**Crustal melting and magma mixing in a continental arc setting: evidence from the Yaloman
intrusive complex in the Gorny Altai terrane, Central Asian Orogenic Belt**

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20 **Abstract:** Granitoids and their hosted mafic enclaves may retain important information on
21 crust-mantle interaction, and thus are significant for study of crustal growth and
22 differentiation. An integrated petrological, geochronological and geochemical study on the
23 granitoid plutons of the Yaloman intrusive complex from the Gorny Altai terrane,
24 northwestern Central Asian Orogenic Belt, was conducted to determine their source nature,
25 petrogenesis and geodynamics. Mafic enclaves are common in the plutons, and a zircon U-Pb
26 age ($389 \text{ Ma} \pm 4 \text{ Ma}$) indicates that they are coeval with their granitoid hosts (ca. 393-387
27 Ma). Petrographic observations reveal that these mafic enclaves probably represent
28 magmatic globules commingled with their host magmas. The relatively low SiO_2 contents
29 (46.0-60.7 wt. %) and high $\text{Mg}^\#$ (38.9-56.5) further suggest that mantle-derived mafic melts
30 served as a crucial component in the formation of these mafic enclaves. The granitoid hosts,
31 including quartz diorites and granodiorites, are I-type in origin, possessing higher SiO_2
32 contents (60.2-69.9 wt.%) and lower $\text{Mg}^\#$ (32.0-44.2). Their zircon Hf and whole-rock Nd
33 isotopic compositions indicate that the magmas were dominated by remelting of
34 Neoproterozoic (0.79-1.07 Ga) crustal materials. Meanwhile, the geochemical modeling,
35 together with the common occurrence of igneous mafic enclaves and the observation of
36 reversely zoned plagioclases, suggests that magma mixing possibly contributed significantly
37 to the geochemical variation of the granitoid hosts. Our results imply that mafic magmas
38 from the mantle not only provided substantial heat to melt the lower crust, but also mixed

with the crust-derived melts to form the diverse granitoids.

The oxidizing and water-enriched properties inferred from the mineral assemblages and compositions imply that the granitoid plutons of the Yaloman intrusive complex were possibly formed in a continental arc-related setting, which is also supported by their geochemistry. The Devonian granitoids from the Gorny Altai terrane show remarkable temporal-spatial-petrogenetic affinities to the counterparts from the Altai-Mongolian terrane, indicating that these two terranes were possibly under subduction of the same oceanic plate (i.e., the Ob-Zaisan Ocean). The voluminous granitoids signify significant crustal recycling and growth as a response to the underplating of extensive mantle-derived basaltic melts.

Keywords: Mafic enclaves; Granitoids; Subduction zone; Crustal melting; Magma mixing; Gorny Altai terrane

1. Introduction

The Earth is distinct from other rocky planets by its unique continental crust with an average composition similar to that of andesite (e.g., [Taylor and McLennan, 1985](#); [Rudnick, 1995](#)). It has been proposed that lateral accretion of island arcs and basaltic underplating beneath the lower crust contribute dominantly to the crustal growth (e.g., [Rudnick, 1990, 1995](#); [Frost et al., 2001](#); [Xiao et al., 2010](#)). However, mantle-derived melts, either in

subduction zones or intra-plate settings, are predominantly basaltic in composition ([Rudnick, 1990](#); [Debari and Sleep, 1991](#); [Holbrook et al., 1999](#)). Processes, such as differentiation of basaltic magmas and/or partial melting of basaltic rocks, are therefore required to drive these mafic components toward more evolved composition. This inferred geochemical differentiation is supported by geophysical investigations, showing that the lower part of the continental crust is dominated by mafic rocks, while the proportion of intermediate-felsic rocks increases upward (e.g., [Rudnick and Fountain, 1995](#)). Various models, including partial melting of pre-existing lower crust by underplating of basaltic magmas with/without magma mixing (e.g., [Borg and Michael, 1998](#); [Petford and Gallagher, 2001](#); [Dufek and Bergantz, 2005](#)) and differentiation of basaltic magmas through middle- to high-pressure fractional crystallization (e.g., [Jagoutz et al., 2013](#); [Lee and Bachmann, 2014](#)) or silicate liquid immiscibility (e.g., [Charlier et al., 2011](#)), have been proposed. Granitoids and their volcanic equivalents constitute a major proportion of the upper continental crust, and thus are critical to understand the crustal growth and differentiation.

The Central Asian Orogenic Belt (CAOB) is considered to be the most important site for the Phanerozoic crustal growth on the Earth (e.g. [Sengör et al., 1993](#); [Jahn, 2004](#)). The Gorny Altai terrane, situated southwest off the Siberian continent (present coordinate; [Figs. 1a, b](#)), is dominated by late Neoproterozoic to early Ordovician basaltic rocks, which have been proposed to be formed in an intra-oceanic island arc setting (i.e., the Kuznetsk-Altai arc;

77 [Buslov et al., 1993, 2002, 2013; Simonov et al., 1994; Ota et al., 2007; Safonova et al., 2011;](#)
78 [Utsunomiya et al., 2009; Kruk et al., 2010](#)). After a period of tectono-magmatic quiescence
79 from the Ordovician to early Devonian, the Gorny Altai terrane underwent extensive
80 magmatic activities in the Devonian to early Carboniferous, as recorded by voluminous
81 granitoids and their volcanic equivalents ([Buslov et al., 1993, 2013; Yolkin et al., 1994;](#)
82 [Daukeev et al., 2008; Kruk et al., 2008, 2011; Buslov and Safonova, 2010; Glorie et al., 2011;](#)
83 [Cai et al., 2014; Safonova, 2014](#) and references therein). The contrasting magmatic styles in
84 these two periods make the Gorny Altai terrane an ideal place to test the above-mentioned
85 models in generation of the intermediate-felsic igneous rocks and to understand how
86 continental crust grew and differentiated. However, due to the paucity of geochemical and
87 isotopic data, the petrogenesis and tectonic setting of the Devonian to early Carboniferous
88 magmatic suites in the Gorny Altai terrane are not well understood ([Kruk et al., 2008, 2011;](#)
89 [Glorie et al., 2011; Cai et al., 2014](#)).

90 In this study, we mainly focus on the granitoid plutons of the Yaloman intrusive complex
91 in the south of the Gorny Altai terrane ([Figs. 1b, c](#)). Mafic enclaves are commonly observed
92 ([Fig. 2a](#)). Such an association may provide important constraints on the source nature and
93 mechanism for the formation of these plutons, which in turn might have great significance in
94 understanding the Phanerozoic crustal evolution in the northwestern CAOB. A combined
95 geochemical, zircon U-Pb-Hf and whole-rock Sr-Nd isotopic study was conducted on the

granitoids and their hosted mafic enclaves. Our results show that melting of pre-existing mafic lower crust and mixing with the newly underplated basaltic magmas in a continental arc setting is a plausible process in the formation of these granitoids, signifying both crustal growth and vertical differentiation.

2. Geological setting

The Gorny Altai terrane is a triangle-shaped tectonic unit, whose northern extension is covered by Cenozoic sediments (Buslov and Safonova, 2010; Buslov, 2011). It is in fault contact with the Rudny Altai terrane to the west, the Altai-Mongolian terrane to the south and the West Sayan terrane to the east (Fig. 1a).

The geodynamic evolution of the Gorny Altai terrane began with the initiation of the Kuznetsk-Altai arc adjacent to the southwestern periphery of the Siberian continent (present coordinate) during the late Neoproterozoic to early Cambrian (Buslov et al., 1993, 2001, 2002, 2003, 2013; Ota et al., 2007; Utsunomiya et al., 2009; Buslov and Safonova, 2010; Glorie et al., 2011; Chen et al., 2015). The tholeiitic and minor boninitic volcanic rocks generated in this period, in association with accretionary complexes and ophiolitic mélange, are preserved in the eastern segment of this terrane (Buslov et al., 1993, 2002; Simonov et al., 1994; Ota et al., 2007; Utsunomiya et al., 2009). The Kuznetsk-Altai arc was suggested to evolve toward a mature one in the middle-late Cambrian with dominance of calc-alkaline magmatism,

including volcanic suites and to a much lesser extent of plutonic granitoids (Buslov et al., 2002; Kruk et al., 2011; Glorie et al., 2011 and references therein). However, a recent detrital zircon study on fore-arc sedimentary sequences from the Gorny Altai terrane shows that the Kuznetsk-Altai arc possibly underwent a prolonged subduction-accretion history with dominance of basaltic-andesitic rocks in ca. 640-540 Ma and increased intermediate-felsic magmatic activities in ca. 540-470 Ma (Chen et al., 2015). This island arc was suggested to collide with the Siberian continent during the late Cambrian to early Ordovician (Berzin et al., 1994; Buslov et al., 2001, 2002; Dobretsov et al., 2003). After the early Ordovician, the Gorny Altai terrane was probably under a passive marginal setting till the early Devonian. During this period, thick terrigenous sediments as well as some carbonate deposits were accumulated in a shallow water shelf setting (Yolkin et al., 1994; Buslov et al., 2013 and references therein).

