The Cihai diabase in the Beishan region, NW China: Isotope geochronology, geochemistry and implications for Cornwall-style iron mineralization

Tong Hou a, Zhaochong Zhang a,*, M. Santosh a,b, John Encarnacion c, Meng Wang a

a State Key Laboratory of Geological Process and Mineral Resources, China University of Geosciences, Beijing 100083, China
b Division of Interdisciplinary Science, Kochi University, Kochi 780-8520, Japan
c Department of Earth and Atmospheric Sciences, Saint Louis University, 3642 Lindell Boulevard, St. Louis, MO 63108, USA

A R T I C L E   I N   P R E S S

Article history:
Received 10 June 2012
Received in revised form 26 February 2013
Accepted 18 March 2013
Available online xxxx

Keywords:
Cihai
Cornwall
Iron deposit
Beishan
Geochemistry
Tectonic setting

A B S T R A C T

Diabase dykes in Cihai, Beishan region, NW China are spatially and temporally associated with ‘Cornwall-type’ iron deposits. U–Pb dating of zircons from a diabase dyke using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) yields an age of 128.5 ± 0.3 Ma, indicating an Early Cretaceous crystallization age. Most of the diabases show low Mg-numbers, suggesting evolved magmas. The diabase dykes show typical ophitic or sub-ophitic textures, and are dominantly composed of phenocrysts of plagioclase (40–50%) and clinopyroxene (30–45%), with minor and varying amounts of biotite and hornblende (1–5%), and minor disseminated magnetite (~5%). Their mineralogy reflects magma differentiation under relatively low oxygen fugacity conditions. The diabase dykes are characterized by minor variation in SiO2 (44.67–49.76 wt.%) and MnO (0.14–0.26 wt.%), but show a marked range of Al2O3 (10.66–14.21 wt.%), total Fe2O3 (9.52–13.88 wt.%), TiO2 (0.66–2.82 wt.%) and relatively high MgO (4.87–9.29 wt.%) with an Mg# value [atomic Mg/(Mg + Fe2+)] of up to 66. The Cihai diabases possibly experienced fractional crystallization of olivine + clinopyroxene and minor crustal contamination during the differentiation process. Prominent negative Nb, Ta and Ti anomalies suggest derivation from subduction-modified mantle. Furthermore, the rocks have relatively unradio genic Sr- and Nd-isotopic ratios. These characteristics probably reflect partial melting of a subduction component in the source mantle lithosphere through heat input from an upwelling asthenospheric mantle. Such processes probably occurred within an extensional setting during the Early Cretaceous in the Beishan area. The iron-rich fluids were derived from deep sources, and the iron ores were concentrated through a convection cell driven by temperature gradients established by the intrusion of the diabase sills. The combined processes of subduction-related enrichment in the source, shallow depth of emplacement, and the involvement of large-scale circulation of basinal brines from an evaporitic source are inferred to have contributed to the formation of the ‘Cornwall-type’ mineralization in Cihai.

© 2013 Elsevier Ltd. All rights reserved.

1. Introduction

Iron deposits are generally classified into four types, magmatic, structure-related, sedimentary, and metamorphic (Dill, 2010). The magmatic iron deposits can be further divided into five subgroups: (1) Fe–Ti–(V) deposits related to mafic intrusions; (2) apatite-bearing Fe oxide deposits; (3) apatite-bearing Fe oxide deposits related to alkaline igneous rocks; (4) volcanic-hosted Fe (unmetamorphosed) deposits; and (5) contact metasomatic Fe deposits (Fe skarn). Among these, the Fe skarn-type is one of the most important and major sources of high-grade iron ores (Einaudi et al., 1981). The igneous intrusions related to the Fe skarn deposits are mainly intermediate-acid in composition, with basic intrusions comprising a relatively small proportion. The Fe skarn deposits related to basic intrusions are recognized as ‘Cornwall-type’ deposits, from the typical example of the deposit at Cornwall, Pennsylvania. This deposit occurs at the contact between Triassic diabase and sedimentary rocks including carbonate rocks and biotite schist (Spencer, 1908; Eugster and Chou, 1979). Compared to other types of Fe skarn deposits, the ‘Cornwall-type’ deposits have received little attention after the 1980s (e.g., Rose et al., 1985), mainly because all of the mines closed by 1973, ending 241 years of continuous mining in the Cornwall region. Previous studies have revealed that the ‘Cornwall-type’ deposits are formed by metasomatic replacement (Eugster and Chou, 1979; Rose et al., 1985). However, although most workers agree that the ore-forming fluids are spatially and temporally connected with the diabase intrusion...
(e.g. Eugster and Chou, 1979), some important issues remain. For instance, some authors (Lapham and Gray, 1973; Eugster and Chou, 1979) have suggested that the iron for Cornwall-type ores was derived from diabase because basic rocks generally have high iron contents. However, analyses of diabase samples from a drill hole through the dyke beneath the Cornwall orebody did not support this conclusion as there was no evidence for the derivation of iron from the Cornwall diabases (Lapham and Gray, 1973; Smith, 1973; Smith et al., 1975). Thus, alternate models, based on oxygen isotope studies, proposed that much of the iron could have been derived from the host Triassic red beds (Rose et al., 1985). Furthermore, the tectonic setting for Cornwall-type deposit is also controversial with proposed models invoking island arc to continental rifting settings (cf. Smith et al., 1975). As most of the inferences on ‘Cornwall-type’ iron deposits were derived from the Cornwall deposit in Pennsylvania, it is important to test these models with investigations from other similar deposits. An ideal region for this is the Cihai high-grade iron ore deposit that occurs in the Beishan region in Xinjiang, NW China. The iron ores here are predominantly found in diabase and at the contact between diabase and wall rocks of primarily carbonate rocks and biotite felsic schist (Xue et al., 2000). This deposit provides a rare opportunity to re-examine the proposed ore-forming models for Cornwall-type deposits. The results from such a study may also be significant for addressing the interaction between basic magma and wall rocks, and the mechanism by which iron became concentrated.

In this study we report results from a detailed investigation of the Cihai iron ore deposit and present mineral chemical, bulk-rock major and trace element and Sr–Nd isotopic data of the diabase, coupled with zircon U–Pb geochronology. Our results constrain the nature of the mantle source and magmatic differentiation processes, which in turn provide some key constraints on the petrogenesis of diabase and its tectonic setting. The data presented in this study might also aid in exploring the yet unraveled ore bodies in this region.

2. Regional geology

The Beishan area is located between the Tarim basin and the southern side of the East Tianshan (Fig. 1a). The Tianshan–Tarim regions have been central to recent models related to the closure of the Paleo-Asian Ocean, as well as being relevant to large igneous provinces correlated to mantle plumes (e.g., Xia et al., 2012; Yang et al., 2012; Xiao et al., 2012). The East Tianshan consists mainly of three tectonic units: the Bogeda-Haerlike Belt in the north, the Jueluotage Belt in the center and Middle Tianshan Massif in the south (Fig. 1b; BGRMKXUR, 1993; Xu et al., 2009). The Beishan region exposes well-preserved Neoproterozoic to late Paleozoic sequences and early Paleozoic felsic rocks (Yue et al., 2001; Xu et al., 2009; Zhang et al., 2011; Mao et al., 2012). After the late Paleozoic, the region successively experienced Indosinian (Triassic), Yanshanian (Jurassic–Cretaceous) and Himalayan (Cenozoic) orogenies. Most of the sedimentary rocks are lacustrine deposits (e.g. Guo et al., 2005). During the Late Jurassic to Early Cretaceous, the Beishan region was characterized by an extensive shallow water lake with reduced clastic material input and intense...

