



# Geochronology and Petrogenesis of Early Pleistocene Dikes in the Changbai Mountain Volcanic Field (NE China) Based on Geochemistry and Sr-Nd-Pb-Hf Isotopic Compositions

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Changbai Mountains intraplate volcanism (NE China) developed above the 500 km deep stagnant portion of the Pacific slab in the last 20 Ma. The more recent activity includes a shield-forming stage (2.8–0.3 Ma), the Tianchi cone construction stage (1.5–0.01 Ma), and a caldera-forming stage (0.2 Ma-present). Detailed studies on the petrogenesis of the volcanic products between the first two stages and the possible role of geodynamics and local tectonics in controlling the volcanism, however, are lacking. Here, we present structural and whole-rock geochemical and zircon Hf isotopic data on Pleistocene dikes of the Changbai Mountains at the transition from the shield-forming to the Tianchi stage with the aim to constrain their age and the source(s) of their parental magma. The dikes represent the shallower feeding system of monogenetic cones and have a NW-SE strike, which is also the preferred strike of the major fault affecting the area and along which the Changbai Mountains monogenetic scoria cones align. The dikes have a potassic affinity and a trachybasaltic composition. Their zircon U–Pb age is 1.19–1.20 Ma (Calabrian). The trachybasalts are enriched in Rb, Ba, Th, U, Nb, Ta, K, Pb, and LREE and slightly depleted in Sr, Zr, Hf, Ti, and HREE with a weak negative Eu/Eu\* ( $\delta\text{Eu} = 0.96\text{--}0.97$ ). Trace elements and isotopic compositions are compatible with an OIB-type source with an EMI signature. The calculated ( $^{87}\text{Sr}/^{86}\text{Sr}$ )<sub>i</sub> (=0.705165–0.705324), ( $^{143}\text{Nd}/^{144}\text{Nd}$ )<sub>i</sub> (= 0.512552–0.512607,  $\epsilon\text{Nd}(t) = -0.58$  to  $-1.65$ ), and Hf model ages ( $T_{\text{DM2}}$ ) of 1768–1562 Ma suggest that the trachybasaltic dikes were contaminated by a Mesoproterozoic, relatively basic lower crust. The source of the Calabrian trachybasalts consists of asthenospheric melts modified by a subcontinental lithospheric mantle. These melts upwell from depth and stop at the crust-mantle interface where underplating processes favor the assimilation of ancient lower crust material. During the ascent to the surface along deep-seated crustal discontinuities, these magmas weakly differentiate.

**Keywords:** dikes, intraplate magmatism, OIB-EM1 sources, zircon U–Pb dating, Sr-Nd-Pb-Hf isotopes, Changbai Mountains-Tianchi 3

## HIGHLIGHTS

- 1) Trachybasalt dikes in Changbaishan volcanic area are controlled by a major crustal discontinuity.
- 2) The trachybasalt dikes were formed at the Calabrian stage of the early Pleistocene (1.19–1.20 Ma).
- 3) An OIB source with EM1 signature assimilates Mesoproterozoic crust by underplating processes.
- 4) Intraplate volcanism above stagnant slabs involves three sources: asthenosphere, slab-derived fluids, and ancient crustal rocks.

## INTRODUCTION

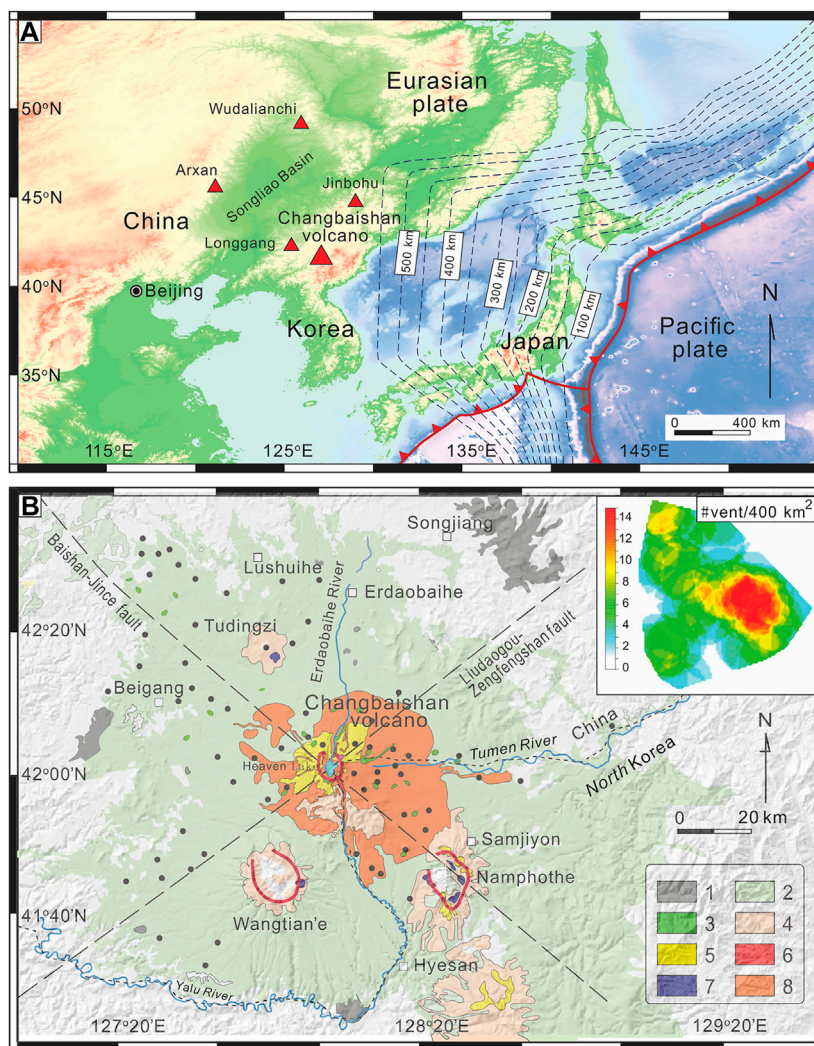
The eastern China block includes the eastern margin of the Central Asian orogenic belt (the Xing'an-Mongolia orogenic belt), the North China Craton (NCC), the Sulu-Dabie high pressure orogeny belt, the Yangtze Craton, and the southeast coastal orogenic belt of China, and it is characterized by Mesozoic volcanic and intrusive rocks (**Figure 1A**). The later Cenozoic volcanic rocks mainly consist of basalts emitted in an intraplate setting (e.g., Wu et al., 2011; Zhang and Luo, 2011; Deng et al., 2017) associated to metallogenic and hydrothermal deposits (Sun et al., 2012, 2013; Deng et al., 2017). Changbai Mountains, which consist of basaltic shields, monogenetic vents, and central-type volcanoes, are located above the 500 km deep stagnant slab of the westward subducting Pacific plate (**Figure 1A**). Three main geodynamic models have been suggested to explain the Changbai Mountain volcanism: an upwelling mantle plume (Mao et al., 1999), an intraplate continental rifting (Gilder et al., 1991), and the Pacific plate subduction (e.g., Mao et al., 2013; Ouyang et al., 2014; Zhang et al., 2014; Xu W. L. et al., 2020; Xu Z.-t. et al., 2020). Previous studies on the Changbai Mountains area mainly focus on the Holocene silicic magmatism (e.g., Wei et al., 2007; Liu et al., 2009; Xu J. D. et al., 2013a). The widely distributed early Pleistocene rocks are less known. The origin of the Pleistocene magmatism is still debated and different hypotheses have been proposed: 1) low degrees of partial melting of the asthenospheric mantle (Xu, et al., 2020a) possibly contaminated by the addition of crustal material (Sun et al., 2015); 2) partial melting of the subducted slab (Ma et al., 2015); 3) involvement of DM and EM1 end-members (Basu et al., 1991) or OIB and EM1 end-members (Xu et al., 2012); 4) an asthenospheric source metasomatized by recycled subducted silicate sediments or a subcontinental lithospheric mantle (EM1-like) and by carbonated eclogites with a MORB signature (Choi et al., 2020); and 5) an EM1-like source originated from the contamination of an OIB asthenospheric mantle by Paleoproterozoic lithosphere (Lee et al., 2021). Therefore, questions regarding the nature of the mantle and possible crustal sources of the Changbaishan Mountain volcanism and of the dynamics of the Pacific plate subduction remain open. In this study, we carried out systematic LA-MC-ICP-MS zircon U–Pb geochronology as well as whole-rock geochemistry, Sr–Nd–Pb isotopic analyses, and zircon Hf isotopic analyses to investigate the petrogenesis of the dikes

related to the early Pleistocene monogenetic activity in area. The results allow us to improve our understanding of the petrogenesis and tectonic setting of the Changbai Mountains magmatism. New data and constraints on the relative role of the asthenosphere, ancient crust, and stagnant slab in the evolution of intraplate volcanism are provided.

## GEOLOGICAL SETTING

The NE China magmatism mainly developed during Mesozoic and Cenozoic (Ge et al., 2003; Wu et al., 2005; Sun et al., 2007) eras. Because of continental rifting processes (Gilder et al., 1996; Zhou and Li, 2000; Zhu et al., 2011), the Changbai Mountain volcanic field (CMVF) is bounded by two NE–NW striking regional fault systems, the Dunhua–Mishan faults to the northeast and the Xar Morron–Changchun faults to the southeast although faults with a NW–SE strike are also locally present (**Figure 1B**). The origin of the latter faults is unclear, and they may represent the response of the NE China terrains to different rates of rifting associated to the Pacific slab roll-back. The basement of CMVF consists of Precambrian metamorphic rocks including Archean tonalite, trondhjemite, granodiorite (2.4–2.5 Ga) and Proterozoic quartzite, schists, and granulite assemblages (1.7–1.9 Ga) (Jin and Zhang, 1994; Wang et al., 2003). The CMVF volcanic rocks were emitted by three stratovolcanoes located at the China/North Korea boundary: the Tianchi caldera (last eruption in 1903) and the now inactive Wangtian'e in China and Namphothe edifices in North Korea. Several monogenetic scoria cones surround these three large stratovolcanoes. They roughly align along a NW–SE direction (inset in **Figure 1**). This direction is the same of the main fault affecting the Tianchi caldera (Baishan–Jince fault in **Figure 1**). After a basaltic plateau forming phase developed between 20 and 3 Ma, the volcanism in the Tianchi area can be roughly divided in three stages: 1) shield-forming stage (2.8–0.3 Ma), 2) the Tianchi cone construction stage (1.5–0.01 Ma), and 3) caldera-forming, explosive eruption stage (0.2 Ma–present) (Fan et al., 1999; Liu et al., 2015; Takeshi et al., 2020). Monogenetic volcanism mainly developed during the shield-forming stage, although some vents formed also during the cone construction stage. The early two stages were characterized by the emission of basaltic lava flows and scorias, while trachytes and rhyolites were emitted as lava flows and pyroclastic fall and flows, and these intermediate-acid magmatism mainly occurred in the last stage of basaltic magmatism.

