



U-Pb Ages and Hf Isotope of Zircons from a Carbonatite Dyke in the Bayan Obo Fe-REE Deposit in Inner Mongolia: its Geological Significance

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Abstract: Detailed studies on U-Pb ages and Hf isotope have been carried out in zircons from a carbonatite dyke associated with the Bayan Obo giant REE-Nb-Fe deposit, northern margin of the North China Craton (NCC), which provide insights into the plate tectonic in Paleoproterozoic. Analyses of small amounts of zircons extracted from a large sample of the Wu carbonatite dyke have yielded two ages of late Archaean and late Paleoproterozoic (with mean ²⁰⁷Pb/²⁰⁶Pb ages of 2521±25 Ma and 1921±14 Ma, respectively). Mineral inclusions in the zircon identified by Raman spectroscopy are all silicate minerals, and none of the zircon grains has the extremely high Th/U characteristic of carbonatite, which are consistent with crystallization of the zircon from silicate, and the zircon is suggested to be derived from trapped basement complex. Hf isotopes in the zircon from the studied carbonatite are different from grain to grain, suggesting the zircons were not all formed in one single process. Majority of $\varepsilon_{\text{Hf}}(t)$ values are compatible with ancient crustal sources with limited juvenile component. The Hf data and their T_{DM2} values also suggest a juvenile continental growth in Paleoproterozoic during the period of 1940–1957 Ma. Our data demonstrate the major crustal growth during the Paleoproterozoic in the northern margin of the NCC, coeval with the assembly of the supercontinent Columbia, and provide insights into the plate tectonic of the NCC in Paleoproterozoic.

Key words: crustal growth, zircon U-Pb age, Hf isotope, Bayan Obo, North China Craton

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1 Introduction

The North China Craton (NCC) has attracted attention in terms of its complex evolutionary history during the Precambrian, especially for preserving important information of the Paleoproterozoic–Mesoproterozoic supercontinent Columbia (Kusky and Li, 2003; Zhai et al., 2003; Zhai and Liu, 2003; Zhai, 2004; Zhao et al., 2004; Zhao et al., 2005; Peng et al., 2010; Santosh et al., 2010; Kusky, 2011; Zhai and Santosh, 2011; Zhai et al., 2011; Peng et al., 2012; Zhao et al., 2012; Zhao and Zhai, 2013; Peng et al., 2014; Zhai et al., 2014b; Guo et al., 2015; Santosh et al., 2015; Wang et al., 2015; Shi and Zhao, 2017) since it was proposed (Rogers and Santosh, 2002; Zhao et al., 2003). The consensus is that the NCC is formed by amalgamation of a number of micro-continental blocks (one opinion was shown in Fig. 1a, after Zhao et al., 2012). However, when and how these blocks were assembled remain controversial, and a variety of models had been proposed for the tectonic subdivision and amalgamation of the craton (Kusky and Li, 2003; Zhai and Liu, 2003; Zhai et al., 2005; Zhao et al., 2005; Santosh et al., 2006; Faure et al., 2007; Kusky et al., 2007; Trap et

al., 2007; Santosh et al., 2010; Zhai et al., 2010; Kusky, 2011; Trap et al., 2012).

The NCC also demonstrates special characteristics, for example, it has experienced a complex geological evolution during Precambrian: multi-stage cratonization, Paleoproterozoic rifting–subduction–accretion–collision event, Late Paleoproterozoic–Neoproterozoic multi-stage rifting, and various and abundant mineralization (Zhao et al., 2005; Santosh et al., 2010; Zhai, 2010; Geng et al., 2012; Zhai and Santosh, 2013; Fan et al., 2014a; Zheng et al., 2014; Li et al., 2015; Wang et al., 2015; Wei et al., 2015; Yang and Santosh, 2015; Yuan et al., 2015; Zhang et al., 2015; Wang, et al., 2017; Zhang et al., 2018; Xiao et al., 2019). The fundamental architecture of the NCC was constructed through the rifting and subduction-accretion-collision tectonics during Paleoproterozoic (Santosh et al., 2015; Zhai and Zhu, 2016; Liu et al., 2018), produced a number of intra-continental rifts (Wan et al., 2003; Zhai and Liu, 2003; Lu et al., 2008; Meng et al., 2011; Zhai et al., 2014a), including the southern marginal Xiong'er rift, central-northern Yanliao rift and northern marginal Zhaertai–Bayan Obo–Huade rift system. However, tectonic settings during early Paleoproterozoic (2.5–1.9

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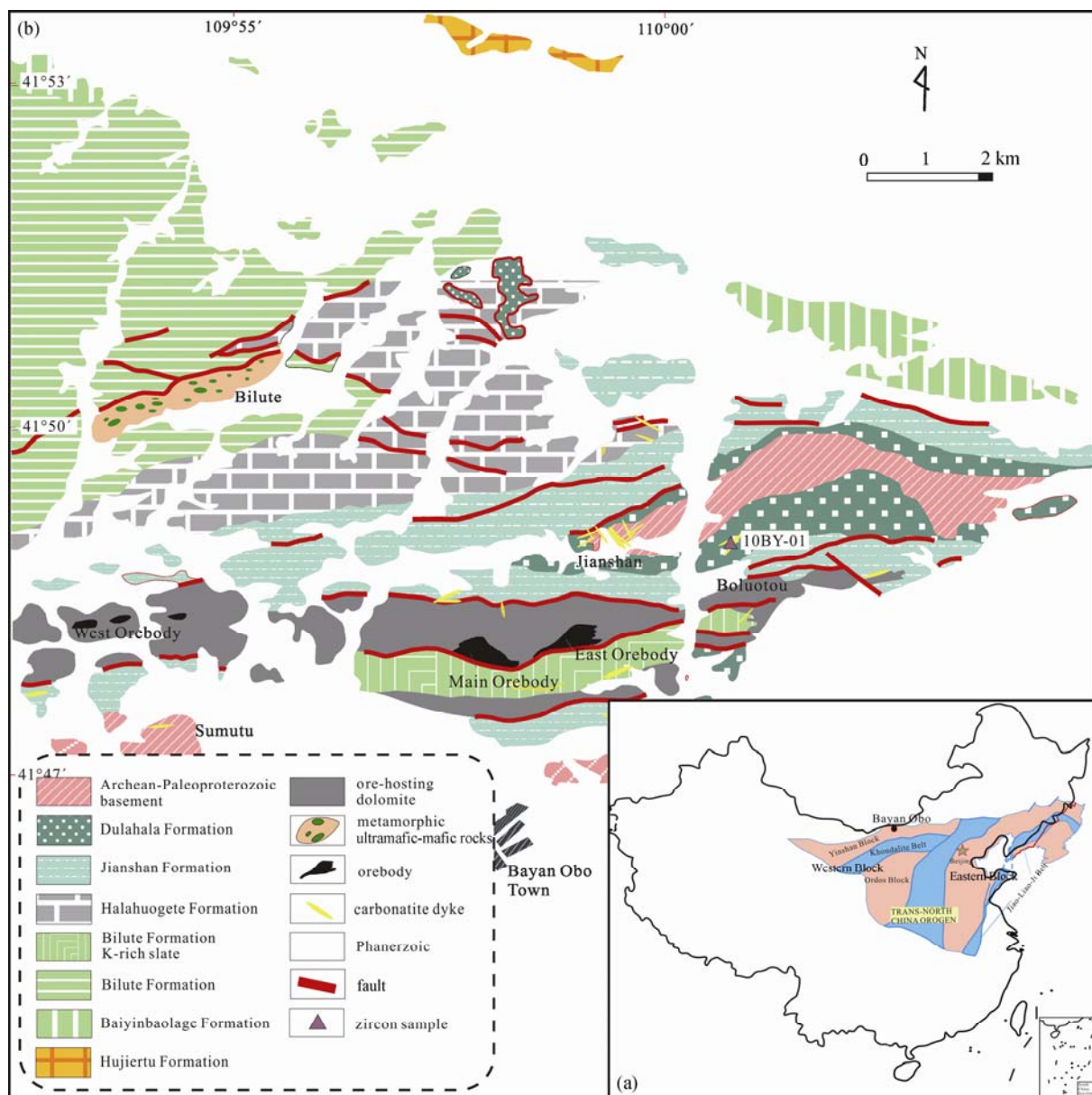


Fig. 1. Geological map of the Bayan Obo deposit.

