The Middle Triassic Meiwu Batholith, West Qinling, Central China: Implications for the Evolution of Compositional Diversity in a Composite Batholith

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ABSTRACT

An integrated study involving whole-rock and Sr–Nd–Hf isotope geochemistry and zircon geochronology and trace element combined with detailed field investigation was carried out for the composite Meiwu batholith in the West Qinling orogenic belt of central China to probe the origins of its compositional diversity and its emplacement history. The batholith is composed of quartz diorite, granodiorite and biotite granite, with abundant mafic magmatic enclaves and minor tonalitic enclaves in the granodiorite. The crystallization age of the batholith is ~240–245 Ma. Geochemical and Sr–Nd–Hf isotopic data indicate that the magmas that formed the quartz diorite and the mafic enclaves were derived by partial melting of enriched lithosphere mantle, followed by variable degrees of hybridization with crustal magmas in deep crustal hot zones. These initially heterogeneous, hybrid magmas successively intruded into the upper crust and coalesced into a large magma chamber. Zircon trace element and Hf isotopic compositions suggest that the outer fine-grained part of the quartz diorite pluton crystallized from a less differentiated magma as a result of rapid cooling and thus preserved its initial heterogeneities, whereas the inner medium-grained part of the quartz diorite pluton crystallized from a convecting, isotopically homogeneous magma that had undergone advanced magmatic differentiation. The Sr–Nd–Hf isotopic compositions of the mafic magmatic enclaves are strikingly similar to those of the host granodiorite, implying their isotopic equilibration. The tonalitic enclaves have high Sr/Y ratios and most probably represent magmas derived from partial melting of thickened mafic lower crust. The Sr–Nd–Hf isotope data suggest that the granodiorite and biotite granite were dominantly derived from isotopically heterogeneous crustal sources. However, the granodiorite also has relatively high Mg#, Cr, and Ni, indicating a minor contribution from a mantle source. The granodiorite was constructed incrementally from a number of discrete melt batches that were generated by partial melting of mafic lower crust under variable water fugacity. These melt batches did not assemble into a large magma chamber and thus preserved their source chemical features. The granodiorite magma was also replenished by mafic and high Sr/Y magmas, resulting in abundant and compositionally diverse magmatic enclaves. The biotite granite formed by the successive accumulation of discrete magma pulses generated by dehydration melting of mafic lower crust under water-absent conditions. These magma pulses coalesced into a small single magma chamber where they underwent fractional crystallization. The various rock types exhibit distinct geochemical variations, indicating that the Meiwu batholith was constructed from multiple injections of magma over a protracted period. Fractional crystallization, assimilation, magma mixing and/or mingling occurred during magma ascent and at the emplacement level. Distinct magma sources...
played a primary role in controlling the chemical diversity of the igneous bodies at pluton and batholith scale.

Key words: batholith; magmatic processes; magma sources; magma chamber; western Qinling

INTRODUCTION

The formation of batholiths involves the transfer of mass and heat both laterally and vertically, operating on crust and upper mantle scales (Petford, 2012). Granitoid batholiths are the most abundant constituent of continental magmatic arcs, and their origin is essential to understanding the evolution and differentiation of the continental crust (Kemp & Hawkesworth, 2003; Menand et al., 2011). Large batholiths are commonly characterized by their geochemical, textural and petrological diversity (e.g. Memeti et al., 2010; Schoene et al., 2012). There are four generic stages in batholith formation: melt generation, melt segregation, magma ascent and magma emplacement (Petford et al., 2000). The compositional diversity of granitoids may be controlled by different source compositions, variable melting conditions (e.g. temperature, pressure/depth, water and oxygen fugacity), the degree of partial melting and fractional crystallization, wall-rock assimilation, magma mixing and mingling, restite unmixing and/or peritectic assemblage entrainment (e.g. DePaolo, 1981; Chappell et al., 1987; Tepper et al., 1993; Patiño Douce, 1999; Barbarin, 2005; Stevens et al., 2007; Clemens et al., 2009; Clemens & Stevens, 2012). Such diversity may originate during each stage of batholith formation, from the source to final emplacement. However, it remains controversial as to which is the dominant process in shaping the chemical variation of granitoid batholiths.

Recently, it has been demonstrated that many plutons grow incrementally by the assembly of successive and relatively small magma pulses over a few thousand to a few million years (e.g. Coleman et al., 2004; Glazner et al., 2004; Michel et al., 2008; Schaltegger et al., 2009; Schoene et al., 2012). The incremental growth model has far-reaching implications for the petrogenesis and the heterogeneity of magma bodies (Coleman et al., 2004; Glazner et al., 2004; Menand et al., 2011). Coleman et al. (2004) and Glazner et al. (2004) suggested that many plutons grow by repeated addition of small magma pulses from distinct sources, and rarely represent the products of crystallization of large magma chambers. However, as an alternative mechanism it has been proposed that large homogeneous magma chambers can exist, at least for some limited time, and that the geochemical variations observed in some plutons could result from shallow magmatic processes (e.g. Lipman, 2007; Annen, 2009; Paterson, 2009; Memeti et al., 2010). The conservation of the initial composition of the separate magma batches versus later homogenization by internal magma chamber processes depends on the emplacement and crystallization rate and the ability of the different magma batches to amalgamate (Clemens et al., 2010; Menand et al., 2011). Rapid magma pulses emplacement and solidification tends to preserve initial geochemical heterogeneities, whereas rapid amalgamation of magma batches and slow crystallization tends to result in large homogeneous magma chambers that subsequently undergo internal differentiation (Annen, 2009, 2011; Clemens et al., 2010). Thus, the impact of emplacement rate and assembly style of repeated magma pulses on the compositional variation and the construction of plutons is critical.

The Meiwu composite batholith, one of the largest batholiths in the West Qinling orogenic belt, is composed of quartz diorite, granodiorite and biotite granite, with abundant mafic magmatic enclaves (MME) and minor tonalitic enclaves (TE) in the granodiorite. The batholith provides an excellent case study with which to assess the causes of compositional diversity in composite batholiths. The granodiorite was previously dated at ~206–212 Ma by K–Ar on biotite (Tao, 1982). The quartz diorite has a U–Pb zircon sensitive high-resolution ion microprobe (SHRIMP) age of 245 ± 6 Ma (Jin et al., 2005). Preliminary geochemical studies have been reported for the granodiorite and biotite granite (Zhang et al., 2007a; Luo et al., 2012b). Here we present an integrated study of laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) U–Pb zircon ages and trace element and major and trace element and Sr–Nd–Hf isotopic compositions for each component of the Meiwu batholith. Combined with detailed field work and the results of previous studies, these data allow us to identify the petrogenetic processes responsible for the compositional diversity of the batholith, and provide important insights into the dynamics of construction of composite batholiths.

GEOLOGICAL SETTING

The Qinling orogen in central China separates the North China Block from the South China Block; it is subdivided into the East Qinling and West Qinling (Fig. 1a). The West Qinling orogenic belt is bounded by the East Kunlun and Qaidam terranes to the west and the Qilian orogen to the north, and to the south it is separated from the Bayan Har/Songpan–Ganzi block by the An’imaque–Mianlue suture zone (Fig. 1b). This suture zone contains abundant ophiolite fragments that are considered to be the relics of the Late Palaeozoic Palaeo-Tethys oceanic slab (Xu et al., 2002; Guo et al., 2007). Precambrian basement is rarely exposed; the sedimentary cover consists mainly of Devonian to Cretaceous sediments (Feng et al., 2002).
Fig. 1. (a) Simplified geological map of China, showing the major tectonic units. (b) Simplified map showing distribution of Early Mesozoic granitoids in the Qinling orogen (after Feng et al., 2002). (c) Geological map of the Meiwu composite batholith. Place names: WQ, West Qinling; EQ, East Qinling; SPGZ, Songpan–Ganzi terrane; DB, Dabie belt; SL, Sulu belt; QD (QDM), Qaidam; QL, Qilian Shan belt; KL (EKL), Kunlun Shan belt; NQL, North Qinling; SQL, South Qinling. Pluton names in the West Qinling: MB, Miba; MSL ; Mishuling; WQ, Wenquan; XH, Changba; CB, Huangzhuguan; HZG, Yeliguan; YLG, Xiahe; XK, Xiekeng; SPX, Shuanpengxi; TR, Tongren; JZ, Jianzha; HMH, Heimahe; WQ, Wenquan. Data sources for zircon U–Pb ages in Fig. 1b: East Qinling from Zhang et al. (2008, and references therein), Qin et al. (2009, 2010b) and Jiang et al. (2010); northwestern margin of South China Block from Li et al. (2007), Zhang et al. (2007b) and Qin et al. (2010a); West Qinling from Feng et al. (2002), Jin et al. (2005), Zhang et al. (2006), Qin et al. (2009), Luo et al. (2012a) and Li et al. (2014); Maixiu volcanic rocks from Li et al. (2013).
Fig. 1. Continued.
Early Mesozoic granitoids are widespread in the Qinling orogen (Fig. 1b). A large amount of geochronological and geochemical data has been reported for the granitoids in the East Qinling and the neighboring northwestern Yangtze Block (YZB) and in the eastern part of the West Qinling (Fig. 1b). These indicate magma crystallization ages of ~227–205 Ma (Late Triassic), coeval with the final amalgamation of the North China and the South China blocks (Zhang et al., 2007a, 2007b, 2008, and references therein; Qin et al., 2009, 2010a, 2010b; Jiang et al., 2010). Some granitoids from the central and western parts of the West Qinling (Fig. 1b) have zircon U–Pb ages of ~245–235 Ma (Feng et al., 2002; Jin et al., 2005; Zhang et al., 2006; Guo et al., 2012; Luo et al., 2012a, 2012b; Li et al., 2014), which are older than the granitoids in the adjacent areas. There are two opinions concerning the generation of these Middle Triassic granitoids in the West Qinling: they may have formed in an active continental margin setting related to subduction (Jin et al., 2005; Meng et al., 2005; Guo et al., 2012; Li et al., 2013, 2014) or in a post-collision continental setting (Zhang et al., 2006, 2008; Luo et al., 2012a, 2012b).