The study area is located in the northern sector of the Tianchi area (**Figure 1B**). Here, the basement rocks include gneiss of the Paleoproterozoic Sidaolazi formation, amphibolite of the Yangjiadian formation, and quartz schist of the Mesoproterozoic Huashan formation. The volcanic rocks consist of Pleistocene and Pliocene basaltic lava flows and minor scorias (**Figure 2**). The products studied here are two dikes representing the shallowest feeding system of two neighbors, partly dismantled cones constituted by oxidized scorias and ash (**Figure 2**). Dike 1 strikes NNW–



**FIGURE 1 | (A)** Schematic map showing location of the Changbaishan and other major volcanoes in NE China. The Japan trench is reported as red solid line with triangles and the depth of the Pacific slab is reported as dashed lines; the number indicates the depth in km (modified from Zhang et al., 2018; Chen et al., 2021; Fan et al., 2021; Sun et al., 2021). **(B)** Geological map of the Changbai Mountains volcanic field showing distribution of polygenetic (Changbaishan (Tianchi), Wangtian'e, and Namphothe) and monogenetic (gray dots) volcanoes. Geological map legend: 1, pre-shield basalts; 2, shield-forming basalts; 3, post-shield basalts; 4, intermediate to silicic volcanic rocks (mainly trachyte with minor trachyandesite); 5, intermediate to silicic volcanic rocks (mainly trachyte with minor trachyandesite); 6, early rhyolite; 7, rhyolite; 8, pyroclastic deposits (primarily trachytic and rhyolitic in composition) erupted by caldera-forming eruptions of Tianchi. The inset on the top right reports the spatial density of monogenetic vents.

SSE; it is 15–20 m long and 0.4–1.8 m thick. Dike 2 strikes NW-SE with dip 80°–85°; it is about 30 m long and 0.7–5.2 m thick with some lateral digitations having length of less than 2 m (Figures 3A,B). These dikes have 5–15 mm wide vesicles testifying the degassing of magma near the surface (Figures 3C,D).

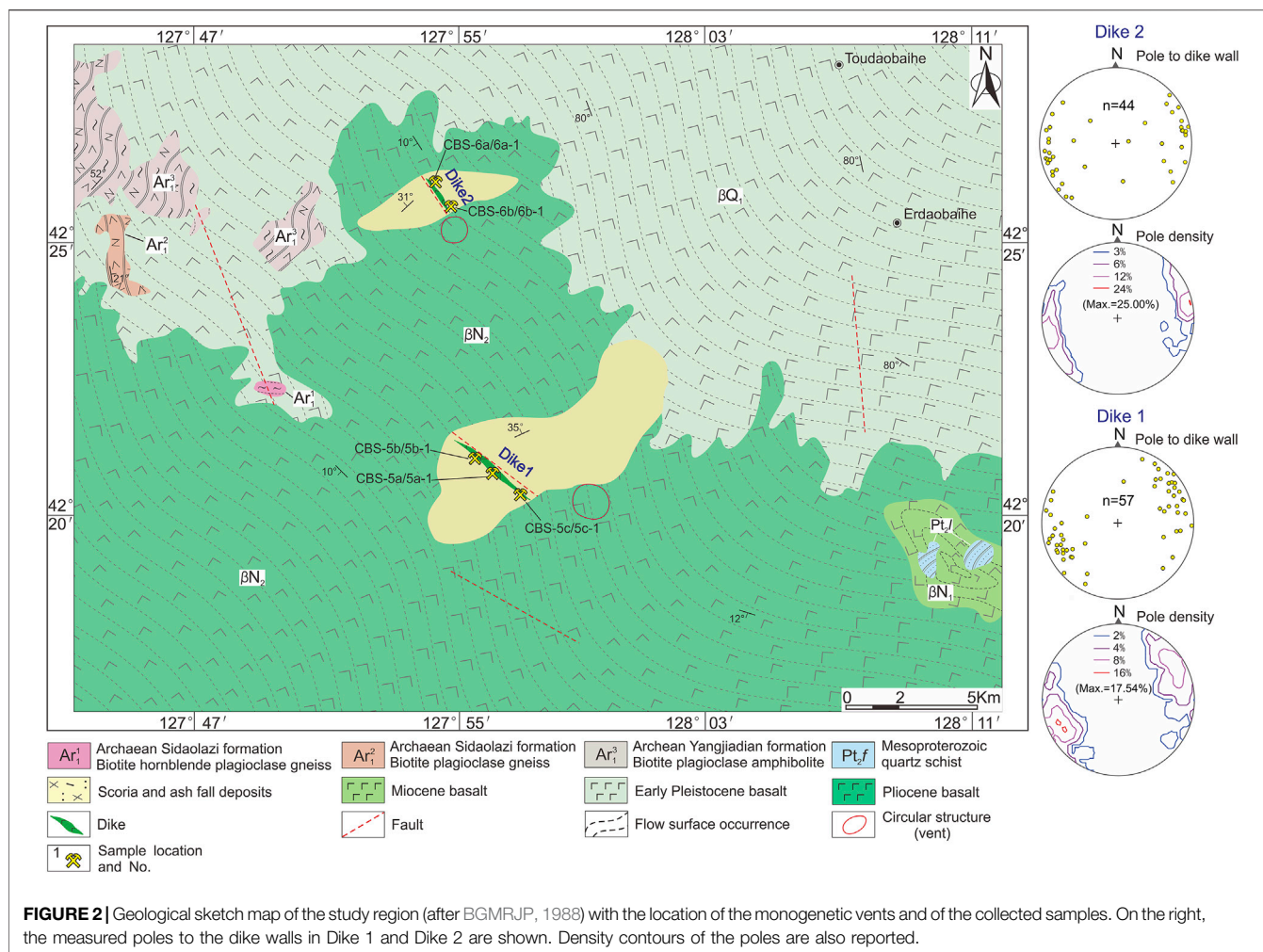
## SAMPLING AND ANALYTICAL TECHNIQUES

The sampling locations are shown in Figure 2. Unaltered, fresh samples were selected for a combined study involving whole-rock major and trace elements analyses, cathodoluminescence (CL)

imaging, zircon U-Pb ages, and whole-rock Sr-Nd-Pb isotopes and zircon Hf isotopic analyses.

## Sample Descriptions

The dikes are dark gray in color, massive to poorly vesiculated, and show a trachytic texture with plagioclases phenocrysts representing 80–85 vol% of the total minerals (Figures 3E,F). Clinopyroxene and biotite account for 5–10 vol% of the phenocrysts. The groundmass includes fine-grained plagioclase, pyroxene, amphibole, zircon (Figures 3G,H), apatite, and magnetite microlites. No significant differences in mineral assemblage, texture, and crystal content have been recognized between the two dikes.



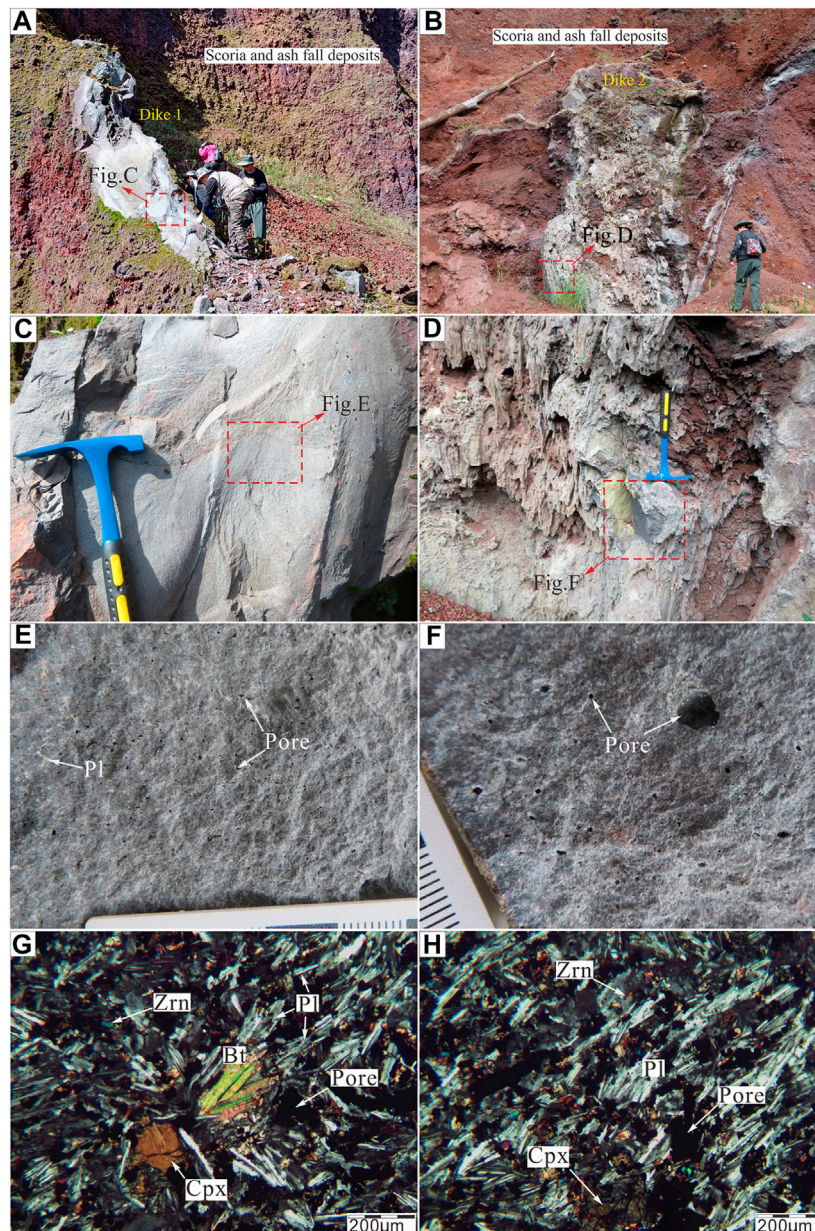
## LA-MC-ICP-MS Zircon U-Pb Dating

Zircon was mounted in epoxy discs, polished to expose the grains, cleaned ultrasonically in ultrapure water, and then cleaned again prior to the analysis using AR grade methanol. Pre-ablation was conducted for each spot analysis using 5 laser shots (~0.3  $\mu\text{m}$  in depth) to remove potential surface contamination. The analysis was performed using 30  $\mu\text{m}$  diameter spot at 5 Hz and a fluence of 2 J/cm<sup>2</sup>. Zircon U-Pb dating was conducted using LA-MC-ICP-MS (laser ablation multiple collector inductively coupled plasma mass spectrometry) in Nanjing Hong Chuang Geological Exploration Technology Service Company Limited (NHGETSCL). The Resolution SE model laser ablation system (Applied Spectra, United States) was equipped with ATLEX 300 excimer laser and a two volume S155 ablation cell. The laser ablation system was coupled to a Neptune Plus MC-ICP-MS (Wrexham, Wales, United Kingdom). Detailed tuning parameters can be seen in Thompson et al. (2018). Iolite software package was used for data reduction (Paton et al., 2010). Zircon 91500 and GJ-1 were used as primary and secondary reference materials, respectively. 91500 was analyzed twice and GJ-1 analyzed once every 10–12 analyses of the sample.

The age of GJ-1 ( $604 \pm 6$  Ma, 1  $\sigma$ ) measured during the experiment was consistent with the recommended values within the uncertainty range, and the preferred value of 91500 is in Wiedenbeck et al. (1995); the <sup>207</sup>Pb/<sup>206</sup>Pb and <sup>206</sup>Pb/<sup>238</sup>U age of 91500 is 1065.4 and 1062.4 Ma. Based on the decay coefficient, Th/U ratio, and the error of the measurements in young zircons, we corrected the age loss caused by <sup>203</sup>Th depletion (Zou et al., 2010). The LA-MC-ICP-MS U-Pb isotopic data are listed in **Table 1**.

