In this review, we describe the geological characteristics and metallogenic–tectonic origin of Fe deposits in the Altay orogenic belt within the Xinjiang region of northwestern China. The Fe deposits are found mainly within three regions (ordered from northwest to southeast): the Ashele, Kelan, and Maizi basins. The principal host rocks for the Fe deposits of the Altay orogenic belt are the Early Devonian Kangbutiebao Formation, the Middle to Late Devonian Altay Formation, with minor occurrences of Lower Carboniferous and Early Paleozoic metamorphosed volcano-sedimentary rocks. The principal mineral-forming element groups of the deposits are Fe, Fe–Cu, Fe–Mn, Fe–P, Fe–Pb–Zn, Fe–Au, and Fe–V–Ti. The Fe deposits are associated with distinct formations, such as volcanic rocks, skarn deposits, pegmatites, granite-related hydrothermal vein mineralization, and mafic pluton-related V–Ti magnetite deposits. The Fe deposits are most commonly associated with volcanic rocks in the upper Kangbutiebao Formation, in the volcano-sedimentary Kelan Basin, and in skarn deposits at several localities, including the lower Kangbutiebao Formation in the volcano-sedimentary Maizi Basin, and the Altay Formation at Jaerbasidao–Kekebulake region. Homogenization temperatures of fluid inclusions in the prograde, retrograde and sulfide stages of the skarn type deposit are mainly medium- to high-temperature (cluster between 200 and 500 °C), medium-temperature (cluster between 200 and 340 °C) and low- to medium-temperature (cluster between 160 and 300 °C), respectively. Ore fluids in the sedimentation period in the volcano-sedimentary type deposit are characterized by low- to medium temperature (with a peak around 190 °C), low to moderate salinity (3.23 to 22.71 wt.% NaCl equiv). Ore fluids in the pregmatite type deposit are characterized by low- to medium temperature (with a peak at 240 °C), low salinity (with a peak around 9 wt.% NaCl equiv). An analysis of the isotopic data for Fe deposits from the Altay orogenic belt indicates that the sulfur was derived from several sources, including volcanic rocks and granite, as well as bacterial reduction of sulfate from seawater. The present results indicate that different deposit types were derived from various sources. The REE geochemistry of rocks and ores from the Fe deposits in the Altay orogenic belt suggests that the ore-forming materials were derived from mafic volcanic rocks. Based on isotopic age data, the timing of the mineralization can be divided into four broad intervals: Early Devonian (410–384 Ma), Middle Devonian (377 Ma), Early Permian (287–274 Ma), and Early Triassic (c. 244 Ma). The ore-forming processes of the Fe deposits are closely related to volcanic activity and the emplacement of intermediate and felsic intrusives. We conclude that Fe deposits within the Altay orogenic belt developed in a range of tectonic settings, including continental arc, post-collisional extensional settings, and intracontinental settings.
1. Introduction

The Altay Mountains are located in the southwestern part of the Central Asian orogenic system (or collage) and traverse northwest to southeast across nearly 500 km of China and parts of Russia, Kazakhstan, and Mongolia. The Altay forms an important component of the Central Asian orogenic belt and is comprised of a number of tectonic assemblages. The orogenic system is a collage of minor continental blocks, island arc fragments, and accretionary complexes that underwent Paleozoic subduction-related crustal accretion and subsequent overprinting during the Mesozoic and Cenozoic by intracontinental orogenic processes (Long et al., 2007; Sengör et al., 1993; Wang et al., 2006; Windley et al., 2002, 2007; Xiao et al., 2009, 2010; Yang et al., 2011a). While the southern margin of the Altay (which lies within the Xinjiang Autonomous Region) is one of China's major nonferrous and rare-metal metallogenic belts, the region also contains important Fe deposits. In addition to the occurrence of Fe ores, VMS (volcanogenic massive sulfide ore)-type deposits occur in Devonian volcano-sedimentary formations, including the Ashile Cu–Zn deposit, the Kekete Pb–Zn deposit, the Tiemuer Pe–Zn deposit, and the Dadonggou Pb–Zn deposit (Geng et al., 2010a; Liu et al., 2008b; Wan et al., 2010, 2011; Wang et al., 2002, 2003a). More than 100 Fe deposits (occurrences) have been discovered throughout the Altay orogenic belt, characterized by small scale, multi-stage, multi-origin and poly-mineral associations. Predominantly among the Fe deposits are large- and medium-scale Fe deposits like Mengku, Tuomoerte, Wutubulake, and Balabakebulak, as well as small-scale Fe–P deposits in Abagong, and Fe–Cu deposits in Qiaxia.

In general, the Fe deposits in the Altay orogenic belt have not been systematically studied; notable exceptions include the Fe deposits in the Mengku, Wutubulake, and Saerbulak regions, and Fe–P deposits in Abagong (Liu et al., 2009, 2010; Xu et al., 2010; Yang et al., 2010; Zhang et al., 1987, 2011). Although the Mengku Fe deposit has been relatively well studied, its genesis remains a subject of debate, and its mineralization type has been variously described as skarn, volcano-exhalative sedimentary, volcano-exhalative sedimentary + superposition–reformation, marine volcanism, and iron mineralization associated with a major strike-slip shear zone (Li et al., 2003a; Wang et al., 2003b; Xu et al., 2010; Yang et al., 2008; Zhang et al., 1987). Previous investigations have (1) explored the genetic relationship between Fe deposits and volcanism along the Abagong–Mengku region (Zhang et al., 1987), (2) classified the Cu–Pb–Fe–Zn metallogenic series that are associated with marine volcanism (Wang et al., 2002), (3) summarized the geological characteristics of typical Fe deposits in the Altay and analyzed the prospects (Zhang, 2003), and (4) discussed the volcanic processes and metallogeny of the Late Paleozoic (Niu et al., 2006). However, only a few papers on this subject have been published in the international literature (e.g., Wan et al., 2012; Xu et al., 2010; Yang et al., 2010), with most of the characteristics of these deposits reported solely in the Chinese literature.

This paper describes and reviews the principal geological characteristics, as well as the temporal and spatial distributions of Fe deposits in the Altay orogenic belt, and integrates the results of previous studies with new microthermometric, geochemical and age data on the deposits (primarily the Mengku, Wutubulake, Abagong, Tuomoerte, Qiaxia, Saerbulake and Jiaerbasidao). We then discuss the metallogenesis and tectonic setting of the Fe deposits.

2. Geological setting

The Xinjiang Altay is comprised of two principal tectonic elements: the Early Paleozoic North Altay continental-margin active belt, and the Late Paleozoic South Altay active continental margins of the Siberian Plate. The northern Altay can be further subdivided into the Devonian–Carboniferous volcano-sedimentary Norte Basin and the Kanasi–Keke-tuohai Paleozoic magmatic arc, while the southern Altay can be subdivided into the Devonian–Carboniferous Kelan backarc basin, the Carboniferous–Permian Ka'erba–Naleimu magmatic arc, and the Carboniferous Xika'erba forearc basin. The Erquis (also known as Ertix or Irtysh)–Burgen suture forms the boundary between the Siberian and Kazakhstan–Junggar plates (He et al., 2004).