It was suggested that the Gorny Altai terrane turned into an active continental margin during the early-middle Devonian, resulting in extensive volcanic activities and coeval volcanoclastic deposits (Fig. 1b; Buslov et al., 1993; Buslov and Safonova, 2010; Kruk et al., 2011; Glorie et al., 2011; Safonova, 2014). These rocks were best represented by the early Devonian Ongudai and middle Devonian Kuratin volcanic complexes (Buslov et al., 1993). The Ongudai complex consists mainly of basaltic-andesitic lavas, andesitic tuffs, and volcanoclastic rocks, while the Kuratin complex is composed predominantly of andesitic to

felsic volcanic rocks, pyroclastic deposits and small subvolcanic intrusions (Buslov et al., 1993; Safonova, 2014). Granitoid intrusions increased significantly within the Gorny Altai terrane (Figs. 1b, c) in the middle-late Devonian and they have been recently dated to be ca. 395-360 Ma old (Glorie et al., 2011; Kruk et al., 2011 and references therein; Cai et al., 2014). These rocks were thought to be generated during the collision between the Gorny Altai and Altai-Mongolian terranes, coupled with the formation of the CTUS suture zone (Buslov et al., 1993, 2001, 2004a, b, 2013; Buslov and Safonova, 2010; Glorie et al., 2011).

The collision between the East European and Siberian continents during the late Carboniferous to Permian induced large-scale strike-slip faulting, which truncated the CTUS suture zone and formed the mosaic structure in the northwestern CAO (e.g., Buslov et al., 2001). The minor ca. 255-220 Ma magmatic rocks documented Permian to Triassic intra-plate magmatism (Glorie et al., 2011), which was suggested to be plume-related (e.g., Vladimirov et al., 2008).

3. Samples and petrography

The granitoid plutons of the Yaloman intrusive complex crop out in the Chiketaman pass district. They intruded the Cambrian to Silurian (volcano-)sedimentary sequences (Figs. 1b, c). Three types of rocks, including quartz diorites, granodiorites and their hosted mafic enclaves, were sampled from four separate plutons (Fig. 2).

3.1 Quartz diorites and their hosted mafic enclaves

Quartz diorites were collected from the Chiketaman and Ust-Chuya plutons. They are all medium-grained and consist mainly of plagioclase (55-65%), amphibole (10-20%), quartz (15-20%), K-feldspar (2-5%), biotite (0-3%) and minor magnetite (1-3%), as well as accessory minerals such as apatite and zircon (Fig. 2b). The plagioclase grains are generally ca. 1-3 mm in length with subhedral to euhedral shapes. Most of them have been strongly altered to clay minerals such as sericites and kaolinites, while the rest fresh grains or relicts show twinning and/or concentric zoning. The amphiboles are mostly euhedral to subhedral in shape and have well-developed cleavages. Some of them contain tiny plagioclase inclusions. In contrast, the K-feldspar and quartz grains are always subhedral to anhedral.

Mafic enclaves are common but unevenly distributed within these plutons. In general, they have ellipsoidal shapes, sizing from a few to tens of centimeters (Fig. 2a). Three mafic enclave samples (C14-1, C14-2 and C14-3) were collected from the Chiketaman pluton; one (HKAM171) was collected from the Ust-Chuya pluton. Sample C14-1 is a medium-grained mafic diorite consisting mainly of amphibole (40-50%) and plagioclase (40-50%), with minor quartz (1-2%), magnetite (1-2%), biotite (1%) and K-feldspar (< 1%). The grain size of amphiboles can reach up to 0.5 cm in length. Voluminous 100-200 µm plagioclase grains are enclosed by coarse-grained amphiboles, showing a poikilitic texture (Fig. 2c). Samples C14-2, C14-3 and HKAM171 are fine-grained, with amphibole and plagioclase grains being generally

ca. 0.5-1.0 mm in length. The former two samples are composed predominantly of amphibole (40-50%) and plagioclase (40-50%), as well as minor magnetite (1-3%), quartz (1-3%) and K-feldspar (0-1%). In contrast, sample HKAM171 has slightly lower contents of amphibole (ca. 30%) and plagioclase (ca. 55%), and higher contents of quartz (ca. 10%) and K-feldspar (ca. 5%). A few coarse-grained plagioclase grains are observed, which could be xenocrysts from the host rock or phenocrysts (Fig. 2d)

3.1 Granodiorites and their hosted mafic enclaves

Granodiorites were collected from the Kadrin and Yaloman plutons. In general, they are medium- to coarse-grained, and show mineral assemblages consisting of plagioclase (40-45%), K-feldspar (20-25%), quartz (25-30%), biotite (2-5%), amphibole (1-3%), magnetite (1-2%) and accessory minerals such as apatite and zircon (Fig. 2e).

Mafic enclave samples were only collected from the Kadrin pluton. These rocks are fine-grained and occasionally contain coarse-grained feldspars that could be either xenocrysts from the granodiorite host or phenocrysts. Two types of mafic enclaves are identified based on their mineral assemblages. The first type, represented by sample HKAM180, consists predominantly of plagioclase (ca. 50%), biotite (ca. 25%), amphibole (ca. 20%) and quartz (ca. 5%). Accessory minerals include magnetite and acicular apatite (Fig. 2f). The second type includes samples HKAM177, HKAM178 and HKAM179. Their mafic minerals

are dominated by amphibole (35-40%), with only minor amounts of biotite (2-5%). Other minerals, including plagioclase (45-50%), quartz (3-5%), magnetite (1-2%) and K-feldspar (0-2%), are similar to those of sample HKAM180.

4. Analytical methods

4.1 Zircon U-Pb and Hf-isotope analyses

In-situ zircon U-Pb dating was conducted, with referring to cathodoluminescence (CL) imaging, by employing a Nu Instruments MC-ICPMS equipped with a Resonetics Resolution M-50-HR Excimer Laser Ablation System at the Department of Earth Sciences, the University of Hong Kong. Analyses were performed with a beam of 30 μm and 6 Hz repetition rate, which yielded a signal intensity of 0.03 V at 238 U for the standard zircon 91500. Typical ablation time was 40 s for each measurement, resulting in pits of 30-40 μm in depth. Masses 232, 208-204 were simultaneously measured in static-collection mode. The detailed analytical procedure can be referred to [Xia et al. \(2011\)](#). Standard zircons 91500 and GJ were used as external standards and were analyzed twice before and after every 10 analyses. Integration of background and analytic signals, time-drift correction, and quantitative calibration for U-Pb dating were performed by software ICPMSDataCal ([Liu et al., 2010](#)). Preferred U-Th-Pb isotopic ratios used for 91500 are from [Wiedenbeck et al. \(1995\)](#). Concordia diagrams and probability density plots were made using Isoplot/Ex_ver3 ([Ludwig,](#)

210 [2003](#)). The U-Pb dating results are presented in [Appendix Table S1](#).

