Late Permian to Early Triassic mafic to felsic intrusive rocks from North Liaoning, North China: Petrogenesis and implications for Phanerozoic continental crustal growth

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Zircon U–Pb dating, whole-rock major oxide, trace element and Sr–Nd–Hf isotopic data are presented for the Late Permian to Early Triassic Shijianfang batholith from Faku in North Liaoning Province, North China. Two main magmatic suites are documented: one, with an emplacement age of ca. 260 Ma, is mafic to intermediate in composition, forming small gabbroic to monzonitic intrusions; the other is felsic and quartz monzonitic to granitic in composition, with an emplacement age of ca. 250 Ma. Geochemically, the rocks from the mafic suite show strong enrichment of LILE and LREE and depletion in HFSE, and also have moderately depleted isotopic compositions, with εNd(t) ranging from 0.7049 to 0.7054, εHf(t) from +2.72 to −1.82 and zircon εHf(t) values from +3.4 to +7.1. These features suggest derivation from high degree partial melting of a subduction-related metasomatized lithospheric mantle source. The rocks from the felsic suite range from 66.7 to 77.2 wt.% SiO2, and differentiate moderately depleted isotopic compositions, with εNd(t) from +2.72 to +0.55 and zircon εHf(t) values from +3.6 to +6.6. This suggests that the parental magma for the felsic suite originated from partial melting of mixed protoliths composed of juvenile basaltic underplate and ancient lower crustal materials. Subsequent fractional crystallization, with overprinting by magmatic hydrothermal fluids, can explain the geochemical variations of the felsic suite.

The juvenile character of both lithospheric mantle and crustal rocks, as recorded by the Shijianfang batholith, suggests that the northern Liaoning block forms part of a Phanerozoic accretionary orogenic belt. This observation indicates that the Chifeng-Kaiyuan Fault likely represents the Mesozoic lithospheric boundary between the North China Craton and the Xing-Meng Orogenic Belt in northern Liaoning. The Shijianfang batholith, together with Late Permian to Middle Triassic adakitic and A-type rocks, mafic-ultramafic cumulates and alkaline intrusions from the neighboring areas, constitutes an important post-collisional to intra-plate anorogenic magmatic province within the continental interior of the newly amalgamated North China–Mongolian plate. The unique mafic and felsic coupling in the Shijianfang batholith provides a good proxy record for the multiple-step vertical continental crustal growth in the newly amalgamated continental interior during the Phanerozoic, triggered by continued magmatic underplating due to lithospheric delamination and hot asthenospheric upwelling within a post-collisional to post-orogenic extensional regime.

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1. Introduction

The generation of continental crust is accomplished through igneous processes that transfer material from the upper mantle to the surface of the Earth at appropriate sites (Reymer and Schubert, 1986; Rudnick, 1995). Thus multiple linkages between crustal generation mechanism, favorable tectonic setting and potential magma proxy record have long been fundamental in fully
characterizing the growth and differentiation of the continental crust. The present near-consensus holds that the continental crust can grow by lateral accretion of arc complexes in subduction zones and vertical addition of mafic underplate at the crust–mantle interface (Rudnick, 1990; Frost et al., 2001a; Kemp and Hawkesworth, 2003). However, their relative importance in different tectonic environments remains debatable. Uncertainties also revolve around the detailed path through which the predominantly mantle-derived basaltic protoliths evolved into differentiated, felsic crust. Clearly, assessing any models for the formation of new crust is impaired by limited documentation of an appropriate magma proxy record.

In this respect, contemporaneous post-orogenic intrusive mafic and felsic magmatism may prove to be more instructive. On the one hand, as one of the common features of many orogens around the world (Bonin, 2004), post-orogenic intrusive mafic magmatism may not only serve as a diagnostic geodynamic tracer for constraining the tectonic evolution of ancient orogenic belts (Dewey, 1988), but also represent an important mechanism of continental growth (Johnson, 1991; Farmer, 2003). The origin of such magmatism is commonly attributed to lithospheric extension by orogenic collapse (Ruppel, 1995), slab breakoff (Davies and von Blanckenburg, 1995), convective thinning (England and Houseman, 1989), delamination of continental lithosphere (Kay and Kay, 1993) or magmatic underplating (Furlong and Fountain, 1986), or some combination of these. On the other hand, post-orogenic, intra-platonic tectonic settings constitute favorable sites for the formation of felsic magmas with juvenile isotopic signatures, as witnessed by numerous Precambrian and Phanerozoic examples, such as the Proterozoic anorogenic granite provinces of the southwestern USA (Frost et al., 1999), Paleozoic granites of southeastern Australia (Turner et al., 1992; Kay et al., 1997; King et al., 1997) and Paleozoic to Mesozoic granites of the Central Asian Orogenic Belt (CAOB) (Wickham et al., 1996; Han et al., 1997; Wu et al., 2000; Jahn et al., 2001; Wu et al., 2002, 2003a; Jahn, 2004; Chen and Arakawa, 2005; Liu et al., 2005). The documentation of such felsic rocks therefore provides significant information on their contribution to the build-up of the middle to upper continental crust (Turner et al., 1992; Frost and Frost, 1997; Frost et al., 2001a; Mushkin et al., 2003).

The Faku dome in northern Liaoning Province occupies a critical tectonic position that links a Phanerozoic orogen to the north, i.e., the Xing-Meng Orogenic Belt (XMOB) of the CAOB, with a Precambrian craton to the south, i.e., the North China Craton (NCC). Based on the notion that the migmatites and metamorphic complexes in the Faku area are of a Proterozoic age (LBGMR, 1989), the Faku dome has long been regarded as a Precambrian terrane along the northern margin of the NCC. However, the 1:50,000 scale geological mapping (LBGMR, 1998) and a recent 40Ar/39Ar geochronological study (Zhang et al., 2005a) revealed that these deformed and metamorphosed complexes, with a variety of protoliths including plutonic intrusions and supracrustal volcanic and sedimentary rocks, were genetically related to Triassic ducite shearing events. Zircon U-Pb dating further identified a number of Permian granitic intrusions (Zhang et al., 2005b) and Middle Triassic gabbroic stocks (Zhang et al., 2009a). Clearly, the tectonic affinity of the northern Liaoning area needs to be re-evaluated.

In this paper, we present new zircon U-Pb age, elemental and Sr–Nd–Hf isotopic data for the largest intrusive complex, the Shijianfang (SJF) batholith, from the Faku dome in order to: 1) document the geochemical characteristics of the rocks from this batholith; 2) track their source and evaluate the juvenile nature of the northern Liaoning block, and 3) address the mechanism of granitoid genesis and implications for Phanerozoic continental crustal growth within the interior of the NCC following its amalgamation with the Mongolian Plate.