(a) Generalized tectonic framework of the North China Craton showing the major crustal blocks and intervening suture zones (modified after Zhao et al., 2005); (b) geological map of the Bayan Obo region (after Yang et al., 2011) and locations of the carbonatite. China basemap after China National Bureau of Surveying and Mapping Geographical Information.

Ga), such as when the pulling-apart stage, subduction and collision occurred, were still unresolved.

Located in the northern marginal basin of the NCC, the regional geological setting and tectonic environment of the Bayan Obo area have attracted wide attention as it hosts the world's largest REE–Nb ore deposit. However, most researchers focused on its metallogenic setting, few studies had carried out on evolutionary history of the basement (Conrad and McKee, 1992; Yuan et al., 1992; Cao et al., 1994; Wang et al., 1994; Liu et al., 2004; Hu et al., 2009; Yang et al., 2009; Campbell et al., 2014; Fan et al., 2014b; Ma et al., 2014; Lai et al., 2015; Smith et al., 2015; Zhu et al., 2015; Lai et al., 2016). According to limited results, basement in the northern margin of the

NCC mainly formed during ca. 1900–2500 Ma (Zhang et al., 2000; Wang et al., 2001; Wang et al., 2002b; Jian et al., 2005; Fan et al., 2010; Zhong et al., 2015), and geochronological and geochemical studies only resulted in a preliminary conclusion that the collision orogenic movement of north margin of the NCC between 2.0 Ga and 1.9 Ga had brought the swarm of diorite-granodiorite magma and intense regional peak metamorphism event (Fan et al., 2010).

There were dozens of carbonatite dykes occurred in the Bayan Obo region (Tao et al., 1998; Fan et al., 2016), and the carbonatite dykes intruded during ca. 1300–1400 Ma (Yang et al., 2011; Fan et al., 2014b), which were considered to have a close relationship with the REE

mineralization (Yang et al., 2000; Yang and Le Bas, 2004; Yang et al., 2011; Lai et al., 2012; Ling et al., 2013). However, zircons in the Bayan Obo carbonatite dykes did not crystallize from carbonatite magma at the time of dyke emplacement, but were considered to be derived from wall rock contamination (Fan et al., 2006; Liu et al., 2008). The ages obtained from the zircons are consistent with ages measured on basement rocks elsewhere in the Bayan Obo region (Fan et al., 2010). Moreover, ages of these zircons ranged from 1.8 Ga to ~2.5 Ga, provided a possible approach to discuss evolutionary history in northern margin of the NCC during Paleoproterozoic.

In this contribution, zircons from the Bayan Obo carbonatite dike (WU-dyke) were dated by SIMS U-Th-Pb, and then analyzed for Hf isotopes, in order to determine the origin of the zircons and thus their geological significance.

2 Geological Settings

The giant Bayan Obo REE-Nb-Fe deposit is located in Inner Mongolia, China, about 2 km north from Bayan Obo town (N41.8°, E109.9°, Fig. 1b). It lies at the northern margin of the NCC, bordering the Paleozoic Central Asian Orogenic Belt to the north (Xiao and Santosh, 2014). The whole deposit is hosted in Palaeo- to Mesoproterozoic sediments of the Bayan Obo Group (Nie et al., 2002; Lai et al., 2016), which unconformably overlies the basement and consists of low grade sandstones, siltstones, slate, limestones and dolomites (Institute of Geochemistry, 1988; Drew et al., 1990; Bai et al., 1996; Zhong et al., 2015; Yang et al., 2017). The Bayan Obo Group can be divided into several formations: Dulahala Formation (H1-H2), consisting of quartzite, quartz sandstone and slate; Jianshan Formation (H3-H5), consisting of slate, quartzite, quartz sandstone and minor limestone; Halahuoqite Formation (H6-H8), consisting of quartz sandstone, dolomite, limestone and minor slate; Bilute Formation (H9-H10), consisting of slate with sandstone intercalations; Baiyinbaolage Formation (H11-H13), consisting of sandstone, siltstone and slate; and Hujertu Formation (H14-H18), consisting of limestone, quartz sandstone and quartzite, from lower to upper parts (Fig. 2).

Igneous rocks in the mine area include Proterozoic anorogenic igneous rocks (Wang et al., 2003), Phanerozoic granite (Ling et al., 2014), and a series of carbonatite dykes (Tao et al., 1998), the age of which was suggested to be 1.3–1.4 Ga (Yang et al., 2011; Zhu et al., 2015; Zhang et al., 2017). The ore bodies are hosted mainly in the H8 dolomite marble, which is 18 km long from east to west and 2–3 km wide from north to south, with a characteristic layered structure. Based on the wide distribution of fenitization and geochemical studies, carbonatitic dykes were considered to be closely associated with mineralization (Chao et al., 1992; Le Bas et al., 1992; Smith et al., 2000; Fan et al., 2004; Yang et al., 2009; Lai and Yang, 2013; Sun et al., 2013; Wang et al., 2019).

There are nearly 100 carbonatite dykes distributed in the Bayan Obo deposit (Fan et al., 2016), and they are small in scale, usually 0.5 to 4.0 meters wide and 6 to 200 meters long (Figs. 3a, 3b). Most of the dykes strike to the

