A rare Phanerozoic amphibolite-hosted gold deposit at Danba, Yangtze Craton, China: significance to fluid and metal sources for orogenic gold systems

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Abstract

The Danba gold deposit is located in a poorly-documented gold province on the north-western margin of the Yangtze Craton. It is sited in Devonian sequences in a high-grade metamorphic terrane that includes an extensional metamorphic core complex. Around the deposit, peak metamorphic conditions of 6 ± 0.5 kbar and 650 ± 50 °C at ca. 193 Ma were followed by retrograde sillimanite-grade conditions of 4.5 ± 0.5 kbar and 550 ± 50 °C. The deposit is hosted in a broadly strata-bound ductile-brittle shear zone with high-T proximal alteration assemblages of biotite-amphibole-plagioclase and ore assemblages dominated by pyrrhotite, but with a strong association between gold and bismuth tellurides. Alteration mineral thermobarometers, together with heating/freezing studies of low-salinity $\text{H}_2\text{O}-\text{CO}_2-\text{CH}_4$ fluid inclusions, indicate P-T conditions of early ore deposition of approximately 4–5 kbar and 500–650 °C at around 185 ± 9 Ma indicated by Re-Os geochronology on ore-related molybdenite. In conjunction, all data demonstrate that Danba represents a Lower Jurassic hypozonal orogenic gold deposit that formed during post-peak metamorphic retrogression. The primary high P-T nature of the deposit, combined with its late-metamorphic timing, negate that the ore fluid was sourced via devolatilization of the hosting supracrustal sequences. A deep externally-derived ore-fluid source is required. The most likely source is the K– $\text{H}_2\text{O}-\text{CO}_2$ and ore-metal fertilized lithospheric mantle that was metasomatized during Neoproterozoic subduction. It is proposed that transition from lithospheric transpression to extension in the Jurassic triggered the devolatilization of this metasomatized lithosphere to cause the formation of this rare Phanerozoic amphibolite-hosted gold deposit at Danba.

Keywords Hypozonal orogenic deposit · Phanerozoic gold · Amphibolite-facies metamorphism · Metasomatized lithosphere · Re-Os age

Introduction

Hypozonal orogenic lode-gold deposits in amphibolite-facies domains are widely considered to be restricted to Precambrian,

largely Archean, terranes (Groves 1993; Knight et al. 1993), perhaps reflecting significantly higher mantle temperatures and resultant continental thermal gradients (Grove and Parman 2004, and references therein). Phanerozoic hypozonal orogenic gold deposits are generally considered to be essentially absent in the geological record (Goldfarb et al. 2001, 2005; Goldfarb and Groves 2015; Kolb et al. 2015).

A fundamental question related to hypozonal orogenic gold deposits is whether they formed syn- to post-peak metamorphism (Groves et al. 1998; Goldfarb et al. 2005) or formed pre-peak metamorphism with subsequent metamorphic overprint (Phillips and Powell 2009, 2010; Tomkins 2010). The former would imply a deep and external fluid source (Goldfarb and Groves 2015), whereas the latter would support a source of auriferous fluids from greenschist- to amphibolite-facies prograde metamorphism of abundant sedimentary sequences (Tomkins and Grundy 2009) from which gold and

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associated metals can be extracted by metamorphism at deeper crustal levels (Goldfarb et al. 2005; Pitcairn et al. 2006). In contrast, the Precambrian hypozonal orogenic gold deposits are largely sited in basalt-dominated sequences from which some ubiquitous gold-related elements such as arsenic cannot be extracted (Pitcairn et al. 2015).

Goldfarb and Groves (2015) proposed that, apart from metamorphism of enclosing sedimentary rocks, orogenic gold deposits could form from other fluid sources (Bierlein et al. 2006), including metamorphic devolatilization of subducting oceanic slab and overlying sediment wedge (Goldfarb and Santosh 2014; Groves and Santosh 2016) or from metasomatized lithosphere (Hronsky et al. 2012). Such source regions could explain the anomalous deposits such as those of the Jiaodong Province, China (Goldfarb and Santosh 2014; Deng and Wang 2016), and the Megashear Zone in northern Mexico (Goldfarb et al. 2007) where young deposits occur in terranes that were metamorphosed hundreds to thousands of million years earlier. Such models can also explain the mixed stable and radiogenic isotope signals of the deposits that indicate deep and extensive auriferous fluid pathways (Ridley and Diamond 2000).

In the context of this ongoing debate, the Jurassic Danba gold deposit of the north-western Yangtze Craton, as described below, is potentially critical as it is a very rare Phanerozoic

orogenic deposit with typical hypozonal characteristics (Gebre-Mariam et al. 1995). If the deposit formed at similar P-T conditions to its amphibolite-facies wall rocks, it would add support to a model in which ore-forming fluids are sourced from external sub-crustal environments even in a Phanerozoic cooler Earth. The Danba deposit is described and discussed below in order to resolve its genesis and define its significance to the ongoing debate on orogenic gold source models.

Geologic setting

Tectonic evolution

The Danba gold deposit is located on the north-western margin of the Yangtze Craton (Fig. 1a). To the north-west is the Songpan-Garzê accretionary prism (Fig. 1b), which is bordered by the Paleotethyan Garzê-Litang suture (Sengör and Natalin 1996; Reid et al. 2007) formed as a result of the closure of the Paleotethyan Ocean by the end of Triassic (Deng et al. 2014, 2017). In the Late Triassic, convergence between the North China, South China, and Qiangtang blocks, subsequent to the closure of the Paleotethyan Ocean, resulted in the filling of a thick (5–15 km) sequence of Triassic flyschoid

Fig. 1 Geological map of the Songpan-Garzê accretionary prism. **a** Simplified geotectonic map. **b** Geological map showing the stratigraphy, magmatic rocks, structures, and gold deposits in the region, modified after Xu et al. (1992), Jolivet et al. (2015), and Billerot et al. (2017). Note that a Mesozoic domal belt exists along the north-western margin of Yangtze Craton. Ages and types of granite are from Roger et al. (2004), Zhang et al. (2006, 2007), Xiao et al. (2007), Yuan et al. (2010), Sigoyer et al. (2014), and Chen et al. (2017)

sedimentary rocks in the Songpan-Garzê relict basin and formation of the accretionary prism (Weislogel et al. 2006; Weislogel 2008; Roger et al. 2010).

The north-western margin of the Yangtze Craton is characterized by a > 1000-km Mesozoic domal domain along the Longmenshan thrust nappe belt (Fig. 1b; Chen and Wilson 1996; Wallis et al. 2003). The domal domain comprises a series of extensional metamorphic core complex formed at ca. 180 to 160 Ma (Zhou et al. 2002, 2008). The Neoproterozoic crystalline basement, which extensively crops out along the domal domain (Fig. 1b), is a product of Triassic metamorphism of a 860–750 Ma Panxi-Hannan arc assemblage (Fig. 1a; Zhou et al. 2002, 2008; Zhao and Zhou 2008) that shows depletion of high field strength elements (HSFE) and enrichment of large ion lithophile elements (LILE) (Zhou et al. 2006a, b). This Neoproterozoic basement is overlain by a thick Silurian-Devonian metasedimentary sequence that itself is overlain by the Triassic flyschoid (Brugier et al. 1997; Roger et al. 2004).

Since the Mesozoic, there have been three widely-recognized tectonic events: 1) Late Triassic compression, in which shortening and thickening of the crust of the Songpan-Garzê accretionary prism induced intense folding, thrust faulting, and Barrovian-type metamorphism (Huang et al. 2003a, b; Weller et al. 2013); 2) Early Jurassic domal extension along the > 1000 km length of the north-western margin of the Yangtze Craton (Fig. 1b); and 3) relatively minor Cenozoic thrusting related to India-Asia collision (Roger et al. 2004).