Within Xinjiang, the Altay is divided into Northern, Central, and Southern provinces (Fig. 1). The Northern Altay is bounded to the south by the Hongsanzui–Nuoye Fault and is composed predominantly of Middle to Late Devonian andesite and dacite, and Late Devonian to Early Carboniferous metasediments. The Devonian volcanics are thought to have formed in an island arc setting with sediments deposited conformably on island arcs in the associated fore-arc basin (Wang et al., 2006). S-type granites of the Northern Altay are mainly Silurian and Devonian in age, with some younger (but less common) intrusions of Permian to Jurassic age.

The Central Altay (specifically in the Kanas–Koktokay region) is bounded by the Hongsanzui in the north, Abagong, and Bazhai faults in the south, and is composed mainly of Early Paleozoic metamorphic rocks. The exposed strata include thick Neoproterozoic to Middle Ordovician low-grade metamorphosed flysch-type rocks (Habahe Group), Upper Ordovician volcanic molasse and turrigenous clastic sequences (Dongxileke and Baihaba Formations), and Middle to Upper Silurian meta-sandstone (Kulumuti Formation). The age of the Habahe Group, which is composed of migmatic and quartz schist, remains controversial. Published interpretations range from Neoproterozoic–Middle Ordovician (Chen and Jahn, 2002; Li et al., 2006; Wang et al., 2002; Windley et al., 2002) to Late Ordovician (Li et al., 2003b) to Middle Ordovician–Early Devonian (Long et al., 2010; Yuan et al., 2007).
Granitoids (gneissic biotite monzogranite, gneissic granite, biotite granite, and two-mica granite) are widespread and are thought to have been emplaced during the Late Silurian to Early Devonian, with a few younger emplacement dates [e.g., the Altai No. 3 pegmatite in Keketuohai is Triassic (220–218 Ma) in age] (Wang et al., 2007; Zhu et al., 2006). The Central Altay is the most important part of the Altay microcontinent (Li et al., 2003b; Xiao et al., 2010).

The Southern Altay is bounded by the Abagong Fault to the north and by the Erqis structural belt in the south. The Southern Altay is composed mainly of Devonian metavolcanic rocks and their dominant units are as noted in the Altay Formation (Yuan et al., 2007). These granitoids (gneissic granite, gneissic granodiorite, and gneissic tonalite) are thought to have been emplaced during the Early Devonian, as evidenced by an age range of 400 ± 6 to 404 ± 8 Ma for the Mengku deposit (Xu et al., 2010; Yang et al., 2010), with some evidence for a less common Permian age of 286.6 ± 2.6 Ma for the Jiaerbasidao deposit. A few intrusions yield ages of Late Ordovician (e.g., Abagong: 461 ± 8 to 458 ± 3 Ma; Tiemierte 459 ± 5 Ma; Chai et al., 2010; Liu et al., 2008a), Late Carboniferous (e.g., Ashele: 318 ± 6 Ma; Yuan et al., 2007), Triassic (e.g., Jiangjunshan: 245 Ma; Wang et al., 2002), and Jurassic (e.g., Shangkelan: 181–177 Ma; Wang et al., 2002). The age data for igneous intrusions show peaks at 460, 400, 375, and 278 Ma (Wang et al., 2006; Yang et al., 2010).

The Erqis Fault, one of the largest transcurrent faults in central Asia, forms the boundary between the Altay orogen and the Kazakhstan–Junggar Plate (He et al., 2004; Sengör et al., 1993; Wang et al., 2006). The fault zone contains an ophiolite with a zircon U–Pb age of 390 Ma (Wang et al., 2003c), and is thought to be the site of early to middle Paleozoic subduction. During the late Paleozoic (290–246 Ma), the zone experienced large-scale sinistral displacement (e.g., Briggs et al., 2009; Laurent-Charvet et al., 2003).
3. Iron genesis

More than 100 Fe-ore-producing localities have been documented in the Altay (Fig. 2), with most of the deposits classified as mineralized occurrences. Only approximately 20% of the ore deposits have been explored. The Fe deposits in the Altay formed primarily in the Late Paleozoic (see the discussion below) have been affected by multiple tectono-hydrothermal events that have resulted in multi-stage metallogenesis. Furthermore, the ore deposits (occurrences) are concentrated in the volcano-sedimentary rock series, indicating a volcanic genesis for the ore bodies or source beds that supplied ore-forming materials.

In this paper, we classify the various genetic types of Fe based mainly on the mineralization. Based on the characteristics of ore-bearing rock series and deposits, and on geochemical data, we recognize five genetic types of Fe deposits in the Altay (Table 1): volcanic rock, skarn, granite-related hydrothermal vein, pegmatite and mafic-rock-related V-Ti-magnetite. Of these types, volcanic rock and skarn are the most common. The volcanic rock type can be further divided into volcano-hydrothermal and volcano-sedimentary subtypes. The characteristics of typical deposits are listed in Table 2.

4. Spatial distribution of iron deposits

The main ore-bearing rock series of the Fe deposits in the Altay are the metamorphosed volcano-sedimentary sequence of the Lower Devonian Kangbutiebao Formation. Other ore-bearing formations include the metavolcano-metasedimentary rocks of the Middle–Upper Devonian Altay Formation, and the metavolcano-metasedimentary association of the Lower Carboniferous Kalaerqis Formation and Lower Paleozoic. There is considerable variation in the spatial distribution of the various deposit types. The deposits of volcanic type occur mainly

---

Table 1

<table>
<thead>
<tr>
<th>Type</th>
<th>Host rocks and ore-controlling structure</th>
<th>Ore deposit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volcanic rock</td>
<td>Volcano-hydrothermal Volcano-sedimentary</td>
<td>Kangbutiebao Fe deposit, Abagong Fe–P deposit, etc</td>
</tr>
<tr>
<td>Skarn</td>
<td>Skarn in the exocontact zone of granite, diorite between limestone, volcaniclastic rocks and volcanic lava; skarn zone</td>
<td>Tuonoerte Fe–(Mn) deposit, Qiaxia Fe–Cu deposit, Boketubai Fe–Mn deposit, etc</td>
</tr>
<tr>
<td>Pegmatite</td>
<td>Pegmatite and contact zone</td>
<td>Mengku Fe deposit, Wutubulake Fe deposit, Jiaerbasitao Fe deposit, etc</td>
</tr>
<tr>
<td>Granite-related hydrothermal vein</td>
<td>Exocontact zone of granite and plagiomphibolite and gneiss; NE-trending fault</td>
<td>Liangkeshu Fe deposit, Tangbenqi Fe deposit</td>
</tr>
<tr>
<td>Mafic rock-related V-Ti-magnetite</td>
<td>Mafic rock; lithology</td>
<td>Kuwei V–Ti–Fe deposit</td>
</tr>
</tbody>
</table>
Table 2
Characteristics of some major Fe deposits in Altay, Xinjiang.

