFSEs (Th, U, Nb, and Ta). However, modelled parental melts from the SW Tarim Craton are in good agreement with the trace-element abundances of the Datong shoshonitic rocks. Thus, we propose that the Datong shoshonitic rocks are more likely derived from crystal fractionation of mafic magmas in the SW Tarim Craton.

Genesis of the Datong monzogranites

The Datong monzogranite is classified as an aluminous A-type granite due to the following characteristics: (1) high Zr+Nb+Ce+Y concentrations (Fig. 7d; 360–428 ppm; Whalen et al., 1987; Frost et al., 2001); (2) high total alkali contents (Fig. S2a; $\text{Na}_2\text{O}+\text{K}_2\text{O} = 8.4\text{--}10.1$ wt.%; Frost & Frost, 2010); and (3) high zircon saturation temperatures (T_{Zr}), apatite saturation temperatures (T_{ap}) and biotite-melt temperatures ($T_{\text{Biotite-melt}}$) (Fig. S2b; $T_{\text{Zr}} = 815\text{--}848^\circ\text{C}$, $T_{\text{ap}} = 922\text{--}967^\circ\text{C}$ and $T_{\text{Biotite-melt}} = 860\text{--}871^\circ\text{C}$). The petrogenesis of A-type granites is still a subject of considerable debate (e.g., Barker et al., 2013; Namur et al., 2011; Papoutsas et al., 2016). However, it is generally accepted that they represent high-temperature, low-pressure magmas that crystallized at relatively low melt- H_2O concentrations (Collins et al., 1982; Frost et al., 2001; Frost & Frost, 2010; King et al., 1997, 2001). Various petrogenetic processes have been proposed, including: (1) direct fractionation of mantle-derived basaltic magma (Eby et al., 1992); (2) mixing between crust- and mantle-derived magmas (Yang et al., 2006); (3) partial melting of granulitic metasedimentary or igneous rocks (Whalen et al., 1987; Huang et al., 2011); and (4) partial melting of pre-existing calc-alkaline granitoids at high temperatures and

shallow depths (Patiño Douce, 1997; King et al., 1997).

Although A-type granites are sometimes associated with large volume of a contemporaneous mafic-intermediate rocks, the Datong examples are dominated by felsic rocks. They have high silica ($\text{SiO}_2 = 69.7\text{--}72.6$ wt.%), low MgO (0.6–0.7 wt.%), Cr (3–14 ppm) and Ni (2–9 ppm), along with evolved Sr–Nd–Hf radiogenic isotope compositions. These argue against significant mantle input. Additionally, there is no evidence of magma mixing between mantle-derived and crustal magmas, such as the presence of mafic xenoliths/enclaves, mingling textures, and bimodal or scattered distributions of zircon $\epsilon\text{Hf}(t)$ values (Fig. 2). Therefore, we propose that the A-type granites in the Datong area were derived from the partial melting of the continental crustal protoliths.

Derivation from high-temperature melting of granulitic metasedimentary rocks (Huang et al., 2011) or from the melting of metasedimentary restites after previous melt extraction (Gao et al., 2014) can be excluded for the following reasons: (1) the Datong A-type granites have high alkalis (8.4–10 wt.%), low A/CNK values (mostly < 1.1) and relatively low $\delta^{18}\text{O}$ values (7.2–8.0 ‰), arguing against derivation from the partial melting of granulitic metasedimentary rocks, (2) a felsic granulite residue after a previous melt extraction, proposed as the source of A-type granites (Collins et al., 1982), would be enriched in Ca and Al and depleted in K and Si (Creaser et al., 1991), and (3) the absence of granulitic terrains in the Datong area. On the other hand, Frost & Frost (2010) proposed that partial melting of granitoids (tonalites and granodiorites) is a plausible genetic mechanism for high-silica, calc-alkaline A-type granites. Importantly, the Nd–Hf isotopic compositions of the Datong A-type granites resemble Neoproterozoic tonalites and Mesoproterozoic granodiorites (Fig. 9a–c), considered as the Precambrian basement in the Tianshuihai terrane and Tarim Craton, respectively (Zhang et al., 2016, 2018c, 2019b). Therefore, once again, both are potential candidates for the crustal sources of the Datong A-type granites.

Thermodynamic and trace element modeling provide additional constraints. The biotite-melts thermometer of Li & Zhang (2022) suggests crystallization at 867–874 °C (average 871 °C) (Table S3), the zircon saturation thermometer of Watson & Harrison (1983) gives a range of

816–848 °C, with an average of 831 °C (Table S8), but the apatite saturation temperatures (T_{ap}) of Watson & Harrison (1983) yield a relatively high range of 922–967 °C (average 944 °C) (Table S8). Considering the absence of inherited zircon in the samples, zircon temperatures represent a minimum estimate (Miller et al., 2003). In contrast, the negative relationship between P_2O_5 and SiO_2 (Fig. 11a) indicates that the parent magma was saturated with phosphorus (Lee & Bachmann, 2014) and thus the apatite saturation temperature is more robust. For estimating emplacement pressure, the biotite-melt geobarometers of Li & Zhang (2022) provide ranges of 5.7–7.0 kbar (average 6.4 kbar). Thus, P-T estimates suggest that the Datong monzogranite was emplaced at a temperature of ~850–950 °C and pressure of 5–7 kbar, representing the minimum temperature and pressure of source melting. Furthermore, experimental studies show that the products of dehydration melting of calc-alkaline granitoids at low pressures ($P \leq 4$ kbar) are in accordance with all of the lithochemical features of A-type granites, whereas most of the geochemical features diagnostic of A-type granites at relatively high pressures (≥ 8 kbar) are lost (Patiño Douce, 1997). Meanwhile, the studied A-type granites have lower $\text{FeO}^{\text{T}}/(\text{FeO}^{\text{T}} + \text{MgO})$ (0.62–0.66) and $10000 \times \text{Ga}/\text{Al}$ ratios (2.04–2.10) compared to typical A-type granites. Thus, experimental studies suggest a maximum pressure of around 8 kbar. Collectively, we argue that the studied A-type granites were formed by the partial melting of granitic rocks (tonalites or granodiorites) at high temperature ($> \sim 950$ °C) and low-pressure (5–8 kbar) conditions.

Based on these P–T estimates, we carried out phase equilibrium modeling for the partial melting of Neoproterozoic tonalites (Zhang et al., 2018c) and Mesoproterozoic granodiorites (Zhang et al., 2019b) from the Tianshuihai terrane and the SW Tarim Craton, respectively (Fig. 14). Initial H_2O contents were set at 2 wt.% and 4 wt.%, to represent water-undersaturated and water-saturated melting (Fig. 14). The residual mineral assemblages, melting degrees and major element compositions of melts were determined using thermodynamic calculations of Connolly (2009) (see Analytical Methods and Table S14). Subsequent trace element modeling was performed using the batch melting model (Shaw, 1970) for Tianshuihai Neoproterozoic tonalites (Sample 15D070H6; Zhang et al., 2018c) and Tarim Mesoproterozoic granodiorites

(2016KL09H10; Zhang et al., 2019b). Results are summarized in Figs. 14 and 15, and indicate that the A-type granites were likely generated by the partial melting of the Precambrian basement in the SW Tarim Craton (Mesoproterozoic granodiorites; Figs. 14a–b and 15a–b) rather than Tianshuihai terrane (Neoproterozoic tonalites; Figs. 14c–d and 15c–d). The calculated residual mineral assemblage from modeling consists of plagioclase, orthopyroxene and ilmenite in proportions of ~6:3:1 (Fig. 14; Table S14). Melting at varying higher pressures (for example, 9 kbar) fails to account for the high HREE contents in our samples (Fig. 15b). In a relatively wet system (4 wt.% H₂O), similar degrees of partial melting can be achieved at lower temperatures (850–900°C) and at 5 to 7 kbar (Fig. 15a–b), but the resulting melts have higher CaO and lower K₂O contents compared to the studied A-type granites (Fig. 15a). Our results therefore demonstrate that partial melting at high temperature ($T > 950^{\circ}\text{C}$), low-pressure ($P = 5\text{--}8$ kbar) and anhydrous ($\text{H}_2\text{O} = \sim 2$ wt.%) conditions of Mesoproterozoic granodiorites in the SW Tarim Craton can generate melts with major and trace element compositions consistent with those of the A-type granites in the southern Kunlun terrane.

Geodynamical Interpretations

Subduction Polarity

It has been noted above that Ordovician magmatic rocks are widespread in the WKOB, but their tectonic context remains controversial (e.g., Xiao et al., 2003; 2005; Yuan et al., 2005; Zhu et al., 2018; Zhang et al., 2018a, 2019e). Some researchers propose that the Proto-Tethys Ocean subducted southward beneath the southern Kunlun terrane and the Tianshuihai terrane (Liao et al., 2010; Liu et al., 2014a; Zhang et al., 2018a, 2019e; Wang et al., 2021a). This interpretation is mainly based on the presence of voluminous early Paleozoic arc magmas and oceanic accretionary complex in the southern margin of the Oyttag-Kudi suture zone (Liao et al., 2010; Zhang et al., 2018a, 2019e; Wang et al., 2021a; Zhang et al., 2021). However, recent high-precise geochronological data have also documented widespread early Paleozoic magmatism during ca. 470–420 Ma in the SW Tarim Craton to the north (Zhang et al., 2019d; Wang et al., 2021a, 2022a), which is consistent with northward subduction (Xiao et al., 2003,

2005; Yuan et al., 2005; Tao et al., 2022) or bi-directional subduction (Zhu et al., 2018; Wang et al., 2021a; Xu et al., 2021).

As discussed in the petrogenesis section, our studies on the early Ordovician (483 Ma) gabbros in the southern Kunlun terrane suggest that they were derived from low-degree partial melting of the enriched Tarim mantle sources metasomatized by subducted sediment melts. Additionally, the geochemical compositions and Nd–Hf isotopic decoupling in the middle Ordovician (461 Ma) shoshonitic rocks exhibit typical arc-type affinities (Fig. 9d), and elemental relationships indicate their likely derivation from these via crystal fractionation of the Tarim mantle-derived basaltic magmas. Thus, the original depleted upper mantle beneath the southern Kunlun terrane must have been partially modified by enriched Tarim mantle sources before the early Ordovician (ca. 483 Ma; Fig. 16a). Thermodynamic, major and trace element modeling of the Datong monzogranites also suggests that these were most likely generated by the partial melting of the Precambrian basement in the SW Tarim Craton (Figs. 15a–b), suggesting that the deep crust beneath southern Kunlun terrane had also been replaced by the Precambrian basement of the SW Tarim Craton in the Ordovician.