Fig. 1. (a) Geologic map of the study area, showing the location of the Siberian, Tarim and Sino-Korean Cratons in relation to the Central Asian Orogenic Belt. (b) Regional geological map of the Beishan and Eastern Tianshan regions (modified after Su et al. (2011)).
evaporation. The evaporite sequences include limestone, marl, gypsum and salt (e.g. Wu et al., 2003; Chen and Wang, 2003; Zhang, 1981; Zhai, 2002).

The late Paleozoic tectonic history in this region is closely related to the complex subduction and closure of the South Tianshan Ocean (e.g., Gao et al., 2009; Yue et al., 2001; Zhang et al., 2002b; Xiao et al., 2009, 2012). The subsequent structural evolution of the Beishan region during Mesozoic to Cenozoic times can be divided into four stages (Guo et al., 2005): Jurassic extensional faulting and intracontinental down-warping, Early Cretaceous extensional subsidence followed by Late Cretaceous compressional uplift, the Oligocene to Miocene piedmont down-warping stage and the foreland stage during the Pliocene to Pleistocene. Mesozoic regional NNE-trending and Cenozoic regional NE-trending faults form the boundaries of the Beishan region, some of which are part of the system of continental-scale strike-slip faults of central Asia (Cunningham, 2005; Cunningham et al., 1996). Some of these large strike-slip faults are reactivated Paleozoic EW-trending faults, many of which are terrane boundaries or tectonic suture (Ao et al., 2010).

Many of the Permian mafic–ultramafic complexes discovered so far, which host magmatic Cu–Ni sulfide mineral deposits, are distributed in the western part of the Beishan region and intrude Proterozoic and Carboniferous strata (e.g., Su et al., 2011). In the Middle Tianshan Massif and Jueluotag Belt in the neighboring Eastern Tianshan area, Permian Cu–Ni-bearing mafic–ultramafic complexes occur extensively such as those of Tulaergen, Xiangshan, Huangshan, Tianyu and Baishiquan (e.g. Tang et al., 2012), and the Triassic Weiya mafic intrusion that hosts Fe–Ti–V-ores (e.g. Wang et al., 2008) (Fig. 2). Widespread Mesozoic A-type granite (Yin, 2008, 2010) and diabase dykes (Liu et al., 2010) have also been recognized in the Beishan region.

3. Mine geology

3.1. Stratigraphy and igneous rocks

Outcrops in the Cihai area are composed of basement of Neoproterozoic Qingbaikou Group unconformably overlain by Lower–Middle Permian Hongliuhe Formation and Upper Permian Luotuogou Formation (BGMRXJAR, 1993). The Neoproterozoic Qingbaikou Group consists of three lithologic sections from bottom upward. The first section is composed of banded marble, and carbonaceous marble intercalated with lenticular limestone. The second section is the major ore bearing horizon, consisting of biotite felsic schist, carbonaceous felsic schist and sericite schist intercalated with lenticular marble. Thick banded silicified marble intercalated with lenticular quartz comprises the uppermost section. The basement rocks are discordantly over lain by Lower Permian Hongliuhe Formation which is composed of Na-rich shallow-marine facies volcanic rocks and can be divided into lower and upper lithologic sections. The lower section is composed of intermediate-acid volcanic rocks, such as andesite, andesitic tuff and dacitic–andesitic breccias, and the upper section is dominantly basalt. The overlying Upper Permian Luotuogou Formation comprises mainly sandstone and silt, intercalated with lenticular limestones (Fig. 2). Four groups of faults trending NW, NS, NNE and EW have been recognized and the NNE trending faults control the distribution of diabases in the Cihai region. The EW and NW trending faults were formed during Paleozoic time and are contractional. Both these fault trends are subsidiary faults and are associated with the Xingxingxia and Liuyuan deep faults which are possibly related to the closure of the South Tianshan Ocean (e.g. Xiao et al., 2009, 2012) as stated above. Near the Cihai deposit, relatively small-scale NNE- and NS-trending faults were formed during the Mesozoic, related to the continent-scale strike-slip faults of central Asia (Cunningham, 2005; Cunningham et al., 1996).

The multiple pulses of dykes of diabase are emplaced into gabbro and Neoproterozoic strata (Fig. 3). These diabase dykes are closely associated with iron oxide mineralization. Late thin dykes of diabase cut across the previously emplaced diabase bodies as well as the iron ore bodies in some places (Fig. 4b and c). All the analyzed samples in this study were collected from the middle portion of the diabase dykes in the open pit to minimize any contamination or alteration. The analyzed diabase samples are dark-colored and show similar characteristics in texture, mineralogy and alteration, except sample CH05 which contains aggregates of diopside possibly formed through the strong skarn effect. These samples show typical ophitic or sub-ophitic textures, and consist
of plagioclase (40–50%), clinopyroxene (30–45%; augite and diopside), varying content of biotite and hornblende (1–5%), and minor disseminated anhedral magnetite crystals in the diabase occur as interstitial phases among the clinopyroxene and plagioclase, indicating that they crystallized from the magma, later than the clinopyroxene and plagioclase. Slightly altered clinopyroxene and labradorite rimmed by hornblende and fresh albite, respectively, is commonly present. Other types of weak alteration are also observed, such as epidotization, sericitization and kaolinization. These samples show extensive post-magmatic alteration in the aureoles around the ore bodies. Plagioclase is commonly replaced by sericite, whereas clinopyroxene is altered into hornblende, epidote, chlorite and biotite. In addition, some euhedral to subhedral granular pyrite occurring as veins have also been recognized.

### 3.2. Ore bodies

The Cihai iron deposit was discovered in the 1970s and hosts an estimated ore reserve of 100 Mt at 48% Fe (Zhao, 2000), with ongoing open cast mining. The iron deposit can be divided into three sections: Cihai, West Cihai and South Cihai (Fig. 2), all of which are covered by red beds of Tertiary Kushui Formation and Quaternary sediments (Fig. 4b). Magnetic surveys and drilling reveal that over 90% of the iron ore reserves identified so far are hosted in the Cihai section (Fig. 4a), whereas the other two sections are relatively small in scale (Tang et al., 2010). Therefore, in the following sections, we only focus on the Cihai section, where the iron ore bodies are intimately associated with intrusive masses of diabase (Wang et al., 2006) and most of them are developed at the contact of the diabase with Neoproterozoic strata comprising limestone (marble) and felsic schist.

Available mine data shows that the Cihai ore section is 1600 m in length and 300–500 m in width, with a depth of more than 900 m. Large masses of intrusive diabase occur adjacent to each of the ore bodies in immediate contact. The iron oxide ore occurs as large and small masses of varying form, either entirely enclosed by the diabases or within the rocks with which the diabases are in direct contact, such as biotite felsic schist (Fig. 3). The iron oxide ore bodies occur as fracture-fillings in the diabases and also at the contact with the host sedimentary rocks, forming more than 200 parallel ore bodies striking NNE and showing stratiform, lensoid, and vein shapes, with length and thickness ranging from 50 to 755 m and 1 to 20 m, respectively (Fig. 3). These are occasionally cut by late diabase dykes.

Four types of ores have been recognized in Cihai: massive, disseminated, banded, and brecciated. Among these, the most important economically are the massive ores and banded ores, which constitute the high grade orebodies. These high grade ore bodies are in general concentrated in the upper part of the deposit, whereas the low grade ores occur disseminated in the lower part. The massive ores contain magnetite, diopside, chlorite, hornblende and calcite, and minor sulfide. The key minerals in the orebodies crystallized in the sequence diopside, magnetite, sulfide and calcite. The disseminated ores typically occur in the margin where the ore bodies thin outward, and contain more gangue minerals such as garnet, diopside and chlorite. Also, the disseminated ores show sharp contact with the massive ores. The banded ores are only found in the west Cihai section, and the gangue mineral bands comprise garnet, diopside and hornblende. The brecciated ores mainly occur in the lower parts of ore bodies. The breccias are predominantly composed of the fragments of diabase and skarn assemblage which formed in the prograde skarn stage. Extensive skarn effects are also observed from the presence of considerable amount of garnet and diopside, forming alteration zones at the contact of the orebodies (Fig. 4f). However, these minerals are not restricted to the contact between ore and carbonated rocks, and are therefore different from the traditional skarns (e.g. Einaudi et al., 1981).