<table>
<thead>
<tr>
<th>Ore deposit</th>
<th>Metal assemblage</th>
<th>Host rocks</th>
<th>Intrusions</th>
<th>Hydrothermal alteration</th>
<th>Ore mineral assemblage</th>
<th>Gangue mineral assemblage</th>
<th>Length, thickness of orebody</th>
<th>Grade</th>
<th>Category</th>
<th>Type</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mengku</td>
<td>Fe–(Cu)</td>
<td>D1 granite, leucoleptite, amphibolite of lower Kangbutiebao Fm.; skarn</td>
<td>Tonalite, granite and diorite; 387, 384, 400 and 404 Ma</td>
<td>Skarn (Garnet, clinoopyroxene, amphibole, tremolite, epidote, calcite, albite), quartz, calcite, sericite, albite</td>
<td>Magnetite, maghemite, pyrite, pyrrhotite, chloropyrite, molybdenite, pyrite</td>
<td>Garnet, clinoopyroxene, amphibole, K-feldspar, plagioclase, epidote, chlorite, calcite, quartz, tremolite</td>
<td>No. 1 orebody 2322 m; 1.9–103.2 m; other orebody 50–763 m; 1.3–110.8 m; 60–960 m; 2.5–20.33 m</td>
<td>TFe average 24 wt.%–58 wt.%; Cu average 0.00 wt.%–1.53 wt.%</td>
<td>MFe average 23.3 wt.%–27.96 wt.%</td>
<td>L</td>
<td>Skarn</td>
</tr>
<tr>
<td>Wutubulake</td>
<td>Fe</td>
<td>D1 granite, amphibolite of lower Kangbutiebao Fm.; skarn</td>
<td>Tonalite; 386–388 Ma</td>
<td>Skarn (garnet, diopside, amphibole, tremolite, epidote, calcite, chlorite), quartz, calcite, albite</td>
<td>Magnetite, maghemite, pyrite, pyrrhotite, molybdenite</td>
<td>Diopside, garnet, amphibole, tremolite, biotite, plagioclase, quartz, epidote, chlorite</td>
<td>20–920 m; 1–40 m</td>
<td>TFe 20 wt.%–65.77 wt.%</td>
<td>M</td>
<td>Skarn</td>
<td>Zhang et al. (2011)</td>
</tr>
<tr>
<td>Balabakebulake</td>
<td>Fe</td>
<td>D1 amphibole plagiogneiss, granulite of lower Kangbutiebao Fm.; skarn</td>
<td>Granite</td>
<td>Skarn (garnet, diopside, amphibole, epidote, scapolite)</td>
<td>Magnetite, pyrite</td>
<td>Diopside, epidote, hornblende, garnet, plagioclase</td>
<td>Quartz, calcite, epidote, garnet, chlorite</td>
<td>80–870 m; 6.47–29.65 m</td>
<td>TFe 27.37 wt.%–68.20 wt.%</td>
<td>S</td>
<td>Skarn</td>
</tr>
<tr>
<td>Jiaerbasidao</td>
<td>Fe</td>
<td>D2–3 marble and metatamaccaresous sandstone of lower Altay Fm.; skarn</td>
<td>Biotite granite; 287 Ma</td>
<td>Skarn (garnet, epidote, actinolite, chlorite), quartz, calcite</td>
<td>Magnetite, pyrite, chloropyrite</td>
<td>Diopside, garnet, amphibole, tremolite, biotite, plagioclase, quartz, epidote, chlorite</td>
<td>200–1800 m; 1.4–16.56 m</td>
<td>TFe average 44.18 wt.%–67.21 wt.%; P₂O₅ average 3.8 wt.%–10.8 wt.%</td>
<td>TFe average 31.31 wt.%–33.69 wt.%; Mn average 4.45 wt.%–7.64 wt.%</td>
<td>M</td>
<td>Volcano-sedimentary</td>
</tr>
<tr>
<td>Abagong</td>
<td>Fe–P</td>
<td>D1 metavolcanic breccia, metatuff, granulite, schist and leucoleptite of upper Kangbutiebao Fm.</td>
<td>Tonalite, granite, amphibolite, tremolite, chlorite, albite, kaolinite, carbonate</td>
<td>Tonalite, amphibolite, tremolite, chlorite, albite, kaolinite, carbonate</td>
<td>Magnetite, maghemite, pyrite</td>
<td>Quartz, apatite, biotite, muscovite, actinolite, albite, calcite</td>
<td>200–650 m; 8.8–15.0 m</td>
<td>TFe average 30.56 wt.%–47.42 wt.%; Cu average 0.25 wt.%–0.44 wt.%</td>
<td>TFe average 29.40 wt.%–32.10 wt.%; average 30.56 wt.%–41.47 wt.%</td>
<td>S</td>
<td>Volcano-sedimentary + skarn</td>
</tr>
<tr>
<td>Tuomoerte</td>
<td>Fe–(Mn)</td>
<td>D1 metamaficous sandstone, metatuff, metabreccia tuff and marble of upper Kangbutiebao Fm.</td>
<td>Granite porphyry dyke; 401 Ma</td>
<td>Actinolite, tremolite, sericite, chlorite</td>
<td>Magnetite, maghemite, pyrite, chloropyrite, pyrrhotite, pyrolysite</td>
<td>Calcite, quartz, biotite, sericite, actinolite, epidote, chlorite</td>
<td>20–650 m; 8.8–15.0 m</td>
<td>TFe average 44.18 wt.%–67.21 wt.%; P₂O₅ average 3.8 wt.%–10.8 wt.%</td>
<td>TFe average 31.31 wt.%–33.69 wt.%; Mn average 4.45 wt.%–7.64 wt.%</td>
<td>M</td>
<td>Volcano-sedimentary</td>
</tr>
<tr>
<td>Qiaxia</td>
<td>Fe–Cu</td>
<td>D1 marble, metatamaccaresous silstone, chlorite quartz schist, tuffaceous metasiltstone of upper Kangbutiebao Fm.</td>
<td>Diorite porphyry dyke</td>
<td>Skarn, quartz, pyrite, barite</td>
<td>Magnetite, chloropyrite, pyrite</td>
<td>Quartz, chlorite, calcite, biotite, sericite, epidote, chlorite, barite</td>
<td>Fe orebody 50–650 m; 2.8–5.8 m; Cu orebody 50 m; 2.3–7.5 m</td>
<td>TFe average 20 wt.%–47.42 wt.%; Cu average 0.25 wt.%–0.44 wt.%</td>
<td>TFe average 29.40 wt.%–32.10 wt.%; average 30.56 wt.%–41.47 wt.%</td>
<td>S</td>
<td>Volcano-sedimentary + skarn</td>
</tr>
<tr>
<td>Liangkeshu</td>
<td>Fe</td>
<td>D2–3 chlorite quartz schist of upper Altay Fm.; skarn; pegmatite</td>
<td>Granite; 377 Ma</td>
<td>Epidote, garnet</td>
<td>Magnetite</td>
<td>Plagioclase, biotite, quartz, epidote, garnet</td>
<td>70 m; average 2.5 m</td>
<td>TFe 29.40 wt.%–32.10 wt.%; average 30.56 wt.%–41.47 wt.%</td>
<td>S</td>
<td>Pegmatite</td>
<td>This paper</td>
</tr>
<tr>
<td>Kuerqis</td>
<td>Fe</td>
<td>C1 plagiognness and amphibolite of Kalaerqis Fm.</td>
<td>Granite; 274 and 279 Ma</td>
<td>Clinopyroxene, chlorite, epidote</td>
<td>Magnetite</td>
<td>Hedenbergite, quartz, epidote, tremolite, biotite, actinolite, carbonate, plagioclase</td>
<td>290 m; 1–12.5 m</td>
<td>TFe average 41.47 wt.%</td>
<td>S</td>
<td>Granite-related hydrothermal</td>
<td>Yang et al. (2012)</td>
</tr>
</tbody>
</table>