Lithostratigraphy and metamorphism

Around the Danba area, Neoproterozoic crystalline basement of the Yangtze Craton is unconformably overlain by Paleozoic cover represented by Silurian, Devonian, and Permian schist; quartzite; marble; gneiss; and amphibolite. In turn, Triassic sandstones, carbonates, and turbidites unconformably overlie this thick Paleozoic cover (Harrowfield and Wilson 2005; Zhou et al. 2008). The Danba gold deposit is hosted in Late Devonian strata (Electronic supplementary material: ESM Fig. 1), which comprise a 5300–3200 m sequence of pelites, carbonate rocks, quartzites and mafic rocks metamorphosed at amphibolite to granulite facies conditions (Fan et al. 2013).

In the Danba area, amphibolite-facies metamorphism peaked during the Late Triassic to Early Jurassic (ESM Fig. 1), as identified by Billerot et al. (2017). After re-interpretation by Weller et al. (2013), Huang et al. (2003a, b) indicated that the Barrovian metamorphism evolved progressively from kyanite-grade conditions of 5.3–8 kbar and 570–600 °C at 205–190 Ma, to sillimanite-grade conditions at 4.8–6.3 kbar and 640–680 °C at 197–180 Ma. These metamorphic P-T estimates were based on Thermocalc modeling (version 2.6) and other thermobarometers, mainly on metapelite and amphibolite from the whole area, with timing of these events

based on monazite U-Pb geochronology. A sillimanite-grade metapelite (9827-1) in the north-west of the area (ESM Fig. 1), very close to the Danba deposit, yielded P-T conditions of 640 ± 25 °C and 4.8 kbar at 193.4 ± 5.2 Ma (Huang et al. 2003a, b), representing estimated peak-metamorphic conditions around the deposit.

Regional doming

The Danba dome and decollement structure (ESM Fig. 1) is represented by exposure of Neoproterozoic basement in a metamorphic core complex to the north-west of Danba (Jolivet et al. 2015). The core of the dome comprises mainly migmatitic granite and granitic gneiss complexes with U-Pb ages of zircon cores of ca. 830 Ma and metamorphic rims of 177 ± 3 Ma, the latter representing exhumation of the lower plate to form the extensional metamorphic core complex (Zhou et al. 2002). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of amphibole associated with foliation provided cooling ages of about 166–159 Ma for the regional doming (Zhou et al. 2008).

Granitic intrusions

Around the Danba area, widespread Late Triassic to Early Jurassic granites (230–170 Ma) intruded into the metamorphic strata and Triassic cover (Fig. 1; ESM Fig. 1). These granites can be subdivided into two age groups (Roger et al. 2004). The first group is represented by late-orogenic, ca. 230–200 Ma granites that have I-type and/or adakitic characteristics (Zhang et al. 2006; Xiao et al. 2007; Yuan et al. 2010). Both have $I_{\text{Sr}} = 0.7050$ to 0.7108 and $\varepsilon_{\text{Nd}}(t) = -1.0$ to -9.5 , and the adakites also have high $\text{Sr}/\text{Y} = 20$ –110 and $(\text{La}/\text{Yb})_{\text{N}} = 25$ –105. The second group comprises late- to post-orogenic A- and S-type granites containing mafic enclaves dated at ca. 210–180 Ma (Zhang et al. 2007; Sigoyer et al. 2014; Jolivet et al. 2015; Chen et al. 2017), consistent with intensive mantle-crustal interaction and lithospheric extension during this period. The A-type granites have $I_{\text{Sr}} = 0.7090$ –0.7123 and $\varepsilon_{\text{Nd}}(t) = -2.72$ to -4.26 with enrichment of HFSE (Sigoyer et al. 2014).

Deposit geology

Geological units

The amphibolite-facies metasedimentary rocks in the Danba gold deposit form a monocline trending 310 – 0° and dipping 45 – 75° south-west to west (Fig. 2; ESM Fig. 2). The immediate wall rocks to the ore bodies are quartzite, sillimanite-garnet-biotite schist, and amphibole-biotite schist (Figs. 2 and 3; ESM Fig. 2). Distal to the ore bodies, there is amphibolite and marble to the north-west, and biotite schist and granulite to the south-west (Fig. 2).

Fig. 2 Geological map of the Danba gold deposit showing the amphibolite-facies metamorphic wall rocks and major ore-controlling shear zones, modified after Fan et al. (2013)

Ore bodies of the Danba gold deposit are mainly contained in a series of N-S trending, bedding- and foliation-parallel shear zones (Fig. 2; ESM Fig. 2). This series of shear zones is cut by a set of later NW-trending reverse faults (Fig. 2). There are thin fine-grained granodiorite dykes in the wall rocks to the Danba ore bodies (Fig. 2), but the closest large granite intrusion is more than 10 km from the mine (ESM Fig. 1). The ore bodies had been mined to a depth of 600 m below the surface and there are no indications of a granite intrusion below the ore bodies. This is confirmed by the results of EH4 magnetotelluric sounding used during exploration at the mine (Fan et al. 2013).

Metamorphic petrography

Metamorphic mineral assemblages in the wall rocks of the Danba gold deposit indicate that peak metamorphism reached

amphibolite-facies conditions with a typical amphibole-plagioclase assemblage together with coarse-grained high-T, high-Ti red-brown biotite (Thomson 2001; Henry et al. 2005) (ESM Fig. 3a, b). In garnet + biotite + sillimanite + plagioclase + quartz assemblages (ESM Fig. 3c, d), sillimanite is common and ubiquitously associated with retrogressive muscovite that has replaced the biotite along its contacts with garnet grains. The biotite wraps around the garnet with a sigmoidal shape, producing distinctive pressure shadows that reflect initial growth during shear-related deformation (ESM Fig. 3c). Fractured garnet grains, with cracks infilled by minerals from the retrogressive metamorphic assemblage (ESM Fig. 3d), are also common.

The P-T conditions of metamorphism in the immediate vicinity of the gold orebodies are discussed in more detail below.

Fig. 3 Field photographs of the Danba gold deposit. **a** Outcrop of the Major Lode in flexures in the shear zone and amphibolite-facies wall rocks. **b** The thickest section of the Major Lode. **c** Auriferous quartz vein in the Major Lode

Alteration

Subtle, millimeter to centimeter scale, symmetrical alteration zones of biotite alteration and silicification typically surround auriferous veins (Fig. 4; ESM Fig. 4a–d), with variable amounts of sulfide, carbonate, and K-feldspar minerals (ESM Fig. 4e). In places, biotite, the main alteration mineral, is intimately intergrown with the ore minerals (ESM Fig. 4f).

Petrographically, alteration zones are characterized by quartz-biotite-amphibole-K-feldspar-plagioclase-calcite-scheelite assemblages. Hydrothermal quartz has extensively replaced the amphibole-plagioclase assemblage in the alteration zone (ESM Fig. 4a). Silicification, represented by milky microcrystalline quartz, is intimately associated with biotite and aggregates of sulfide and gold (Fig. 4b–e; ESM Fig. 4b). Adjacent to this silicification, pyrrhotite, which cuts the schistosity, becomes more abundant compared with unaltered wall rocks (ESM Fig. 4c). Plagioclase, K-feldspar and amphibole (ESM Fig. 4d–e) are intergrown with quartz and biotite.

K-feldspar is restricted to the gold-related alteration zones and is absent from distal wall rocks.