The most common ore mineral in the Cihai deposit is magnetite and the most common gangue minerals are garnet, diopside, tremolite, biotite, calcite, chlorite and quartz. In accordance with their primary mineral assemblages, these ores can be classified into three types: (a) anhydrous silicate-magnetite ore constituting the major ore type, including diopside-magnetite and garnet-diopside-magnetite; (b) hydrated silicate-magnetite ore assemblage comprising tremolite + diopside + magnetite, tremolite + calcite + magnetite, quartz + chlorite + magnetite and garnet + magnetite + tremolite; and (c) sulfide-bearing magnetite assemblage.
Fig. 4. (a) Overview of the Cihai open-cast mine. (b) Dykes of diabase intrude earlier diabase dykes. (c) The iron ores, cut by dykes of diabase. (d) Diabassic texture displayed in Cihai diabase. Cpx = clinopyroxene, Pl = plagioclase. (e) Porphyritic texture displayed in dykes of Cihai diabase. (f) Skarn minerals in altered diabase. Di = diopside, Ep = epidotite.
predominantly composed of high grade ores, such as pyrrhotite + chalcopyrite + magnetite, pyrite + magnetite and pyrrhotite + pyrite + magnetite (e.g. Xue et al., 2000). The magnetite grains are fractured and often replaced by the sulfide minerals. The presence of significant cobalt has been reported from pyrite, chalcopyrite, siegenite, safflorite, and cobaltite in previous studies, which are important by products of the primary iron mining enterprise (Xu et al., 2009).

4. Analytical methods

4.1. Zircon U–Pb dating

Zircons were separated from a diabase dyke by heavy liquid and magnetic techniques, and then handpicked under a binocular microscope. Zircon grains were mounted on adhesive tape then enclosed in epoxy resin and polished to about half of their diameter. In order to observe textures of the polished zircons, CL imaging was carried out using a Hitachi S3000-N scanning electron microscope (SEM) with Mono CL3 Cathodoluminescence System for high resolution imaging and spectroscopy at the Institute of Geology Chinese Academy of Geological Sciences.

Zircon U–Pb dating was performed by Finnigan Neptune multicollector ICP-MS with a Newwave UP213 laser-ablation system at Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing. Helium was used as the carrier gas to enhance the transport efficiency of the ablated material. The analyses were conducted with a beam diameter of 25 lm with a 10 Hz repetition rate and a laser power of 2.5 J/cm² (Hou et al., 2009). The masses 206Pb, 207Pb, 208Pb, 204(Pb + Hg) and 202Hg were measured by multi-ion-counters, while the masses 206Pb, 232Th, 238U and 238U were collected by Faraday cup. Zircon GJ1 was used as the standard and zircon Plesovice was used to optimize the machine. U, Th and Pb concentrations were calibrated using 29Si as the internal standard and zircon M127 (U: 923 ppm; Th: 439 ppm; Th/U: 2.05, Nasdala et al., 2008) as the external standard. 207Pb/206Pb and 206Pb/238U ratios were calculated using the ICPMSDataCal 4.3 program (Liu et al., 2008). The common-Pb was not corrected because of the high 206Pb/204Pb ratios (>1000). Data with abnormally high 206Pb counts were deleted. The zircon Plesovice is dated as un

4.2. Mineral chemistry

Mineral chemical analysis was performed by wavelength dispersive analysis (WDS) using standard procedures on the JXA-8200 electron microprobe at Washington University in St. Louis (MO, USA), using “Probe for Windows” for data reduction (see http://www.probesoftware.com/). Electron microprobe analyses were determined for clinopyroxene, plagioclase and magnetite, and the measured data were corrected using CITZAF after Armstrong (1995). Oxide and silicate standards were used for calibration (e.g. Amelia albite for Na, Si; microcline for K; Gates wollastonite for Ca; Alaska Anorthite for Al; synthetic fayalite for Fe; synthetic forsterite for Mg; synthetic TiO2 for Ti; synthetic Mn-olivine for Mn; synthetic Cr2O3 for Cr). Precision on oxide abundances is better than 1%.

4.3. Major and trace elements

After screening under the microscope, relatively fresh samples were selected and sawed into slabs and the central parts were used for whole-rock analyses. Specimens were crushed in a steel mortar and ground in a steel mill to powders of ~200 mesh. Major elements and ground were acquired through the analysis of fused glass discs using a scanning wavelength dispersion X-ray fluorescence (XRF) spectrometer at the Key Laboratory of Orogenic Belts and Crustal Evolution, Ministry of Education, School of Earth and Space Sciences, Peking University. The analytical uncertainties are less than 1%, estimated from repeated analyses of two standards (andesite GSR-2 and basalt GSR-3). Loss on ignition was determined gravimetrically after heating the samples at 980 °C for 30 min.

Trace elements were determined by solution ICP-MS performed at the ICP-MS Laboratory at the National Research Centre for Geoanalysis, Beijing. After complete dissolution, powders (~40 mg) were dissolved in distilled HF + HClO4 in 15 ml Savillex Teflon screw-cap breakers. Precision for most elements was typically better than 5% RSD (relative standard deviation), and the measured values for Zr, Hf, Nb and Ta were within 10% of the certified values. Two standards (granite GSR-1, basalt GSR-3) were used to monitor the analytical quality.

4.4. Sr and Nd isotopes

Rb–Sr and Sm–Nd isotopic compositions were obtained using a Finnigan Triton TI mass spectrometer at the Department of Earth Sciences, Nanjing University. Sr and Nd isotopic ratios were normalized against the Nd/Sm ratio of 0.1194 and 146Nd/144Nd = 0.7219, respectively. 87Sr/86Sr for the NIST87 Sr standard was 0.710268 ± 0.000007 (2σ standard deviation, n = 8), and 146Nd/144Nd for the La Jolla Nd standard was 0.511842 ± 0.000006 (2σ standard deviation, n = 6). Total blanks were 100 pg for Sr and 80 pg for Nd, and negligible for the determination of isotopic compositions.

5. Results

5.1. LA-ICP-MS U–Pb dating

Zircon grains separated from the Cihai diabase dyke are generally transparent, euhedral, and prismatic, with clear oscillatory zoning (Fig. 5) and high Th/U ratios (0.77~1.23, Table 1), typical of igneous zircons (Corfu et al., 2003). A group of 13 zircon grains define a concordia age of 128.5 ± 0.3 Ma (Fig. 6), interpreted as the crystallization age for the Cihai diabase dyke. The age from one spot (CH2–5) shows large error, correlated to the unusually low 235U and 232Th contents of the spot (Table 1).