Notes: L—large deposit class with Fe ore > 100 Mt; M—medium with Fe ore between 10 Mt and 100 Mt; S—small with Fe ore < 10 Mt.
in the upper Kangbutiebao Formation of the Kelan Basin, with ore bodies generally aligned parallel to beds within the formation. The deposits are medium to small in scale. Fe deposits of the volcano-hydrothermal type occur mainly around Kangbutiebao and Tiemiertie, villages, as well as along Abagong–Xitieshan region, and are composed predominately of Fe and P. These deposits occur in close spatial association with Pb–Zn deposits, as exemplified by the Abagong Fe–P and the Talate Pb–Zn–(Fe) deposits. The volcano-sedimentary type ore bodies are distributed along Qiaxia–Tuomoerte region and are dominated by Fe, Mn, and Cu deposits. This type of ore body is associated spatially with Pb–Zn deposits, as exemplified by the Tuomoerte Fe–(Mn), Qiaxia Fe–Cu, and Tiemiertie Pb–Zn deposits.

The skarn-type deposits are found mainly in the Maizi Basin and at Jiaerbasidao–Kekebulak region. The deposits at Jiaerbasidao–Kekebulak region occur in skarns and marbles in a contact zone between granites and limestones of Lower Paleozoic units and the Altay Formation. The limestones occur mainly as relicts in granites, and mafic dikes often occur near the contact zones between these rock types. The ages of the granites range from Early Devonian to Early Permian (see the discussion below). Documented Fe deposits include the Jiaerbasidao, Saerbulak, Mingjin-1, Kekebulak, and Tiemulike deposits. These occurrences are characterized by small-scale deposits which generally contain only Fe. The skarn-type deposits in the Maizi Basin occur mainly in the lower Kangbutiebao Formation, and the ore bodies are generally distributed parallel to beds in the formation. In both the ore bodies and nearby skarns, the dominant mineral assemblage is garnet, pyroxene, amphibole, actinolite, tremolite, epidote, and chlorite. The mineral assemblages within the skarns show considerable variability between the different deposits. The skarns are the products of limestone and volcanic rock (lava and clastic rock) metasomatism by magmatic thermal fluids. Fe mineralization is distributed in skarns of the exocontact zone between plutons, limestones, and volcanic rocks. The deposits are characterized by superposition mineralization. While earlier volcanic activities provided the source beds, the main metallogenic processes were related to retrograde alteration of the skarns (e.g., the Mengku Fe deposit). The formation of skarn is thought to have been related to Early–Middle Devonian intrusions of tonalite, granite, granodiorite, and diorite, and the ore bodies and skarns formed prior to regional metamorphism (Yang et al., 2008). The mineralization is dominated by Fe (locally associated with Cu) and the deposits are mainly medium to large in size (e.g., the large Mengku Fe deposit and the medium Wutubulak and Balabakebulak Fe deposits).

The granite-related hydrothermal vein type deposits are distributed throughout the Erqis Fault zone near the Fuyun County Town. Mineralization occurs in the exocontact zone of the granites, in Lower Carboniferous amphibolites and plajo-amphibolite gneisses, and in the country rocks. The distribution of mineralization appears to be controlled by the contact zone and fault geometry, and the ore bodies show an en echelon distribution. The mineralization is mainly vein-like and contains only Fe (e.g., the Kuerquis Fe deposit).

Pegmatite-type deposits are found throughout the southeastern Maizi Basin and near Liangkeshu region. These deposits occur in pegmatites of the Altay Formation (primarily within the contact zone between granites, and schists and granulites) and in metavolcano-metasedimentary rocks of the Kangbutiebao Formation. Fe mineralization typically occurs as banded and disseminated veins and patches, spatially and genetically related to pegmatites. Mineralization is dominated by Fe, and pegmatite-type muscovite deposits occur throughout the mining district (e.g., the Liangkeshu and Tablexier Fe deposits).

5. Geological features of iron deposits in the Altay

5.1. Mengku Fe deposit

The Mengku deposit in Fuyun County, located in the Maizi Basin, is the largest known Fe deposit in Xinjiang and lies approximately 67 km NW of Fuyun County Town. Discovered in 1953, its reserves are thought to be at least 200 million tons, the bulk of which (150 million tons) is found in the western section (Nos. 1–9 ore bodies) (Li et al., 2010). The deposit is located in the Third Member (Early Devonian) of the lower Kangbutiebao Formation, which is characterized by hornblende granulite, banded granulite, hornblende plagiogneiss, amphibolite, biotite schist, meta-ryholite, meta-sandstone, and leucoleptite (Fig. 3). Intrusive rocks are well developed in the ore district, and the ages of gneissic biotite, granites, tonalite, and biotite diorite are 400, 404, 378, and 384 Ma, respectively (Xu et al., 2010; Yang et al., 2010; this paper).

The mineralized zone of the Mengku ore district (which contains more than 40 ore bodies) covers an area of 5.5 km in length and about 400 m in width (Fig. 3). The host rocks of the Nos. 1–6 ore bodies (Fig. 3) in the western section of the ore district are primarily hornblende granulite and leucoleptite, and they contain skarn relicts with garnet and epidote. Marble is present as blocks, both inside and adjacent to the country rocks of the ore bodies. The host rocks of the Nos. 7–22 ore bodies in the eastern section are composed

---

![Fig. 3. Geological map of the Mengku Fe ore district (modified after Yang et al., 2010).](image-url)
mainly of garnet skarn and secondarily of hornblende granulite, leucoleptite, and marble. The country rock located north of the No. 1 and No. 9 ore bodies is gneissic granite and gneissic tonalite vein, respectively. The largest of the ore bodies (the No. 1 ore body, which includes the previously defined No. 7 and Fe-11 bodies) extends 2322 m in length, varies from 1.92 to 103.18 m in thickness (average thicknesses of 12.99–44.26 m), and extends to a depth of 580 m. The smaller ore bodies have dimension of 50 to 763 m in length and 1.3 to 110.8 m in width (Li et al., 2010). The shape of the ore bodies is complex, varying from bedded to lenticular, chambered, and irregular (Fig. 4). The ore bodies are mostly conformable with the surrounding country rock, with the occasional unconformable contacts attributed to skarn formation.

