Phoenix: an Early Devonian granite-related tungsten deposit from the eastern Lachlan Orogen, New South Wales

Abstract

The Phoenix tungsten deposit, 97 km northwest of Goulburn, lies within the regionally extensive Frogmore Fault Zone bounding the Yass Shelf and the Hill End Trough. That deposit is hosted by fine-grained quartzofeldspathic rocks of the Early to Middle Ordovician Abercrombie Formation and felsic volcanic rocks of the Early Silurian Hawkins Volcanics. Two vein systems are present within the mine area: the Main vein system — which includes Main vein, Back vein, North East vein and Eastern lode; and the Western vein. These trend 330° and dip easterly at 50° to 60°. Tungsten-bearing quartz veins contain both scheelite and wolframite, together with a complex assemblage including pyrrhotite, magnetite, chalcopyrite, pyrite, native bismuth and sphalerite. Tourmaline, biotite, white mica and chlorite are present within the veins, as well as locally forming selvages.

Stage 1 of the Phoenix mineral paragenesis included emplacement of quartz–tourmaline veins with or without muscovite and/or fine-grained white mica. This was followed by a pyrrhotite–magnetite–wolframite assemblage (Stage 2). Scheelite was deposited in a quartz–scheelite–muscovite–biotite assemblage, with scheelite partially replacing wolframite in Stage 3. The final stage (Stage 4) involved deposition of a chalcopyrite–pyrite–native bismuth–bismuthinite–sphalerite assemblage with minor fine-grained white mica, chlorite, biotite, calcite and apatite.

Isotope data for lead and sulfur from sulfides were collected for 11 samples. Sulfur isotope values for sulfide vary between 3.3‰ and 5.8‰, supporting the interpretation for Phoenix that sulfur was largely derived from a magmatic reservoir. The initial ratio lead isotope data support the interpretation that lead was derived from an Early to Middle Devonian crust-derived reservoir.

\(^{40}\text{Ar}/^{39}\text{Ar}\) dating of biotite from quartz–biotite–scheelite–sulfide veins at Phoenix indicates that the mineralisation formed at 403.2 ± 1.1 Ma. The timing of mineralisation at Phoenix indicates that the deposit formed coincident with a wider Early Devonian magmatic event in the Eastern Subprovince of the Lachlan Orogen and increases the prospectivity of the Goulburn 1:250 000 map sheet area.

Keywords: Lachlan Orogen, Frogmore Fault Zone, mineral paragenesis, lead isotopes, sulfur isotopes, \(^{40}\text{Ar}/^{39}\text{Ar}\) dating, intrusion-related tungsten mineralisation, Phoenix mine

AUTHORS

P.M. Downes¹, P. Lennox², S. Wood³, D. Phillips⁴ & A. Dunlop⁵

¹Geological Survey of New South Wales, Industry & Investment NSW, Maitland
²School of Biological, Earth and Environmental Sciences, University of New South Wales, Kensington, NSW 2052
³Oil Search Limited, GPO Box 2442, Sydney, NSW 2001
⁴School of Earth Sciences, University of Melbourne, Parkville, Victoria, 3010

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Contact: john.greenfield@industry.nsw.gov.au

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Introduction

The Phoenix deposit, 97 km northwest of Goulburn, lies within the Palaeozoic Eastern Subprovince of the Lachlan Orogen (Figure 1). That deposit is within the regionally extensive Frogmore Fault Zone, which...
extends along the margin of the Yass Shelf and older basement units interpreted by Scott et al. (2004) as forming basement to the southern part of the Hill End Trough (Figure 2).

A number of tungsten-bearing quartz veins are present at Phoenix. These include the Main vein system (Main vein, Back vein, North East vein and Eastern lode) and the Western vein that crops out approximately 150 m west-southwest of the Main vein system (Figure 3). In general, the mineralised veins are poorly exposed at the surface. Samples for this study were collected from drillcore stored at the Industry & Investment NSW WB Clarke Geoscience Centre at Londonderry, mine dumps at the Phoenix mine and from surface outcrop because underground workings are inaccessible.