5.2. Mineral chemistry

The chemical compositions of representative clinopyroxene, plagioclase and magnetite from the Cihai samples are listed in Tables 2–4, respectively. Phenocryst clinopyroxene from the Cihai diabase dykes shows Mg# values [Mg/(Mg + Fetot)] in atoms per formula unit of 0.80–0.83, variable Al2O3 and TiO2 contents (4.38–2.09 wt.% and 0.78–0.41 wt.%, respectively) and is relatively rich in Cr2O3 (up to 0.94 wt.%). The clinopyroxene phenocrysts show minor compositional variation with Wo41.56–41.54En45.09–47.80 Fs7.92–10.08. classifying as diopside. Compared to the phenocryst clinopyroxene, that from the groundmass [Wo46.91–48.54En34.27–39.29 Fs7.92–10.08] has lower Mg# value (0.78–0.70), and Al2O3 and TiO2 contents (0.23–0.53 wt.% and 0.03–0.12 wt.%, respectively). In particular, the Mg# and Al2O3 values of clinopyroxene show a rough positive correlation (figure not shown). Additionally, the clinopyroxene are characterized by significantly low Fe# and Fe2+/Fe3+ ratios. All analyzed plagioclases can be classified as labradorite; however, the An content of plagioclase from groundmass (An40–58) is lower than those from phenocrysts (An40–64). The magnetites in Cihai
iron ores contain FeO (91.58–93.55 wt.%), Al₂O₃ (0.77–1.62 wt.%) and Cr₂O₃ (0.04–0.10 wt.%), and are characterized by low TiO₂ content (0.04–0.12 wt.%; Table 4).

5.3. Bulk-rock major and trace elements

Geochemical data for the Cihai diabase are listed in Table 5. One of the samples (CH05) shows high concentration of CaO (21.33 wt.%), MnO (0.11–0.45 wt.%), and TiO₂ (0.62–2.82 wt.%), and a relatively large range of Al₂O₃ (8.34–14.21 wt.%), total Fe₂O₃ (14.25 wt.%), and MgO (4.87–9.29 wt.%). On the AFM and SiO₂ vs. FeO T/MgO diagrams, the Cihai samples display thorough leucitic affinity (Fig. 7a and b). On Harker diagrams, the Cihai diabases show broad negative correlations between SiO₂ and TiO₂, TFe₂O₃, total Fe₂O₃ (14.25 wt.%), and MgO (4.87–9.29 wt.%). On the primitive mantle-normalized incompatible element patterns, which are distinct from the volcanic rocks belonging to Mid-Oceanic Ridge Basalts (MORBs) and Oceanic Island Basalts (OIBs) typically characterized by no or insignificant HFSE anomalies (Sun and McDonough, 1989). Moreover, Zr and Hf are slightly enriched over Nb and Ta, which are in turn slightly enriched over Y and HREE.

5.4. Sr and Nd isotopes

Strontium and neodymium isotopic data from representative samples of the Cihai diabases are given in Table 6. The rocks show variable initial ⁸⁷Sr/⁸⁶Sr ratios (0.706–0.711), with relatively constant εNd(t) values (+2.7 to +6.1) which suggest a common source region. Moreover, the Sr isotopic compositions of the Cihai diabases exhibit positive correlation with LOI (loss of ignition) (not shown), suggesting that the Sr concentrations have been considerably affected by alteration, possibly through fluids (Sano et al., 2002), which is also consistent with the presence of secondary biotite in the diabases. Accordingly, the Sr–Nd isotopic ratios of our samples are plotted above the mantle array (Fig. 10a). Moreover, the εNd(t) values show slightly positive correlation with the MgO content (Fig. 10b), and the values are higher than those of the Weiyia V–Ti–Fe-bearing mafic intrusion (Wang et al., 2008). The plots overlap with the field of the Permian Ni–Cu-bearing ultramafic–mafic intrusions both in Beishan (Jiang et al., 2006; Ao et al., 2010; Su et al., 2011) and Eastern Tianshan (Tang et al., 2012; Chai et al., 2008; Zhou et al., 2004).

6. Discussion

6.1. Alteration effects

Before we use elemental and isotope compositions to discuss the petrogenesis, we evaluate the possible effects of hydrothermal alteration. The samples from the Cihai diabases are variably altered, as seen in thin sections, and this is reflected in, for example, their LOI (weight loss on ignition to 1100 °C) values, which range from 1.49 to 4.45 wt.%. In most cases, the high-field-strength elements (HFSE), rare earth elements (REE), Th and transition elements are immobile during the most intense hydrothermal alteration (e.g. Hawkesworth et al., 1997; Zhou, 1999). As will be shown below, the abundances of these elements in our rocks show coherent variation trends. Thus, in general, the Cihai diabases still preserve the signature of their original trace element compositions. In addition, some major elements such as Ti and P are not readily transported by hydrothermal alteration (e.g. Zhou, 1999), but Ca, Na, K, Al and some of the large ion lithophile elements (LILE; e.g. Sr, Ba and Rb) are considerably mobile (Smith and Smith, 1976).

Table 2

Representative analyses of clinopyroxene from Cihai diabases.

<table>
<thead>
<tr>
<th></th>
<th>CH-11-01</th>
<th>CH-11-02</th>
<th>CH-11-03</th>
<th>CH-11-04</th>
<th>CH-11-05</th>
<th>CH-11-06</th>
<th>CH-09-01</th>
<th>CH-09-02</th>
<th>CH-09-03</th>
<th>CH-09-04</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>51.68</td>
<td>53.46</td>
<td>52.38</td>
<td>52.77</td>
<td>52.20</td>
<td>51.67</td>
<td>54.01</td>
<td>53.11</td>
<td>53.86</td>
<td>53.42</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.72</td>
<td>0.41</td>
<td>0.77</td>
<td>0.64</td>
<td>0.74</td>
<td>0.77</td>
<td>0.78</td>
<td>0.03</td>
<td>0.07</td>
<td>0.12</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.14</td>
<td>0.15</td>
<td>0.19</td>
<td>0.17</td>
<td>0.17</td>
<td>0.17</td>
<td>0.21</td>
<td>0.21</td>
<td>0.25</td>
<td>0.24</td>
</tr>
<tr>
<td>MgO</td>
<td>16.05</td>
<td>17.04</td>
<td>16.15</td>
<td>16.68</td>
<td>15.82</td>
<td>16.57</td>
<td>14.21</td>
<td>12.27</td>
<td>13.49</td>
<td>12.45</td>
</tr>
<tr>
<td>CaO</td>
<td>21.11</td>
<td>21.40</td>
<td>20.81</td>
<td>21.00</td>
<td>21.45</td>
<td>20.26</td>
<td>23.62</td>
<td>23.54</td>
<td>23.96</td>
<td>24.05</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.37</td>
<td>0.32</td>
<td>0.32</td>
<td>0.34</td>
<td>0.33</td>
<td>0.31</td>
<td>0.33</td>
<td>0.52</td>
<td>0.23</td>
<td>0.30</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.03</td>
<td>0.09</td>
<td>0.02</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>Total</td>
<td>100.62</td>
<td>100.28</td>
<td>100.02</td>
<td>100.72</td>
<td>100.16</td>
<td>100.73</td>
<td>100.49</td>
<td>100.79</td>
<td>100.73</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 6. LA-ICP-MC U–Pb zircon concordia diagrams for Cihai diabase.
implying that their original concentrations were likely modified by alteration effects. Mg may also exhibit mobility during alteration (Zhou, 1999). Thus, these mobile elements will not be employed for petrogenetic interpretations in our work. Previous studies have shown that only severe alteration by interaction with seawater can affect the $\frac{143\text{Nd}}{144\text{Nd}}$ ratios (Mahoney et al., 1998) and the $\frac{87\text{Sr}}{86\text{Sr}}$ ratios can be markedly modified (e.g., Menzies et al., 2000). Thus, these mobile elements will not be employed for petrogenetic interpretations in our work. Previous studies have shown that only severe alteration by interaction with seawater can affect the $\frac{143\text{Nd}}{144\text{Nd}}$ ratios (Mahoney et al., 1998) and the $\frac{87\text{Sr}}{86\text{Sr}}$ ratios can be markedly modified (e.g., Menzies et al., 2000).