Importantly, the source characteristics of Ordovician magmatic rocks in the southern Kunlun terrane exhibit a systematic spatial variation, with the influence of sub-continental lithospheric mantle beneath the Tarim gradually increasing from south to north (Tao et al., 2024). These features imply that Ordovician magmatic rocks in the southern Kunlun terrane were related to mantle or crustal sources from the Tarim craton. Therefore, we consider that the Ordovician magmatic rocks in the southern Kunlun terrane represent the magmatic arc of an ocean basin (e.g., southern Proto-Tethys Ocean) developed in the southern margin of the southern Kunlun terrane, and that northward subduction was responsible for the consumption of the southern Proto-Tethys Ocean during the Ordovician (Fig. 16). This explanation is consistent with previous studies on the Ordovician–Silurian sedimentary, ophiolitic mélange rocks and structural records within the southern Kunlun terrane (Li et al., 2018; Chen et al., 2019; Zhu et al., 2021a; Wang et al., 2021b; Chang et al., 2023; Tao et al., 2024). For example, the late Cambrian–early Ordovician sedimentary strata in the southern Kunlun terrane exhibit

a single detrital zircon age population of 490–520 Ma (Zhu et al., 2021a; Wang et al., 2021b), indicating their primary derivation from local Cambrian magmatic rocks. In contrast, detrital zircons from Ordovician–early Silurian (ca. 481–430 Ma) sedimentary rocks show a predominant age population of 450–470 Ma, with two subordinate clusters around 840–780 Ma and 1100–1140 Ma (Li et al., 2018; Chen et al., 2019; Chang et al., 2023). These age patterns are similar to those of detrital zircons from the SW Tarim Craton (Wang et al., 2014b) but differ from those in the Tianshuihai terrane (Yang et al., 2022). This change indicates a major provenance shift from a single to multiple source regions, suggesting that the final closure of the northern Proto-Tethys Ocean along the Oyttag-Kudi suture occurred between the 490 and 481 Ma (Tao et al., 2024).

In contrast, the N-MORB type ophiolitic rocks with U-Pb zircon ages of 455–446 Ma in the Subashi area (Zha et al., 2022), and 465 Ma OIB seamount assemblage in the Kangxiwa area (Yang et al., 2022), indicate the existence of a mature ocean basin (i.e., southern Proto-Tethys Ocean) on the south of the southern Kunlun terrane during the Ordovician period. Furthermore, structural analysis demonstrates that the Ordovician subduction complex is marked by a series of north-dipping ductile thrusts along the southern margin of the southern Kunlun terrane, also consistent with contemporaneous northward subduction of the southern Proto-Tethys Ocean (Xiao et al., 2002, 2003; Wang, 2004).

Some researchers have also argued that the Proto-Tethys Ocean subducted southward beneath the Tianshuihai terrane during the Ordovician, based on the presence of Early Paleozoic arc magmatism in the Tianshuihai terrane (e.g., ca. 531 Ma monzogranites, Yin et al., 2020; ca. 490 Ma gabbro, Liu et al., 2019). Hence, a bidirectional subduction model was further proposed to explain the generation of magmatic rocks in the WKOB during the Ordovician period (Zhu et al., 2018; Wang et al., 2021a; Xu et al., 2021). However, the absence of Early Ordovician–Early Silurian arc magmatism in the Tianshuihai terrane (Fig. 1c; Yang et al., 2022; Tao et al., 2024), coupled with the middle-late Ordovician sedimentary succession (e.g., Dongguashan Group) in the Tianshuihai terrane as passive margin sediments (Liu, 2015) indicates that the Proto-Tethys Ocean did not subduct southward beneath the Tianshuihai terrane during the

Ordovician period (Yang et al., 2022; Zha et al., 2022). In addition, the youngest known arc magmatic rocks are the ca. 490 Ma gabbros (Liu et al., 2019) in the Tianshuihai terrane, implying that the southward subduction of the Proto-Tethys Ocean terminated around 490 Ma. Thus, it is reasonable to infer that the northward subduction of the southern Proto-Tethys Ocean gave rise to the southern Kunlun terrane and SW Tarim Craton arc magmatic system during Ordovician.

Tectonic and Magmatic Evolution

Given the development of multiple early Paleozoic ophiolite suture zones in the WKOB, including three approximately NW-trending zones (from north to south): the ca. 525–494 Ma Oyttag-Kudi (Huang et al., 2014; Li & Zhang, 2014), the ca. 487–480 Ma Menggubao-Pushou-Qimanyute (Cui et al., 2007; Zhang et al., 2021) and the 455–446 Ma Mazha-Kangxiwa-Subashi suture zones (Fig. 1b; Zha et al., 2022; Yang et al., 2022), the WKOB probably involved episodic accretion and collision processes as a result of successive closure of branches of the Proto-Tethys Ocean (Xiao et al., 2003, 2005; Tao et al., 2024).

The northern Proto-Tethys Ocean represented by the Oyttag-Kudi ophiolites is considered to have subducted southward since ca. 530 Ma (Fig. 4A1), based on the ca. 525–494 Ma Kudi SSZ-type ophiolite (Li & Zhang, 2014) and arc magmatism in the south (Xiao et al., 2002; Jiang et al., 2008; Zhang et al., 2019a) as well as the paucity of ca. 530–490 Ma subduction-related records in the northern Kunlun terrane. Subsequently, the northern Proto-Tethys Ocean was closed at ca. 490–485 Ma (Tao et al., 2024), resulting in formation of a series of northwest-verging imbricated thrusts along the northern margin of the southern Kunlun terrane (Xiao et al., 2003, 2005). In contrast, the magmatic sequence of plagiogranites (ca. 494 Ma), boninites (ca. 487 Ma) and Nb-enriched gabbros (ca. 485 Ma) in the Qimanyute Ophiolite indicates the formation of a new subduction zone within the Proto-Tethys Ocean during the early Ordovician (Zhang et al., 2021). It has been demonstrated above that the parental magmas of the Pushou gabbros most likely originated from enriched mantle, metasomatized by fluids derived from the

subducted oceanic slab. Therefore, the Pushou gabbros and the Qimanyute Ophiolite reveal that the northward subduction of the southern Proto-Tethys Ocean initiated at ~485 Ma (Fig. 16a).

The modeling described above suggests that the middle Ordovician (ca. 470 Ma) A-type granites were likely generated by partial melting of the Precambrian basement in the SW Tarim Craton at high temperature ($T > \sim 950$ °C), low-pressure ($P = 5\text{--}8$ kbar) and anhydrous ($\text{H}_2\text{O} = \sim 2$ wt.%) conditions. A-type granites are commonly associated with extensional settings, such as post-collision, intra-plate extension and/or subduction-related extension (i.e., slab roll-back; Dall'Agnol et al., 2012 and references therein). According to the discussion above, the A-type granites in the southern Kunlun terrane are related to the evolution of the southern Proto-Tethys Ocean, so slab roll-back and breakoff of the Proto-Tethys Ocean crust are feasible geodynamic mechanisms that could trigger partial melting of the low crust and widespread high-temperature, low-pressure magmatism (Xiao et al., 2003, 2005; Gao et al., 2013; Wu et al., 2021a).

In general, slab roll-back involves the descending slab migrating backwards, creating the necessary rheology for the asthenosphere to ascend to the subduction corner. In this case, the resulting magmatic activity becomes younger toward the trench (Yin et al., 2020; Wang et al., 2021a). However, the relatively narrow, linear zone of magmatism in the middle Ordovician (Fig. 1b) in the northern margin of the southern Kunlun terrane suggests that the A-type granites might not be related to slab roll-back. Additionally, slab roll-back would have released water, fertilizing the mantle wedge and leading to water-flux melting, producing a significant amount of calc-alkaline arc-type magmas, in contrast to the widespread middle Ordovician alkaline felsic rocks observed (e.g., Datong and Yirba pluton; Yuan et al., 2003; Li et al., 2018; Zhang et al., 2019d, 2019e; Wu et al. 2021a). Further, the widespread absence of the Middle-Upper Ordovician strata in the SW Tarim Craton probably signifies a significant unconformity, possibly in response to uplift and a contractional events in the SW Tarim Craton during the middle Ordovician (Wu et al., 2021b). This suggests a dynamic mechanism that can generate crust thinning and extension in a contractional collision environment.

Therefore, a slab break-off model is more likely responsible for the generation and emplacement of the A-type granites and associated granitoids in the southern Kunlun terrane

(Fig. 16b). Slab break-off typically occurs in the early stages of continental collision, due to a decrease in the subduction rate induced by the positive buoyancy of the continental lithosphere being introduced into the subduction zone. Since the Ordovician accretionary complex associated with the subduction of the southern Proto-Tethys Ocean contains abundant continental fragments and seamounts (e.g., Kudi gneiss-schist complex; Xiao et al., 2002; Yang et al., 2022; Sang et al., 2022), we further suggest that a continental fragment or seamount-continental margin collisional model might be appropriate. Subsequently, these processes can cause subducted slab break-off when the subducted oceanic lithosphere exceeds the positive buoyancy of the attached continental/seamount lithosphere. As a result, the upwelling of hot asthenosphere through the slab window will trigger partial melting of Precambrian basement rocks, producing intensive high-temperature, low-pressure, and anhydrous A-type granites (e.g., studied A-type granite, Yirba pluton; Yuan et al., 2003; Li et al., 2018; Wu et al., 2021a). Moreover, driven by collision dynamics and/or the physical differences between continental/seamount and oceanic crust, the residual southern Proto-Tethys oceanic lithosphere began to sink into the subcontinental lithospheric mantle and induced arc magmatism (e.g., Datong syenites; Fig. 16c) on the southern margin of the southern Kunlun terrane. Meanwhile, this hypothesis can also explain the transition of Ordovician magmatism from the north to the south of the Pushou-Qimanyute Ophiolites (Figs. 1b and 16c) both before and after the slab break-off (ca. 470 Ma).

CONCLUSIONS

Three lithological types of Ordovician igneous rocks have been investigated in the southern Kunlun terrane of the WKOB, NW Tibet: (1) Pushou gabbro, (2) Datong syenite, and (3) Datong monzogranite. LA-ICP-MS zircon U–Pb dating indicates that these igneous rocks were emplaced in the Early Ordovician (483 ± 3 Ma), Middle Ordovician (461 ± 2 Ma), and Middle Ordovician (469 ± 2 Ma), respectively.

The Pushou gabbros exhibit typical arc-like geochemical signatures (e.g., Nb–Ta–Ti anomalies, and Rb–U peaks), enriched Sr–Nd–Hf isotopic compositions, as well as anomalous

zircon $\delta^{18}\text{O}$ values. They were derived from low-degree ($\sim 15\%$) partial melting of subduction-modified Tarim mantle source (with an additional 5% sediment melts) under the spinel-garnet facies.

The Datong syenites are characterized by moderate SiO_2 , with high-K to shoshonitic affinity and enriched Sr–Nd–Hf isotopic compositions, coupled with relatively low $\text{Mg}^\#$ values, Cr and, Ni contents. They were likely derived from the fractional crystallization of mantle-derived basaltic magma in the Tarim Craton, involving plagioclase, amphibole, and Fe-Ti oxides (e.g., magnetite).

The Datong monzogranites exhibit high SiO_2 , $\text{FeO}^{\text{T}}/(\text{FeO}^{\text{T}} + \text{MgO})$, HFSE concentrations, and high zircon saturation temperatures (T_{Zr}), typical of A-type granite, coupled with enriched Sr–Nd–Hf isotope compositions and high zircon $\delta^{18}\text{O}$ values, formed by water-deficient (~ 2 wt.%) partial melting of the Precambrian basement (Mesoproterozoic granodiorites) in the SW Tarim Craton at high temperature ($T > \sim 950^\circ\text{C}$), low-pressure ($P = 5\text{--}8$ kbar) conditions.