FIELD RELATIONSHIPS AND PETROGRAPHY

The Meiwu batholith, with an area of ~423 km², intruded into Carboniferous and Permain sediments, and is elongated NW–SE parallel to regional tectonic structures (Fig. 1c). It consists of quartz diorite, granodiorite and biotite granite (Fig. 1c), and contains abundant MME and minor TE in the granodiorite. The granodiorite constitutes the bulk of the batholith. Quartz diorite is exposed in the northeastern and southeastern marginal zones and was intruded by the granodiorite. The contact between them is gradual, indicating that they might be coeval. The biotite granite, exposed as a stock, intrudes the granodiorite in the southwestern corner of the pluton. The contact between the biotite granite and the granodiorite is sharp (Fig. 2a) and suggests that the biotite granite may have intruded slightly later than the granodiorite. MME are ubiquitous in the granodiorite but rare or absent in the biotite granite and quartz diorite (Fig. 2b–e). Scarc TE are also observed in the granodiorite (Fig. 2f). The MME and TE were thus incorporated into their host granodiorite slightly earlier than the emplacement of the biotite granite. The contacts between the magmatic enclaves and their host rocks are commonly sharp, although diffuse contacts are also observed (Fig. 2a–c).

Quartz diorite

The quartz diorite was sampled at the northern margin of the Meiwu batholith (Fig. 1c). It shows grain sizes ranging from 0.2 to 3 mm (Fig. 3a and b) with a progressive transition from a fine-grained margin to a medium-grained interior, suggestive of rapid chilling. The mineralogy is dominated by plagioclase (50–55 vol. %), quartz (10–15 vol. %), biotite (10–15 vol. %), amphibole (8–15 vol. %) and K-feldspar (2–5 vol. %); accessory minerals include titanite, apatite, zircon and magnetite. Plagioclase is lath-shaped with oscillatory zoning and polysynthetic twinning. K-feldspar is mostly anhedral; rare crystals are subhedral with tartan twinning. Amphibole and biotite are typically anhedral to subhedral and tend to form clots. Quartz is typically interstitial.

Mafic magmatic enclaves

The MME are generally ovoid, ellipsoidal, elongated, and lenticular (Fig. 2a and b). Their sizes commonly range from 10 cm to several meters. They contain the same mineral assemblages as the quartz diorite and their host granodiorite, but have different mineral proportions and textures. They are finer grained than their host-rock, with texture variation from equigranular to porphyritic. They are composed of plagioclase (40–45 vol. %), amphibole (10–15 vol. %), biotite (10–15 vol. %), K-feldspar (10–15 vol. %), кварц (3–5 vol. %), with minor apatite, titanite, zircon, and magnetite. Acicular apatite is common as inclusions in plagioclase and quartz (Fig. 3d and e). K-feldspar phenocrysts commonly have ovoid or irregular shapes, oscillatory zoning and simple twinning. Some K-feldspar phenocrysts occur as poikilitic crystals containing fine-grained plagioclase, biotite and amphibole (Fig. 3f, g and j). Amphibole and biotite are typically <1 mm in size, showing skeletal crystal and ragged flakes (Fig. 3e–g), respectively. Two types of plagioclase can be observed: lath-like crystals (0.2–0.6 mm) and less abundant larger anhedral crystals (1–2 mm) (Fig. 3e and i). Quartz xenocrysts with amphibole-rich rims are also present (Fig. 3h). Myrmekitic intergrowths of quartz and alkali feldspar are common (Fig. 3j).

Tonalitic enclaves

The TE are scarce in the granodiorite; they have the same shapes and sizes as the MME (Fig. 2f). They are fine-grained (0.2–1.5 mm) and equigranular in texture (Fig. 3k). They consist mainly of plagioclase (45–50 vol. %), quartz (20–25 vol. %), K-feldspar (10–15 vol. %), and biotite (7–10 vol. %), with accessory zircon, titanite, apatite, and Fe–Ti oxides. Plagioclase (0.5–1.5 mm) is mainly anhedral to subhedral, with polysynthetic twinning and oscillatory zoning. Biotite (0.2–0.5 mm) is euhedral to subhedral. Anhedral quartz (0.4–0.8 mm) commonly occurs interstitially.

Granodiorite

The granodiorite is fine- to medium-grained and consists of plagioclase (30–45 vol. %), quartz (20–25 vol. %), K-feldspar (10–15 vol. %), biotite (3–6 vol. %), amphibole (3–5 vol. %) and accessory minerals (titanite, apatite, zircon and magnetite). Plagioclase (1–2.5 mm) mainly forms subhedral euhedral lath-shaped crystals, and shows albite twinning and oscillatory zoning. K-feldspar
(1.5–3 mm) is anhedral to subhedral, and with a perthitic texture or tartan twinning. Locally, K-feldspar mega-crysts (generally >2 cm long) are present (Fig. 2b). Interstitial quartz is generally <1 mm. Amphibole occurs as euhedral rhombic crystal (1–1.5 mm) and subhedral prismatic crystals (2–3 mm) (Fig. 3c). Biotite (0.8–2 mm) occurs as lamellae and tabular crystals.

**Biotite granite**

The biotite granite is fine-grained and locally shows a weak flow orientation (Fig. 2d). It consists mainly of plagioclase (30–40 vol. %), K-feldspar (25–35 vol. %), quartz (25–30 vol. %) and biotite (3–6 vol. %), with accessory zircon, apatite and Fe–Ti oxides (Fig. 3l).

Plagioclase (1–1.5 mm) is mainly anhedral to subhedral, with polysynthetic twinning or oscillatory zoning. K-feldspar (0.5–1 mm) is anhedral with tartan twinning. Some K-feldspar crystals are partially resorbed and embayed by quartz (Fig. 3l). Quartz is anhedral with a diameter of 1–1.5 mm. Biotite commonly occurs interstitially.

**ANALYTICAL METHODS**

Fresh rock samples were crushed in a steel crusher and then powdered in an agate mill to a grain size < 200 mesh. Major elements were analyzed at the Hubei Geological Analytical Center, Wuhan. The analytical uncertainty is generally <5%. Trace elements, including
rare earth elements (REE), were measured using an Agilent 7500a ICP-MS system at the State Key Laboratory of Geological Processes and Mineral Resources (GPMR), China University of Geosciences, Wuhan. The sample-digestion procedure for ICP-MS analyses and analytical precision and accuracy are the same as those described by Liu et al. (2008c). Whole-rock Sr and Nd isotopic ratios were measured by thermal ionization mass spectrometry (TIMS) on a Triton system at GPMR. During the period of analysis the NBS987 standard gave an average $^{87}\text{Sr} / ^{86}\text{Sr}$ value of $0.710239 \pm 0.000010$ (2σ) and the BCR-2 standard an average $^{143}\text{Nd} / ^{144}\text{Nd}$ value of $0.512620 \pm 0.000004$ (2σ). Details of the Sr and Nd isotopic analytical procedures have been given by Gao et al. (2004).

Zircons were separated using conventional techniques. Cathodoluminescence (CL) images, taken at Northwest University, Xi’an, were used to check the internal textures of single zircons and to guide U-Pb dating and Hf isotope analysis. U-Pb dating and trace element analyses of zircon were carried out synchronously by LA-ICP-MS on an Agilent 7500 equipped with a

Fig. 3. Photomicrographs of representative samples of the Meiwu batholith. (a) Fine-grained quartz diorite (09104); (b) medium-grained quartz diorite (09108); (c) granodiorite (0907); (d) mafic magmatic enclave (MME) (0911), with abundant needle-likeapatite; (e–g) MME (09101) with abundant needle-like apatite (Ap); K-feldspar (Kfs) usually occurs as poikilitic crystals containing fine-grained plagioclase (Pl), biotite (Bt) and amphibole (Amp); (h) MME (09103); quartz (Qz) xenocrysts with hornblende-rich rims can be observed; (i) MME (09103); biotite is included in poikilitic plagioclase; (j) MME (09103), granophyric texture with myrmekitic intergrowths of quartz and alkali feldspar; (k) tonalite enclave (09113); (l) biotite granite (0912).
193 nm ArF excimer laser at GPMR. Operating conditions for the laser ablation system and the ICP-MS instrument and data reduction methods were the same as those described by Liu et al. (2010). A beam diameter of 32 μm was used. Zircon 91500 was used as an external standard. Common Pb correction was made by using the program ComPbCorr#3-17 (Andersen, 2002). Trace elements were calibrated against NIST SRM 610 and using ZrO2 (66-1%) as an internal standard (Liu et al., 2010). Zircon U–Pb data were processed using the ISOPLOT program (Ludwig, 2003).

Zircon Hf isotope analysis was carried out in situ using a Neptune multicollector (MC)-ICP-MS system at GPMR. Analytical spots were located close to or on top of the LA-ICP-MS spots. The instrumental conditions and data acquisition were described by Hu et al. (2012). The analyses were conducted with a beam diameter of 44 μm for larger grains and 32 μm for small grains, at a rate of 8 Hz, and laser energy of 60 mJ. The mass fractionations of Hf and Yb were calculated using an exponential law and values of 0.7325 for $^{179}\text{Hf}/^{177}\text{Hf}$ and 1.1248 for $^{173}\text{Yb}/^{171}\text{Yb}$, respectively (Blichert-Toft et al., 1997). Ratios used for the corrections were $^{176}\text{Lu}/^{175}\text{Lu} = 0.02656$ (Blichert-Toft et al., 1997) and $^{176}\text{Yb}/^{173}\text{Yb} = 0.7876$ (McCulloch et al., 1977), respectively. To monitor the accuracy of this correction, every 5–7 sample analyses were followed by analysis of the reference zircons. Reference zircons 91500, GJ-1, Mud Tank and Temora were analyzed as unknowns. During the analysis, the standard zircons gave $^{176}\text{Hf}/^{177}\text{Hf}$ ratios.
of 0.282302 ± 3 (2σ; MSWD = 1.4, n = 36) for 91500, 0.282007 ± 5 (2σ; MSWD = 3.1, n = 39) for GJ-1 and 0.28248 ± 5 (2σ; MSWD = 4.5, n = 41) for Mud Tank and 0.282698 ± 6 (2σ; MSWD = 3.1, n = 25) for Temora (Supplementary Data Table S1; supplementary data are available for downloading at http://www.petrology.oxfordjournals.org). These values are consistent with the recommended 176Hf/177Hf ratios within analytical error (Supplementary Data Fig. S1).

