Petrogenesis and tectonic settings of the Late Carboniferous Jiamantieliek and Baogutu ore-bearing porphyry intrusions in the southern West Junggar, NW China

Ping Shen a,*, Wenjiao Xiao a, Hongdi Pan b, Lianhui Dong a,c, Chaofeng Li a

a Key Laboratory of Mineral Resources, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing 100029, China
b College of Earth Sciences, Chang'an University, Xi'an 710054, China
c Xinjiang Bureau of Geology and Mineral Resources, Urumqi 830000, China

Keywords: Zircon U–Pb age Sr–Nd–Pb isotope Jiamantieliek Baogutu West Junggar Xinjiang

1. Introduction

The Central Asia Orogenic Belt (CAOB) is the largest Phanerozoic juvenile crustal growth orogenic belt in the world (Sengör and Natal'in, 1996; Jahn et al., 2000, 2004; Xiao et al., 2008, 2009, 2010) and have plentiful mineral sources (Seltmann and Porter, 2005; He and Zhu, 2006; Zhu et al., 2007; Shen et al., 2010a,b, 2013). As a part of the CAOB, the West Junggar terrain in Xinjiang (NW China) is economically important, not only as a potential target for Cu–Au–Mo–W–Cr exploration (Shen et al., 1993; He and Zhu, 2006; Zhu et al., 2007; Shen et al., 2010a,b, 2012, 2013), but also as a critical area to study the subduction-accretion history of the CAOB.

The West Junggar terrain largely comprises Palaeozoic volcanic arcs in the northern West Junggar and accretionary complexes in the southern West Junggar (e.g., Windley et al., 2007; Xiao et al., 2008, 2009, 2010; Zhang et al., 2011a,b), which were accreted onto the Kazakhstan plate as the Tarim, Kazakhstan and Siberian plates converged (Jahn et al., 2000; Chen and Jahn, 2004; Chen and Arakawa, 2005; Xiao et al., 2008). This geodynamic process led to the formation of voluminous granitoids and small ore-bearing stocks. The former are I- and A-type granites with highly-depleted isotopic signatures (εNd(t) = +6.4 to +9.2) (Jahn et al., 2000; Chen and Jahn, 2004; Chen and Arakawa, 2005; Han et al., 2006). The latter are diorites that are associated with the porphyry copper deposits (e.g. Baogutu and newly-discovered Jiamantieliek) in the southern West Junggar (Fig. 1b, Shen et al., 2009, 2013). Previous studies were mainly concentrated on the voluminous granitoids in the area (e.g. Jahn et al., 2000; Hu et al., 2000; Chen and Jahn, 2004; Chen and Arakawa, 2005; Han et al., 2006; Zhou et al., 2008), with few attention to these ore-bearing stocks. In addition, the intrusions tectonic setting of the southern West Junggar is controversial. They have been attributed either to subduction-related sources in an island arc setting (Xiao et al., 2008, 2009; Shen et al., 2009; Geng et al., 2009, 2011; Tang et al., 2009, 2010; Zhang et al., 2011a) or to depleted mantle contributions in a post-collisional extensional setting...
These models clearly have significantly different implications for intrusion petrogenesis and associated mineralisation. In this paper, we first report the major elements, trace elements and Sr–Nd–Pb isotopic geochemical data for whole rocks and for U–Pb zircon SHRIMP ages from the Jiamantieliek ore-bearing intrusion in the Barluk Mountains. For comparison, we have also added new results for the Baogutu intrusion in the Kelamay Region relating to 10 whole-rock major and trace element analyses, and four Sr–Nd–Pb isotope ratios. Our aim was to provide constraints to the tectonic setting and associated ore-bearing magma in the West Junggar, Xinjiang. All these factors have significant implications for understanding the continental growth and associated mineralisation in the CAOB.

2. Geological outline

The West Junggar terrain is bound by the Altai orogen to the north and by the Tian Shan orogen to the south and extends westward to the Junggar–Balkhash region in adjacent Kazakhstan and eastward to the Junggar Basin in North Xinjiang, NW China. The southern West Junggar is located between latitudes 45°15′ and 46°15′ north and longitudes 82°15′ and 86°00′ east (Fig. 1b). Geologically, the southern West Junggar is characterised by several
northeast-trending faults, including the Barluk, Mayile and Darbut faults, and fault-bounded accretionary complexes. This is very different from the northern West Junggar where the major faults and fault-bounded blocks are mainly E–W oriented (Fig. 1b, Xiao et al., 2008; Shen et al., 2012a).

The Barluke Mountains are located in the western part of the southern West Junggar (Fig. 1b). The fossil-dated Devonian strata occur widely across the Barluke Mountains, close to the border with China–Kazakhstan (Fig. 1b; BGMRXUAR, 1993), and comprise the volcano-sedimentary strata of the Middle Devonian Barluk and Tielieketi Groups. The Barluk Group is composed of siltstone, tuff and minor basalt; the Tielieketi Group consists of greywacke, tuffaceous mudstone and tuffaceous siltstone. These Devonian sequences are intruded by Carboniferous–Permian intrusions in the Barluk Mountains. The Carboniferous intrusions include peridotite, gabbro, diorite and quartz diorite stocks, and the Permian intrusions include diorite, quartz diorite and adamellite stocks [BGMRXUAR, 1993]. Structurally, the Barluk Mountains are characterised by the northeast-trending Barluke fault. There are NW-trending faults that are almost orthogonal to the Barluke fault. The Jiamantieliek intrusion is found in the western part of the Barluk fault (Fig. 1b).