The petrogenesis of the gabbro, syenite and monzogranite provides robust evidence to infer that the original upper mantle and lower crust in the southern Kunlun terrane may have been modified by Tarim subcontinental lithospheric mantle and Precambrian basement, respectively, indicating that Ordovician magmatism in the southern Kunlun terrane was generated by the northward subduction of the southern Proto-Tethys Ocean during the Ordovician.

Northward subduction of the southern Proto-Tethys Ocean began at ~ 485 Ma. A-type granites represent the magmatic expression of the break-off of the southern Proto-Tethys Ocean slab, caused by the collision of continental fragment/seamount with the SW Tarim Craton during the middle Ordovician. Subsequent generation of late Ordovician Datong syenites occurred in response to resumed northward subduction of the southern Proto-Tethys Ocean along the southern margin of the southern Kunlun terrane.

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Data Availability Statement

Supplementary data are available at Journal of Petrology online or Mendeley Data repository (<https://data.mendeley.com/drafts/ds235n9sj7>).

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Figure captions

Fig. 1. (a) Tectonic map of eastern Pamir region in Western Tibet, showing the distribution of the main suture zones and faults (modified after Tang et al. (2020)), (b) Geological map of West Kunlun orogenic belt (WKOB; modified after Li et al., 2018), showing the main tectonic units and sample locations of compiled ages, (c) Zircon U-Pb geochronological framework of Paleozoic magmatism in the WKOB (modified after Tao et al., 2024).

Fig. 2 (a) Simplified distribution map of plutons in the Datong area, WKOB (after Li et al., 2018); (b) Simplified distribution map of plutons in the Pushou area, WKOB.

Fig. 3. Field and microscope photographs of (a, b) the Pushou gabbro; (c, d) Datong monzogranite, and (e, f) Datong syenite from the southern Kunlun terrane in the WKOB. Mineral abbreviations: Px, pyroxene; Amp, amphibole; Bt, biotite; Pl, plagioclase; Kf, potassic feldspar; Q, quartz.

Fig. 4 Mineral composition and classification diagrams for Pushou gabbro and Datong monzogranite. (a) Clinopyroxenes from the Pushou gabbro plotted on the enstatite–ferrosilite–diopside–hedenbergite quadrilateral (Morimoto, 1988); (b) Classification of biotite (Foster, 1960); (c) Plots of feldspar compositions (Smith & Brown, 1988); (d) Plots of Cl vs. F of apatite (Zhang et al., 2020).

Fig. 5 U–Pb isotope data for representative samples. (a) Pushou gabbro (sample 21DW04); (b) Datong monzogranite (sample WK1610); (c) Datong syenite (sample WK1612). (d) Th/U–Age diagram of zircons for the representative samples.

Fig. 6. Magmatic zircon trace element plots; (a) Eu/Eu* vs. Hf content; (b) Yb/Gd ratio vs. Hf content; (c) Th/U ratio vs. Hf content; (d) Ti vs. Hf content.

Fig. 7. Petrochemical classification and geochemical characteristics of samples from the southern Kunlun terrane; (a) TAS classification diagram (Le Bas et al., 1986); (b) K₂O vs. SiO₂ diagram (Gill, 1981); (c) A/NK vs. A/CNK diagram (Frost et al., 2001); (d) (Na₂O+K₂O)/CaO ratios vs. Zr+Nb+Ce+Y contents diagram (Whalen et al., 1987). Southern Kunlun Terrane: SKT; Northern Kunlun Terrane: NKT. The literature data for the Ordovician mafic rocks in the southern Kunlun terrane are from Zhang et al. (2019e); (2021) and Wang et al. (2021a); Ordovician granitic rocks in the northern Kunlun terrane are from Zhang et al. (2019d); the Datong shoshonitic rocks are from Liao et al. (2010); Wang et al. (2017); Zhu et al. (2018); Zhang et al. (2021).

Fig. 8. Chondrite-normalized REE patterns and primitive mantle-normalized multi-element patterns for the southern Kunlun terrane samples. Normalization values from Sun & McDonough (1989) for chondrite and primitive mantle. Background data sources are consistent with those in Fig. 7.

Fig. 9. (a) Zircon Hf isotopic characteristics for igneous rocks in the WKOB; (b) Plot of $\delta^{18}\text{O}$ vs. $\epsilon_{\text{Hf}}(t)$ for zircons of igneous rocks from the southern Kunlun terrane. The blue curve represents the mixing between enriched mantle peridotite and subducted sediments, and black curve represents the mixing between enriched mantle peridotite and depleted mantle. The enriched mantle source is represented by $\epsilon_{\text{Hf}}(t) = -5.4$; $\delta^{18}\text{O} = 7.5\text{‰}$ and oceanic sediment is represented by $\epsilon_{\text{Hf}}(t) = -12$; $\delta^{18}\text{O} = 20\text{‰}$. (c) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. $\epsilon_{\text{Nd}}(t)$ ($t = 485\text{ Ma}$) diagram; The curve represents mixing between depleted mantle-derived melts (as represented by the Qimanyute ophiolitic rocks) and subducted sediments (as represented by the global oceanic sediments (GLOSS) ($(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7161$; $\epsilon_{\text{Nd}}(480\text{ Ma}) = -4.62$) from Plank & Langmuir (1998); (d) Plot of $\epsilon_{\text{Hf}}(t)$ vs. $\epsilon_{\text{Nd}}(t)$ for the igneous rocks from the southern Kunlun terrane. Literature data for the Qimanyute Ophiolitic mélange rocks in the southern Kunlun terrane are from Zhang et al. (2021); the Ordovician mafic rocks in the southern Kunlun terrane are from Wang et al. (2021a), Zhang et al. (2019e) and Zhang et al. (2021); Neoproterozoic mafic rocks in the Tarim Craton from Zhang et al. (2019b) and Zhang et al. (2019f); ca. 1.8~1.2 Ga granitic rocks in the SW Tarim Craton are from Zhang et al. (2019b) and ca. 0.8 Ga tonalites in the Tianshuihai are from Zhang et al. (2018c). Other data sources as in Fig. 7.

Fig. 10. (a) La/Sm vs. La (ppm); (b) $(\text{La}/\text{Sm})_N$ vs. SiO_2 (wt.%); (c) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. SiO_2 (wt.%); (d) $\epsilon_{\text{Nd}}(t)$ vs. SiO_2 (wt.%). Background data sources are as in Fig. 7.

Fig. 11. Variations of selected major and trace elements and ratios vs. SiO_2 and MgO for the magmatic rocks from the southern Kunlun terrane. (a) Representative modeled residual liquids formed by equilibrium crystallization of dry and hydrous parental basalts, using Rhyolite-

MELTS (Lee & Bachmann, 2014); (b) Schematic illustration showing the effects of magmatic H₂O on the liquid line of descent (LLD; after Deng et al., 2019); (c-f) The Rhyolite-MELTS (Gualda et al., 2012; Ghiorso & Gualda, 2015) modeling of LLDs for fractional crystallization illustrates the effects of initial H₂O content, pressure, and oxygen fugacity buffers on magma differentiation; (g-i) Trace element ratios vs. SiO₂; Background data sources as in Fig. 7.

Fig. 12. (a) Rb/Nb vs. Th/Yb diagram; (b) Rb/Y vs. Th/Yb diagram; (c) Dy/Yb vs. La/Yb diagram; (d) Trace-element modeling results for nonmodal batch melting of spinel-garnet peridotite. The enriched lithospheric mantle (Tarim mafic dyke vs. Tianshuihai gabbro) mixed with ~5% subducted sediments to produce the metasomatized mantle wedge source of the Pushou gabbro. The melting curve for spinel-garnet stability field and spinel field, in terms of Dy/Yb vs. La/Yb ratios, was modeled as nonmodal batch melting using the trace element partition coefficients as those in Pilet et al. (2011). The average compositions of sub-continental lithospheric mantle as starting compositions (McDonough, 1990). See Supplementary Table S12 for details of the mixing and melting model as well as different end-member compositions on the mixing lines. Background data sources are as in Fig. 7.

Fig. 13. (a) Rhyolite-melts software modeling illustrating fractional crystallization of Ordovician mafic rocks in the southern Kunlun terrane; (b) Modeling trace element patterns for parental magmas from the Tianshuihai terrane and SW Tarim Craton, respectively.

Fig. 14. Simplified P-T phase diagram for the composition of (a, b) Mesoproterozoic granodiorites (Sample 2016KL09H10; SW Tarim Craton; Zhang et al., 2019b) and (c, d) Neoproterozoic tonalites (Sample 15D070H6; Tianshuihai terrane; Zhang et al., 2018c), calculated with water contents of 2 wt.% and 4 wt.%, corresponding to water-deficient and water-fluxed melting, respectively.

Fig. 15. Major and trace element patterns of granitic melts calculated at different P-T-X(H₂O)

conditions for the (a, b) Mesoproterozoic granodiorites (Sample 2016KL09H10; Zhang et al., 2019b) and (c, d) Neoproterozoic tonalites (Sample 15D070H6; Zhang et al., 2018c) from the SW Tarim Craton and the Tianshuihai terrane, respectively. Individual mineral K_d values are from Bédard, (2006), Philpotts & Schnetzler (1970) and Bea et al. (1994).

Fig. 16. Schematic block diagrams illustrating the Early–Late Ordovician tectonic evolution of the southern Proto-Tethys Ocean in the WKOB. OKOB: Oyttag-Kudi Ophiolitic Belt; PQOB: Pushou-Qimanyute Ophiolitic Belt. See the text for details.