6.2. Petrogenesis: crustal assimilation vs. source enrichment

The similarity in petrography and major and trace elements compositions between Cihai diabase samples suggest that they are co-magmatic. The diabases have generally high MgO (max. $9.29\text{wt.\%}$; $\text{Mg}^\#$ = 66) contents, indicating a dominant magma source from the upper mantle. However, these rocks are also marked by "crustal-like" trace element and isotopic features such as the enrichment in large ion lithophile element (LILE) and light rare earth element (LREE), and depletion in the high field strength elements. The fractionation of Ti-bearing minerals is also supported by the correlation between MgO and $\text{FeO}_\text{tot}$, which are not observed (Figs. 8–10). Mohr (1987) suggested that mantle-derived magmas are subjected to some degree of contamination during their ascent and/or temporary residence in crustal magma chambers. Our samples exhibit low Ta/La ratios (0.02–0.04) (cf. Ta/La = 0.06 for primitive mantle; Wood et al., 1979), and relatively large variation in Nd isotopes and slightly positive correlation between MgO and $\text{Sm}_{\text{eq}}(t)$ values (Fig. 10b), indicating that they have been affected by crustal contamination. Nevertheless, the absolute concentrations of incompatible elements (including Nb, Ta, Zr, Sm, etc.) are much higher than those typically observed in crustal rocks (Taylor and McLennan, 1985). Their positive $\text{Sm}_{\text{eq}}(t)$ values and relatively constant Nb/La ratios with increasing $\text{SiO}_2$ contents (not shown) suggest that the amount of crustal contamination was not significant or minor. Therefore, we infer that the majority of geochemical and isotopic features of the Cihai diabases including their low Nb–Ta content and the enrichments in the LILE and LREE correlate with derivation from an enriched mantle source beneath the Cihai area. Notably, their overall geochemical characteristics are consistent with an origin from enriched lithospheric mantle sources, similar to the Permian basalt–andesite in the Pyrenees (Lago et al., 2004).

6.3. Fractional crystallization

The Cihai diabases have variable Mg$^\#$ (45–66) and compatible element contents, such as Cr (15.78–293.25 ppm) and Ni (19.28–79.16 ppm) (Table 5), suggesting that they were not derived from a primary melt that was in equilibrium with mantle peridotites and have undergone fractional crystallization within a magma chamber or en route to the surface (Cox, 1980; Prestvik and Goles, 1985). For these diabases, $\text{SiO}_2$ shows a rough negative correlation with MgO, total FeO$_2$ and TiO$_2$, and positive correlation with Al$_2$O$_3$ (Fig. 8a, b, c, d). These features, combined with the low Ni and Cr contents, probably suggest the fractionation of olivine-, pyroxene- and Fe–Ti oxides (e.g., rutile, ilmenite, titanite, etc.). Additionally, the fractionation of Ti-bearing minerals is also supported by the slight enrichment of Zr–Hf over Nb–Ta, because of their high compatibility for HFSE and their consequent capability of altering HFSE signatures of equilibrium. For example, rutile has been shown to have partition coefficients for Nb and Ta exceeding that for Zr, implying that their original concentrations were likely modified by alteration effects. Mg may also exhibit mobility during alteration (Zhou, 1999). Thus, these mobile elements will not be employed for petrogenetic interpretations in our work. Previous studies have shown that only severe alteration by interaction with seawater can affect the $\frac{143\text{Nd}}{144\text{Nd}}$ ratios (Mahoney et al., 1998) and the $\frac{87\text{Sr}}{86\text{Sr}}$ ratios can be markedly modified (e.g., Menzies et al., 2000).

6.2. Petrogenesis: crustal assimilation vs. source enrichment

The similarity in petrography and major and trace elements compositions between Cihai diabase samples suggest that they are co-magmatic. The diabases have generally high MgO (max. $9.29\text{wt.\%}$; $\text{Mg}^\#$ = 66) contents, indicating a dominant magma source from the upper mantle. However, these rocks are also marked by "crustal-like" trace element and isotopic features such as the enrichment in large ion lithophile element (LILE) and light rare earth element (LREE), and depletion in the high field strength elements (HFSE; Ta and Nb) (Table 5; Fig. 9). At least three processes are likely to account for the geochemical features of the Cihai diabases. These are: (1) binary mixing between mantle and crust-derived magmas; (2) crustal assimilation, i.e., mantle-derived melts that assimilated wall rocks during magma ascent; and (3) metasomatism of an older source material, that is, enrichment in the mantle source region via geodynamic processes such as subduction. The magma mixing model is not favored in the present case because this should generate mixing curves in the isotopic correlation diagrams and in plots between isotopic ratios and certain elements (e.g., MgO or SiO$_2$), which are not observed (Figs. 8–10). Mohr (1987) suggested that mantle-derived magmas are subjected to some degree of contamination during their ascent and/or temporary residence in crustal magma chambers. Our samples exhibit low Ta/La ratios (0.02–0.04) (cf. Ta/La = 0.06 for primitive mantle; Wood et al., 1979), and relatively large variation in Nd isotopes and slightly positive correlation between MgO and $\text{Sm}_{\text{eq}}(t)$ values (Fig. 10b), indicating that they have been affected by crustal contamination. Nevertheless, the absolute concentrations of incompatible elements (including Nb, Ta, Zr, Sm, etc.) are much higher than those typically observed in crustal rocks (Taylor and McLennan, 1985). Their positive $\text{Sm}_{\text{eq}}(t)$ values and relatively constant Nb/La ratios with increasing $\text{SiO}_2$ contents (not shown) suggest that the amount of crustal contamination was not significant or minor. Therefore, we infer that the majority of geochemical and isotopic features of the Cihai diabases including their low Nb–Ta content and the enrichments in the LILE and LREE correlate with derivation from an enriched mantle source beneath the Cihai area. Notably, their overall geochemical characteristics are consistent with an origin from enriched lithospheric mantle sources, similar to the Permian basalt–andesite in the Pyrenees (Lago et al., 2004).
this has presumably been affected by alteration associated with the skarn formation (Meinert, 1984). Moreover, the fractional crystallization of plagioclase is supported by the presence of some plagioclase as phenocrysts in the diabase dykes. However, the degree of fractional crystallization of plagioclase ought to be insignificant, derived from lithospheric mantle or asthenospheric magmas with asthenosphere (Fig. 9). Thus, these diabases might have been derived from OIB-like asthenospheric mantle. Geochemically, the enrichment of the LILEs (Rb, K, Th, U, Sr, Pb), depletion of the high-field strength elements (HFSEs) and Pb), depletion of the high-field strength elements (HFSEs) (Table 5), which are much higher than that of OIB (5.83), suggesting that they were not derived from OIB-like asthenospheric mantle. None of the geochemical characteristics are similar to MORB type asthenosphere (Fig. 9). Thus, these diabases might have been derived from lithospheric mantle or asthenospheric magmas with significant involvement of lithospheric mantle materials. Alternatively, a supra-subduction zone magma source is also possible, as discussed below.

Geochronologically, the enrichment of the LILEs (Rb, K, Th, U, Sr, and Pb), depletion of the high-field strength elements (HFSEs) (Nb and Ti), and the HREE (Yb) are characteristic features of mag-
Fig. 7. (a) AFM diagram showing geochemical variations of Cihai diabases, NW China. The tholeiitic and calc-alkaline trends are after Wilson (1989). (b) SiO$_2$ vs. FeO$_t$/MgO diagram (after Miyashiro, 1975).