2. Geological setting

The NCC is bounded on the south by the Paleozoic to Triassic Qinling–Dabie–Sulu orogenic belt (Meng and Zhang, 2000) and on the north by the XMOB (Davis et al., 2001) (Fig. 1a). It is one of the world’s oldest cratons, as evidenced by the presence of ≥3.6 Ga crustal remnants exposed at the surface (Liu et al., 1992) and in lower crustal xenoliths (Zheng et al., 2004). The craton consists of two Archean continental blocks, the Eastern and the Western blocks, separated by the Proterozoic Trans-North China Orogen (Fig. 1a; Zhao et al., 2001). The craton was stabilized during the late Paleoproterozoic and subsequently covered by a thick sequence of Proterozoic to Paleozoic sediments (Zhao et al., 2001). However, it experienced intense tectonothermal reactivation after the Paleozoic, mainly due to the effect of the circum-cratonic orogenic belts (e.g., Davis et al., 2001; Kusky et al., 2007). To the north, the XMOB, which constitutes the eastern segment of the Altait Tectonic Collage (Sengor et al., 1993) or CAOB (Jahn, 2004), is located between the North China and Siberian cratons (Fig. 1b). It is a complex orogenic belt formed through successive accretion of arc complexes, accompanied by emplacement of voluminous subduction zone granitic magmas, mainly during the Paleozoic (Davis et al., 2001; Xiao et al., 2003). During this period, multiple Mongolian arc terranes were assembled to the active margins of the NCC (Davis et al., 2001). The Solonker suture marks the closure of the paleo-Asian ocean and the collision between the NCC and Mongolian composite terranes (Yin and Nie, 1996; Davis et al., 2001; Xiao et al., 2003). With the gradual closure of this ocean basin the NCC and the southern Mongolian terranes were amalgamated and behaved as a combined North China–Mongolian plate (Davis et al., 2001).

As the eastern segment of the CAOB, the XMOB of Northeast China is composed of a collage of micro-continental blocks: the Jiamusi Block in the southeast, the Songliao Block in the middle and the Xing’an and Erguna blocks in the northwest, all separated by major faults (Fig. 1b).

The Songliao block is composed of the Songliao Basin and the Zhangguangcai Range, with the majority of the block occupied by Phanerozoic granites. Local sub-greenstone to amphibolite facies meta-volcanic and meta-sedimentary rocks have been identified and designated as the Hulan Group, with a maximum deposition age of 287 ± 6 Ma as constrained by detrital zircon U–Pb ages (Wu et al., 2007). The Songliao basin developed in the late Mesozoic over a basement of Phanerozoic granites and Paleozoic strata (Wu et al., 2001).

Located to the south of the Songliao block, the northern Liaoning block mainly consists of three tectonic units: the Zhezhong depression in the northwest, the Faku dome in the middle and the Tieling Basin in the east (Fig. 1c). The Faku dome is mainly composed of igneous and sedimentary rocks variably metamorphosed to greenschist facies. They were once regarded as part of the Precambrian basement of the NCC (LBGMR, 1998). However, 1,50,000 scale geological mapping (LBGMR, 1998) revealed that these so-called basement complexes turned out to be deformed Phanerozoic intrusions and meta-volcanic and meta-sedimentary rocks (Fig. 1d). The latter can be correlated with the Upper Paleozoic Hulan Group of the Songliao block in terms of main lithological assemblage, metamorphic and deformational style, and the well-preserved coral fossils in the constituent limestone unit (LBGMR, 1998; Wu et al., 2007). The lower meta-sedimentary formation, striking approximately NNE, is mainly composed of sub-greenstone facies feldspar-quartz schist, mica-quartz schist, slate and dolomitic marble. The upper meta-volcanic formation mainly consists of fine-grained felsic gneiss, biotite/hornblende-plagioclase gneiss and mica schist, with a protolith interpreted as andesite to dacite by the LBGMR (1998). It was intruded by a number of Permian granitic intrusions (Zhang et al., 2005b).

Since Mesozoic times, the northern Liaoning block has become part of the eastern China active tectonic belt, which experienced geodynamic transition from the paleo-Asian to paleo-Pacific tectonic realms (Davis et al., 2001; Kusky et al., 2007). This led to widespread Jurassic–Cretaceous granitic magmatism, and the development of a series of NE- to NNE-trending strike-slip faults and basins in the Cretaceous (Xu et al., 2000).
3. Analytical methods

3.1. Mineral analysis

Major element compositions of minerals were measured on a Cameca SX-51 electron microprobe at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGGCAS), Beijing. The operating conditions were: 15 kv accelerating voltage, 20 nA beam current and a 3 μm spot diameter. A PAP matrix correction program was used in data processing (Pouchou and Pichoir, 1991). Natural and synthetic mineral standards were applied for calibration.

3.2. Zircon U–Pb dating

Zircons were separated from a monzodiorite (FK04-19), a quartz monzonite (FK06-39) and a granite (FK04-24) sample using standard density and magnetic separation techniques and purified by hand-picking under a binocular microscope. Cathodoluminescence (CL) images were obtained prior to analyses, using the Cameca SX-51 microprobe at the IGGCAS to reveal their internal structures.

Zircons from the monzodiorite were dated using the SHRIMP II (sensitive high-resolution ion-microprobe at the Beijing SHRIMP Center in the Institute of Geology, Chinese Academy of Geological
3.3. Major and trace element analysis

For geochemical and isotopic analyses, fresh samples were ground in an agate mill to ~200 mesh. Major elements were analyzed with a Phillips PW 2400 X-Ray fluorescence spectrometer (XRF) at the IGGCAS. The analytical uncertainties are generally within 1–5%. Loss on ignition (LOI) was determined by routine procedures. Trace element abundances were measured by inductively coupled mass spectrometry (ICP-MS) at the IGGCAS using a Finnigan Mat Element spectrometer. Samples were dissolved in distilled HF+HNO₃ in 15 ml high-pressure Teflon bombs at 120 °C for 6 days, dried and then diluted to 50 ml for analysis. A blank solution was also prepared. The total procedural blank was <50 ng for all trace elements. Indium was used as an internal standard to correct for matrix effects and instrument drift. Precision for all trace elements is estimated to be 5% and accuracy is better than 5% for most elements by analyses of the GSR-3 standard (see Zhang et al., 2008a).

3.4. Whole-rock Sr-Nd isotopic analysis

Sr and Nd isotopic compositions were measured on a Finnigan Mat 262 thermal ionization mass spectrometer at the IGGCAS, following the procedure described in Zhang et al. (2008a). Procedural blanks were <100 pg for Sm and Nd and <500 pg for Rb and Sr. ¹⁴⁶Nd/¹⁴⁴Nd was corrected for mass fractionation by normalization to 146Nd/144Nd = 0.7219, and ⁸⁷Sr/⁸⁶Sr ratios were normalized to ⁸⁷Sr/⁸⁶Sr = 0.1194. The measured values for the La Jolla and BCR-2 Nd standards and NBS-987 Sr standard were 143Nd/144Nd = 0.511853 ± 0.000011 (2σ, n = 3) and 0.512613 ± 0.000012 (2σ, n = 3) and ⁸⁷Sr/⁸⁶Sr = 0.710251 ± 0.000011 (2σ, n = 5) during the period of data acquisition.

3.5. In situ Hf isotopic analysis

In situ zircon Hf isotopic analyses were conducted using the Neptune MC-ICP-MS, equipped with a 193-nm laser at the IGGCAS. Spot size of 32 μm was used for analysis, with a laser repetition rate of 10 Hz at 100 mJ. The detailed analytical procedure and correction for interferences follow those described by Wu et al. (2006a). During analyses, the ¹⁷⁶Hf/¹⁷⁷Hf and ¹⁷⁶Lu/¹⁷⁷Hf ratios of the standard zircon (91500) were 0.282270 ± 0.000023 (2σ, n = 15) and 0.000028, similar to the commonly accepted ¹⁷⁶Hf/¹⁷⁷Hf ratio of 0.282284 ± 0.000003 (1σ) measured using the solution method (Woodhead et al., 2004).

4. Field relations and petrography

Field mapping reveals that the SJF batholith intruded into Upper Paleozoic meta-sedimentary and meta-volcanic rocks (LBGMR, 1998). It is composed of two main igneous suites (Fig. 1e); a mafic suite that consists of monzogabbro, monzodiorite and monzonite, and a felsic suite that comprises massive K-feldspar granite, granodiorite (quartz monzonite) and monzogranite. Representative electron microprobe analyses of minerals from various rock types are presented in an online table.