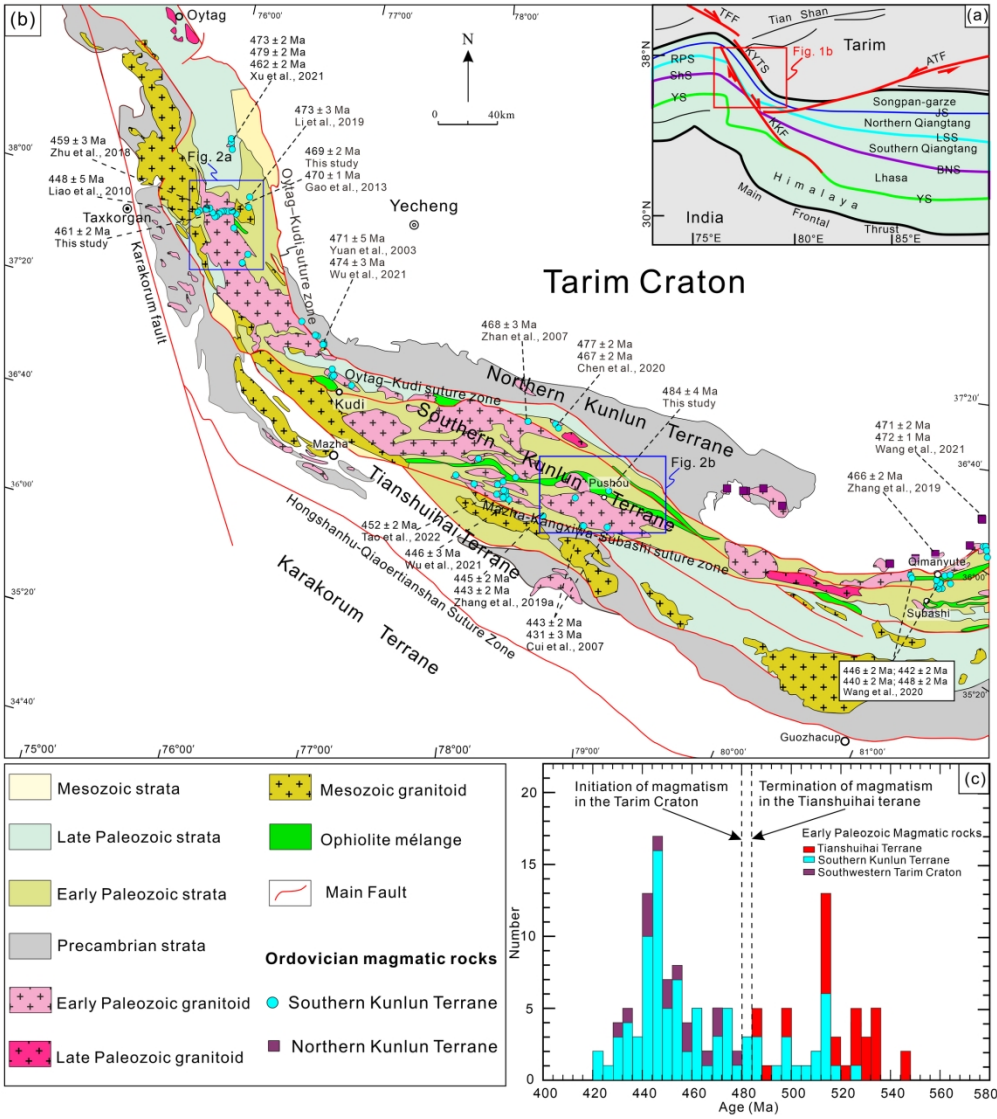


Figure 1

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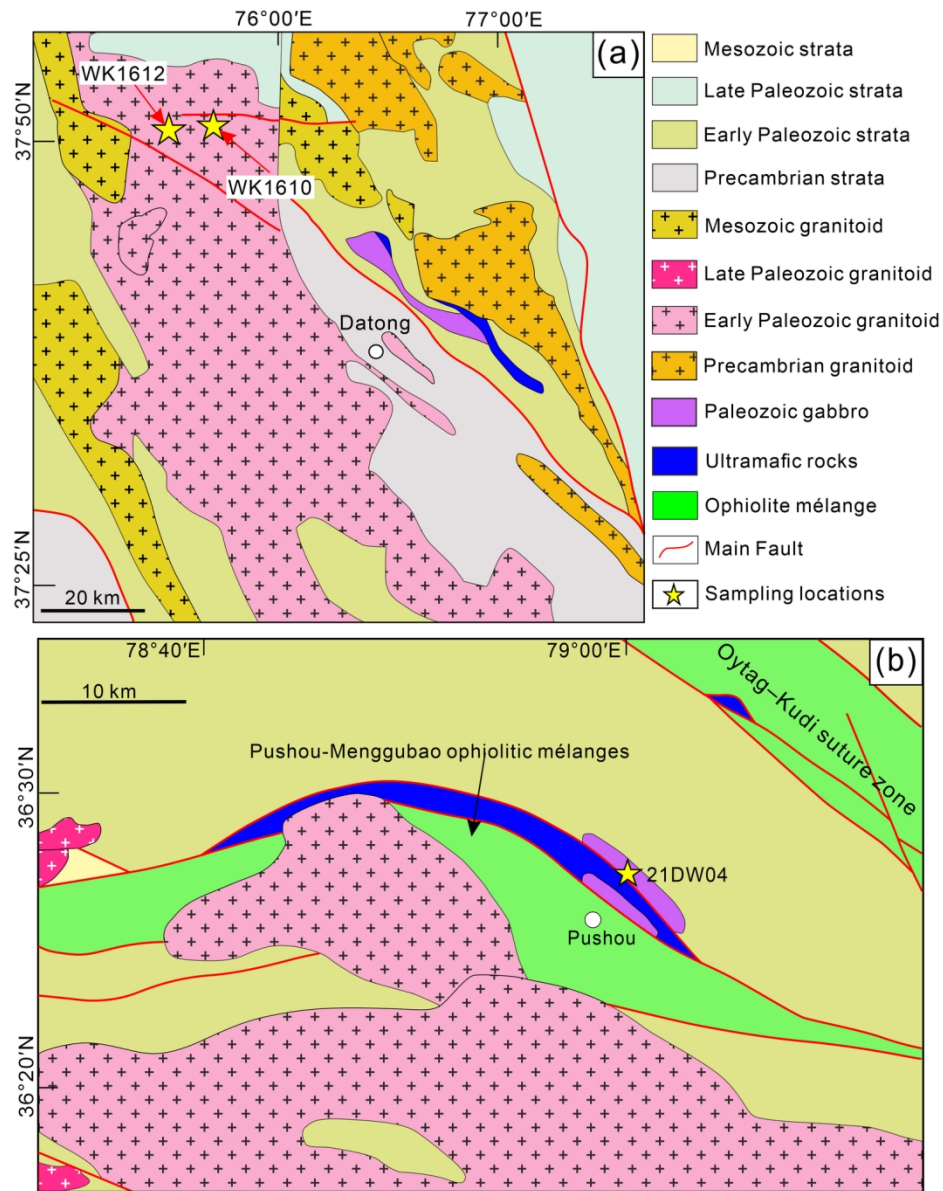


Figure 2

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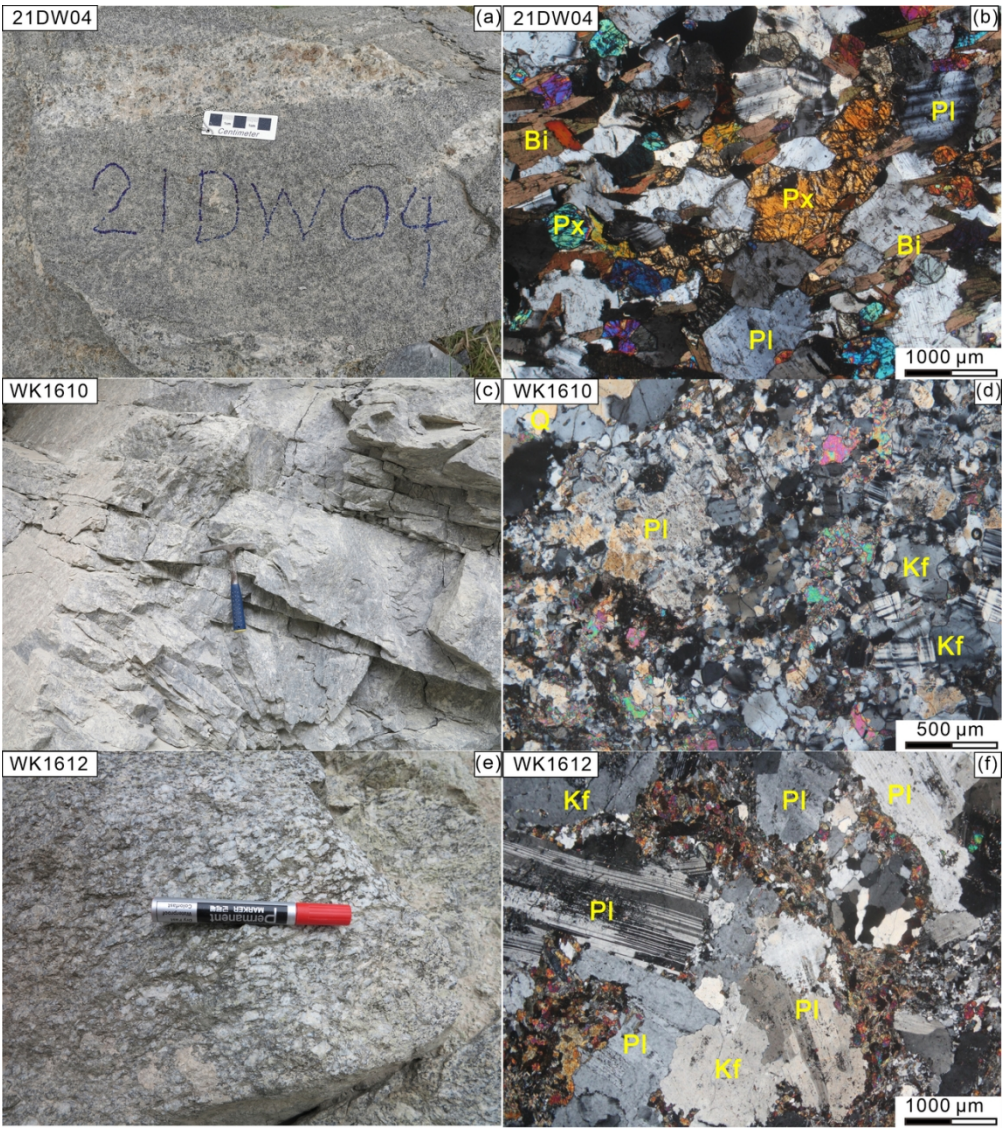


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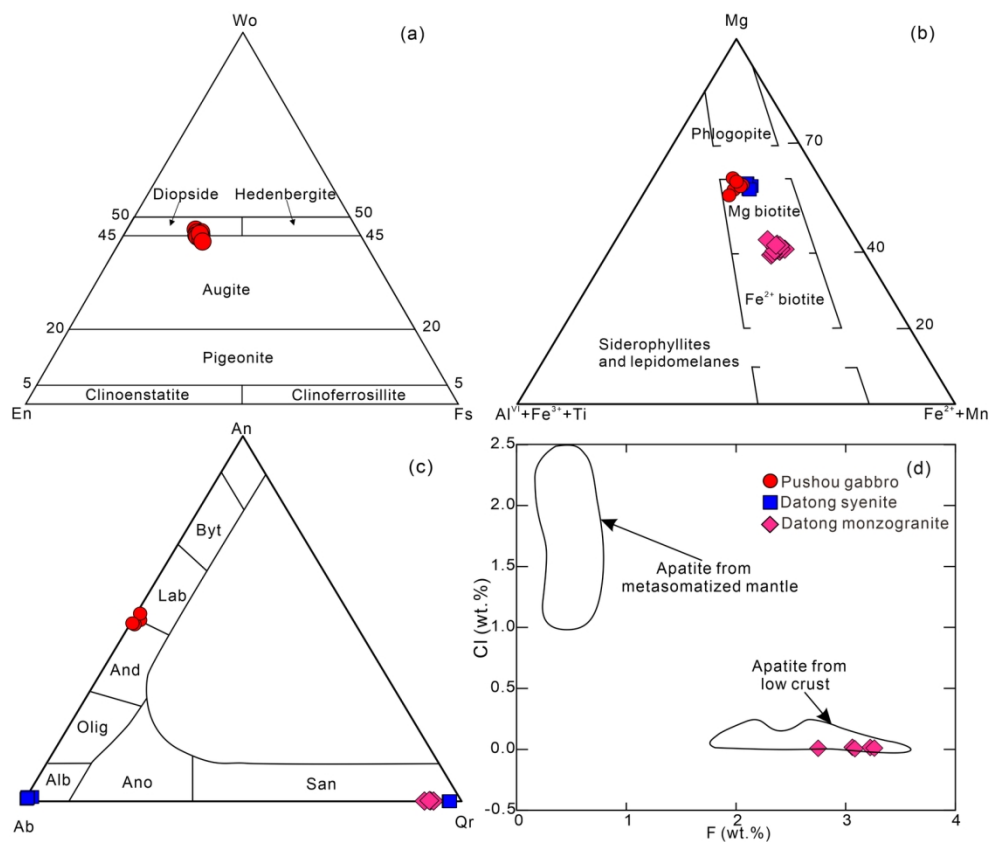