## Zircon Lu-Hf Isotopic Analyses

Experiments of *in situ* Hf isotope ratio analysis were conducted using the Teledyne Cetac Technologies Analyte Excite laser ablation system (Bozeman, Montana, United States) and the Nu Instruments Nu Plasma II MC-ICP-MS (Wrexham, Wales, United Kingdom) that were hosted at the State Key Laboratory for Mineral Deposits Research, Nanjing University. The 193 nm ArF excimer laser, homogenized by a set of beam delivery systems, was focused on zircon surface with fluence of 6.0 J/cm<sup>2</sup>. Ablation protocol employed a spot diameter of 50  $\mu\text{m}$  at 8 Hz repetition rate for 40 s. Helium



**FIGURE 3** | Field photographs (A–D), typical hand specimens (E, F), and photomicrographs (G, H) of the trachybasalt dikes in the study area. Abbreviations: Bt = biotite; Cpx = clinopyroxene; Pl = plagioclase; Zrn = zircon.

was applied as carrier gas to efficiently transport aerosol to MC-ICP-MS; the results of the zircon Hf isotopic analyses are given in **Table 2**. The precision of sample test is 0.000005–0.000012 (0.015–0.04‰, 2s), and the accuracy is better than 0.000 012 (~0.04‰).

### Major and Trace Element Determinations

The geochemical sample analyses were performed at the State Key Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics, Chinese Academy of Sciences. The major element oxides were analyzed by a Panalytical

PW4400 X-ray fluorescence spectrometer (AxiosmAX, Holland). The detailed sample-digesting procedure was as follows: 1) sample powder (200 mesh) were placed in an oven at 105°C for drying of 12 h. 2) ~1.0g dried sample was accurately weighted and placed in the ceramic crucible and then heated in a muffle furnace at 1000°C for 2 h. After cooling to 400°C, this sample was placed in the drying vessel and weighted again in order to calculate the loss on ignition (LOI). 3) 0.6 g sample powder was mixed with 6.0 g cosolvent and 0.3 g oxidant in a Pt crucible, which was placed in the furnace at 1150°C for 14 min. Then, the sample was

**TABLE 1** | LA-MC-ICP-MS zircon U–Pb data of trachybasalt dike from the Changbai Mountain volcanic rock area.

Sample No.	U				Th/U				Isotope ratio				Age/Ma							
	Th	U	ppm	Th/U	<sup>207</sup> Pb/ <sup>235</sup> U	1σ	<sup>206</sup> Pb/ <sup>238</sup> U	1σ	<sup>206</sup> Pb/ <sup>207</sup> Pb	1σ	<sup>206</sup> Pb/ <sup>235</sup> Th	1σ	<sup>207</sup> Pb/ <sup>235</sup> U	1σ	<sup>207</sup> Pb/ <sup>206</sup> Pb	1σ	<sup>206</sup> Pb/ <sup>235</sup> Th	1σ		
Trachybasalt dike 1, CBS-5a, mean = 1.19 ± 0.0057 Ma, n = 15																				
CBS-5a-1	5893	2373		2.5	0.001271	0.000029	0.0001874	0.0000010	5336.179	28.47481	0.04900	0.0011	0.0000629	0.0000004	1.29	0.03	1.21	0.01	137.00	1.27
CBS-5a-2	2046	1336		1.5	0.001382	0.000035	0.0001870	0.0000010	5347.594	28.59676	0.05360	0.0014	0.0000641	0.0000005	1.40	0.04	1.21	0.01	309.00	1.30
CBS-5a-3	4131	2138		1.9	0.001247	0.000028	0.0001862	0.0000010	5371.723	27.98975	0.04880	0.0011	0.0000626	0.0000004	1.27	0.03	1.20	0.01	120.00	1.27
CBS-5a-4	3684	1842		2.0	0.001326	0.000039	0.0001836	0.0000013	5446.623	38.56541	0.05250	0.00150	0.0000617	0.0000005	1.35	0.04	1.18	0.01	274.00	1.25
CBS-5a-5	3028	1887		1.6	0.001259	0.000027	0.0001837	0.0000009	5443.066	27.25681	0.05000	0.0011	0.0000622	0.0000004	1.28	0.03	1.18	0.01	171.00	1.26
CBS-5a-6	3780	1841		2.1	0.001331	0.000030	0.0001864	0.0000010	5385.958	28.50657	0.05170	0.0012	0.0000624	0.0000004	1.35	0.03	1.20	0.01	251.00	1.26
CBS-5a-7	1977	1501		1.3	0.001307	0.000034	0.0001865	0.0000011	5361.930	31.62533	0.05100	0.0014	0.0000633	0.0000006	1.33	0.04	1.20	0.01	207.00	1.28
CBS-5a-8	654.4	734		0.9	0.001350	0.000069	0.0001847	0.0000016	5414.185	46.90144	0.05280	0.0027	0.0000636	0.0000012	1.37	0.07	1.19	0.01	230.00	1.29
CBS-5a-9	1554	1363		1.1	0.001409	0.000043	0.0001865	0.0000017	5361.930	48.87550	0.05480	0.0018	0.0000644	0.0000008	1.43	0.04	1.20	0.01	365.00	1.30
CBS-5a-10	2761	1757		1.6	0.001303	0.000026	0.0001855	0.0000010	5390.254	27.89265	0.05070	0.0010	0.0000628	0.0000004	1.32	0.03	1.20	0.01	214.00	1.27
CBS-5a-11	3285	2139		1.5	0.001384	0.000036	0.0001856	0.0000011	5387.931	31.93278	0.05420	0.0014	0.0000631	0.0000005	1.41	0.04	1.20	0.01	355.00	1.28
CBS-5a-12	3640	2065		1.8	0.001282	0.000031	0.0001826	0.0000015	5476.451	44.98728	0.05100	0.0013	0.0000623	0.0000006	1.30	0.03	1.18	0.01	218.00	1.26
CBS-5a-13	2051	1457		1.4	0.001212	0.000034	0.0001829	0.0000012	5467.469	35.87185	0.04820	0.0014	0.0000628	0.0000006	1.23	0.03	1.18	0.01	88.00	1.27
CBS-5a-14	1662	1276		1.3	0.001261	0.000036	0.0001840	0.0000012	5434.783	35.44423	0.0500	0.0015	0.0000626	0.0000006	1.28	0.04	1.19	0.01	149.00	1.27
CBS-5a-15	1889	1348		1.4	0.001248	0.000043	0.0001827	0.0000015	5473.454	44.93804	0.04960	0.0017	0.0000619	0.0000006	1.27	0.04	1.18	0.01	144.00	1.25
Trachybasalt dike 2, CBS-6a, mean = 1.20 ± 0.0048 Ma, n = 10																				
CBS-6a-01	3104	1829		1.7	0.001326	0.000030	0.0001856	0.0000010	5387.060	28.7302100	0.05190	0.00120	0.0000638	0.0000005	1.35	0.03	1.20	0.01	250.00	1.29
CBS-6a-02	7090	3433		2.1	0.001266	0.000021	0.0001838	0.0000011	5440.696	32.5613000	0.05005	0.00085	0.0000614	0.0000004	1.29	0.02	1.18	0.01	182.00	1.24
CBS-6a-03	1268	1231		1.0	0.001359	0.000037	0.0001863	0.0000014	5367.687	40.3368800	0.05320	0.00150	0.0000653	0.0000007	1.38	0.04	1.20	0.01	285.00	1.32
CBS-6a-04	2543	1671		1.5	0.001272	0.000030	0.0001867	0.0000011	5356.186	31.5576100	0.04950	0.00120	0.0000632	0.0000005	1.29	0.03	1.20	0.01	144.00	1.28
CBS-6a-05	879	1073		0.8	0.001361	0.000044	0.0001854	0.0000013	5383.743	37.8202100	0.05330	0.00170	0.0000649	0.0000009	1.38	0.05	1.20	0.01	286.00	1.31
CBS-6a-06	1552	1229		1.3	0.001266	0.000046	0.0001861	0.0000015	5373.455	43.3110300	0.04950	0.00160	0.0000631	0.0000006	1.29	0.05	1.20	0.01	138.00	1.28
CBS-6a-07	1429	1139		1.3	0.001317	0.000046	0.0001854	0.0000015	5393.743	43.6887000	0.05140	0.00180	0.0000624	0.0000008	1.34	0.05	1.19	0.01	222.00	1.26
CBS-6a-08	1198	1218		1.0	0.001410	0.000041	0.0001852	0.0000013	5399.568	37.9019400	0.05330	0.00160	0.0000637	0.0000008	1.43	0.04	1.19	0.01	368.00	1.29
CBS-6a-09	2466	1577		1.6	0.001354	0.000034	0.0001863	0.0000011	5367.687	31.6932600	0.05290	0.00130	0.0000633	0.0000005	1.37	0.03	1.20	0.01	289.00	1.28
CBS-6a-10	4087	2573		1.6	0.001316	0.000025	0.0001874	0.0000010	5336.179	28.4748100	0.05100	0.00100	0.0000638	0.0000004	1.34	0.03	1.21	0.01	216.00	1.29