A decay constant for 176Lu of 1.865 × 10^{-11} a^{-1} was adopted (Scherer et al., 2001). Initial 176Hf/177Hf ratios, denoted as εHf(t), were calculated relative to the chondritic reservoir with a 176Hf/177Hf ratio of 0.282772 and 176Lu/177Hf of 0.0332 (Blichert-Toft et al., 1997). Single-stage Hf model ages (TDM1) were calculated relative to the depleted mantle, which is assumed to have a linear isotopic growth from 176Hf/177Hf = 0.279718 at 4.55 Ga to 0.283250 at the present day with a 176Lu/177Hf ratio of 0.0384 (Vervoort & Blichert-Toft, 1999). Two two-stage Hf model ages (TDM2) were calculated relatively to the depleted mantle, which is assumed to have a linear isotopic growth from 176Hf/177Hf = 0.279718 at 4.55 Ga to 0.283250 at the present day with a 176Lu/177Hf ratio of 0.0384 (Vervoort & Blichert-Toft, 1999). 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Fig. 4. Representative CL images of zircons from the Meiwu batholith. (a) Fine-grained quartz diorite (09104); (b) medium-grained quartz diorite (09108); (c) granodiorite (MR-1); (d) granodiorite (SLG-2); (e) mafic magmatic enclave (MME) (09101); (f) MME (0911); (g) tonalitic enclave (09113); (h) biotite granite (09116). The continuous-line circles show LA-ICP-MS dating spots and corresponding U–Pb ages (Ma), and the dashed-line circles show Lu–Hf isotope analysis spots and corresponding $e_{Hf}(t)$ values. White bars represent 100 $\mu$m.
Fig. 5. Zircon U–Pb concordia diagrams. (a) Fine-grained quartz diorite (09104); (b) medium-grained quartz diorite (09108); (c) granodiorite (MR-1); (d) granodiorite (SLG-2); (e) mafic magmatic enclave (MME) (09101); (f) MME (0911); (g) tonalitic enclave (09113); (h) biotite granite (09116). Insets in (a)–(e), (g) and (h) show weighted average ages. Inset in (f) shows probability density plot of ages. Shaded and dashed ellipses are not included in the age calculation. Ellipses represent 1σ uncertainty for single analyses.
in zircon are usually interpreted to reflect the cooling and fractional crystallization of the host magma; the Eu/Eu* variation may be related to either feldspar in the residue or to fractionation of the magma (Wooden et al., 2007; Claiborne et al., 2010).

The crystallization temperatures of zircon were calculated using the Ti-in-zircon thermometer ($T_{TiZ}$) (Ferry & Watson, 2007), assuming $a_{SiO_2} = 1$ and $a_{TiO_2} = 0.7$ as suggested by Claiborne et al. (2010). The $T_{TiZ}$ for sample 09104 (fine-grained quartz diorite) ranges from 775 to 924°C, with 64% of the analyses showing $T_{TiZ}$ greater than 827°C (Fig. 7). The $T_{TiZ}$ for 09108 (medium-grained quartz diorite) shows a limited range of variation between 795 and 827°C, except for one spot, which gave an anomalously high temperature (909°C) (Fig. 7). Samples MR-1 and SLG-2 from the granodiorite gave the lowest $T_{TiZ}$ (~600–750°C) (Fig. 7). MME sample 09101 has a larger $T_{TiZ}$ range (683–852°C), with 64% of the analyses above the highest $T_{TiZ}$ of the host granodiorite (750°C) (Fig. 7). Zircons from MME 0911 were not considered here because their trace elements have been affected by late-stage processes. The TE also have higher $T_{TiZ}$ (743–908°C) than the host-rock (Fig. 7). The biotite granite 09116 gives $T_{TiZ}$ (757–822°C) between that of the granodiorite and the quartz diorite (Fig. 7).

All the zircons from the Meiwu batholith show common trends in which Hf and Yb/Gd increase and Th/U and Eu/Eu* decrease with decreasing $T_{TiZ}$ (Fig. 7). However, zircons from different units have distinct geochemical features. Zircons from 09104 (fine-grained quartz diorite) have lower Hf [(9.9–10.6) $\times 10^3$ ppm] and Yb/Gd (5.2–6.9), and higher Th/U and Eu/Eu* (0.22–0.29) corresponding to higher $T_{TiZ}$ than zircons from 09108 (medium-grained quartz diorite) (Fig. 7). Zircons from the two granodiorites (MR-1 and SLG-2) have the highest Hf [(10.6–11.8) $\times 10^3$ ppm] and lowest Th/U, corresponding to the lowest $T_{TiZ}$, and display larger variation in Yb/Gd (4.7–26.9) and Eu/Eu* (0.06–0.40) (Fig. 7). Compared with zircons from the host granodiorite, zircons from the MME (09101) have relatively low Hf [(9.9–11.1) $\times 10^3$ ppm] but show similar Th/U, Yb/Gd and Eu/Eu* variations (Fig. 7). Zircons from the TE (09113) have the lowest Hf [(5.6–7.7) $\times 10^3$ ppm] and Yb/Gd (5.4–7.1), and the highest Th/U and Eu/Eu* (0.31–0.53) relative to higher $T_{TiZ}$ (Fig. 7). Zircons from the biotite granite (09116) show variations in Hf, Th/U, Yb/Gd and Eu/Eu* that correlate either positively or negatively with $T_{TiZ}$. As $T_{TiZ}$ decreases, Hf [(10.6–12.6) $\times 10^3$ ppm] and Yb/Gd (9.6–12.5) increase, and Th/U and Eu/Eu* (0.18–0.11) decrease (Fig. 7).

**Whole-rock major and trace elements**

Whole-rock major and trace element data for the Meiwu batholith (Table 2) show considerable variation from calc-alkaline to peraluminous and shoshonitic compositions.

**Quartz diorite**

The quartz diorites have SiO$_2$ of 58.46–62.55 wt % and are medium-K to high-K calc-alkaline (Fig. 8), with K$_2$O/Na$_2$O ratios between 0.23 and 0.91 (Fig. 9e). The rocks are characterized by high MgO (45.8–6.82 wt %), Mg#
They show enrichment of large ion lithophile elements (LILE) (e.g. Rb, Ba, Th, U, K) and light REE (LREE), but depletion of Nb, Ta, P and Ti (Fig. 10a). They display moderately fractionated REE patterns (LaN/YbN = 9.2–13.2) and weak negative Eu anomalies (Eu/Eu* = 0.71–0.95) (Fig. 10b).

**Mafic magmatic enclaves**

The MME have SiO2 controls of 56.72–62.08 wt %, are high-K to shoshonitic (Fig. 8a), and have K2O/Na2O ratios of 0.66–1.53 (Fig. 9e). Relative to the quartz diorites, the MME have similar high MgO, Mg#, Cr and Ni, but are more enriched in K2O, Na2O, P2O5, Rb, Ba, and Zr (Fig. 9; Table 1). The MME also have higher P2O5, Ba, Zr, Sm and Y, and lower Sr/Y and Eu/Eu* than the host granodiorites (Fig. 9 and Table 1). However, the contents of K2O, Na2O and Rb in the MME overlap with those of the host granodiorites (Figs 8a and 9d, i). The MME are enriched in LILE and show negative anomalies of Nb, Ta, P and Ti (Fig. 10a and c). They show convex REE patterns, and display more enriched middle REE (MREE) and heavy REE (HREE) and more pronounced negative Eu anomalies (Eu/Eu* = 0.25–0.57) than those of the quartz diorite and the host granodiorite (Fig. 10b and d).

**Tonalitic enclaves**

The TE analysed is medium-K calc-alkaline and weakly peraluminous (Fig. 8). It has SiO2 = 65.87 wt %, within the range of the host granodiorite, but lower MgO, Mg#, Cr and Ni than the granodiorite at this SiO2 content. The noteworthy differences between the TE and the host granodiorite are that the former has higher Al2O3 (17.1 wt %), Na2O (3.91 wt %), Sr (548 ppm), and Sr/Y (82), and lower K2O (2.43 wt %), K2O/Na2O (0.62), Yb (0–78 ppm) and Y (8–77 ppm) values (Fig. 9). The TE also shows enrichment of LILE and depletion of Nb, Ta, P and Ti (Fig. 10c). REE patterns show HREE depletion.
with a high LaN/YbN ratio (31) and a weak positive Eu anomaly (Eu/Eu* = 1.11) (Fig. 10d).

**Granodiorite**

The granodiorites, with SiO$_2$ contents ranging from 65.18 to 69.93 wt %, are metaluminous to weakly peraluminous, and plot in the high K-calc-alkaline field (Fig. 8). They have K$_2$O/Na$_2$O ratios between 0.98 and 1.85 (Fig. 9e) and lower MgO (1.47–3.24 wt %), Mg# (50–58), Cr (46–93 ppm) and Ni (12–50 ppm) than the quartz diorites and the MME, TiO$_2$, MgO, CaO and P$_2$O$_5$, and Cr and Ni are negatively correlated with SiO$_2$, whereas Al$_2$O$_3$, Na$_2$O, Rb, Sr, Zr, Th, REE, Sr/Y and Eu/Eu* scatter and show no correlation with SiO$_2$ (Fig. 9). In Fig. 10c, all samples show negative Nb, Ta, P and Ti anomalies and are enriched in Rb, Ba, Th, U and K. They show strongly fractionated REE patterns with LaN/YbN = 11.5–23.7 and variable Eu anomalies (Eu/Eu* = 0.39–0.92) (Fig. 10d).