The mafic suite constitutes about 15% of the total outcrop area of the batholith. It mainly occurs as thick massive dikes with fine-grained margins chilled against the country rocks, or as enclaves within the felsic suite (Fig. 1d). Among the main rock types, gabbros are massive, fine- to medium-grained with sub-ophitic textures, and consist mainly of hornblende (30–55 modal %), pyroxene (5 modal %), biotite (0–5 modal %), plagioclase (30–50 modal %), orthoclase (5–10 modal %), rare quartz and accessory magnetite, zircon, apatite, titanite and secondary epidote. Pyroxene is dominantly diopside, with a reconstructed composition Wo₄₀En₅₀Fs₁₀. Hornblende is the most abundant mafic mineral and mainly occurs as subhedral to euhedral phenocrysts. It shows a narrow compositional range of X₅₀ = (Mg/ (Fe+Mg)) = 0.67–0.70 and mainly belongs to magnesio-hastingsite according to the classification scheme of Leake et al. (1997). Plagioclase (An₇₀₋₅₀Ab₂₋₅0Or₁₋₂) generally forms an interlocking framework with interstitial perthitic orthoclase (Or₅₀Ab₅₀). Monzodiorite is fine- to medium-grained, and is composed of 40–65 modal % plagioclase (An₅₀₋₃₅Ab₄₅₋₃₀Or₁₋₂), 10–25 modal % hornblende (magnesio-hastingsite, X₅₀ = 0.55–0.69), 5–10 modal % biotite (X₅₀ = 0.55–0.53), 10–20 modal % orthoclase (Or₅₀₋₃₅Ab₂₋₅0), 0–5 modal % quartz, and minor magnetite, apatite, zircon, titanite and ilmenite. Monzonite typically consists of 45 modal % plagioclase (An₄₁₋₂₇Ab₅₇₋₄₂Or₁₋₂), 40 modal % orthoclase (Or₅₀₋₃₅Ab₂₋₅0), 10 modal % hornblende (magnesio-hastingsite, X₅₀ = 0.46–0.54), with subordinate biotite (X₅₀ = 0.52–0.55) and quartz, and accessory zircon, titanite, magnetite and apatite. Locally, monzodiorite can be observed grading into fine-grained monzogranite.

The felsic suite forms the bulk of the SJF batholith (Fig. 1d) and field observations show that it has clear intrusive relations with the mafic suite (LBGMR, 1998). The felsic rocks are all weakly gneissic and are characterized by subhorizontal foliations associated with a N–S mineral stretching lineations and top-to-the-NNE sense of shear (Hao et al., 1996; Zhang et al., 2005a). Contacts between the various rock types appear to be gradational. Granodiorites and quartz monzodiorites are medium- to coarse-grained with hypidiomorphic inequigranular textures. Major constituents include plagioclase (An₅₋₅₀Ab₂₋₇₅Or₂₋₅), 42–52 modal %), quartz (15–25 modal %), perthitic orthoclase (Qr₅₀₋₃₀Ab₂₋₅0Or₁₋₂, 15–25 modal %), biotite (5–10 modal %) and hornblende (3 modal %), with accessory zircon, apatite, magnetite, allanite, titanite and ilmenite. Both plagioclase and orthoclase commonly occur as phenocrysts. Quartz occurs as both phenocrysts and interstitial grains. Hornblende commonly occurs as subhedral crystals and they are classified as ferrohornblende, magnesio-hastingsite and ferro-edenite, with X₅₀ ranging from 0.49 to 0.56. Biotite occurs both interstitially and as occasional larger ragged subhedral flakes, with X₅₀ ranging from 0.48 to 0.53. With the amount of K-feldspar increasing, granodiorite grade into monzogranite. Typical samples of monzogranite comprise orthoclase (and microcline) (35–40 modal %), plagioclase (30–35 modal %), quartz (25–30 modal %), biotite (3–7 modal %), and accessory zircon, apatite, magnetite and allanite. K-feldspar granite is fine- to medium-grained with a characteristic interlocking texture of alkali-feldspar and anhedral quartz. Major constituents are K-feldspar (50–70 modal %) quartz (25–35 modal %), plagioclase (10–15%) and subordinate biotite (5%). Accessory minerals include zircon, apatite, magnetite, allanite and ilmenite. K-feldspar occurs as both perthitic orthoclase (Qr₅₀₋₃₀Ab₂₋₅0) and microcline (Qr₅₀₋₃₀Ab₂₋₅0An₁) which typically form large subhedral to anhedral grains. Plagioclase (An₇₋₅₀Ab₂₋₅0Or₁₋₂) is chiefly present as coarse albite exsolution within
alkali-feldspar. Quartz mainly forms aggregates of large phenocrysts. Subhedral to anhedral biotite ($X_{\text{Mg}} = 0.15–0.5$) is interstitial to K-feldspar and quartz, and is commonly altered to chlorite; other alteration products are sericite, epidote and hematite. The SJF batholith is intruded by the Middle Triassic Xiaofangshen gabbro stock and the Late Jurassic Hongtuqiangzi pluton. The Xiaofangshen gabbro has a zircon $U$–$Pb$ age of 241 Ma (Zhang et al., 2009a), and is medium- to coarse-grained with intergranular textures. Typical major constituents are plagioclase, hornblende, pyroxene and biotite, with minor amounts of quartz, magnetite, zircon and apatite. The Hongtuqiangzi pluton is mainly composed of biotite monzogranite that exhibits an inequigranular to porphyritic texture, with coarse K-feldspar phenocrysts set in a fine- to medium-grained matrix of plagioclase, K-feldspar, quartz, biotite, minor hornblende, and accessory titanite, magnetite, apatite and zircon. Zircon $U$–$Pb$ dating documented an emplacement age of 159 Ma (Zhang et al., 2005a).

5. Analytical results

5.1. Zircon $U$–$Pb$ data

The results of SHRIMP zircon $U$–$Pb$ analysis of the monzodiorite sample (FK04-19) are listed in Table 1. The zircons are mostly clear, euhedral to subhedral, stubby to elongate prisms (Fig. 2a). They are between 60 and 150 μm in length, with length-to-width ratios ranging from 1.2 to 3. Fifteen analyses were conducted on 15 grains during a single analytical session. Measured $U$ concentrations vary from 235 to 1216 ppm, Th ranges from 160 to 1544 ppm and Th/$U$ ratios range from 0.68 to 1.91. The results define a weighted mean $^{206}\text{Pb} / ^{238}\text{U}$ age of 261 ± 2 Ma with an MSWD of 1.59 (Fig. 2b).

Samples FK04-24 and FK06-39 are from the felsic suite in the central part of the SJF batholith and the results of LA-ICP-MS zircon $U$–$Pb$ analyses are also listed in Table 1. Zircons from both samples are colorless or light brown with no inclusions. They are 70 to 170 μm in size and mostly subhedral–euhedral prismatic crystals. In CL images, they commonly show internal oscillatory zonation (Fig. 2c, e). Twenty one analyses were conducted on 21 zircon grains from the sample FK04-24. Measured $U$ concentrations vary from 155 to 2219 ppm, Th ranges from 126 to 2375 ppm and Th/$U$ ratios range from 0.25 to 1.98. Nineteen analyses form a coherent group and define a weighted mean $^{206}\text{Pb} / ^{238}\text{U}$ age of 249 ± 2 Ma with an MSWD of 1.8 (Fig. 2d); two analyses were omitted because of their discordance. Meanwhile, fourteen analyses were conducted on 14 zircon grains from the sample FK06-39. Measured $U$ concentrations vary from 62 to 1398 ppm, Th ranges from 47 to 195 ppm and Th/$U$ ratios range from 0.30 to 1.17. All analyses form a coherent group and define a weighted mean $^{206}\text{Pb} / ^{238}\text{U}$ age of 248 ± 2 Ma with an MSWD of 0.66 (Fig. 2f).