Seven ore types have been distinguished in the Mengku deposit based on mineral assemblages (Zhang et al., 1987): diopside–magnetite, garnet–magnetite, amphibole–magnetite, quartz–albite/quartz–magnetite/quartz–hematite, diopside–amphibole–magnetite, quartz–pyrite–magnetite, and apatite–magnetite. The structure of the ore bodies is predominantly massive and disseminated, and to a lesser degree banded, brecciated, spotted, and veined. The texture of the ores is primarily granoblastic, metasomatic–residual, palimpsest, and cataclastic. Ore minerals are primarily magnetite, with lesser amounts of maghemite (with minor pyrite), pyrrhotite, chalcopyrite, molybdenite, and hematite. Gangue minerals include garnet, diopside, amphibole, and feldspar, minor biotite, quartz, calcite, chlorite, epidote, tremolite, and sericite, and rare scapolite and apatite. The average

---

**Fig. 4.** Combinational geological section of No. 1 orebodies in the Mengku Fe deposit (after Li et al., 2010).
Fe₂O₃ grade of the ore bodies is 24 wt.%–58 wt.% with a peak in the distribution between 35 wt.% and 48 wt.%. Cu mineralization (grades of 0.001 wt.%–1.53 wt.%) occurs between prospect lines 110 and 118 in the western section of the No. 1 ore body (Li et al., 2010).

Wall-rock alteration types include skarn, quartz, carbonate, sericite, and albite. Skarn is more intensively developed in the eastern part of the mining area. Relics of garnet, clinopyroxene, and chlorite are found in ore body No. 1, and extensive skarns have been discovered in ore body No. 9 and its eastern extension. The dominant skarn minerals are clinopyroxene (mainly diopside, with minor pyroxene), garnet (mainly andradite, with minor grossular), amphibole, chlorite, albite, and epidote. Three periods of hydrothermal mineralization have been identified on the basis of field evidence and petrographic analyses: skarn, regional metamorphism, and tectono-magmatic hydrothermal. The skarn period corresponds to the main Fe mineralization of the Mengku deposit and can be further divided into three mineralization stages: (1) prograde stage: clinopyroxene + garnet + albite + scapolite + apatite; (2) retrograde stage: magnetite + amphibole + epidote + chlorite + tremolite + K-feldspar + plagioclase + quartz; and (3) sulfide stage: pyrite + pyrrhotite + quartz + calcite. The regional metamorphism stage resulted in the deformation of the ore bodies (with the magnetite ores undergoing recrystallization) and the formation of (epidote + chlorite + garnet) quartz and (quartz, garnet) calcite veins. In contrast, the tectono-magmatic hydrothermal stage is represented by the occurrence of large molybdenite–pyrite–chalcopyrite–quartz veins.

5.2. Abagong Fe–P deposit

The Abagong deposit, located in the Kelan volcano-sedimentary basin (about 25 km southeast of Altay City), is a small Fe deposit. The deposit occurs in the Second Member of the upper Kangbutiebao Formation and is composed of meta-tuff, meta-rhyolite, meta-volcanic breccia, breccia-bearing schist, amphibolite, and granulite (Fig. 5). Adjacent country rocks are primarily amphibolite, meta-tuff, biotite–quartz schist, quartz–albite porphyrite, and metabreccia. Voluminous breccias that occur near the ore bodies contain crypto-explosive, clastic, and magnetized breccias. Breccia pipes (related to Fe–P mineralization) are thought to occur in the mining district in a similar manner to that observed in the Washan and Gushan magnetite–apatite deposits of the Middle to Lower Yangtze River Valley (Hou et al., 2011; Mao et al., 2011). SHRIMP zircon analyses yield dates of 413 and 407 Ma for the meta-rhyolite (Chai et al., 2009).

The mineralization belt extends about 5 km along a NW–SE trend and can be differentiated into two distinct ore segments (NW and SE; the NW segment is also called the Xiaotieshan or Xitieshan deposit). Three main ore bodies within the belt have lenticular and vein forms, the geometry of which is controlled by volcanic faults. The veins obliquely cut structural horizons and have a sharp contact with wall rocks. A common feature of the ore bodies is branching and merging along strike and a wedge-shaped vertical distribution (Fig. 6). Specific dimensions of the principal ore bodies (Fe1, Fe2 and Fe3) are as follows: (Fe1) 850 m in length and 1.3–43.2 m in thickness (average thickness of 16.6 m); (Fe2) 1800 m in length and 5.0–16.7 m in thickness (average thickness of 7.5 m); and (Fe3) 200 m in length and 1–2 m in thickness (average thickness of 1.4 m). Four mineral assemblages have been identified in the belt: quartz–magnetite, apatite–magnetite, quartz–apatite, quartz–magnetite, and muscovite–quartz–magnetite. The ore types of the Abagong Fe deposit are dominated by fine-grained massive magnetite and dense disseminated magnetite, rare banded magnetite, and brecciated magnetite. In the brecciated magnetite ores, the breccia is developed in tuff (clasts of which are highly variable in size and shape) cemented by magnetite. The dominant ore minerals include magnetite, maghemite, and minor pyrite. Gangue minerals include quartz, apatite, calcite, biotite, muscovite, actinolite, sphene, and albite. The grade of Fe₂O₃ varies from 44.18 wt.% to 67.21 wt.%, and that of P₂O₅ varies from 3.8 wt.% to 10.8 wt.%. The total REE

---

**Fig. 5.** Simplified geological map of the Abagong Fe–apatite deposit.
The Tuomoerte Fe–(Mn) deposit, chlorite, epidote, and carbonate in the hanging wall, and epimoi- narse garnet. The alteration shows vertical zonation, with quartz, biotite, chlorite, epidote, albite, apatite, pyrite, quartz, and calcite, with Wall-rock alteration in the deposit consists mainly of tremolite, actinolite, and chlorite quartz schist, meta-rhyolitic tuff, meta-tuffaceous sandstone, meta-volcanic breccia. The host rocks contain chlorite schist, meta-tuffaceous sandstone, meta-calcareous siltstone, and marble. Meta-rhyolite from the upper Kangbutiebao Formation yields a SHRIMP zircon age of 407 Ma (Chai et al., 2009). The Fe ore body is cut by granite porphyry dikes that yield an LA-ICP-MS zircon age of 401 Ma.

Wall-rock alteration of the Tuomoerte Fe–(Mn) deposit is dominated by quartz, sericite, chlorite, garnet, actinolite, and tremolite. Three periods of mineralization can be distinguished on the basis of field evidence and petrographic analyses: volcancosedimentary, magmatic–hydrothermal, and regional metamorphism. The majority of the metallogenic activity occurred during the volcanic-sedimentary period, during which the Fe ore bodies formed (with associated Mn mineralization). During the magmatic–hydrothermal period, a biotite granite porphyry dike intruded the Fe ore bodies, resulting in the formation of veins of pyrite and magnetite (mainly networks of veinlets and other disseminated forms with some quartz veinlets and epidote). Magnetite-bearing skarns occur in areas of limestone wall rock. Similarly, Cu-bearing quartz veins and chalcopyrite occur in areas of sillstone and limestone. The regional metamorphism period was characterized by the deformation of the ore bodies and wall rocks, and the Fe minerals that formed during the volcanic-sedimentary period were replaced by magnetite, accompanied by grain coarsening via recrystallization.