Ore body characteristics

The deposit is characterized by a thick ore zone known as the Major Lode with a length of about 2 km and dip of 50–90° west (Figs. 2 and 3; ESM Fig. 2), which is mined to a current depth of 300 to 600 m in different sectors of the mine. Exploration data show that the Major Lode has a true thickness range of 0.8–27.3 m (mean 8.0 m) and a gold grade range of 2.6–26.3 g/t (mean 7.0 g/t). The total gold resource, including the Major Lode and other minor ore bodies, is about 50 t (1.6 Moz) Au (Fan et al. 2013). It is thickest closer to the surface and generally thins with depth (ESM Fig. 2). Locally, the thickest gold lode completely occupies the shear zone (Fig. 2; ESM Fig. 2), which developed mainly in planes of weakness in the wall rocks, especially those between

Fig. 4 Wall rock alteration in Adit 4155 m of exploration line 14 (ESM Fig. 2). **a** Cross section illustrating thin alteration zone (Bio, biotitization) between the auriferous quartz vein (Qz) and wall rocks. **b** Silicification (Si) along the schistosity of biotite schist. **c** Biotite alteration within an auriferous quartz vein. **d** Silicification along the border of quartz vein and biotite schist. **e** Sulfide aggregate and silicification along the border of quartz vein with biotite schist. Note that the alteration is mainly proximal with very limited distal alteration

relatively incompetent biotite schist and competent quartzite (Figs. 2 and 3; ESM Fig. 2).

Mineralization types include quartz veins, quartz stockworks, disseminated and massive sulfides, and intermediate types (e.g.,

Fig. 5a). The gold-bearing sulfides are distributed both along the border of quartz veins with wall rocks and within the quartz veins (Fig. 4). In places, small late ore-stage, fine to medium grained, milky quartz veins cut early ore-stage, fine to coarse

Fig. 5 Petrography of ore and mineral assemblages. **a** Mixed ore with intergrown biotite (Bt) in pyrrhotite (Po) and coarse-grained quartz (Qz). **b** Scheelite (Sch) aggregate in quartz vein. **c** Intergranular native gold between pyrrhotite grains and quartz. **d** Coexistence of pyrrhotite, pyrite, and chalcocite (Ccp) with minor sphalerite inclusions along fractures. **e** Late-stage fine-grained milky quartz cross-cutting early-stage medium-grained transparent quartz. **f** Euhedral molybdenite (Mo) intergrown with pyrrhotite and quartz and accompanied by late-stage pilsenite (Pil). **g** Late-stage quartz and pilsenite cutting early-stage pyrrhotite. **h** Native gold associated with pilsenite. **i** Minor galena (Gn) intergrown with pilsenite

grained translucent quartz veins (Fig. 5e). Carbonate veins or veinlets are virtually absent in the quartz veins.

Ore minerals

Gold grains occur as inclusions in pyrrhotite and quartz (Fig. 5c), in fractures within them, or along grain margins. They are commonly intergrown with bismuth tellurides (Fig. 5h, i). Gold grains vary in diameter from $>5\ \mu\text{m}$ to 1 mm (mainly 25–200 μm). Gold fineness ranges from 778 to 896 (mean 821; Fan et al. 2013), which is lower than the gold fineness of >900 that typifies most Archean amphibolite-facies gold deposits (Morrison et al. 1991; Knight et al. 1993). The ore minerals at the Danba deposit can be described in terms of two distinct assemblages within a single mineralization event: 1) an early ore-stage with dominant pyrrhotite, scheelite, minor molybdenite, pyrite, chalcopyrite, sphalerite, and trace gold; and 2) a late ore-stage with dominant Bi-tellurides, trace chalcopyrite, sphalerite, galena, pyrite, and major gold.

In the early ore-stage, anhedral pyrrhotite, the dominant ore-related sulfide (Fig. 5a, c–g), occurs within the quartz veins and also along the foliation in wall rocks immediately adjacent to the ore and alteration zones. Pyrrhotite contains some native gold. Fine- to coarse-grained scheelite aggregates occur both in quartz veins and within massive polymetallic sulfide aggregates (Fig. 5b). Chalcopyrite and fine-grained pyrite are everywhere distributed along fractures in other minerals, commonly associated with pyrrhotite (Fig. 5d). Sphalerite is normally contained as small inclusions in chalcopyrite and pyrite. Minor euhedral molybdenite is intergrown with pyrrhotite and is sited on the boundaries between sulfides and quartz (Fig. 5f). Some coarse-grained quartz coexists with the sulfides (Fig. 5a), whereas most quartz distal to sulfide aggregates is fine to medium-grained and translucent with a greasy luster (Fig. 5e).

In the late ore-stage, bismuth tellurides, absent in early-stage ore, are sited in fractures in milky white medium-grained quartz (Fig. 5g) and commonly intergrown with coarse-grained native gold (Fig. 5h). The most common bismuth telluride mineral is euhedral pilsenite ($\text{Bi}_{2+x}\text{Te}_{3-x}$) rather than tetradyte (Bi_2TeS_2) or joseite ($\text{Bi}_4\text{Te}_{2-x}\text{S}_{1+x}$) (Fan et al. 2013), suggestive of low $f\text{S}_2$ at the late ore-stage. Rare pyrite and galena grains are sited on the margins of pilsenite grains (Fig. 5i). In many places, pilsenite has replaced pyrrhotite, suggesting that the telluride-bearing assemblage formed late in the paragenetic sequence. In a few cases, pilsenite is intergrown with late ore-stage chalcopyrite, galena, and pyrite.

Ore geochemistry

Silver, Bi, Cu, Pb, Te, and Zn are characteristically co-enriched with Au at Danba, with high Bi and Te most strongly

associated with high Au (Fan et al. 2013), compatible with the petrography of the ore assemblage. In general, Te and Bi show a strong positive correlation with Au as well as with each other, whereas W appears independent of other metals. Arsenic, with concentrations below 10 ppm, is a noticeable absentee from the ore body.

Analytical methods

Twelve core samples of wall rocks and 116 samples of gold ores and wall rocks from underground workings at Danba were collected for petrography, mineralogy, geochronology, fluid-inclusion analysis, and sulfur isotopic studies. Descriptions of all these samples are given in Table 1 of the electronic supplementary material.

Re-Os geochronology

Five ore-related molybdenite samples from the deposit were handpicked under a binocular microscope after crushing, cleaning, and sieving to 30–60 mesh. Osmium and rhenium were then separated by distillation and extraction following the procedures described in Du et al. (2004). The Re and Os isotope ratios for each sample were determined using TJA Plasmaquad ExCell inductively-coupled plasma-mass spectrometry (ICP-MS) at the Re-Os Laboratory of the National Research Center of Geoanalysis, Chinese Academy of Geosciences.

Repeated analyses of molybdenite standard HLP from a carbonate vein-type Mo-Pb deposit in the Jinduicheng-Huanglongpu area of Shaanxi province, China, were performed in order to test analytical reliability (Stein et al. 1997). The ^{187}Re decay constant of $1.666 \times 10^{-11}\ \text{y}^{-1}$ was used for calculating molybdenite ages. The uncertainty in each individual age determination is about 1.4%, comprising the uncertainty of the decay constant of ^{187}Re , uncertainty in isotope measurement, spike calibration for ^{185}Re and ^{190}Os , and individual weighing and analytical random errors.

Mineral chemistry (EPMA)

Compositions of metamorphic and alteration minerals (biotite, amphibole, and garnet, in particular), bismuth tellurides, and gold in 16 samples were analyzed using a JEOLJXA-8230 electron probe micro-analyzer combined with an INCAX-ACT energy spectrometer at Wuhan University of Technology. Different synthetic minerals and relevant standard minerals (from SPI Corp., United States) were selected for instrument calibration. In terms of wavelength-dispersive mode, the electron beam had an acceleration voltage of 15 kV and current of 10 nA, with 1 μm diameter and counting time of 30 s on each element. To improve the precision of Ti for Ti-thermometer calculation, the

counting time was increased to 60 s. The estimated precision for each element is better than $\pm 2\%$.