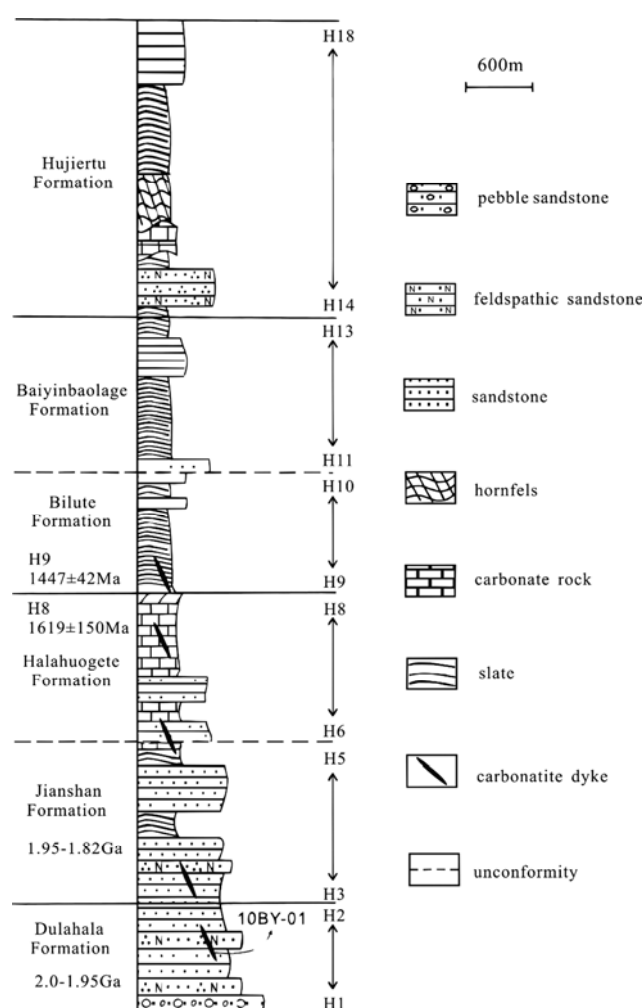


Fig. 2. Strata column of the Bayan Obo Group in the Bayan Obo area, northern margin of the NCC (compiled from Bai et al., 1996 and Zhang et al., 2017).

Ages of Dulahala and Jianshan Formation were cited from Zhong et al., 2015; age of H8 was cited from Lai et al., 2016; age of H9 was cited from Liu et al., 2016.

northeast or northwest (Tao et al., 1998; Wang et al., 2002a), which intrude Paleoproterozoic metamorphic basement rocks in the Kuangou syncline and the Proterozoic Bayan Obo Group near the ore body (Fig. 1b). It is significant that these dykes have metasomatised the country rocks on their contact zones, producing fenitization (Figs. 3a–3d), characterized by the presence of sodic amphiboles, aegirine, albite and phlogopite (Le Bas et al., 1992). Carbonatite dykes were grouped into coarse-grained and fine-grained facies based on the grain size (Le Bas et al., 1997; Yang et al., 2003; Yang and Le Bas, 2004). In addition, carbonatite dykes were segregated based on mineralogical compositions: dolomite, coexisting dolomite-calcite and calcite types (Figs. 3e, 3f) (Tao et al., 1998; Yang et al., 2011; Fan et al., 2016).

Geochronologic studies of carbonatite near the Bayan Obo deposit have been carried out using various types of methods (Fig. 1b), and yielded a wide range of ages (see discussion below). Zircons selected from carbonatite dykes were derived from wall rock contamination (Fan et

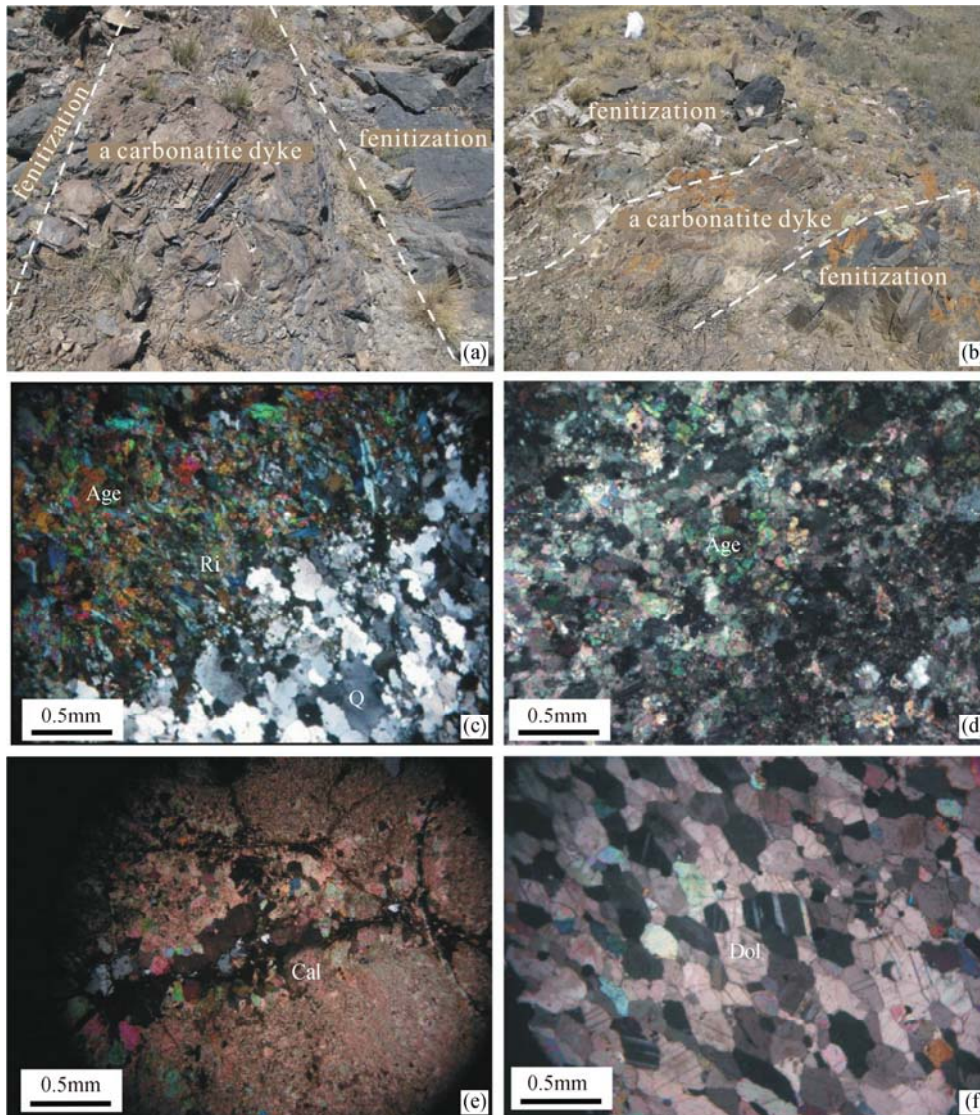


Fig. 3. Outcrops and photomicrographs of carbonatite dykes in the Bayan Obo region.

(a) No.1 carbonatite dyke (named Wu dyke by Le Bas, 1992) in the East Orebody, which caused strong fenitization in the contact rocks-quartz sandstone (H1-Jianshan Formation); (b) another carbonatite dyke in the Buoloutou; (c) fenitization in the contact rocks- quartz sandstone; (d) aegirine bearing calcite carbonatite (Wu-dyke) close to fenitization; (e) calcite carbonatite (Wu dyke) with calcite vein; (f) dolomite carbonatite in the Buoluolou. abbreviations: Aeg-aegirine; Cal-calcite; Dol-dolomite; Q-quartz; Ri-riebeckite.

al., 2006; Liu et al., 2008). The ages, obtained from the zircons ranging from 1.8 Ga to ~2.5 Ga, are consistent with ages measured on basement rocks elsewhere in the Bayan Obo region (Fan et al., 2010). Since the most detailed studies were carried out on the No.1 carbonatite dyke in the Bayan Obo, the carbonatite dyke studied here (10BY-01) was collected from Dulahala, which is also known as the Wu-dyke (Le Bas et al., 1997).