The Kelamay Region, located in the southeastern section of the West Junggar, is characterised by the occurrence of Lower Carboniferous volcano-sedimentary strata (Fig. 1b). There are three Early Carboniferous stratigraphic units: from oldest to youngest, they are the Tailegula, Baogutu and Xibeikulasi groups (Shen and Jin, 1993). The Tailegula group consists of a succession of basic volcanic and volcaniclastic rocks intercalated with chert. The Baogutu group includes tuffaceous siltstone, silt tuff and felsic tuff. The Xibeikulasi group consists of greywacke and sandstone. These Early Carboniferous sequences were intruded by ore-bearing stocks at about ~320 Ma (Fig. 1b; Shen et al., 2012b; Tang et al., 2009). The Baogutu intrusion is placed to the south of the Darbut fault; the main structures in this area are a series of approximately N-trending faults and folds, which are almost orthogonal to those found to the north of the Darbut fault (Fig. 1b). The Xibeikulasi, Baogutu and Tailegula groups have a synclinal structure, in which the Xibeikulasi group forms the core and the Baogutu and Tailegula groups form the flanks. The studied Baogutu intrusion occurs within the eastern flank of the syncline.

3. Ore-bearing intrusions

3.1. Jiamantieliek

Some granitoids that occur in the Barluk Mountains are represented by the Jiamantieliek intrusion hosting the Jiamantieliek porphyry copper deposit (Fig. 2). The Jiamantieliek intrusion was intruded into the volcano-sedimentary strata of the Middle Devonian Barluk and Tielieketi Groups and now occupies localised dilatant sites created by a structural intersection of NW- and NE-trending faults (Fig. 2). The intrusion is irregular in shape at the surface, and its total area is 9 km².

We established a profile of over 400 m in length through the Jiamantieliek ore-bearing intrusion (Fig. 2). There were at least two intrusive phases within this intrusion, based on cross-cutting relationships; these were the main-stage stock, in which occurred....
abundant alteration and mineralisation, and the minor late-stage dikes.

Main-stage stock is widespread in the study area (Fig. 2a), consisting of diorite and quartz diorite. Diorite and quartz diorite have a granular to weakly porphyritic texture (Fig. 4a). Diorite contain plagioclase (60–65%), hornblende (15–20%), biotite (5–10%), and minor quartz (<5%) and accessory opaque oxides, zircon, and apatite. Quartz diorites have similar mineralogies to the diorites, but are distinguished by more quartz (<10% by volume, up to 15–20% by volume). Plagioclases (An = 30–40) of the diorite and quartz diorite show typically andesite zoning (Fig. 4a), corresponding to andesine. The main-stage stock makes abrupt contacts with the wall rocks of the intrusion. Several small bodies of porphyry cutting the main-stage diorite stock have been mapped (Fig. 2b).

Fig. 3. A Geological map of the Kelamay Region in the southern West Junggar region (modified after Shen and Jin, 1993), showing the location of the studied Baogutu intrusion and associated deposit. Fig. 3B. Geological map of the Baogutu intrusion. Dots indicate the position of drill holes, and pentagons show the location of the samples used in this study (Shen et al., 2010a).

Fig. 4. Photomicrographs of the main rocks from Jiamantieliek and Baogutu intrusions. (A) Jiamantieliek main-stage diorite with granular texture; (B) Jiamantieliek late-stage diorite porphyry with porphyritic texture; (C) Baogutu main-stage diorite with granular texture; (D) Baogutu late-stage diorite porphyry with porphyritic texture. All photomicrographs were captured under transmitted lights. Abbreviations: Pl: plagioclase; Hb: hornblende.
They are referred to here as late-stage diorite porphyry and quartz diorite porphyry. Sharp contacts between the late-stage and main-stage rocks are exposed mainly on the eastern rim of the Jiamantieliek intrusion (Figs. 2a and b). The late-stage porphyries are characterised by a porphyritic texture. Phenocrysts of plagioclase, hornblende, and biotite in diorite porphyry and quartz diorite porphyry are typically 2–3 mm in diameter and account for 20–30 vol.% of the rock. Microcrystalline plagioclase, hornblende and minor quartz define the groundmass, which occupies 70–80 vol.% of the porphyries (Fig. 4b).

3.2. Baogutu

Our previous work includes descriptions of the petrology, mineralisation and geochronology of the Baogutu deposit (Shen et al., 2010a,b, 2012b), the main characteristics of which are summarised in the following sections.

The Baogutu intrusion contained metal, at levels of 63 × 10^4 tonnes Cu, 1.8 × 10^4 tonnes Mo and 14 tonnes Au, and is associated with a Carboniferous intrusion that was emplaced into the Lower Carboniferous volcano-sedimentary sequences of the Baogutu and Xibeikuliji Groups. Two intrusive phases at Baogutu are the main-stage diorite stock and minor late-stage diorite porphyry dikes. Sharp intrusive contacts have been observed between the main-stage diorites and late-stage diorite porphyries (Fig. 3b). Alteration and mineralisation were closely related to the main-stage diorites.