211 In-situ zircon Hf-isotope compositions were measured using a Neptune Plus MC-ICP-MS
212 with connection to a Geolas 2005 excimer ArF laser ablation system at the State Key
213 Laboratory of Geological Processes and Mineral Resources, China University of Geosciences
214 (Wuhan). Helium was used as the carrier gas within the ablation cell and was merged with
215 argon (makeup gas) after the ablation cell. For the 193 nm laser a consistent 2-fold signal
216 enhancement was achieved in helium rather than in argon gas ([Hu et al., 2008a](#)). We used a
217 simple Y junction downstream from the sample cell to add small amounts of nitrogen (4 ml
218 min⁻¹) to the argon makeup gas flow ([Hu et al., 2008b](#)). The addition of nitrogen in
219 combination with the use of the newly designed X skimmer cone and Jet sample cone in
220 Neptune Plus improved the signal intensity of Hf, Yb and Lu by a factor of 5.3, 4.0 and 2.4,
221 respectively. All data were acquired on zircon in single spot ablation mode at a spot size of 44
222 μm and the energy density of laser ablation is about 5.3 J/cm². Each measurement consisted
223 of 20 s of acquisition of the background signal followed by 50 s of ablation signal acquisition.
224 Detailed operating conditions for the laser ablation system and the MC-ICP-MS instrument
225 and analytical method are the same as description by [Hu et al. \(2012\)](#). We applied the
226 directly obtained βYb value from the zircon sample itself in real-time ([Liu et al., 2010](#)). The
227 $^{179}\text{Hf}/^{177}\text{Hf}$ and $^{173}\text{Yb}/^{171}\text{Yb}$ ratios were used to calculate the mass bias of Hf (βHf) and Yb
228 (βYb), which were normalized to $^{179}\text{Hf}/^{177}\text{Hf}=0.7325$ and $^{173}\text{Yb}/^{171}\text{Yb}=1.1248$ ([Blichert-Toft et](#)

al., 1997) using an exponential correction for mass bias. Interference of ^{176}Yb on ^{176}Hf was corrected by measuring the interference-free ^{173}Yb isotope and using $^{176}\text{Yb}/^{173}\text{Yb} = 0.7876$ (McCulloch et al., 1977) to calculate $^{176}\text{Yb}/^{177}\text{Hf}$. Similarly, the relatively minor interference of ^{176}Lu on ^{176}Hf was corrected by measuring the intensity of the interference-free ^{175}Lu isotope and using the recommended $^{176}\text{Lu}/^{175}\text{Lu} = 0.02656$ (Blichert-Toft et al., 1997) to calculate $^{176}\text{Lu}/^{177}\text{Hf}$. We used the mass bias of Yb (βYb) to calculate the mass fractionation of Lu because of their similar physicochemical properties. Off-line selection and integration of analyte signals, and mass bias calibrations were performed using ICPMSDataCal (Liu et al., 2010). The Hf isotopic compositions are presented in Appendix Table S2.

4.2 Whole-rock major-, trace-element and Sr-Nd isotopic analyses

Major elements were measured using X-ray fluorescence (XRF) on fused glass beads at the Department of Earth Sciences, the University of Hong Kong. Accuracies of the XRF analyses are estimated to be 2% for most major elements and 1% for SiO_2 .

Trace elements were analyzed on a Quadrupole ICP-MS at the State Key Laboratory of Ore Deposit Geochemistry, Institute of Geochemistry, Chinese Academy of Sciences, Guiyang. About 50 mg powders for each sample were digested in a mixture of 1 ml HF and 0.5 ml HNO_3 within closed beakers in high-pressure bombs. A standard solution containing single element Rh was used for external calibration, and standards OU-1 and AMH-1 were used as

reference materials. The analytical accuracies are better than 5% for elements with concentration more than 200 ppm and 5-10% for those less than 200 ppm. Detailed analytical method can be referred to [Qi et al. \(2000\)](#). The whole-rock major- and trace-element compositions are presented in [Appendix Table S3](#).

Sr-Nd isotopic compositions were measured at the State Key Laboratory of Isotope Geochronology and Geochemistry, Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, following the procedures described by [Li et al. \(2006\)](#). Sr and rare earth elements (REEs) were separated through cation columns. Nd fractions were then further separated by HDEHP-coated Kef columns. The Sr-Nd isotopic compositions were then measured on a Micromass Isoprobe multi-collector ICP-MS. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively. The reported $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of unknown samples were adjusted to the NBS987 standard $^{87}\text{Sr}/^{86}\text{Sr} = 0.71025$ and Shin Etsu JNdi-1 standard $^{143}\text{Nd}/^{144}\text{Nd} = 0.512110$, respectively. The Sr-Nd isotopic compositions are presented in [Appendix Table S4](#).

4.3 Mineral compositions

Back-scattered electron (BSE) images and mineral compositions were performed on a JEO LIXA-8100 Electron Microprobe at the State Key Laboratory of Isotope Geochronology and Geochemistry, Guangzhou Institute of Geochemistry, Chinese Academy of Sciences. The

accelerating voltage, current and beam diameter are 15 kV, 10 nA, and 1-2 μm , respectively.

The mineral compositions are presented in [Appendix Table S5](#).

5. Results

5.1 Zircon U-Pb dating and Hf-isotope compositions

Zircons from four representative samples, including two quartz diorites (samples CG-71 and HKAM172), one granodiorite (sample CG-79) and one mafic enclave (sample HKAM180), were separated for U-Pb isotopic dating. All the zircon grains show euhedral to subhedral prismatic shapes. Zircons from the former three samples are mostly 150-250 μm long and have length/width ratios of ca. 1.5-3.0 ([Figs. 3a, b](#)), while those from the mafic enclave are mostly 100-200 μm long and have greater length/width ratios predominantly of ca. 2.5-4.0. All the analyzed zircons show high Th/U ratios (0.19-1.96; [Appendix Table S1](#)). Integrated with the well developed oscillatory zoning ([Figs. 3a, b](#)), these zircons are considered to be magmatic in origin.

Sample CG-71 (from the Chiketaman pluton): Eighteen zircon grains were analyzed for U-Pb isotopic compositions. Except for the zircon grain CG-71-2, others yield concordance of 97-99% and have $^{206}\text{Pb}/^{238}\text{U}$ ages varying from 396 to 384 Ma with a weighted mean of 389 ± 2 Ma (MSWD = 0.50, $n=17$; [Fig. 3a](#)). Twelve out of the eighteen dated zircons were chose for in-situ Hf-isotope analysis. They yielded uniform Hf isotopic compositions with $\epsilon_{\text{Hf}}(t)$ values of

8.8-10.0 (mean = 9.4 ± 0.2 , MSWD = 1.09) and two-stage Hf model ages (T_{DM2}) from 1004 to 895 Ma (mean = 948 ± 28 , MSWD = 0.55; [Appendix Table S2](#); [Fig. 3e](#)).

Sample CG-79 (from the Yaloman pluton): Seventeen out of the analyzed twenty zircons are concordant or nearly concordant. Most of them give $^{206}\text{Pb}/^{238}\text{U}$ age ranging from 395 to 377 Ma with a weighted mean of 387 ± 3 Ma (MSWD = 0.86, $n = 16$; [Fig. 3b](#)). Zircon grain CG-79-19 has a slightly older $^{206}\text{Pb}/^{238}\text{U}$ age of 406 ± 8 Ma. Twelve representative dated zircons were analyzed for Hf-isotope compositions and yielded $\epsilon_{\text{Hf}}(t)$ values varying from 8.0 to 11.1 (mean = 9.5 ± 0.6 , MSWD = 6.8; [Appendix Table S2](#); [Fig. 3f](#)). Correspondingly, the T_{DM2} values range from 1072 to 792 Ma with a weighted mean of 935 ± 57 (MSWD = 3.4).

Sample HKAM172 (from the Ust-Chuya pluton): The analyzed thirteen representative zircons are all concordant or nearly concordant. They have $^{206}\text{Pb}/^{238}\text{U}$ ages varying from 395 to 382 Ma, with a weighted mean value of 389 ± 4 Ma (MSWD = 0.34; [Fig. 3c](#)).

Sample HKAM180 (from the Kadrin pluton): Ten representative zircons were analyzed and nine of them are concordant. These zircons yielded a restrict range of $^{206}\text{Pb}/^{238}\text{U}$ ages (396-386 Ma), which gave a weighted mean of 389 ± 4 Ma (MSWD = 0.54; [Fig. 3d](#)). This age is consistent with the emplacement age of the granodiorite host (393 ± 10 Ma; [Cai et al., 2014](#)).

