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**Origin of mafic microgranular enclaves from the Late Triassic
Cuojaoma granitic pluton: implications for the tectonic evolution of
the Yidun arc belt, eastern Tibetan Plateau**

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Abstract: *In-situ* U-Pb ages and Hf-O isotopes of zircon are frequently used to study the petrogenesis of mafic microgranular enclaves (MMEs) in granitoids. However, whether zircons in MMEs have recorded the primary isotopes of their host MMEs is still questionable. In this study, MMEs and host granites from the Cuojaoma granitic pluton in the Triassic Yidun arc belt, eastern Tibetan Plateau, are studied. LA-ICP-MS zircon U-Pb dating of studied MMEs yields an age of 213 ± 2 Ma. Model calculations indicate that the whole-rock composition of the MMEs was different from the melt composition from which the zircon grains grew. The MMEs crystallized during magma mixing based on analyses of zircon trace element and Zr-Hf isotopes. None of zircons in this study have recorded the primary isotope signature of the mafic endmember of MMEs because they are different from the host granite in terms of zircon morphology, ϵ_{Hf} values and ΔQFM . The Zr isotopic data demonstrate open-system magmatic crystallization for the host granite and that magma mixing for MMEs play key role in controlling Zr isotopic variations. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of An-rich plagioclase crystals in groundmass from MMEs range from 0.7027 to 0.7058, which are significantly lower than host granites (0.7080 – 0.7089). The lowest initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7027 – 0.7028) are also lower than arc volcanic rocks and similar to a depleted mantle. Combined with previous studies, we propose that MMEs were generated during post-collision extension by mixing of mafic magma of Xiaxiaoliu basalt and felsic magma from host granite.

Keywords: MMEs, in situ Sr-Hf-Zr isotope, Yidun arc belt, Eastern Tibetan Plateau

1. Introduction

Mafic microgranular enclaves (MMEs) are common in granitic plutons (Didier and Barbarin, 1991). They have the potential to provide significant information on the nature of the source rocks, the mechanism and geodynamics of generating granitic melts, and the interaction between continental-derived and mantle-derived magmas (Barbarin, 2005; Kemp et al., 2007; Jiang et al., 2013; Wu et al., 2020). Several mechanisms have been suggested for their petrogenesis, such as restite separation (Chappell and White, 1992), wall-rock assimilation (Shellnutt et al., 2010), cumulates at the early stage of magma fractionation (Shellnutt et al., 2010), and magma mixing (Vernon, 1984). The primary sources of the MMEs and their host granitoids are the

key issues for distinguishing between these models. Compared with whole-rock Rb-Sr and Sm-Nd isotopic compositions, which are easily equilibrated between the MMEs and their host granitoids (Barbarin, 2005; Yang et al., 2007), *in-situ* mineral analysis can retain their initial isotopic compositions. Thus, most recent studies have focused on *in-situ* mineral chemistry, principally using zircon, alkali and plagioclase feldspars, apatite, titanite, monazite and perovskite (Kemp et al., 2007; Paton et al., 2007; Gregory et al., 2009; Yang et al., 2008, 2009) to trace the origin and petrogenetic processes of MMEs. Zircon has been extensively analyzed to determine their U-Pb ages and Hf-O isotopes (Griffin et al., 2002; Kinny and Maas, 2003; Yang et al., 2007; Wu et al., 2020). However, mantle-derived mafic magma is initially Zr-unsaturated, and the zircons from MMEs will not record the primary isotopes of the magma if magma mixing or crustal contamination occurred before zircon crystallization (Wu et al., 2020). Thus, it is still not well constrained whether the zircons in MMEs actually record the primary isotope signature of mantle-derived magma, and the source region and petrogenesis of MMEs cannot be distinguished by zircon Hf-O isotopes. Recent developments of bulk and high precision *in-situ* analytical methods have permitted studies on mass-dependent Zr isotope fractionation in terrestrial samples (Inglis et al., 2018) and on Zr isotopic composition of zircons (Zhang et al., 2019). Large mass-dependent Zr isotopic variations exist at both mineral (>5‰) and bulk-rock scales (ca. 0.6‰) (Guo et al., 2020, Ibañez-Mejia and Tissot, 2019, Inglis et al., 2018, Zhang et al., 2019), suggesting stable Zr isotopes as a potential tracer of igneous processes. Furthermore, An-rich plagioclase with oscillatory zoning is also common in MMEs. Such plagioclase usually crystallises early in a mafic magma chamber, and so may have recorded the primary isotopic composition of the mantle-derived magma (Perugini et al., 2005; Staby et al., 2008; Wu et al., 2020). The

newly developed fs-LA-MC-ICP-MS method has been successfully used to investigate *in-situ* Sr isotopes of An-rich plagioclase crystals in MMEs (Zhang et al., 2018). Hence, an integrated study of *in-situ* Zr isotopes of zircon and *in-situ* Sr An-rich plagioclase may shed new light on the petrogenesis of MMEs.

The Late Triassic Garzê-Daocheng granitic belt contains several large granitic batholiths and lies in the eastern part of the Yidun arc belt (Fig. 1). This granitic belt therefore provides an ideal window for exploring the evolution of the Garzê-Litang Paleo-Tethys Ocean (Dong et al., 2022; Zhan et al., 2021). However, a range of tectonomagmatic models have been suggested including an arc setting, related to eastward subduction of the Jinshajiang Ocean (Reid et al., 2005; Roger et al., 2010) or westward subduction of the Garzê-Litang Ocean (Hou et al., 2003; Wang et al., 2018; Dong et al., 2022), and a post-orogenic extensional setting (Peng et al., 2014; Wu et al., 2017b; Zhan et al., 2021). The source region and petrogenesis of the granitoids also remain ambiguous. The enriched isotopic compositions of these rocks have led to interpretations that they were generated either by partial melting of ancient crustal rocks with/without a minor contribution from mantle-derived magmas (Reid et al., 2005; Peng et al., 2014; Wu et al., 2017b; Dong et al., 2022; Xie et al., 2022), large scale mixing between mantle and crustal-derived magmas (Hou et al., 2001, 2003), or were produced by crystal fractionation of mantle-derived mafic magmas with (Dong et al., 2022) or without assimilation (Zhan et al., 2021), which hampers our understanding of the evolution of the whole Yidun arc belt.

The Cuojaoma batholith is one of the largest Triassic granitic plutons (~1800 km²) in the northern Yidun arc belt, and provides an opportunity to constrain the evolutionary history of the Yidun arc belt. It is generally accepted that mantle-derived mafic magma had played a crucial role in the formation of the Cuojaoma and other

granitoids in Garzê-Daocheng granitic belt (Peng et al., 2014; Wu et al., 2017b; Zhan et al., 2021; Xie et al., 2022). However, the nature of the mantle-derived mafic magma and the details of the interaction between mafic and felsic magmas are poorly constrained. Various-shaped mafic microgranular enclaves (MMEs) are widespread within the Cuojaoma batholith (Wu et al., 2020; Zhan et al., 2021; Xie et al., 2022). Thus, a study of the Cuojaoma batholith, especially on the petrogenesis of the MMEs, will help us not only to understand the source of Cuojaoma batholith and decipher its geodynamic setting (Barbarin and Didier, 1991; Barbarin, 2005; Yang et al., 2015), but will also shed new light on the petrogenesis of MMEs and their host granites. In this contribution, we present new zircon U-Pb ages, *in-situ* Hf-Zr isotopes of zircon and *in-situ* Sr isotopes of An-rich plagioclase, along with mineral and whole-rock geochemical data of the MMEs and host granites from the Cuojaoma granitic pluton in the northern part of YAB (Fig. 2), with the aim of understanding the petrogenesis of the MMEs and host granites from the Cuojaoma granitic pluton, and its regional geodynamic setting.