Figure 4

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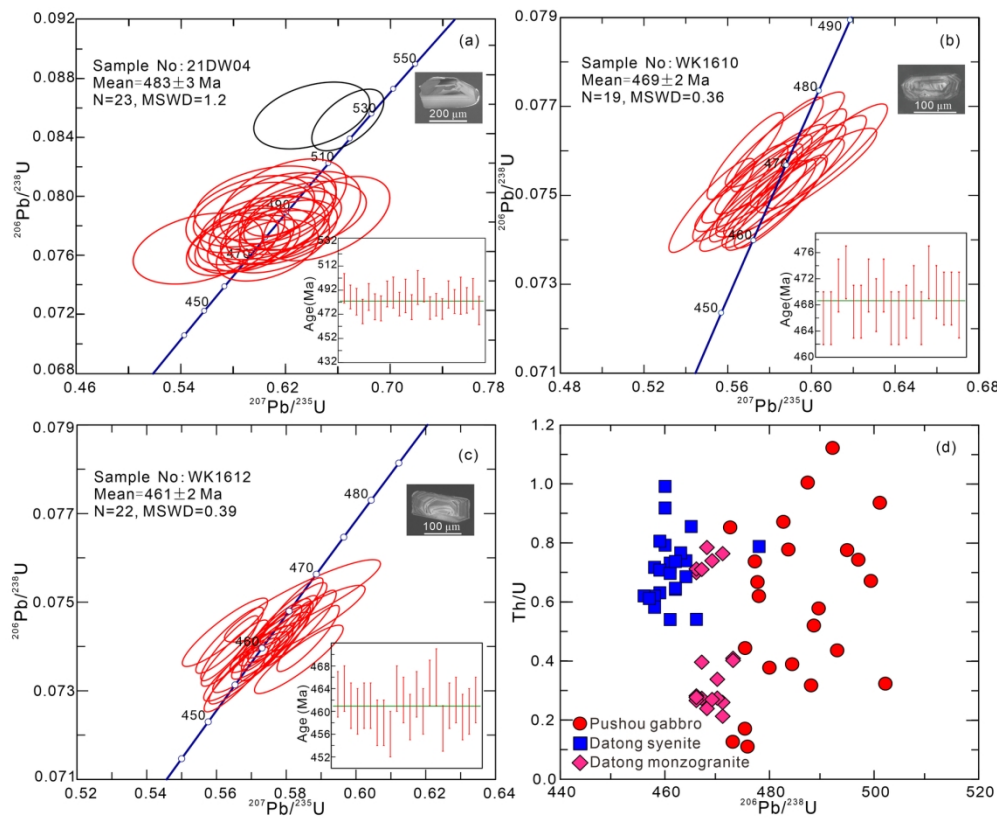


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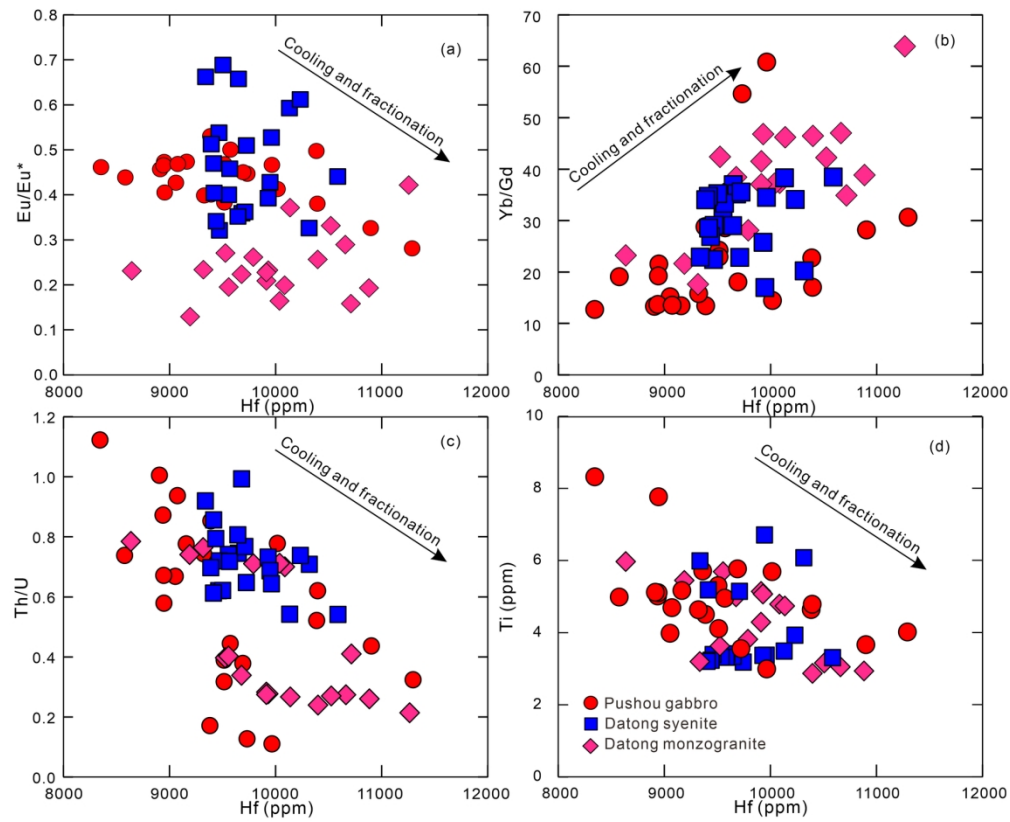


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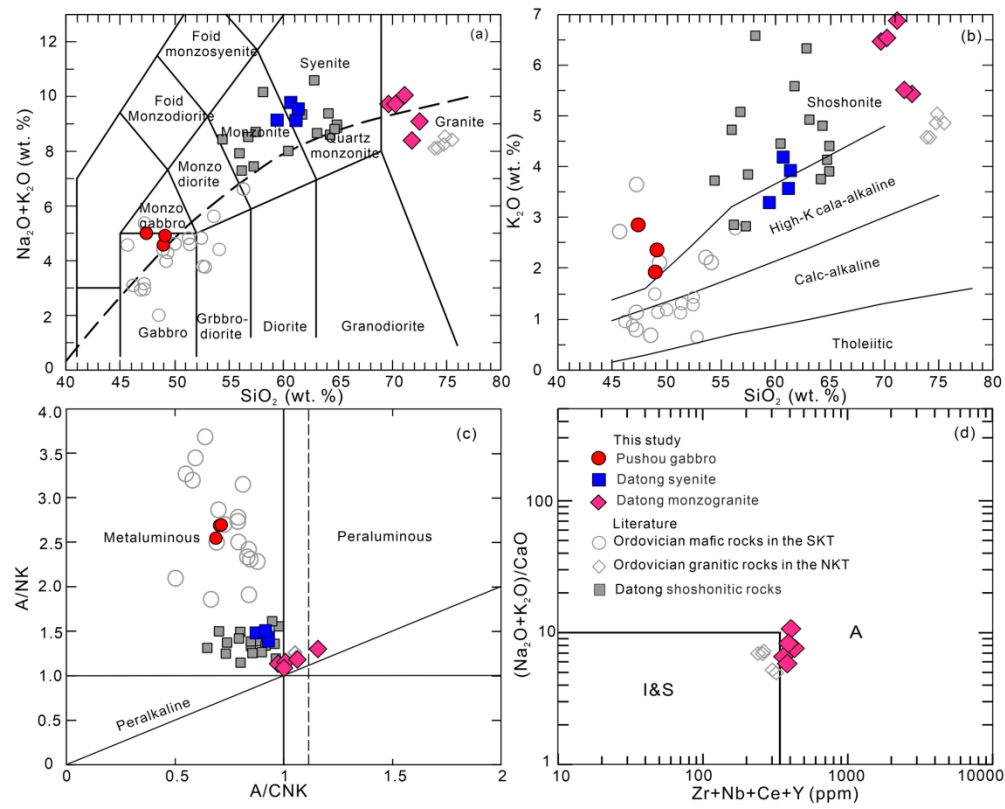


Figure 7

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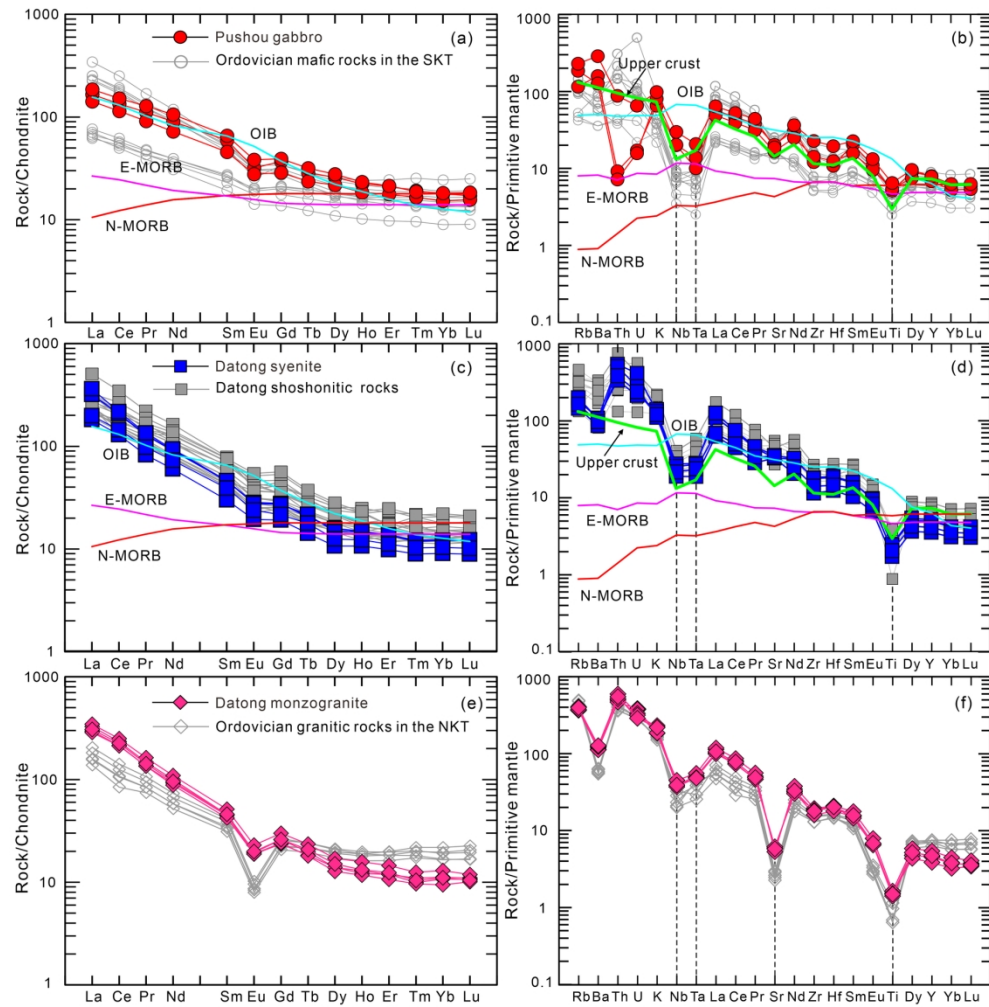