3 Sampling and Analytical Methods

3.1 Zircon U-Pb analysis

The carbonatite (Wu-dyke) sample for zircon analysis was collected from Dulahala, northeast of the Eastern iron ore-body, which intruded into the H1 formation of the Bayan Obo Group (Fig. 1b). The dyke metasomatised the

country rocks on their contact zones, producing fenitization (Figs. 3a–3d). Carbonatite studied here were calcite carbonatite, one type of carbonatite classified by Yang et al. (2011). It is dark grey in color, dense massive (Figs. 3a, 3b), and fine-grained under microscopic observation (Figs. 3d, 3e). Mineral compositions of the carbonatite dyke are mostly calcite or dolomite (Figs. 3e, 3f), as well as some aegirine and calcite veins (Figs. 3d, 3e).

Zircon grains extracted from the carbonatite, with no obvious veins or fenite and together with zircon standard 91500, were mounted in epoxy mounts which then polished to section the crystals in half for analysis. All zircons were documented with transmitted and reflected light photographs as well as cathodoluminescence (CL) images, to reveal their internal structure, and the mount

was vacuum-coated with high-purity gold prior to secondary ion mass spectrometry (SIMS) analysis.

Measurements of U, Th and Pb were conducted using the Cameca IMS-1280 SIMS at the Institute of Geology and Geophysics, Chinese Academy of Sciences in Beijing. U-Th-Pb ratios and absolute abundances were determined relative to the standard zircon 91500 (Wiedenbeck et al., 1995), analyses of which were interspersed with those of unknown grains, using operating and data processing procedures similar to those described before (Li et al., 2009; Li et al., 2010). Uncertainties on individual analysis in data tables are reported at a 1σ level; mean ages for pooled U/Pb (and Pb/Pb) analyses are quoted with 95% confidence interval. Data reduction was carried out using the isoplot v3.0 program (Ludwing, 2003).

3.2 Zircon Hf isotope analysis

After Zircon U-Pb isotope measurement, the insitu analyses of Hf isotopes were conducted at the State Key Laboratory of Continental Dynamics in Northwest University, Xi'an, using a Nu Plasma HR multiple collector inductively coupled plasma mass spectrometer (MC-ICP-MS), which equipped to a GeoLas2005 193 nm excimer ArF laser-ablation system. The instrumental conditions and data acquisition were described in detail in Yuan et al. (2008). In calculation of $\varepsilon_{\text{Hf}}(t)$, the recommended decay constant value of ^{176}Lu is $1.867 \times 10^{-11} \text{ y}^{-1}$, the $^{176}\text{Lu}/^{177}\text{Hf}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ values of primitive mantle are 0.0384 and 0.28325 respectively (Bouvier et al., 2008).

4 Results

4.1 SIMS zircon U-Pb dating

The carbonatite sample yielded merely few zircons, which were dated by SIMS method (Fig. 4). Twenty analyses were performed on 20 zircons from carbonatite

sample 10BY-01 and the results are listed in Table 1 and the age data are plotted on concordia diagram in Fig. 5.

The zircons extracted from the carbonatite dyke are consisted of medium to coarse (70–200 μm diameter), clear, subhedral to euhedral, rounded prismatic grains. CL imaging has shown most grains possess simple growth texture (Fig. 4).

The U-Th-Pb analyses yield a range of U and Th concentrations (114–867 ppm and 42–328 ppm, with an average of 389 and 185 ppm, respectively), and Th/U ratios ranging from 0.13 to 1.48 (with an average of 0.57), similar to typical magmatic zircons (Hoskin and Black, 2000). U depletion is a characteristic for zircons from carbonatitic or kimberlitic systems, Th/U ratios in carbonatite are high, reaching up to 10000 (Belousova et al., 2002). High Th/U ratio is also the characteristic of H8 dolomite and zircons of carbonatite in the Bayan Obo (Ling et al., 2013; Zhang et al., 2017), therefore, zircons in this study did not come from H8 dolomite or carbonatite.

The U-Pb analyses yield concordant or nearly concordant results within analytical error, with the $^{207}\text{Pb}/^{206}\text{Pb}$ apparent ages ranging from 1848.9 to 2529.5 Ma (Table 1). Most zircons have typical oscillatory zones (Fig. 4), yielding a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1921 ± 14 Ma (Fig. 5), even in one relict core (spot 9, Fig. 4). Three other relict cores (spot 13, 18, 19, Fig. 4) have a relatively older age of 2521 ± 25 Ma (Fig. 5). All ages are similar to the age of basement rocks (Fan et al., 2010; Wang et al., 2001), indicating zircon inheritance.

4.2 Zircon Lu-Hf isotopic compositions

The Hf isotopic compositions are listed in Table 2 and Fig. 6. The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios range from 0.280908 to 0.281602, and the calculated $\varepsilon_{\text{Hf}}(t)$ values are from -15.9 to 2.0 (Table 2 and Fig. 6a). The positive initial $\varepsilon_{\text{Hf}}(t)$ values are consistent with derivation from a long-term depleted mantle source, whereas the negative initial $\varepsilon_{\text{Hf}}(t)$

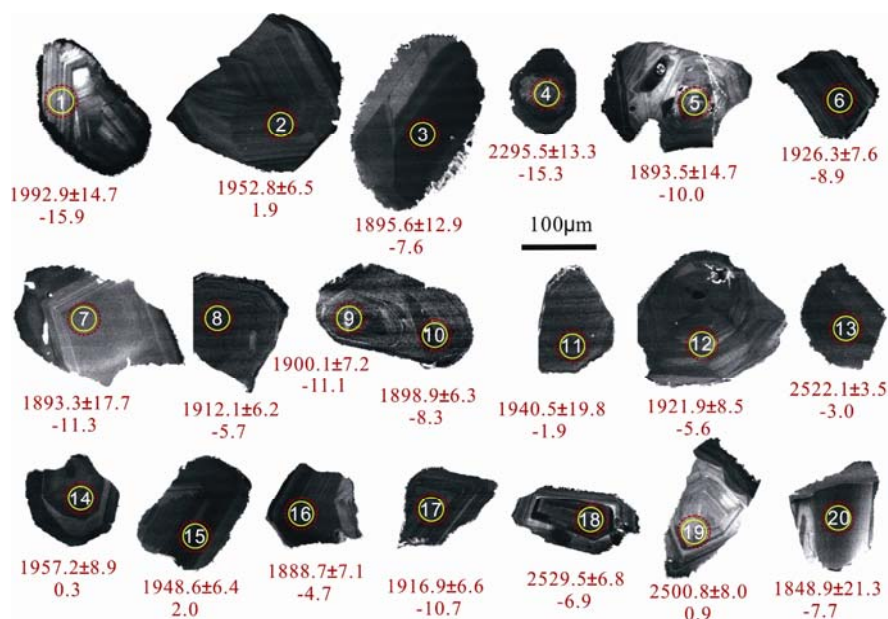


Fig. 4. Cathodoluminescence images of zircons in carbonatite from carbonatite. Analytical spots with U-Pb ages in Ma (yellow solid circle) and $\varepsilon_{\text{Hf}}(t)$ value (red broken circle) are also shown.