2. Geological setting and Petrography

The Yidun arc belt (YAB), which is located in the northeastern Sanjiang orogeny, eastern Tibetan Plateau, lies between the Qiangtang terrane to the west, Songpan-Garzê Fold Belt to the northeast and Yangtze Craton to the southeast. It is bounded to the west by the Jinshajiang suture belt and to the east by the Garzê-Litang suture belt. The Garzê-Litang Ocean opened in the early (Yan et al., 2005) or late Permian (Song et al., 2004; Xiao et al., 2008) as a result of either a back arc basin caused by eastward subduction along the Jinshajiang suture (Roger et al., 2008, 2010) or a plume beneath the western margin of Yangtze Craton (Xiao et al., 2008). The

Jinshajiang Ocean began to subduct during the late Triassic to form the Yindun arc belt (Hou et al., 2001, 2004).

The YAB can be divided into two domains: the western and eastern Yindun arc belt (Reid et al., 2005). The western YAB is also known as Zhongza Massif and is composed of carbonate-rich Paleozoic metasedimentary rock intercalated with mafic volcanic rocks with a Neoproterozoic basement dominated by granitic gneisses and metavolcanic rocks (BGMRS, 1991; Wu et al., 2017b). The eastern YAB is composed of N-S trending Late Triassic sedimentary and igneous rocks, and Late Triassic granitoids resulting from the westward subduction of the Garzê-Litang Oceanic crust (Leng et al., 2012; Wang et al., 2014). Recently, it was proposed that an ancient basement existed beneath the eastern YAB (Reid et al., 2007; Peng et al., 2014; Wu et al., 2017b). There are widespread Late Triassic bimodal volcanic suites and arc-type volcanic rocks with large Late Triassic granitic batholiths (He et al., 2013; Wu et al., 2017b) and multiple sulfide deposits (Hou et al., 2003; Wang et al., 2013) in the YAB.

Magmas in the YAB were principally emplaced during the Late Triassic and Cretaceous. The Late Triassic intrusions are mainly arc-related calc-alkaline I-type granites with negative to positive zircon $\epsilon_{\text{Hf}}(t)$ and negative bulk-rock $\epsilon_{\text{Nd}}(t)$ values which formed between 237 Ma and 206 Ma (He et al., 2013; Hou et al., 2001; Reid et al., 2007; Wu et al., 2017b). The Cretaceous granitoids with variable negative $\epsilon_{\text{Nd}}(t)$ and $\epsilon_{\text{Hf}}(t)$ values were formed between 107 Ma and 77 Ma and most show A-type granite affinities (Li et al., 2017; Gao et al., 2017). The distribution of the Late Triassic and Late Cretaceous magmatic rocks and their associated deposits are controlled by the Late Triassic NW-trending and NE-trending faults, which are regarded as the main migration channel for magma and ore-forming fluid (Li et al.,

2017).

The Cuojiaoma pluton, covering an area of ~4000 km², is one of the largest granitoids in the YAB, and is located approximately 50 km south of the Garzê city. It intrudes the Upper Triassic Lamaya (T₃l_m) and Lanashan (T₃l) formations. The rock types are mainly porphyritic monzogranite with a small amount of granodiorite (Fig. 3). The porphyritic monzogranite contains phenocrysts (10 – 20 vol.%) mainly of perthitic alkali feldspar, with minor quartz and plagioclase. The groundmass comprises quartz (20 – 35 vol.%), plagioclase (25 – 35 vol.%), potassium feldspar (20 – 30 vol.%), biotite and amphibole (total 5 – 10 vol.%) with apatite and zircon as accessory phases (Wu et al., 2017b). Mafic microgranular enclaves (MMEs) are widespread in the Garzê-Daoheng granitic belt (Fig. 3). Typically, MMEs are scattered in the felsic plutons (Fig. 3a), but locally they are densely distributed (Fig. 3b). They range in length from 2cm to 80 cm and display lenticular or angular shapes, and contain noticeably finer-grained (minerals ranging in size from 0.1 to 0.5 mm) than the host felsic rocks in the field. MMEs are variably elongated without solid-state deformation (Fig. 3), suggesting enclaves and host rocks represent compositionally distinct magmas (Vernon, 1991; Kumar et al., 2004). This is supported by the alignment of hornblende, biotite and elongate laths of plagioclase (Fig. 4a) without signs of solid-state deformation. This is because such a preferred dimensional orientation of crystals is likely to have formed where enough melt was present to allow extensive rotation of crystals (Chen et al., 2009). The MMEs are composed of gabbroic diorite, diorite and monzodiorite, but are dominated by diorite. They commonly have sharp contacts with the host granites, however chilled margins (e.g., decrease in grain size) on the MMEs have not been observed. The MMEs from the Cuojiaoma pluton consist mainly of plagioclase and amphibole, with minor perthitic

alkali feldspar, quartz and biotite (Fig. 4), and accessory magnetite, sphene, apatite and zircon. Textures vary from equigranular, fine-grained to porphyric. Plagioclase laths in the groundmass show compositional zoning and typical magmatic textures (Fig. 4a). Normal and oscillatory zoning in plagioclase is widely developed and can be divided into two groups based on the in-situ Sr isotopes (Fig. 4a-c). Type 1 has homogeneous initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7049 – 0.7057); type 2 has dark cores with the lowest initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7027 – 0.7028) (Fig. 4b, c). The plagioclase has suffered minor argillization and sericitization. Amphibole occurs as subhedral grains whereas quartz and alkali feldspar are interstitial.

3. Analytical Methods

Three MME samples were analyzed for whole rock major and trace element compositions. Whole-rock major element oxides (wt.%) are measured on fused glass discs, using an X-ray fluorescence at the Hubei Geological Research Laboratory (HGRL). Whole-rock trace elements are measured by an Agilent 7700 inductively coupled plasma-mass spectrometer (ICP-MS) at the State Key Laboratory of Geological Processes and Mineral Resources (GPMR), China University of Geosciences, Wuhan. The MME sample CJ-3 is selected for zircon U-Pb dating and Hf isotope analysis. *In-situ* U-Th-Pb isotopes of zircons were measured by laser-ablation (LA)-ICP-MS using an Agilent 6500a ICP-MS system attached to a GeoLas 2005 excimer ArF laser ablation system at the GPMR. *In-situ* zircon Hf isotope measurements were conducted at the GPMR, by using a Neptune Plus MC-ICP-MS (Thermo Fisher Scientific, Germany) in combination with a Geo-las 2005 excimer ArF laser ablation system (Lambda Physik, Göttingen, Germany). Mineral geochemistry of plagioclase was determined by LA-ICP-MS at GPMR. The

laser ablation system is a GeoLas HD excimer ArF laser ablation system (Coherent, Göttingen, Germany), which is connected to an Agilent 7500a ICP-MS. Sr isotope ratios of feldspar were dated by a Neptune Plus MC-ICP-MS (Thermo Fisher Scientific, Germany) in combination with an NWR Femto^{UC} femtosecond system (New wave Research, Fremont, CA, U.S.A.) at GPMR. Zr isotope ratios of zircons were also analyzed by the Neptune Plus MC-ICP-MS in combination with a Geolas HD excimer ArF laser ablation system (Coherent, Göttingen, Germany) at GPMR. Detailed analytical methods of whole-rock major and trace element compositions, LA-ICP-MS zircon U-Pb dating, mineral geochemistry of plagioclase, *in-situ* zircon Zr-Hf isotopes, and *in-situ* Sr isotope of plagioclase are given in the supplementary material ([Appendix 1](#)).