The Phoenix deposit was worked intermittently between 1914 and 1955. Recorded production totals 23 948 t of ore assaying 1.86% WO₃ (with a recovered grade of 1.48% WO₃) (Meszaros 1996). Most of this production came from stoping on the central 160 m portion of the 350 metre-long Main vein system (Figure 3). The Main vein was stoped from the surface to No. 5 level (97.5 m below surface). Additional limited stoping was undertaken on the adjacent Back vein between the surface and No. 3 level (48.1 m below surface). However, only minor development and stoping was undertaken on the Eastern lode (7.5 m east of Main vein) on No. 5 level. The Western vein was traced by surface trenching for 150 m, but the grades were ‘poor’ (Mulholland 1943). A drive from No. 5 level was commenced to test the Western vein at depth but was abandoned before it reached the lode position. Aside from limited underground diamond drilling of five horizontal holes ranging between 11 m and 41 m from No. 5 level (Tungsten Consolidated Frogmore mine undated plan), only two surface diamond drillholes have been completed in the mine area (Figure 3). Diamond drillhole DP3N-1 intersected minor mineralisation (1.0 m assaying 0.24% W — Willett 1979) 160 m below surface.

A number of workers, including Mulholland (1943), Gibbons (1960), and Felton and Wood (1977), have previously described aspects of the geology and mineralisation at Phoenix. Mulholland (1943) briefly described the mining history, extent of underground workings and potential resources in the mine area. Gibbons (1960) mapped the Frogmore area, including the Phoenix mine, as part of a BSc honours project. He briefly described a number of deposits — including the Phoenix mine, where he investigated the mineralogy of the tungsten-bearing veins, proposed a paragenesis for the mineralisation and interpreted the Phoenix deposit as granite-related. Felton (1977) reviewed the metallogenesis of the Goulburn 1:250 000 map sheet area, briefly described the mineralogy of the Phoenix vein system and suggested that the deposit was related to the Wyangala Batholith.

The aim of this paper is update the mineral paragenesis of Gibbons (1960), to present new ⁴⁰Ar/³⁹Ar dating results and the results of a reconnaissance sulfur- and lead-isotope study.

**Geological setting**

The Phoenix tungsten deposit lies within the Eastern Subprovince of the Lachlan Orogen in eastern New South Wales. This orogen is part of a >1000 km-wide orogenic system that developed along the palaeo-Pacific margin of the Australian craton during Palaeozoic time (Foster et al. 1999) and is an integral part of the Tasmanides (Scheibner & Basden 1996; Foster & Gray 2000; Glen 2005).

Two distinct packages of rocks dominate the Ordovician to early Silurian stratigraphy of the Eastern Subprovince of the Lachlan Orogen. One package includes quartz-rich, turbiditic sedimentary rocks of the Adaminaby Group and overlying black shale units with thin-bedded mudstones and rare turbidites of the Late Ordovician Bendoc Group.
**Figure 3.** Local geology and extent of underground workings of the Phoenix mine. The geology is adapted from Wood (2001). The extent of underground workings are from Mulholland (1943) and unpublished Tungsten Consolidated Frogmore mine plans (from Meszaros 1996).
The second package includes ultramafic (Wyborn 1992) to intermediate volcanic rocks and associated intrusions — as well as intercalated sedimentary rocks, including limestones, of the Early Ordovician to earliest Silurian Macquarie Arc (Glen et al. 1998). The Macquarie Arc is interpreted to be related to a subduction zone that formed outboard of the palaeo-Pacific margin of east Gondwana (Meffre et al. 2007).

Deformation in the early Silurian (Benambran Orogeny) resulted in juxtaposition of the Adaminaby Group and the Macquarie Arc (Glen et al. 2007). Following that, the eastern part of the Lachlan Orogen underwent prolonged crustal extension in response to slab rollback at the leading edge of the palaeo-Pacific oceanic plate and a retreating subduction boundary (Collins 2002). This latter event resulted in rifting and fragmentation of the eastern part of the Lachlan Orogen to form a number of north-trending, elongate palaeogeographic highs with intervening back-arc basins, including the Hill End Trough. The Hill End Trough, initiated in the late Wenlock, is a north-trending, back-arc rift basin (Scheibner & Basden 1998; Packham 2003) flanked on its western side by the palaeogeographic Molong High and Yass Shelf. Core to the Molong High are units of the Molong Volcanic Belt (Figure 1) that are part of the Macquarie Arc.