Fluid inclusion analyses

At Danba, all generations of representative quartz from quartz veins, stockworks, and disseminated ores were examined. Fluid inclusions were described and analyzed from samples representative of both the early and late-ore stages.

Twenty-four polished sections about 100 μm thick were prepared for fluid inclusion analyses. Cooling and heating experiments were carried out using a Linkam THMSG 600 heating–freezing stage (-198 $^{\circ}\text{C}$ to 600 $^{\circ}\text{C}$) attached to a Leitz transmitted-light microscope at the Fluid Inclusion Laboratories, China University of Geosciences, Beijing. Synthetic fluid inclusions of known compositions were used to calibrate the stage. Estimated accuracy from reproducibility of measurements is ± 0.2 $^{\circ}\text{C}$ for clathrate-melting temperatures ($T_{\text{m,cla}}$), melting temperatures of the carbonic phase (T_{mCO_2}) and CO_2 homogenization temperatures (T_{hCO_2}) with a heating rate of 0.5 – 1 $^{\circ}\text{C}/\text{min}$. The accuracy is ± 2 $^{\circ}\text{C}$ for homogenization temperatures (T_{htot}) with a heating rate of 1 – 2 $^{\circ}\text{C}/\text{min}$. At other times, heating/cooling rates were restricted to be < 10 $^{\circ}\text{C}/\text{min}$. Laser Raman spectroscopic analysis was conducted using a Renishaw-inVia spectrometer at China University of Geosciences, Beijing. The Ar+ laser wavelength was 514 nm, laser power 20 mW, diameter of laser beam spot 2 μm , and spectrometer resolution 2 cm^{-1} .

Fluid inclusion assemblages (FIAs), the concept of Goldstein and Reynolds (1994), were used to verify the consistency of the microthermometric data. The mean values of each FIA were used as representative values.

Sulfur isotope analyses

Twenty pyrrhotite samples from the Danba gold deposit were used for sulfur isotope analysis. Pyrrhotite grains were carefully handpicked individually under a binocular microscope after the samples were crushed, cleaned, and sieved to 200 mesh, to attain over 99% purity pyrrhotite separates. The EA-ISOPRIME100 mass spectrometer was used for sulfur isotope measurement, at the analytical laboratory of China University of Geosciences, Beijing. Sulfur isotope analyses were carried out utilizing standard samples GBW04414 and GBW04415, according to the method DZ/T 0184.14-1997. Environmental conditions were 25 $^{\circ}\text{C}$ and 15% humidity. The temperature was 1150 $^{\circ}\text{C}$ in the oxidized column and 850 $^{\circ}\text{C}$ in the reduction furnace. The sulfur isotope values, with analytical precision of about $\pm 0.2\text{‰}$, are reported using the δ notation in per mil, relative to the Cañón Diablo Troilite (CDT) standard.

Geochronology

The results for Re–Os isotope dating of euhedral molybdenite inter-grown with ore-related sulfides (e.g., Fig. 5f) are shown in Fig. 6 and listed in ESM Table 2. Analyses of five molybdenite samples from the Danba gold deposit yield a ^{187}Re – ^{187}Os isochron age of 185.3 ± 9.4 Ma (MSWD = 1.4) at the 95% confidence level. As the isochron age is more reliable than model ages, it is chosen to represent the formation age of the early ore-stage of the deposit.

Metamorphism and wallrock alteration

Contrasts between metamorphic and alteration mineral assemblages

Although some minerals, such as biotite, amphibole and plagioclase, occur in both the gold-related alteration zones and the more distal metamorphosed wall rocks, there are clear compositional differences between them (Fig. 7; ESM Tables 3 and 4). Metamorphic amphiboles (ESM Fig. 3a, b) are normally magnesia hornblende, with minor ferropargasite or, more rarely, pargasite (Fig. 7a). In contrast, amphibole intergrown with quartz (ESM Fig. 4a) from amphibole-biotite schist within the alteration zone is mainly edenite (Fig. 7a). Metamorphic plagioclase in unaltered amphibolite is typically oligoclase ($\text{An}_{0.24}$), whereas labradorite ($\text{An}_{0.65}$) dominates in alteration zones (Fig. 7b). Iron-rich biotite dominates in unaltered garnet-biotite schist and amphibolite wall rocks (ESM Fig. 3a–d), whereas Mg-rich biotite is abundant in alteration zones (Fig. 7c), and especially where layers of coarse-grained euhedral

Fig. 6 Molybdenite Re–Os isochron for the Danba gold deposit. The data are listed in ESM Table 2

Fig. 7 Classification and differentiation of metamorphic and alteration minerals from Danba using calculated cations per formula unit (pfu). **a** Classification of calcic amphibole (S1, 2 and 6) based on Leake et al. (1997). **b** Classification of plagioclase (S1, 2 and 6). **c** Classification of biotite (S1, 2 and 4–9) based on Abdel-Rahman (1994)

hydrothermal biotite occur at quartz vein margins or as aggregates within those veins (ESM Fig. 4b–d, f).

Peak metamorphic P-T conditions

P-T estimates of metamorphism of the wall rocks distal to the ore bodies, together with ore-related wall rock alteration, in the Danba mine were calculated using three methods: 1) amphibole-plagioclase thermometer (Holland and Blundy 1994) and corresponding calibration barometer (Bhadra and Bhattacharya 2007), 2) garnet-biotite-plagioclase-quartz thermobarometer (Berman 1991), and 3) Ti-in-biotite thermometer (Henry et al. 2005).

Metamorphic temperature estimates for peak sillimanite-grade metamorphism in unaltered amphibolite (ESM Fig. 3a, b) were obtained from the amphibole-plagioclase thermometer (Holland and Blundy 1994) which has good precision (± 40 °C) in the range 400–1000 °C and 1–15 kbar over a broad range of bulk compositions. It is based on the quartz-saturation dependent reaction:

The estimated metamorphic temperature is 660–760 °C (mean 710 °C) for the magnesia hornblende-oligoclase assemblage in wallrock amphibolite (ESM Fig. 5a; ESM Table 5). Using these temperatures, pressure estimates were calculated using the barometer of Bhadra and Bhattacharya (2007), which is a garnet-free amphibole-plagioclase barometer based on the reaction below.

Calculated metamorphic pressures of wall rock amphibolite are 5.7–6.9 kbar at 690–750 °C.

The Ti-in-biotite thermometer (Henry et al. 2005) was used to check the temperature calculated above. As the biotite schist in the deposit is either a metapelite or altered metapelite that contains peraluminous garnet and Ti-saturation minerals, such as ilmenite in biotite (ESM Fig. 3d), the thermometer is ideally suitable for this calculation. The calibration range for this thermometry is $X_{Mg} = 0.275$ –1.000, $Ti = 0.04$ –0.60 pfu, $P =$ roughly 4–6 kbar, and temperature = 480–800 °C. Precision of the thermometer is estimated to be ± 12 °C for higher temperatures. The final estimated metamorphic temperature is

640–670 °C (mean 660 °C) for Fe-rich biotite in wall rock amphibolite (ESM Fig. 5c; ESM Table 5). This is slightly lower than the temperatures calculated from the amphibole-plagioclase thermometer but still well within the range of amphibolite-facies conditions.