Figure 8

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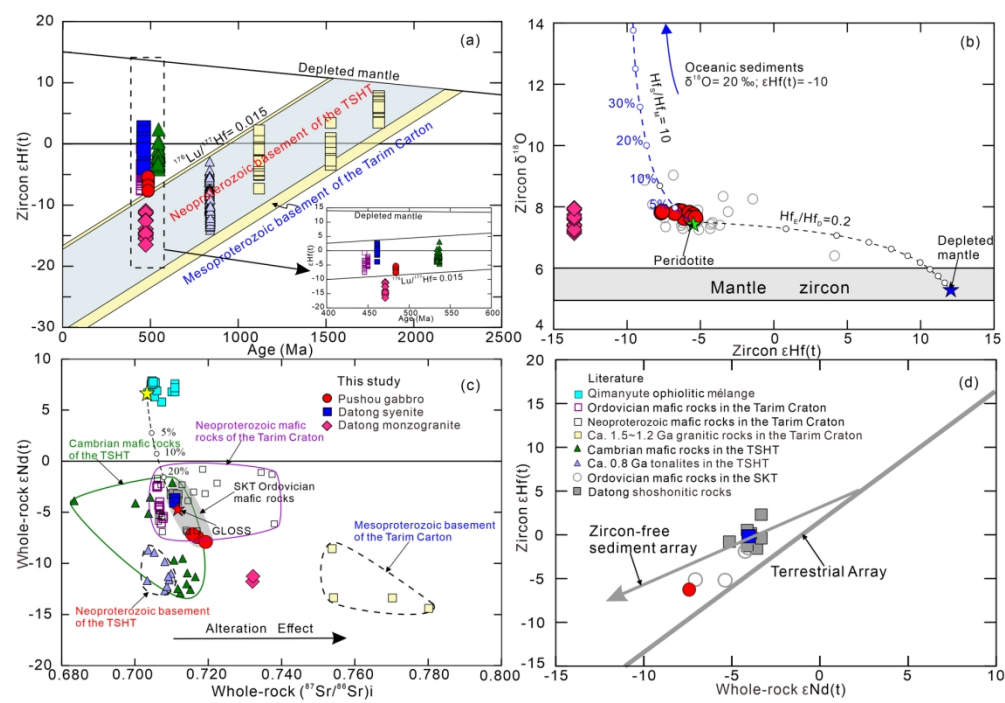


Figure 9

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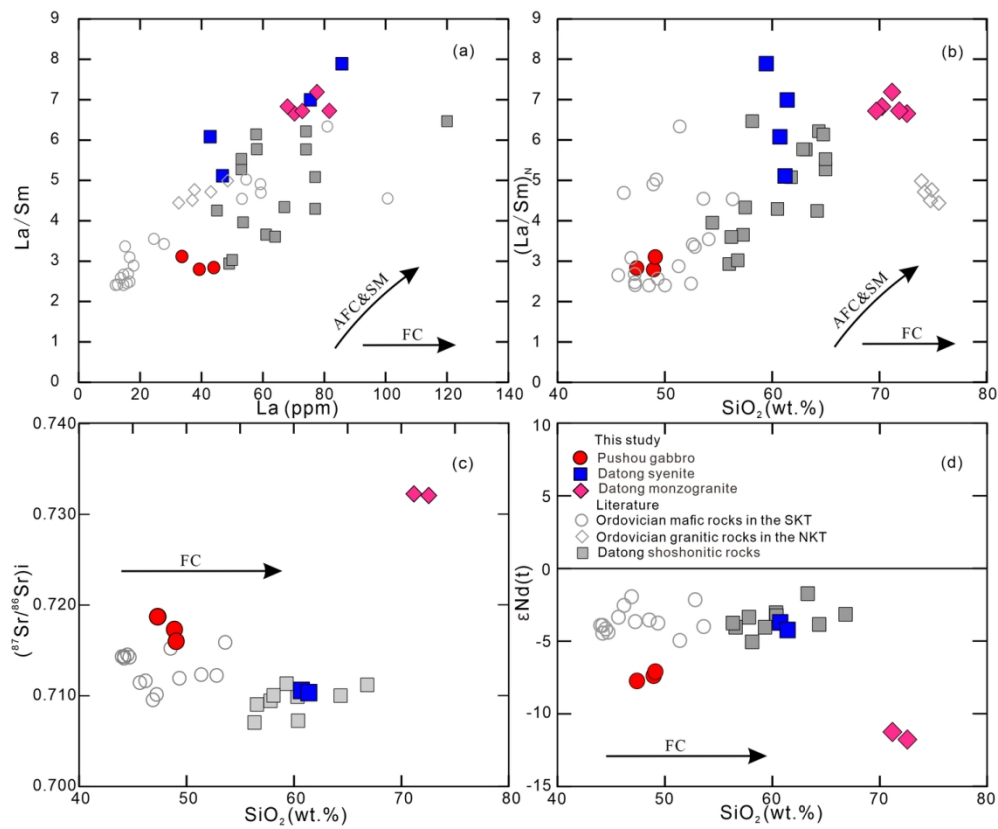


Figure 10

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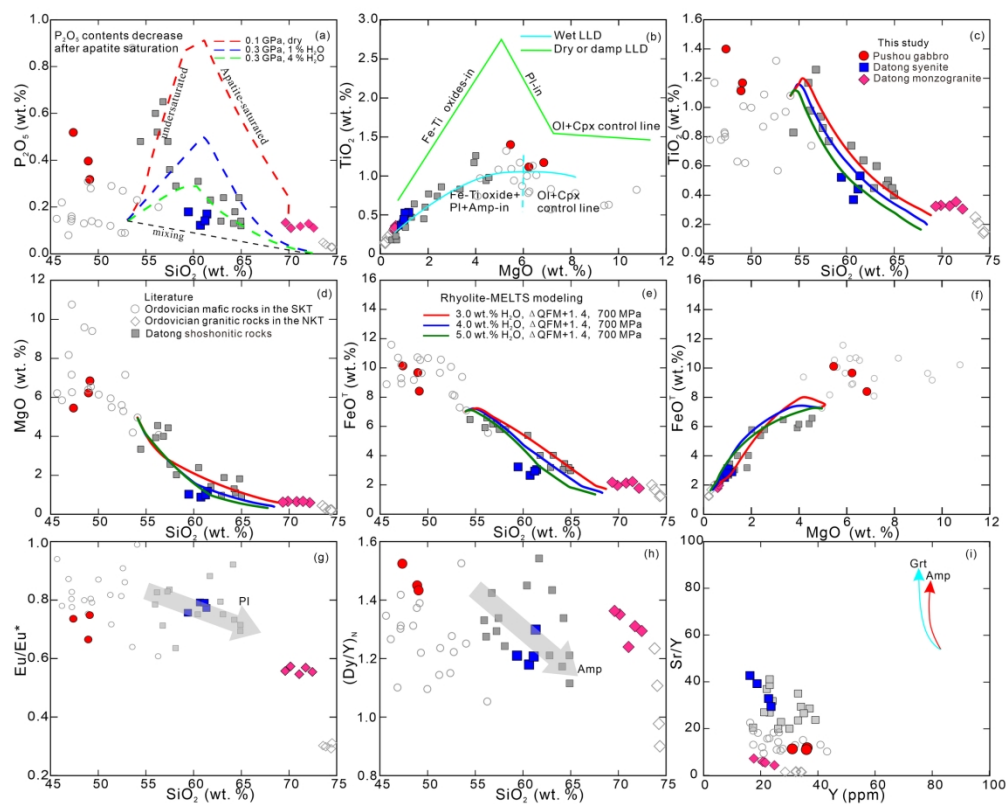


Figure 11

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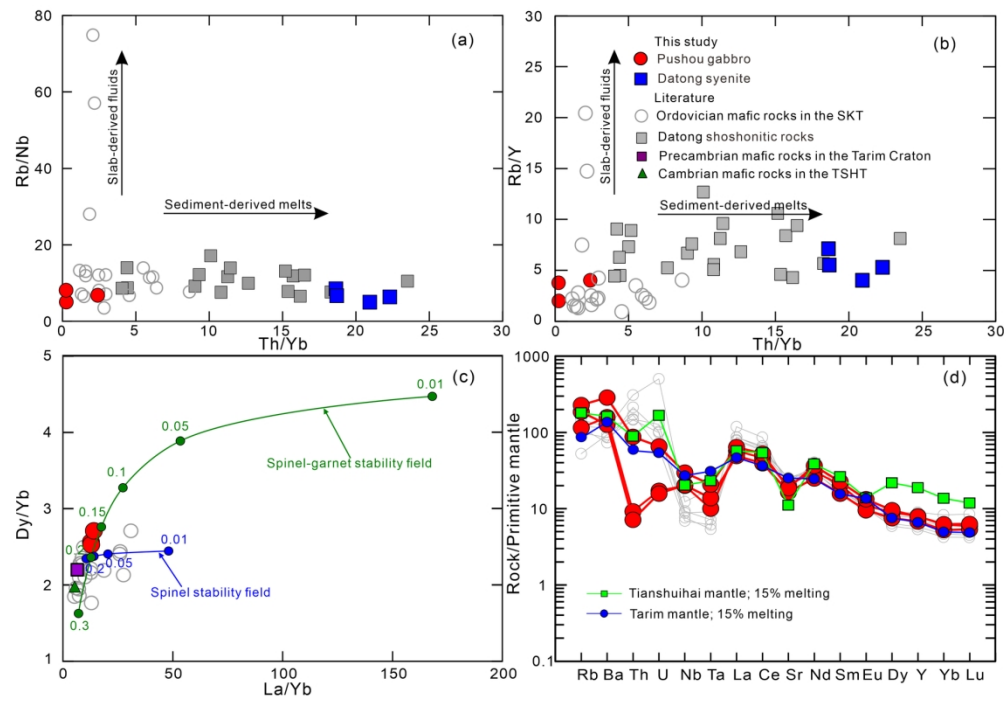


Figure 12

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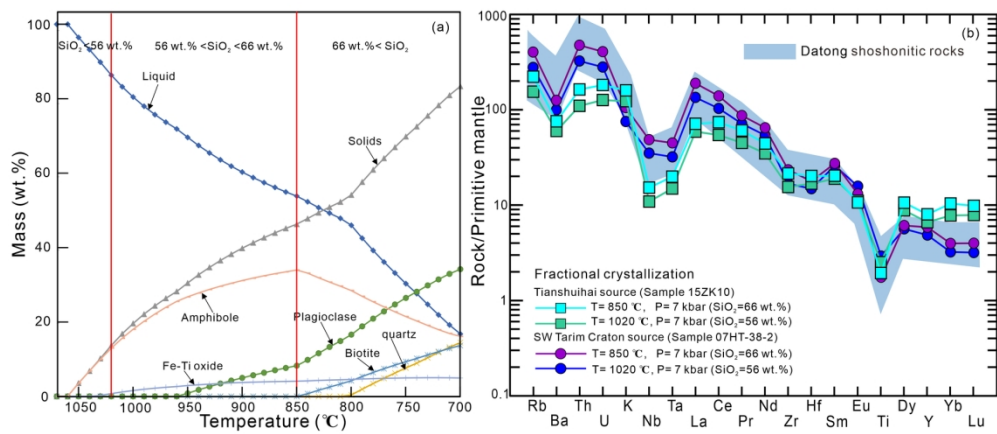