5.2 Whole-rock major- and trace-element compositions

Quartz diorites from the Chiketaman and Ust-Chuya plutons yield SiO_2 and $\text{Na}_2\text{O} + \text{K}_2\text{O}$

contents of 60.2-63.3 wt.% and 3.96-5.55 wt.%, respectively (Appendix Table S3), plotting in the diorite field in the SiO_2 versus $\text{Na}_2\text{O} + \text{K}_2\text{O}$ diagram (Fig. 4a; Middlemost, 1994). These rocks have relatively high contents of TiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3^{\text{T}}$, MgO and CaO , but low contents of Na_2O and K_2O (Figs. 5a-g). The low total alkali contents suggest that the quartz diorites belong to the sub-alkaline series (Fig. 4a; Irvine and Baragar, 1971). Additionally, these rocks have low $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$ contents and $\text{FeO}^{\text{T}}/(\text{FeO}^{\text{T}} + \text{MgO})$ ratios, indicative of calcic to calc-alkalic and magnesian geochemical affinities (Figs. 4b-c; Frost et al., 2001). The A/CNK ($\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$, molar ratios) values of the quartz diorites are 0.94-1.01, indicating metaluminous composition (Fig. 4d). All these rocks show parallel rare earth elements (REEs) patterns in the chondrite normalized diagrams ($\Sigma\text{REEs} = 80.4\text{-}117$ ppm; Appendix Table S3), with enrichment of light rare earth elements (LREEs, $(\text{La/Yb})_{\text{N}} = 4.95\text{-}7.33$), nearly flatten heavy rare earth elements (HREEs, $(\text{Gd/Yb})_{\text{N}} = 1.15\text{-}1.22$) and moderate negative Eu anomalies ($\delta\text{Eu} = 0.70\text{-}0.81$) (Figs. 6a, c). In the primitive mantle normalized trace-element diagram, the quartz diorites show pronounced enrichment in Pb and Sr, as well as depletion in Nb, Ta and Ti (Figs. 6b, d).

Granodiorites from the Kadrin and Yaloman plutons yield higher SiO_2 (67.3-69.9 wt.%), Na_2O and K_2O but lower TiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3^{\text{T}}$, MgO and CaO contents (Appendix Table S3; Figs. 4a, 5a-g), in comparison with the quartz diorites. In the SiO_2 versus $\text{Na}_2\text{O} + \text{K}_2\text{O}$ diagram, these granodiorites are plotted within the sub-alkaline field (Fig. 4a). The low $\text{Na}_2\text{O} + \text{K}_2\text{O} -$

324 CaO contents and $\text{FeO}^{\text{T}}/(\text{FeO}^{\text{T}} + \text{MgO})$ ratios further suggest that these rocks belong to the
325 calcic to calc-alkalic and magnesian series (Figs. 4b-c; Frost et al., 2001). The A/CNK values of
326 the granodiorites are 1.01 to 1.05, showing weakly peraluminous compositions (Fig. 4d). All
327 these rocks are featured by parallel chondrite normalized RREs patterns ($\Sigma\text{REEs} = 92.5\text{-}117$
328 ppm), enrichment of LREEs ($(\text{La}/\text{Yb})_{\text{N}} = 6.13\text{-}8.18$), negligible differentiation of HREEs
329 ($(\text{Gd}/\text{Yb})_{\text{N}} = 1.19\text{-}1.35$) and moderate negative Eu anomalies ($\delta\text{Eu} = 0.67\text{-}0.87$) (Fig. 6e;
330 Appendix Table S3). Enrichment in Pb and Sr, together with depletion in Nb, Ta and Ti, are
331 observed in the primitive mantle normalized trace-element diagram (Fig. 6f).

332 Mafic enclaves have lower SiO_2 contents of 46.0-56.1 wt.% (except for the sample
333 HKAM177 with SiO_2 content of 60.7 wt.%), but higher TiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3^{\text{T}}$, MgO and CaO
334 contents, than their granitoid hosts (Figs. 5a-g). In the SiO_2 versus $\text{Na}_2\text{O} + \text{K}_2\text{O}$ diagram, they
335 plot in the gabbro, gabbroic diorite and diorite fields (Fig. 4a). These rocks show relatively
336 high $\text{Mg}^{\#}$ ($\text{MgO}/(\text{MgO} + \text{FeO}^{\text{T}})$, molar ratios) of 38.9-47.2 and low A/CNK values of 0.76-0.97
337 (Fig. 4d; Appendix Table S3). Distinct from other mafic enclaves, sample HKAM171 has lower
338 Al_2O_3 and higher MgO content (Figs. 5b, d). The mafic enclaves show REEs contents varying
339 from 98.7 to 147 ppm, with weak to moderate LREEs enrichment ($(\text{La}/\text{Yb})_{\text{N}} = 1.96\text{-}4.87$),
340 pronounced negative Eu anomalies ($\delta\text{Eu} = 0.42\text{-}0.75$, except for sample C14-1 with such a
341 value of 1.05), and nearly flat HREEs patterns ($(\text{Gd}/\text{Yb})_{\text{N}} = 1.03\text{-}1.30$) (Figs. 6a, c, d). These
342 samples show trace-element patterns similar to their granitoid host, including depletion in

Nb, Ta and Ti, and enrichment in Pb and Sr (Figs. 6b, d, f).

5.3 Whole-rock Sr-Nd isotopic compositions

The seven representative granitoids, including quartz diorites and granodiorites, yield initial $^{87}\text{Sr}/^{86}\text{Sr}$ (390 Ma) ratios from 0.704476 to 0.705436, ϵ_{Nd} (390 Ma) values of 1.9-3.0 and two-stage Nd model ages (T_{DM2}) of 989-899 Ma (Appendix Table S4). The five representative mafic enclaves yield a slightly wider range of initial $^{87}\text{Sr}/^{86}\text{Sr}$ (390 Ma) ratios from 0.703374 to 0.705675, but have ϵ_{Nd} (390 Ma) values of 2.4-3.3 and T_{DM2} values of 951-876 Ma largely overlapping with those of the granitoid hosts.

5.4 Mineral compositions

5.4.1 Amphibole

Amphiboles from the granitoid and their hosted mafic enclaves yield indistinguishable chemical compositions (Appendix Table S5). They have Ca and (Na + Ca) in the B-site varying from 1.61 to 1.76 a.p.f.u. and 1.79 to 1.92 a.p.f.u., respectively, therefore belonging to the calcic sub-group (Leake et al., 1997). The (Na + K) values in the A-site are 0.23-0.50 a.p.f.u. and Si^{IV} in the T-site range from 6.70-7.16 a.p.f.u.. These amphiboles yield $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratios of 0.50-0.72 and they can be further classified into magnesiohornblende (Fig. 8a) based on the nomenclature scheme proposed by Leake et al. (1997).

5.4.2 Plagioclase

Plagioclases from the diorite mostly show normal zoning. The cores of these plagioclases vary from andesine to labradorite, with X_{An} ($X_{An} = Ca/(Ca + Na + K)$, molar ratios) values ranging from 0.42 to 0.63 (Appendix Table S5). The rims show lower calcium contents ($X_{An} = 0.21-0.36$) and are grouped into oligoclase to andesine. In contrast to the diorites, plagioclases from the granodiorite generally yield lower CaO contents. Their cores and rims show X_{An} values varying from 0.35 to 0.48 (i.e., andesine) and 0.07-0.31 (albite to oligoclase), respectively. Besides, a few plagioclase grains show core-mantle-rim structures (Fig. 2h), with an increase of calcium from the core ($X_{An} = 0.37-0.39$) to the mantle ($X_{An} = 0.38-0.46$) and a subsequent decrease to the rim ($X_{An} = 0.29-0.32$). The plagioclases from the mafic enclaves are characterized by sharp boundary between the cores and rims in BSE images (Fig. 2g). The cores have X_{An} values varying from 0.40 to 0.57 and accordingly they are grouped into andesine to labradorite. In contrast, the rims possess much lower calcium contents ($X_{An} = 0.15-0.37$) and compositionally vary from oligoclase to andesine.

5.4.3 Biotite

Biotites only occur in the granodiorites and their hosted mafic enclaves (Fig. 3) and they show indistinguishable geochemical compositions (Appendix Table S5). These biotites show moderate $Mg/(Mg + Fe^{2+} + Mn^{2+})$ ratios of 0.40-0.46 and they plot in the transitional area

between Mg-biotite and Fe-biotite (Foster, 1960; Fig. 8b). All the biotites are characterized by moderate enrichment of Al_2O_3 (13.5-14.1) and MgO (8.97-10.3) and depletion in FeO (21.6-23.9), which are typical of those from metaluminous calc-alkaline granitoids (Figs. 8c, d; Abdel-Rahman, 1994).