Post-peak metamorphic P-T conditions

P-T calculations for retrograde assemblages in sillimanite-garnet-biotite schist (ESM Fig. 3c, d) were performed for mineral pairs on garnet reaction rims, using the garnet thermobarometer, thermodynamic (TWQ version 2.4.3), and the internally consistent database of (Berman 1991; updated to 2007), to outline a small range in P-T space which is defined by the reactions (ESM Fig. 5b):

P-T estimates reflect the retrograde sillimanite-grade metamorphism of the wall rocks of the Danba deposit with a pressure of 4–5 kbar and temperature of 540–580 °C (ESM Fig. 5b; ESM Table 5).

Calculations using the Ti-in-biotite thermometer (Henry et al. 2005) yield 560–640 °C (mean 590 °C) for sillimanite-garnet-biotite schist, with eight of the nine data points concentrating in the range 560–600 °C (ESM Fig. 5c; ESM Table 5), in close agreement with temperatures derived from the garnet thermobarometer.

P-T conditions of wall rock alteration

The calculated temperature for wall rock alteration, based on the amphibole-plagioclase thermometer (Holland and Blundy 1994), is 590–630 °C (mean 610 °C) for the edenite-labradorite assemblage in amphibole-biotite schist from the proximal alteration zone (ESM Figs. 4a and 5a; ESM Table 5). In good agreement is the temperature of 600–650 (mean 620 °C) calculated using the Ti-in-biotite thermometer (Henry et al. 2005) on assemblages from the same sample (ESM Fig. 5c).

Temperatures calculated using the Ti-in-biotite thermometer are 560–670 °C (mean 640 °C) for Mg-rich biotite from the margin of the alteration zone (ESM Figs. 4b–d and 5c; ESM Table 5) and 580–660 (mean 630 °C) for Mg-rich biotite intimately intergrown with ore minerals in the proximal alteration zone (ESM Fig. 4f).

Fluid inclusion data

Fluid inclusion petrography

Fluid inclusions were studied in both early-stage and late-stage quartz (e.g., Fig. 5c). Four types of fluid inclusions are defined: abundant liquid-dominated H₂O–CO₂–NaCl inclusions (Fig. 8a–c, e); minor vapor-dominated CO₂–H₂O–NaCl inclusions (Fig. 8d); rare vapor-only CO₂ ± CH₄ inclusions; and aqueous inclusions. Both the medium to coarse-grained subhedral and euhedral early ore-stage quartz and turbid milky fine to medium-grained late ore-stage quartz contain major liquid-dominated and minor vapor-dominated fluid inclusions.

At room temperature, liquid-dominated inclusions contain both a liquid H₂O phase and liquid CO₂ phase (10–60 vol% of the inclusion). These inclusions are 5–30 µm in diameter, appearing both as clusters and discrete inclusions. Vapor-dominated inclusions, similar to liquid-dominated inclusions, also contain two phases, but the volume of liquid CO₂ is > 60%. They are both discrete inclusions or in coexistence with liquid-dominated inclusions. Vapor-only inclusions contain only CO₂. Other inclusions comprise two phases (liquid and vapor H₂O). The liquid phase CO₂ in all inclusions separates into liquid CO₂ and vapor CO₂ when temperature decreases to below about 20 °C (Fig. 8d).

Fluid inclusions as clusters in growth zones (Fig. 8a, c) and isolated inclusions are likely primary, while those in short intra-crystal trails are likely pseudo-secondary (Chi et al. 2017). Only inclusions considered to be primary or pseudo-secondary were examined, and obvious secondary fluid inclusions (Fig. 8c), distributed along fractures, are excluded.

Microthermometry of fluid inclusions

A total of 61 fluid inclusions assemblages (FIA) and isolated inclusions (a total of 206 measurements) were studied for microthermometry (ESM Tables 6 and 7). The melting temperatures of the CO₂ phase (T_mCO₂), clathrate-melting temperatures (T_m.cla), CO₂ homogenization temperature (ThCO₂), and final homogenization temperature (T_{htot}) were determined for liquid-dominated inclusions in early-stage and late-stage quartz. Liquid-dominated inclusions homogenized to the liquid phase through the disappearance of the vapor phase (Fig. 9a).

Homogenization temperatures of 300–436 °C (mean 355 °C) for the early ore-stage inclusions and 230–350 °C (mean 300 °C) for the late ore-stage inclusions were obtained (Fig. 9a). The highest temperature of 400–436 °C is at the lower end of the temperature range of 400–575 °C at which pyrite converts into pyrrhotite during metamorphism of graphitic schist (Ferry 1981; Yang et al. 2016).

Fig. 8 Photomicrographs (all in plane polarized light except f) showing various occurrences of fluid inclusions in quartz at Danba. **a** Growth zone in coarse-grained subhedral quartz of early ore-stage. **b** Typical liquid-dominated fluid inclusions of early ore-stage. **c** Growth zone and secondary fluid inclusions along fractures in late ore-stage quartz. **d** Vapor-dominated fluid inclusion of late ore-stage separating into triple phases below room temperature. **e** A cluster of liquid-dominated fluid inclusions in late ore-stage quartz constituting a fluid inclusion assemblage (FIA). **f** Raman spectra on vapor phase in a fluid inclusion (S4075-3-5) of the late ore-stage

The salinity (equiv. wt% NaCl), bulk composition and density of the two types of inclusions in the early and late ore-stages were estimated using the program FLINCOR of Brown (1989) with internal equations from Brown and Lamb (1989) and Bowers and Helgeson (1983). The relatively consistent results are presented in ESM Table 6. Overall, the aqueous-carbonic fluid inclusions have a total XCO₂ of 0.09 and salinity of 2.6 wt% NaCl equiv. For early ore-stage fluid inclusions, the mean XCO₂ is 0.10 and salinity is 1.6 wt% NaCl equiv. For late ore-stage inclusions, the mean XCO₂ is 0.09 and salinity is 3.4 wt% NaCl equiv.

Volatile compositions of fluid inclusions

The results of laser Raman spectroscopy spot analyses on the CO₂ phases in fluid inclusions show that they can contain both CO₂ and CH₄. In some samples, concentrations of CH₄ are

high and some exceed those of CO₂ (Fig. 8f). This is in accordance with microthermometric results that indicate melting temperatures of the CO₂ phase are depressed from the CO₂ triple point (−56.6 °C) to −65.6 to −56.8 °C by a species such as CH₄. The CH₄ content in the CO₂ + CH₄ phase was estimated from the V-X phase diagram of the CO₂–CH₄ system (Thiery et al. 1994). In inclusions from the early ore-stage, the carbonic phase has higher CH₄ contents (range 2–18 mol%, mean 5 mol%) than inclusions from the late ore-stage (range 1–7 mol%, mean 4 mol%).

Pressure correction for early ore-stage inclusions

In general, the homogenization temperatures of fluid inclusions commonly do not reflect the true temperatures of the ore fluid that formed the deposit. The temperatures of the early ore-stage fluid require estimation using an isochore P-T

Fig. 9 Fluid inclusion data of ore-related quartz from the Danba gold deposit. **a** Histograms of homogenization temperatures. **b** Representative isochores (dashed dark lines, designated in CO₂ volume percent: vol%, a value estimated in the laboratory) based on Brown and Lamb (1989) for pressure correction of the early ore-stage H₂O–CO₂–NaCl fluid; Calculated pressures from FLINCOR of Brown (1989) are listed in ESM Table 7

diagram with a known trapping pressure derived from the program FLINCOR of Brown (1989) (Fig. 9b). The pressure range is consistent with contemporarily regional metamorphic conditions of 4.8–6.3 kbar (Huang et al. 2003a, b) and 5 to 6.5 kbar (Billerot et al. 2017).