Table 1 SIMS zircon U-Pb isotopic data of the Wu carbonatite dyke from Bayan Obo

Sample/spot	Concentrations (ppm)				Isotopic ratios						Age (Ma)					
	U	Th	Pb*	Th/U	$^{207}\text{Pb}/^{235}\text{U}$	$\pm\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm\sigma$	$^{208}\text{Pb}/^{232}\text{Th}$	$\pm\sigma$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm\sigma$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm\sigma$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm\sigma$
10BY-01@1	129	191	71	1.48	6.10913	1.72	0.3617	1.51	0.10	8.00	1992.9	14.7	1991.6	15.1	1990.3	25.9
10BY-01@2	453	57	182	0.13	5.76163	1.54	0.3489	1.50	0.09	8.09	1952.8	6.5	1940.7	13.5	1929.3	25.1
10BY-01@3	124	95	55	0.77	5.37731	1.67	0.3362	1.51	0.09	8.04	1895.6	12.9	1881.2	14.4	1868.3	24.5
10BY-01@4	386	149	182	0.39	7.59783	1.69	0.3783	1.50	0.10	8.48	2295.5	13.3	2184.6	15.3	2068.5	26.6
10BY-01@5	177	150	79	0.85	5.35683	1.72	0.3353	1.51	0.09	8.07	1893.5	14.7	1878.0	14.8	1863.9	24.5
10BY-01@6	350	176	152	0.50	5.61490	1.56	0.3451	1.50	0.10	8.00	1926.3	7.6	1918.4	13.6	1911.1	24.9
10BY-01@7	196	151	88	0.77	5.38609	1.80	0.3372	1.50	0.09	8.01	1893.3	17.7	1882.6	15.5	1873.0	24.4
10BY-01@8	563	176	232	0.31	5.53676	1.54	0.3430	1.50	0.09	8.00	1912.1	6.2	1906.3	13.3	1901.0	24.8
10BY-01@9	660	273	276	0.41	5.45353	1.55	0.3401	1.50	0.09	7.98	1900.1	7.2	1893.3	13.4	1887.1	24.6
10BY-01@10	867	328	358	0.38	5.45584	1.54	0.3405	1.50	0.09	8.03	1898.9	6.3	1893.7	13.3	1888.9	24.6
10BY-01@11	114	42	49	0.37	5.78225	1.87	0.3526	1.50	0.10	8.13	1940.5	19.8	1943.8	16.3	1946.9	25.3
10BY-01@12	341	160	147	0.47	5.60771	1.58	0.3455	1.50	0.09	8.04	1921.9	8.5	1917.3	13.7	1913.0	24.9
10BY-01@13	775	321	469	0.42	10.87512	1.52	0.4739	1.50	0.13	7.97	2522.1	3.5	2512.5	14.2	2500.6	31.2
10BY-01@14	243	119	107	0.49	5.79331	1.59	0.3499	1.50	0.09	8.01	1957.2	8.9	1945.4	13.8	1934.3	25.2
10BY-01@15	515	246	225	0.48	5.77300	1.54	0.3504	1.50	0.09	8.01	1948.6	6.4	1942.4	13.4	1936.5	25.1
10BY-01@16	419	176	176	0.42	5.42557	1.55	0.3405	1.50	0.09	7.99	1888.7	7.1	1888.9	13.4	1889.1	24.6
10BY-01@17	540	270	232	0.50	5.54242	1.55	0.3424	1.50	0.09	7.99	1916.9	6.6	1907.2	13.4	1898.3	24.7
10BY-01@18	283	169	180	0.60	11.04300	1.56	0.4791	1.50	0.13	7.99	2529.5	6.8	2526.8	14.6	2523.4	31.5
10BY-01@19	149	178	104	1.20	10.72283	1.58	0.4732	1.51	0.12	8.08	2500.8	8.0	2499.4	14.8	2497.7	31.4
10BY-01@20	495	268	199	0.54	5.17257	1.92	0.3319	1.51	0.06	8.82	1848.9	21.3	1848.1	16.5	1847.4	24.3

Table 2 Zircon Hf isotope data for carbonatite in the Bayan Obo

No.	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm(2\sigma)$	Age (Ma)	$\epsilon_{\text{Nd}}(t)$	$\pm(2\sigma)$	$T_{\text{DM1}}(\text{Ma})$	$\pm(2\sigma)$	$f_{\text{LW/Hf}}$	$T_{\text{DM2}}(\text{Ma})$	$\pm(2\sigma)$
10BY-01@1	0.010037	0.000380	0.281083	0.000008	1992.9	-15.9	0.2	2972	11	-0.99	3577	18
10BY-01@2	0.000779	0.000025	0.281595	0.000006	1952.8	1.9	0.1	2264	8	-1.00	2461	14
10BY-01@3	0.012528	0.000492	0.281380	0.000011	1895.6	-7.6	0.2	2581	14	-0.99	3002	23
10BY-01@4	0.010160	0.000400	0.280908	0.000009	2295.5	-15.3	0.2	3207	12	-0.99	3770	19
10BY-01@5	0.017841	0.000688	0.281323	0.000010	1893.5	-10.0	0.2	2672	13	-0.98	3143	21
10BY-01@6	0.018387	0.000756	0.281334	0.000009	1926.3	-8.9	0.2	2662	12	-0.98	3104	19
10BY-01@7	0.014128	0.000548	0.281281	0.000011	1893.3	-11.3	0.2	2719	14	-0.98	3224	23
10BY-01@8	0.006330	0.000261	0.281414	0.000008	1912.1	-5.7	0.2	2521	11	-0.99	2899	18
10BY-01@9	0.013532	0.000568	0.281282	0.000015	1900.1	-11.1	0.3	2720	20	-0.98	3219	33
10BY-01@10	0.011326	0.000446	0.281357	0.000010	1898.9	-8.3	0.2	2610	14	-0.99	3047	22
10BY-01@11	0.008061	0.000307	0.281507	0.000008	1940.5	-1.9	0.2	2399	11	-0.99	2684	18
10BY-01@12	0.012018	0.000479	0.281420	0.000009	1921.9	-5.6	0.2	2527	12	-0.99	2899	19
10BY-01@13	0.017323	0.000767	0.281124	0.000008	2522.1	-3.0	0.2	2946	11	-0.98	3201	18
10BY-01@14	0.000533	0.000017	0.281545	0.000007	1957.2	0.3	0.1	2330	10	-1.00	2566	16
10BY-01@15	0.002594	0.000092	0.281602	0.000010	1948.6	2.0	0.2	2259	14	-1.00	2454	23
10BY-01@16	0.009351	0.000376	0.281462	0.000008	1888.7	-4.7	0.2	2464	11	-0.99	2819	18
10BY-01@17	0.014004	0.000584	0.281282	0.000009	1916.9	-10.7	0.2	2720	11	-0.98	3208	18
10BY-01@18	0.027572	0.001211	0.281034	0.000011	2529.5	-6.9	0.2	3104	16	-0.96	3439	25
10BY-01@19	0.008350	0.000386	0.281232	0.000007	2500.8	0.9	0.1	2774	10	-0.99	2943	16
10BY-01@20	0.013043	0.000552	0.281409	0.000011	1848.9	-7.7	0.2	2547	14	-0.98	2973	23