**Biotite granite**

The biotite granites have SiO$_2$ contents of 72.07–77.10 wt % and are weakly peraluminous and high-K calc-alkaline (Fig. 8) with K$_2$O/Na$_2$O = 1.54–2.72 (Fig. 9e). They have the lowest MgO, Mg#, Cr and Ni relative to the other rock types. SiO$_2$ increases as TiO$_2$, Al$_2$O$_3$, MgO (Mg#), CaO, Na$_2$O, P$_2$O$_5$, Cr, Ni, Ba, Sr, Sr/Y and Eu/Eu* decrease and K$_2$O, K$_2$O/Na$_2$O, Rb, REE, Th (U)
Table 2: Continued

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| SiO₂ | 66.72 | 65.86 | 68.20 | 65.18 | 65.55 | 67.84 | 62.08 | 56.72 | 59.22 |
| TiO₂ | 0.44 | 0.74 | 0.43 | 0.54 | 0.53 | 0.40 | 0.48 | 0.69 | 0.57 |
| Al₂O₃ | 15.77 | 13.57 | 14.92 | 16.03 | 16.29 | 16.05 | 15.52 | 14.42 | 15.87 |
| FeO | 2.88 | 5.48 | 3.21 | 3.77 | 3.61 | 2.72 | 4.55 | 6.86 | 5.47 |
| MnO | 0.05 | 0.06 | 0.05 | 0.06 | 0.06 | 0.05 | 0.12 | 0.19 | 0.13 |
| MgO | 2.22 | 3.24 | 1.77 | 2.75 | 2.45 | 2.02 | 4.69 | 6.33 | 5.32 |
| CaO | 3.41 | 3.21 | 3.52 | 4.31 | 4.11 | 3.39 | 4.65 | 6.16 | 6.88 |
| Na₂O | 3.48 | 2.50 | 2.79 | 3.36 | 3.38 | 3.30 | 3.27 | 3.03 | 3.03 |
| K₂O | 3.56 | 3.39 | 3.93 | 3.44 | 3.46 | 3.51 | 4.01 | 3.25 | 4.65 |
| P₂O₅ | 0.12 | 0.13 | 0.10 | 0.13 | 0.15 | 0.11 | 0.12 | 0.40 | 0.25 |
| LOI | 0.92 | 0.92 | 0.88 | 0.70 | 0.50 | 0.65 | 0.60 | 1.79 | 0.92 |
| Total | 99.63 | 99.76 | 99.80 | 99.46 | 99.87 | 99.72 | 99.57 | 99.73 | 100.53 |
| K₂O/Na₂O | 1.02 | 1.85 | 1.41 | 1.02 | 1.02 | 0.98 | 1.23 | 1.07 | 1.53 |
| A/CNK | 1.00 | 0.95 | 0.98 | 0.94 | 0.97 | 1.01 | 0.85 | 0.73 | 0.84 |
| Mg# | 58 | 51 | 50 | 57 | 55 | 57 | 65 | 62 | 63 |
| Cr | 90.0 | 92.8 | 53.4 | 92.8 | 74.0 | 69.5 | 251 | 318 | 255 |
| Co | 10.2 | 20.2 | 8.3 | 69.5 | 71.9 | 70.7 | 77.8 | 22.7 | 66.2 |
| Ni | 25.7 | 49.6 | 22.9 | 32.1 | 26.8 | 21.9 | 105 | 110 | 112 |
| Nb | 853 | 281 | 361 | 442 | 463 | 536 | 412 | 345 | 461 |
| Y | 25.7 | 27.2 | 20.7 | 17.4 | 15.0 | 12.9 | 24.8 | 51.4 | 51.0 |
| Zr | 155 | 208 | 161 | 141 | 148 | 118 | 178 | 243 | 231 |
| Nb | 10.9 | 15.39 | 9.72 | 10.0 | 9.37 | 8.72 | 8.12 | 12.2 | 11.1 |
| Ba | 826 | 775 | 614 | 754 | 830 | 825 | 1771 | 794 | 1484 |
| Hf | 4.27 | 5.58 | 4.91 | 4.16 | 4.18 | 3.39 | 4.65 | 6.78 | 6.32 |
| Ta | 1.13 | 1.70 | 1.30 | 1.02 | 1.05 | 1.07 | 1.12 | 1.09 | 1.04 |
| Pb | 26.4 | 15.1 | 28.3 | 25.8 | 36.1 | 38.1 | 46.6 | 29.0 | 37.2 |
| Th | 22.11 | 33.58 | 16.94 | 20.86 | 20.87 | 14.62 | 9.53 | 8.74 | 15.02 |
| U | 2.62 | 5.91 | 5.53 | 5.93 | 5.76 | 5.69 | 5.93 | 5.98 | 10.4 |
| La | 41.0 | 49.9 | 40.0 | 38.4 | 42.9 | 23.3 | 18.7 | 28.1 | 37.3 |
| Ce | 80.7 | 97.6 | 78.1 | 75.3 | 82.8 | 46.4 | 52.8 | 79.1 | 83.3 |
| Pr | 8.57 | 9.86 | 7.49 | 8.02 | 8.67 | 5.07 | 7.05 | 12.95 | 10.65 |
| Nd | 30.5 | 40.0 | 30.8 | 29.8 | 31.1 | 18.9 | 29.9 | 57.4 | 44.5 |
| Sm | 5.54 | 7.68 | 5.60 | 5.37 | 5.59 | 3.81 | 6.82 | 13.9 | 10.0 |
| Eu | 1.16 | 0.93 | 1.00 | 1.18 | 1.24 | 1.07 | 1.20 | 1.05 | 1.32 |
| Gd | 4.31 | 6.55 | 5.17 | 4.53 | 4.28 | 3.16 | 5.77 | 11.6 | 8.23 |
| Tb | 0.58 | 0.93 | 0.67 | 0.61 | 0.58 | 0.44 | 0.84 | 1.69 | 1.21 |
| Dy | 3.07 | 4.46 | 3.31 | 3.31 | 2.86 | 2.37 | 4.68 | 9.29 | 6.57 |
| Ho | 0.61 | 0.86 | 0.69 | 0.59 | 0.55 | 0.42 | 0.85 | 1.75 | 1.18 |
| Er | 1.56 | 2.59 | 1.59 | 1.61 | 1.56 | 1.21 | 2.49 | 4.67 | 3.28 |
| Tm | 0.23 | 0.38 | 0.31 | 0.24 | 0.22 | 0.17 | 0.38 | 0.67 | 0.47 |
| Yb | 1.63 | 2.43 | 1.88 | 1.41 | 1.30 | 1.05 | 2.43 | 4.34 | 2.81 |
| Lu | 0.23 | 0.35 | 0.31 | 0.21 | 0.20 | 0.14 | 0.38 | 0.60 | 0.41 |
| Eu/Eu* | 0.70 | 0.39 | 0.56 | 0.71 | 0.74 | 0.92 | 0.57 | 0.25 | 0.43 |
| La₈N/Yb₈N | 18.1 | 14.7 | 15.2 | 19.5 | 23.7 | 16.0 | 5.5 | 4.7 | 9.5 |