4. Results

4.1 Geochronology

The Cuojiaoma granite has yielded a range of U-Pb zircon ages: 215.1 ± 2.0 Ma to 224.7 ± 2.0 Ma ([Weislogel, 2008](#)), 219 ± 1 Ma ([Wu et al., 2017b](#)), 217.1 ± 1.5 Ma ([Tian et al., 2019](#)), 236 ± 1.9 Ma ([Wang et al., 2017](#)), 214 ± 2.0 - 219 ± 2.0 Ma ([Zhan et al., 2021](#)), and 216.7 ± 2.2 Ma – 221.8 ± 1.4 Ma ([Xie et al., 2022](#)). This geochronological data is listed in [Table 1](#), and U-Pb Concordia diagrams are shown in [Fig. 5](#). The representative zircons CL images are shown in [Fig. S1](#). For this study we selected one MME from Cuojiaoma (CJM-3) for zircon dating ([Fig. 2b](#)). Zircons from Cuojiaoma granitic rock (CJM-6, [Wu et al., 2017](#)) are prismatic and many have well-developed oscillatory zoning ([Fig. S1b](#)). However, the zircons from MMEs (CJM-3) are dark anhedral - subhedral granular crystals ([Fig. S1a](#)). Oscillatory zoning is absent or weak in these crystals which are 50 – 120 μm in length with length/width ratios of 1 – 1.8.

Th/U ratios range from 0.53 to 3.00, indicating a magmatic origin (Wu and Zheng, 2004). Zircon in CJM-3 is divided into two types: type 1 zircons are represented by dark anhedral - subhedral granular with obvious and relatively light planar or fan-shaped crystalline bands in the core area (e.g., spot 1, 2, 3, 7, 8, 12 and 15) or are homogeneous with no difference from core to rim (e.g., spot 4, 5, 6, 9, 10, 17 and 18); type 2 zircons show dark anhedral granular with light inherited core (e.g., spot 11, 13, 14, 16 and 20).

Based on the Hf isotopes, Type 1 zircons have relatively lower Hf isotopes with ϵ_{Hf} values ranging from -1.8 to +1.3 (mean ϵ_{Hf} value of -0.2) than those of Type 2 zircons (ϵ_{Hf} values of +1.0 to +2.1 and mean value of +1.5). Compared with Type 2 zircons, Type 1 zircons have $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios of 4 to 14 (average $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratio of 7.9, Ballard et al., 2002) with ΔQFM ranging from -2.3 to +1.6 and mean value of -0.3. Type 2 zircons have $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios ranging from 9 to 50 (average $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratio of 29.7) and ΔQFM of -0.8 to -0.7 (mean value of -0.75). However, the distinct differences in trace elements and Ti-in-zircon temperature (T_{zircon}) between both type zircons were not discovered.

Twenty zircons from sample CJM-3 were dated and fourteen sites of type 1 zircons record $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 212 Ma to 214 Ma, and define a weighted mean age of 213 ± 2 Ma ($N = 14$, $\text{MSWD} = 0.05$) (Fig. 5c), which is interpreted as the crystallization age of the studied MMEs. This result is consistent with the age from Zhan et al. (2021). The other six grains of type 2 zircons yielded concordant ages ranging from 236 Ma to 251 Ma and are interpreted as inherited zircons.

4.2 Whole-rock geochemistry

Three MMEs samples were selected for whole-rock major and trace element

analysis and the results are presented in Table 2. The MMEs exhibit low – intermediate SiO_2 contents of 53.8 – 60.3 wt.%, and classify as gabbroic diorite, monzodiorite and diorite on a SiO_2 vs. $\text{Na}_2\text{O}+\text{K}_2\text{O}$ diagram (Fig. 6a). In contrast, the host granites are characterized by high SiO_2 contents of 71.2 – 73.4 wt.% (Fig. 6a). MMEs have variable MgO contents (2.6 – 9.8 wt.%), and high FeO^{T} (5.7 – 7.9 wt.%), compared to the host granites which are characterized by lower MgO (0.6 – 1.1 wt.%) and FeO^{T} (2.2 – 3.4 wt.%) contents (Table 2). The A/CNK of the host granites and MMEs range from 1.04 to 1.08 and from 0.60 to 1.05 respectively, showing that the granite is metaluminous-peraluminous and the MME are metaluminous to slightly peraluminous (Fig. 6b). The host granites have Na_2O of 2.6 – 3.1 wt.%, K_2O of 3.2 – 3.8 wt.%, and the MMEs possess a wider range of Na_2O (1.9 – 3.8 wt.%) contents and lower K_2O (1.3 – 2.3 wt.%). The granites and the MME are part of the medium- to high-K calc-alkaline series (Fig. 6c).

The light rare earth elements (LREEs), are enriched in host granitic rocks and MMEs relative to the heavy rare earth elements (HREEs) (Fig. 7a). The host rocks have steeper REE patterns ($(\text{La}/\text{Yb})_{\text{N}} = 10.3 – 14.5$) with pronounced negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.28 – 0.33$), while the MMEs have $(\text{La}/\text{Yb})_{\text{N}}$, ranging from 3.5 to 8.8 with variable Eu/Eu^* ratios of 0.36 to 1.00 (Fig. 7a). The host granites and MMEs have similar normalized trace element patterns with enrichment in large lithophile elements (LILEs. e.g., Rb, Th, and U) and depletion in high field strength elements (HFSEs, e.g., Nb) (Fig. 7b). The host granite has negative Ba, Nb Sr and Eu anomalies, while the MMEs have similar Ba and Nb troughs to the host granite, and either no, or negative, Sr and Eu anomalies (Fig. 7b).

4.3 Trace elements and Lu-Hf isotopes of zircon

Trace elements in zircon are shown in Table S3. All Late Triassic zircons from MME (CJM-3) have positive chondrite-normalized Ce anomalies, marked negative to positive Eu anomalies (δEu ranges from 0.51 to 1.92), and enrichments in HREEs (Fig. S1). These trace element patterns, and ΣREE abundances of 476 to 2690 ppm are typical of igneous zircons ($\Sigma\text{REE} = 250 - 5000$ ppm) (Table S3). The $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios of the zircons were calculated using the equations of Ballard et al. (2002) and yielded ratios between 4 and 12, with an average of 10 ($n=10$ with the exclusion of CJM-3-10 and 12; Table S1).

Zircon Lu-Hf isotopes were also analyzed on the same CL zircon domains that were dated by U-Pb, and the results are presented in Fig. 8a and Table 3. The initial Hf isotopic ratios are calculated at emplacement age using the $^{176}\text{Lu}/^{177}\text{Hf}$ decay constant published by Söderlund et al. (2004). The $^{176}\text{Lu}/^{177}\text{Hf}$ values of the Late Triassic zircons used in the age calculation are low, within the range of 0.0012-0.0025 (<0.003), indicating there was very little radiogenic ingrowth of ^{176}Hf . Their $^{176}\text{Hf}/^{177}\text{Hf}$ ratios range between 0.282608 and 0.282695. Eight spot analyses were obtained and yielded $\epsilon\text{Hf}(t)$ values between -1.8 and +1.3. Whereas the $^{176}\text{Lu}/^{177}\text{Hf}$ values of the 236 Ma – 251 Ma zircons range from 0.0010 to 0.0033, with relatively higher $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282665 to 0.282711 and $\epsilon\text{Hf}(t)$ values of +1.0 to +2.1 (Fig. 8a).

4.4 The composition of plagioclase and its in-situ Sr isotopes

The compositions of plagioclase and its *in-situ* Sr isotopes are shown in Table 4 and 5, respectively. Six plagioclase grains from sample CJM-11 were analyzed, and they are andesine labradorite and bytownite in composition with An values ranging from

41 to 73 (Fig. 8b). The Sr concentrations have a negative correlation with An contents (Fig. 8f).