6. Sources of ore-forming materials

6.1. Source of ore sulfur

We interpret that the ore sulfur was derived from volcanic rocks in the ore district, based on an analysis of 69 pyrite samples from the Mengku Fe deposit, which yield positive δ34S values ranging from 1.9‰ to 13.98‰ (mainly 3‰–10‰, with peak values of 4.5‰ and 7‰) (Fig. 9). These values differ significantly from those of VMS-type Keketabe Pb–Zn deposit (also located in the Maizi Basin), which have values of −11.1‰ to 4.5‰. The δ34S values of the stockwork zone are −1.0 to 4.5‰, and those of banded or massive ores range from −11.1 to −8.8‰, indicating that hydrothermal fluids may have had multiple sources and that bacterial sulfate reduction probably played an important role (Wan et al., 2010). The δ34S values for 20 sulfide samples from the Wutubulak Fe deposit vary from 3.7‰ to 10.2‰ (peak value of 6.5‰), similar to those of the Mengku deposit. The δ34S values for 20 sulfide samples from the Saerbulak Fe deposit vary between −3.2‰ and 16.9‰; this is a typical range for granites, which suggests that the sulfur was derived from granites located near the ore bodies. The δ34S values of 15 pyrite samples from the Ku’ertixi Fe deposit range between −7.2‰ and 0.7‰ (mainly between −3.9‰ and 0.7‰; with peak near 0.5‰); again, this is similar to δ34S values for mantle-sourced sulfur (0 ± 3‰, Hoefs, 1997), indicating that the sulfur was derived from mineralization-related granites.

The δ34S values for 19 pyrite samples and a single pyrrhotite measurement from the Abagong Fe–P deposit range between −4.3‰ and 5.2‰ (peak at −3.5‰), which implies that the sulfur was derived from volcanic rocks within the ore district. The δ34S values of 18 sulfide samples from the Tuomoerte Fe–(Mn) deposit are highly variable, from −20‰ to 13.1‰ (mainly −2‰ to 1‰ and 6‰ to 13‰, with peaks at −0.5‰ and 10.5‰). The sulfur isotope data suggest that the sulfur has multiple sources. For example, δ34S values in the range of −2‰ to 1‰ are indicative of mantle-sourced sulfur (and indicate an association with the biotite granite porphyry dikes that

![Schematic geological cross section of No. 52 prospecting line at the Abagong Fe-apatite deposit.](image)
intrude the ore bodies), whereas $\delta^{34}$S values in the range of 6‰ to 13‰ are indicative of sulfur derived from volcanic rocks. The $\delta^{34}$S value for one sample is strongly negative ($-20‰$), which is characteristic of biogenic sulfur. The $\delta^{34}$S values of 13 pyrite samples from the Qiaxia Fe–Cu deposit are all negative ($-14.2‰$ to $-1.1‰$ without a well-defined peak value), which is similar to the $\delta^{34}$S values for the Tiemierte Pb–Zn deposit (also located in the Kelan Basin). Negative $\delta^{34}$S values are indicative of light sulfur enrichment (typical of biogenic sulfur) and suggest that the sulfur originates from bacteria-reduced seawater. Such $\delta^{34}$S values are also evidence for sedimentation during the formation of the deposit.

In summary, we recognize four basic features of the sulfur isotope geochemistry of the Altay Fe deposits: 1) sulfur of the Mengku, Wutubulak, and Abagong deposits is derived mainly from volcanic wall rocks; 2) sulfur of the Saerbulak and Ku’erix deposits originates mainly from granites; 3) sulfur of the Qiaxia deposit is primarily from bacteria-reduced seawater; and 4) sulfur of the Tuomoerte deposit has a complex origin, being derived from volcanic rocks, granites, and to a lesser degree bacteria-reduced seawater.

### 6.2. Rare earth elements

The total REE content of apatite from the Abagong Fe–P deposit ranges from 1352.96 to 6986.33 ppm (average 3717.70 ppm), with $(\text{La/Yb})_N = 1.37$–$9.77$ and $\delta\text{Eu} = 0.22$–$0.30$. The data show enrichment in light REE, depletion in heavy REE, relatively weak fractionation of LREE, and moderately negative Eu anomalies. The REE characteristics of the Abagong deposit are similar to those of the Ningwu porphyrite Fe deposit in China (Yu and Mao, 2002) and the Kiruna-type Fe deposit in Sweden (Frietsch and Perdahl, 1995), suggesting that these three deposits have a common genesis (Liu et al., 2009). The REE patterns of apatite are very similar to those of volcanic rock within the ore district (Fig. 10), indicating that Fe mineralization was related to volcanism. The total REE contents of amphibolite in the Wutubulake Fe deposit are high, ranging from 174.42 to 324.19 ppm, with LREE/HREE = 4.61–9.69, $(\text{La/Yb})_N = 6.48$–$11.06$, $(\text{La/Sm})_N = 2.54$–$4.55$, and $\delta\text{Eu} = 2.31$–$3.04$. The REE patterns of skarns and ores show enrichment in LREE, depletion in HREE, and moderate positive Eu anomalies. The total REE contents of the ores and skarns range from 4.39 to 156.94 ppm, with LREE/HREE = 0.69–5.14, $(\text{La/Yb})_N = 0.14$–$6$, $(\text{La/Sm})_N = 0.29$–$2.79$, and $\delta\text{Eu} = 0.92$–$4.45$. The similar chondrite-normalized REE patterns of the skarns and ores indicate that they are genetically related (Fig. 10). $(\text{La/Yb})_N$, $(\text{La/Sm})_N$, and $\sum\text{La–Nd}$ vs. $\sum\text{Sm–Ho}$ vs. $\sum\text{Er–Lu}$ data show a relation between ores, skarns, and amphibolite, indicating that Fe mineralization was related to skarns, and that the Fe was derived from mafic volcanic rocks (amphibolite).
The total REE contents of ores in the Mengku Fe deposit range from 10.75 to 258.77 ppm, with LREE/HREE = 1.32–15.38, (La/Yb)_N = 0.93–31.65, and δEu = 1.33–2.96. The chondrite-normalized REE patterns of ores show enrichment in LREE, depletion in HREE, and moderate positive Eu anomalies. The REE patterns of ores are very similar to those of amphibolite within the ore district (Fig. 10), indicating that the Fe was derived from mafic volcanic rocks.

7. Characteristics of ore fluids

Homogenization temperatures and salinities of fluid inclusions in Fe deposits in Altay, Xinjiang are summarized in Table 3 and Fig. 11. Homogenization temperatures of fluid inclusions (in garnet and diopside) in the prograde stage of the skarn type deposits (e.g. Mengku, Wutubulake, Saerbulake and Jiaerbasitao) range mainly from 160 to 580 °C (cluster between 200 and 500 °C), and salinity ranges from 1.23 to 40.18 wt.% NaCl equiv (cluster between 7.6 and 15 wt.% NaCl equiv). Homogenization temperatures of fluid inclusions (in epidote, tremolite and chlorite) in the retrograde stage of the skarn type deposits (e.g. Mengku, Wutubulake, Saerbulake and Jiaerbasitao) range mainly from 160 to 550 °C (cluster between 200 and 340 °C), and salinity ranges from 2.07 to 21.75 wt.% NaCl equiv (cluster between 3.5 and 12.5 wt.% NaCl equiv). Homogenization temperatures of fluid inclusions (in calcite and quartz) in the sulfide stage of the skarn type deposits (e.g. Wutubulake, Saerbulake and Jiaerbasitao) range mainly from 160 to 466 °C (cluster between 160 and 300 °C), and salinity ranges from 0.71 to 45.33 wt.% NaCl equiv (cluster between 2 and 9.5 wt.% NaCl equiv).