6 Discussions

6.1 Estimation of oxygen fugacity, pressure, temperature and water-in-melt

Oxygen fugacity ($f\text{O}_2$) exerts great control on the compositions of mafic minerals (Wones, 1981; Clowe et al., 1988; Ridolfi et al., 2010). For instance, amphiboles crystallized from oxidizing magmas would possess higher $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ and lower $\text{Fe}/(\text{Fe} + \text{Mg})$ ratios than those from the reducing ones (Wones, 1981). The amphiboles from the granodiorites, quartz diorites and their hosted mafic enclaves in this study yield indistinguishable compositions (Fig. 8). They are characterized by relatively high $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratios of 0.13-0.33 and low $\text{Fe}/(\text{Fe} + \text{Mg})$ ratios of 0.35-0.55 (Appendix Table S5), indicating crystallization under a relatively oxidizing condition (Fig. 9; Wones, 1981; Clowe et al., 1988).

The total Al contents in amphiboles largely rely on the pressure when they crystallized (e.g., Schmidt, 1992; Anderson and Smith, 1995). Anderson and Smith (1995) proposed a calibration on the amphibole Al-barometer considering the effect of temperature, which can be applied to relatively oxidizing conditions (i.e., $\text{Fe}/(\text{Fe} + \text{Mg})$ and $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Fe}^{2+})$ ratios in

ranges of 0.40-0.65 and above 0.20, respectively). The amphibole-plagioclase thermometer of [Holland and Blundy \(1994\)](#) is applied to calculate temperature, which is a function of pressure. Eight amphibole-plagioclase pairs from the quartz diorites, granodiorites and their hosted mafic enclaves are selected to calculate the emplacement pressure and temperature ([Appendix Table S5](#)). The quartz diorites (samples CG-74, CG-76 and HKAM170) yield crystallization temperature and pressure varying from 667 to 685 °C and from 2.22 to 3.68 kbar, respectively. Assuming that 1 kbar is equivalent to ca. 3.5 km of the crust, the emplacement depth is estimated to be ca. 7.77-12.9 km. In contrast, the granodiorite (sample HKAM175) show slightly lower crystallization temperature varying from 625 to 651 °C but higher pressure of 3.35-3.97 kbar. Accordingly, they were emplaced at ca. 11.7-13.9 km in depth. The mafic enclave (sample HKAM180) hosted in the granodiorites yields crystallization pressure of 2.99-3.30 kbar (equivalent to depth of 10.5-11.6 km) and crystallization temperature of 658-677 °C. Since all the calculations are based on the compositions of amphibole and plagioclase rims in contact with each other, the obtained pressures and temperatures possibly represent the final emplacement conditions.

Experimental data reveal that the aluminium index of amphiboles, which is a function of major cations, is well correlated to the H₂O content of the melts that they are crystallized from ([Ridolfi et al., 2010](#)). Accordingly, these researchers proposed an empirical formula to estimate the H₂O content dissolved in magmas solely based on the compositions of

crystallized amphiboles, showing a maximum relative error of ca. 15% (Ridolfi et al., 2010). Since this method has only been calibrated for volcanic rocks, constraints from other observations are combined in judging the validity of the estimation. Calculation shows that the quartz diorites, granodiorites and their hosted mafic enclaves yield broadly similar, but a slight increasing trend of, water contents of 5.03-6.25 wt.%, 5.39-6.70 wt.% and 5.91-7.17 wt.%, respectively (Appendix Table S5). A common feature is that mafic minerals of these rocks, especially those mafic enclaves and quartz diorites, are dominated by amphibole (Figs. 2b-d), the stability of which largely relies on the H₂O content in the magma (e.g., Tang et al., 2012 and references therein). The crystallization of amphibole, instead of orthopyroxene, probably indicates an oxidizing environment with relatively high water contents in melts (> 5 wt.%; e.g., Bogaerts et al., 2006; Tang et al., 2012 and references therein). This is consistent with the estimations of oxygen fugacity and water contents via the compositions of amphiboles.

6.2 Petrogenesis of the mafic enclaves

The mafic enclaves in this study are generally round to sub-round in shape and show homogeneous microstructures with no sign of recrystallization and deformation (Figs. 2a, c, d, f). They have mineral assemblages comparable to the granitoid hosts but contain higher abundance of amphiboles ($\pm$ biotites) as a result of their more mafic compositions. Except for

sample C14-1, other mafic enclaves all show finer-grained textures than their granitoid hosts. Previous studies revealed that residues formed after melt extraction would always experience high pressure-temperature metamorphism and thus have coarse-grained textures and/or inhomogeneous microstructures (e.g., banded structures; [Vernon, 1984](#); [Maas et al., 1999](#)). Likewise, minerals formed by early fractional crystallization are generally coarse-grained with euhedral shapes ([Vernon, 1983](#)). All these characteristics suggest that the investigated mafic enclaves (except for the sample C14-1) are unlikely to represent residues or early-formed crystal aggregates. Additionally, acicular apatites are common in the mafic enclaves ([Fig. 2f](#)), indicating crystallization in a quenched environment ([Vernon, 1984](#)). This is generally the case when a small volume of hotter mafic magmas commingled with cooler felsic ones, resulting in rapid crystallization of minerals in the mafic magmas during the thermal equilibration ([Vernon, 1984](#); [Sparks and Marshall, 1986](#)). The plagioclase grains in the mafic enclaves are commonly zoned, with relatively high calcium contents in the cores ($X_{An} = 0.40-0.57$) and abrupt decreases of this element in the thin rims ($X_{An} = 0.15-0.37$; [Fig. 2g](#)). Crystallization in a closed magma system cannot account for such kind of compositional discontinuities. We propose that the plagioclase cores were probably crystallized from a mafic melt while the rims from a more felsic one. Zircon U-Pb dating shows that the mafic enclave sample HKAM180 was crystallized at 389 ± 4 Ma ([Fig. 3d](#)), which is nearly identical to the emplacement age of the granodiorite host (ca. 387-393 Ma; [Cai et al., 2014](#) and this

study). All the data above suggest that the studied mafic enclaves probably represent coeval mafic magmas commingled with the granitoid host magmas. The coarse-grained mafic enclave sample C14-1 is dominated by amphiboles and plagioclases, and the amphiboles contain abundant fine-grained plagioclase inclusions (ca. 40-70 μm ; Fig. 2c). The positive Eu anomaly ($\delta\text{Eu} = 1.05$; Appendix Table S3) may imply that this enclave possibly represents mineral aggregates formed during the early crystallization process. The difference in grain sizes between the amphiboles and enclosed plagioclases can be possibly attributed to their different nucleation rates in a relatively hydrous crystallization environment.

Experimental studies have shown that magmas generated by partial melting of crustal rocks, even the most mafic ones (i.e., basaltic rocks), generally have $\text{Mg}^\#$ lower than 40 (e.g., Rapp et al., 1991; Rapp and Watson, 1995; Patiño Douce, 1999). The mafic enclaves of this study are characterized by low SiO_2 contents (mostly of 46.0-56.1 wt.% except for sample HKAM177 with 60.7 wt.%), coupled with relatively high $\text{Mg}^\#$ (38.9-56.5) and transition elements (e.g., V, Sc) contents (Appendix Table S3). This indicates that mantle-derived mafic melts probably made a significant contribution to form these mafic enclaves. However, the X_{An} values of plagioclase cores are lower than those from primitive basaltic rocks (e.g., Kuritani et al., 2014), implying that the mantle-derived melts possibly already underwent fractional crystallization and/or assimilation of crustal materials to some extent before the plagioclase were crystallized. The Nd-isotope compositions of the mafic enclaves ($\epsilon_{\text{Nd}}(390 \text{ Ma})$

= 2.4-3.3) are much lower than that of the contemporaneous depleted mantle (Fig. 7b). This was possibly related to two different petrogenetic processes. Firstly, the mantle-derived basaltic melts possibly assimilated some enriched crustal components before they were entrained within the granitoid host magmas. Secondly, certain amount of the granitoid host magmas was involved during the formation of these mafic enclaves, as revealed by the petrographic observation. We note that the Nd-isotope compositions of the mafic enclaves are similar to those of their granitoid hosts in each individual pluton of the Yaloman intrusive complex (Fig. 10 inset; Appendix Table S4). This could be attributed to the rapid diffusive isotopic equilibrium after entrainment (e.g., Leshner, 1990; Fourcade and Javoy, 1991).