An abundance of plagioclase megacrysts are found in the MMEs (Fig. 3), which provides good evidence of physical mixing between mafic and felsic magmas (Yang et al., 2015). Plagioclase megacrysts from both MMEs and host rocks show similar petrographic features, indicating that the plagioclase megacrysts from MMEs may be xenocrysts. Therefore, plagioclase in groundmass from MMEs have been analysed as they may better constrain the magma source of MMEs. Plagioclases in groundmass from enclaves have an irregular, calcium-rich core (An₆₀₋₇₃), surrounded by relatively Ab-rich rim (An₄₁₋₄₆), and the compositional change between the two is abrupt.

In sample CJM-11 from Cuojiaoma granitic pluton, two types of plagioclase crystals were recognized (Fig. 4). Type 1 has dark cores with light rims and show similar initial ⁸⁷Sr/⁸⁶Sr ratios from the cores to the rims (0.7052-0.7057, Table 5). Type 2 plagioclase has dark cores with light rims, however, the dark cores have much lower initial ⁸⁷Sr/⁸⁶Sr ratios (0.7027-0.7028, Table 5) than the rims (0.7049-0.7057, Table 5). In general, the initial ⁸⁷Sr/⁸⁶Sr ratios of plagioclase from Cuojiaoma sample CJM-11 (0.7027-0.7057) are much lower than those of the host granites (0.7080 – 0.7089) (Wu et al., 2017b).

4.5 Zr isotopes of zircons

Zircons from the host granite (sample CJM-6) and MME (sample CJM-3) were measured for Zr stable isotopes and the results are listed in Table 6 and presented on Fig. 9. Twelve zircons from sample CJM-3 were analyzed. The cores of five zircons have $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values ranging from $-0.31 \pm 0.03\text{‰}$ to $+0.1 \pm 0.03\text{‰}$, whereas those of the zircon rims range from $-0.43 \pm 0.03\text{‰}$ to $+0.41 \pm 0.03\text{‰}$ (Fig. 9). Sixteen

zircons from sample CJM-6 from the host granite of the Cuojaoma pluton were analyzed. Six cores have $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values ranging from $-0.25 \pm 0.05\text{‰}$ to $+0.07 \pm 0.04\text{‰}$, and nine rims have $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values ranging from $+0.03 \pm 0.04\text{‰}$ to $+0.15 \pm 0.04\text{‰}$, respectively.

5. Discussions

5.1 Formation of zonal plagioclase texture in MMEs

The zonal texture of plagioclases records the physical and chemical conditions of magma evolution during crystallization (Ghaffari et al., 2015; Alves and Janasi, 2020). As shown in Fig. 4a, plagioclase in groundmass from MMEs always has an embayed calcium-rich core (An_{60-73}). The core is in turn abruptly mantled by an idiomorphic, sodium-rich plagioclase (An_{41-46}) (Fig. 4a-c). This indicates that the geological environment in which MMEs formed has undergone significant changes since the crystallization of plagioclases. The reason for this change either that the plagioclase crystals have been transferred into higher temperature or more basic magma (Li et al., 1999); or that mafic magma with plagioclase crystals has entered into the host granitic magma, resulting in a sudden decrease in pressure and release of large amounts of water and volatile components from the magma (Castro, 2001; Landi et al., 2004). Considering that the core of plagioclase with An_{60-73} is the product of crystallization of mantle source magma at high temperature, it is unlikely that the plagioclase crystals have been transferred into a higher temperature, or more basic magma. Furthermore, during the Late Triassic, the tectonic setting was post-collisional setting (see section 5.4). Thus, the second explanation is more convincing.

Our explanation to the MMEs in this study is as follows. At first stage, the mafic end-member magma evolved and crystallized. Then the mantle-derived mafic magma

ascended rapidly along the tectonic channel and invaded the shallow crust. Simultaneously, the calcium-rich plagioclases were resorbed to form irregular resorbed zones due to a sudden decrease in pressure. The mafic magma was injected into the host granitic magma at shallow crustal level, and they mixed to form the enclave-forming dioritic magma. The previously crystallized plagioclase from the mafic magma was transferred into the enclave-forming dioritic magma. Such a resorbed plagioclase acts as the site of new mineral growth, resulting in the formation of magmatic overgrowth of plagioclase (with An₄₁₋₄₆) from the enclave-forming magma. Such a textural and compositional disequilibrium resulted from rapid crystallization and incomplete magma mixing (Donaire et al., 2005; Chen et al., 2009). This conclusion is also supported by the *in-situ* Sr isotopes of plagioclase that the core and rims of plagioclase has relatively lower and higher initial ⁸⁷Sr/⁸⁶Sr ratios of 0.7027-0.7028 and 0.7052-0.7057, respectively (Fig. 4b, Table 5), suggesting a magma mixing process.

5.2 Petrogenesis of the MMEs

5.2.1 Genetic link between the host granite and its mafic enclaves

Four mechanisms have been proposed for the genesis of MMEs: (1) restites that form by partial melting of source rocks which are then entrained by granitic melt (Chappell and White, 1991); (2) wall rock xenoliths (Maas et al., 1997); (3) cumulates produced by fractional crystallization (FC) of early magmas or cognate melts (Shellnutt et al., 2010); and (4) generated by magma mingling when coeval mafic melts are injected into an intermediate or felsic magma chamber (Vernon, 1991; Barbarin and Didier, 1992).

The MMEs associated with the Cuojiaoma granites are ubiquitous in outcrop and

generally more mafic than the host. Based on the classification diagrams and trace element variations, it appears that the enclaves exhibit wide compositional variations. It is therefore difficult to assess their relationships with their host compositions because of possible element diffusion and chemical equilibration (Eberz and Nicholls, 1990; Holden et al., 1991; Manu Prasanth et al., 2023). Higher concentrations of certain transition metals (e.g. Mn, Nb, Ta, Table S2) in enclaves suggest they are more primitive than the host granites. If diffusion is dominant, the transition metals tend to diffuse faster than the HFSEs, pointing out that diffusion may not be the dominant process that imparts significant compositional variations in the enclaves (Manu Prasanth et al., 2023). Although the compositional gap between mafic and felsic end-members is not obvious (Fig. 10), which indicate that they may derived from a common mafic parental melt. Overall depletion of HFSEs in the enclaves would seem to suggest magma differentiation within an evolving silicic liquid, however, the differences in the trace and rare earth element patterns between enclaves and host granites (Fig. 7) indicate that the precursors were possibly not derived from a similar mafic parental source (Manu Prasanth et al., 2023). The overlapping isotopic ages between enclaves and host granites described above, indicate that the mafic enclaves are unlikely to be formed by xenoliths or melting restites. Zircons from the MMEs and host monzogranites display different Lu-Hf compositions (Fig. 8a and c), and simple fractional crystallization in a closed system would not change the zircon Lu-Hf isotopic composition (Shaw and Flood, 2009). The zircon $\epsilon_{\text{Hf}}(t)$ values (-16.4 to -1.4) in the host granites are distinct from those ($\epsilon_{\text{Hf}}(t) = -1.8$ to +1.3) of the MMEs (Table 3 and S5), again precluding a simple origin by closed-system fractional crystallization (Griffin et al., 2002). Although the MMEs have similar zircon ages (213 Ma) to their host granites (214-224.7 Ma), the observation of reverse and oscillatory zones in

plagioclase, and the fact that some plagioclases have much lower initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the host granites preclude a cogenetic cumulate model (Fig. 4b).