**TABLE 2** | MC-LA-ICP-MS zircon Lu–Hf isotopic data for the trachybasalt dikes of the Changbai Mountain volcanic area.

No.	Age (Ma)	$^{176}\text{Yb}/^{177}\text{Hf}$	2 $\sigma$	$^{176}\text{Lu}/^{177}\text{Hf}$	2 $\sigma$	$^{176}\text{Hf}/^{177}\text{Hf}$	2 $\sigma$	$\epsilon\text{Hf}(0)$	$\epsilon\text{Hf}(t)$	TDM1 (Ma)	TDM2 (Ma)	fLu/Hf
Trachybasalt dike 1, CBS-5a, $n = 15$												
CBS-5a-01	1.21	0.062707	0.001568	0.002052	0.000040	0.282700	0.000018	−2.6	−2.5	805	1768	−0.94
CBS-5a-02	1.21	0.055832	0.000777	0.001865	0.000028	0.282757	0.000017	−0.5	−0.5	719	1585	−0.94
CBS-5a-03	1.20	0.073038	0.000130	0.002444	0.000007	0.282764	0.000017	−0.3	−0.3	720	1562	−0.93
CBS-5a-04	1.18	0.038649	0.001056	0.001411	0.000035	0.282752	0.000017	−0.7	−0.7	718	1603	−0.96
CBS-5a-05	1.18	0.029207	0.000286	0.000992	0.000008	0.282720	0.000017	−1.8	−1.8	754	1702	−0.97
CBS-5a-06	1.20	0.050335	0.000730	0.001737	0.000023	0.282714	0.000015	−2.0	−2.0	778	1722	−0.95
CBS-5a-07	1.20	0.074336	0.001120	0.002543	0.000040	0.282722	0.000022	−1.8	−1.7	783	1696	−0.92
CBS-5a-08	1.19	0.050710	0.001263	0.001737	0.000046	0.282720	0.000016	−1.8	−1.8	769	1703	−0.95
CBS-5a-09	1.20	0.029297	0.000238	0.001044	0.000007	0.282720	0.000014	−1.8	−1.8	755	1703	−0.97
CBS-5a-10	1.20	0.080004	0.000779	0.002703	0.000026	0.282720	0.000018	−1.8	−1.8	790	1703	−0.92
CBS-5a-11	1.20	0.069372	0.000220	0.002339	0.000007	0.282762	0.000015	−0.4	−0.3	721	1570	−0.93
CBS-5a-12	1.18	0.016243	0.000180	0.000588	0.000006	0.282727	0.000014	−1.6	−1.6	736	1680	−0.98
CBS-5a-13	1.18	0.063841	0.000410	0.002154	0.000012	0.282741	0.000017	−1.1	−1.1	748	1638	−0.94
CBS-5a-14	1.19	0.062606	0.000248	0.002097	0.000011	0.282742	0.000016	−1.1	−1.0	745	1632	−0.94
CBS-5a-15	1.18	0.062604	0.000729	0.002075	0.000020	0.282755	0.000014	−0.6	−0.6	726	1593	−0.94
Trachybasalt dike 2, CBS-6a, $n = 10$												
CBS-6a-01	1.20	0.043957	0.000447	0.001522	0.000011	0.282731	0.000017	−1.5	−1.4	749	1668	−0.95
CBS-6a-02	1.18	0.017150	0.001408	0.003769	0.000035	0.282766	0.000015	−0.2	−0.2	745	1558	−0.89
CBS-6a-03	1.20	0.068196	0.001944	0.002352	0.000062	0.282729	0.000016	−1.5	−1.5	769	1674	−0.93
CBS-6a-04	1.20	0.072148	0.001162	0.002389	0.000043	0.282750	0.000020	−0.8	−0.8	740	1608	−0.93
CBS-6a-05	1.20	0.046462	0.000438	0.001613	0.000014	0.282729	0.000013	−1.5	−1.5	754	1674	−0.95
CBS-6a-06	1.20	0.031677	0.000113	0.001078	0.000004	0.282746	0.000014	−0.9	−0.9	720	1622	−0.97
CBS-6a-07	1.19	0.039444	0.000217	0.001324	0.000005	0.282745	0.000016	−0.9	−0.9	725	1622	−0.96
CBS-6a-08	1.19	0.038749	0.000527	0.001339	0.000016	0.282769	0.000013	−0.1	−0.1	692	1548	−0.96
CBS-6a-09	1.20	0.047389	0.000692	0.001586	0.000023	0.282750	0.000014	−0.8	−0.7	723	1607	−0.95
CBS-6a-10	1.21	0.085789	0.001011	0.002845	0.000033	0.282747	0.000017	−0.9	−0.9	754	1618	−0.91

quenched with air for 1 min to produce flat discs on the 5fire brick for the XRF analyses. Trace elements and rare earth elements (REEs) were analyzed by Perkin-Elmer ELAN 300D ICP-MS (United States). The detailed sample-digestion procedure refers to Sun et al. (2013). The analytical

precisions for major elements and trace elements were better than 3 and 5%, respectively, and the major and trace element compositions of the trachybasalt dikes (CBS-5a, CBS-5b, CBS-5c, CBS-5a-1, CBS-5b-1, CBS-6a, and CBS-6b) are given in **Table 3**.

**TABLE 3 |** Major (wt%) and trace element (ppm) compositions of the trachybasalt dikes of the Changbai Mountain volcanic area.

Sample No.	Trachybasalt dyke 1			Trachybasalt dyke 2			
	CBS-5a	CBS-5b	CBS-5c	CBS-6a-1	CBS-6b-1	CBS-6a	CBS-6b
SiO <sub>2</sub>	49.32	49.35	49.20	49.28	49.36	49.08	49.14
Al <sub>2</sub> O <sub>3</sub>	16.00	16.05	15.94	16.04	15.94	15.67	15.75
TiO <sub>2</sub>	3.36	3.36	3.33	3.36	3.34	3.28	3.26
Fe <sub>2</sub> O <sub>3</sub>	13.79	13.85	13.75	13.86	13.72	13.70	13.65
FeO	7.19	7.01	7.15	7.08	7.11	5.17	7.35
TFe <sub>2</sub> O <sub>3</sub>	19.60	19.47	19.52	19.55	19.46	17.50	17.50
MnO	0.17	0.17	0.17	0.17	0.17	0.17	0.17
MgO	4.24	4.27	4.24	4.25	4.22	4.00	4.00
CaO	6.80	6.81	6.78	6.84	6.80	6.69	6.65
Na <sub>2</sub> O	3.82	3.83	3.79	3.82	3.80	3.54	3.77
K <sub>2</sub> O	2.55	2.54	2.55	2.51	2.51	2.34	2.45
P <sub>2</sub> O <sub>5</sub>	0.74	0.74	0.73	0.74	0.74	0.70	0.70
K <sub>2</sub> O + Na <sub>2</sub> O	6.37	6.38	6.34	6.34	6.31	5.88	6.22
LOI	-0.57	-0.54	-0.59	-0.68	-0.64	0.08	-0.28
Total	100.24	100.43	99.88	100.20	99.95	105.13	105.13
K <sub>2</sub> O/Na <sub>2</sub> O	0.67	0.66	0.67	0.66	0.66	0.66	0.65
A/CNK	1.21	1.22	1.21	1.22	1.22	1.25	1.22
Mg*	27.85	28.09	27.92	27.93	27.88	28.96	27.27
Li	3.53	3.52	3.74	3.63	3.63	3.62	3.72
Be	2.29	2.27	2.34	2.26	2.26	2.27	2.31
K	21167.87	21109.79	21159.57	20860.85	20811.06	20860.85	20811.06
P	3282.73	3273.85	3256.08	3287.18	3264.96	3287.18	3264.96
Sc	17.16	18.00	18.28	17.60	17.97	18.97	18.99
Ti	20768.96	20901.33	21122.11	21030.50	21379.13	19823.08	19716.12
V	227.44	229.28	232.56	231.29	232.57	229.28	232.56
Cr	7.66	7.88	7.76	7.58	7.80	7.88	7.79
Co	38.20	38.68	39.30	38.69	39.10	38.68	39.32
Ni	26.60	26.65	27.33	26.78	26.74	26.65	27.34
Cu	18.19	18.32	18.75	18.33	18.53	18.32	18.76
Zn	155.77	155.14	160.57	157.30	161.23	155.14	160.57
Ga	27.24	27.49	27.63	26.76	27.25	27.49	27.63
Rb	46.27	54.18	57.13	52.21	54.94	54.18	57.13
Sr	629.96	649.46	655.50	638.34	650.66	649.46	655.50
Zr	310.96	311.58	317.01	313.45	312.83	311.58	317.01
Nb	49.05	49.30	50.20	49.49	49.27	49.30	50.20
Mo	2.15	2.13	2.17	2.11	2.12	2.13	2.17
Cd	0.30	0.29	0.30	0.29	0.29	0.29	0.30
In	0.10	0.10	0.09	0.10	0.10	0.10	0.09
Sb	0.05	0.04	0.05	0.04	0.05	0.04	0.05
Cs	0.34	0.35	0.36	0.35	0.33	0.35	0.36
Ba	767.78	798.94	813.59	776.41	807.94	798.94	812.59
Hf	7.42	7.41	7.43	7.35	7.57	7.41	7.52
Ta	2.99	2.95	2.96	2.91	3.00	2.95	2.97
W	0.94	0.90	0.90	0.90	0.92	0.90	0.91
Tl	0.03	0.03	0.03	0.03	0.03	0.03	0.29
Pb	6.49	7.37	7.77	6.82	7.62	7.37	7.76
Bi	0.01	0.01	0.01	0.00	0.01	0.01	0.01
Th	5.15	5.13	5.16	5.02	5.30	5.13	5.22
U	1.13	1.12	1.14	1.11	1.16	1.12	1.15
Y	34.17	34.22	34.78	34.18	34.62	34.22	34.68
La	45.38	46.23	46.96	45.23	46.86	46.23	46.86
Ce	93.81	94.12	95.39	92.97	95.57	94.12	95.40
Pr	11.36	11.41	11.61	11.22	11.65	11.41	11.61
Nd	45.91	45.99	46.54	45.13	46.90	45.99	46.56
Sm	9.77	9.70	9.84	9.53	9.92	9.77	9.88
Eu	3.00	2.98	3.00	2.93	3.05	2.93	3.04
Gd	8.98	8.94	9.01	8.82	9.09	8.80	9.07
Tb	1.28	1.27	1.28	1.25	1.29	1.27	1.28
Dy	6.78	6.72	6.80	6.64	6.89	6.66	6.88
Ho	1.29	1.28	1.30	1.26	1.32	1.26	1.32
Er	3.25	3.23	3.24	3.17	3.31	3.17	3.31
Tm	0.45	0.44	0.45	0.43	0.45	0.43	0.45
Yb	2.66	2.65	2.67	2.60	2.72	2.75	2.73

(Continued on following page)



**TABLE 3 |** (Continued) Major (wt%) and trace element (ppm) compositions of the trachybasalt dikes of the Changbai Mountain volcanic area.

Sample No.	Trachybasalt dyke 1			Trachybasalt dyke 2			
	CBS-5a	CBS-5b	CBS-5c	CBS-6a-1	CBS-6b-1	CBS-6a	CBS-6b
Lu	0.40	0.40	0.40	0.39	0.41	0.40	0.42
ΣREE	234.31	235.36	238.47	231.56	239.43	284.78	297.78
LREE	209.23	210.43	213.33	207.01	213.96	265.66	273.66
HREE	25.08	24.93	25.14	24.56	25.47	24.72	25.48
LREE/HREE	8.34	8.44	8.48	8.43	8.40	8.51	8.37
La <sub>N</sub> /Yb <sub>N</sub>	11.49	11.75	11.87	11.71	11.63	11.35	11.61
δEu	0.96	0.96	0.96	0.96	0.97	0.95	0.98
δCe	0.97	0.96	0.96	0.97	0.96	0.96	0.96
(Na <sub>2</sub> O + K <sub>2</sub> O)/δEu	6.61	6.63	6.62	6.60	6.52	6.70	6.64

\*Mg (Magnesium index) = mole [100 \* (MgO/40.3044)/(MgO/40.3044 + (FeO + Fe<sub>2</sub>O<sub>3</sub> \* 0.8998)/71.844)].