Correspondingly, the P–T diagrams for a H₂O–CO₂–NaCl fluid of Brown and Lamb (1989) were used to calculate the pressure correction, as it is applicable for high P–T fluid conditions ($P > 350$ °C; $P = 2$ –10 kbar). Their representative isochores ($p\text{CO}_2 = 0.8$ and 6 wt% NaCl) were chosen to simulate the fluid conditions of the deposit ($p\text{CO}_2 = 0.76$ and < 5.86 wt% NaCl).

Employing this correction, average trapping P–T conditions in the early ore-stage at Danba are approximately 500–550 °C at 4.5–5 kbar (Fig. 9b, blue dashed lines): 4.5 kbar is the trapping pressure at the transition from early to late ore stages, and 5 kbar approximates to the lower limit of the regional metamorphic conditions. Temperature conditions during the early ore stage at Danba are estimated using the maximum CO₂ volume percent isochore of the early ore-stage fluid inclusions (50%, S930-10-1 and S680-1-1; ESM Table 6). As salinity has a limited influence on the isochores compared with CO₂ density, the widely applied pressure-correction system for H₂O–NaCl fluid from Potter (1977) can also be used to verify the above result. In this system, the temperature would be corrected over a broad range from 350 to 400 °C to over 500–550 °C.

The P–T estimates thus fall within amphibolite-facies metamorphic conditions in accordance with more robust P–T conditions calculated from wall rock alteration silicate minerals via thermos-barometers.

Sulfur isotope ratios

The $\delta^{34}\text{S}$ data for pyrrhotite from the Danba gold deposit are presented in Fig. 10 and listed in ESM Table 8. Twenty analyses range from 3.1 to 9.9‰ with a mean of 7.8‰ and median of 8.3‰. The first and third quartiles are 7.2‰ and 8.9‰, respectively.

Discussion

Review of critical age data for mineralization

The linearity in the ^{187}Re – ^{187}Os plot in Fig. 6 suggests that the mineralized assemblages evolved as an isotopically closed system: the Re–Os system has not been reset (Gannoun et al. 2003). Molybdenites, especially those having high Re concentrations, such as those with up to 10^5 ng/g at Danba (ESM Table 2), are difficult to disturb (Selby et al. 2002), even after solid-state recrystallization at granulite facies metamorphic P–T conditions and in fluid-present deformation environments (Stein et al. 1998, 2001, 2003; Bingen and Stein 2003). The Re–Os thus represents the timing of gold mineralization under post-peak metamorphic conditions.

Classification of Danba gold deposit

The Danba gold deposit is situated adjacent to a ductile-brittle shear zone within an amphibolite-facies metamorphic terrane. Local wall rocks experienced peak sillimanite-grade P–T conditions of 650 ± 50 °C and 6 ± 0.5 kbar and post-peak sillimanite-grade metamorphism at conditions of 550 ± 50 °C and $4.5 \pm$

Fig. 10 “Box and whisker” plots of $\delta^{34}\text{S}$ values of gold-bearing pyrrhotite from the Danba gold deposit, and gold-related pyrite from the nearby Yanzigou gold deposit. These are compared with variations in the sulfur isotopic compositions of sulfides in global sediment-hosted orogenic gold deposits through geologic time that relate to the time-dependant marine sulfate curve (Goldfarb et al. 1997; Chang et al. 2008). The $\delta^{34}\text{S}$ values for the Danba deposit are listed in ESM Table 8

0.5 kbar (Fig. 11). The Danba gold deposit formed broadly within the P-T range from peak to post-peak metamorphism. Best-estimate robust temperature conditions for wall rock alteration assemblages are 600 ± 50 °C, although pressure has not been determined. Best estimates for less-robust P-T conditions based on fluid inclusion analyses are 525 ± 25 °C and 4.5 ± 0.5 kbar (Fig. 11). These P-T conditions are consistent with the dominance of pyrrhotite rather than pyrite and the absence of gold-related carbonates in the ore assemblage.

These relationships between P-T conditions of regional metamorphism and gold mineralization are broadly compatible with available age constraints that indicate that gold was introduced at ca. 185 ± 9 Ma after regional amphibolite-facies metamorphism at ca. 193 ± 5 Ma (Huang et al. 2003a, b), with all events overlapping within geochronological constraints and probably occurring within a 10-My time frame (Fig. 11). This period of regional metamorphism and gold mineralization also overlaps the age range of the late- to post-orogenic A- and S-type granites dated at about 200–180 Ma (ESM Fig. 6).

The high temperature derived from alteration assemblages and fluid inclusions, combined with the presence of gold-related bismuth tellurides, raises the possibility that Danba is

Fig. 11 P-T diagram illustrating the peak metamorphic P-T conditions (amphibolite), retrograde P-T evolution (sillimanite-gamet-biotite schist), and estimated P-T conditions for formation of gold mineralization at the Danba deposit. Aluminosilicate triple point is from Holdaway and Mukhopadhyay (1993). And = andalusite; Ky = kyanite; Sil = sillimanite. The age of peak metamorphism is from Huang et al. (2003a, b)

an intrusion-related gold deposit (IRGD formed from $\text{H}_2\text{O}-\text{CO}_2 \pm \text{CH}_4$ fluids: Lang et al. 2000; Baker 2002). However, there are a number of lines of evidence against this hypothesis. Importantly, the nearest exposed granite intrusion is over 10 km from Danba and there is no evidence of a significantly proximal intrusion within the deposit environment from either geological or geophysical evidence. In support, zircons from the small dykes within the deposit (Fig. 2) are hydrothermally altered (Wang QF, unpublished data), indicating that the dykes were emplaced prior to the deposit. In addition, the minimum pressure estimate of about 4 kbar is outside the depth range of documented IRGDs. The sulfur isotope data (Fig. 10; ESM Table 8) are also inconsistent with a magmatic sulfur source, the $\delta^{34}\text{S}$ of which is normally considered to be around 0‰ (Gemmell and Large 1992).

The possibility that Danba represents a pre-existing gold deposit that was metamorphosed during the regional metamorphism, discussed above in terms of its Re-Os isotopic age, must be considered further in view of controversies related to high T-P deposits in high metamorphic environments, as summarized by Goldfarb and Groves (2015). A number of lines of evidence again argue against this model. First, the regional peak metamorphism is indistinguishable in age from that of the deposit. Second, hydrothermal quartz extensively replaced peak amphibole-plagioclase assemblage in alteration zones (ESM Fig. 4a). As elegantly demonstrated by Stanton (1972), the textures of mineral grains during solid-state recrystallization are controlled by their interfacial free energy, such that pyrrhotite, for example, should everywhere be interstitial to the silicate alteration minerals, but this is clearly not the case. The hydrothermal gold-related pyrrhotite crosscuts silicate minerals in the alteration zone (ESM Fig. 4c), whereas

pyrrhotite in wall rock is interstitial to silicate minerals along the schistosity (ESM Fig. 3a, b). Third, deposits formed under greenschist-facies conditions in fault zones universally have distal alteration zones that are far more extensive than proximal alteration zones (Eilu et al. 1999; Eilu and Groves 2001) and this geometry should be preserved during subsequent metamorphism. In contrast, Danba has essentially no discernable distal alteration (Fig. 7; ESM Fig. 2), implying that ore fluid was channeled along the hosting ductile-brittle shear zone. Finally, the presence of a strong Au-Bi-Te association is also consistent with formation of the Danba deposit under high-T conditions by analogy to their association in intrusion-related deposits, as discussed above. Such an association is very rare in mesozonal orogenic gold deposits, the probable precursor deposit type in any metamorphic overprint model.