The following lines of evidence indicate that the Late Triassic MMEs from the Cuojaoma granitic pluton were most likely produced by the mixing or mingling of mafic and felsic magmas:

(1) The MMEs are randomly distributed in the host granites and are notably finer-grained and dark-colored than the host rocks (Fig. 3). They are generally ellipsoidal and occasionally lenticular or irregular in shape (Fig. 3) and show sharp contacts with the host granites. Their elliptical and lenticular shapes and different sizes indicate that they were probably produced by magma mingling (Wang et al., 2017). Furthermore, the MMEs often contain feldspar xenocrysts (Fig. 3), which is probably due to the hybridization between the enclave-forming magma and host magma (Jiang et al., 2013).

(2) The MMEs in this study display typical igneous textures and lack cumulate, metamorphic or sedimentary textures (Fig. 4a, d) which is difficult to relate to a residual source origin or wall-rock xenoliths.

(3) Geochemical signatures of the MME and host granite support a magma mixing origin. Characteristic linear or hyperbolic mixing arrays on Figs. 11a-d, lend further support to a significant role for magma mixing in their petrogenesis. Geochemical modeling of batch partial melting, fractional crystallization and magma mixing can be used to investigate the processes which exert an influence on the trace-element variations of cogenetic magmas (Schiano et al., 2010). Partial melting and fractional crystallization would produce a straight line with a slope and a horizontal line, respectively, in plots of two incompatible elements with different partition coefficients (e.g., Th vs. Th/Nd, La vs. La/Sm; Fig. 11f and g). This conclusion is also supported

by the linear correlation in the La vs. Th diagram (Fig. 11h).

(4) The MMEs have variable $Mg^\#$ (Mean value of 42) with enrichment of LREEs (e.g., Rb, K) and depletion of HFSES (e.g., Nb, Ta) and their Nb/Ta ratios range from 6.5 to 12.5 (Table 2 and S2). These are intermediate between those of lower crust (8.3, Rudnick and Gao, 2003) and primitive mantle (17.5; Jochum et al., 1989), indicating that the MMEs were generated by mixing between the depleted mantle-derived and lower crust-derived magmas.

(5) The relative oxidation states of magma inferred from Ce(IV)/Ce(III) have been calculated, and the result show that Cuojiaoma MME (CJM-3) has Ce^{4+}/Ce^{3+} ratios of 4 to 14 (average Ce^{4+}/Ce^{3+} ratio of 7.9, Ballard et al., 2002) with ΔQFM ranging from -2.3 to +1.6 (average ΔQFM -0.3 with CJM-3-10 and 12 excluded, Table S4, Loucks et al., 2020). In contrast the host granites have average Ce^{4+}/Ce^{3+} ratios of 18 (Ballard et al., 2002; Pan et al., 2020) and ΔQFM ranging from -1.3 to +4.6 (average ΔQFM +0.8, Table S4, Loucks et al., 2020). This indicates that the MMEs were probably formed by magma mixing processes.

Thus, we propose that the MMEs represent contemporaneous magmas which were derived from mixing of magmas from two distinct source regions, namely depleted mantle-derived and lower crust-derived magmas.

5.2.2 Origins of zircons in the MMEs

Zircon Hf-O isotopes have been extensively used in studying the origin of MMEs (Griffin et al., 2002; Yang et al., 2007; Zhan et al., 2021), based on the assumption that zircons from MMEs can preserve the primary Hf-O isotopes of the mafic magma and are resistant to subsequent geological processes including magma mixing and assimilation of wall-rock (Yang et al., 2007; Wu et al., 2020). However, in most cases,

zircons from both MMEs and the host granites show similar Hf-O isotope compositions (e.g., [Zi et al., 2012](#); [Jiang et al., 2013](#); [Chen et al., 2016](#); [Wu et al., 2020](#)). Previous studies also point out that if any geological processes like magma mixing or crustal contamination occurred before zircon crystallization, then zircons will not record the primary isotopes of the magma ([Wu et al., 2020](#)). It therefore needs to be evaluated whether the zircons investigated in this study actually record the primary isotope compositions of the host MMEs.

The Zr solubility model by [Watson and Harrison \(1983\)](#) and apparent fractionation index temperature methods by [Kirkland et al. \(2015\)](#) are adopted to evaluate whether the composition of the zircons is in equilibrium with the whole-rock composition of their host MMEs ([Watson and Harrison, 1983](#); [Boehnke et al., 2013](#); [Wu et al., 2020](#)). The result suggests that the required minimum Zr concentration for zircon saturation is 643 ppm if the magma temperature of sample CJM-03 was between 800 °C and 900 °C. The measured Zr concentration from the bulk rock of sample CJM-03 is 174 ppm. Thus, the zircons are unlikely to have crystallized directly in the enclave. Additionally, the calculated fractionation index temperatures of the MMEs (950 – 1650 °C) are apparently higher than their zircon saturation temperatures of 677 to 1377 °C ([Fig. 8d](#); [Table S4](#)), suggesting that the whole rock composition was different from the melt composition in which the zircon grains grew.

Generally, the zircons in MMEs can be formed in three ways: (1) crystallization in an evolved mafic magma before magma mixing ([Yang et al., 2007](#); [Liu et al., 2013](#)); (2) crystallization in the mixed magma ([Wu et al., 2020](#)) and (3) entrainment of earlier zircons in the granitic magma chamber ([Wu et al., 2020](#)). Zircons from the MMEs have similar or slightly higher $\varepsilon_{\text{Hf}}(t)$ values (-10.8 to +1.3, [Table S5](#)) than their host granites (-16.4 to -1.4, [Table S5](#)) ([Fig. 8c](#)), indicating that the zircons did not

crystallize in the evolved mafic magma before magma mixing. In the second scenario, a mafic magma mixed with a volume of granitic magma becomes Zr-saturated which enables growth of zircon. In this case, the zircons will have similar or slightly more depleted isotopic compositions than those of the host granite, and they may still differ from the host granite in terms of zircon morphology and crystallization temperature (Wu et al., 2020). Zircons of MMEs from the Cuojaoma granitic pluton have similar or slightly higher $\epsilon_{\text{Hf}}(t)$ values (Fig. 8a), which suggests that they couldn't have simply crystallized in the mafic magma under a closed system. As revealed by *in-situ* Sr isotopes of the An-rich plagioclase, the mafic magma may be derived from depleted mantle and would crystallize zircons with depleted mantle-like Hf isotopes. Zircons from host granite are basically euhedral prismatic with well-developed oscillatory zoning, but the zircons from MMEs display dark anhedral - subhedral granular.

In the third case, zircons will be indistinguishable from those of the host granite. Several plagioclase megacrysts have been observed in the MMEs (Fig. 3), providing evidence for the physical mixing between mafic and felsic magmas. Numerous studies have shown significant differences in distribution patterns of the rare earth elements between the mantle-derived and crustal-derived magmatic zircons (e.g., Siebel et al., 2009; Melnik et al., 2022). Mantle-derived magmatic zircons are featured by low LREEs abundances and show no or weak negative Eu anomalies, whereas crustal-derived magmatic zircons display pronounced negative Eu anomalies with high LREEs because of the fractional crystallization of plagioclase. Compared with zircons from the host granites, Type 1 zircons from MMEs have higher Th/U ratios (0.41 – 2.47) and are smaller in grain size than those of the host granite (Fig. 8e). They also have lower abundances of LREEs, and show obvious enrichment in Ce and

depletion in Pr with no or weak Eu negative anomalies (0.51-1.92, the values of spot 10 and 12 of trace elements from the CJM-3 MME) (Table S3; Fig. S1-S3). Whereas zircons from host granite (CJM-6) have pronounced negative Eu anomalies (0.04-0.18) (Table S3; Fig. S2). Thus, the geochemical differences between MMEs and host granites are so significant (e.g., negative Eu anomalies Fig. S3) that zircons from MMEs cannot represent earlier zircons that have been entrained in the host granitic magma. Zircons (Type 1) have dark anhedral - subhedral granular with obvious and relatively light planar or fan-shaped crystalline bands in the core area, which are totally different from zircons with distinct narrow oscillatory zoning from the host granite (Fig. S1), although they have similar ages.