5.2. Chemical analyses

Major and trace element analyses are presented in Table 2. As shown on the $K_2O + Na_2O$ vs. SiO$_2$ diagram (Fig. 3a), the rocks display a broadly bimodal geochemical distribution, with a compositional gap in SiO$_2$ from 59.9 to 66.7%. The mafic suite falls in the fields of monzogabbro, monzodiorite and monzonte, while the felsic suite mostly lies within the fields of quartz monzonite/granodiorite and granite. The rocks from the mafic suite range from 46.6 to 59.9% SiO$_2$. The aluminium saturation index $[\text{ASI} = \text{molar Al}_2\text{O}_3 / (\text{CaO} + K_2\text{O} + Na_2O)]$ increases with SiO$_2$ content from 0.67 to 1.04 (Fig. 3b). Applying the classification of Frost et al. (2001b), the mafic rocks mainly belong to the magnesian series (Fig. 3c). On the alkali-lime index (MALI = Na$_2$O + KO – CaO by wt) plot (Fig. 3d), they span the calcic to alkaline fields, with the majority being alkaline-calcic. In most major element Harker plots (Fig. 4), most samples from the suite define a continuous evolution trend: CaO, Al$_2$O$_3$, MgO, TiO$_2$, TiO$_2$, MnO, and P$_2$O$_5$ are negatively correlated with SiO$_2$, while K$_2$O and Na$_2$O are positively correlated, though the latter is more variable. Likewise, compatible elements such as Sc, Co and V decrease with increasing SiO$_2$ (Fig. 5), a feature also shared by Sr and Y. Rb and Pb are positively correlated with SiO$_2$ (Fig. 5). Abundances of some high field strength element (HFSEs) such as Zr, Hf, Nb, Ta and rare earth element like La and Nd remain almost constant with increasing SiO$_2$ (Fig. 5). It is notable that some samples deviate from the evolution lines in some plots due to their lower Al$_2$O$_3$ and higher MgO contents, consistent with the relatively higher amount of pyroxene in these samples. Sample FK06-45 is distinguished by its higher Al$_2$O$_3$, K$_2$O and Na$_2$O. This, together with its strong positive Eu anomaly (Eu/Eu* = 2.46), is compatible with its dominant mineral constituents of two kinds of feldspars.

Monzogabbro and monzodiorite samples display moderate light REE enrichment (La$_N$/Yb$_N$ = 6.9 to 13.2) and small negative or positive Eu anomalies (Eu/Eu* = 0.85–1.12 Fig. 6a). On a primitive mantle-normalized spidergram (Fig. 6b), they are enriched in large ion lithophile elements (LILE), with positive Ba, Sr, Th and U anomalies, and are depleted in HFSEs with negative Nb, Ta, Zr and Hf anomalies. The monzonite samples show moderate to high LREE enrichment (La$_N$/Yb$_N$ = 6.9 to 24.1), negligible negative to moderate positive Eu anomalies (Eu/Eu* = 0.92–2.46) (Fig. 5a), moderate enrichment of LILE (e.g., Th, U, LREE), and depletion of HFSEs (Fig. 6b). The felsic suite varies from 66.7 to 77.2% SiO$_2$ and spans the metaluminous–peraluminous boundary with A/CKN of 0.88 to 1.12 (Fig. 3b). Using the classification of Frost et al. (2001b), low-silica members are predominantly magnesian, while most high-silica members are ferroan (Fig. 3c). On the alkali-lime index plot (Fig. 3d), both groups span the calcic to alkaline fields. As with the mafic suite, the rocks from the felsic suite also define an evolution trend in most major element Harker plots (Fig. 4): CaO, Al$_2$O$_3$, MgO, TiO$_2$, Fe$_2$O$_3$, TiO$_2$, ZrO$_2$, MnO, and P$_2$O$_5$ are negatively correlated with SiO$_2$, while K$_2$O is weakly positively correlated.

Most low-silica (SiO$_2$ <70%) members, except sample FK06-44, and some high-silica (SiO$_2$ >70%) members are characterized by relatively unfractonated REE patterns with small negative Eu anomalies (Eu/Eu* = 0.71–0.96) (Fig. 6c, e), high Ba, Sr, Zr, Hf, and weak negative anomalies for Nb and Ta (Fig. 6d). Sample FK06-44 is distinctive with its huge negative Eu anomaly (Eu/Eu* = 0.06) (Fig. 6c), strong depletion in Ba, Sr, Eu and Ti, and enrichment in HFSEs (such as LREE, Zr, Hf, Nb and Ta) (Fig. 6d). This is typical of the geochemistry of the highly evolved I-type granites with tetrad REE patterns and non-charge-and-radius-controlled (non-CHARAC, Bau, 1996) trace element behavior from NE China (e.g., Jahn et al., 2001, 2004). Such features are also shared by most high-silica members, with ‘seagull’ REE patterns (Fig. 6e), huge negative Eu anomalies (Eu/Eu* = 0.11–0.24), pronounced depletion of Ba, Sr, and Ti, and positive anomalies of Th, U, Pb, Nb, Ta, Zr, Hf, Y and HREE (Fig. 6f).

5.3. Whole-rock Nd–Sr isotopic compositions

Whole-rock Sr–Nd isotope analytical data, together with the calculated initial Sr–Nd isotopic compositions, are presented in Table 3. As shown in a plot of $\varepsilon_{\text{Nd}}$ vs. $t$ (Fig. 7), the monzogabbro to monzonte samples show slightly depleted isotopic compositions, with rather restricted $I_\text{Nd}$ ratios ranging from 0.7049 to 0.7054, $\varepsilon_{\text{Nd}}(t)$ from +2.72 to −1.82 and model ages ($t_{\text{DM}}$) of 807 to 1150 Ma. The felsic rocks display indistinguishable Nd compositions from the mafic ones, with $\varepsilon_{\text{Nd}}(t)$ values varying from +2.28 to −0.55 and model ages ($t_{\text{DM}}$) of 834 to 1064 Ma (Table 3, Fig. 7). However, their Sr isotopic compositions are more scattered, with some high-silica samples having very high $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios, while the relatively unfractonated felsic samples yield a range of $I_{\text{Sr}}$ ratios from 0.7043 to 0.7057.
between 650 and 769 Ma (Fig. 8). The Hf model ages (from 0.28272 to 0.28282 and 0.28279, $\varepsilon_{\text{DM2}}$ between 644 and 758 Ma. Zircons from the quartz monzonite sample show a range of initial $^{176}$Hf/$^{177}$Hf ratios from 0.28271 to 0.28281 and $\varepsilon_{\text{DM2}}$ of these zircons mainly range from 3.4 to 6.1 (Table 4).