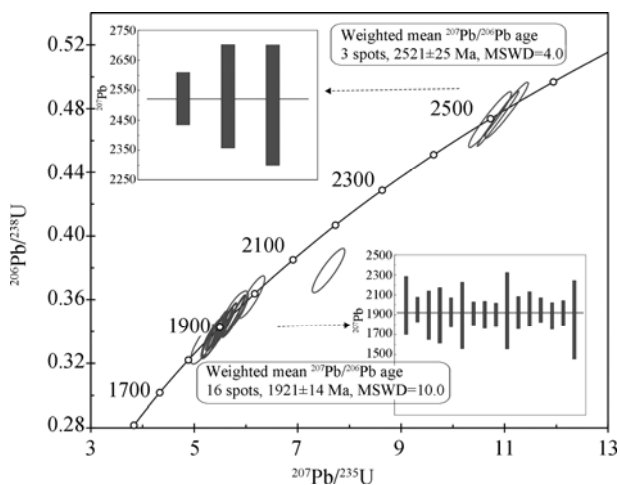


Fig. 5. Zircon concordia diagram of the Wu carbonatite in the Bayan Obo.

values could imply the involvement of crustal components into the mantle source (Jiang et al., 2006). The Hf model ages (T_{DM2}) are also listed in the Table 2 and Fig. 6b, ranging from 2454.0 Ma to 3769.9 Ma, most of them are older than the age of basement, few are consistent with the Pb-Pb ages, at ~2.5 Ga (Table 1).

5 Discussion

5.1 The age of carbonatite

Precise dating of the carbonatites is extremely important for understanding the evolution of the Zhaiertai-Bayan Obo-Huade rift system (Li et al., 2007; Yang et al. 2011; Zhong et al., 2015; Zhu et al., 2015). As shown in Table 3, previous geochronological studies on the carbonatite dykes and their minerals in the Bayan Obo region yielded a wide range of ages from ~2.1 Ga to ~1.2 Ga. Most researchers proposed that emplacement of carbonatite dykes in the Bayan Obo region occurred during the Mesoproterozoic period at ~1.3 Ga, which were probably

related to the final breakup of the Columbia Supercontinent (Yang et al., 2011; Zhong et al., 2015; Zhu et al., 2015; Zhang et al., 2017; Liu et al., 2018). Those younger ages (1.10–1.25 Ga) were probably related to the REE mineralized process (Lai et al., 2015).

It must be noted that zircon U-Pb dating of carbonatitic dyke is difficult, since it is difficult to separate zircon from carbonatitic samples due to the low contents of Zr and U

in the carbonatite (Amelin and Zaitsev, 2002; Rodionov et al., 2012; Campbell et al., 2014). In this case, the measured Th-Pb ages are more precise than the U-Pb ages (Amelin and Zaitsev, 2002; Rodionov et al., 2012). Th/U ratios of zircons in this study are far lower than typical carbonatite and Bayan Obo Th/U values, indicating an origin of basement contamination (Liu et al., 2008). More details are discussed below.

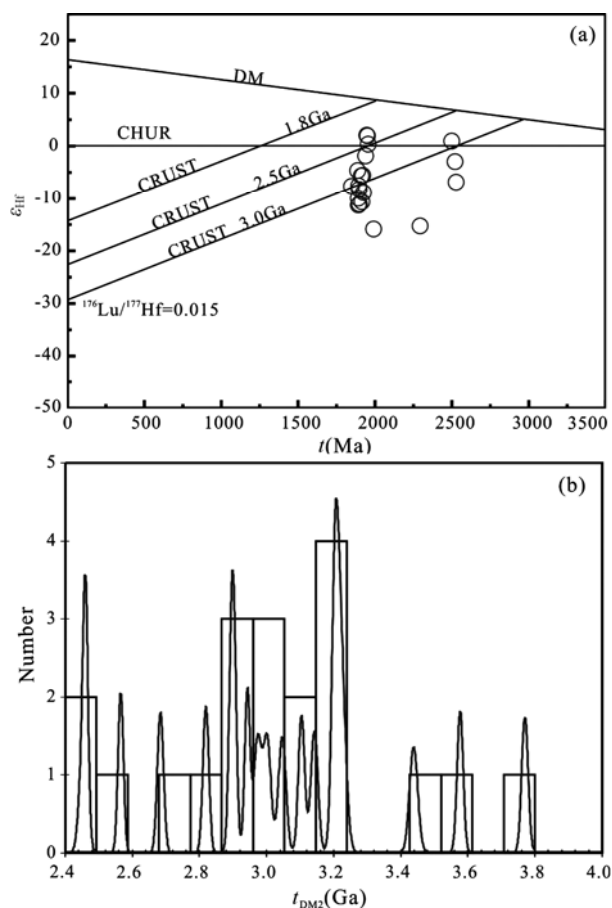


Fig. 6. Diagram of zircon Hf model ages (T_{DM2}) and $\epsilon_{Hf}(t)$ for slate in the Bayan Obo.

a- calculated $\epsilon_{Hf}(t)$ values; b- zircon two-stage Hf model ages (T_{DM2}).

5.2 The origin of the zircon

The age of 2070 ± 33 Ma for zircon from a carbonatite dyke was at first interpreted as the intrusion time of the carbonatite (Fan et al., 2002), then it has been re-interpreted that 2070 Ma is the age of the basement complex, not the age of the carbonatite dyke, since this age is much older than the Sm-Nd isochron ages and zircon is usually found near the margin of the intrusions (Le Bas, 2006). A concordant zircon age of 1925 ± 8 Ma and an upper intercept age of 1416 ± 77 Ma were obtained from the Wu carbonatite dike using U-Pb ID-TIMS. The age of 1925 ± 8 Ma was interpreted as the age of trapped wall rock, and the age of 1416 ± 77 Ma was interpreted as the intrusion age of the carbonatite (Fan et al., 2006). Although zircon is rare in carbonatite, it has been reported and studied from carbonatites in other places around the world (Heaman et al., 1990; Claesson et al., 2000; Amelin and Zaitsev, 2002; Rodionov et al., 2012), and sometimes zircon crystals in carbonatites can be huge and well-developed (Belousova et al., 2002).

Zircons from the Wu carbonatite dyke yielded a mean age of 1921 ± 14 Ma, as well as apparent age of 2521 ± 25 Ma for three older zircon relict cores (Table 1 and Fig. 5). Both two age groups are close to the previously published U-Pb age of 1894 ± 27 Ma and ~ 2.41 Ga for zircons from the same carbonatite dykes, which were considered to be derived from wall rock contamination (Liu et al., 2008). Basement complex at Bayan Obo, Inner Mongolia, are composed of Neoproterozoic mylonitic granite-gneiss (2588 ± 15 Ma), Paleoproterozoic syenite and granodiorite (2018 ± 15 Ma), and biotite granite-gneiss & garnet-bearing granite-gneiss (~ 1890 Ma) (Fan et al., 2010). In addition, detrital zircon ages of wall rock of carbonatite dykes (Jianshan, Dulahala and Halahuogete formation) were