Figure 13

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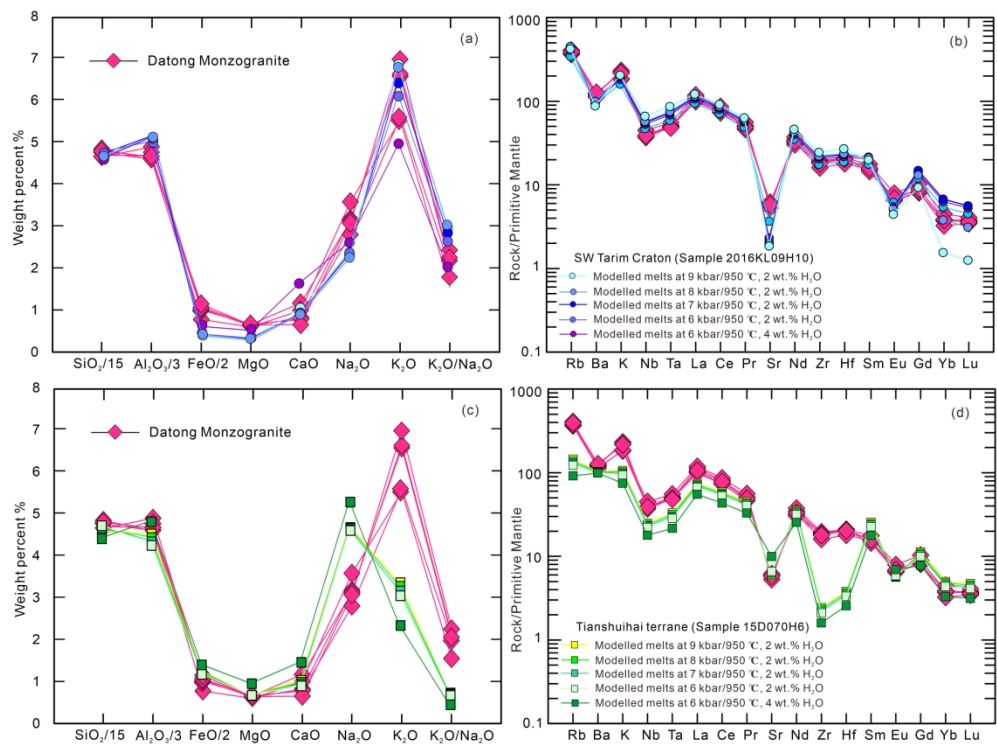


Figure 15

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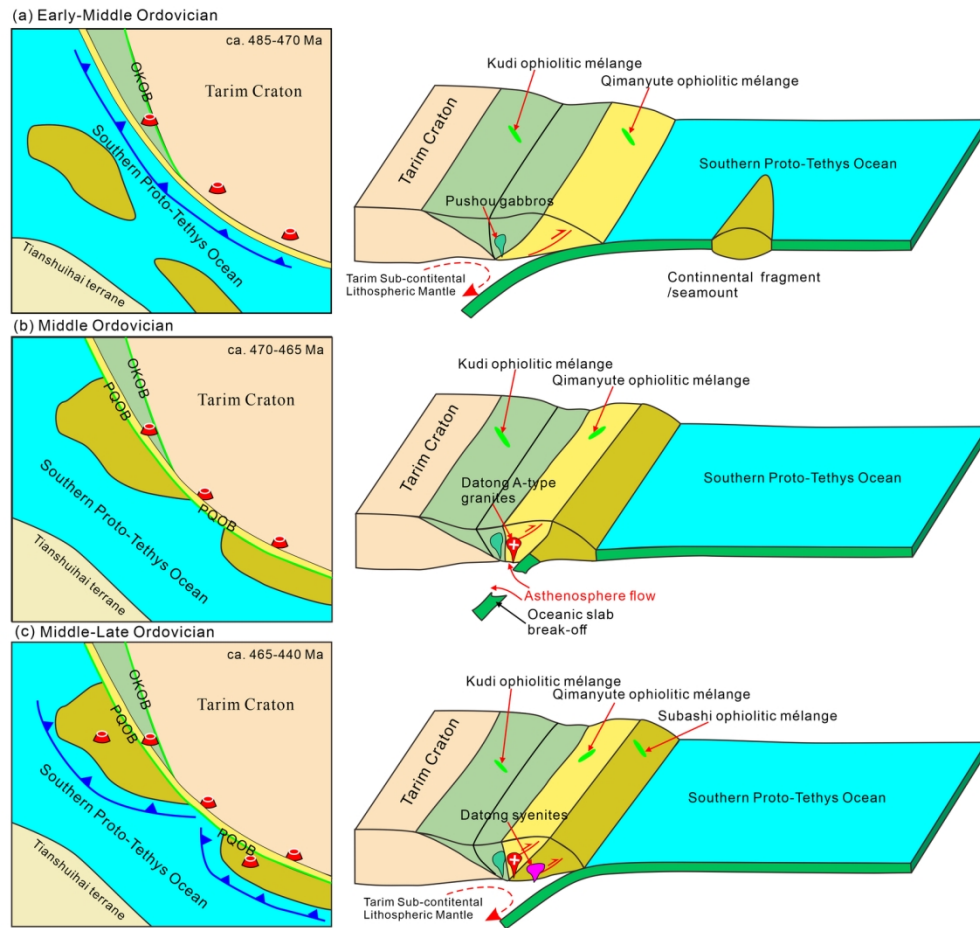


Figure 16

154x145mm (300 x 300 DPI)

Table 1 Summary of sample localities and ages of the intrusions in the West Junggar oceanic arc

Samples	Locality	Longitude (E)	Latitude (N)	Lithology	Age (Ma)	error
ZK003-233	Baogutu	-	-	Diorite	321	2
BCK2-1	Baogutu	-	-	Granodiorite porphyry	310	2
XMST08	Xiemisitai Mts	85°31'07"	46°42'39"	K-feldspar granite	419	2
XMST14	Xiemisitai Mts	85°31'07"	46°42'39"	Diorite	422	2
XMST02	Xiemisitai Mts	85°31'07"	46°42'39"	Alkaline granite	417	4
XMST01	Xiemisitai Mts	85°30'25"	46°35'05"	K-feldspar granite	420	4
XMST03	Xiemisitai Mts	85°29'09"	46°34'44"	Rhyolite-Porphyry	420	5
HZ	Xiemisitai Mts	85°52'28"	46°36'42"	Rhyolite-Porphyry	414	5
XGSL03	Xiemisitai Mts	85° 29'09"	46° 34'44"	Rhyolitye-porphyry	420	5
TW1	Xiemisitai Mts	83°57'43"	46°19'11"	Granodiorite	411	3
HLSL	Saier Mts	85°48'44"	46°44'16"	Felsite	405	4
CGKL	Saier Mts	85°53'39"	46°45'56"	K-feldspar granite	410	3
HBK	Saier Mts	85°40'42"	46°46'32"	K-feldspar granite	417	5
TLG	Hebukesaier	85° 10'26"	46° 47'11"	Granite	263	6
KLMS	Wuerkashier Mts	84°07'00"	46°17'58"	K-feldspar granite	302	2
SLK	Wuerkashier Mts	84°34'52"	46° 22'45"	K-feldspar granite	290	4
LST	Sawur Mts	85°54"	47°11'15"	K-feldspar granite	321	5
ZQS	Tarbgatay Mts	83° 00'30"	47°06'07"	Granodiorite	325	3
DYS	Tarbgatay Mts	84° 21'16"	46° 53'50"	Pyroxene diorite	333	2
BEG	Tarbgatay Mts	83°04'58"	47°04'23"	Granodiorite	332	3
ABDL	Tarbgatay Mts	83°16'58"	46°59'12"	Monzonitic granite	332	3
QS-1	Xiemisitai Mts			quartz syenite	418	5
D-1	Xiemisitai Mts			diorite	424	5
MG-1	Xiemisitai Mts			monzogranite	418	5
S-1	Xiemisitai Mts			syenite	392	6
HG-1	Xiemisitai Mts			hornblende gabbro	422	5
12XJ37-A	Xiemisitai Mts	85°07'20"	46°38'30"	monzogranite	429	1
12XJ05-B	Xiemisitai Mts	85°10'10"	46°36'40"	Quartz diorite	440	3
12XJ14-A	Xiemisitai Mts	85°10'15"	46°34'20"	Quartz monzodiorite	445	6
KM-9918-3	Keramay	83°45'05"	45°39'06"	K-feldspar granite	296	7
AK154-2	Akebastaw	84° 45'33"	45° 24'39"	K-feldspar granite	305	4
MG 164-2	Miaoergou	83°57'00"	45°31'3"	Chamockite	305	3
MG9811-1	Miaoergou	84° 01'45"	45° 32'45"	Chamockite	296	3
MG 123-2	Miaoergou	83°52'20"	45°30'55"	Chamockite	305	3
MG136-1	Miaoergou	83° 56'26"	45° 29'25"	K-feldspar granite	308	6
MG9951-1	Miaoergou	83°52'33"	45°31'57"	K-feldspar granite	298	4
Gn-71	Sawur Mts			K-feldspar granite	338	4
Gn-40	Tarbgatay Mts			Granodiorite	325	4
Gn-14	Tarbgatay Mts			K-feldspar granite	303	4
KLMY21	Keramay			K-feldspar granite	295	6
KLMY3	Keramay			Granite	318	5
KLMY16	Akebastaw			K-feldspar granite	276	5
MEG5	Miaoergou			Granite	327	7
TGL1	Taergen			K-feldspar granite	287	5
KL-19	Keramay			Quartz diorite	316	3
ALT-49	Barleik Mts			diorite	324	2
ALT-87	Barleik Mts			diorite	321	2
ALT-52	Barleik Mts			diorite	320	2
ALT-74	Barleik Mts			quartz syenite	314	5
ALT-130	Barleik Mts			quartz syenite	282	2
ALT-36	Barleik Mts			quartz syenite	272	2
ALT-101	Barleik Mts			granitic porphyry	274	2
ALT-116	Barleik Mts			granitic porphyry	270	2
ALT-84	Barleik Mts			granitic porphyry	259	5
9LB-50	Saleinuohai Mts	82°56'37.7"	45°22'42.3"	plagiogranite	515	3
09LB-53	Saleinuohai Mts	82°58'43.9"	45°22'20.0"	plagiogranite	514	3
09LB-77	Saleinuohai Mts	82°50'17.4"	45°21'41.6"	plagiogranite	509	2