### Table 4

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<th>Pb (ppm)</th>
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5.4. Zircon Hf isotopes

As presented in Table 4, zircons from the mafic sample (FK04-19) show a range of initial $^{176}$Hf/$^{177}$Hf ratios from 0.28271 to 0.28281 and $\Delta_t$ (values from 3.4 to 7.1 (Fig. 8). The Hf model ages ($T_{\text{DM2}}$) of these zircons mainly range between 618 and 758 Ma. Zircons from the granite sample (FK04-24) show a range of initial $^{176}$Hf/$^{177}$Hf ratios from 0.28272 to 0.28282 and $\Delta_t$ (values from 3.6 to 6.1 (Table 4, Fig. 8). The Hf model ages ($T_{\text{DM2}}$) for these zircons mainly range between 644 and 758 Ma. Zircons from the quartz monzonite sample (FK06-39) exhibit a range of initial $^{176}$Hf/$^{177}$Hf ratios from 0.28271 to 0.28279, $\Delta_t$ (values from 3.4 to 6.2 and the Hf model ages ($T_{\text{DM2}}$) between 650 and 769 Ma (Fig. 8).

### 6. Discussion

6.1. Petrogenesis

6.1.1. Origin of the mafic rocks

The U-Pb age of 261 ± 2 Ma for magmatic zircons from the monzodiorite supports the field evidence that the mafic suite intruded into the Upper Paleozoic meta-sedimentary formations, and indicates that the mafic suite of the SJF batholith was emplaced in the Late Permaid. Petrographically, the hornblende-rich character of the SJF mafic rocks is reminiscent of that of the high-level, hornblende-rich mafic intrusions of the Mesozoic Sierra Nevada batholith (Sisson et al., 1996) and mafic
plutonic rocks from the Variscan Pyrenees calc-alkaline complex (Roberts et al., 2000). However, their occurrence as thick, massive dikes and sub-ophitic texture argue against a cumulate origin. This is further substantiated by the MgO, Cr and Ni contents of the gabbros, which are far below those expected for the rocks with a cumulate origin (MgO > 15%, Cr > 2000 ppm, Ni > 300 ppm) (e.g., Roberts et al., 2000).

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Fig. 2. (a) [c] and (e) Cathodoluminescence (CL) images of the dated zircons, and (b) [d] and (f) zircon U-Pb concordia diagrams for mafic sample Fk04-19 and felsic samples Fk04-24 and Fk06-39, respectively, from the Shijianfang batholith.
The SJF mafic rocks have high Fe₂O₃ (up to 12.6 wt.%), MgO (up to 9.0 wt.%), Cr (246–517 ppm), and V (144–262 ppm) contents at low-silica values (SiO₂ = 46.6–54.1%), distinct from those of any crustal materials (Rudnick and Gao, 2003) or crustally-derived melts (e.g., Patiño Douce and Beard, 1995; Patiño Douce, 1997). This suggests that their parental magma was derived by partial melting of a mantle source.

Furthermore, these mafic rocks are characterized by selective enrichment of LILs and light REE, and depletion in HFSEs (e.g., Nb, Ta and Ti), thus leading to high La/Nb (2.5–3.6), Ba/Nb (38–95), and Zr/Nb ratios (10.8–20.6) and low Ce/Pb (4.6–13.5). These values bear close resemblance to those of arc volcanic rocks worldwide (e.g., Stern, 2002).

In general, such geochemical features can be attributed either to partial melting of an enriched mantle source that was previously spiked in LILE and LREE by slab-derived hydrous fluids or melts prior to magma generation, or to extensive crustal contamination of MORB-like magmas during magma ascent and emplacement.

Crustal contamination would result in positively correlated εNd(t) and Sr. However, these features are not observed in the SJF mafic rocks (Fig. 9), thus suggesting that crustal contamination did not play a significant role in their formation.
Since both fractional crystallization and crustal assimilation can reduce the Mg#, this potentially allows parental melts to be identified using the relative Mg# (e.g., McMillan et al., 2003). Following this argument, samples FK04-18, FK06-24 and FK06-26 collected from the presence of primary hornblende and biotite is consistent with a wet, using the relative Mg# (e.g., McMillan et al., 2003). Following this argument, samples FK04-18, FK06-24 and FK06-26 collected from the presence of primary hornblende and biotite is consistent with a wet,
which was tectonically linked at this time (Fig. 7, Zhang et al., 2008a), the asthenospheric melts (see the following discussion) have a much more depleted isotopic composition similar to MORB.

This is not the case for the SJF mafic rocks. Instead, their evolved isotopic compositions are consistent with an enriched sub-continental lithospheric mantle. Specifically, the elevated Ba, K, and Sr abundances and high Ba/La ratios (Fig. 10a) provide evidence for a fluid component from a subducted slab (McCulloch and Gamble, 1991), while the high Th/Yb ratios (Fig. 10a) indicate a potential hydrous melt component from subducted sediments (Pearce et al., 1999; Woodhead et al., 2001).

The lack of HREE and Y depletion reflects a garnet-absent source, and the relatively low (La/Yb)\text{CN} (6.9–12.4) and (Tb/Yb)\text{CN} (1.5–1.9) further suggest relatively high degree melting of a mantle source in the spinel stability field at depths shallower than 70 km (Watson and McKenzie, 1991).

The Sr–Nd–Hf isotopic data can provide further information on the nature of the mantle source region for the SJF mafic rocks. As presented above, these rocks exhibit slightly negative to weakly positive $\varepsilon$\text{Nd}(t) values, Neoproterozoic Nd model ages and moderate $I$\text{Sr} ratios (0.7049–0.7055). These features are similar to those found in
numerous late Paleozoic to early Mesozoic mantle-derived igneous rocks from the XMOB (Fig. 7. Chen et al., 2000; Wu et al., 2000, 2002, 2004) and are consistent with the juvenile nature of the lithospheric mantle beneath the XMOB, as also constrained by mantle xenolith evidence (Zhang et al., 2000; Wu et al., 2003b), but distinct from those of Archean to Paleoproterozoic sub-continental lithospheric mantle beneath the northern margin of the eastern NCC during the Late Paleozoic to Early Mesozoic (Wu et al., 2006b; Zhang et al., 2008c). On the other hand, as shown on the \( \varepsilon_{Hf}(t) \) vs. emplacement age plot (Fig. 8), all zircon data points fall in the field of the XMOB as defined by igneous zircons extracted from Phanerozoic igneous rocks in that region (Yang et al., 2006). Therefore, we suggest that the parental magma for the SJF mafic rocks originated from typical juvenile lithospheric mantle beneath the XMOB.

Although the exact age of this juvenile lithosphere, i.e., the timing of the mantle enrichment in the source of the SJF mafic rocks, is unclear, it is probably of Early to Middle Paleozoic times for the following reasons. First, Os isotope data suggest a young lithospheric mantle (0.4–0.70 Ga) beneath NE China (Wu et al., 2003b). Second, the Early Permian asthenospheric melts from the neighboring region have young Nd model ages (\( T_{DM} = 0.39–0.56 \text{ Ga} \)) (Zhang et al., 2008a) that reconcile with the Middle–Late Paleozoic ophiolites in the region (Miao et al., 2007; Jian et al., 2008).Third, as with the mafic–ultramafic complexes displaying Neoproterozoic Nd model ages (mainly 0.79–1.25 Ga) from NE China (e.g., Wu et al., 2004), the Neoproterozoic Nd model ages for the SJF mafic rocks (0.81–1.18 Ga) may result from the mixing of a significant amount of depleted mantle or juvenile component and limited amount of old Precambrian crustal component.