6.3. Petrogenesis of the granitoid hosts

6.3.1 The role of magma mixing and fractional crystallization

The largely varied SiO₂ contents of the granitoids and their hosted mafic enclaves form linear arrays with most other major elements, as well as trace elements such as Zr, Th and V (Fig. 5). Such linear compositional variations can be attributed to either fractional crystallization or mixing between two contrasting melts (e.g., basaltic and felsic; Lee and Bachmann, 2014). Zirconium (Zr) and phosphorus (P) have been recently applied in modeling the fractional crystallization and magma mixing processes (Lee and Bachmann, 2014). These two elements are generally under-saturated in basaltic magmas and can be progressively

495 enriched in the evolved ones till the saturation of zircons and apatites, when Zr and P will be
496 significantly removed. Accordingly, a magmatic suite formed by fractional crystallization from
497 a basaltic magma would be generally featured by kink-shaped patterns in the SiO₂ versus Zr
498 and P₂O₅ diagrams (Figs. 5h, i; Lee and Bachmann, 2014). However, the SiO₂ contents of the
499 granitoids and their mafic enclaves within the Gorny Altai terrane mostly form linear
500 correlations with Zr and P₂O₅ (Figs. 5h, i), indicating that magma mixing, instead of fractional
501 crystallization, probably exerted a major control in the formation of these rocks. Furthermore,
502 the mafic enclaves from the Kadrin and Yaloman plutons show much more pronounced
503 negative Eu anomalies ($\delta\text{Eu} = 0.42\text{-}0.73$) than those of their host granodiorites ($\delta\text{Eu} =$
504 $0.68\text{-}0.87$; Appendix Table S3; Fig. 6e). This indicates that these granodiorites were unlikely
505 to be formed through fractional crystallization from the mafic melts that represented by the
506 mafic enclaves. On the contrary, the broadly negative correlation between the whole-rock
507 $\epsilon_{\text{Nd}}(390 \text{ Ma})$ values and SiO₂ contents of the granitoids from different plutons (Fig. 10) favors
508 that at least two kinds of magmas with contrasting Nd-isotope compositions were possibly
509 involved. This conclusion is further supported by the observation that some plagioclase
510 grains from the granitoids show reverse compositional zoning, with X_{An} values increasing
511 from the core to mantle and then decreasing to the rim (Fig. 2h). Importantly, the X_{An} values
512 of plagioclase mantles (0.38-0.46) are comparable to those of the plagioclase cores
513 (0.40-0.57) from the mafic enclaves (Appendix Table S5). This implies that mantle-derived

basaltic melts, which served as a crucial component of the mafic enclaves, possibly made an important contribution to the formation of these granitoids.

However, the whole-rock $\epsilon_{\text{Nd}}(390 \text{ Ma})$ values and SiO_2 contents of the granitoids do not strictly follow a mixing trend (Fig. 10), implying that some degree of fractional crystallization possibly contributed to their diverse compositions ranging from diorite to granodiorite (Fig. 4a). The Dy/Yb ratios slightly decrease while the La/Yb ratios increase from quartz diorites to granodiorites (Fig. 11a). Additionally, one granodiorite sample (i.e., HKAM182) shows weakly concave-up REEs patterns (Fig. 6e). These geochemical features could be attributed to the fractionation of amphiboles to some extent, which possess high partition coefficients for MREEs (Bottazzi et al., 1999; Davidson et al., 2009). The investigated quartz diorites and granodiorites yield comparable δEu values of 0.70-0.81 and 0.67-0.87 (Appendix Table S3; Fig. 11b), respectively, implying negligible fractionation of plagioclase. As shown in Appendix Table S5, biotites from the granodiorites in this study are featured by peraluminous compositions with A/CNK values of 1.29-1.35. Meanwhile, biotites generally have much higher partition coefficients for vanadium (V) over thorium (Th) (Bea et al., 1994). As a result, the fractionation of this mineral would give rise to decrease of both V/Th ratios and A/CNK values. However, this is not the case for our granitoids (Fig. 11c), therefore arguing against obvious fractionation of biotites in their formation.

To conclude, the geochemical variation among the granitoids was mainly driven by

533 magma mixing, with minor fractionation of amphibole.

534

535 6.3.2 Crustal signature of the granitoid magmas

536 The granitoid samples possess metaluminous to slightly peraluminous geochemical
537 compositions (Fig. 4d). This is consistent with the observation that amphibole ($\pm$ biotite) is
538 the dominant mafic mineral in these rocks, while aluminum-enriched minerals, such as
539 muscovite, cordierite and garnet, are absent (Figs. 2b, d, e). In addition, the granitoids have
540 $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios higher than 1.0 and show negative correlation between P_2O_5 and SiO_2
541 contents (Fig. 5h). The geochemical features, together with the mineral compositions,
542 suggest that these rocks were unlikely to be formed by partial melting of meta-sedimentary
543 rocks (i.e., S-type granitoids) but belong to the I-type sub-group with meta-igneous sources
544 (Chappell and White, 1992).

545 The granitoids are featured by relatively low SiO_2 contents (60.2-69.9 wt.%) and
546 $\text{Al}_2\text{O}_3/(\text{FeO}^{\text{T}} + \text{MgO} + \text{TiO}_2)$ ratios, coupled with relatively high $\text{Mg}^{\#}$ (32.3-44.2) and $\text{Al}_2\text{O}_3 +$
547 $\text{FeO}^{\text{T}} + \text{MgO} + \text{TiO}_2$ contents (Appendix Table S3), indicating a mafic source (e.g.,
548 amphibolites; Fig. 12; Beard and Lofgren, 1991; Rapp et al., 1991; Rapp and Watson, 1995;
549 Patiño Douce, 1999). Experimental studies revealed that either low degree of dehydration
550 partial melting or water-saturated partial melting of amphibolites would leave behind
551 substantial amount of amphiboles in the residues, generating peraluminous

intermediate-felsic melts (Beard and Lofgren, 1991; Rapp et al., 1991; Rapp and Watson, 1995). The granitoids in this study have low A/CNK values of 0.94-1.05 (Appendix Table S3), without or with only slight depletion of MREEs (Figs. 6a, c, e) that can be preferentially hosted in amphiboles (Bottazzi et al., 1999; Davidson et al., 2009). This suggests that these granitoids were probably formed by relatively large degree of dehydration melting of amphibolites. The relatively high HREEs contents, moderate enrichment of LREEs and low Sr/Y ratios (Fig. 6; Appendix Table S3) further imply that the partial melting was likely to occur at a comparatively low pressure setting (< 12 kbar), leaving behind a residue without garnet. Moreover, the moderately negative Eu anomalies ($\delta\text{Eu} = 0.67\text{-}0.87$; Appendix Table S3) suggest that plagioclase, a mineral with high Eu partition coefficient, existed in the residue after partial melting.

The granitoids show whole-rock $\epsilon_{\text{Nd}}(t)$ values of 1.9-3.0 and correspondingly two-stage Nd model ages of 0.90-0.99 Ga (Appendix Table S4). Similarly, zircons from these rocks give $\epsilon_{\text{Hf}}(t)$ values and two-stage Hf model ages of 8.1-11.2 and 0.79-1.07 Ga, respectively (Appendix Table S2). These data suggest that the granitoids were probably dominated by remelting of Neoproterozoic crust. In the latest Mesoproterozoic, the Rodinia supercontinent broke up with the birth of the Paleo-Asian Ocean (Khain et al., 2002; Dobretsov et al., 1995). The coeval oceanic crustal relicts are partly preserved as some ophiolites in the northern CAO, from which the plagiogranites were dated with zircon U-Pb ages of ca. 1.02 Ga (Khain

[et al., 2002](#)). Close to the southwestern margin of the Siberian continent (present coordinate), the Kuznetsk-Altai arc was probably initiated in the late Neoproterozoic ([Buslov et al., 1993, 2002, 2013](#); [Ota et al., 2002, 2007](#); [Glorie et al., 2011](#); [Chen et al., 2015](#)), generating primitive island arc rocks and associated accretionary complexes (e.g., [Ota et al., 2002, 2007](#); [Safonova et al., 2004, 2008, 2011](#); [Utsunomiya et al., 2009](#); [Kruk et al., 2010](#)). Recent detrital zircon studies showed that the subduction continued till ca. 470 Ma ([Buslov et al., 2002](#); [Chen et al., 2015](#)). During this prolonged subduction process, voluminous oceanic crustal relicts from the oceanic plate and paleo-seamounts/oceanic plateaus were possibly accumulated and buried in depth. The juvenile isotopic signature suggests that these rocks likely served as the main sources in the generation of the granitoids ([Utsunomiya et al., 2009](#); [Kruk et al., 2010](#); [Kruk et al., 2011](#); [Cai et al., 2014](#)).