The Phoenix area is bisected by the crust-scale Frogmore Fault Zone that separates the Yass Shelf from the southern Hill End Trough (Scott et al. 2004). Thomas and Pogson (in prep.) suggested that the Frogmore Fault Zone formed during the Benambran Orogeny as a north-northeast-trending fault system — with further movement during both the Tabberabberan Orogeny (as north-trending faults) and the Kanimblan Orogeny (on north-northwest-trending faults). Furthermore, Downes et al. (2005) noted that this zone is a relatively high-strain, imbricate, east-dipping, north-northwest-trending thrust system with tight folds and locally intense penetrative to scaly cleavages with S–C fabrics, abundant kinks and quartz veins.

Wood (2001) identified three deformation-related events in the Phoenix mine area. The initial deformation resulted in early sinistral strike-slip movement along the Frogmore Fault that was then followed by later dextral strike-slip movement. In addition, Wood (2001) mapped a macroscopic antiform, with its axis trace trending 030°, that had developed between the Frogmore Fault and Main Shear (Figure 2).

Within the Phoenix area (Johnston, Pogson et al. in prep.) are quartz-rich, fine- to coarse-grained turbiditic sedimentary rocks of the Ordovician Abercrombie Formation (Adaminaby Group) that outcrop extensively to the east of and immediately west of the Phoenix mine area. To the southwest of the mine are mafic volcaniclastic and sedimentary rocks of the Ordovician Kenyu Formation (part of the Molong Volcanic Belt) (Figure 2). The Phoenix mine itself lies within the Frogmore Fault Zone. In the mine area are hornfels and metasedimentary rocks of the Abercrombie Formation and felsic volcanic rocks and related sedimentary rocks of the middle Silurian (Wenlock in Thomas & Pogson in prep.) Hawkins Volcanics that have been preserved as a fault slice within the Frogmore Fault Zone (Figure 3).

A number of plutons intrude the area (Figure 2). These include the Ballyhooley Granite, Reids Flat Granite and Licking Gully Granite that are part of the early to middle Silurian Hovells Suite of the Wyangala Supersuite. The Ballyhooley Granite, which crops out 3 km east of the Phoenix mine (Figure 2), is a variably foliated, coarse-grained, equigranular S-type muscovite–biotite granite to biotite granite that is weakly reduced and moderately to strongly fractionated (SiO$_2$ > 73%, very high Rb/Sr ratio and very low Ba — Thomas & Pogson in prep.). U–Pb SHRIMP dating of zircons has provided an age of 425 ± 3 Ma for the Ballyhooley Granite (Lennox & Zwingmann 2007). In addition, a poorly exposed deformed and altered muscovite-rich granite (Gibbons 1960; Wood 2001) crops out 300 m to 1 km northeast of Phoenix mine. Based on geochemical data, Wood (2001) suggested that the granite, which is enriched in MgO and Al$_2$O$_3$, is a marginal phase of the Ballyhooley Granite.

**Sample preparation and laboratory studies**

Samples for ore microscopy, isotope studies and age dating were collected. Samples were from diamond drillcore (stored at the Industry & Investment NSW WB Clarke Geoscience Centre at Londonderry), outcrop and from mine dumps at the Phoenix mine. Initially, samples were described in hand specimen and using a binocular microscope. Selected samples were prepared as polished and standard thin sections for ore microscopy and petrology.

Twelve sulfide-rich powders were analysed for their sulfur isotope composition. The powders were obtained using a microdrill and a binocular microscope. Contamination of mineral separates by other sulfide species was minimised by selecting coarse-grained material. Prior to isotopic analysis, the vacuum line at the University of Newcastle was used to convert sulfide samples to sulfur dioxide gas (at 950°C, with cuprous oxide (Cu$_2$O) as the oxidant) following the procedure of Robinson and Kusakabe (1975). Isotopic analysis was undertaken on a Finnigan 252 gas source isotope mass spectrometer at the CSIRO Centre for Isotope Studies, North Ryde. Data were reported to an accuracy of ± 0.2‰, relative to Cañon Diablo Troilite (CDT) sulfide and a variety of secondary standards.
Two samples were analysed for their lead-isotope composition at the CSIRO Exploration and Mining laboratories as part of the present study with repeated analyses being carried out on one sample. Carr et al. (1995) described the preparation of samples for lead-isotope analyses at CSIRO Exploration and Mining. After crushing, whole-rock samples were digested using a mixed 7N nitric + 7N hydrochloric acid solution prior to ion exchange. Lead was further purified by microelectrode deposition onto Pt electrodes. Samples were then analysed using a VG ISOMASS 54E solid source thermal ionisation mass spectrometer run in fully automatic mode. Ratios were normalised to accepted values of international standard NBS SRM 981. Precision estimates representing two standard deviations are 0.05 percent for the $^{207}$Pb/$^{206}$Pb ratio and 0.1 percent for all other ratios (Gulson et al. 1998). Analytical precision was based on over 1400 analyses (Carr et al. 1995).