As shown above, the intra-suite rocks of the SJF mafic suite show systematic variations in their geochemical characteristics. With increasing \( \text{SiO}_2 \), ASI, K2O, Na2O, Rb, Pb, Th and U increase and CaO, Al2O3, MgO, Fe2O3, TiO2, MnO, P2O5, Sc, Co, V, Sr and Y decrease (Figs. 4 and 5). These variations could be explained by fractionation of phases such as olivine, pyroxene, amphibole, plagioclase, apatite and titanite. This can also be reflected by the Cr–Ni fractionation vector plot (Fig. 10b).

6.1.2. Petrogenesis of the felsic suite

6.1.2.1. Classification and magmatic affinity. The U–Pb ages of 248 ± 2 Ma and 249 ± 2 Ma for the magmatic zircons from the quartz monzonite and granite samples together with its intrusive contact with the mafic suite, suggests that the felsic suite of the SJF batholith was emplaced in the earliest Triassic.

Among approximately 20 different granite classification schemes that have evolved over the past decades, one widely used classification scheme is the Alphabetic classification, which subdivides granites into I-, S-, M- and A-type according to their protolith (Pitcher, 1993). However, distinction between different types is not always straightforward. The reason lies in the fact that similar granitic compositions can be produced by partial melting of a variety of sources or can be achieved by a number of different processes (e.g., Frost et al., 2001b).

This is especially so for the distinction between A-type granites and highly fractionated I-type ones. As exemplified by the Lachlan Fold Belt (Whalen et al., 1987; King et al., 1997), NE China (Jahn et al., 2001; Wu et al., 2002, 2003a) and the NCC (Zhang et al., 2008b; Jiang et al., 2009), high Ga/Al ratios and Zr + Nb + Y + Ce values, two of the
most diagnostic features of A-type granites, can also be achieved in highly fractionated I-type and S-type granites. Moreover, a higher FeOt/MgO ratio has usually been adopted to discriminate A-type granites from I-types. In reality, the difference in FeOt/MgO between A-type and I-type granites is only manifest at SiO2 < 70%, and when SiO2 > 70% the FeOt/MgO of the two types overlap significantly (Frost et al., 2001b). It is clear that various chemical discrimination diagrams cannot effectively distinguish fractionated A-type granites from highly-fractionated I-type granites (Jiang et al., 2009). Therefore, classification of highly felsic rocks can only be made with confidence when there is an association with less evolved rocks within the same suite (King et al., 1997).

In the case of the SJF felsic rocks, they straddle the fields of A-, and M-, I-, S-types in the K2O + Na2O, FeOt/MgO and Nb vs. Ga/Al diagrams of Whalen et al. (1987) (Fig. 11a–c). In the FeOt/MgO and (K2O + Na2O)/CaO vs. (Zr + Nb + Ce + Y) diagrams of Whalen et al. (1987) (Fig. 11d, e), most low-silica members plot on the unfractonated granite field, while high-silica members plot on both the fractionated felsic I-type and A-type granite fields.

Recognizing the drawbacks of previous schemes, Frost et al. (2001b) proposed a non-genetic, non-tectonic geochemical classification scheme that attempted to incorporate the best qualities of the previous schemes. As shown above, the SJF felsic rocks span the entire spectrum, from calc-alkalic to alkaline on the SiO2 vs. MALI diagram from this classification scheme (Fig. 3d). These compositional trends bear a close resemblance to those of Caledonian-type post-collisional granites of Great Britain (Fig. 3c, d), and in particular to the post-orogenic calc-alkaline magmas of the Scottish Highlands as represented by the 425 Ma Arrochar and Garabal Hill-Fyne complexes (Clemens et al., 2009). As pointed out by Clemens et al. (2009), such crossing-field distribution trends are commonly manifest in post-collisional magmatic suites, and may reflect petrogenetic scenarios of magma mixing or partial melting of mixed protoliths.

This is also consistent with their distribution in the Rb vs. Y + Nb diagram of Pearce et al. (1984). As shown in Fig. 11f, the SJF felsic rocks plot in both the VAG (volcanic arc granite) and WPG (within-plate granite) fields and also mostly fall in the post-collisional field of Pearce (1996). Such ambiguities are typical for post-collisional granitoid suites in general (Pearce, 1996), and suggest a transition from calc-alkaline to alkaline magmatic series in orogenic to post-orogenic tectonic settings (Barbarin, 1999). Moreover, the negative anomalies in Nb and Ta in most low-silica members are more typical of subduction-related and post-collisional I-type granites than A-type, anorogenic types.

Petrographically, the SJF felsic rocks contain hornblende, titanite, allanite and biotite, but lack primary muscovite, garnet and monazite. Such mineral signatures are typical of I-type granitoids, but not of S-types. Therefore, we suggest that the SJF felsic rocks can be assigned to the post-collisional calc-alkaline to alkaline I-type association. Such a tectonic affiliation is also compatible with the geodynamic setting of the SJF batholith, as discussed in the following section.
(2) partial melting of pre-existing igneous rocks at intracrustal levels with subsequent fractional crystallization and/or restite crystal fractionation (e.g., Chappell, 1996).

Looking at the SJF batholith overall, although coexisting mafic and felsic suites may point to contrasting mantle and crustal sources, magma mixing and AFC process during ascent and emplacement appear to have been insignificant for the reasons stated below. For instance, there is considerable overlap of the initial isotope ratios between felsic and mafic suites (Fig. 9) and a distinctly chaotic nature in the distribution of the values among the intra-suite rocks are inconsistent with what might be expected in the case of magma mixing and crustal assimilation. On the other hand, undercooled magmatic inclusions (e.g., Poli and Tommasini, 1991) and xenocrystic inherited zircons, typical indicators for magma mixing and crustal assimilation, are rarely observed within the SJF felsic suite.

The SJF felsic rocks could have been produced by extreme differentiation of contemporaneous mantle-derived parental basaltic melt, even though such volumes of mafic rock are not presently exposed. However, there are several arguments against this possibility. First, the geochemical characteristics of most low-silica and some

![Fig. 5. Plots of various trace elements vs. SiO$_2$ for the mafic and felsic suites from the Shijianfang batholith. Symbols as for Fig. 3.](image-url)
high-silica members from the felsic suite, such as the relatively unfractuated REE patterns, depletion in HFSEs and no significant negative Eu anomaly, seem to preclude an origin by simple fractionation of basaltic melt. Secondly, the SJF batholith lacks a continuum in rock types and composition that is generally expected for a magma differentiation model (Frost et al., 1999) as in the cases of the alkaline magmatic suite of the Amram Massif (Mushkin et al., 2003) and the Karamay suite from the West Junggar fold belt (Chen and Arakawa, 2005). Third, recent experimental work has shown that extreme fractionation (96–97 vol.%) would be required for a mafic parental magma to yield residual liquids of potassic, low-silica rhyolitic composition (≤68% wt.% SiO2), with the most voluminous products being intermediate in composition (Sisson et al., 2005; Whitaker et al., 2008). The relative scarcity of intermediate rocks in the study area is obviously at odds with a simple magma differentiation model, after all, intermediate liquids would have lower viscosity than granitic liquids that intruded to the present level (Clemens et al., 2009). Lastly, the crossing-field distribution trends of the SJF felsic rocks on the SiO2–K2O (Fig. 4) and SiO2–MALI (Fig. 3d) diagrams do not reflect a differentiation process, but imply that there existed mixed magma sources or that magma mixing processes were involved (Roberts and Clemens, 1993; Clemens et al., 2009).