SLN-19	Saleinuohai Mts	82°55'43.0"	45°23'21.2"	plagiogranite	511	4
09LB-81	Saleinuohai Mts	82°50'4.3"	45°21'29.0"	quartz plagioclase	500	2
LBD-09	Saleinuohai Mts	83°03'36.7"	45°20'35.3"	plagiogranite	493	2
HJK-13	Saleinuohai Mts	82°52'28.8"	45°20'53.7"	carbonated felsic rock	495	9
HJK-14	Saleinuohai Mts	82°58'20.7"	45°19'13.4"	gabbro	498	2
HJK-18	Saleinuohai Mts	82°56'42.1"	45°17'45.0"	porphyritic granite	492	2
D07057	Taergen	83°32'35"	45°45'55"	K-feldspar granite	296	3
AS3	Akebastaw			K-feldspar granite	303	3
MI	Miaoergou			K-feldspar granite	305	2
06XJ145	Baogutu	84°27'10"	45°29'57"	Quartz dioritic porphyry	315	4
06XJ153	Baogutu	84°26'25"	45°23'54"	Dioritic porphyry	314	4
06XJ156	Baogutu	84°26'14"	45°23'24"	Dioritic porphyry	310	3
06XJ182	Keramay			Granodiorite	316	3
12WJ08	Xiemisitai	85°08'31"	46°41'21"	Syenitic porphyry	412	4
11WJ70	Xiemisitai	85°33'45"	46°41'11"	Granite	412	5
11WJ85	Xiemisitai	85°30'01"	46°42'48"	Alkali granite	416	5
12WJ03	Xiemisitai	85°07'46"	46°41'16"	Syenitic granite	418	7
11WJ94	Xiemisitai	85°27'31"	46°42'30"	Syenitic granite	424	6
11WJ07	Miaoergou	83°54'32"	45°30'25"	Aalkali-feldspar granite	294	3
11WJ23	Miaoergou	83°57'00"	45°31'04"	Charnokite	297	4
11WJ26	Hatu	84°32'45"	46°06'39"	Biotite granite	298	4
11WJ06	Miaoergou	83°57'00"	45°30'54"	Charnokite	298	2
11WJ16	Miaoergou	83°51'18"	45°32'31"	Aalkali-feldspar granite	300	5
11WJ12	Miaoergou	83°51'03"	45°32'50"	Charnokite	305	7
06XJ136	Keramay	84°39'43"	45°40'26"	Diorite	306	5
06XJ138	Baogutu	84°27'02"	45°29'56"	Diorite	310	4
06XJ120	Keramay	84°50'54"	45°43'40"	Aalkali-feldspar granite	311	4
06XJ148	Keramay	84°32'25"	45°28'25"	Granodiorite	312	4
15XJ07	Keramay	85°13'33"	46°18'04"	Diorite	315	5
15XJ01	West Keramay	84°31'59"	45°39'33"	Diorite	316	7
06XJ143	Baogutu	84°27'05"	45°29'56"	Gabbro	332	4
WJ2009	Mayile Mts	83°01'27"	46°01'39"	Quartz Monzonite	270	2
19BLK03	Mayile Mts	83°23'08.03"	45°47'43.87"	Granodiorite	432	4
WJ1103	Alashankou	82°44'43"	45°18'33"	Alkali granite	269	2
WJ1115	Bieluagaxi	84°24'52.3"	46°02'56.4"	Granodiorite	299	2
WJ1603	Suyunhe	82°29'13"	45°47'20"	Monzonitic granite	302	2
18BGT02	Baogutu	84°28'07.06"	45°29'51.43"	Diorite	315	2
18EK01	Keramay	84°42'40.62"	45°40'26.45"	Monzonitic granite	311	1
18MG03	Miaoergou	83°56'58.85"	45°31'03.32"	Charnockite	306	1
18MG04	Miaoergou	83°52'18.98"	45°30'54.84"	Charnockite	306	1
WJ1309	Xiemisitai	85°28'23.6"	46°36'6.6"	A-type granite	307	1
WJ1315	Sawur Mts	86°15'21.7"	47°26'36.6"	A-type granite	293	1
WJ1325	Sawur Mts	86°07'45.7"	47°08'4.9"	I-type granite	336	1
WJ1326	Sawur Mts	86°10'32.0"	47°13'27.8"	A-type granite	298	1
WJ2001	Xiemisitai Mts	86°06'45.33"	46°45'31.56"	Rhyolite	425	5
WJ2002	Xiemisitai Mts	86°05'26.09"	46°44'33.85"	Rhyolite	418	4
WJ2004	Xiemisitai Mts	86°02'06.51"	46°42'17.51"	Diorite Porphyry	438	4
WJ2005	Xiemisitai Mts	86°22'57.28"	46°38'23.69"	Alkali granite	304	3
WJ1118	Tarbgatay Mts	84°35'38.05"	46°45'27.26"	Alkali granite	277	1
WJ1121	Tarbgatay Mts	84°21'17.0"	46°41'18.4"	Alkali granite	277	5
WJ1142	Xiazijie	86°10'45.2"	46°34'32.4"	Granodiorite	315	2
09BGT-18	Baogutu			Quartz diorite	313	3
09MYL-19	Mayile Mts	83°13'22"	45°34'09"	Diorite	501	3
09MYL-145	Mayile Mts	83°11'46"	45°34'26"	Gabbro	493	5
09MYL-147	Mayile Mts	83°12'16"	45°34'31"	Granite	493	5
08MYL-10	Mayile Mts	83°16'32"	45°40'55"	Diorite	488	5
08MYL-19	Mayile Mts	83°16'24"	45°40'48"	Diorite	485	2
09AEST-8	Mayile Mts	83°23'15"	45°47'47"	Granodiorite	444	2
09AEST-9	Mayile Mts	83°23'15"	45°47'47"	Granodiorite	439	3

09MYL-56	Mayile Mts	83°15'51"	45°40'22"	Diorite	318	4
08MYL-53	Mayile Mts	83°19'16"	45°40'57"	Granodiorite	308	2
08MYL-54	Mayile Mts	83°20'36"	45°41'43"	Granodiorite	313	3
09MYL-37	Mayile Mts	83°16'49"	45°36'45"	Granodiorite	307	2
09JBL-4	Mayile Mts	83°30'01"	45°38'56"	Granodiorite	311	1
09JBL-3	Mayile Mts	83°27'33"	45°40'43"	K-feldspar granite	309	2
08TLG-1	Taergen			K-feldspar granite	309	4
09TK-18	Barleik Mts	83°12'19"	45°52'24"	Trondhjemite	503	3
10TK-84	Barleik Mts	83°13'14"	45°51'36"	diorite	509	3
10TK-88	Barleik Mts	83°13'26"	45°51'56"	diorite	503	3
10TK-3	Barleik Mts	83°11'59"	45°53'13"	Granodiorite	372	2
10TST-2	Barleik Mts	82°44'22"	45°55'15"	K-feldspar granite	277	2
10QE-3	Barleik Mts	82°27'05"	45°43'52"	Granodiorite	311	2
WJ1107	Tarbgatay Mts	84°21'21"	46°53'52"	Gabbro	343	2
WJ1109	Tarbgatay Mts	84°40'54"	46°56'38"	Biotite Granite	329	2
WJ1133	Hongguleleng	86°15'20.7"	46°40'41.3"	Quartz Monzonite	297	3
WJ1135	Hongguleleng	86°31'22.9"	46°41'48.8"	Quartz Monzonite	282	2
WJ1143	Xiazijie	86°14'0.7"	46°34'22.3"	Alkali granite	302	2
WJ1122	Xiemisitai Mts	85°32'20.2"	46°41'36"	Gabbro	420	3
WJ1125	Xiemisitai Mts	85°30'20.0"	46°42'38.3"	K-feldspar granite	420	1
WJ1126	Xiemisitai Mts	85°27'34.9"	46°42'28.7"	Syenitic granite	418	2
WJ1128	Xiemisitai Mts	85°28'30.0"	46°34'44.3"	Rhyolite	419	2
WJ1132	Xiemisitai Mts	85°36'60"	46°39'32.4"	Granodiorite	420	2
WJ1311	Xiemisitai Mts	85°06'54.7"	46°42'23.9"	Granodiorite	419	2
WJ1312	Xiemisitai Mts	85°19'36.6"	46°42'56.1"	K-feldspar granite	420	2
WJ1136	Saier Mts	85°53'18.5"	46°45'43"	Syenitic granite	407	3
WJ1314	Xiemisitai Mts	85°47'36.1"	46°38'41.2"	monzogranite	414	3
18AK01	Akebasitao	84°21'37.80"	45°43'09.21"	A-type granite	305	1
18HATU	Hatu	84°14'24.52"	46°01'06.60"	A-type granite	302	1
18TCG01	Tiechanggou	84°31'16.54"	46°08'19.97"	A-type granite	307	2
18NKM01	Northern Keramay	84°50'42.84"	45°43'56.67"	A-type granite	305	1
WJ1144	Miaoergou	83°54'44.5"	45°31'51.8"	A-type granite	302	2
WJ1117	Wuerkashier	84°03'57.26"	46°17'44.75"	A-type granite	298	2
18MY01	Mayile Mts	83°13'22.36"	45°34'09.57"	Alkali granite	502	2
18MY02	Mayile Mts	83°16'13"	45°40'47"	Diorite	488	2
18MY03	Mayile Mts	82°50'02.08"	45°21'37.99"	Diorite	507	2
19BLK01	Mayile Mts	83°13'05.88"	45°51'33.46"	Diorite	502	3
19BLK02	Mayile Mts	83°13'14.04"	45°51'53.66"	Granodiorite	503	3
19SLH07	Mayile Mts	82°56'39.13"	45°22'41.84"	Trondhjemite	511	3
19MYL01	Mayile Mts	83°14'34.04"	45°34'28.12"	Gabbro	490	4
19SLH02	Mayile Mts	82°58'21.48"	45°19'12.85"	Gabbro	503	2
OK-4	Sawur Mts	86° 17'51"	47° 01'01"	Adamellite	324	6
STS-1	Sawur Mts	86° 01'41"	47° 04'23"	Adamellite	328	6
KE-18	Sawur Mts	86° 21'20"	47° 12'33"	K-feldspar granite	303	8
KY-5	Sawur Mts	86°17'32"	47°13'28"	Alkali granite	298	5
QQH-14	Sawur Mts	86° 15'22"	47°26'38"	Alkali granite	291	9

Method	Data sources
LA-ICPMS	Cao et al., 2016
LA-ICPMS	Cao et al., 2016
SHRIMP II	Chen et al., 2010
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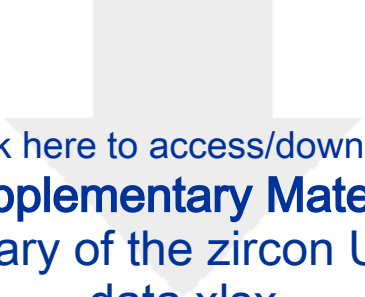
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SHRIMP	Zhou et al., 2008
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Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

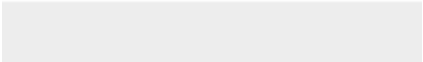
☐ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:



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Supplementary Material

Table S1 A summary of the zircon U-Pb, Hf-O isotope
data.xlsx





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Supplementary Material

Table S2 In situ zircon Hf-O isotopes.xlsx