Table 3 Summary of the published ages of carbonatites in the Bayan Obo region

Age (Ma)	Dating method	Dating object	Sample locality	References
1223	Sm-Nd isochron	Whole rock	Dlahala	Zhang et al., 1994
1176	Sm-Nd isochron	Whole rock	north of the Main orebody	Le Bas, 2007
1354	Sm-Nd isochron	Whole rock	Jianshan	Yang et al., 2011
1286	Sm-Nd isochron	Whole rock	Kuangou and north of the Main orebody	Zhu et al., 2015
1358	Sm-Nd isochron	Whole rock	data cited from published works	Lai et al., 2015
1260	Ar-Ar	Riebeckite in dyke	Northeast of the east orebody	Conrad and McKee, 1992
1236	Pb-Pb isochron	Whole rock	north of the Main orebody	Liu et al., 2006
1871	LA-ICP-MS U-Pb	Zircons in the carbonatite	Dongjielegele	Wang et al., 2002
1416	LA-ICP-MS U-Pb	Zircons in the carbonatite	Dulahala	Fan et al., 2006
2085	DTIMS U-Pb			
1934		Zircons in the carbonatite	north of the Main orebody	Liu et al., 2006
1984	SHRIMP U-Pb			
2035				
1418	SHRIMP U-Pb	Zircons in the carbonatite	Dulahala	Fan et al., 2014
1894	SHRIMP and			
1944	LA-ICP-MS U-Pb	Zircons in the carbonatite	Dulahala and Jianshan	Liu et al., 2008
1956				
1301	LA-ICP-MS Th-Pb	Zircons in the carbonatite	east and southeast to the East ore body	Zhang et al., 2017

consistent with the zircon ages of basement rocks in the Bayan Obo region (Yang et al., 2012; Zhong et al., 2015). Our zircon ages from carbonatite are consistent with the age of basement rocks, indicating an inheritable origin. Moreover, as mentioned above, U depletion is a characteristic for zircons from carbonatitic or kimberlitic systems, resulted in high Th/U ratios reaching up to 10000 (Belousova et al., 2002). Th/U ratios of zircon in this case range from 0.13 to 1.48, with an average of 0.57, which is far lower than typical carbonatite and Bayan Obo Th/U values (usually >200 for whole rocks and up to >10000 for monazite) (Claesson et al., 2000; Belousova et al., 2002; Liu et al., 2008; Ling et al., 2013; Zhang et al., 2017). Therefore, zircons for carbonatite in this study were not from H8 dolomite or carbonatitic magma, which is further supported by mineral inclusions studies in zircon by Raman spectroscopy (Fig. 7).

All evidence assembled above indicate that zircon grains from the Wu carbonatite dyke were derived from trapped basement complex. The U-Pb ages of 1921 ± 14 Ma and 2521 ± 25 Ma represent the ages of components within the basement rocks in the Bayan Obo region, which are consistent with the previous work (Liu et al., 2008). Interestingly, despite the age similarity, $\varepsilon_{\text{Hf}}(t)$ values in zircon from the studied carbonatite are different from grain to grain (Table 1, 2), suggesting that zircons were not formed by a single process.

5.3 Tectonic implications

The NCC behaved as a stable continent block since ~2.5 Ga cratonization throughout the assembly of micro-blocks

and reworking of old metamorphic basement (Zhai, 2014). The fundamental architecture of the NCC was constructed in the Paleoproterozoic through the collision of the major crustal blocks (Zhao et al., 2002; Zhai and Santosh, 2011; Santosh et al., 2013; Zhao and Zhai, 2013), subsequent to which continental rifting was initiated during ~2350–2000 Ma, similar to other cratons in the world (Condie and Kroner, 2008). The next major imprint in the NCC is the Paleoproterozoic orogenic events during 2.35–1.97 Ga which involved rifting followed by subduction-accretion-collision processes (Zhai and Santosh, 2013). Another crustal reworking took place in ~1.9–1.8 Ga, which was caused by the Paleoproterozoic Hutuo Movement (Zhai, 2014).

The Yinshan Block is located in the northern part of the Western Block, and comprised of a metamorphosed Archean basement and a Mesoproterozoic cover (Zhao et al., 2005; Jian et al., 2012). The Yinshan Block has witnessed major tectonic events including magmatism at ~2.5 Ga (Li et al., 2007; Liu et al., 2014; Zhang et al., 2014) and a series of magmatism during the period 2.1–1.8 Ga, when global scale collisional events led to the assembly of the Earth's first coherent supercontinent Columbia (Rogers and Santosh, 2002; Zhao et al., 2005; Santosh, 2010; Meert, 2012; Zhao et al., 2012; Nance et al., 2014).

In the northern margin of the Yinshan Block, a subordinate magmatism has been recorded in the Zhaertai-Huade-Bayan Obo Group in the period of 2.15–1.7 Ga, with a peak age of ~1.9 Ga (Wang et al., 2003; Li et al., 2007; Fan et al., 2010; Liu et al., 2014; Ma et al., 2014;

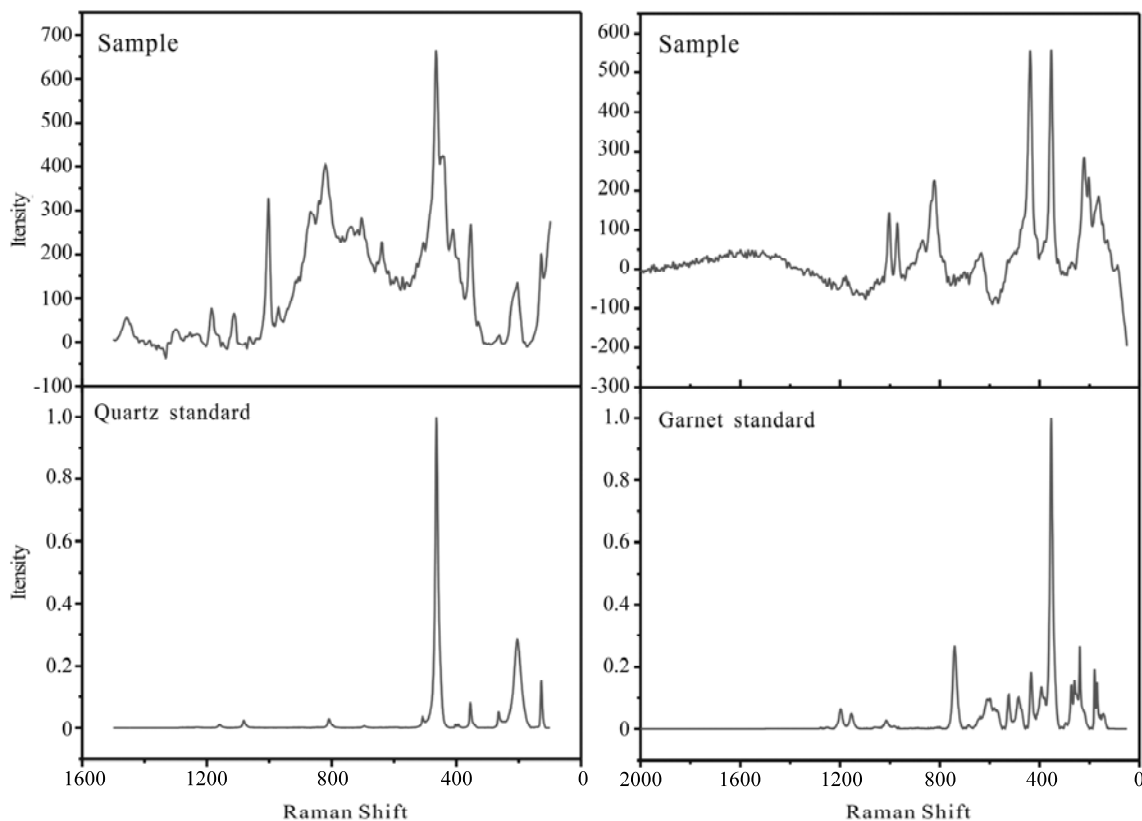


Fig. 7. Raman spectrum of mineral inclusions for spot 5 in zircon from Bayan Obo Wu carbonatite dyke.