Since our preceding reasoning does not favor magma mixing, it seems most likely that the SJF felsic magmas were formed directly by
The compositions of the source play a critical role in determining the major element chemistry of the melt (Patiño Douce and McCarthy, 1997) and partial melting products by dehydration melting experiments are characterized by distinct chemical signatures, so partial melting from compositionally different protoliths may be distinguished using appropriate major element plots. As shown in Fig. 12a–d, the less-differentiated low-silica members from the SJF felsic suite seem to be broadly compatible with an origin by dehydration melting from meta-basaltic and -andesitic sources.

Given the continuation of basaltic underplating from Late Permian to Middle Triassic, as witnessed by the SJF mafic suite and the Xiaofangshen gabbro stock (Zhang et al., 2009a), a favorable scenario has been sustained for dehydration melting of the sources involving previously-underplated material within a post-collisional extensional environment. This is similar to the Mesoproterozoic Sherman granites from the New England Fold belt of Eastern Australia (Landenberger and Collins, 1996), Late Jurassic to Early Cretaceous granitoids of Inner Mongolia of North China (Liu et al., 2005) and the Late Permian granodiorites in the Songliao block of NE China (Liu et al., 2010). In this scenario, mantle-derived magmatic underplating during post-collisional thermal relaxation or rifting provides both continued heat supply and newly mantle-derived material to the base of the lower crust. They differentiate to form gabbros and monzodiorites and thus constitute juvenile lower crust. The heat supply from further underplating due to rising asthenosphere during continued rifting is capable of elevating lower crustal temperatures to above 850–900 °C (Clemens et al., 1986) or 900–950 °C (Huppert and Sparks, 1988). This leads to partial melting of a mixed protolith, which contain the newly underplated basaltic lower crust and old lower crustal materials, to yield parental magmas for the felsic suite with similar whole-rock Sr–Nd and zircon Hf isotopic compositions. This is supported not only by the zircon saturation temperature (Watson and Harrison, 1983) range from 939 to 821 °C obtained for the low-silica members from the SJF felsic suite, but also by the recent experimental study on underplated arc basaltic rocks which demonstrates that geochemically important volumes of common granitic liquids (SiO$_2$ > 66%) can form near the solidus of diverse medium to high-K basaltic compositions at mid-to-lower crustal pressures (Sisson et al., 2005).

![Fig. 7. εNd(t) vs. Nd(t) for the Shijianfang batholith. The field of Paleozoic kimberlites and mantle xenoliths from the eastern North China Craton are from Wu et al. (2006b). DM depleted mantle is from Zindler and Hart (1986).](image)
As in the case of the Phanerozoic granitoids in NE China (Wu et al., 2002, 2003a; Jahn, 2004; Liu et al., 2010), we can use a simple two-component mixing model to estimate the proportions of juvenile and ancient crustal components involved in the genesis of the SJF felsic rocks. In the model, the newly underplated mafic crust and the ancient lower crust constitute two major components. Modeling results show that the juvenile crustal component that resulted from mantle-derived magmatic underplating has played an essential role in the genesis of the SJF felsic rocks (Fig. 13a). As cautioned by Wu et al. (2003a) and Jahn (2004), this by no means indicates that the felsic rocks were formed by mixing basaltic and lower crustal melts in such proportions. Rather, it suggests that the felsic magmas were produced by melting of a mixed lithology containing lower crustal material that was intruded or underplated by a basaltic magma in such a proportion.

Most high-silica members of the SJF felsic suite show huge negative Eu anomalies in their REE patterns (Fig. 6f). With increasing SiO2, the abundances of Ti, Al, Fe, Mg, Ca, P, Ba, Sr, Y, Zr, Hf and Sc decrease, whereas those of K, Rb, Pb, Th and U increase (Figs. 4 and 5). These systematic variations indicate that advanced fractional crystallization, mostly probably of plagioclase, K-feldspar, hornblende, biotite, zircon, ilmenite and apatite, has taken place during the formation of these granites. This is consistent with slightly lower zircon saturation temperature of 865–732 °C obtained for this suite.

As shown above, one low-silica and some high-silica members show tetrad REE patterns and non-CHARAC trace element behavior. As briefly reviewed by Jahn et al. (2001), the tetrad REE effect was first recognized by two chemists (Peppard et al., 1969) and successively documented in natural samples from marine and terrestrial geochemical systems (Masuda and Ikeuchi, 1979; Kawabe, 1995; Bau, 1996). In recent years, this geochemical behavior has been widely observed in highly evolved magmatic systems with strong hydrothermal interaction and it is considered to form through intense interaction of the residual melts with fluorine-bearing hydrothermal solutions.

### Table 4

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</tbody>
</table>

As in the case of the Phanerozoic granitoids in NE China (Wu et al., 2002, 2003a; Jahn, 2004; Liu et al., 2010), we can use a simple two-component mixing model to estimate the proportions of juvenile and ancient crustal components involved in the genesis of the SJF felsic rocks. In the model, the newly underplated mafic crust and the ancient lower crust constitute two major components. Modeling results show that the juvenile crustal component that resulted from mantle-derived magmatic underplating has played an essential role in the genesis of the SJF felsic rocks (Fig. 13a). As cautioned by Wu et al. (2003a) and Jahn (2004), this by no means indicates that the felsic rocks were formed by mixing basaltic and lower crustal melts in such proportions. Rather, it suggests that the felsic magmas were produced by melting of a mixed lithology containing lower crustal material that was intruded or underplated by a basaltic magma in such a proportion.

Most high-silica members of the SJF felsic suite show huge negative Eu anomalies in their REE patterns (Fig. 6f). With increasing SiO2, the abundances of Ti, Al, Fe, Mg, Ca, P, Ba, Sr, Y, Zr, Hf and Sc decrease, whereas those of K, Rb, Pb, Th and U increase (Figs. 4 and 5). These systematic variations indicate that advanced fractional crystallization, mostly probably of plagioclase, K-feldspar, hornblende, biotite, zircon, ilmenite and apatite, has taken place during the formation of these granites. This is consistent with slightly lower zircon saturation temperature of 865–732 °C obtained for this suite.
fluids (Bau, 1996; Irber, 1999; Jahn et al., 2001, 2004; Wu et al., 2004; Liu and Zhang, 2005; Zhang et al., 2008b). This suggests that late-stage fluid-melt interaction could have played a role in the petrogenesis of the SJF felsic suite. It may accentuate the above elemental enrichment or depletion tendency and lead to the K$_2$O scatter (Poitrasson and Pin, 1998), unusually high concentrations of Fe and Nb in some samples, as in the case of the highly evolved Baerzhe granites from NE China (Jahn et al., 2001).

6.2. Tectonic affinity and geodynamic setting

The tectonic affinity of the Faku dome has been a controversial but important issue, given its critical locality between a Phanerozoic accretionary orogen and a Precambrian craton. Previous geochronological studies have recognized a few Permian granitic intrusions and Middle Triassic mafic stocks from what was once regarded as Precambrian crystalline basement (Zhang et al., 2005a, b, 2009a). These late Paleozoic to early Mesozoic ages, coupled with their primitive geochemical signatures, lead us to suggest that no large-scale Precambrian crystalline basement existed in the Faku dome. This echoes similar views for the nature of the basement in regions such as the Xing’an block (Miao et al., 2003), the Songliao Basin (Wu et al., 2001; Pei et al., 2007) and the Jamusi Block (Wilde et al., 2003). This is also consistent with the contrast in lithospheric structure revealed by systematic geological–geophysical sections: Western Liaoning is characterized by a thick lithosphere with high rigidity and strength, whereas Songliao and northern Liaoning are characterized by relatively thin lithosphere with low rigidity and strength (Xu et al., 2000).
The occurrence of the Late Permian to Early Triassic SJF mafic to felsic intrusions, with their Nd model ages falling within the "juvenile" range as defined by the Phanerozoic igneous rocks from the CAOB (Fig. 13a) and their ultimate derivation from a juvenile subcontinental lithospheric mantle, provides further evidence indicating that the extensive late Paleozoic to early Mesozoic magmatism in the Faku region represents significant addition of juvenile material from the lithospheric mantle to the crust. Such crust/mantle interaction implies that the northern Liaoning block has a tectonic affinity with the Phanerozoic accretionary orogenic belt. This observation further leads to the possibility that the surface suture between the NCC and the XMOB in northern Liaoning might be located along the Chifeng-Kaiyuan Fault (F2 on Fig. 1c), not the Xilamulunhe Fault (F1 on Fig. 1c) as previously advocated (LBGMR, 1989) and still maintained in the most recent geological atlas of China (Ma, 2002).