6.3.3 Implication on the geodynamic evolution

Granitoids and their volcanic equivalents can form in subduction, collisional and post-collisional settings, yielding diverse geochemical compositions as a result of distinct source rocks and physical conditions (e.g., [Pearce et al., 1984](#); [Batchelor and Bowden, 1985](#); [Harris et al., 1986](#); [Barbarin, 1999](#); [Patiño Douce, 1999](#)).

It was suggested that the early-middle Devonian andesite-dacite-rhyolite suites, represented by the Ongudai and Kuratin volcanic complexes, were generated under an active

continental margin plunging beneath the Gorny Altai terrane, while the middle-late Devonian (ca. 395-360 Ma) granitoid plutons, including those of the Yaloman intrusive complex in this study, resulted from the collision with the Altai-Mongolian terrane (Buslov et al., 1993; Glorie et al., 2011; Safonova, 2014 and references therein). However, the field occurrence, petrographic characteristics and whole-rock geochemical compositions argue against the collisional origin for the granitoids of Yamolan intrusive complex. Previous studies have shown that collision-related granitoid intrusions, such as the Himalayan leucogranites, are generally small in volume, coupled with relatively strong deformation and peraluminous compositions (e.g., Harris et al., 1986; France-Lanord and Le Fort, 1988; Nalini Jr. et al., 2008; Searle et al., 2009). These rocks possibly represent pure crustal melts and are remarkably devoid of mafic enclaves with igneous origins (e.g., Le Fort et al., 1987). However, the granitoids of the Yaloman intrusive complex crop out at least ca. 700 km² (Figs. 1b, c) and do not show visible deformation in both macro- and micro-scales (Fig. 2). Importantly, these granitoid plutons are exclusively metaluminous to weakly peraluminous in composition (Fig. 4) and contain various proportions of igneous mafic enclaves epitomizing the input of mantle-derived melts (Fig. 2). These features are quite different from those counterparts generated in collision zones but compatible with granitoid batholiths in arc settings (e.g., Barbarin, 1999). Estimation from the mafic enclaves shows that the mantle-derived mafic magmas in generation of the investigated granitoids are featured by high water contents and

oxygen fugacity (see details in section 6.1). All these characteristics strongly argue for a continental arc-related setting for the Yaloman intrusive complex. This conclusion can be further supported by the tectonic discrimination diagrams, showing that nearly all the granitoids plot within the fields of typical volcanic arcs or destructive active plate margins (Fig. 13; Pearce et al., 1984; Batchelor and Bowden, 1985).

It should be noted that the Devonian granitoids and their volcanic equivalents with strong geochemical affinities to the subduction zone magmas crop out ubiquitously within the Altai-Mongolian terrane (e.g., Wang et al., 2006, 2009; Yuan et al., 2007; Sun et al., 2008, 2009; Cai et al., 2011, 2014, 2015), south of the CTUS suture zone (Figs. 1a, b). The remarkable similarities in temporal-spatial-petrogenetic relationship to the counterparts from the Gorny Altai terrane implies that the Devonian magmatic suites in both terranes were possibly generated by subduction of the same oceanic plate (Filippova et al., 2001 and references therein; Windley et al., 2007; Cai et al., 2014). This is consistent with recent investigations on the early Devonian sedimentary sequences from the Gorny Altai terrane and meta-sedimentary complexes along the CTUS suture zone, suggesting that the Gorny Altai and Altai-Mongolian terranes possibly had already amalgamated prior to the early Devonian (Chen et al., 2015; unpublished data). According to the paleogeographical reconstruction by Filippova et al. (2001), the Kazakhstan continent was possibly located in the central of the Paleo-Asian Ocean in the early Devonian that was divided into four oceanic

basins, namely the Ob-Zaisan, Uralian, Turkestan and Junggar-Balkhash Oceans (Fig. 14a). The subduction of Ob-Zaisan oceanic crust due to the approaching of the Kazakhstan continent probably resulted in the widespread magmatism along the Siberian continental margins (e.g., the Gorny Altai and Altai-Mongolian terranes).

6.4 An integrated model in formation of the granitoids and their hosted mafic enclaves

The occurrence of igneous mafic enclaves within the granitoid plutons of the Yaloman intrusive complex from the Gorny Altai terrane provides direct evidence for the interaction between the upper mantle and lower crust. Our data suggest that partial melting of pre-existing amphibolitic lower crust (Petford and Gallagher, 2001; Dufek and Bergantz, 2005), coupled with mixing with the underplated basaltic magmas from the mantle, is a plausible way to generate metaluminous to weakly peraluminous granitoid magmas in a continental arc setting. Based on the discussions above, we propose an integrated model to interpret the formation of the investigated granitoids and their hosted mafic enclaves as follows (Fig. 14b). At the beginning, substantial water-enriched basaltic magmas were underplated beneath the Gorny Altai terrane as a result of the subduction of the Ob-Zaisan Ocean. Some of them intruded in the lower crust and formed a hot zone (Annen et al., 2006). The heat and water released during the crystallization of these water-enriched magmas could significantly facilitate dehydration melting of surrounding crustal materials (i.e., amphibolites;

647 [Annen et al., 2006](#)), generating relatively large volumes of hydrous crustal melts.
648 Subsequently, these melts migrated upward, and pooled in the upper crust (ca. 7.77-13.9
649 km). The injection of the evolved mafic magmas, which were formed after incomplete
650 crystallization with/without assimilation of crustal materials during the emplacement, into
651 the crustal melts induced mixing between them to some extent. Locally, the unmixed evolved
652 mafic magmas are preserved as mafic enclaves. We consider that water played a crucial role
653 in these processes. Firstly, it can decrease the subsolidus temperature of the amphibolitic
654 lower crust (e.g., [Annen et al., 2006](#) and references therein), therefore making relatively
655 large-scale partial melting possible. Secondly, the viscosities of both the mafic and crustal
656 melts would be decreased due to their water-enriched properties, promoting migration of
657 these melts to the upper crust. Lastly, when water-enriched mafic magmas were emplaced
658 beneath the more felsic crust-derived melts in a high-level magma chamber, the
659 crystallization of the mafic magmas due to heat transfer would make the mafic magmas have
660 bulk density equal to or even lower than the overlying felsic ones ([Herbert et al., 1982](#)). The
661 resultant overturning would cause intimate mixing between these two kinds of magmas
662 ([Herbert et al., 1982](#)). To conclude, the mantle-derived basaltic magmas not only provided
663 substantial heat to melt the lower crust, but also mixed with the crust-derived melts to form
664 diverse granitoids.
665

7. Conclusions

(1) The mafic enclaves within the granitoid plutons of the Yaloman intrusive complex (ca. 390 Ma) from the Gorny Altai terrane were igneous in origin and represented coeval magmatic globules commingled with their host magmas.

(2) The granitoids were predominantly derived from remelting of pre-existing Neoproterozoic crustal materials due to underplating of mantle-derived mafic magmas. Subsequent mixing of crust-derived melts with the evolved mafic melts in the upper crust was possibly responsible for the geochemical diversity of the granitoids.

(3) The Devonian granitoids from the Gorny Altai terrane were probably generated in a continental arc-related tectonic setting, where significant vertical crustal growth and differentiation were achieved as a result of underplating of subduction-related mafic magmas.

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1020

1021 **Figure captions:**

1022 Fig. 1 (a) Simplified tectonic architecture of the northwestern Central Asian Orogenic Belt
1023 (CAOB, modified after [Buslov and Safonova \(2010\)](#) and [Buslov \(2011\)](#)). (a) Inset shows that
1024 the CAO is situated among the Siberian, Tarim and North China continents. (b) Distribution
1025 of intrusive rocks in the Gorny Altai terrane (GA), northern segment of the Altai-Mongolian
1026 terrane (AM) and the Charysh-Terekta-Ulagan-Sayan (CTUS) suture zone between these two

1027 terranes. Complexes along this suture zone are also shown. This map is modified after
1028 [Daukeev et al. \(2008\)](#), [Glorie et al. \(2011\)](#) and [Kuibida et al. \(2014\)](#). (c) The investigated
1029 granitoid plutons of the Yaloman intrusive complex and surrounding volcano-sedimentary
1030 units (modified after the Russian geological map). Abbreviations: SC, Siberian continent; TC,
1031 Tarim continent; NCC, North China continent; WS, West Sayan; B, Barguzin; TM,
1032 Tuva-Mongolian terrane; Cam., Cambrian; O, Ordovician; S, Silurian; D, Devonian; C,
1033 Carboniferous; Mes., Mesozoic; Cen., Cenozoic.