**TABLE 4 |** Lead isotope compositions of trachybasalt dikes in the study area.

Sample No.	<sup>206</sup> Pb/ <sup>204</sup> Pb	1σ	<sup>207</sup> Pb/ <sup>204</sup> Pb	1σ	<sup>208</sup> Pb/ <sup>204</sup> Pb	1σ	<sup>207</sup> Pb/ <sup>206</sup> Pb	1σ	<sup>208</sup> Pb/ <sup>206</sup> Pb	1σ
CBS-5a-1	17.7047	0.00024	15.5525	0.00020	38.0870	0.00069	0.8788	0.00001	2.1517	0.00003
CBS-5b-1	17.7164	0.00018	15.5549	0.00016	38.0933	0.00063	0.8786	0.00001	2.1512	0.00003
CBS-5c-1	17.7241	0.00026	15.5580	0.00021	38.0929	0.00058	0.8784	0.00001	2.1502	0.00002
CBS-6a-1	17.7071	0.00084	15.5553	0.00029	38.0958	0.00098	0.8791	0.00003	2.1525	0.00008
CBS-6b-1	17.7172	0.00025	15.5564	0.00023	38.0993	0.00076	0.8786	0.00001	2.1514	0.00003

## Sr-Nd-Pb Isotopic Analyses

Whole-rock Pb isotopic measurements were conducted by a MAT-261 thermal ionization mass spectrometer. All chemical preparations were performed on class 100 work benches within a class 1000 over-pressured clean laboratory. Sample digestion is as follows: 1) sample powder (200 mesh) was placed in an oven at 105°C for drying of 12 h; 2) 50–200 mg sample powder was accurately weighed and placed in a Teflon bomb; 3) 1–3 ml HNO<sub>3</sub> and 1–3 ml HF were added into the Teflon bomb; 4) Teflon bomb was put in a stainless steel pressure jacket and heated to 190 °C in an oven for >24 h; 5) after cooling, the Teflon bomb was opened and placed on a hotplate at 140°C and evaporated to incipient dryness, and then 1 ml HNO<sub>3</sub> was added and evaporated to dryness again; and 6) the sample was dissolved in 1.0 ml of 1.0 M HBr. After centrifugation, the supernatant solution was loaded into an ion-exchange column packed with AG resin. After complete draining of the sample solution, columns were rinsed with 1.0 M HBr to remove undesirable matrix elements. Finally, the Pb fraction was eluted using 6.0 M HCl and gently evaporated to dryness prior to mass-spectrometric measurement. The sample analyses were performed at the Analytical Laboratory of the Beijing Research Institute of Uranium Geology (BRIUG), the precision of <sup>206</sup>Pb/<sup>204</sup>Pb is (0.002–0.025%, 1σ), and the accuracy is better than 0.03%. The results of the whole-rock Pb isotopic analyses are given in **Table 4**.

The results of Sr–Nd isotope analyses are reported in **Table 5**. All chemical preparations were performed on class 100 work benches within a class 1000 over-pressured clean laboratory; the experimental process refers to Pan et al. (2020).

The REE solution was used to separate the Nd fraction by the Nd-column method. Sr and Nd isotopes analyses were performed on a Nu Plasma II MC-ICP-MS (Wrexham, Wales, UK) at the Analytical Laboratory of the BRIUG, Beijing, China. The precision of the test sample = 0.000005–0.00002 (0.01–0.05‰, 1σ), better than 0.000025 (~0.05‰). Raw data of isotopic ratios were corrected for mass fractionation by normalizing to <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194 for Sr, <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219 for Nd, and <sup>205</sup>Tl/<sup>203</sup>Tl = 2.3885 for Pb with exponential law. International isotopic standards (NIST SRM 987 for Sr, JNdi-1 for Nd, and NIST SRM 981 for Pb) were periodically analyzed to correct instrumental drift. Geochemical reference materials of USGS BCR-2, BHVO-2, AVG-2, and RGM-2 were treated as quality control.

## RESULTS

### Zircon U–Pb Dating

Zircons from the trachybasalt dikes (sample CBS-5a and CBS-6a) are light gray with no inclusions. They are euhedral–subhedral columnar crystals with size of 50–200 μm (aspect ratio of 1:1 to 3:1). In cathodoluminescence (CL) images, the zircons exhibit low, homogeneous luminescence and occasionally display oscillatory zoning (**Figure 4**). Th/U is 0.8–2.5, a range of values consistent with a magmatic origin (Hoskin and Schaltegger, 2003). The 25 analyses of zircon grains from CBS-5a and CBS-6a give weighted mean <sup>206</sup>Pb/<sup>238</sup>U ages of 1.19 ± 0.0057 Ma (MSWD = 2.00, n = 015) and 1.20 ± 0.0048 Ma (MSWD = 0.89, n = 10), respectively (**Figures**

TABLE 5 | Sr–Nd isotopic data of trachybasalt dikes in the study area.

Sample No.	t Ma	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$		$^{87}\text{Sr}/^{86}\text{Sr}$		$^{86}\text{Sr}/^{87}\text{Sr}$		$1\sigma$	$^{87}\text{Sr}/^{86}\text{Sr}$		$1\sigma$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	Sm ppm	Nd ppm	$^{147}\text{Sm}/^{144}\text{Nd}$		$^{143}\text{Nd}/^{144}\text{Nd}$		$1\sigma$	$(^{143}\text{Nd}/^{144}\text{Nd})_i$	$\epsilon\text{Nd}(t)$	TDM1 Ma	TDM2 Ma	fSm/Nd
				$1\sigma$	$1\sigma$	$1\sigma$	$1\sigma$	$1\sigma$	$1\sigma$																	
CBS-5a-1	1.19	46.27	629.96	0.207480	0.705328	0.000006	0.006783	0.000001	0.056809	0.000008	0.705324	9.77	45.91	0.133612	0.512553	0.000006	0.512552	-1.65	1139	953	-0.32					
CBS-5b-1	1.19	54.18	649.46	0.235616	0.705290	0.000005	0.006780	0.000001	0.056786	0.000007	0.705286	9.70	45.99	0.132364	0.512567	0.000005	0.512566	-1.37	1095	931	-0.33					
CBS-5c-1	1.19	57.13	655.50	0.246157	0.705269	0.000006	0.006781	0.000001	0.056792	0.000008	0.705264	9.84	46.54	0.132682	0.512572	0.000004	0.512571	-1.28	1091	924	-0.33					
CBS-6a-1	1.20	52.21	638.34	0.231031	0.705240	0.000004	0.006804	0.000001	0.056894	0.000007	0.705236	9.53	45.13	0.132496	0.512608	0.000003	0.512607	-0.58	1021	867	-0.33					
CBS-6b-1	1.20	54.94	650.66	0.238469	0.705169	0.000005	0.006803	0.000001	0.056974	0.000006	0.705165	9.92	46.90	0.132773	0.512581	0.000004	0.512580	-1.10	1075	910	-0.32					

Note: the initial ratio of Sr–Nd isotopes and the model age are based on  $\lambda\text{Rb} = 1.43 \times 10^{-11}$  years $^{-1}$  and  $\lambda\text{Sm} = 6.540 \times 10^{-12}$  years $^{-1}$  (Peucat et al., 1988).

5B,D). Because the magmatism in the Tianchi area occurred in the late Cenozoic, the content of common Th, U, and Pb in the Th–U–Pb system of zircons may be considered high, and the content of  $^{235}\text{U}$  decays to  $^{207}\text{Pb}$  is low. Therefore,  $^{206}\text{Pb}/^{208}\text{U}$  is selected as the weighted average age. Tera–Wasserburg concordia diagrams were used to calculate the concordant age ( $1.19 \pm 0.0030$  Ma and  $1.20 \pm 0.0038$ , respectively, Figures 5A,C). This age is taken to represent the emplacement age of the studied dikes.

## Zircon Lu–Hf Isotopic Data

The spots used for LA–MC–ICP–MS U–Pb dating are the same as those for *in situ* zircon Lu–Hf isotopic analysis. Twenty-five analyses of the trachybasalt dikes (sample CBS-5a and CBS-6a) yielded  $^{176}\text{Hf}/^{177}\text{Hf}$  of 0.282700–0.282769,  $\epsilon\text{Hf}(t) = -2.5$  to  $-0.1$ , and  $T_{\text{DM2}} = 1548$ –1768 Ma (Figure 8A).

## Whole-Rock Geochemistry

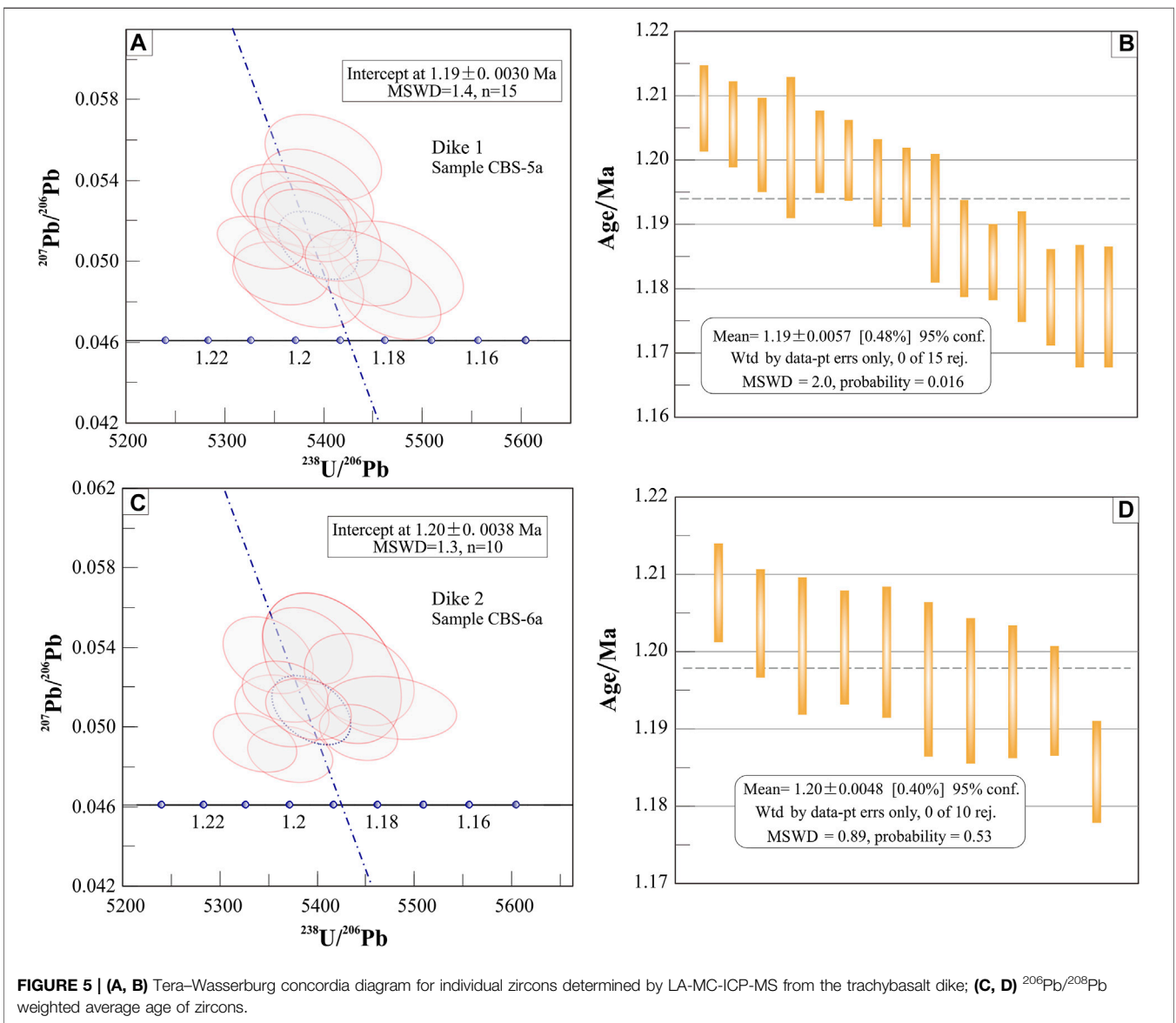
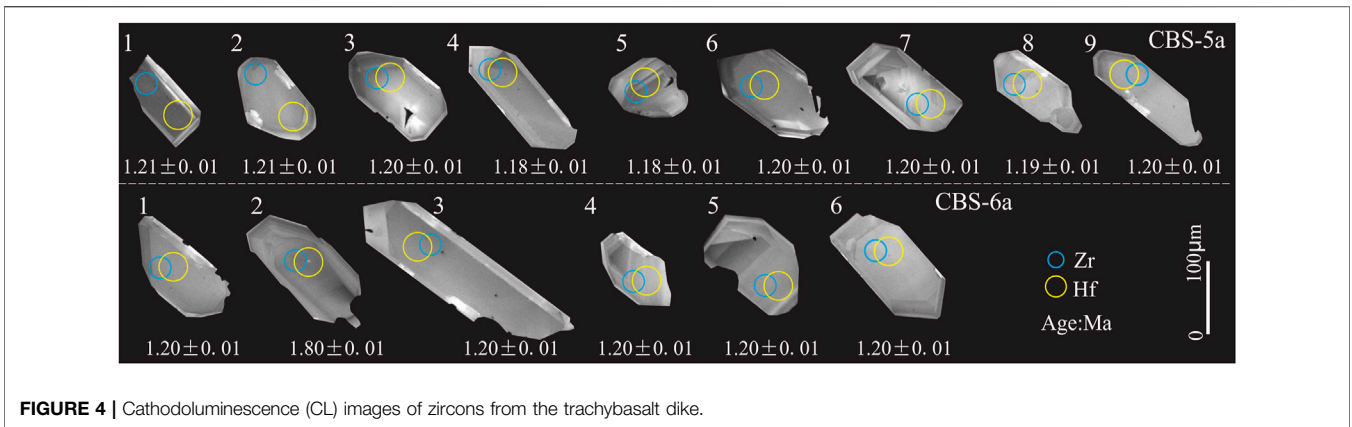
The dikes show a trachybasaltic composition ( $\text{SiO}_2 = 49.08$ – $49.35$  wt%;  $\text{Na}_2\text{O} = 3.54$ – $3.83$  wt%, and  $\text{K}_2\text{O} = 2.34$ – $2.55$  wt%) with high  $\text{Al}_2\text{O}_3$  (15.67–16.05 wt%) and  $\text{P}_2\text{O}_5$  (0.70–0.74 wt%) contents and have a potassic affinity (Figures 6A,B). Mantle-normalized trace element (REE) and chondrite-normalized rare earth element patterns are reported in Figures 7A,B. The samples display parallel trace element patterns. They show relative enrichments in LILE and HSFE. The Zr/Nb (6.31–6.35), La/Nb (0.91–0.95), Ba/Nb (15.65–16.21), Ba/Th (149.22–157.57), Rb/Nb (0.94–1.14), K/Nb (421.50–431.55), Th/Nb (0.10–0.11), and Ba/La (16.92–17.33) values are higher than those of MORB, whereas the chondrite-normalized REEs show enrichments in LREE and depletion in HREE; LREE/HREE ranges from 8.34 to 8.48. The dikes are characterized by a low differentiated REE pattern ( $\text{La}_N/\text{Yb}_N = 11.49$ – $11.75$ ) and weak Eu negative anomalies ( $\text{Eu}/\text{Eu}^* = \delta\text{Eu} = 0.96$ – $0.97$ ).