The ore-element associations, potassic wall rock alteration, and low-salinity $\text{H}_2\text{O}-\text{CO}_2-\text{CH}_4$ ore fluid, combined with the strong structural control and late orogenic timing of gold mineralization, are all consistent with the classification of Danba as an orogenic gold deposit (Groves et al. 1998; Goldfarb et al. 2005). The P-T conditions fall within the range of those of slightly post-peak metamorphic, hypozonal orogenic gold deposits in amphibolite-facies metamorphic terranes elsewhere, as overviewed by Gebre-Mariam et al. (1995): $\sim 475^\circ\text{C}$ at 3 kbar to $\sim 700^\circ\text{C}$ at 6 kbar (Smith 1996). Such hypozonal orogenic gold deposits are relatively common in well-documented Archean greenstone belts from Western Australia (Bloem et al. 1994; Napier et al. 1998; Knight et al. 2000; Vielreicher et al. 2002), where they form an integral part of a crustal continuum (Groves 1993). Archean hypozonal orogenic gold deposits elsewhere, such as the Renco deposit in Zimbabwe (Kolb and Meyer 2002), Hutti deposit in India (Kolb et al. 2005), and New Consort deposit in South Africa (Dziggel et al. 2010) also formed under post-peak amphibolite-facies metamorphic conditions on a retrograde P-T path.

Danba appears to be an exceptionally rare, Phanerozoic example of a hypozonal orogenic gold deposit with substantial gold production, with previously recorded deposits almost universally representing mesozonal to epizonal orogenic deposits in sub-greenschist to greenschist-facies environments (Goldfarb et al. 2005; Goldfarb and Groves 2015).

Source of ore fluid

As discussed by Goldfarb and Groves (2015) and Groves and Santosh (2016), orogenic gold deposits that formed in terranes that were metamorphosed millions of years before the gold mineralization event, or were deposited under amphibolite-facies P-T conditions, cannot have a fluid source derived from regional metamorphism of enclosing continental rock sequences: the most commonly proposed ore-fluid source (Kerrick and Fyfe 1981; Phillips and Groves 1983; Colvine

et al. 1984; Cox et al. 1991; Powell et al. 1991; Bierlein and Crowe 2000; Goldfarb et al. 2001, 2005; Dubé and Gosselin 2007; Phillips and Powell 2009, 2010; Tomkins 2010).

At Danba, the previously devolatilized amphibolite-facies wall rocks of the gold deposit already appear to have been on a retrograde path during gold mineralization. Hence, generation of the fluids responsible for gold mineralization must have been external to the hosting continental rock sequences (Otto et al. 2007; Kolb et al. 2015), and must be deeply derived (Ridley and Diamond 2000).

The major possibilities for such a deep fluid source appear to be devolatilization of the sediment wedge above a subduction zone (Goldfarb and Santosh 2014; Groves and Santosh 2016) or lithosphere metasomatized during subduction-related fluid release (Hronsky et al. 2012; Goldfarb and Santosh 2014).

In order to determine if either of these is a viable ore-fluid source for the Danba gold deposit, it is necessary to examine the tectonic history of the hosting terrane.

As discussed above, the Danba gold deposit formed in a Mesozoic post-collisional transpressional to extensional regime after the Late Triassic closure of Paleotethys. The regional granites (ESM Fig. 6) in the Songpan-Garzê accretionary prism were generally emplaced later than the magmatic rocks in the Yindun arc that were related to subduction of the Paleotethyan Ocean (Reid et al. 2007). The regional hybrid granites evolved from late-orogenic adakites and I-type granites (230–200 Ma) to late- to post-orogenic A- and S-type granites (210–180 Ma) that overlap the period of regional amphibolite-facies metamorphism and gold mineralization (ESM Fig. 6). Importantly, the occurrence of mafic enclaves in the granite indicates a mixing between mantle-derived melts and the granitic magma (Sigoyer et al. 2014; Chen et al. 2017). The marked diversity of granite types in the late- to post-collisional environment, especially the occurrence of A-type granite, has been interpreted to be related to delamination of lower lithospheric mantle or other similar processes (Zhang et al. 2007; Yuan et al. 2010; Sigoyer et al. 2014; Chen et al. 2017), all involving asthenosphere upwelling.

In the context of this tectonic model, the most probable source region for the Danba ore fluid, in the absence of broadly synchronous subduction, is metasomatized lithosphere beneath the Neoproterozoic basement (Fig. 12). This metasomatism was most likely related to the subduction of oceanic crust from ca. 950 to ca. 750 Ma in the north-western Yangtze Craton, which resulted in the partial melting of the mantle wedge by addition of fluids and melts from the subducted oceanic plate (Zhou et al. 2002; Zhao and Zhou 2008). The resultant melts were largely emplaced in the crust to form the igneous assemblage in the Panxi-Hannan arc (Fig. 1a), now Neoproterozoic basement, while underplating of the oceanic crust and overlying sediment, created a metasomatized lithospheric mantle wedge.

Fig. 12 Schematic model for formation of the Danba hypozonal gold deposit in Paleozoic amphibolite-facies metasedimentary sequences. The ore fluid and metal source is interpreted to be underlying hydrated lithospheric mantle which was metasomatized during Neoproterozoic subduction. Upwelling asthenosphere in the Upper Triassic to Lower

Jurassic reactivated the pre-enriched lithosphere and triggered advection of high P-T auriferous fluid into a trans-crustal fault, interpreted to be the Xianshuihe Fault, with seismic pumping of over-pressured fluid into a shear zone on the margin of a metamorphic core complex. References for the ages are given for Fig. 1 and ESM Fig. 1

This metasomatized Neoproterozoic lithosphere appears to be the only viable source of ore fluid for the Danba orogenic gold deposit. The deeply derived fluid could be advected into the crust, probably via the strike-slip Xianshuihe Fault, and focused into the shear zone at Danba (Fig. 12). The mechanisms for migration into the crust by fluids created through devolatilization of metasomatized lithospheric mantle have been rarely discussed (Kennedy et al. 1997; Burnard and Polya 2004; Finlay et al. 2010; Klemperer et al. 2013), with seismic pumping along crustal-scale faults one logical mechanism (Cox 2016). Rapid passage of such deeply-sourced ore fluid through faults with zones of local water saturation may prevent the ore fluids from being consumed through partial melting (Schrauder and Navon 1994; Bureau and Keppler 1999; Klein-BenDavid et al. 2011; Rospabé et al. 2017).

The $\delta^{34}\text{S}$ range of +3.1 to +9.9 ‰ falls within the normal range between about 0 and +10 ‰ for orogenic gold deposits (Kerrick 1987, 1989; Golding et al. 1990; Nesbitt 1991; Partington and Williams 2000). The limited concentration range of +7.2 to +8.9 ‰ for most samples is also indicative of a single source as changes in redox and other chemical parameters at the site of gold deposition can only shift sulfur isotope compositions by a few per mil (Goldfarb and Groves 2015). Importantly, the nearby Yanzigou gold deposit, that is also contained in Devonian metasedimentary rocks, has $\delta^{34}\text{S}$ values of +7.47 to +9.35 ‰ for five pyrite samples in ore (Hou 2010), implicating a similar sulfur source to Danba. Although inconclusive, a comparison

between the sulfur isotope data for Danba and sea-water sulfate plus global sediment-hosted orogenic gold deposits through time (Fig. 10) shows that the sulfur in the ore sulfides of the Danba deposit could be derived from Neoproterozoic age sulfur in a subduction-related sediment wedge, whose devolatilization metamorphosed adjacent lithosphere, although Neoproterozoic sulfur in the basement or in the thick (5.3–3.2 km) Devonian hosting sequence are also possibilities.