1034

1035 Fig. 2 The field occurrence (a) and thin section photomicrographs (b-h) of representative
1036 granitoids and their hosted mafic enclaves from the Yaloman intrusive complex. (a) A mafic
1037 enclave about 20 cm long is enclosed by the quartz diorite. (b) Sample CG-72 is a
1038 medium-grained quartz diorite predominantly composed of plagioclase (Pl), amphibole
1039 (Amp), quartz (Qtz) and K-feldspar (Kfs). (c) Sample C14-1 is a medium-grained mafic enclave
1040 with mineral assemblages mainly of amphibole and plagioclase. Note quite a lot tiny
1041 plagioclase grains are enclosed by the coarser-grained amphiboles. (d) HKAM171 is a
1042 fine-grained mafic enclave with dominance of plagioclase, amphibole, quartz and K-feldspar.
1043 A few coarse-grained plagioclases can be observed. (e) Sample HKAM174 is a
1044 medium-grained granodiorite consisting mainly of plagioclase, K-feldspar, quartz, biotite (Bi)
1045 and amphibole. (f) Sample HKAM180 is a fine-grained mafic enclave dominated by

1046 plagioclase, biotite, amphibole and quartz. Acicular apatites (Ap) are commonly observed. (g)
1047 A backscattered electron (BSE) image of plagioclase grains from the mafic enclave sample
1048 HKAM180, showing an abrupt decrease of calcium content from the core to the rim ($X_{An} =$
1049 $Ca/(Ca + Na + K)$, atomic ratios). (h) A BSE image of a plagioclase grain from the granodiorite
1050 sample CG-80, exhibiting a core-mantle-rim structure as a result of variation in calcium
1051 contents. The symbols “(-)” and “(+)” in the upper right angles of figures b-f denote
1052 plan-polarized and perpendicular polarized light, respectively. The X_{An} values in (g) and (h)
1053 can be referred to [Appendix Table S5](#).

1054

1055 Fig. 3 U-Pb concordia diagrams and plots of age (Ma) versus $\epsilon_{Hf}(t)$ for zircons from the
1056 granitoids and their hosted mafic enclaves from the Yaloman intrusive complex. CL images of
1057 representative zircon grains are shown. The circles are 32 μm in diameter, representing the
1058 spots for zircon U-Pb analyses.

1059

1060 Fig. 4 Classification of the granitoids and their hosted mafic enclaves from the Yaloman
1061 intrusive complex. (a) The total alkalis ($Na_2O + K_2O$) versus SiO_2 diagram is after [Middlemost](#)
1062 [\(1994\)](#). The three mafic enclave samples in the gabbro field are actually mafic diorite without
1063 pyroxenes. The dashed line separating the alkaline and sub-alkaline fields is after [Irvine and](#)
1064 [Baragar \(1971\)](#). (b) $Na_2O + K_2O - CaO$ versus SiO_2 and (c) $FeO^T/(FeO^T + MgO)$ versus SiO_2

1065 diagrams are after [Frost et al. \(2001\)](#). (d) A/CNK versus A/NK diagram. The I-S line is after
1066 [Chappell and White \(2001\)](#).

1067

1068 Fig. 5 Binary plots showing the compositional variation of the granitoids and their hosted
1069 mafic enclaves from the Yaloman intrusive complex. In (f), (g) and (i), the mafic enclaves are
1070 featured by slightly higher Na₂O, K₂O and Rb, which are possibly due to the entrainment of
1071 alkali feldspar. The fractional trends of basaltic rocks under three different conditions (i.e., (1)
1072 0.1 Gpa and no volatiles in the basaltic magma; (2) 0.3 Gpa with 1 wt.% H₂O in the basaltic
1073 magma and (3) 0.3 Gpa with 4 wt.% H₂O in the basaltic magma) in (h) and (i) are after the
1074 Rhyolite-MELTS modeling results by [Lee and Bachmann \(2014\)](#). The trend of I-type granitoids
1075 in (h) is after [Chappell and White \(1992\)](#). Symbols same as in Fig. 4.

1076

1077 Fig. 6 Chondrite-normalized REEs patterns and primitive mantle-normalized spider diagrams
1078 of the granitoids and their hosted mafic enclaves from the Yaloman intrusive complex. Data
1079 of the chondrite and primitive mantle are from [Sun and McDonough \(1989\)](#). Symbols same
1080 as in Fig. 4.

1081

1082 Fig. 7 Plots of ⁸⁷Sr/⁸⁶Sr(390 Ma) versus ε_{Nd}(390 Ma) values (a) and crystallization ages (Ma)
1083 versus ε_{Nd}(390 Ma) values (b) for the granitoids and their hosted mafic enclaves from the

1084 Yaloman intrusive complex. Symbols same as in Fig. 4.

1085

1086 Fig. 8 Compositions of amphiboles and biotites from the granitoids and their hosted mafic
1087 enclaves from the Yaloman intrusive complex. Plots (a) and (b) are after [Leake et al. \(1997\)](#)
1088 and [Foster \(1960\)](#), respectively, while (c) and (d) are after [Abdel-Rahman \(1994\)](#). Symbols
1089 same as in Fig. 4.

1090

1091 Fig. 9 Al^{IV} values versus $Fe/(Fe + Mg)$ ratios of amphiboles from the granitoids and their
1092 hosted mafic enclaves from the Yaloman intrusive complex. The low, intermediate and high
1093 oxygen fugacity fields are after [Anderson and Smith \(1995\)](#). Symbols same as in Fig. 4.

1094

1095 Fig. 10 SiO_2 (wt.%) versus $\epsilon_{Nd}(390 \text{ Ma})$ values for the granitoids from the Yaloman intrusive
1096 complex. The mafic enclaves are included in the figure inset for comparison. Symbols same
1097 as in Fig. 4.

1098

1099 Fig. 11 Plots of La/Yb versus Dy/Yb , SiO_2 (wt.%) versus Eu/Eu^* , A/CNK versus V/Th for the
1100 granitoids and their hosted mafic enclaves from the Yaloman intrusive complex. Symbols
1101 same as in Fig. 4.

1102

1103 Fig. 12 Compositional comparison between the granitoid hosts in this study with those
1104 partial melts from meta-pelites, meta-greywackes and amphibolites (modified after [Patiño](#)
1105 [Douce, 1999](#)). Symbols same as in Fig. 4.

1106

1107 Fig. 13 Tectonic discrimination diagrams showing that the granitoid hosts in this study were
1108 probably formed in a destructive active plate margin or a volcanic arc setting. The fields for
1109 granitoids from various tectonic settings in (a) and (b) are referred to [Batchelor and Bowden](#)
1110 [\(1985\)](#) and [Pearce et al. \(1984\)](#), respectively. Symbols same as in Fig. 4.

1111

1112 Fig. 14 (a) A paleogeographic reconstruction diagram (modified after [Filippova et al. \(2001\)](#)
1113 and [Windley et al. \(2007\)](#)) showing that the Gorny Altai and Altai-Mongolian terranes were
1114 under a coherent active continental margin in ca. 390 Ma as a result of shrinking of the
1115 Ob-Zaisan Ocean (a branch of the Paleo-Asian Ocean). Abbreviations: AM, Altai-Mongolian
1116 terrane; BL, Barlyk arc; ChTS, Chinese Tien Shan; CTS, Central Tien Shan; EJ, East Junggar; GA;
1117 Gorny Altai terrane; KHM, Khanty-Mansi; NTS, North Tien Shan; MG, Magnitogorsk; NU,
1118 North Urals; STS, South Tien Shan; WS, West Sayan terrane. (b) A schematic diagram
1119 (modified after [Annen et al. \(2006\)](#) and [Jacob et al. \(2015\)](#)) showing the mechanism in
1120 formation of the granitoids and their hosted mafic enclaves from the plutons of the Yaloman
1121 intrusive complex.

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