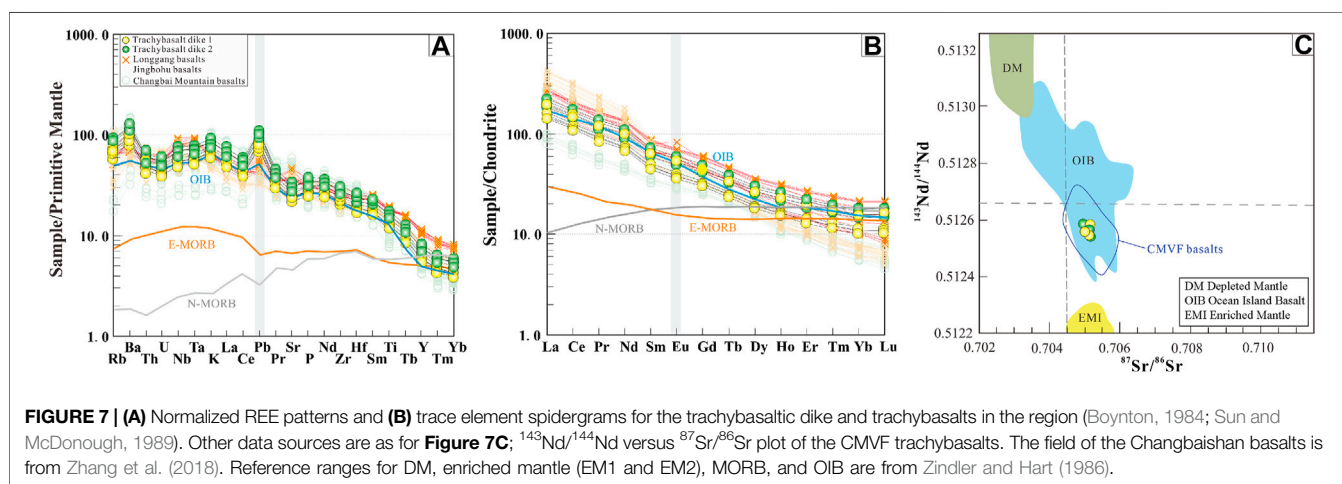
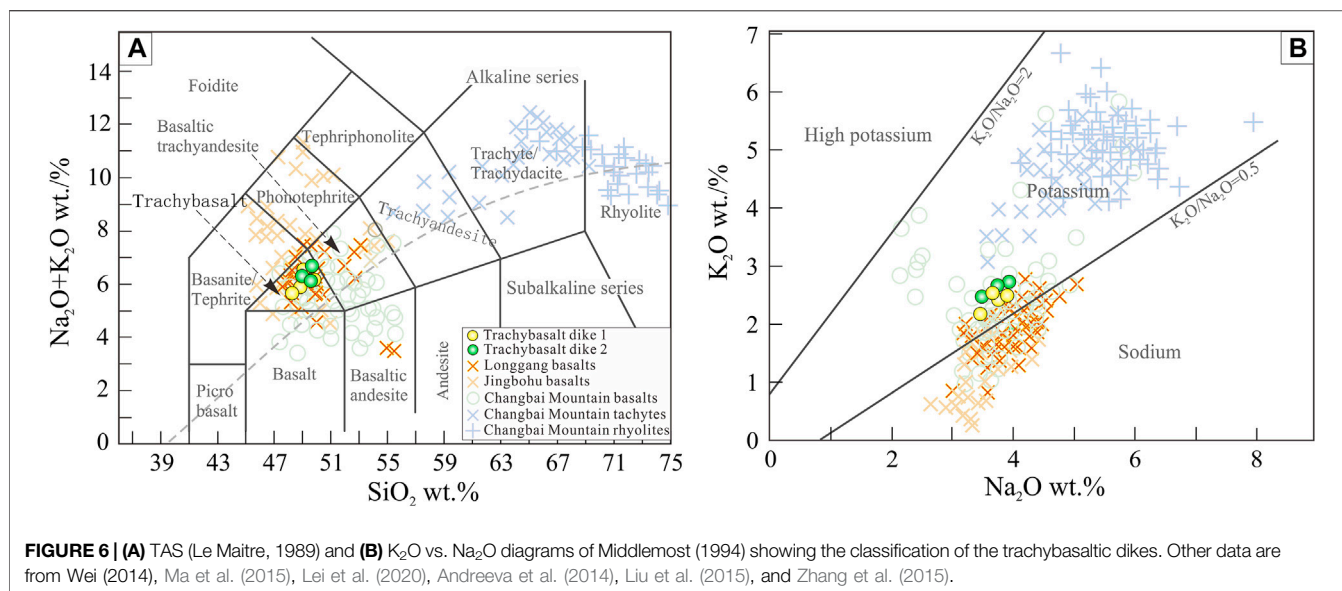
## Lead Isotope Composition

The dikes have moderately high  $^{206}\text{Pb}/^{204}\text{Pb}$  (17.7047–17.7241) values with  $^{207}\text{Pb}/^{204}\text{Pb} = 15.5525$ – $15.5580$  and  $^{208}\text{Pb}/^{204}\text{Pb} = 38.0870$ – $38.0993$ . In the  $^{207}\text{Pb}/^{204}\text{Pb}$  vs.  $^{206}\text{Pb}/^{204}\text{Pb}$  plots (Figure 8B), they lie within or close to the values representative of the lower crust.

## Sr–Nd Isotope Composition

Five samples of the trachybasalt dikes were analyzed for their whole-rock Sr–Nd isotopic composition. They have evolved radiogenic isotope compositions with  $(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.705165$ – $0.705324$  and  $(^{143}\text{Nd}/^{144}\text{Nd})_i = 0.512552$ – $0.512607$  corresponding to  $\epsilon\text{Nd}(t)$  values of  $-0.58$  to  $-1.65$ . According to the data reported in Figure 8C, the Sr–Nd isotope composition of the trachybasaltic dikes approaches that of the coeval basaltic rocks in the CMVF region.