The main-stage diorite stock includes granular to weakly porphyritic diorite (Fig. 4c) and weakly porphyritic quartz diorite. Diorite contains plagioclase (40–50%), hornblende (25–30%), biotite (5–10%), minor clinopyroxene (<5%) and quartz (<5%), and rare titanite, rutile, apatite, magnetite, and zircon. Quartz diorites have similar mineralogies to the diorites, but are distinguished by less pyroxene (<3–5% by volume) and more quartz (<10% by volume, up to 15% by volume). The late-stage diorite porphyry dikes are diorite porphyry and quartz diorite porphyry, which have the porphyritic texture (Fig. 4d). Phenocrysts are plagioclase, hornblende, and minor biotite, which are 1–2 mm in diameter and account for 10–20 vol.% of the rock. The groundmass is microcrystalline plagioclase, hornblende and minor quartz which occupies 80 vol.% of the porphyries.

4. Methods and results

4.1. Methods

Seventeen representative samples from the Jiamantieliek intrusion were petrographically selected for chemical analyses. The location of samples is shown in Fig. 2b. They are the main-stage dikes.
diorite and quartz diorite and late-stage diorite porphyry and quartz diorite porphyry. The six samples from the main-stage diorite stock were selected for Rb, Sr, Sm and Pb isotope compositions analyses. We also selected two samples (YJ1-21, YJ1-175) from the main-stage diorite stock and late-stage diorite porphyry dikes at Jamantieliek, respectively, for zircon separation (the location of samples is shown in Fig. 2b). For comparison, ten representative samples from the Baguotu main-stage stock and late-stage diorite porphyry dikes were selected for chemical analyses. The four samples were selected for Rb, Sr, Sm, Nd and Pb isotope compositions analyses. All measurements were carried out at the Institute of Geology and Geophysics, Chinese Academy of Sciences in Beijing, except for SHRIMP zircon U–Pb dating, which were carried out in the Beijing SHRIMP Center, Chinese Academy of Geological Sciences in Beijing.

4.1.1. Major and trace elements

About 0.5–1.0 kg samples were crushed and quartered. 100-g sample was milled and analysed for major oxides, rare earth elements (REE), selected trace elements and Rb, Sr, Sm, Nd and Pb isotope compositions. Major elements were analysed by XRF-1500 Sequential x-Ray Fluorescence Spectrometer, with wet chemical determination of FeO and loss-on-ignition. Analytical precision based on certified standards and duplicate analyses are expressed in terms of relative percentages, which range from ±0.01% to ±0.20%. A PQ2 Turbo inductively Coupled Plasma Mass Spectrometer (ICP-MS) was used to analyse trace elements and RREE. 100-g of sample was digested in beakers using a hot HF + HNO3 mixture followed by a HF + HNO3 + HClO3 mixture to ensure complete dissolution and then taken up in 1% HNO3 (dilution factor of ~500). The measurement error and drift were controlled by regular analysis of standard samples with a periodicity of 10%. Analysed uncertainties of ICP-MS data at the ppm level are better than ±3%–10% for trace elements, ±5%–10% for RREE.

4.1.2. Sr–Nd–Pb isotopes

Rb, Sr, Sm and Nd isotope compositions were analysed according similar procedure that described by Chen et al. (2002). Procedural blanks were <100 pg for Sm and Nd and <50 pg for Rb and Sr. The 87Sr/86Sr ratios were normalised to 86Sr/88Sr = 0.1194 and the 142Nd/144Nd ratios to 144Nd/146Nd = 0.7219. Typical within-run precision (2σm) for Sr and Nd was estimated to be ±0.000010 and ±0.000013, respectively. The measured values for the JMC Nd standard and NBS987 Sr standard were 143Nd/144Nd = 0.511937 ± 7

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>ZK106-1</td>
<td>1</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-2</td>
<td>2</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-3</td>
<td>3</td>
<td>505</td>
<td>1.5</td>
<td>4.7</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-4</td>
<td>4</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-5</td>
<td>5</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-6</td>
<td>6</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-7</td>
<td>7</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-8</td>
<td>8</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-9</td>
<td>9</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-10</td>
<td>10</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-11</td>
<td>11</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-12</td>
<td>12</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-13</td>
<td>13</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-14</td>
<td>14</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-15</td>
<td>15</td>
<td>506</td>
<td>1.5</td>
<td>4.8</td>
<td>2.9</td>
<td>1.9</td>
<td>1.5</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
<tr>
<td>ZK106-16</td>
<td>16</td>
<td>502</td>
<td>1.1</td>
<td>4.1</td>
<td>2.4</td>
<td>1.4</td>
<td>1.2</td>
<td>1.8</td>
<td>1.7</td>
<td>1.7</td>
<td>1.8</td>
<td>1.9</td>
<td>1.9</td>
<td>1.9</td>
<td>1.8</td>
<td>1.7</td>
<td>1.6</td>
</tr>
</tbody>
</table>

Data sources: ‘Shen et al. (2009); the other this study. Abbreviations: D: diorite; QD: quartz diorite; DP: diorite porphyry; QDP: quartz diorite porphyry.’
(2σm, n = 12) and 86Sr/88Sr = 0.710226 ± 12 (2σm, n = 12), respectively, during the period of data acquisition. Pb isotopic ratios were analysed with the same mass spectrometer; analytical precision is better than ±0.1‰.