(continued)
weighted mean of $-1.8 \pm 1.4$ and two-stage Hf model ages $T_{DM2}$ of 1.08–1.61 Ga, whereas zircons from 09108 (medium-grained quartz diorite) have homogeneous $\varepsilon_{Hf}(t)$ of $-2.1 \text{ to } -1.4$ with a weighted mean of $-1.8 \pm 0.3$ and $T_{DM2}$ of 1.36–1.46 Ga (Supplementary Data Table S4 and Fig. 12a). Two granodiorites (MR-1 and SLG-2) have uniform $\varepsilon_{Hf}(t)$ ($-5.0 \text{ to } -2.3$) with a weighted mean of $-3.7 \pm 0.4$ and $T_{DM2}$ of 1.39–1.58 Ga, consistent with the host granodiorite (Supplementary Data Table S4 and Fig. 12a). The tonalite enclave has a zircon $\varepsilon_{Hf}(t)$ of $-2.0 \text{ to } 0.0$ with a weighted mean of $-1.2 \pm 0.4$ and $T_{DM2}$ of 1.27–1.40 Ga. The biotite granites have $\varepsilon_{Hf}(t)$ of $-4.2 \text{ to } -2.6$ with a weighted mean of $-3.2 \pm 0.4$ and $T_{DM2}$ of 1.43–1.54 Ga, comparable with the granodiorites (Supplementary Data Table S4 and Fig. 12a), suggesting that initial $^{176}Hf/^{177}Hf$ ratios remain almost unaffected, even when the U–Pb isotopic system has been disturbed (Bomparola et al., 2007). Both MME have uniform $\varepsilon_{Hf}(t)$ ($-5.0 \text{ to } -2.3$) with a weighted mean of $-3.7 \pm 0.4$ and $T_{DM2}$ of 1.39–1.58 Ga, consistent with the host granodiorite (Supplementary Data Table S4 and Fig. 12a). The tonalite enclave has a zircon $\varepsilon_{Hf}(t)$ of $-2.0 \text{ to } 0.0$ with a weighted mean of $-1.2 \pm 0.4$ and $T_{DM2}$ of 1.27–1.40 Ga. The biotite granites have $\varepsilon_{Hf}(t)$ of $-4.2 \text{ to } -2.6$ with a weighted mean of $-3.2 \pm 0.4$ and $T_{DM2}$ of 1.43–1.54 Ga, comparable with the granodiorites (Supplementary Data Table S4 and Fig. 12a), suggesting that initial $^{176}Hf/^{177}Hf$ ratios remain almost unaffected, even when the U–Pb isotopic system has been disturbed (Bomparola et al., 2007). Both MME have uniform $\varepsilon_{Hf}(t)$ ($-5.0 \text{ to } -2.3$) with a weighted mean of $-3.7 \pm 0.4$ and $T_{DM2}$ of 1.39–1.58 Ga, consistent with the host granodiorite (Supplementary Data Table S4 and Fig. 12a). The tonalite enclave has a zircon $\varepsilon_{Hf}(t)$ of $-2.0 \text{ to } 0.0$ with a weighted mean of $-1.2 \pm 0.4$ and $T_{DM2}$ of 1.27–1.40 Ga. The biotite granites have $\varepsilon_{Hf}(t)$ of $-4.2 \text{ to } -2.6$ with a weighted mean of $-3.2 \pm 0.4$ and $T_{DM2}$ of 1.43–1.54 Ga, comparable with the granodiorites (Supplementary Data Table S4 and Fig. 12a), suggesting that initial $^{176}Hf/^{177}Hf$ ratios remain almost unaffected, even when the U–Pb isotopic system has been disturbed (Bomparola et al., 2007). Both MME have uniform $\varepsilon_{Hf}(t)$ ($-5.0 \text{ to } -2.3$) with a weighted mean of $-3.7 \pm 0.4$ and $T_{DM2}$ of 1.39–1.58 Ga, consistent with the host granodiorite (Supplementary Data Table S4 and Fig. 12a). The tonalite enclave has a zircon $\varepsilon_{Hf}(t)$ of $-2.0 \text{ to } 0.0$ with a weighted mean of $-1.2 \pm 0.4$ and $T_{DM2}$ of 1.27–1.40 Ga. The biotite granites have $\varepsilon_{Hf}(t)$ of $-4.2 \text{ to } -2.6$ with a weighted mean of $-3.2 \pm 0.4$ and $T_{DM2}$ of 1.43–1.54 Ga, comparable with the granodiorites (Supplementary Data Table S4 and Fig. 12a).
et al. Calmus et al. HREE and Y depletions, and high Sr, Sr/Y and LaN/YbN characterized by strong LILE and LREE enrichments, type HMA and Baja California HMA (bajaites) are formed. Archean sanukitoids, Western Aleutians Adak-diorites also exhibit some distinct geochemical differences. Archean sanukitoids and Cenozoic HMA, the quartz diorites have high MgO, Mg#, Cr and Ni contents, which are generally similar to those of Archean sanukitoids suites (e.g. Shirey & Hanson, 1984; Smithies & Champion, 2000) and Cenozoic high-Mg arc andesites (HMA) (e.g. Western Aleutians, Yogodzinski et al., 1994, 1995; Mt. Shasta, Baker et al., 1994; Grove et al., 2002; Setouchi, Tatsumi, 2006; Baja California, Calmus et al., 2003). However, compared with the Archean sanukitoids and Cenozoic HMA, the quartz diorites also exhibit some distinct geochemical differences. Archean sanukitoids, Western Aleutians Adak-type HMA and Baja California HMA (bajaites) are characterized by strong LILE and LREE enrichments, HREE and Y depletions, and high Sr, Sr/Y and LaN/YbN values (Yogodzinski et al., 1995; Smithies & Champion, 2000; Calmus et al., 2003). However, the Meiwu quartz diorites in contrast show lower LREE, Sr, Sr/Y and LaN/YbN and higher HREE; they are characterized by higher LILE and LREE (Fig. 10a and b), higher SiO₂, and lower MgO, Mg#, Cr, and Ni (Fig. 13), and are distinct from the the Setouchi HMA in western Japan (Shimoda et al., 1998; Tatsumi, 2006) and Mt. Shasta HMA in the western USA (Baker et al., 1994; Grove et al., 2002). The Meiwu quartz diorites have similar MgO, Mg#, Cr and Ni to the Western Aleutian Pip-type HMA (Fig. 13), but have higher LILE and LREE and LILE/LREE ratios (Fig. 10a and b) (Yogodzinski et al., 1994). Because there are both similarities to and differences from these mantle-derived rocks, it is important to determine whether the Meiwu high-Mg quartz diorites could also represent a primitive mantle-derived magma.

All Meiwu quartz diorites have similar mineral assemblages, whole-rock geochemistry, and homogeneous ln Sr (0.707–0.708) and εNd(t) (~4.0); the two studied samples also have consistent average zircon εHf(t) values (~1–8), suggesting that they are cognate and derived from the same source. However, zircons from the fine-grained sample 09104 show a much larger range of εHf(t) variation (up to ~8 ε units) compared with that (only < 1 ε units) of the medium-grained sample 09108. This excludes the possibility that the quartz diorites could represent a uniform primitive magma. Instead, they reflect open-system magmatic processes, involving mixing of melts from different sources. Thus, the Meiwu quartz diorites are not comparable with Archean sanukitoids and Cenozoic arc HMA, plotting within the field of the east Setouchi and Oto-Zan andesitic rocks in Japan (Fig. 13). These rocks have been suggested to be formed by mixing between mantle-derived mafic magmas (HMA or basaltic magmas) and crustal-derived felsic magmas (Kawabata & Shuto, 2005; Tatsumi et al., 2006).

The quenched margins (fine-grained) of the quartz diorite pluton indicate rapid cooling, whereas the inner part of the pluton is medium-grained, suggesting relatively slow cooling and crystallization. Zircons may record changes in ambient conditions during their growth and thus may provide direct evidence for such cooling.
Zircons from the fine-grained sample 09104 have a smaller size than those from the medium-grained sample 09108 (Fig. 4), also suggesting that the zircons in the margins crystallized more rapidly. Zircons from the medium-grained sample display sector zoning, typical for high-$T$ melts, indicating that they may have grown over a prolonged period (Schaltegger et al., 2009, and references therein).

Zircons from the fine-grained sample 09104, which have higher $T_{\text{RZ}}$, Th/U and Eu/Eu*, lower Hf and Yd/Gd
Fig. 10. Primitive mantle (PM) normalized trace element patterns (a, c, e) and chondrite-normalized rare earth element patterns (b, d, f) for the Meiwu batholith. Chondrite and PM values are from Sun & McDonough (1989). Archean high-Mg sanukitoids and Cenozoic high-Mg andesites from worldwide localities are shown for comparison. Data sources: Archean high-Mg sanukitoids and Western Aleutians Adak-type HMA from Smithies & Champion (2000); Baja California HMA (bajaites) from Calmus et al. (2003); Western Aleutians Adak-type HMA from Yogodzinski et al. (1995); Western Aleutians Pip-type HMA from Yogodzinski et al. (1994) and Setouchi sanukitoid from Tatsumi (2006).
The MME show typical igneous textures, and do not show a common chemical relationship with the host granodiorite. This clearly contradicts the restite model (Collins, 1998). The fine-grained texture and presence of acicular apatite in the MME (Fig. 3d and e) are indicative of rapid cooling and incompatible with the model of early cumulates (Dahlquist, 2002). Any disrupted fine-grained chilled margins would be incorporated late into the host granodiorite as totally or partially solidified bodies, and should be subangular to subrounded, with sharp and planar contacts (Donaire et al., 2005), which we did not observe here. The elongated and lobate shapes of the Meiwu MME, and lack of evidence of solid-state deformation or felsic veins (Fig. 2c–e), rather suggest that their parental mafic magma was injected into the host granodiorite magma while it was still partially molten. Furthermore, the presence of poikilitic K-feldspar megacrysts, plagioclase showing corrosion and overgrowths, and quartz xenocrysts with amphibole rims (Fig. 3e–j) indicates that the Meiwu MME are globules of mafic magma that were injected into and mingled with the host magma. These mafic globules were initially quenched and then slowly cooled in the host granitic magma (Vernon, 1990).

The absence of any isotopic contrast between the MME and their host granitoid suggests that there could be genetic connection between them (Figs 11 and 12). The parent magmas of the MME and their granitoid host could be derived either from a common source or from two distinct source materials that have reached

(Fig. 7), and a much larger εNd(t) variation (Fig. 12), may have crystallized from a high-temperature, less differentiated magma that cooled rapidly, thus preserving the initial heterogeneities. In contrast, zircons from the medium-grained sample 09108, with lower TIR, Th/U and Eu/Eu′, higher Hf and Yd/Gd (Fig. 7), and a restricted range of εNd(t) variation (Fig. 12), may have crystallized from a more homogeneous convecting magma chamber that had undergone large-scale magmatic differentiation (Wooden et al., 2007; Claihorne et al., 2010). Zircons from these two samples have strong negative Eu anomalies (Fig. 7d), indicating that they crystallized after significant feldspar fractionation. Zircons from sample 09108 have more pronounced negative Eu anomalies (Eu/Eu′ = 0.17–0.23) than those of sample 09104 (Eu/Eu′ = 0.22–0.29), implying that the magma chamber interior had experienced more advanced feldspar fractionation than at its border. This is also consistent with the whole-rock Eu anomalies (09104, Eu/Eu′ = 0.88 vs 09108, Eu/Eu′ = 0.72).

Mafic magmatic enclaves: magma mixing/mingling and diffusional exchange

Several hypotheses have been proposed for the origin of MME: (1) refractory residual source rock (Chappell et al., 1987); (2) mantle-derived mafic magmas (Barbarin, 2005); (3) autoliths, including cognate more mafic melt (Vernon, 1984), early cumulates (Dahlquist, 2002) and disrupted chilled margins (Donaire et al., 2005).
complete isotopic equilibration (Barbarin, 2005). The Meiwu MME have relatively high MgO, Mg#, Cr and Ni, indicating that their parent magma was dominantly mantle-derived. Many geochemical characteristics of the MME are similar to those of the quartz diorites, suggesting that the Meiwu quartz diorites may also represent their magmatic parent. Zircons from the two studied MME have similar zircon $t_{Hf}^d$ values to those from the host granodiorite (Fig. 12). Li et al. (2012) have reported a similar phenomenon from the Gushan granite in North China and suggested that zircons in the MME crystallized from the magma after considerable chemical equilibration with the host magma. A similar interpretation could also apply to our samples. Zircons from MME 09101 have similar internal structures and trace element ratios to those of the host granodiorite (Fig. 12). Li et al. 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host rocks, which could support this conclusion (Figs 4 and 7).