Homogenization temperatures of fluid inclusions in the regional metamorphism period in the Mengku deposit range mainly from 140 to 513 °C, centering on peaks at approximately 190, 240 and 350 °C, and salinity ranges from 1.23 to 60.3 wt.% NaCl equiv (with a peak at 1.5 and 9.5 wt.% NaCl equiv).

In summary, homogenization temperatures of fluid inclusions in the prograde stage of the skarn type deposit are mainly medium- to high-temperature, and thus are mainly medium-temperature in the retrograde stage, and thus are mainly low- to medium temperature in the sulfide stage. The three metallogenic stages of the skarn type deposit have similar salinities, which range mainly from 1 to 19 wt.% NaCl equiv, with some minor 30 to 57 wt.% NaCl equiv. and the cluster between 8 and 12 wt.% NaCl equiv, does not change significantly. The fluids in the regional metamorphism period in the Mengku deposit show medium- to high-temperature and low-salinity features. Ore fluids in the sedimentation period in the volcano-sedimentary type deposit are characterized by low- to medium temperature (142 to 355 °C, with a peak around 190 °C), low to moderate salinity (3.23 to
22.71 wt.% NaCl equiv). Ore fluids in the pegmatite type deposit are characterized by low- to medium temperature (170 to 367 °C, with a peak distribution at 240 °C), low salinity (0.18 to 18.72 wt.% NaCl equiv, with a peak around 9 wt.% NaCl equiv).

8. Timing of iron deposits in the Altay Mountains

Table 4 provides a summary of isotopic age data for Fe deposits in the Altay. In addition to Re–Os age data of molybdenite in the Mengku and Wutubulake deposits, most of these dates are based on zircon SHRIMP and LA-ICP-MS U–Pb ages obtained from analyses of granitoids related to Fe mineralization. Although the absolute timing of Fe mineralization is poorly constrained by these isotopic dates, an approximate mineralization age may be inferred using this approach. We further acknowledge that there are considerable differences in mineralization ages among the deposits.

The Tuomoerte Fe–(Mn) deposit is a volcano-sedimentary deposit located in the middle-upper parts of the Second Member of the upper Kangbutiebao Formation. Fe mineralization in this deposit was controlled by the host volcano-sedimentary rocks. The age of the meta-rhyolite of the Third Member of the upper Kangbutiebao Formation is 406.7 Ma (Chai et al., 2009), and the age of a granite porphyry dike that intruded the Fe ore body is 401.6 Ma. We therefore propose an Early Devonian age (402–407 Ma) for the Fe–Mn mineralization.

Molybdenite Re–Os model ages of magnetite ores from the Wutubulake deposit are 243.6 ± 4.1 and 244.2 ± 4.2 Ma, indicating that mineralization occurred in the Early Triassic.

The immediate host rocks of ore bodies from the Mengku Fe-deposit include skarn, granulite, leucoleptite, and amphibolite. The Fe mineralization is related to retrograde skarns. The formation of skarns was closely related to the intrusion of biotite granite, tonalite, and biotite diorite in the Mengku ore district. Zircon SHRIMP and LA–ICP–MS U–Pb ages for the biotite granite, tonalite, and biotite diorite are 408 ± 8, 400 ± 6, and 384.1 ± 1.2 Ma, respectively (Xu et al., 2010; Yang et al., 2010; this paper). These ages indicate Early Devonian timing for emplacement of the granitoids. The ages of Fe mineralization are slightly younger than those of the intrusions (<408 vs. 381 Ma, respectively). Molybdenite from a large pyrite–chalcopyrite–molybdenite quartz vein within the No. 1 ore body yields a Re–Os isochron age of 261 ± 6.9 Ma (Yang et al.,...
Table 3

Microthermometric data for Fe deposits in Altay, Xinjiang.

<table>
<thead>
<tr>
<th>Name of deposits</th>
<th>Tuomoerte</th>
<th>Mengku</th>
<th>Wutubulake</th>
<th>Saerbulake</th>
<th>Jiaerbasidao</th>
<th>Liangkeshu</th>
</tr>
</thead>
</table>

9. Geodynamic–metallogenic framework for the iron deposits of the Altay

9.1. Metallogenesis of Fe deposits in an island arc setting

The Fe deposits in the Altay formed as a consequence of tectonic interactions between the Kazakhstan–Junggar and Siberian plates. During the Early–Middle Ordovician (500–462 Ma), northward subduction of the Paleo-Asian oceanic plate beneath the Altay microcontinent of the Siberian Plate resulted in the formation of a continental arc along the southern margin of the central Altay (Windley et al., 2002). Between 460 and 415 Ma, fluids generated by dehydration of the subducted plate moved through the hot mantle wedge, resulting in melting and the formation of granites. Examples of such granites include the Qiemuqieke granites (462 Ma) (Wang et al., 2006) and the Abagongbei–Tiemierte granites (462–458 Ma) from the Ordovician (Chai et al., 2010; Liu et al., 2008a), and 436 ± 4 Ma volcanic rocks (meta-crystalline tuff) in the Chonghuer Basin (Zeng et al., 2009).

Between 415 and 380 Ma, the Paleo-Asian oceanic plate continued to subduct, forming a series of continental-margin extensional fault-depression basins along the southern margin of the Altay. P–T conditions in the subduction zone eventually reached those of the eclogite facies, and the subducted plate tore and subsided, causing asthenospheric upwelling and resulting in heating and melting of the plate and associated sediments. The melt generated from subducted plate and sediments, along with asthenospheric mantle and mantle wedge, underplated the lower crust, resulting in melting of crustal materials and the formation of acidic volcanic rocks of the Kangbutiebao Formation. The depleted magma of the underplated material eventually formed the mafic volcanic rocks of this formation. The widespread granitic magmatism that occurred around 400 Ma has a geochemical signature characteristic of a continental arc (Wang et al., 2006). Between 415 and 400 Ma, Fe source beds formed via volcano-sedimentation processes in the lower Kangbutiebao Formation of the Maizi Basin. This was followed (between 400 and 384 Ma) by the formation of a suite of skarn mineral assemblages in the Mengku–Tielulike region, associated with interaction between a magmatic intrusion and surrounding hydrothermally altered and limestones and volcanic rocks. A large amount of magnetite was formed during retrograde metamorphism of the early skarns. This event was the main metallogenic stage for the deposits found in Mengku and Tielulike. Contact-metasomatic skarns and Fe mineralization occurred in response to granitic intrusions around 410 Ma in the Saerbulak area. In the Kelan Basin, volcanic activity...
Table 4
Summarized geochronological data for Fe deposits in Altay, Xinjiang.