In preparation for $^{40}$Ar/$^{39}$Ar dating, biotite grains were separated from sample DP4N–1A, collected from a quartz–biotite–scheelite–sulfide filled fracture at the Phoenix mine. Standard crushing, sieving, de-sliming and magnetic separation methods were utilised. Hand-picked single biotite grains were then washed in deionised water and acetone. Prior to irradiation, the grains were wrapped in aluminium packets and placed into aluminium irradiation canisters with aliquots of the flux monitor GA1550 (Age = 98.8 ± 0.5 Ma; Renne et al. 1998). The samples were irradiated in position 5C at the McMaster reactor, Hamilton, Ontario, Canada. Following irradiation grains were removed from their packaging and loaded into a copper sample holder. $^{40}$Ar/$^{39}$Ar analyses were undertaken at the University of Melbourne, using procedures described previously by Phillips and Miller (2006). Aliquots of 2–4 biotite grains were step-heated by a Spectron CW Nd:YAG laser. Argon isotopes were analysed on a MM5400 mass spectrometer, equipped with a Daly detector. Mass discrimination was monitored by analyses of standard air volumes. Interference correction factors were: $(^{36}\text{Ar}^{37}\text{Ar})_{\text{Ca}} = 2.89 (\pm 0.05) \times 10^{-4}$; $(^{39}\text{Ar}^{37}\text{Ar})_{\text{Ca}} = 6.80 (\pm 0.20) \times 10^{-4}$; $(^{39}\text{Ar}^{39}\text{Ar})_{\text{K}} = 0.0004 (\pm 0.0004)$. Ca/K ratios were calculated from the following relation: Ca/K = 1.9 $\times$ $^{39}\text{Ar}^{37}\text{Ar}$. The reported isotopic data have been corrected for system backgrounds, mass discrimination, isotopic interferences, fluence gradients and atmospheric contamination. Unless otherwise stated, errors associated with the age determinations are one sigma uncertainties and exclude uncertainties in the J-value, age of the fluence monitor GA1550 and the decay constants. The decay constants are those of Steiger and Jager (1977). The $^{40}\text{Ar}^{39}\text{Ar}$ dating technique was described in detail by McDougall and Harrison (1999).

Mine geology and mineralogy

A number of mineralised zones are present at Phoenix. These include the Main vein system (Main vein, Back vein, North East vein and Eastern lode) and the Western vein. The Main vein system is hosted by cleaved siltstones to very fine-grained quartz-rich sandstones of the Abercrombie Formation whereas the Western vein is hosted by quartz–lithic sandstones of the Hawkins Volcanics (Figure 3). These mineralised structures trend approximately 330° and dip easterly at 50° to 60° (Mulholland 1943).

Mineralisation at Phoenix consists of discontinuous masses of wolframite and subhedral scheelite largely within quartz and quartz–muscovite–tourmaline veins that both Mulholland (1943) and Gibbons (1960) noted as exhibiting boudinage structures. Minor pyrite, pyrrhotite, chalcopyrite and trace arsenopyrite, bismuthinite, bornite and native bismuth are also present. In addition, Hart and Dunkin (1940) recorded trace galena and sphalerite, while Gibbons (1960) noted minor magnetite and enargite. Alteration and gangue minerals associated with mineralised veins include quartz, tourmaline and muscovite, together with minor biotite, chlorite and late calcite. In thin section, quartz grains in quartz–scheelite veins display well-developed undulose extinction with grains often being slightly elongated in a variety of directions — indicating that these veins have been deformed (probably post-mineralisation).

Results of the present study support Gibbons’ (1960) observation that both wolframite and scheelite were early in the paragenesis — and that scheelite replaced wolframite. Gibbons (1960) proposed that the initial ore assemblage was pyrite, pyrrhotite, magnetite and wolframite. Scheelite then replaced both wolframite and pyrite. This was followed by chalcopyrite, a second-generation wolframite with an unidentified grey mineral. The final stages of ore deposition included enargite followed by late-stage veins, crosscutting earlier vein assemblages, with trace disseminated native bismuth with bismuthinite rims (Gibbons 1960).