Zhong et al., 2015). Bayan Obo is located on the north margin of the Yinshan Block, and the zircon ages obtained here are consistent with this local geological background. Two ages obtained here (2521 ± 25 Ma and 1921 ± 14 Ma) are similar to the age of basement complex of the Bayan Obo (mylonitic granite-gneiss, with an age of 2588 ± 15 Ma, and biotite granite-gneiss & garnet-bearing granite-gneiss, with an age of ~ 1890 Ma, Fan et al., 2010).

The results of Hf isotope have indicated that $\varepsilon_{\text{Hf}}(t)$ values of zircons from Zhaertai-Bayan Obo-Huade Group and their basement rocks are dominantly positive, suggesting juvenile sources with limited crustal participation (Li et al., 2007; Liu et al., 2014; Ma et al., 2014; Liu et al., 2017), similar to the results elsewhere in the Yinshan Block (Ma et al., 2013; Zhang et al., 2014; He et al., 2017). However, unlike their U-Pb ages, Hf isotope results are much different from previous work, their dominantly negative $\varepsilon_{\text{Hf}}(t)$ values (low to -15.9 , Table 2) do indicate a crustal source. As mentioned above, zircon grains from the carbonatite dyke here were derived from trapped basement complex, the Hf isotopic characteristics provide an approach to discuss evolutionary history in northern margin of the NCC during Paleoproterozoic.

There are four groups of data according to their age, $\varepsilon_{\text{Hf}}(t)$ value and crustal residence age (T_{DM2}) (Table 1 and Table 2). Group 1, ranging from 2529.5 Ma to 2500.8 Ma with $\varepsilon_{\text{Hf}}(t)$ values of -6.9 to 0.9 (with an average of -3.0) and T_{DM2} of 3439 Ma to 2943 Ma, represents the cratonization of the NCC at the end of Neoproterozoic by recycle of juvenile crustal rocks, accompanied by voluminous intrusion of crustally-derived granitic melts leading to the construction of the basic tectonic framework of the NCC (Zhai et al., 2005; Li et al., 2018; Liu et al., 2018). Group 2, ranging from 2295.5 Ma to 1992.9 Ma with $\varepsilon_{\text{Hf}}(t)$ values of -15.3 to -15.9 (with an average of -15.6) and T_{DM2} of 3439 Ma to 2943 Ma, represents the Paleoproterozoic orogenic imprint (Zhai and Santosh,

values indicate protoliths were dominantly crustal recycling. Group 3, ranging from 1957.2 Ma to 1940.5 Ma with $\varepsilon_{\text{Hf}}(t)$ values of -1.9 to 2.0 (with an average of 0.56) and T_{DM2} of 2454 Ma to 2684 Ma, was similar to the time that the amalgamation of the Yinshan Block and the Ordos Block along the Khondalite Belt forming the Western Block (Zhao et al., 2005; Santosh, 2010), and positive $\varepsilon_{\text{Hf}}(t)$ values indicate protoliths were mainly from juvenile components (Santosh et al., 2015; Shi et al., 2018; Song et al., 2018; Sun et al., 2018a,b; Zhou et al., 2018), probably in a subduction tectonics. Group 4, ranging from 1926.3 Ma to 1848.9 Ma with $\varepsilon_{\text{Hf}}(t)$ values of -4.7 to -11.3 (with an average of -8.3) and T_{DM2} of 2819 Ma to 3224 Ma, was broadly coeval with the global assembly of the Earth's first coherent supercontinent Columbia (Rogers and Santosh, 2002; Zhao et al., 2003; Wu et al., 2018). Negative $\varepsilon_{\text{Hf}}(t)$ values indicate a convergence process by ancient crustal recycling with asthenospheric injection related to slab-break off (Santosh et al., 2012).

According to the previous work, the NCC have experienced a major phase of continental growth at ca. 2.7 Ga, the amalgamation of micro-blocks and cratonization at ca. 2.5 Ga, and Paleoproterozoic rifting and subduction-

accretion-collision tectonics during ca. 1.95–1.82 Ga accompanied the subduction-collision process and suturing of continental blocks within the Paleoproterozoic supercontinent Columbia (Zhai and Zhu, 2016, and references therein). Our studies support the theory that cratonization of the NCC took place at the end of Neoproterozoic, and the similarity of the single stage Hf model ages and the two-stage model ages (Group 1, Table 2) implied that the protoliths were juvenile crustal rocks. Our studies are also consistent with the opinion that the NCC experienced a major collision process during the Paleoproterozoic (~ 2300 – 1970 Ma) (Zhai and Santosh, 2013; Zhai and Zhu, 2016). However, in addition to the existing models of juvenile continental growth in the NCC, there were two major pulses at 2.7 and 2.5 Ga, with only crustal recycling thereafter (Wan et al., 2011; Geng et al., 2012; Zhai, 2014), another juvenile crustal generated during 1957.2–1940.5 Ma, according to their positive $\varepsilon_{\text{Hf}}(t)$ values (with an average of 0.56) and similar T_{DM1} and T_{DM2} (Table 2), probably in a subduction tectonics. These reveal that the extent of crust generation and recycling through time during Archean and Paleoproterozoic still request more constraint, and multiple subduction zones might have generated during the assembling of Columbia (Santosh et al., 2015). Afterwards, a large area of khondalite series and S type granites have developed due to the formation of Inner Mongolian-north Hebei orogenic belt (Li et al., 2000; Wan et al., 2006; Zhong et al., 2007; Yin et al., 2011; Dong et al., 2013; Chen et al., 2017; Zhang et al., 2019), and the evolution of the NCC into a stable continental platform was finally accomplished.

It is suggested that the NCC experienced a subduction-collision orogenic process after ~ 2.5 Ga cratonization. A detailed study on various zircons of basement complex in the Bayan Obo area along the northern margins of the NCC provides insights into the primeval plate tectonic in Paleoproterozoic, and also reveals Paleoproterozoic crustal growth from both crustal recycling and juvenile components through time.

6 Conclusions

SIMS U-Pb dating of zircon from a carbonatite dyke in Bayan Obo Fe-REE deposit yielded mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1921 ± 14 Ma and 2521 ± 25 Ma. Given that zircons derived from carbonatitic magma generally have high Th/U, the low to moderate Th/U values in this studied from the carbonatite dyke are inconsistent with crystallization from carbonatitic magma at the time of dyke emplacement, and zircons are suggested to be derived from trapped basement complex.

The Lu-Hf data on zircons varies greatly with dominantly negative $\varepsilon_{\text{Hf}}(t)$ values, suggesting ancient crustal sources with limited juvenile component. The old crustal residence ages indicate that melting of older components beneath the continent also contributed to the magma genesis. The Hf data also suggests a juvenile continental growth in Paleoproterozoic during the period of 1940–1957 Ma.

The northern margin of NCC has witnessed active plate tectonics and major crustal growth during Archean-

Paleoproterozoic. Detailed zircon studies provide insights into the plate tectonic in Paleoproterozoic, and also reveals Paleoproterozoic crustal growth from both crustal recycling and juvenile components through time.

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