Most of major and trace elements in the MME do not display linear co-variations, and are distinctly different from the trends in the host rocks (Fig. 9). This rules out the likelihood that two-component magma mixing has played a significant role in shaping the composition of the MME. Physical mingling may explain the transference of some quartz, K-feldspar megacrysts and plagioclase xenocrysts from the host magma into the MME, but fails to account for the enrichment of K$_2$O, Na$_2$O, P$_2$O$_5$, Rb, Ba, Zr, Y, MREE and HREE in the MME relative to the host granodiorite. Experimental studies have shown that some mobile elements (K, Na, P, Rb and Ba) tend to migrate from felsic magma to mafic magma (Blundy & Sparks, 1992). The inward diffusion of K can lead to the crystallization of biotite that can act as a sink for K (Rb and Ba), which is in agreement with the enrichment of biotites in the MME. The enrichment of P then may have resulted in the formation of abundant apatite crystals (Fig. 3d and e).

The Meiwu MME exhibit convex REE patterns with enrichment of MREE and HREE and strong negative Eu anomalies compared with the quartz diores and the host granodiorite. Several studies have suggested that MME in felsic rocks may exhibit geochemical zonation and display MREE and HREE enrichment and more negative Eu anomalies from core to rim (Fig. 14a–c) (e.g. Eberz & Nicholls, 1990; Jiang et al., 2005; Farner et al., 2014). This indicates that the centres of large MME may represent the least modified chemical composition. The REE patterns of the bulk Meiwu MME are parallel to those of the MME margins, suggesting that the MME had experienced a high degree of chemical exchange with the host magma. Because REE diffusion rates in felsic melts are slower than in mafic melts, many of the REE partition preferentially into minerals that are abundant in the MME, such as amphibole and/or apatite (Blundy & Sparks, 1992; Barbarin, 2005). To constrain this issue further, the REE patterns of amphibole and apatite from the Meiwu MME are shown in Fig. 14d and Supplementary Data Table S5. The results show that a combination of apatite and amphibole could explain the REE patterns of the MME (Fig. 14d). The good negative correlation between P$_2$O$_5$ and Eu/Eu* in the MME suggests that apatite may play a more important role than amphibole (Fig. 14d). Simple mass-balance calculations show that an additional 0.5% apatite in the mafic parent magma could result in the REE patterns of the Meiwu MME (Fig. 14d). This is consistent with the presence of abundant acicular apatite in the MME.

In summary, we conclude that the Meiwu MME were produced by a two-stage process. The pre-injection stage parental magmas of the MME, by analogy with the quartz diores, were mainly formed at depth by mixing between a mantle-derived mafic magma and crustal felsic magmas. After injection into the host magma, the MME magmas experienced local mechanical mingling and diffusional exchange with their host magma.

**Tonalitic enclaves: partial melting of the thickened mafic crust**

The Meiwu TE has a distinct chemical and Sr–Nd–Hf isotopic composition compared with its host granodiorite, implying a distinct magma source. It has high Na$_2$O, Al$_2$O$_3$, Sr, Sr/Y (>40) and La$_{N}$/Yb$_{N}$ (>20), and low HREE (Yb < 1.8) and Y (<18), and lacks a negative Eu anomaly. High Sr/Y in evolved magmatic rocks can be generated by several processes, including partial melting of subducted ocean crust (Defant & Drummond, 1990), partial
melting of thickened mafic lower crust (Atherton & Petford, 1993), melting of delaminated mafic lower crust (Gao et al., 2004) or fractionation of garnet and amphibole from basaltic magmas (Macpherson et al., 2006).

Slab-derived high Sr/Y rocks from modern arcs have mid-ocean ridge basalt (MORB)-like Sr–Nd isotopic signatures (Defant & Drummond, 1990). The Sr–Nd isotope composition of the Meiwu TE is distinct from those of the MORB-type rocks from the Mian-Lue and A’nimaque ophiolites (Fig. 11a) (Xu et al., 2002; Guo et al., 2007), indicating that the TE magma could not have been derived from melting of a subducted oceanic slab (Defant & Drummond, 1990). The TE has lower K2O/Na2O, MgO, Mg#, Cr, and Ni than adakitic rocks that have been generated by partial melting of delaminated lower crust (Gao et al., 2004), excluding this possibility. The limited number of whole-rock samples makes it difficult to evaluate the possibility of a fractional crystallization origin for the TE. Zircon trace element contents and element ratios may provide additional constraints on this process. Wooden et al. (2007) and Claiborne et al. (2010) have suggested that Hf and Yb/Gd should increase and Th/U decrease with fractional crystallization towards lower temperatures. However, the lack of any correlation of $T_{TiZ}$ vs Hf, Th/U and Yb/Gd in the zircons from sample 09104 argues against fractional crystallization (Fig. 7). The studied TE yields a $T_{TiZ}$ temperature range from $\sim$740 to 910°C, which straddles the solidus ($\sim$820°C) for partial melting of arc basalts to basaltic andesites under fluid-absent conditions (Moyen & Stevens, 2006). According to its geochemical and Sr–Nd–Hf isotopic composition (Figs 11 and 12), the TE parent magma was most probably derived by partial melting of thickened, older, mafic lower crust. The TE has high Sr/Y and La/Yb, low HREE (Y) and lacks a negative Eu anomaly, which together

**Fig. 13.** Bulk-rock compositions: (a) MgO vs SiO2; (b) Mg# vs SiO2; (c) Cr vs SiO2; (d) Ni vs SiO2. Also shown for comparison are fields for average primitive arc basalts (Kelemen et al., 2003), Baja California HMA (Calmus et al., 2003), Western Aleutians Adak-type HMA (Yogodzinski et al., 1995), Western Aleutians Pip-type HMA (Yogodzinski et al., 1994), Mt. Shasta HMA (Baker et al., 1994; Grove et al., 2002), Setouchi HMA (Tatsumi, 2006), East Setouchi and Oto-Zan andesitic rocks (Kawabata & Shuto, 2005; Tatsumi et al., 2006), XK diorites (Guo et al., 2012; Luo et al., 2012a) and metabasaltic and eclogite experimental melts (0.7–3 GPa) (Rapp & Watson, 1995; Johannes & Holtz, 1996; Sisson et al., 2005).
with the concave-up REE pattern suggests the presence
of residual amphibole and minor garnet and the ab-
sence of plagioclase in its source at pressures of 10–12
kbar equivalent to a crustal thickness of 30–40 km (Rapp
& Watson, 1995; Moyen & Stevens, 2006).

Granodiorite: melting of mafic rocks under
variable water fugacity
The Meiwu granodiorites show distinct compositional
trends from the other rock types of the batholith (Fig. 9),
suggesting that they were not produced by fractional
crystallization of mafic magma (e.g. the quartz diorite)
or by simple mixing between a mafic magma and a
more felsic magma (e.g. the biotite granite). For the
granodiorites, negative correlations of MgO and P2O5,
and Cr and Ni vs SiO2 could be explained by fraction-
arion crystallization. However, many other elements
such as K2O, Na2O, Al2O3, Rb, Sr, Zr, Th and REE show
trends versus SiO2 that are too scattered (Fig. 9) to be
compatible with a fractionation model.

The Meiwu granodiorites are characterized by variable
Na2O, Al2O3 and Sr contents, and Sr/Y and Eu/Eu* ratios, which are most probably controlled by either
plagioclase fractionation at the emplacement level or
variable proportions of residual plagioclase in the
source. Their Sr/Y and Eu/Eu* ratios show no obvious
correlation with CaO (Fig. 15), suggesting that the vari-
able Sr/Y and Eu/Eu* ratios were inherited from the
source region rather than acquired at the emplacement
level. The consumption of plagioclase in the source
would increase the Sr/Y and Eu/Eu* ratio and CaO con-
tent in the resultant magma. Experimental studies on
melting of mafic rocks have shown that plagioclase sta-
bility depends not only on pressure, but also on water
content (Moyen & Stevens, 2006). Under conditions of
fluid-absent melting, plagioclase stability is strongly
pressure dependent, and plagioclase would gradually
disappear at pressures above 10 kbar. Under fluid-
present conditions, plagioclase begins to disappear at
lower temperatures than that of fluid-absent melting.
Both the zircon saturation temperature (~750–790°C)

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Fig. 14. Comparison of chondrite-normalized REE patterns of the Meiwu MME with those of other plutons and volcanic rocks: (a) the Swifts Creek Pluton, SE Australia (Eberz & Nicholls, 1990); (b) the Xiangshan volcanic rocks, SE China (Jiang et al., 2005); (c) the Bernasconi Hills Pluton, North America (Farner et al., 2014). (d) Chondrite-normalized REE patterns of apatite and amphibole in the Meiwu MME. The light-green dashed line in (d) represents a simple mass-balance calculation (95% average quartz diorite + 0.5% apatite). Inset shows the relation between P2O5 and Eu/Eu* for the Meiwu MME and the quartz diorite.
(Watson & Harrison, 1983) and $T_{\text{TiZ}} \sim 600–750^\circ \text{C}$ of the Meiwu granodiorites are lower than the solidus $\sim 820–900^\circ \text{C}$ for the onset of fluid-absent dehydration melting of amphibole at low pressure (Moyen & Stevens, 2006). Thus, we suggest that the Meiwu granodiorites were most probably generated by melting of mafic rocks under fluid-present conditions at low pressure. The variable Na$_2$O, Sr and Sr/Y and negative Eu anomalies in the granodiorites were mainly controlled by variable water fugacity.