Fig. 8. Variation diagrams for major oxides vs. SiO$_2$ contents for the Cihai diabases. Symbols and data source are same as Fig. 7.
mas generated in supra-subduction zone settings (e.g., Wilson, 1989; Polat et al., 2011; Yellappa et al., 2012). The Cihai diabases show a slight enrichment in LREE with $\text{(La/Yb)}_{N\text{r}}$ ranging between 1.50 and 8.77 and incompatible trace element levels 10 times as those of primitive mantle values (Sun and McDonough, 1989) (Fig. 9a and b). In terms of Sr and Nd isotopic values $[(87\text{Sr}/86\text{Sr})_i = 0.706–0.711$ and $e\text{Nd}(t) = +2.7$ to +6.1], although the Sr isotopic values were probably influenced by hydrothermal fluids as stated above, the Cihai diabases extend from DM towards the EMII end-member components (Fig. 10a). Since the EMII is considered to be a mantle contaminated by terrigenous sediments or subducted continental crust (Weaver, 1991; Greenough et al., 2005), it appears that these diabases were derived from modified enriched mantle influenced by a crustal/subduction component. In the Nb/Ta vs. Zr/Sm diagram, our sample plots overlap the arc magma field (Fig. 11a), likely suggesting the contribution of an arc-like source (Saunders and Tarney, 1984; Foley et al., 2002). Moreover, the Nb/Yb vs. Th/Yb plot in Fig. 11b shows that samples from the Cihai samples plot above the MORB–OIB array (Pearce and Peate, 1995), clearly indicating a subduction component in their sources as fluid/melt released from a subducted slab is known to result in high Th and low Nb in the upper mantle wedge (Pearce et al. 1995). Previous studies (e.g., Seghedi et al., 2001) have indicated that Ba is the most mobile incompatible element in arc magmatism caused by fluids released from the subducted slab, whereas Th is preferentially enriched in melts derived from the subducted slab. On the Ba–Nb/Yb diagram (Fig. 11c) proposed by Kepezhinskas et al. (1997) and the Th/Yb–Sr/Nd diagram (Fig. 11d) presented by Woodhead et al. (1998), the Cihai samples suggest that the enrichment of the lithospheric mantle beneath the Beishan region was possibly related to fluids from a subducted slab rather than melts derived from it.

Table 6

<table>
<thead>
<tr>
<th>Samples</th>
<th>Rb</th>
<th>Sr</th>
<th>$^{87}\text{Rb}/^{86}\text{Sr}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
<th>$2\sigma$</th>
<th>Sm</th>
<th>$^{147}\text{Sm}/^{144}\text{Nd}$</th>
<th>$^{143}\text{Nd}/^{144}\text{Nd}$</th>
<th>$2\sigma$</th>
<th>$(^{87}\text{Sr}/^{86}\text{Sr})_i$</th>
<th>$e\text{Nd}(t)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH07</td>
<td>42.2</td>
<td>370.1</td>
<td>0.3297</td>
<td>0.710014</td>
<td>0.00010</td>
<td>6.87</td>
<td>23.47</td>
<td>0.1762</td>
<td>0.512833</td>
<td>0.000005</td>
<td>0.70876</td>
</tr>
<tr>
<td>CH10</td>
<td>35.8</td>
<td>321.0</td>
<td>0.3227</td>
<td>0.708841</td>
<td>0.00012</td>
<td>3.80</td>
<td>12.77</td>
<td>0.1790</td>
<td>0.512874</td>
<td>0.000016</td>
<td>0.70761</td>
</tr>
<tr>
<td>CH11</td>
<td>19.5</td>
<td>209.3</td>
<td>0.2692</td>
<td>0.706649</td>
<td>0.00012</td>
<td>3.87</td>
<td>12.33</td>
<td>0.1888</td>
<td>0.512946</td>
<td>0.000000</td>
<td>0.70562</td>
</tr>
<tr>
<td>CH19</td>
<td>12.2</td>
<td>328.8</td>
<td>0.1076</td>
<td>0.708440</td>
<td>0.00008</td>
<td>7.14</td>
<td>26.92</td>
<td>0.1596</td>
<td>0.512802</td>
<td>0.000008</td>
<td>0.70803</td>
</tr>
<tr>
<td>CH25</td>
<td>69.1</td>
<td>200.0</td>
<td>0.9981</td>
<td>0.713250</td>
<td>0.00012</td>
<td>4.94</td>
<td>17.68</td>
<td>0.1684</td>
<td>0.512869</td>
<td>0.000008</td>
<td>0.70944</td>
</tr>
</tbody>
</table>

Note: Chondrite uniform reservoir (CHUR) values ($\left(\frac{^{143}\text{Sm}}{^{144}\text{Nd}}\right)_{\text{CHUR}} = 0.512638$, $\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_{\text{CHUR}} = 0.512638$) are used for the calculation. $\lambda_{\text{Sm}} = 1.42 \times 10^{-11}/$year (Steiger and Jäger, 1977), $\lambda_{\text{Rb}} = 1.42 \times 10^{-11}/$year (Steiger and Jäger, 1977), $\lambda_{\text{Sm}} = 6.5 \times 10^{-12}/$year (Lugmair and Hart, 1978). $(^{87}\text{Sr}/^{86}\text{Sr})_i$ and $e\text{Nd}(t)$ were calculated at 128 Ma.

Fig. 9. Chondrite-normalized rare earth element patterns and primitive-mantle-normalised multi-element diagrams for the mafic dykes. REE abundances for chondrites and trace element abundances for primitive mantle are after Sun and McDonough (1989).

Table 6

<table>
<thead>
<tr>
<th>Samples</th>
<th>Rb</th>
<th>Sr</th>
<th>$^{87}\text{Rb}/^{86}\text{Sr}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
<th>$2\sigma$</th>
<th>Sm</th>
<th>$^{147}\text{Sm}/^{144}\text{Nd}$</th>
<th>$^{143}\text{Nd}/^{144}\text{Nd}$</th>
<th>$2\sigma$</th>
<th>$(^{87}\text{Sr}/^{86}\text{Sr})_i$</th>
<th>$e\text{Nd}(t)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH07</td>
<td>42.2</td>
<td>370.1</td>
<td>0.3297</td>
<td>0.710014</td>
<td>0.00010</td>
<td>6.87</td>
<td>23.47</td>
<td>0.1762</td>
<td>0.512833</td>
<td>0.000005</td>
<td>0.70876</td>
</tr>
<tr>
<td>CH10</td>
<td>35.8</td>
<td>321.0</td>
<td>0.3227</td>
<td>0.708841</td>
<td>0.00012</td>
<td>3.80</td>
<td>12.77</td>
<td>0.1790</td>
<td>0.512874</td>
<td>0.000016</td>
<td>0.70761</td>
</tr>
<tr>
<td>CH11</td>
<td>19.5</td>
<td>209.3</td>
<td>0.2692</td>
<td>0.706649</td>
<td>0.00012</td>
<td>3.87</td>
<td>12.33</td>
<td>0.1888</td>
<td>0.512946</td>
<td>0.000000</td>
<td>0.70562</td>
</tr>
<tr>
<td>CH19</td>
<td>12.2</td>
<td>328.8</td>
<td>0.1076</td>
<td>0.708440</td>
<td>0.00008</td>
<td>7.14</td>
<td>26.92</td>
<td>0.1596</td>
<td>0.512802</td>
<td>0.000008</td>
<td>0.70803</td>
</tr>
<tr>
<td>CH25</td>
<td>69.1</td>
<td>200.0</td>
<td>0.9981</td>
<td>0.713250</td>
<td>0.00012</td>
<td>4.94</td>
<td>17.68</td>
<td>0.1684</td>
<td>0.512869</td>
<td>0.000008</td>
<td>0.70944</td>
</tr>
</tbody>
</table>

Fig. 10. (a) Initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $e\text{Nd}(t)$ diagram of Cihai diabases, compared with Weyya V–Ti–Fe-bearing mafic intrusion (Wang et al., 2008), Ni–Cu-bearing ultramafic–mafic intrusions both in Beishan (Jiang et al., 2006; Su et al., 2011) and Eastern Tianshan (Chai et al., 2008; Tang et al., 2012). (b) MgO vs. $e\text{Nd}(t)$ variation of Cihai diabase. Symbols and data source are same as Fig. 7. See text for detailed discussion.