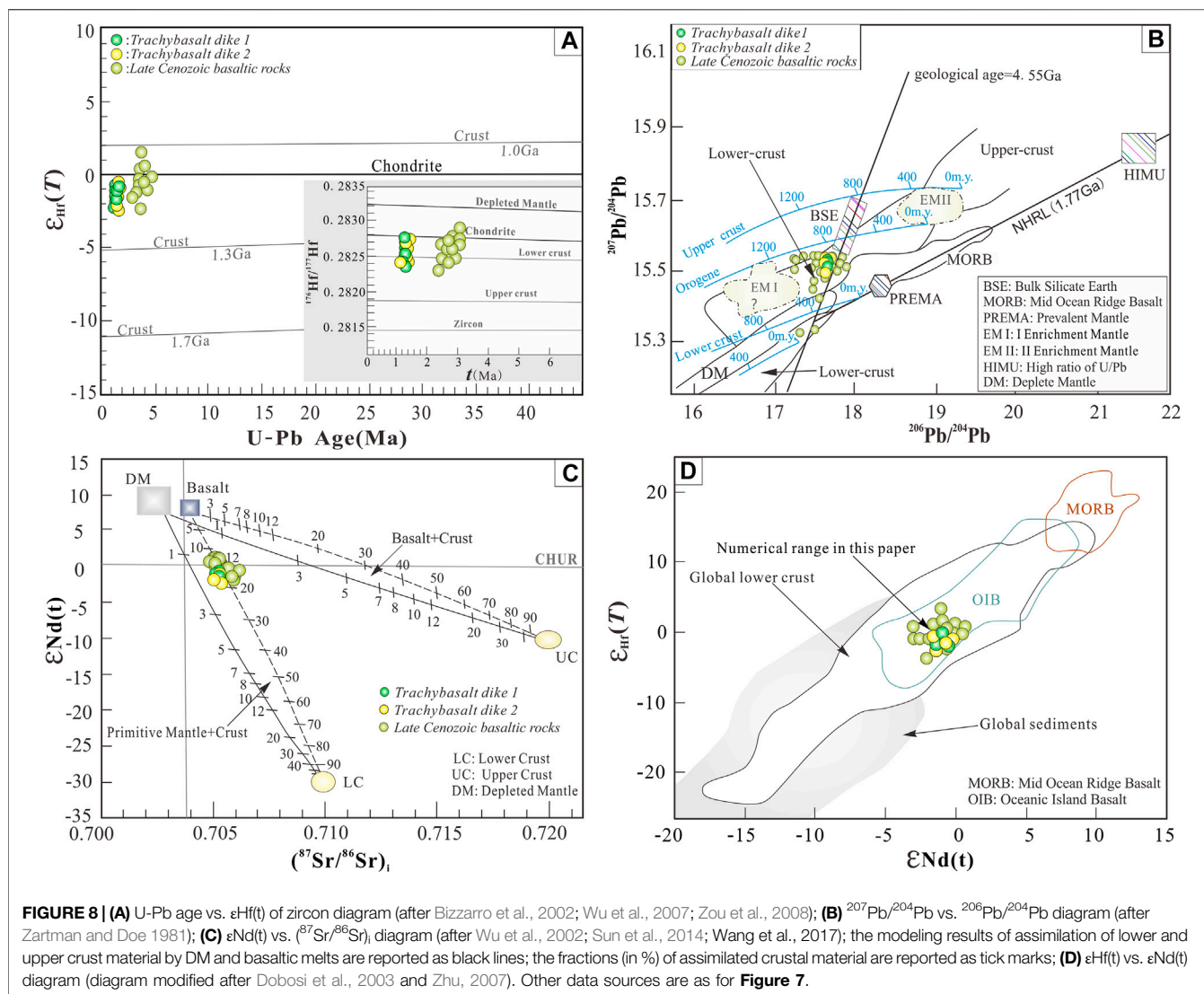


## DISCUSSION

### Timing and Tectonic Control of Volcanism at CMVF

Application of U–Pb disequilibrium dating of zircons with contrasting stabilities and compositions can provide a unique perspective to study the evolution of magma within the framework of absolute time (Wu et al., 2004). All dike outcrops in the field consist of hypabyssal basalts (Figures 3A,B). All dated zircons from the dikes are euhedral–subhedral columnar crystals and display oscillatory zoning; their Th/U is 0.8–2.5, a value consistent with a magmatic origin. Although the zircon crystals of the two dikes could be interpreted as antecrysts, however, their petrographic features do not support such hypothesis. The growth of zircons in poorly evolved magmas may be due to different processes including (Borisova et al., 2020) the partial melting of

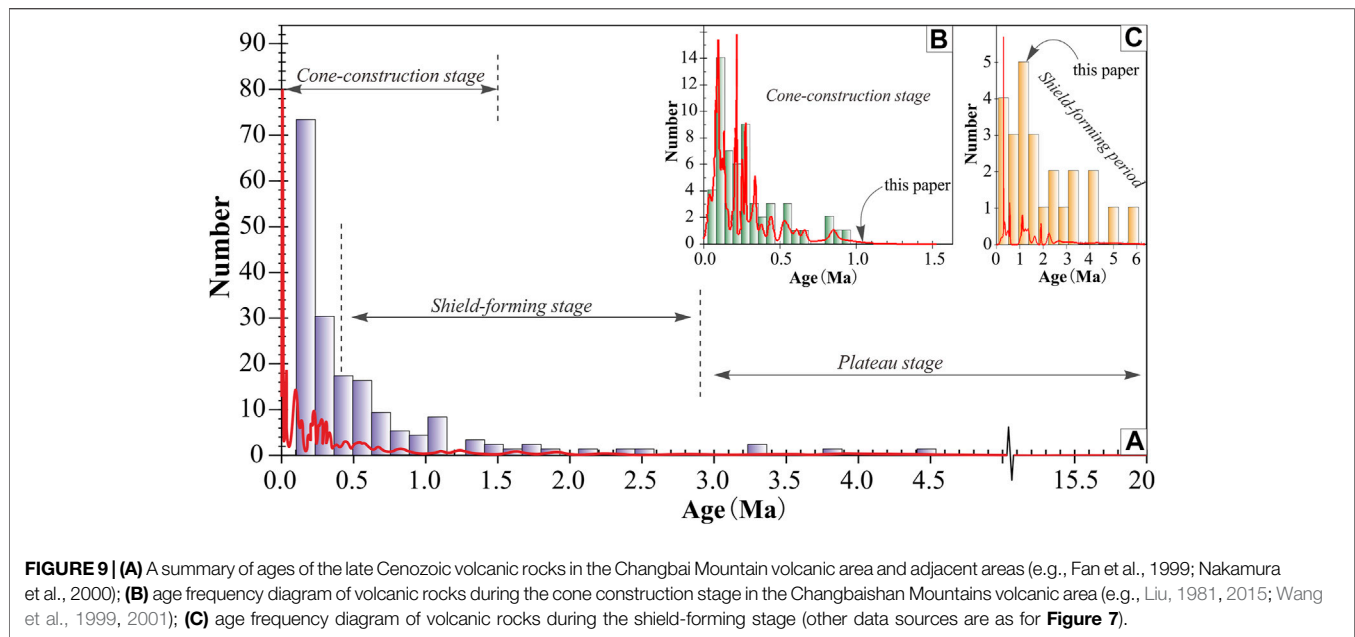
hydrated peridotite, the felsic crustal involvement/recycling or zircon sealing, total zircon dissolution, and reprecipitation and/or a contribution of metasomatic Si–F-rich fluid. Our data do not allow us to better constrain the origin of the CMVF zircons studied here. As preliminary hypothesis, crustal recycling (see Discussion) or metasomatic processes in a hypabyssal environment could be responsible for the formation of the zircon crystals of CMVF dikes. Field observations suggest that dikes were formed by basaltic magma hypabyssal process (Figures 3A,B), consistent with the euhedral–subhedral columnar morphology and oscillatory zoning of all the dated zircon crystals. Single grain LA–MC–ICP–MS zircon U–Pb concordance ages of 1.19–1.20 Ma ( $n = 25$ ) and  $^{206}Pb/^{238}U$  ages (1.18–1.20 Ma) are here obtained from the trachybasaltic dikes and confirm that the age is 1.18–1.20 Ma. We conclude that studied dikes emplaced in the Calabrian stage of Pleistocene. As previously reported, the last three main stages of the Tianchi



volcanism are 2.8–0.3 Ma (shield stage), 1.5–0.01 Ma (cone construction stage), and 0.2 Ma–present (caldera-forming stage) (Cheong et al., 2019; Zou et al., 2021). Therefore, our new geochronological data show that the trachybasalt dikes emplaced between the shield and cone forming stages (Figures 9A–C). The trachybasalt dikes, which are coeval with the less evolved basaltic magmatism, evidence an early phase of evolution processes within the CMVF magmatic system, at least in the Tianchi norther sector. Our field data better constrain the role of regional tectonics in controlling the CMVF volcanism during Calabrian. In fact, the NW–SE strike of dikes mirrors that of the major structural discontinuity affecting the CMVF area, i.e., the Baishan–Jince fault, and the preferred alignment of the monogenetic volcanoes (Figure 1). As a result, the CVMF Calabrian magmatism developed along a deep discontinuity probably affecting the whole crust along which basaltic magma upwelled and partly differentiated. The  $\text{SiO}_2$  content and the values of A/CNK, MgO,  $\text{Fe}_2\text{O}_3$ ,  $\text{TiO}_2$ ,  $\text{P}_2\text{O}_5$ , and V of the dikes confirm that they represent poorly evolved products of mantle melts.

## Petrogenesis

As known, intraplate magmatism can be due to different processes including 1) asthenosphere upwelling carried by mantle plume associated or not to a hot spot, 2) lithosphere thinning, continental stretching, and consequent upwelling of the lithospheric mantle, 3) fluid release from a subducting plate and mantle metasomatism, and 4) crust–mantle interaction (Defant and Drummond, 1990; Puffer, 2001; Sisson et al., 2005). The Sr, Nd, Pb, and Hf isotopic ratios (Figure 7C; Figures 8B, D), Zr/Hf (41.35–42.69), Nb/Ta (16.41–16.98), Th/U (4.53–4.56), Th/La (0.11), Nb/U (43.49–44.58), and Lu/Hf (0.05), and the ratio of other trace elements (Figure 7A, B; Figure 8; Zhao, 2016) of the Calabrian CVMF dikes (Figure 8B) are consistent with an OIB-type mantle source (Figure 10B). In particular, the values of the Sr and Nd isotopic ratios (Figure 7C) overlap those of the CVMF basaltic rocks, for which an OIB source has been recognized (Zhang et al., 2018; Choi et al., 2020). However, the trace element pattern indicates that the source is more complex. The  $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\delta\text{Eu}$  (6.52–6.63), Nb/La (1.05–1.09), La/Yb (10.36–17.61), Th/Nd (0.11–0.15), and  $\delta\text{Ce}$  (0.96–0.97) values and incompatible