Further observations made as part of the present study include:

- an early quartz–tourmaline with or without fine-grained white mica assemblage predated the initial wolframite stage of Gibbons (1960)
- a quartz–scheelite–muscovite–biotite assemblage (Photograph 1)
- fractures filled with pyrite, bismuthinite and chalcopyrite crosscut earlier mineralised structures, including quartz–scheelite veins
- a chlorite–biotite–calcite–apatite assemblage that is also associated with chalcopyrite
• pyrite, native bismuth and bismuthinite were deposited coeval with chalcopyrite
• chalcopyrite is intergrown with sphalerite
• enargite, which was recorded by Gibbons (1960), was not observed.

The present study indicates that the paragenetic sequence differs from that of Gibbons (1960) and consists of four stages. \textbf{Stage 1} includes the emplacement of quartz–tourmaline with or without muscovite and/or fine-grained white mica. \textbf{Stage 2} is the deposition of pyrrhotite–magnetite–wolframite with the probable inclusion of arsenopyrite. \textbf{Stage 3} includes scheelite deposition as a quartz–scheelite–muscovite–biotite assemblage in places replacing earlier wolframite. \textbf{Stage 4} is dominated by a chalcopyrite–pyrite–native bismuth–bismuthinite–sphalerite assemblage with minor fine-grained white mica, chlorite, biotite, calcite and apatite. The presence of bismuth minerals in Stage 4 suggests that ore-forming fluids cooled over time.

The presence of pyrrhotite, wolframite and minor arsenopyrite indicates that the initial ore-forming fluids were moderately reduced. Furthermore, the presence of muscovite/white mica, minor magnetite and late calcite suggests that the ore-forming fluids were near-neutral pH to slightly alkaline. Enargite, as noted by Gibbons (1960), is characteristic of acidic and strongly oxidised mineral systems — such as high sulfidation epithermal deposits. It is suggested that enargite is unlikely to be present at Phoenix and that Gibbons (1960) may have mis-identified this mineral.

\textbf{Isotopic data and comments}

\textbf{Sulfur isotopes}

Sulfur isotope values for sulfides from Phoenix lie in a narrow range between 3.3‰ and 5.8‰ (mean 4.6‰ — 12 analyses) with values for chalcopyrite ranging between 3.3‰ and 5.0‰ (mean 4.5‰ — 10 analyses). Single analyses of pyrite and pyrrhotite yielded values of 5.8‰ and 3.9‰, respectively. Table 1 and Figure 4 summarise the sulfur-isotope results.

The mineralisation at Phoenix contains pyrrhotite and arsenopyrite in addition to pyrite and base-metal sulfides, suggesting that the initial (Stage 2) ore-forming fluids were reduced. For reduced fluids, the isotopic composition of sulfides should correspond closely to the isotopic composition of total dissolved sulfur in the mineralising fluids ($\delta^{34}$S$_{\text{fluid}}$ — Ohmoto & Goldhaber 1997). Thus, the S-isotope composition of pyrrhotite from Stage 2 should reflect the sulfur isotope composition of the sulfur source. Because sulfur isotope values for sulfides from Stage 4 overlap with the $\delta^{34}$S value for Stage 2 pyrrhotite, it is suggested that there was little change in the $\delta^{34}$S composition of the ore-forming fluids between these two stages. Values as low as 3.3 $\delta^{34}$S‰ are close to the range of values for normal igneous sulfur (0 ± 2 $\delta^{34}$S‰ — Ohmoto & Rye 1979). This suggests that the majority of sulfur in the Phoenix deposit was derived from a magmatic source, either as a direct magmatic contribution accompanying felsic magmatism or, indirectly, through dissolution and recycling of rock sulfide from an older magmatic reservoir. Values as high as 5.8 $\delta^{34}$S‰ suggest that minor sulfur was also derived from a reduced seawater sulfate reservoir — such as the host metasedimentary sequence.