4.1.3. SHRIMP zircon U–Pb dating

Zircon grains were separated using standard techniques involving heavy liquid and magnetic separation for SHRIMP study. Representative grains were handpicked using a binocular microscope, mounted in an epoxy resin disc, and then polished and coated with a gold film. All zircons were documented with transmitted and reflected light micrographs as well as cathodoluminescence (CL) images to reveal their internal structures, and the mount was vacuum-coated with high-purity gold prior to SHRIMP analysis.

U, Th, and Pb analyses of the two samples (YJ1-21, YJ1-175) were carried out using SHRIMP II. The operating conditions followed those of Williams (1998). Handpicked zircons were mounted in epoxy together with Temora (417 Ma) and SL13 (U content 238 ppm, 572 Ma) zircon standards, cut in half, and polished. U–Th–Pb isotopic ratios were determined relative to the TEMORA standard zircon (Black et al., 2003), and their concentrations were calibrated relative to SL13 reference zircon. The instrumental techniques are similar to those described by Compston et al. (1992). Data were reduced according to the procedure of Williams (1998), using ISOPLOT (Ludwig, 1999) and SQUID (1.02) software. The analytical data are presented as 1σ error boxes on the concordia plots, and uncertainties in mean ages are quoted at the 95% confidence level (2σ). The decay constants used in age calculations are 238U = 0.155125 Ga−1 and 235U = 0.98485 Ga−1.

4.2. Results

4.2.1. Major and trace elements

Major and trace element concentrations of the studied samples are listed in Tables 1 and 2 and plotted in Figs. 5–7, respectively. The loss on ignition (LOI) for all rocks analysed in this study ranged from 0.48 wt.% to 4.74 wt.%, which probably resulted from postmagmatic fluid-rock interactions. All major oxides were LOI-free normalised before petrogenetic interpretation. The investigated rocks from the Jiamantieliek intrusion have a sub-alkaline character and a limited range of SiO2 content (56–66%) except for one outlier of 68% (Table 1; Fig. 5A). In the diagram for SiO2 vs. K2O (Fig. 5B), samples were shown to have a relatively wide range of K2O, from medium-K calc-alkaline to high-K calc-alkaline.

Na, K and low-field-strength elements are mostly mobile and susceptible to change during alteration (e.g., Humphris and Thompson, 1978), whilst the high-field-strength elements and REE are essentially immobile during all but the most severe seafloor-hydrothermal alterations (e.g., Pearce, 1975; Wood et al., 1979) and during high-salinity, oxidised fluid-rock interactions at temperatures >400 °C (Dongen et al., 2010). At Baogutu, the ore-bearing rocks were subjected to fluid-rock interactions at temperatures <400 °C (Shen et al., 2010a), while at Jiamantieliek the fluid-rock interactions took place in temperatures conditions of <400 °C (unpublished data). Thus, immobile elements are referred to in the following classification.

Most samples from the Jiamantieliek were plotted in the gabbro to diorite fields with sub-alkaline character in the Zr/TiO2 vs. Nb/Y diagram (Fig. 5C); this was consistent with the above-mentioned results for the use of transmitted light petrography. Minor samples are plotted in the diorite field and close to the boundary between...
the diorite and granodiorite. Together with the results of the detailed thin section petrography (quartz <20%), these samples are quartz diorites. Therefore, it was possible to classify the Jiamantieliek intrusion, by its petrography and immobile elements, which ranges from diorite to quartz diorite. In addition, all samples showed a high Mg$^#$ character, with Mg$^#$ >0.4 (Fig. 5D).

Fig. 6. Chondrite normalised (Nakamura, 1974) REE distribution and patterns of trace elements normalised (Sun and McDonough, 1989) to E-MORB for the Jiamantieliek intrusions.

Fig. 7. Chondrite normalised (Nakamura, 1974) REE distribution and patterns of trace elements normalised (Sun and McDonough, 1989) to E-MORB for the Baogutu intrusions.
Table 3
Nd-Sr-Pb isotopic data for the Jiamantieliek and Baogutu intrusions.

<table>
<thead>
<tr>
<th>Intrusions</th>
<th>Jiamantieliek</th>
<th>Baogutu</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample</td>
<td>ZK2-260</td>
<td>001-191</td>
</tr>
<tr>
<td>Rocks</td>
<td>D</td>
<td>D</td>
</tr>
<tr>
<td>ZK211-263</td>
<td>28.31</td>
<td>9.75</td>
</tr>
<tr>
<td>ZK201-616</td>
<td>0.1233</td>
<td>0.1193</td>
</tr>
<tr>
<td>ZK202-295</td>
<td>0.12779</td>
<td>0.12812</td>
</tr>
<tr>
<td>BGT-600</td>
<td>12</td>
<td>10</td>
</tr>
<tr>
<td>BGT-5-1&quot;</td>
<td>0.51252</td>
<td>0.51256</td>
</tr>
<tr>
<td>BGT-5-2&quot;</td>
<td>2.8</td>
<td>3.4</td>
</tr>
<tr>
<td>BGT-5-3&quot;</td>
<td>610</td>
<td>544</td>
</tr>
<tr>
<td>BGT-5-4&quot;</td>
<td>3.81</td>
<td>3.86</td>
</tr>
<tr>
<td>U</td>
<td>1.91</td>
<td>7.79</td>
</tr>
<tr>
<td>Th</td>
<td>23.81</td>
<td>18.25</td>
</tr>
<tr>
<td>Pb</td>
<td>15.70</td>
<td>15.41</td>
</tr>
<tr>
<td>(205Pb/204Pb)m</td>
<td>39.96</td>
<td>37.80</td>
</tr>
<tr>
<td>(205Pb/204Pb)i</td>
<td>17.41</td>
<td>17.89</td>
</tr>
<tr>
<td>(206Pb/204Pb)m</td>
<td>15.34</td>
<td>15.33</td>
</tr>
<tr>
<td>(206Pb/204Pb)i</td>
<td>37.74</td>
<td>37.31</td>
</tr>
</tbody>
</table>