Ti-in-zircon (T_{zircon}) is used to estimate the magma temperature at the time of zircon crystallization (Watson et al., 2006). By using the T_{zircon} thermometry of Watson and Harrison (2005) (Fig. 8e), the T_{zircon} of the host granites range from 698 to 975 °C (the 1337 °C is excluded due its abnormally high value; Table S4) which are similar to those of MMEs (677 to 1006 °C), mostly higher than that the wet granite solidus (~650 °C; Scaillet et al., 2016). Therefore, Type 1 zircons in MMEs most likely crystallized during magma mixing.

However, the inherited cores from Type 2 zircons have older ages (236 Ma to 251 Ma) than that of the host granite, indicating that the inherited cores may represent the entrainment of earlier zircons from (1) minor parental magma of the Late Triassic batholiths in the Yidun Terrane that had been generated by at least ca. 251 Ma during subduction, or (2) captured from wall rock. Thus, some zircons of MMEs might be inherited from the entrainment of earlier zircons in the granitic magma chamber.

5.2.3 End members of MMEs

It has been proposed that the end-member source of mafic magma of Late Triassic magmatic rocks in the Yidun Terrane could be either the subcontinental lithospheric mantle (SCLM) or asthenospheric mantle (Zhan et al., 2021). The Late Triassic Gacun tholeiitic basalts (231-230 Ma) from Changtai area in the northern Yidun Terrane, with a relatively low positive $\epsilon\text{Nd}(t)$ values (+1.5 to +1.1; Wang et al., 2013), are derived from an enriched SCLM source. Whereas the Late Triassic basalts (216 Ma; $\epsilon\text{Nd}(t)$ values of +6.9 to +5.3) from the Xiaxiaoliu area in the southern Yidun Terrane (Chen et al., 2017) and Ganluogou ferrodiorite in the northern Yidun Terrane (depletions in Nb, Ta, Zr and Hf, and enrichments in Sr-Nd isotopes with -3.8 to -2.4, Wu et al., 2016) represent asthenospheric melts.

In-situ Sr isotopes of An-rich plagioclase crystals in MMEs can provide insights into the details of the different magma batches in a magmatic system, similar to Hf isotopes in zircon (Davidson et al., 2001; Griffin et al., 2002). The *in-situ* Sr isotopes of the An-rich plagioclases have shown that their cores have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.7027 and 0.7057 (Fig. 4b), indicating that the mafic magma was probably derived from depleted mantle. Two types of plagioclases have been identified based on the *in-situ* Sr isotopes: for those grains with homogeneous initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7049 – 0.7057; type 1), which are also lower than those of host granites (Fig. 4b, e) ($(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7080 - 0.7089$; Wu et al., 2017b), we propose that they were formed after magma mixing. The dark cores with the lowest initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7027 – 0.7028; type 2; Fig. 4b, c) may record the mafic end-member composition. Thus, the primary magma of sample CJM-11 was Zr-unsaturated and was derived from the depleted mantle rather than enriched SCLM.

After mixing with felsic magma, the mixed magma reached Zr-saturation and

began to crystallize zircons. During this mixing process, the rims of plagioclases crystallized and these had higher Sr isotope signatures (0.7049 to 0.7057, Table 5). Additionally, the cores of some plagioclase crystals have similar initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7027 to 0.7028) to those of Xiaxiaoliu basalts ($(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7037$ to 0.7062) which were derived from depleted mantle.

Thus, considering that the MMEs were derived from the mixing of depleted mantle-derived magmas and the host granitic magmas, we adopted the mixing model from Ma et al. (2000) to calculate the proportions of mantle contribution. The felsic end-member magma component is represented by CJM-6H with the highest SiO_2 content (73.44%) (Wu et al., 2017b), and the depleted mantle component is modeled using the Xiaxiaoliu basalt based on seven basaltic samples with mean value of initial $^{87}\text{Sr}/^{86}\text{Sr}=0.7053$ and $\text{SiO}_2 = 46.54\%$ from Chen et al. (2017). Most of the Late Triassic igneous rocks in the Yidun Terrane (except some granitoids or volcanic rocks such as Lannitang volcanic rocks – quartz monzonite porphyry, Pulang (quartz) diorite (porphyry) – quartz monzonite (porphyry) and Xiawacu quartz monzonite – granite that show distinct fractional crystallization processes), plot on or around the mixing line on Fig. 12. Due to the lack of Sr-Nd isotope data on the MMEs, we plotted *in-situ* Sr isotopes of the cores and rims of the plagioclases in MMEs on a $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. SiO_2 diagram (Fig. 12). The mixing lines apply to the cores and rims of plagioclase formed after magma mixing, and the result shows that ~10 – 40% Xiaxiaoliu basalt mixed with ~90 – 60% Cuojiaoma granite can produce the compositions of the MMEs. The cores of plagioclase have even lower $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios of 0.7027-0.7028 than those of the mean values of Xiaxiaoliu basalts, but they are close to the sample 10XXL-17 ($(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.7037$) of Xiaxiaoliu basalts, which indicate that the asthenospheric mantle may not homogeneous.

5.3 Constraints from zircon Zr isotopes

Zircon is an accessory mineral that has been widely used to assess the formation and evolution of continental crust (Kemp et al., 2006). Compared to bulk analysis, spatially resolved analyses of zircon Zr isotopes can provide more details of the chemical and thermal evolution of magmas (Guo et al., 2020). Stable Zr isotopes may be effective in tracing magma differentiation (Guo et al., 2020, 2023; Inglis et al., 2019; Wu et al., 2020; Yuan et al., 2022). It has been investigated that the Zr isotopes of the continuous composition of basalt - andesite - rhyolite series formed by magma differentiation in the Hekla region of Iceland and found that Zr isotope ratios become heavier with the increase of SiO₂ content in rocks, with a rapid increase occurring after 65%, indicating that the continuous crystallization of zircons in magma results in a heavier isotopic composition of Zr in the melt (Inglis et al., 2019). Yuan et al. (2022) analyzed the Zr isotope of Shidao alkaline complex in the Jiaodong area, and obtained that highly evolved samples possess relatively heavier Zr isotopes. Research on zircons of 90~96 Ma hornblende gabbro, biotite diorite and mafic enclaves from Gangdise batholith in southern Qinghai-Tibet Plateau, show that the core of zircons usually has a lighter Zr isotope composition, while the rims have a relatively heavier composition. The characteristic of "light core and heavy rim" can be explained as that when crystallization of zircons occur, lighter Zr isotopes are more likely to enter into zircons, resulting in the heavy isotopic composition in the residual melt (Guo et al., 2020). The average value of the calculated fractional coefficient is positively correlated with the average zircon Ti temperature, showing that compositional changes during magma crystallization process can be indicated by the differential fractionation of Zr isotopes in single zircon (Guo et al., 2020). Wu et al. (2020) investigated the Zr isotopes of zircons from 220 Ma granite and MMEs in Yidun

magmatic arc, China, and found that the rims of zircon have relatively heavier Zr isotopic composition, suggesting that the residual melt tends to be more enriched in heavy Zr isotopes with the continuous evolution of magma. Thus, zircon Zr isotopes have the potential to reveal zircon crystallization kinetics and magma dynamics. Integrated with other geochemical and isotopic tools, stable Zr isotopes may provide new insight into the dynamic history of diverse igneous systems through time (Guo et al., 2023).