The Solonker zone has been regarded as the site of final closure of the paleo-Asian ocean (e.g., Tang, 1990; Sengör et al., 1993; Xiao et al., 2003), but there is much controversy concerning the timing of the suturing. Some authors propose that it took place during the Permian to Early Triassic (Sengör et al., 1993; Chen et al., 2000; Xiao et al., 2003) as previously advocated (LBGMR, 1989) and still maintained in the most recent geological atlas of China (Ma, 2002).

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However, our recent documentation of the Early Permian post-collisional bimodal volcanism in Central Inner Mongolia suggests that the NCC and Mongolian micro-continents had amalgamated by the earliest Permian, forming the combined Mesozoic North China–Mongolian Plate (Zhang et al., 2008a). The widespread occurrence of the Late Permian–Middle Triassic post-orogenic intrusive suites along the western segment of the northern margin of the NCC supports this proposition (Zhang et al., 2009b), as does the recent characterization of Late Permian (262 Ma) post-orogenic granitic and associated mafic rocks from northeast China (Liu et al., 2010).

As reviewed by various authors (e.g., Liégeois, 1998; Vanderhaeghe and Teyssier, 2001; Bonin, 2004), an orogenic cycle typically features a pre-collisional period characterized by subduction leading to oceanic basin closure and terrane docking, a period of arc-continent or continent–continent collision accommodated by crustal thickening and post-collisional to post-orogenic periods. Marotta et al. (1998) identified four stages of mantle ‘unrooting’ during this process: orogenic growth, initiation of gravitational instability until lithospheric failure, sinking of the detached lithosphere and relaxation of the system.

When evaluated within this general context of thermal and mechanical evolution of the continental crust during orogenesis, the northern margin of the newly amalgamated North China–Mongolian Plate was dominated by post-orogenic extensional regimes from the Late Permian to Early Triassic, possibly corresponding to the third stage, i.e., lithosphere delamination, in terms of mantle evolutional process of Marotta et al. (1998). During this post-orogenic stage, repetitive generation of water-bearing magmas resulted in an increasingly depleted and dehydrated continental lithosphere (Bonin, 2004). This led to the thermal and mechanical instability of the thickened lithospheric keel and, coupled with the weak link with the crust (Meissner and Mooney, 1998), induced delamination of the sub-continental lithospheric mantle and subsequent crustal extension (Marotta et al., 1998). This extensional tectonic regime enables rapid upwelling of asthenosphere, and triggers concomitant decompressional melting of the mantle and magmatic underplating at the crust–mantle boundary.

Alternatively, given the coincidence of the SJF batholith in timing with Permian superplume events like the 251 Ma Siberian Traps of Russia (Campbell et al., 1992) and the 260 Ma Emeishan Large Igneous Province in Southwestern China (Xu et al., 2001), it is
Dehydration of the thinning lithosphere results ultimately in a shift towards the continental interior of the newly amalgamated North China–Mongolian Plate.

6.3. Implication for Phanerozoic continental growth

Phanerozoic granitoids and high-silica rhyolites with juvenile isotopic signatures occur worldwide in areas such as the New England and Lachlan Fold Belts in southeastern Australia (Turner et al., 1992; Keay et al., 1997; King et al., 1997), the Cordillera of North America (Samson et al., 1989; Pickett and Saleeby, 1999), the Newfoundland Appalachians in Canada (Whalen et al., 1996), the Niger–Nigerian province (Kinnaird and Bowden, 1987) and the Arabian–Nubian shield (Mushkin et al., 2003) and, have been regarded as the most eloquent witness of continental crustal growth during the Phanerozoic. This is especially the case in the CAOB, which is the most important site of Phanerozoic continental growth (e.g., Jahn, 2004) as evidenced by the voluminous emplacement of such felsic magmas in the hinterland of the orogen, including the Russian Transbaikalia (Wickham et al., 1996), Kazakhstan (Heinhorst et al., 2000), northern Xinjiang of NW China (Han et al., 1997; Chen and Arakawa, 2005) and NE China (Wu et al., 2000; Jahn et al., 2001; Wu et al., 2002, 2003a).

Given the tectonic affinity of the northern Liaoning block with the Xing’an and Songliao blocks from NE China, we adopted the equation of DePaolo et al. (1991), that was previously applied to the Phanerozoic granitoids in NE China (Wu et al., 2003a; Jahn, 2004), to estimate the proportions of mantle components in the SJF felsic rocks. Modeling results show that the mantle component represents about 71–82% (Fig. 13b). This high proportion is comparable to ca. 80–90% of juvenile crust in the granites from the Xing’an block and ca. 60–90% for those from the Songliao block (Wu et al., 2003a).

In contrast to lateral continental growth through accretion of arc complexes and subduction zone magmatism prior to final collision of major crustal blocks (Sengör et al., 1993), this study, together with our recent documentation of Early Permian bimodal volcanism along the Solonker suture zone (Zhang et al., 2008a), attests to the increasingly recognized importance of post-collisional vertical accretion in Phanerozoic continental crustal growth (Jahn, 2004). Specifically, these two case studies exemplify a crustal continuum accretion model that involves basaltic underplating, subsequent fractionation of mantle-derived magmas and partial melting of mixed lithologies of juvenile lower crust (basaltic underplates and their differentiates) and ancient lower crust.

This model differs from the differentiation model of Chen and Arakawa (2005) with reference to the late Carboniferous granitoids from West Junggar foldbelt (NW China). The latter visualizes that high-silicic magmas evolved from contemporaneous basaltic magmas by protracted fractional crystallization during post-collisional extension and thus represent production of juvenile continental crust in the Phanerozoic. This indicates that both mechanisms can convert a basic crust into a felsic one and thus result in voluminous Phanerozoic continental growth in the CAOB.

These case examples from the CAOB, together with similar ones from Australia (e.g., Turner et al., 1992) and North America (e.g., Frost and Frost, 1997), all point to the importance of a favorable post-collisional to post-orogenic extensional regime for worldwide continental crustal growth during the Phanerozoic. Such consanguinity is not an accident, but hinges on the synchronicity between evolved thermal and mechanical phases under post-collisional to post-orogenic settings and essential multiple-step processes for continental crustal growth. In the case of the XMOB, involving a much smaller scale of collision than, for instance, in Tibet (Zhang et al., 2008a), the first episode of magma generation occurred when post-collisional extensional collapse caused the removal of parts of the thickened lithosphere. Sequential asthenosphere upwelling perturbed the original thermal gradient (Bonin, 2004), resulting in asthenospheric melts, as represented by the ca. 280 Ma Xilinhot tholeiitic lavas.
Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.lithos.2010.03.005.

References