The consistent ^{187}Re and ^{187}Os values that plots on a linear array in Fig. 6 suggest a common age and an isotopically homogeneous source (Gannoun et al. 2003). The high initial $^{187}\text{Os}/^{188}\text{Os}$ values 11 ± 14 also indicate a probable source from metasomatized lithospheric mantle (McInnes et al. 2008). The subducted crustal components incorporated into the lithosphere have provided large amount of radioactive Os and have increased $^{187}\text{Os}/^{188}\text{Os}$ ratios (Gannoun et al. 2003). Both the high Re concentrations of molybdenites (up to 10^5 ng/g; ESM Table 2) and $^{187}\text{Re}/^{187}\text{Os}$ ratios (up to ~300) support this explanation (Stein et al. 2001; Gannoun et al. 2003; Çelik et al. 2018).

The Re-Os data, although equivocal, are therefore at least consistent with a fluid and metal source derived from metasomatized lithospheric mantle.

Potential analogs for the Danba gold deposit

The Danba gold deposit is located on the edge of a metamorphic core complex during doming of the enclosing belt (Fig. 1;

ESM Fig. 1), related to post-orogenic transpression to extension and related Jurassic asthenosphere upwelling. Some other Phanerozoic orogenic gold deposits in high-grade metamorphic rocks are also related to decollement structures during metamorphic core complex formation related to regional extension caused by asthenosphere upwelling. Most of these are mesozonal orogenic gold deposits, deposited under P-T conditions similar to those deposits in Phanerozoic greenschist-facies domains, but they formed in Precambrian rocks that were metamorphosed many hundreds to thousands of million years before the gold mineralization event, as summarized by Goldfarb et al. (2001, 2007). These include the well-documented Jiaodong gold province of China (Goldfarb and Santosh 2014; Deng et al. 2015; Deng and Wang 2016) and the less-well documented deposits of the Megashear Zone of northern Mexico (summarized in Goldfarb et al. 2007).

However, the late Carboniferous to early Permian (315–285 Ma) gold deposits of the Variscan belt in the Massif Central of France, defined as orogenic gold deposits by Bouchot et al. (2005), are an exception. They include deposits with significant gold production in the Saligne (117 t = 4 Moz gold) and St-Yreix (37 t = 1.3 Moz gold) districts. Here, the time gap between early subduction and compressional orogenesis and gold mineralization related to post-orogenic transpressional to extensional doming and exhumation (Olivier et al. 2004) was less than 200 My, with gold mineralized domes parallel to the orogen (Whitney et al. 2004). The deposits described as “deep-seated gold deposits” by Bouchot et al. (2005) fit the classification of hypozonal orogenic gold deposits and show strong similarities to Danba. As summarized by Bouchot et al. (2005) they have an early ore-stage dominated by pyrrhotite and arsenopyrite with some loellingite, common in Archean hypozonal orogenic gold deposits from Western Australia (Neumayr et al. 1993), followed by a late ore-stage with gold, base metals and Bi-bearing minerals. Homogenization temperatures of low-salinity H₂O–CO₂–CH₄ fluid inclusions range from 260 to 450 °C, with interpreted P-T depositional conditions of up to 4.0–5.5 kbar and 450–500 °C, slightly lower than those for Danba, and commensurate with the presence of some pyrite and carbonate minerals in the ore bodies.

Other exceptions include some ~370 Ma lode gold deposits hosted in turbidite sequences in the Paleozoic Meguma Group, Nova Scotia, Canada (Kontak et al. 1990; Ryan and Smith 1998). Although most gold deposits are sited in greenschist-facies domains, several, including Beaver Dam and Cochrane Hill, are located in amphibolite-facies domains and are interpreted by Kontak et al. (1990) to be associated with retrogression during regional doming. They have ore mineral assemblages of pyrrhotite-scheelite-molybdenite-Bi-Te minerals, fluid temperatures approaching 450–500 °C, and sulfide $\delta^{34}\text{S}$ values of $9 \pm 1\%$. On the basis of these parameters, Kontak et al. (1990) suggest derivation of ore components from a sub-crustal source.

The hypozonal orogenic gold deposits of the French Massif Central appear to be the best analogs of Danba in terms of tectonic evolution, structural association, nature of economic ore bodies, P-T conditions of gold deposition, and interpreted fluid and metal sources, but several of the Meguma deposits are also probable analogs.

Conclusions

The anomalous Danba gold deposit is located in a poorly-documented gold province on the north-western margin of the Yangtze Craton. The province is characterized by a > 1000-km Mesozoic domal domain along the Longmenshan thrust nappe belt. The conjunction of a variety of research data from structural geology, metamorphic petrology, ore and alteration petrology, geochronology, and fluid inclusion and sulfur isotope studies demonstrate that Danba represents a Lower Jurassic hypozonal orogenic gold deposit formed within a P-T range of 4–5 kbar and 500–650 °C. It most likely formed during post-peak sillimanite-grade retrogression (4.5 ± 0.5 kbar and 550 ± 50 °C) subsequent to peak sillimanite-grade metamorphism (6 ± 0.5 kbar and 650 ± 50 °C). A combination of textural relationships between ore-related sulfides and silicate alteration minerals, the Au-Bi-Te association, and the virtual restriction of wall rock alteration to the proximal zone strongly indicate that the Danba deposit formed during the regional metamorphic event and was not metamorphosed after formation.

The primary high P-T nature of the deposit, combined with its late-metamorphic timing, make it highly unlikely that the ore fluid was sourced via devolatilization of the hosting rock sequences. A deep externally-derived fluid source is required. The timing of tectonic events in the region negates a direct source of ore fluid via devolatilization of the sediment wedge above a down-going subduction slab, as has been suggested for anomalous orogenic gold deposits such as those of the Jiaodong Province. The most likely source that meets both the constraints of the tectonic evolution of the area and the geological and isotopic constraints for the deposit is lithospheric mantle that was metasomatized during a Neoproterozoic subduction-related event. The lithospheric mantle was heated and reactivated by asthenosphere upwelling during a major Early Jurassic event involving a transition from transpressional to extensional tectonics. This asthenosphere upwelling was responsible for granite intrusion, regional metamorphism, doming, including the generation of metamorphic core complexes, and devolatilization of fertilized lithosphere to provide the Danba ore fluid. This is interpreted to have been advected into the crust, probably via the crustal-scale Xianshuihe Fault, and focused into the shear zone at Danba by seismic pumping.

Danba appears to be part of only the third well-documented and well-endowed Phanerozoic hypozonal orogenic gold district in an amphibolite-facies metamorphic terrane. The best analog is

represented by the Variscan gold districts of the French Massif Central, but several deposits in the Meguma gold districts of Nova Scotia also have strong similarities. Danba is similar in many respects to the majority of Archean hypozonal orogenic deposits which have been attributed by some authors (Kolb et al. 2015) to higher thermal gradients of 40–60 °C/km and locally up to 80 °C/km when the thermal regime of the early Earth was greater, due to significantly higher mantle temperatures. At Danba, as for the deposits of the Massif Central, such a thermal regime was probably derived more locally by widespread asthenosphere upwelling during a tectonic transition to extension.

As for the Archean examples of hypozonal deposits, a deep external source is required. For the Archean deposits, devolatilization of the sedimentary wedge above a subducting slab meets tectonic and crustal-scale architectural constraints. For Danba, a more indirect origin via devolatilization of lithosphere that was metasomatized in an earlier subduction-related event is the only reasonable model that meets all geological and geochronological constraints.

The key question, as was asked for the Jiaodong gold deposits (Groves and Santosh 2016), is whether Danba and its Phanerozoic analog deposits are simply exceptions in terms of Phanerozoic auriferous-fluid source or whether it has a more fundamental importance and questions acceptance of the widely-accepted Phanerozoic model of metamorphic devolatilization of crustal rock sequences, just as the Archean analogs question that model for the genesis of Precambrian orogenic gold deposits?

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