Three types of zircons can be recognized based on Zr isotopic variation and crystal size generally (Guo et al., 2023). Type I zircons are those megacrysts which are > 1 mm in length including standard reference zircons and natural megacryst zircons (Zhang et al., 2019). Their intragrain isotopic variations are generally between 0.03‰ and 0.15‰ (Zhang et al., 2019; Tompkins et al., 2020). Type II zircons are represented by the Duluth FC1 gabbro zircons which are small (typically < ~100 μm in length) and have extremely large isotopic variations of ~5‰ and are from a mid-continental rift regime (Ibañez-Mejia and Tissot, 2019). Type III are mainly igneous zircons, with lengths of ~100-300 μm and isotopic variations in one sample usually < 1‰, most of which are internally zoned and characterized by Zr isotopically light cores and heavier rims (Guo et al., 2020, 2023).

5.3.1 Constraints on Cuojiaoma host granites

Zircons from Cuojiaoma host granite have an overall $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ variation of ~0.4‰ and also possess Zr isotopically light cores (-0.25‰ to 0.07‰) and heavier rims (0.03‰ to 0.15‰) (Fig. 9b; Table 6). They show similarities to the Type III zircons in the Gangdese arc, (Guo et al., 2020), which are thought to be formed in low supersaturation conditions (Guo et al., 2023).

Mass-dependent equilibrium isotope fractionation (EIF) can only explain small Zr isotopic variations, i.e., when the predicted zircon-melt EIF is between -0.032‰ and -0.081‰ at 700 °C (Chen et al., 2020; Méheut et al., 2021). Large Zr isotopic variations in natural samples are most likely due to kinetic isotope fractionation (KIF) which comprises diffusion-limited crystallization (DLC) and diffusion-triggered crystallization (DTC) (Chen et al., 2020; Méheut et al., 2021). With DLC, zircon rims which are crystallized later inherit the heavier Zr isotopic composition from the residual melt (Guo et al., 2020). In contrast to DLC, the growth of Zr-poor minerals during DTC tends to produce early crystallized zircons with heavier Zr isotopic compositions than the initial melt (Chen et al., 2020; Méheut et al., 2021). Small intragrain variations of $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ (0.02‰ to 0.29‰) and significant individual variations of $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ (0.40‰ to 0.84‰) are observed in zoned zircons from both MMEs and host granites from the Cuojiaoma granitic pluton, with zircon $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ increasing from cores to rims (Fig. 5a and 9). Thus, both zircons in MMEs and host granites were probably formed by DLC rather than EIF or DTC processes (Guo et al., 2023).

Cuojiaoma host granite contains abundant plagioclases which are primary Zr-poor minerals with Zr partition coefficient (K_D) of 0.036 – 0.36 (GERM database, <https://kdd.earthref.org/KdD>). Diffusive processes related to the growth of Zr-poor minerals results in Zr enrichment at the solid-melt interface and thus zircon saturation (Chen et al., 2020; Méheut et al., 2021). If diffusive processes are applied to the Cuojiaoma host granites, the early crystallized zircons at the high concentration of Zr would have high $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values and be characterized by heavy cores and light rims. However, such results cannot be reconciled with the fact that zircons in the Cuojiaoma granites have light cores and heavy rims (Fig. 5a).

Considering the different predictions of zircon Zr isotopic variations with time between the DLC and DTC processes, a negative $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}} - \text{Zr/Hf}$ correlation indicates diffusion limited zircon crystallization, while a positive correlation suggests diffusion-triggered zircon crystallization (Guo et al., 2020). The host granites have relatively negative $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}} - \text{Zr/Hf}$ correlations (Fig. 13b) in single grains which is consistent with the prediction of DLC processes. Decreasing Th/U ratios with increasing $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values in zircons are consistent with the progressive magma differentiation because the highly evolved magma usually contains low Th/U (Zhu et al., 2023). However, the zircon data show relatively constant $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values with decreasing Th/U ratios in zircons (Fig. 13d). Additionally, the U abundances (Harrison et al., 2007) and T_{zircon} (Ferry and Watson, 2007) are a good indicator of the temperature of zircon crystallization. The constant U contents and T_{zircon} in zircon (Fig. 13f, h) indicate that cooling of the magma has little impact on $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values. Thus, we propose that the DLC process is the predominant mechanism for the variations of $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values in the zircons of the Cuojiama host granite.

Decreasing Zr/Hf ratios are associated with larger negative Eu anomalies and increasing Yb/Gd ratios in zircons during magma differentiation (Fig. 13j, l). The reduction of Eu/Eu* and elevation of Yb/Gd ratios are controlled by the segregation of plagioclase (Nathwani et al., 2021) and crystallization of apatite (Reid et al., 2011). However, according to the ab initio calculations, plagioclase segregation would incorporate heavy Zr isotopes and lead to lighter Zr isotopic compositions in the evolved melt (Chen et al., 2020), which conflicts with the fact that $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ tends to become higher with magmatic differentiation. The Zr concentrations of apatite are generally lower than 5 ppm (Bea et al., 2006), indicating that it has a negligible influence on the $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ of the evolving silicate melts. In contrast, zircon

contains the highest concentrations of Zr (~500,000 ppm; [Finch and Hanchar, 2003](#)). Nearly all studied zircons have light-to-heavy $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ from cores to rims, suggesting the preferential incorporation of lighter Zr isotopes into zircon. Thus, the progressive crystallization of zircons rather than plagioclase or other accessory mineral results in evolved melts having heavy Zr isotopic compositions.

Based on the Sr-Nd-Hf isotopes of Cuojiaoma host granites, they were generated by the partial melting of meta-basaltic rocks of Kangding Complex with assimilated ancient crustal materials ([Wu et al., 2017b](#); [Xie et al., 2022](#)), followed by significant fractional crystallization ([Zhan et al., 2021](#); [Xie et al., 2022](#)). There are two ways that crystal fractionation can induce magmatic compositional evolution during fractional crystallization: either mineral phases which are separated from the magma reservoir accumulate in the open system ([Siegel et al., 2018](#)) or crystalized minerals which are not removed from the closed system result in in-situ differentiation of the adjacent melt ([Langmuir, 1989](#); [Edmonds et al., 2019](#)). The low Sr and large negative Eu anomaly (generally < 0.4) are features of extracted melt after crystal segregation because these elements are compatible with feldspars ([Deering and Bachmann, 2010](#); [Wu et al., 2017a](#)), which is consistent with the host granites containing low Sr (153-200 ppm; [Wu et al., 2017b](#)) and clear negative Eu anomalies (0.28-0.33; [Wu et al., 2017b](#)). The evolved melt after fractionation of saturated zircons would exhibit low Zr/Hf ratios (typically < 30) ([Deering and Bachmann, 2010](#)), which is consistent with the Zr/Hf ratios (26-30; [Wu et al., 2017b](#)) of Cuojiaoma host granites. We thus consider that the Cuojiaoma granites represent open-system crystallization with removal of feldspars and zircons.

5.3.2 Constraints on Cuojaoma MMEs

In-situ Zr isotopes in zircons can be used to explore open-system magma chamber processes in the mush region and provide insights into continental crust evolution (Zhu et al., 2023). Open-system processes such as crystal-melt segregation, magma mixing and crustal assimilation could be involved in the crystal mushes (Edmonds et al., 2019). It remains unclear how large Zr isotope fractionation can be produced during disequilibrium conditions (e.g., mixing) (Guo et al., 2023). A study on zircons in MMEs from Daocheng and Dongcuo plutons in Yidun magmatic arc reveal that the earlier crystallized zircons may contain lighter Zr isotopic compositions but those crystallized later would have heavier Zr isotopic compositions (Wu et al., 2020). Guo et al. (2020) suggested a temperature dependence of zircon-melt isotope fractionation controlled by differences in supersaturation conditions based on the MME in tonalite from Gangdese continental arc. This MME has petrographic features that relate to fast cooling which is the major cause of high supersaturation /undercooling (Guo et al., 2020; Sorokin et al., 2022).