The Meiwu granodiorites also have relatively high MgO, Mg#, Cr (50–58), and Ni (12–50 ppm). Their MgO and Mg# are higher than those of experimental melts produced by melting of metabasalt and eclogite (Fig. 13a and b) (Johannes & Holtz, 1996; Rapp & Watson, 1995; Sisson et al., 2005), indicating that they are unlikely to represent pure crustal melts and probably have a minor mantle-derived magma contribution. The Sr–Nd–Hf isotope compositions of the granodiorites are, however, consistent with those of other crustal melts in the adjacent area (Luo et al., 2012a). This indicates that the isotopic signature of a small amount of mantle-derived magma was mixed with that of crustal melts in the deep magma reservoirs (DePaolo, 1981; Hildreth & Mooibath, 1988).

**Biotite granite: fractional crystallization**

The biotite granites have high SiO$_2$ and low MgO, Mg#, Cr and Ni, with high $I_{\text{Sr}}$ and negative $\varepsilon_{\text{Nd}}(t)$ and zircon $\varepsilon_{\text{Hf}}(t)$ values, indicating that their parental magma was predominantly derived from crustal materials, without a significant mantle contribution. As illustrated in Fig. 9, most of the selected elements in the biotite granites show linear and/or slightly curvilinear trends with SiO$_2$, which can be attributed to either fractional crystallization or partial melting.

The mineral assemblage of the biotite granites consists of plagioclase, biotite and K-feldspar with minor zircon, Fe–Ti oxides and apatite. P$_2$O$_5$ shows a negative correlation with SiO$_2$ (Fig. 9f) implying apatite fractionation. Separation of Fe–Ti oxides resulted in depletion in Ti, Nb and Ta (Fig. 10e). CaO, Na$_2$O, Sr, Sr/Y and Eu/Eu* decrease with increasing SiO$_2$, which, combined with the positive relationship between Eu/Eu* and Sr/Y versus CaO (Fig. 15), is indicative of plagioclase fractionation. The biotite granites display increasing K$_2$O and Rb with increasing SiO$_2$ (Figs 8a and 9i), precluding K-feldspar and biotite fractionation. Zircons from 09116 show increasing Hf and Yb/Gd and decreasing Th/U and Eu/Eu* with decreasing temperature ($T_{\text{TiZ}}$) (Fig. 7), indicating that they crystallized from an evolving magma coexisting with plagioclase. Thus, the chemical variation of the biotite granites was controlled by fractional crystallization.

To understand further the origin of the biotite granites, we consider the least evolved sample (WQ116) as the parent magma to the more evolved members. This sample is weakly peraluminous with a high K$_2$O/Na$_2$O ratio ($\geq 1$) and a strongly fractionated REE pattern (La$_N$/Yb$_N > 25$) including a moderate negative Eu anomaly (Eu/Eu* = 0.55). This suggests that the biotite granite parental magma was most probably generated by partial melting of medium- to high-K basaltic protoliths with residual plagioclase and amphibole (Sisson et al., 2005).

**Multiple magma sources for the Meiwu batholith**

The contrasts in isotopic composition among the various units of the Meiwu batholith suggest that at least three components are involved in the magma source (Figs 11 and 12).

The quartz diorites have high MgO, Mg#, Cr and Ni, indicating that their parent magma was dominantly mantle-derived. Their high $I_{\text{Sr}}$, intermediate $\varepsilon_{\text{Nd}}(t)$ values and variable $\varepsilon_{\text{Hf}}(t)$ values could be acquired by melting of either an enriched mantle source or a depleted mantle source, followed by crustal contamination. To further constrain their mantle source characteristics, we used a two-component mixing model to evaluate their Nd isotopic compositions (DePaolo & Wasserburg, 2002).
In this modeling, MORB-type basalt from the A’nimaque suture zone was used to represent the depleted mantle end-member, with Nd = 9 ppm, $\varepsilon_{Nd}(t) = +12$ (Fig. 1b and Guo et al., 2007); the adjacent and coeval Xiexen (XK) pyroxene diorite to represent the enriched mantle end-member, with Nd = 15 ppm, $\varepsilon_{Nd}(t) = -1.1$ (Fig. 1b and Luo et al., 2012a); and the Meiwu biotite granite to represent the crustal end-member, with Nd = 35 ppm, $\varepsilon_{Nd}(t) = -8.5$. Our calculations suggest that the mafic end-member of the Meiwu quartz diorites could be derived only by melting of an enriched mantle source (Supplementary Data Fig. S2). A depleted mantle origin would require incorporation of ~45% crustal components (Supplementary Data Fig. S2), but such extensive assimilation would significantly reduce their high MgO, Cr and Ni contents and Mg#. For an enriched mantle origin, only ~25% of crustal components would be required (Supplementary Data Fig. S2). Therefore, the quartz dikes probably originated from melting of an enriched-mantle source, and the resultant melts were subsequently contaminated by ~25% crustal melt.

The Meiwu quartz diorite and the XK pluton have positive zircon $\varepsilon_{zr}(t)$ values up to +3 to +5 (Fig. 12) (Guo et al., 2012; Luo et al., 2012a), suggesting that they share a similar mafic end-member derived from an enriched mantle source. However, the Meiwu quartz dikes also contain zircon with $\varepsilon_{zr}(t)$ of -5.4 to -4.4 and $T_{DM2}$ of 1.5–1.6 Ga. The Meiwu biotite granites and the SPX granodiorites (Luo et al., 2012a) have similar ranges of zircon $\varepsilon_{zr}(t)$ and $T_{DM2}$, which may represent the felsic end-member magma. The Meiwu MME have major element compositions close to the quartz dikes, and have isotopic compositions similar to those of the host granodiorites, suggesting that the Meiwu MME share a similar magma source with the quartz dikes, but that their primary source signature has been eliminated during prolonged and complex interaction with the host granodiorite.

The Meiwu TE, biotite granites and granodiorites have higher $I_{Sr}$, and negative $\varepsilon_{Nd}(t)$ and $\varepsilon_{Hf}(t)$ values (Figs 11 and 12), with older crustal model ages, suggesting that their magma sources involved reworking of crustal basement materials. Because of the rare exposure of the West Qinling basement, studies of the basement have relied on the geochemical and Sr–Nd–Pb isotopic compositions of Triassic crust-derived granitoids (Zhang et al., 2007a) and zircon xenocrysts within the Neogene basalts probably entrained from the deep crust (Zheng et al., 2010). Zhang et al. (2007a) concluded that the Triassic granitoids were derived from partial melting of Meso- to Neoproterozoic high-K (Rb) basaltic basement rocks in the West Qinling orogenic belt. Zheng et al. (2010) suggested that the West Qinling basement had experienced juvenile crustal growth at ~2.7–2.4, 2.05–1.9 and 1.1–0.8 Ga and crustal reworking at ~1.8 and 1.4 Ga (Fig. 16), and that the West Qinling basement has an affinity with that of the Yangtze Block.

Assemblage sequence and emplacement rates of the Meiwu batholith

Our LA-ICP-MS zircon U–Pb data cannot resolve the detailed emplacement timescale for the Meiwu batholith. However, the relative assembly sequences for the various units of the batholith are well constrained by their field relationships, progressing from the quartz dikes, through the granodiorite, the MME and the TE to the biotite granite. The mantle source contribution decreases from the mafic to the felsic magmatism. Field relations, petrographic characteristics and geochemical
data clearly indicate that the Meiwu batholith cannot be interpreted as the product of crystallization of a single large magma chamber, but was constructed from multiple magma injections over a protracted period.

The maximum emplacement period (~4 Myr) and estimate of the volume of the batholith permit a rough estimate of average emplacement rates. The area of the batholith is ~423 km² (Fig. 1c). A thickness of between 3 and 6 km can be estimated using the empirical power-law formula for plutons (McCaffrey & Petford, 1997). The long-term average emplacement rate calculated for the batholith ranges from $3 \times 10^{-4}$ to $6 \times 10^{-4}$ km² a⁻¹. Taking the uncertainties in our LA-ICP-MS zircon dating results and the estimated volumes into account, the long-term average magma emplacement rate of the batholith is within an order of magnitude of previous estimates for other plutons worldwide, which mostly vary from $10^{-4}$ to $10^{-2}$ km³ a⁻¹ (de Saint Blanquat et al., 2011).

### Comparison with other intrusive systems

The pluton incremental growth model has initiated vigorous debates about the link between contemporaneous volcanic rocks and plutons, the dynamics of crustal magma chambers and the causes of chemical compositional variations in granitic magmas.

The Tuolumne intrusive suite in North America is one of the best studied large batholiths. Coleman et al. (2004) and Glazner et al. (2004) suggested that it was assembled mainly by small, dike-like increments over ~10 Myr and that its chemical diversity was mainly inherited from the source. However, Memeti et al. (2010) suggested that the single lobes (~10–60 km²) of the batholith each represent the minimum size of the magma chamber that was fed by either a single pulse or several pulses of magma and that chemical variation was related to in situ crystal–liquid separation. Repeated magma pulses are also able to form a large main magma chamber, such as the Cathedral Peak granodiorite (Burgess & Miller, 2008). Coleman et al. (2012) also suggested that the chemical variations on kilometer scale repeated cycles in the western Half Dome granodiorite were governed by in situ crystal–liquid separation, but that chemical trends at the intrusion scale were dominated by variation in the magma source.

Compared with the Tuolumne intrusion, the construction and evolution of the Meiwu batholith is more complex. The Meiwu quartz diorite may represent a
large magma chamber, like the Cathedral Peak granodiorite of the Tuolumne batholith (Burgess & Miller, 2008). It was constructed from initially heterogeneous, hybrid magma batches that were generated in deep crustal hot zones (Annen et al., 2006) or MASH (melting, assimilation, storage and homogenization) zones (Hildreth & Moorbath, 1988). In the outer part of the quartz diorite pluton single magma batches cooled rapidly and preserved their primary isotope diversity, whereas the inner part formed from single magma batches that coalesced into a larger homogeneous magma chamber that underwent differentiation. The Mount Whitney batholith and the Wooley Creek batholith show similar assembly processes and geochemical features (Hirt, 2007; Coint et al., 2013).