Rb, Sr, Sm, Nd, U, Th and Pb in ppm. (m) is the measured value, (i) is the idealised crystallization age. The 87Sr/86Sr ratios are normalised to 86Sr/88Sr = 0.1194 and the 143Nd/144Nd ratios to 146Nd/144Nd = 0.7219. Data sources: Shen et al. (2009); the other this study. **Abbreviations**: D: diorite; QD: quartz diorite; DP: diorite porphyry.
Comparison of the Baogutu and Jiamantieliek intrusions revealed that the Baogutu rocks are similar to those of Jiamantieliek except that they are less enriched in SiO₂ and alkline, and contain lower KO₂ and variable Mg# (Figs. 5A–D). The Baogutu can be classified as ranging from diorite to quartz diorite.

All samples from the Jiamantieliek intrusion exhibited coherent chondrite-normalised (Nakamura, 1974) patterns of rare earth elements (REE), characterised by a relative enrichment of light rare earth elements (LREE) and nearly flat heavy rare earth element (HREE) segments, without Eu anomalies (Figs. 6A and C). They also have similar enriched mid-oceanic ridge basalt (E-MORB)-normalised (Sun and McDonough, 1989) trace element patterns, characterised by a negative Nb anomaly and enrichment in large ion lithophile elements (LILE) (Figs. 6B and D).

The Baogutu rocks have similar REE patterns and E-MORB-normalised trace element patterns to those of the Jiamantieliek rocks, except that they have more variable REE patterns and a slightly low ΣREE and LILE (Fig. 7). All rocks displayed a negative Nb-anomaly.

**Fig. 8.** Initial (313 Ma) isotope characteristics (after DePaolo, 1981; Alibert, 1985; Deckart et al., 2005) of the Jiamantieliek and Baogutu intrusions from this work and our previous work (Shen et al., 2009). (a) εNd(t) vs. (87Sr/86Sr)i diagram (the early to middle Proterozoic crust from Hu et al., 2000). (b) (143Nd/144Nd)i vs. (87Sr/86Sr)i, (c) (143Nd/144Nd)i vs. (206Pb/204Pb)i, (d) (87Sr/86Sr)i vs. (206Pb/204Pb)i, (e) (207Pb/204Pb)i vs. (206Pb/204Pb)i, and (f) (208Pb/204Pb)i vs. (206Pb/204Pb)i. The diagram shows mixing lines with percentages of mixing between N-MORB and EM1 or EMII end-members. DM, depleted mantle; BSE, bulk silicate Earth; EM1 and EM2, enriched mantle; HIMU, mantle with high U/Pb ratio. Symbols are the same as in Fig. 5.
4.2.2. Sr–Nd–Pb isotopes

The measured and initial (back-calculated to 310 Ma) isotopic ratios are reported in Table 3. All samples from the Jiamantieliek and Baogutu intrusions showed a limited range in their $^{143}$Nd/$^{144}$Nd and $^{87}$Sr/$^{86}$Sr ratios (Fig. 8). In details, initial $^{87}$Sr/$^{86}$Sr ratios for the Jiamantieliek intrusion ranged from 0.70369 to 0.70401; their Nd isotopes displayed small variations of between 0.51247 and 0.51256, and $\epsilon_{Nd}(t)$ from +1.6 to +3.4. Their Pb initial ratios were less variable ($^{206}$Pb/$^{204}$Pb = 17.41–18.23, $^{207}$Pb/$^{204}$Pb = 15.31–15.41 and $^{208}$Pb/$^{204}$Pb = 37.31–37.74). Initial $^{87}$Sr/$^{86}$Sr ratios of the Baogutu main-stage diorite and quartz diorite were from 0.70355 to 0.70388, their Nd isotopes from 0.51257 to 0.51261 and the $\epsilon_{Nd}(t)$ from +4.1 to +6.0. Their initial Pb ratios were also less variable ($^{206}$Pb/$^{204}$Pb = 17.43–18.03, $^{207}$Pb/$^{204}$Pb = 15.32–15.46, and $^{208}$Pb/$^{204}$Pb = 37.75–37.85).

4.2.3. Zircon U–Pb ages

All analyses illustrated the variability in U concentrations, which ranged from 65 to 408 ppm. Thorium ranged from 27 to 436 ppm and Th/U ratios varied between 0.41 and 1.07. All of the zircon Th/U ratios were greater than 0.1, therefore indicating a magmatic origin for the zircons.

Sample YJ1-175: this was a diorite selected from the main-stage stock of the Jiamantieliek intrusion. Analyses of 14 magmatic zircons were obtained from this sample (Table 4). The calculated weighted mean age was 313 ± 4 Ma (Fig. 9b). This age was interpreted as the crystallisation age of the main-stage stock.