The relatively high HREE abundances of the diabases indicate a garnet-free source. Hence, the parental magmas were likely derived from a spinel peridotite source region, which is further supported by flat HREE patterns (Xu, 2001). Additionally, absence of garnet in the source is further supported by the results obtained from the mantle melting modeling for the Cihai diabases using La/Sm versus Sm/Yb ratios. The results from experimental petrology suggest that sub-alkaline basaltic magma may be generated by 15–30% melting of upper mantle peridotite (Green and Ringwood, 1967; Green, 1973). Small degrees of melting (<10%) of mantle peridotite yield alkali-rich basaltic magma (Hirschmann et al., 1998) and show an enriched LREE pattern (Cullers and Graf, 1984). LREE concentration decreases rapidly with increased melting up to 30% and further melting does not significantly change the REE concentration. All our samples plot along the batch-melting curve of spinel peridotite (Fig. 12a) and show slight enrichment in LREE concentration (Fig. 9b), suggesting that these diabases are derived from basaltic magma that resulted from a lower percentage of melting of a garnet-free, enriched sub-continental lithospheric mantle (SCLM), characterized by spinel peridotite.

6.5. Tectonic implications

Some of the geochemical characteristics of the diabases, such as low Nb–Ta content and the enrichments in the LILE and LREE show an island arc signature, consistent with the results obtained from the tectono-magmatic discrimination diagrams (Fig. 13b–d). However, these characteristics were more likely inherited from the enriched mantle source and successive differentiation processes. As inferred, the Cihai diabases were derived from modified enriched mantle influenced by a subduction component. The samples plot derived from a garnet-bearing source are characterized by a steep HREE pattern (Xu, 2001). Additionally, absence of garnet in the source is further supported by the results obtained from the mantle melting modeling for the Cihai diabases using La/Sm versus Sm/Yb ratios. The results from experimental petrology suggest that sub-alkaline basaltic magma may be generated by 15–30% melting of upper mantle peridotite (Green and Ringwood, 1967; Green, 1973). Small degrees of melting (<10%) of mantle peridotite yield alkali-rich basaltic magma (Hirschmann et al., 1998) and show an enriched LREE pattern (Cullers and Graf, 1984). LREE concentration decreases rapidly with increased melting up to 30% and further melting does not significantly change the REE concentration. All our samples plot along the batch-melting curve of spinel peridotite (Fig. 12a) and show slight enrichment in LREE concentration (Fig. 9b), suggesting that these diabases are derived from basaltic magma that resulted from a lower percentage of melting of a garnet-free, enriched sub-continental lithospheric mantle (SCLM), characterized by spinel peridotite.

The relatively high HREE abundances of the diabases indicate a garnet-free source. Hence, the parental magmas were likely derived from a spinel peridotite source region, which is further supported by flat HREE patterns (Fig. 9b), whereas the magmas
in the within-plate basalt field on discrimination diagrams such as Zr/Y vs. Zr (Pearce and Norry, 1979; Fig. 13a) further indicating that these rocks were emplaced into a continental setting rather than an active arc.

The LA-ICPMS zircon data suggests the Cihai diabases were emplaced during Early Cretaceous (Fig. 6). As brought out in previous studies, the Beishan region and surrounding area were affected by intra-continental extension during the Early Cretaceous period (Fig. 14; Zheng et al., 1996; Wu et al., 2003). The argument in favor of the extensional tectonic regime is the formation of within-plate basalts in the Eastern part of Beishan (e.g. Guo et al., 2005; Peng et al., 2011) and extensive occurrence of A-type granites (Yin, 2008, 2010).

The extension regime in Beishan could be part of the long-lived India-Asia collision (Matte et al., 1997; Gaetani et al., 1993; Zhang et al., 2002a, 2004; Kapp et al., 2005). However, the NNE trending extensional fault system in Beishan region developed during Mesozoic indicates that the north–south India–Asia collision was not a likely candidate for the mechanism of the regional extension. Alternatively, the Jurassic to Cretaceous Mongolia–Okhotsk mountain building event in the northeastern part of the Eurasian plate (Krivchinsky et al., 2002; Tomurtogoo et al., 2005) is coeval to the emplacement of Cihai diabase. This event resulted in extensive rifting during the Late Mesozoic between the linked fault system striking NNE and the Mongolia–Okhotsk mountain belt as indicated by widespread intra-plate magmatism and basin formation (e.g., Graham et al., 2001). Therefore, we propose that Mongolia–Okhotsk mountain building event might have triggered the formation of Early Cretaceous extension in the Beishan region.

As mentioned previously, the Late Paleozoic tectonic history of the Beishan region is closely related to the subduction of the South Tianshan Ocean (e.g., Gao et al., 2009; Yue et al., 2001; Xiao et al., 2009). Thus, the lithospheric mantle source beneath the Beishan region was possibly influenced by the Paleozoic South Tianshan Ocean subduction (Fig. 14). This inference could well explain the observed geochemical signatures including the enrichment of LREE and LILE, and depletion of HFSE. Therefore, we suggest the mantle source for Cihai diabases in Beishan region might have been modified by Paleozoic subducted crustal components prior to the formation of the diabases. Combined fluid mobile/immobile trace element and Sr–Nd isotope ratios suggest that the subduction components are mainly fluids derived from the altered oceanic crust with minor contributions from the subducted sediments. It is possible that partial melting of lithospheric mantle modified by Paleozoic subducted-related fluids could be attributed to upwelling of asthenospheric mantle in response to regional extension.

6.6. Constraints on mineralization

6.6.1. Cornwall type mineralization?

Geological investigations have revealed only few gabbros and granites in the mining area. These do not display any close spatial relations with iron ore bodies. Most of the iron ore bodies occur within diabase or between diabase and wall rocks, suggesting that ore genesis is spatially and temporally associated with the Cihai diabases. This conclusion is also supported by the similar trace element compositions in magnetite from diabases and skarn iron ores.
6.6.2. Metallogenesis

Although similar to the Cornwall-type deposit characterized by post-magmatic hydrothermal activity, the origin of Cihai iron oxide deposit is still debated. Some workers proposed that the Cihai iron ores is closely related to the alteration of diabase, coeval with formation of skarns (Zhao et al., 2004), whereas others argue that the ore-forming fluids formed in depth and ascended into the fractures and contact zone between diabase and sedimentary rocks (Xue et al., 2000).

The similarity in trace elements patterns among magnetite of iron ores, disseminated ores, massive ores and the diabases, suggest that the ore-forming fluids have close genetic link with the diabase intrusion (e.g. Wang et al., 2006). With respect to the iron source, it is difficult to explain the enrichment of iron by leaching of the schist and carbonated rocks which have low iron contents. Thus, much of the iron ought to be derived from the magma. Particularly, the tholeiitic evolution trend (Fig. 7) under relatively low oxygen fugacity condition could produce elevated iron concentration in the residual magma (Osborn, 1959). The positive correlation between total Fe$_2$O$_3$ and loss-on-ignition (LOI) of our samples (Fig. 15a) indicates post-magmatic hydrothermal enrichment of iron in the diabase rather than removal of iron. Thus, leaching of diabase is unlikely to account for the enrichment of iron in the ores. Considering that most of the ore bodies occur as dykes and veins, and show strata bound affinity, a source at depth is considered to be most reasonable. The common occurrence of Cl-rich hornblende and other hydrous minerals such as biotite, apatite and chlorite in the diabase (Sheng, 1985) indicates supercritical aqueous chloride solutions as carriers of dissolved Fe (Bell and Simon, 2011).