The Meiwu granodiorite was assembled from small magma pulses that never formed a large pluton-size chamber and thus preserved their source signature. The MME and TE provide evidence for mingling of distinctive magma compositions. The Meiwu biotite granite has a small area (~6 km²) compared with the magmatic lobes (Memeti et al., 2010) and small cycles (Coleman et al., 2012) of the Tuolumne batholith and the Pedrobernardo pluton (Bea, 2010); this represents a smaller, single magma chamber. Small-scale isotope heterogeneity in the biotite granite also suggests that it was not the product of crystallization of a single magma pulse, but was formed by several small heterogeneous magma pulses.

Implications for the evolution of composite batholiths

Based on the above discussion we can see that the spatial and temporal history of incremental construction and evolution of composite batholiths in different locations and tectonic settings does not follow a consistent pattern and is more complex than previously thought. There is no single process or end-member model that can account for the chemical diversity of composite batholiths. Such diversity can be related to heterogeneous source materials, variable melting conditions (e.g. temperature, pressure/depth, and water fugacity) and processes in the source (e.g. magma mixing and MASH processes) or during later evolution (e.g. assimilation and fractional crystallization). The distinct chemical trends among the various units of the Meiwu batholith and of other batholiths support the view that the magma source has a primary control on the chemical diversity of single intrusive units at pluton or batholith scale (e.g. Coleman et al., 2012). During the construction of batholiths successive magma pulses may cool rapidly and preserve their source chemical signatures. Multiple magma chambers of varying size could exist in which the signature of the source was modified by magma chamber processes.

The different assembly styles of the various units of the Meiwu batholith and other batholiths may result from the contrasting thermal evolution of successive magma pulses, which mainly depends on their emplacement and crystallization rates and their capacity to amalgamate (Hirt, 2007; Annen, 2009, 2011). If incrementally assembled magma pluses cool rapidly, they can solidify before the injection of the next pulse and preserve their primary composition. If, however, successive intrusion pulses cool slowly, they would maintain their temperature above the solidus for longer, allowing mixing of new magma batches to occur in a more homogeneous magma chamber.

Mafic magmas have higher temperatures and lower viscosity than felsic magmas; thus the former have higher migration and aggregation capabilities than the latter (Petford et al., 2000; Clemens et al., 2009). We propose that the Meiwu quartz diorites were generated in the deep crust by mixing between hydrous basaltic magma and crustally derived magma. Such hydrous hybrid mafic magma pulses would tend to amalgamate as a single large homogeneous magma chamber.

Zircons from the border and interior facies of the Meiwu quartz diorite provide an insight into the complex process of pluton amalgamation. The fine-grained quartz diorite at the border of the pluton is characterized by relatively higher Ti/Zr, Th/U and Eu/Eu*, lower Hf and Yd/Gd, and variable Hf isotope compositions in zircon, whereas the medium-granited quartz diorite in the interior has relatively lower Ti/Zr, Th/U and Eu/Eu*, higher Hf and Yd/Gd, and homogeneous Hf isotope compositions in zircon. These data indicate that initially isolated, isotopically heterogeneous, magma batches can progressively amalgamate to form a large convecting homogeneous magma chamber. Similar processes have also been reported for volcanic rocks from the NE German Basin (Pietranik et al., 2013) and from the Snake River Plain–Yellowstone Plateau (Wotzlaw et al., 2014). Thus, zircon trace element and isotope compositions can also provide an excellent tool for tracing how magma chambers evolve.

Clearly, granitic magma chambers exist. Annen (2009) concluded that long-term average magma emplacement rates of ~10⁻⁴ km³ a⁻¹ are too slow to form a large-volume granitic magma chamber; a minimum transient emplacement rate of ~10⁻³ km³ a⁻¹ is required to form a granitic magma chamber with a diameter of 4–8 km. Gelman et al. (2013) proposed a critical value [(<5–8) x 10⁻³ km³ a⁻¹] for the formation of large magma reservoirs. Our own estimate (~10⁻⁴ km³ a⁻¹) for the Meiwu batholith is an order of magnitude lower than the likely value (~10⁻³ km³ a⁻¹) for the formation of a stable silicic magma chamber. However, LA-ICP-MS zircon dating cannot resolve the emplacement rates for single units of the Meiwu batholith. For better understanding of the assembly and evolution history of the batholith, high-precision geochronology and thermal modeling are needed.

Tectonic implications

Several geodynamic models have been proposed to account for the Middle Triassic magmatism in the West
Fig. 17. Schematic illustration of the sequential evolution of the assembly of the Meiwu composite batholith. (a) The quartz diorite was formed by partial melting of an enriched lithospheric mantle source, followed by variable degrees of hybridization with crustal melts in a deep crustal hot zone. The initially heterogeneous hybrid magmas then intruded into the upper crust and coalesced to form a large single magma chamber. (b) The granodiorite was constructed incrementally from a number of discrete melt batches that were generated by the melting of mafic lower crust under variable water fugacity. (c) Hybrid magmas similar to the quartz diorites were injected into the host granodioritic magma, producing the MME. Melts derived from the thickened mafic lower crust were subsequently intruded into the host granodioritic magma and disaggregated to form the TE. (d) The biotite granite consists of successive amalgamations of discrete magma batches formed by dehydration melting of mafic lower crust. These magma pulses coalesced into a small single magma chamber. (See text for detailed explanation.)
Qinling orogen: (1) an active continental margin related to subduction of the A’nimaque ocean slab (Jin et al., 2005; Meng et al., 2005; Guo et al., 2012; Li et al., 2013, 2014); (2) an early stage post-collisional setting induced by lithosphere delamination (Zhang et al., 2008); (3) slab break-off (Zhang et al., 2006; Luo et al., 2012a, 2012b). Yin & Zhang (1998) and Chen et al. (2010) suggested that a regional angular unconformity in the East Kunlun between the Late Permian Gequ Formation and the underlying Middle Permian Shuweimenke Formation represents the disappearance of the A’nimaque Ocean and subsequent continental collision event. The Gequ Formation basal molasse has a biostratigraphic age of Changxingian (254–251 Ma), indicating that the onset of collision occurred in the late Permian (Chen et al., 2010). Thus, we consider that the Middle Triassic magmatism occurred in an early stage, post-collision setting.

Both slab break-off and lithosphere delamination can provide the necessary heat to trigger partial melting of the overlying lithosphere to generate post-collisional magmatism (Bonin, 2004). Generally, lithospheric delamination results in a broad region of magmatism and uplift above the delaminated region (e.g. Lastowka et al., 2001), whereas slab break-off produces a narrow, linear zone of magmatism and uplift parallel to the suture zone (e.g. Rogers et al., 2002). The plutons in the central and western parts of the West Qinling orogen exhibit a linear NW–SE trend (Fig. 1b), supporting the slab break-off model. The abrupt change from Middle Triassic shallow marine sedimentary facies to Upper Triassic terrestrial facies and the absence of the upper part of the Middle Triassic (Ladinian) within the NW of the West Qinling belt may indicate that rapid uplift and extensive erosion occurred during the Middle Triassic (Yin et al., 1992; Meng et al., 2005).

CONCLUSIONS: AN INTEGRATED TECTONIC AND PETROGENETIC MODEL FOR BATHOLITH CONSTRUCTION

The Meiwu batholith is composed of quartz diorite, granodiorite, and biotite granite, with abundant mafic enclaves and minor tonalitic enclaves within the granodiorite. The magma crystallization age is ~240–245 Ma. The various units of the batholith exhibit distinct geochemical fingerprints, indicating that it was constructed from repeated magma pulses over a protracted period, rather than by crystallization of a single large batch of magma. The geochemical signatures of each unit of the batholith were mainly inherited from the parental magma sources. Although fractional crystallization, wall-rock assimilation, magma mixing and mingling may occur during magma ascent and at the emplacement level, these processes play only a subordinate role in the modification of the primary chemical variations.

Here, an integrated tectonic and petrogenetic model for the construction and evolution of the Meiwu batholith is proposed as follows.

1. During the Middle Triassic, detachment of the subducted A’nimaque oceanic slab caused hot asthenospheric mantle to upwell through the resultant slab window, and promoted partial melting of underlying enriched lithospheric mantle to generate hydrous basalts. These high-T hydrous basaltic magmas were emplaced into the lower crust and provided the necessary heat and volatiles to induce crustal melting, forming a deep crustal hot zone (Fig. 17a).

2. Mixing between residual evolved basaltic magma and crustal melts generated hybrid magmas; these hybrid magma batches then intruded the upper crust and coalesced as a single magma body to form the quartz diorite pluton. The borders of the quartz diorite cooled rapidly and preserved the primary isotopic diversity, whereas the interior developed into a larger, homogeneous magma chamber that underwent internal differentiation and crystallized more slowly (Fig. 17a).

3. Volatiles released from the hydrous basalts induced fluid-present partial melting of the lower crust. The crystal melts mixed with minor amounts of evolved mafic magma and were successively intruded into the upper crust to form the granodiorite (Fig. 17b).

4. Magmas similar to the quartz diorite were injected into the granodioritic magma, undergoing local mechanical mingling and diffusive exchange with the host magma, producing the various types of MME (Fig. 17c). Melts derived from partial melting of the thickened mafic lower crust also intruded the granodioritic magma and disaggregated to form the TE (Fig. 17c).

5. Following a general trend of progressively decreasing mantle source contribution, dehydration melting of the lower crust produced the biotite granite through heat transfer from the cooling mafic magmas. These magma pulses intruded into the upper crust and coalesced into a small single magma chamber (Fig. 17d).

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SUPPLEMENTARY DATA

Supplementary data for this paper are available at Journal of Petrology online.

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