Sample YJ1-21: this was a diorite porphyry selected from the late-stage dike of the Jiamantieliek intrusion. Analyses of 18 magmatic zircons were obtained from this sample (Table 4). Excluding one outlaying age result of 250.4 ± 7.0 Ma, the calculated weighted mean age for the 17 remaining samples was determined to be 310 ± 5 Ma (Fig. 9d). This age was interpreted as the crystallisation age of the late-stage porphyry.

5. Discussion

5.1. Ages

The Jiamantieliek intrusion includes the main-stage stock and late-stage dikes. The geochronological data provide direct support for the occurrence of two intrusive events. The U–Pb zircon SHRIMP age of the diorite is 313 ± 4 Ma and the diorite porphyry is 310 ± 5 Ma. In our previous work, we determined the U–Pb zircon SIMS ages of the Baogutu intrusion at 313 ± 2 Ma and 312 ± 2 Ma for the main-stage diorite and late-stage diorite porphyry, respectively. These data indicate that the Jiamantieliek and Baogutu ore-bearing intrusions were the product of a short-lived magmatic activity. Tang et al. (2009) have determined a zircon LA-ICPMS U–Pb age of 314 ± 2 Ma and 313 ± 3 Ma for Stock II and III, respectively, in the Kelamay Region (Fig. 3A). Our Re–Os dating of molybdenite from the Baogutu deposit indicates that mineralisation occurred at around 312 Ma (Shen et al., 2012b).

Based on these data, we suggest that the ore-bearing magma and associated Cu-Au-Mo mineralisation in the southern West Junggar occurred in the Late Carboniferous. This is consistent with the ore-bearing magma and mineralisation ages in the North Balkhash Region, Kazakhstan, determined as 316 Ma by SHARMPI zircon dating of the ore-bearing porphyry and 315 Ma by Re–Os dating of molybdenite from the Borly porphyry Cu–Mo deposit (Fig. 1a; Chen et al., 2010). The southern West Junggar in China may be correlated with the North Balkhash porphyry copper ore belt in Kazakhstan.
5.2. Petrogenesis

5.2.1. Jiamantieliek intrusion

The Jiamantieliek intrusion consists of diorite and diorite porphyry, with SiO$_2$ contents ranging from 56 to 68 wt%, Al$_2$O$_3$ from 14 to 16 wt% and MgO from 1.78 to 4.71 wt% (Table 1). High Sr concentrations ranged from 551 to 915 ppm, except for one value of 371 ppm and low $^{87}$Sr/$^{86}$Sr ratios between 0.70369 and 0.70401. The Yb content ranged from 1.3 to 1.9 ppm, except for one outlier of 2.6 ppm, and the Y value ranged from 12.9 to 17.9 ppm except for two outliers of 19.0 ppm and 24 ppm (Table 1). Compositionally, samples from the Jiamantieliek intrusion are similar to those of modern adakites (Defant et al., 1991). However, compared to adakites, the Jiamantieliek intrusion has slightly lower abundances of REE and LILE (Fig. 6) and variable Sr/Y ratios (Fig. 10). They also contain a high content of MgO and a low SiO$_2$ content (Fig. 5) relative to adakites (Defant et al., 1991). In the Sr/Y vs. Y (Defant and Drummond, 1993) and (La/Yb)$_n$ vs. Yb$_n$ (Defant and Drummond, 1990) diagrams, the Jiamantieliek intrusion plots in a transitional adakite to normal arc field (Fig. 10). The Jiamantieliek rocks have less variable Zr contents, but variation in Zr/Nb ratios, suggesting that partial melting process played a role in their petrogenesis (Fig. 11). In addition, the Jiamantieliek intrusion is enriched in LILEs and LREEs, with a marked negative Nb anomaly (Fig. 6). The above data demonstrate that the Jiamantieliek intrusion has a genetic affinity between the adakites and diorites occurred in the normal arc. The former implies that the melt of the Jiamantieliek intrusion could be a product of the partial melting of a Junggar oceanic slab. The latter indicates that they were derived from a subduction-modified mantle source. The melt of the Jiamantieliek intrusion was therefore a product of the partial melting of a Junggar oceanic slab and later interaction with the subduction-modified mantle during ascent.

Experimental studies have demonstrated that Mg$^{#}$ is a useful index in discriminating melts purely derived from the crust from those involved in the mantle. Melts from the basaltic lower crust are characterised by low Mg$^{#}$ (<0.4) regardless of melting degree, whereas those with Mg$^{#}$ >0.4 can only be obtained when a mantle component is involved (Rapp and Watson, 1995). The Jiamantieliek dioritic rocks have a relatively high Mg$^{#}$ content (0.61–0.70) (Table 1; Fig. 5D), indicating the involvement of mantle components.
This result is supported by the isotopic data. Based on Nd–Sr–Pb isotope systematics, the Jiamantieliek intrusive magmas generally lie on a mixing line between the N-MORB and EM-I or EM-II component (Figs. 8a–d). Moreover, modelling based on the Sr, Nd and Pb isotopic composition indicated that the composition of all samples can be explained by mixing between about 70–80 wt% of melts with N-MORB characteristics (from a DM source) and 20–30 wt% of a sedimentary end-member, isotopically similar to EM-I or EM-II (Figs. 8b–d). Lead isotopic ratios of the samples indicated an EM-I source (Figs. 8e and f), which can be considered as an end-member of the mixing process in the genesis. Therefore, the Jiamantieliek intrusive rocks were derived from a depleted mantle source mixed with subducted EM-I. The lower positive $\varepsilon_{Nd}(t)$ value, which corresponds to Nd mean crustal residence ages in the range of 544–688 Ma (Table 3) for the Jiamantieliek samples, indicated an involvement of Neoproterozoic primitive crust in the formation of the Barluke magmatic arc.