<table>
<thead>
<tr>
<th>Ore deposit</th>
<th>Dated minerals/rocks</th>
<th>Dating method</th>
<th>Ages/Ma</th>
<th>Data sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mengku Fe deposit</td>
<td>Biotite granite</td>
<td>Zircon SHRIMP U–Pb</td>
<td>408 ± 8</td>
<td>Xu et al. (2010)</td>
</tr>
<tr>
<td></td>
<td>Tonalite</td>
<td>Zircon SHRIMP U–Pb</td>
<td>400 ± 6</td>
<td>Yang et al. (2010)</td>
</tr>
<tr>
<td></td>
<td>Biotite diorite dyke</td>
<td>Zircon LA-ICP-MS U–Pb</td>
<td>384.1 ± 1.2</td>
<td>This paper</td>
</tr>
<tr>
<td></td>
<td>Molybdenite from the large pyrite–quartz veins in No. 1 orebody</td>
<td>Re–Os (isochron ages)</td>
<td>261 ± 6.9</td>
<td>Yang et al. (2011b)</td>
</tr>
<tr>
<td>Wutubulake Fe deposit</td>
<td>Molybdenite from the ore</td>
<td>Re–Os (model age)</td>
<td>243.6 ± 4.1</td>
<td>This paper</td>
</tr>
<tr>
<td>Tiemulike Fe deposit</td>
<td>Granodiorite</td>
<td>Re–Os (model age)</td>
<td>244.2 ± 4.2</td>
<td>This paper</td>
</tr>
<tr>
<td>Tuomoerte Fe–(Mn) deposit</td>
<td>Meta–rhyolite of upper Kangbutiebao Fm.</td>
<td>Zircon SHRIMP U–Pb</td>
<td>389.3 ± 5.7</td>
<td>This paper</td>
</tr>
<tr>
<td></td>
<td>Granite porphyry dyke intrude into Fe orebody</td>
<td>Zircon LA-ICP-MS U–Pb</td>
<td>401.6 ± 0.6</td>
<td>This paper</td>
</tr>
<tr>
<td>Jiaerbasidao Fe deposit</td>
<td>Granite</td>
<td>Zircon SHRIMP U–Pb</td>
<td>286.6 ± 2.6</td>
<td>This paper</td>
</tr>
<tr>
<td>Saeerbulake Fe deposit</td>
<td>Granite</td>
<td>Zircon SHRIMP U–Pb</td>
<td>401.6 ± 0.6</td>
<td>This paper</td>
</tr>
<tr>
<td>Liangkesu Fe deposit</td>
<td>Granite</td>
<td>Zircon LA-ICP-MS U–Pb</td>
<td>376.7 ± 1.3</td>
<td>Jiao et al. (2011)</td>
</tr>
<tr>
<td>Kuerqis Fe deposit</td>
<td>Granite</td>
<td>Zircon LA-ICP-MS U–Pb</td>
<td>274.1 ± 0.5</td>
<td>This paper</td>
</tr>
</tbody>
</table>

Fig. 11. Histogram of homogenization temperatures from Fe deposits in Altay.
(upper Kangbutiebao Formation) resulted in the formation of the Abagong volcano-hydrothermal type Fe–P deposit, the Tuomoerte volcano-sedimentary type Fe–Mn deposit, and Qiaxia volcano-sedimentary Fe–Cu deposits.

Between 380 and 354 Ma, the fault-depression basin along the southern margin of the Altay remained in extension. The underthrust plate (a relic of the plate splitting discussed above) continued to perturb the mantle, resulting in asthenospheric upwelling that produced pillow basalts and bimodal volcanic rocks of the Altay Formation. In the Liangkeshu region, granites (377 Ma) intruded Altay Formation, simultaneously forming pegmatite veins and skarns, and resulting in Fe mineralization in and around the pegmatites.

9.2. Metallogenesis of Fe deposits in an anorogenic setting

During the Early Permian (287–274 Ma), the Altay orogenic belt was in a post-collisionsal extensional tectonic setting. The Jaerbasidao granites (287 Ma) intruded the Altay Formation, forming skarns at the contact with limestone. This skarn development resulted in the skarn-type Fe deposits. In the Kuerquis area, within the large-scale Erisqi shear zone, granites (278–274 Ma) intruded the Lower Carboniferous amphibolite and plagiogneiss gneiss, forming magmatic–hydrothermal related Fe deposits in the exoskarn contact zone. In the Mengku ore district, tectonomagmatic (?)-hydrothermal activity occurred in the Permian (261 Ma; Yang et al., 2011b), producing large chalcopyrite–pyrite–molybdenite–quartz veins; however, the Cu is sub-economic.

In the Triassic, the Altay orogenic belt was in an intracratonic tectonic setting. In addition to the development of granite-related pegmatite-type metal–muscovite deposits (Wang et al., 2002), skarn Fe deposits developed in volcano-sedimentary rock of the Kangbutiebao Formation in the Maizi Basin (e.g., the Wutubulake Fe deposit).

10. Conclusions

Based on the analysis and discussion presented above, we draw the following conclusions from this study.

(1) The host strata of the Fe deposits in the Altay are dominated by two formations: the Lower Devonian Kangbutiebao Formation and the Middle to Upper Devonian Altay Formation. Fe deposits also occur within a few Lower Carboniferous and Early Paleozoic volcano-sedimentary rock series. The genetic types of the Fe deposits are dominantly volcanic rock and skarn formations, with minor granite-related hydrothermal vein, pegmatite, and mafic–rock-related V–Ti–magnetite.

(2) The genetic types of Fe deposits are associated with particular formations: the volcanic rock-type deposits are mainly distributed in the upper Kangbutiebao Formation of the Kelan Basin; the skarn-type deposits are found in the lower Kangbutiebao Formation of the Maizi Basin, in the Altay Formation along the Jaerbasidao–Kekebulak; and the granite-related hydrothermal vein deposits are distributed in the Erisqi Fault belt. Other genetic types are distributed throughout various formations and show little systematic correlation with particular formations.

(3) We identified four epochs of Fe ore formation based on geochronology data: Early Devonian (410–384 Ma), Middle Devonian (377 Ma), Early Permian (287–274 Ma), and Early Triassic (c. 244 Ma), with a majority of the Fe deposits dating to the Early Devonian. Volcanic–rock-type deposits were emplaced in the Early Devonian, while skarn-type deposits formed throughout these four age ranges. Fe deposits in the Altay developed in a variety of tectonic regimes, including continental arc, post-collisional extensional, and intracontinental settings.

(4) Isotopic data indicate that sulfur in Fe deposits within the Altay originated from a variety of sources, including volcanic rocks, granites, and the bacterial reduction of sulfate from seawater. The source of the sulfur depends on the genetic type of the deposit. In particular, REE data from Fe deposits in the Altay suggest that most of the ore-forming material was derived from mafic volcanic rocks.

(5) Homogenization temperatures of fluid inclusions in the prograde, retrograde and sulfide stages of the skarn type deposit are mainly medium- to high-temperature, medium-temperature and low- to medium temperature, respectively. Ore fluids in the sedimentation period in the volcano-sedimentary type deposit are characterized by low- to medium temperature, low to moderately salinity. Ore fluids in the pegmatite type deposit are characterized by low- to medium temperature, low salinity.

Acknowledgments

This research was jointly supported by the National Key Technologies R&D Program (2006BAB070802-01, 2011BAB06B03-02), the Ministry of Land and Resources public welfare industry special funds for scientific research project (201211073), and the Geological Survey of China Projects (1212010786006). We are grateful to the leaders of the State 305 Project, No. 4 Geological Party of the Xinjiang Bureau of Geology and Mineral Exploration and Development, and No. 706 Geological Team of the Xinjiang Nonferrous GeoeXploration Bureau for their great logistical and moral support.

References