Although there is a positive correlation between the $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ and Zr/Hf ratios in zircons of studied MMEs (Fig. 13a) which suggests a DTC process, these MMEs show an overall $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ variation of $\sim 1\text{‰}$ (Table 6). Furthermore, they have U-shape Zr isotopic profiles which are characterized by Zr isotopically light cores (-0.31‰ to 0.10‰) and gradually heavier rims (-0.39‰ to 0.41‰) (Fig. 5a and 9; Table 6), indicating that they may be produced by a DLC process. They are 50 – 120 μm in length, smaller than those of zircons in host granite. Rapid magma cooling produces high supersaturation conditions which may result in fast nucleation and growth of small zircon grains with relatively large Zr isotopic variations (Guo et al., 2023). This is consistent with lower T_{zircon} and higher U contents in zircons, reflecting

a decrease in temperature (Fig. 13e, g). Thus, we suggest that magma mixing between basaltic and felsic magmas may result in a temperature reduction and an increase of Zr/Hf and Th/U ratios (Fig. 13a, c). This is consistent with the Zr/Hf ratios (31.2-32.5, Table S2) of the MMEs, formed by magma mixing, which are between Zr/Hf ratios of felsic (26.4-29.9, Table S2) and basaltic magma (36.3 to 38.0, Table S2). There is no positive correlation between Zr/Hf and Eu/Eu*, indicating no segregation of plagioclase (Fig. 13i). This is consistent with the mineral assemblages of plagioclase, hornblende, quartz and alkali feldspar in CJM-11 (Fig. 4a, b and d). Decreasing Zr/Hf are associated with increasing Yb/Gd ratios in zircons during magma differentiation (Fig. 13k), suggesting the crystallization of apatite. However, apatite crystallization has little impact on the $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ of the melt because the rocks contain very few apatites.

Therefore, the zircons from MMEs from the Cuojiaoma granitic pluton crystallized during magma mixing. However, none of them have recorded the primary Lu-Hf isotopes of the mafic end-member, so the nature of the mafic end-member of MMEs cannot be investigated by simply using zircon Hf isotopes.

5.3.3 Zircon growth in MMEs

Core and rim analyses of zircons in the Cuojiaoma host granite, (Fig. 9b) show that the cores generally have lighter Zr isotopes than the rims generally, which may support the view that early crystallized zircons will have lighter Zr isotopes. However, the cores of some zircons from host granite CJM-6 have heavier Zr isotopes than their rims (Table 6; Fig. 9b). A possible explanation for this “reverse zoning” is that these zircons have inherited cores which is not distinct in the CL images (Wu et al., 2020). The rims of zircons from MMEs CJM-3 display large variations in Zr isotopes with

$\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values ranging from $-0.43 \pm 0.03\text{‰}$ to $+0.41 \pm 0.03\text{‰}$, whereas those of the host granite CJM-6 have positive values from $+0.03 \pm 0.06\text{‰}$ to $+0.15 \pm 0.06\text{‰}$ (Table 6; Fig. 9a). The zircon with low $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values may have crystallized in a less evolved fraction of the magma. Some zircons in MMEs have higher $\delta^{94/90}\text{Zr}_{\text{IPGP-Zr}}$ values than those of the host granite and these zircons may have crystallized in later interstitial melts. Thus, the relatively large variation of Zr isotopes from MMEs suggests that these zircons crystallized under rather complicated conditions, supporting the idea that they were generated during magma mixing. Interestingly, the cores of some zircons from MMEs also have heavier Zr isotopes than their rims (Table 6; Fig. 9a) and show “reverse zoning”, indicating that these zircons were formed during the early stage of crystallization in the host granite magma chamber.

Based on the above discussion, some of the zircons from MMEs (CJM-3) crystallized at an early stage, and some are actually crystals formed during magma mixing. However, none of zircons have recorded the signature of the primary Lu-Hf isotopes of the mafic end-member and it is therefore not possible to investigate the nature of the mafic source region of MMEs by zircon Hf isotopes. A schematic model summarizing our interpretations is given in Fig. 14.

5.4 Tectonic implications

The tectonic setting of the Triassic granitic belt in the Yidun Terrane has long been debated, as both arc and post orogenic extensional settings have been suggested to explain the generation of these rocks distributed parallel to the Garzê-Litang suture belt (Hou et al, 2003; Peng et al., 2014; Wu et al., 2016, 2017b; Zhan et al., 2021; Dong et al., 2022). Some plagioclases from the MMEs have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios much lower than the host granites and the arc volcanic rocks in the study area, but

763 similar to depleted mantle (Fig. 4b and 12a), implying that the associated
764 mantle-derived magma was not of arc origin, and may alternatively have been derived
765 from depleted mantle in a post-collisional setting.

766 The geochemical characteristics of the studied MMEs are consistent with a
767 post-collisional setting (Fig. 15). The igneous rocks from the YAB show an
768 eastward-younging trend (Wu et al., 2017b) corresponding to the migration of
769 magmatism to the trench of Garzê-Litang Ocean which is usually induced by slab
770 roll-back after initial continental collision (Zhu et al., 2011; Dong et al., 2022). The
771 tectonic setting of the study area switched from subduction, post-subduction to a
772 post-collision extensional setting, as well as a slab-break-off prior to or at 216 Ma
773 (Fig. 16) (Wang et al., 2011; Wu et al., 2016, 2017b; Zhan et al., 2021). Thus, the
774 MMEs were probably formed in a post-collisional regime. From 220 to 216 Ma, the
775 Garzê-Litang oceanic basin finally closed and slab break-off occurred. The tectonic
776 setting of the Yidun Terrane switched from subduction to a post-collision extensional
777 setting. The high temperature and decompression caused by the upwelling of
778 asthenosphere led to the formation of the Cuojiaoma granitic pluton. Simultaneously,
779 the partial melting of asthenospheric mantle also formed the primitive magma of the
780 Xiaxiaoliu basalts. The Cuojiaoma MMEs have similar but younger age (213 Ma)
781 than the host granite and Xiaoxiaoliu basalt (216 Ma, Chen et al., 2017), indicating
782 that they are coeval. The Sr isotopes of An-rich plagioclase therefore suggest that the
783 MMEs have experienced complex magmatic processes including crystallization and
784 fractionation with physical and chemical magma mixing. Mafic magma with
785 plagioclase crystals from an underlying magma chamber was injected into the granitic
786 magma chamber. Zircons in the magma of the MMEs crystallized during or after the
787 mixing between the mafic and felsic magmas.

6. Conclusions

(1) Zircons of the MMEs from the Cuojiaoma granitic pluton have similar U-Pb zircon ages of 213 Ma which is close to their host granite. However, these zircons have higher $\varepsilon_{\text{Hf}}(t)$ values, Th/U ratios and larger variations in Zr isotopes than those of their host granite.

(2) Zircons in the MMEs crystallized during magma mixing and none of them record the primary isotopic composition of the MMEs.

(3) An-rich plagioclase in the MMEs shows complicated core-rim relations with reverse and oscillatory zoning. Their variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios record interaction between the mantle-derived and crust-derived magmas.

(4) MMEs were generated by mixing of mafic magma and felsic magma under high temperature and decompression caused by asthenosphere upwelling in post-collision extensional setting.

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Declaration of interest statement

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

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