5.2.2. Baogutu intrusion

The Baogutu dioritic rocks also have a genetic affinity with the adakites and diorites. In the Sr/Y vs. Y and (La/Yb)n vs. Ybn diagrams, the samples plot in the transitional adakite to normal arc field (Fig. 10). In the Zr/Nb–Zr diagram (Fig. 11), the Baogutu dioritic rocks display a comparable trend to the partial melting process (Fig. 11). In comparison with the Jiamantieliek intrusion, the Baogutu rocks are derived from a more depleted mantle source, with minor mixing of subducted EM-I (Figs. 8a and b). The isotopic signature of the Baogutu intrusion ($\varepsilon_{Nd}(310) = +4.1$ to +6.0; $I_{Sr} = 0.70355–0.70385$ and the corresponding Nd model ages (474–537 Ma, except for one outlier of 618 Ma, Table 3) both support less involvement of Neoproterozoic primitive crust in the formation of the Kelamay magmatic arc.

The Jiamantieliek and Baogutu intrusions have similar characteristics, indicating that a relatively uniform and integrated source region has existed in the southern West Junggar since the Palaeozoic. This suggests that this source region is related to the long-lasting residual oceanic basin and related lithospheres since the Palaeozoic in North Xinjiang. A greater contribution from an enriched mantle source and more involvement of the Neoproterozoic crust would be required to have generated the Jiamantieliek intrusion in the Barluke Mountains.

5.3. Tectonic setting and mineralisation implications

5.3.1. Tectonic setting

Confirming which series a rock belongs to is very important in discriminating the tectonic setting of intrusive rocks. In the SiO2 vs. K2O diagram (Fig. 5B), most samples from the Jiamantieliek intrusion plot on the medium-K calc-alkaline field and to a minor degree on the high-K calc-alkaline field. Calc-alkaline rocks are typical constituents of mature island arcs (e.g. Miyashiro, 1974). Most samples were enriched in LREE (Fig. 6) had Nb/Yb ratios between E-MORB and OIB (Fig. 12A). On discrimination diagrams these samples plotted on the volcanic arc granitoid field (Fig. 12B). Their $I_{Sr}$ values ranged from 0.70369 to 0.70401 and $\varepsilon_{Nd}(t)$ from +1.6 to +3.4, with Nd model ages ranging between 544 and 688 Ma. Their Pb initial ratios were less variable ($^{206}Pb/^{204}Pb = 17.41–18.23$, $^{207}Pb/^{204}Pb = 15.31–15.41$ and $^{208}Pb/^{204}Pb = 37.31–37.74$). Based on these characteristics, we inferred that the rocks from the Jiamantieliek intrusion formed in an island arc setting.

The Baogutu intrusive rocks have a transitional calc-alkaline to tholeiite character (Fig. 5B). Tholeiitic rocks may be associated with emerging island arcs (e.g. Miyashiro, 1974), mid-ocean ridges and back-arc basin spreading centres (e.g. Gill, 1976). Most diorites were enriched moderately in LREE (Figs. 6 and 7), low and E-MORB-like Nb/Yb ratios (Fig. 12A). Their isotopic composition ($\varepsilon_{Nd}(t) = +4.4$ to +6.0, $I_{Sr} = 0.70368–0.70385$) matched that of a depleted mantle source. Thus, it is very likely that the Baogutu intrusive rocks were also formed within an island arc setting.

The Baogutu intrusive rocks have a transitional calc-alkaline to tholeiite character (Fig. 5B). Tholeiitic rocks may be associated with emerging island arcs (e.g. Miyashiro, 1974), mid-ocean ridges and back-arc basin spreading centres (e.g. Gill, 1976). Most diorites were enriched moderately in LREE (Fig. 7) and had E-MORB-like Nb/Yb ratios (Fig. 12A). Their isotopic composition ($\varepsilon_{Nd}(t) = +4.4$ to +6.0, $I_{Sr} = 0.70368–0.70385$) matched that of a depleted mantle source. Thus, it is very likely that the Baogutu intrusive rocks were also formed within an island arc setting.

5.3.2. Tectonic evolution

In comparison to the Jiamantieliek intrusion, the Baogutu intrusion has a transitional tholeiite to calc-alkaline character (Fig. 5B), with moderate enrichment in LREE and LILE (Figs. 6 and 7), low and E-MORB-like Nb/Yb ratios (Fig. 12A) and a high $\varepsilon_{Nd}(t)$ (+4.4 to +6.0), suggesting that the Jiamantieliek intrusion could have formed in a normal island arc setting, while the Baogutu intrusion formed in a
relatively immature island arc setting. Therefore, there is a development of magmatic arc maturation towards the northwestern part of the southern West Junggar.