Secondly, we infer that the mantle source of Cihai diabases has been enriched by subduction-related fluids (Fig. 14). Thus, a relatively higher content of volatile would be expected in the mantle-derived magma (e.g. Johnson et al., 2009). In addition, the fractional crystallization of anhydrous minerals, such as olivine and clinopyroxene, is also capable of enhancing the volatile content of the residual magma. This is consistent with the presence of hydrous minerals (hornblende and biotite) in the diabase. The volatile-rich diabase magma is favorable to form considerable hydrothermal fluids by exsolution or degassing during the post magmatic evolution (e.g. Webster et al., 1999; Halter and Webster, 2004) especially when the magma is emplaced at shallow depth (e.g. Woods and Pyle, 1997). However, these processes are inadequate to account for the extensive albite alteration in the diabase since the fresh diabase is not significantly sodium-rich. As mentioned before, a sequence including limestone, marl, gypsum and salt was formed during Jurassic to Cretaceous in the Beishan region (Fig. 14). There might be a genetic link, probably involving large-scale circulation of basinal brines from an evaporitic source, induced by magmatic heat, with the diabase intrusion buried in this case, similar to the scenario described by Barton and Johnson (1996). This inference is supported by previous studies (Ren, 1985) on the Cihai deposit that show that the clinopyroxene in the iron ores possesses higher $\delta^{18}$O value (8.24–10.05‰; avg. 9.26‰) than those of the diabases (5.46–7.8‰; avg. 6.45%), and higher $\delta^{34}$S value (4.35–6.81‰; avg. 5.53‰) of the magnetite from the iron ores than those in diabase (3.2–4.29‰). Moreover, Cl-rich minerals, such as biotite, apatite and hornblende, are widely recognized in the Cihai deposit as stated earlier. These minerals could also be considered as evidence that an evaporitic source was involved in the mineralization (Cai, 1980). The hydrothermal fluids were possibly formed by the mixture of magmatic fluids, meteoric water and the basinal brine. Such a fluid source is currently observed in seafloor hot springs, such as those in the middle of the Dead Sea (Barton and Johnson, 1996). The evaporites were a source of Cl, the typical Na alteration, and the high oxidation state of

![Fig. 14. Structure, magmatic evolution and metallogenic model of Cihai area. See text for details.](Image)
hydrothermal Fe oxide deposits (Fig. 14). A similar mechanism has been widely invoked to explain the origin of hydrothermal Iron Oxide–Copper–Gold ‘IOCG’ clan (Hitzman, 2000; Hitzman et al., 1992), ‘Kiruna-type’ iron deposits in Bafq district, Central Iran (Torab and Lehmann, 2007), ‘porphyry-type’ and skarn iron deposits in Middle-Lower Yangtze River valley (e.g. Ningwu Research Group, 1978; Institute of Geochemistry of Chinese Academy of Sciences, 1987) and the skarn iron deposits in Laiwu in Shandong Province (e.g. Zong et al., 2011).

An important feature is that the occurrence of skarn minerals in Cihai is not restricted to the contact between ore and carbonated rocks. Instead, considerable iron ores occur at the contact between ore and schist, or in fractures within the diabase (Fig. 3), different from the traditional skarn. The initial positive correlation of MgO with the increasing LOI suggests that the ore-forming fluids might have carried considerable amounts of Mg, whereas the sharp decrease of MgO at higher LOI values probably corresponds to the formation of skarn minerals (Fig. 15b). Since extensive fractures occur in the diabases, it is likely that this model is based on a convection cell driven by temperature gradients established by the intrusion of the diabase sills (Rose et al., 1985). Circulation of fluids is set in motion by the fracturing of the diabase and permeability (Eugster and Chou, 1979). Therefore we propose a depositional model based on convecting, supercritical aqueous chloride solutions of dissolved Fe-Mg. The solutions circulate to the cold end of the cell, where they come in contact with CaCO3 in the wall rocks. Magnetite and pyrite are precipitated and the solutions become enriched in CaCl2 and CO2, some of which may descend with the return flow (Eugster and Chou, 1979). Although some CaCl2 in the return flow could be removed by precipitation of Ca-bearing minerals, such as plagioclase or calcite, dilution by ground water is also possible in an open circulation cell, which could account for the formation of skarn minerals in the fractures of diabase. This hypothesis requires the presence of a “primary magma” that gave rise to both the diabase and the iron-rich solution, as separate events.

The magnetite in all these ores share similar trace element compositions (Wang et al., 2006), suggesting a common origin. Furthermore, the Cihai iron ores contain low concentrations of TiO2 (avg. 0.12 wt.%), Cr (avg. 45.58 ppm), V (avg. 28.4 ppm) (Tang et al., 2010), indicating that they likely formed by post-magmatic hydrothermal activity closely related to the emplacement of the diabases. Different types of iron ores could form during the flow of the hydrothermal fluids. In our model, the driving force for the ore solution was a convection cell with the ‘hot’ end of the cell at depth and the ‘cold’ end being the Neoproterozoic schist and marble. This is supported by the shallow emplacement depth of the diabase. The iron chlorides ride the convection current to the ‘cold’ end of the cell where they come into contact with the wall rocks. As stated above, the massive ores predominantly occur in the central parts of ore bodies, and the disseminated ores always occur on the margin where the ore bodies thin outward, and the banded ores contain gangue mineral bands composed of garnet, diopside and hornblende whereas the brecciated ores mainly occur in the lower parts of ore bodies. Thus the ore body shows evidence of formation via deposition of magnetite to form massive ores (Doyle, 2001), via replacement of wall rocks to form banded and disseminated ores (Forster et al., 2004; Mueller et al., 2012), and via filling in the fracture zone to form brecciated ores (Jebraik, 1997). In summary, the combination of subduction-related enrichment in the source, shallow depth of emplacement, and the involvement of large-scale circulation of basin brines from an evaporitic source can account for formation of ‘Cornwall-type’ mineralization in the Cihai area.

7. Conclusions

The close relation between iron ores and Early Cretaceous diabases in Cihai in the Beishan region is similar to the relations in ‘Cornwall-type’ iron deposits. The geochemical features suggest partial melting of subduction components in the source mantle lithosphere by conductive heating of upwelling asthenospheric mantle in an extensional setting during the Early Cretaceous. The iron ores were formed by iron-rich hydrothermal fluids that were possibly derived from a mixture of magmatic fluids, meteoric water and basin brines, formed by a convection cell driven by temperature gradients established by the intrusion of the diabase sills. A combination of the subduction-related enrichment in the source, shallow depth of emplacement, and the involvement of large-scale circulation of basin brines from an evaporitic source are inferred to have generated the ‘Cornwall-type’ mineralization in Cihai. This model has potential implications for the origin of other ‘Cornwall-type’ deposit elsewhere in the world.

Acknowledgements

We are grateful to Paul Carpenter of Washington University in St. Louis for assistance with the microprobe work. We thank an anonymous reviewer and Editor-in-Chief Prof. Bor-ming Jahn for their thoughtful and constructive comments. Financial support for this work was supported by 973 Program (2012CB416806), the Special Fund for Scientific Research in the Public Interest (200911007-25), the “Fundamental Research Funds for the Central Universities”, the 111 Project (B07011), and PCSIRT.