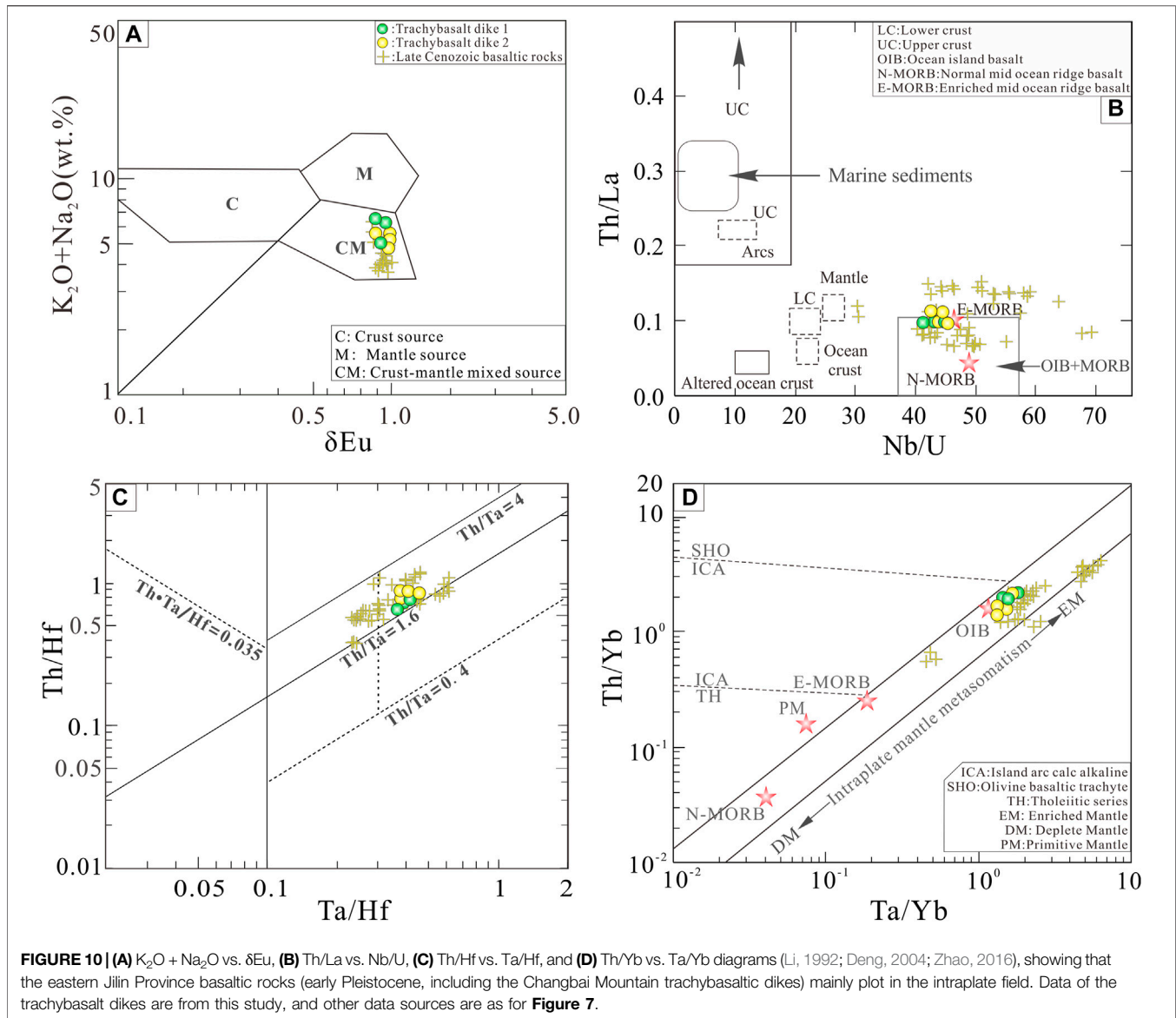


trace element abundances such as Th, U, Nb, and Ta suggest the involvement of an enriched mantle type I (EMI) (**Figures 10C,D**) and some interactions with crustal material (**Figure 10A**). The CVMF Pleistocene basaltic rocks consist of early sodium-rich rocks to later potassic rocks, and the data obtained in this study indeed indicate an alkaline potassic affinity of the rocks emplaced during Calabrian. According to Wang et al. (2017) and Lei et al. (2020), the EMI signature of the NE China potassic volcanism could be due to fluids released by the Pacific sediments of the subducted Pacific slab. More in general, potassic rocks as those studied here are generally considered to be derived from the partial melting of a mantle modified by slab-derived fluids or melts from the subcontinental lithospheric mantle (SCLM) (Foley and Peccerillo, 1992). The last stage of the Pacific subduction plate between NE China and Japan may favor the dehydration of the oceanic crust allowing the partial melting of the mantle. In this framework, the interaction between hydrous fluid and the mantle can produce a phlogopite-bearing source whose partial melting may produce potassic magmas (Wyllie and Sekine, 1982; Condamine and Médard, 2014). Generally, melts in equilibrium with phlogopite are expected to have higher Rb/Sr ( $>0.1$ ) and lower Ba/Rb ( $<20$ ) values than those formed from amphibole-bearing mantle sources (Furman and Graham, 1999), and the early Pleistocene basaltic rocks of the study area have low Rb/Sr (0.07–0.09) and Ba/Rb (14.24–16.59) values. This suggests that potassic amphibole is the main K-rich phase. Because the mantle material is depleted of Zr and Hf, it is impossible for the mantle derived magma to crystallize zircon without the addition of crustal material. As a result, the occurrence of zircon in the mantle derived magma indicates that some crust-mantle interactions occurred during the magma emplacement (Zheng et al., 2006). The studied trachybasaltic dikes are characterized by negative  $\epsilon\text{Hf}(t)$  values ( $-0.1$  to  $-2.5$ ),  $^{176}\text{Hf}/^{177}\text{Hf} = 0.000588\text{--}0.003769$ , and the Mesoproterozoic two-stage Hf

model ages are from 1548 Ma to 1768 Ma (**Figure 6A**). This indicates that there is some crustal material involved in the magma generation processes, i.e., crustal contamination. On the other hand, the consistent negative Nd isotopic compositions ( $\epsilon\text{Nd}(t) = -0.58$  to  $-1.67$ ) of the trachybasalt dikes preclude their derivation from the melting of an old lower continental crust ( $T_{\text{DM1}} = 1021\text{--}1139$  Ma). The variation ranges of the  $(^{87}\text{Sr}/^{87}\text{Sr})_t$  and  $\epsilon\text{Nd}(t)$  values in the mafic rocks in the region are 0.70485–0.70619 and  $+0.02$  to  $-1.65$ , respectively. These values indicate an initial magmatic source consisting of OIB-EMI type melts (**Figure 8C**). Furthermore, the Pb-Pb isotopic data (**Figure 8B**) of the trachybasalt dikes and CVMF basaltic rocks support a mixed source consisting of mantle and ancient lower crust components. In the  $\epsilon\text{Hf}(t)\text{--}\epsilon\text{Nd}(t)$  diagram (**Figure 8D**), all the samples fall into the overlapping area of the global lower crust and OIB. The results of the modeling of assimilation of lower crust material in **Figure 8C** suggest that the studied trachybasalts assimilated about 10–20% of crustal material.

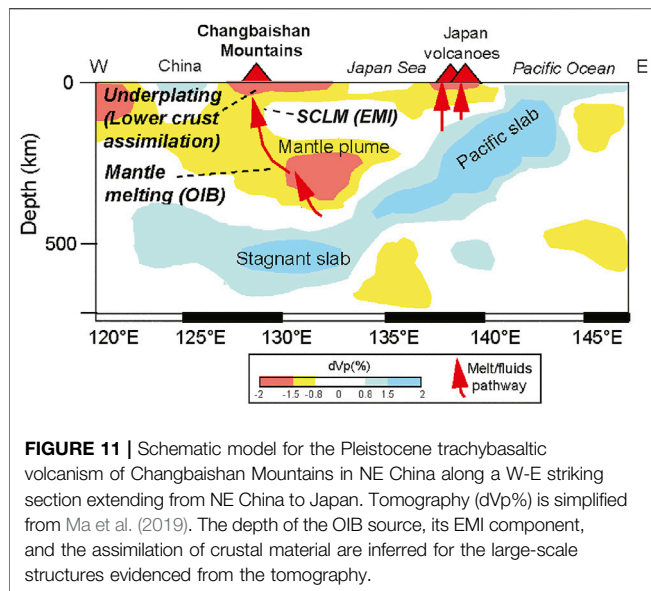
## Geodynamic Implications

The late Cenozoic basaltic rocks in the NE China continental margin have geochemical compositions consistent with OIB-type and EMI sources (Zhou and Armstrong, 1982; Xu et al., 2005; Zhi et al., 1990; Liu et al., 2015; Wang et al., 2015; Mao and Wang, 2000; Liu et al., 2008; Chen et al., 2009; Andreeva et al., 2014; Zhang et al., 2009; Ge et al., 2001; Wang et al., 2011). However, the CVMF Calabrian dikes studied here have moderate  $\text{Mg}^\#$  (27.85–28.09) and higher  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  contents with respect to the less evolved CVMF basaltic rocks (**Figure 6**). Therefore, the dikes do not represent primary magmas and experienced some crystal fractionation. Despite these features, they show geochemical features, e.g., values of  $\text{Na}_2\text{O}/\text{K}_2\text{O}$  and isotopic ratios, and some trace element contents, comparable with



those of the CMVF less evolved rocks (**Figure 6**, **Figure 7B**, and **Figures 8B–D**). This allows us to discuss the significance of the OIB and EMI signatures of the Pleistocene studied dike in the intraplate geodynamic context of NE China. Seismic tomography studies show that there is a high velocity anomaly in the mantle transition zone (MTZ) below NE China at 410–660 km depth, which is interpreted as the slab of the Pacific plate in the MTZ, and a low velocity body above the flat portion of the slab interpreted as a big mantle wedge (BMW) (Li and Hilst, 2010; Tian et al., 2016; Wang et al., 2020). From the middle Cenozoic, the subducted plate gradually subsided into the MTZ and partly melts. The resulting water-rich melt and the slab-derived fluids may metasomatize the overlying mantle. However, the EMI signature of the studied CMVF dikes cannot be explained by slab-derived sediments and associated metasomatizing fluids (Xu et al., 2012) because of their moderate Sr and low Pb isotopic ratios. In accordance with the geochemical signature of the gas

emitted in NE China (Xu S. et al., 2013b) and of the Cenozoic basalts (Zhang et al., 2009), the EM1 source could be the subcontinental lithospheric mantle (SCLM) (**Figure 10D**; Zhu et al., 2011). Regarding the source of the late Cenozoic initial OIB magma in the continental margin of NE China, this source could be a response to the back-arc extension and rifting caused by the roll-back of the Pacific plate (Li and Hilst, 2010; Zheng et al., 2018). Some geophysical studies suggest, however, that the Pacific slab is broken and an OIB-type mantle plume upwells from the lower mantle crossing the slab and entering MTZ (Tang et al., 2014). This model is not supported by shear-wave splitting studies (Lynner and Long, 2014) which show trench-parallel sub-slab fast splitting directions and not, as expected for a mantle plume from the lower mantle, a vertical direction. Based on geochemical data, Kuritani et al. (2009) propose that a back-arc type rifting associated to the slab retreat and



progressive flattening may favor the decompression melting of the asthenosphere. This model could explain the OIB signature of the CMVF basalts including the trachybasaltic dikes studied here. In this scenario, the CMVF OIB melts upwell from the asthenosphere to the lithosphere, where they mix to SCLM melts. These melts accumulate at the base of the lower crust in the early Pleistocene, where they may assimilate lower crust material of the ancient Asia continent (Figure 8A and Figure 10A). However, assimilation processes could affect the mantle melts during their upwelling and storage in the crust. Tomographic data in NE China (Zhu et al., 2019) clearly show a thickening of the Moho due to the accumulation of mantle melts at the base of the crust below CMVF. These data suggest active underplating process where these mantle melts may promote the melting of the lower crust. Therefore, we propose that most of the assimilation of crustal material underwent by the CMVF trachybasalts occurs at the Moho depth and not during their ascent and storage in the crust.

## CONCLUSION

Our results indicate that trachybasaltic dikes emplaced in the Calabrian stage of Pleistocene at 1.19–1.20 Ma, i.e., in a time

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period marked by the transition from a monogenetic-type volcanism to the Tianchi cone construction, central-type activity. The Calabrian volcanism is associated to the emplacement of scoria cones whose shallower feeding systems are controlled by regional tectonics and, in particular, by NW-SE striking structures. Geochemical and Sr-Nd-Pb-Hf isotopic data show that the trachybasaltic dikes have a potassic affinity and their parental magma is constituted by an OIB-type asthenospheric source with an EM I signature related to SCLM melts associated to back-arc and rifting processes linked the slab retreat (Figure 11). The magma upraising from the mantle stops at the Moho and assimilates Mesoproterozoic lower crust material by underplating (Figure 11).

## DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article; further inquiries can be directed to the corresponding author.

## AUTHOR CONTRIBUTIONS

ML: formal analysis, and writing-original draft. ZX: investigation and data curation. GV: writing-review and editing. XP: project administration and formal analysis. DH: supervision and writing-review and editing. GG: writing-review and editing. DY: investigation and formal analysis. BP: data and formal analysis. JF: drafting the work and revising manuscript critically.

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