Most recently, some researchers consider that the Carboniferous Junggar Ocean may have been northwestward subducting beneath the Kelamay arc in the Early Carboniferous (Tang et al., 2009, 2010; Geng et al., 2009, 2011; Yin et al., 2010; Yang et al., 2012). On the basis of detailed field mapping and ages of the ophiolitic mélanges and coherent sedimentary rocks, Zhang et al. (2011a, 2011b) proposed there were double subduction systems in the Darbut area of the southern West Junggar in the Late Carboniferous. Our study about the Early Carboniferous volcanic rocks (submitted for publication) in the Kelamay region (southern West Junggar) supported the double subduction model. Therefore, an alternative model for their formation invokes a double subduction model, as proposed by Zhang et al. (2011a, 2011b).

In this model (Fig. 13) subduction along the Darbut fault was synchronous towards the northwest and southeast during the Late Carboniferous (310–313 Ma), creating the Kelamay intra-oceanic arc in the southeastern Darbut fault and the Barluk intra-oceanic arc in the northwestern Darbut fault. In this scenario, (1) melting of the subducted slab released melts/fluids which metasomatized the lithospheric mantle, creating an enriched mantle source; (2) mixing between the depleted- and enriched- mantle produced diorites and quartz diorites of the Jiamantieliek and Baogutu intrusions.

This geodynamic process led to the formation of subduction-related intrusions and associated porphyry copper deposits in the southern West Junggar (Shen et al., 2009, 2013), represented by the Baogutu ore-bearing intrusions and associated deposits in the Kelamay Region and the newly-discovered Jiamantieliek ore-bearing intrusions and associated deposits in the Barluk Mountains (Fig. 1b).

5.3.3. Mineralisation implications

In general, most porphyry copper deposits are generated in magmatic arc settings and associated with calc-alkaline magma (Hedenquist and Lowenstern, 1994; Richards, 2005; Sillitoe, 2010). In this contribution, we recognise that the ore-bearing intrusions of the southern West Junggar are calc-alkaline intermediate rocks and proposed that they form in an island arc setting. This island arc setting and associated with calc-alkaline magma is favourable for the formation of the porphyry Cu ore deposits in the Kelamay Region and the Barluk Mountains. Furthermore, the maturation of the magmatic arc and the increase in calc-alkaline magma content from the Kelamay Region to the Barluk Mountains are favourable for the formation of the Jiamantieliek porphyry-type ore deposit, although the Jiamantieliek deposit is still being explored.

6. Conclusions

(1) Both Jiamantieliek and Baogutu intrusions are the intermediate intrusions that comprise main-stage diorite stock and late-stage diorite porphyry dikes. The Jiamantieliek intrusion formed in 313 ± 4 Ma and 310 ± 5 Ma and the Baogutu intrusion formed in 313 ± 2 Ma and 312 ± 2 Ma.
(2) The Jiamantieliek rocks are calc-alkaline rocks and enriched in LREE and LIL with distinct negative Nb anomaly, suggesting generation in a normal island arc setting. Baogutu rocks are similar to those of the Jiamantieliek intrusion except for moderate enrichment in LREE and LIL and E-MORB-like Nb/Yb ratios with tholeiite signature, indicative of a relative immature island arc setting.

(3) The Jiamantieliek intrusion isotopic compositions ($^{143}Nd/^{144}Nd = 0.70369$ to $0.70401$, $^{207}Pb/^{206}Pb = 15.31$–15.41) match a mixing source of depleted mantle and EM-I, whereas the Baogutu intrusion isotopic compositions ($^{143}Nd/^{144}Nd = 0.70368$–0.70385) reveal a more depleted mantle source. They formed in an intra-oceanic arc setting.

(4) The Jiamantieliek and Baogutu intrusions have similar characteristics, indicating that a relatively uniform and integrated source region has existed in the southern West Junggar since the Palaeozoic. The magmatic arc maturation developed from east to west in the southern West Junggar, and is favourable for the formation of the porphyry-type ore deposits in the Barluk Mountains.

Acknowledgements

We are very grateful to Editor-in-Chief Bor-ming Jahn, associate editor Irene Yao and two anonymous reviewers for their comments, which significantly improved the paper. This study was supported by research grants from the Innovative Project of the Chinese Academy of Sciences (Grant No. KZCX-EW-LY02), National Natural Science Foundation of China (Grant No. 41272109, 40972064, 41232027, 40725009, 41190071, 41190072), National International cooperation in science and technology project (Grant No. 2010DFB23390), and National 305 Project (Grant No. 2011B060801).

References


Verlag, Berlin, pp. 1–189.


**Xiao, W.J., Huang, B.C., Han, C.M., Sun, S., Li, J.L., 2010. A review of the western part of the Altaiids: a key to understanding the architecture of accretionary orogens. Gondwana Research 18, 253–273.**


**Zhang, J.E., Xiao, W.J., Han, C.M., Ao, S.J., Yuan, C., Sun, M., Geng, H.Y., Cai, K.D., 2010. Late Carboniferous high-Mg dioritic dikes in Western Junggar, NW China: geochemical features, petrogenesis and tectonic implications. Gondwana Research 17, 145–152.**


**Zhang, J.E., Xiao, W.J., Han, C.M., Mao, Q.G., Ao, S.J., Guo, Q.Q., Ma, C., 2011b. A Devonian to Carboniferous intra-oceanic subduction system in Western Junggar, NW China. Lithos 125, 592–606.**