\textbf{Lead isotopes}

Two low-lead samples from the Phoenix mine were analysed for their lead isotope composition as part of the present study (Table 2). The analysis for sample DUP8 has a value of high quality for the data (data quality Q = 1 — Table 2), plots adjacent to the crustal growth curve and within the Devonian granite field of Carr et al. (1995) (Figure 5). Although containing low levels of lead (78 ppm Pb) this analysis is interpreted to be an initial ratio — i.e. the ratio has not been modified by the addition of radiogenic lead (Downes 2009 provided additional data). This analysis plots adjacent to and to the right of the 400 Ma crust–mantle isochron of Carr et al. (1995) supporting the interpretation of an Early to Middle Devonian lead model age (lead model age of 382 ± 11 Ma using PbGraph — Cumming & Richards (1975) model). Three lead isotope composition analyses were carried out on a second low lead (9 ppm Pb) sample (DP3N-1, Table 2). These analyses support the interpretation of a
<table>
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<tr>
<th>Sample No.</th>
<th>Location</th>
<th>mGAE</th>
<th>mGAN</th>
<th>Zone</th>
<th>Description</th>
<th>Mineral</th>
<th>δ$^{34}$S ‰</th>
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<td>DD102A</td>
<td>DP3N-1 — 101.98 m–102.10 m</td>
<td>672094</td>
<td>6208655</td>
<td>55</td>
<td>Early pyrite and dark chlorite</td>
<td>pyrite</td>
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<tr>
<td>DD102B</td>
<td>DP3N-1 — 101.98 m–102.10 m</td>
<td>672094</td>
<td>6208655</td>
<td>55</td>
<td>Early Cu-mineralised quartz vein folded due to slippage along cleavage</td>
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<td>DP159A</td>
<td>DP3N-1 — 159.34 m–159.44 m</td>
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<td>6208655</td>
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<td>Quartz vein with disseminated chalcopyrite</td>
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<td>DP3N-1 — 159.34 m–159.44 m</td>
<td>672094</td>
<td>6208655</td>
<td>55</td>
<td>Quartz vein with disseminated chalcopyrite</td>
<td>chalcopyrite</td>
<td>4.6</td>
</tr>
<tr>
<td>DP165</td>
<td>DP3N-1 — 165.38 m–165.42 m (main lode)</td>
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<td>6208655</td>
<td>55</td>
<td>Metavolcanic rock with chalcopyrite</td>
<td>chalcopyrite</td>
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<td>DUP1</td>
<td>dump</td>
<td>671978</td>
<td>6208511</td>
<td>55</td>
<td>Minor sulfides (inc. chalcopyrite) in metasiltstone/slate adjacent to a mineralised quartz vein (see DUP2)</td>
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<td>55</td>
<td>Minor disseminated chalcopyrite in a quartz vein, minor sulfides also present in the adjacent wall rock (see DUP1)</td>
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<td>dump</td>
<td>671978</td>
<td>6208511</td>
<td>55</td>
<td>Altered phyllite with weathered tourmaline, quartz veins and minor sulfides</td>
<td>chalcopyrite</td>
<td>4.3</td>
</tr>
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<td>DUP4</td>
<td>dump</td>
<td>671978</td>
<td>6208511</td>
<td>55</td>
<td>Disseminated chalcopyrite and wolframite in a quartz vein — vein may have been part of a larger vein or fracture filling</td>
<td>chalcopyrite</td>
<td>4.3</td>
</tr>
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<td>dump</td>
<td>671978</td>
<td>6208511</td>
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<td>Chalcopyrite in quartz vein material near the contact with metasediments — slate</td>
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<td>4.8</td>
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<td>Pyrrhotite associated with a quartz vein in hornfels</td>
<td>pyrrhotite</td>
<td>3.9</td>
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</table>

**Notes.** The grid used is GDA 94 MGA Zone 55

**Figure 4.** Frequency histogram of sulfur isotope values ($\delta^{34}$S) for the Phoenix tungsten deposit
Figure 5. Distribution of $^{207}$Pb/$^{204}$Pb and $^{208}$Pb/$^{204}$Pb data for the Phoenix deposit. Data for DUP8 from Phoenix mine are compared to the signature of Ordovician to Carboniferous metallogenic events and Lachlan Orogen crustal and mantle growth curves after Carr et al. (1995). The ellipse in the upper left-hand corner of the ratio plot is the analytical precision at 95% confidence.

(a) Plot of $^{208}$Pb/$^{204}$Pb vs $^{206}$Pb/$^{204}$Pb data for the area where $^{206}$Pb/$^{204}$Pb lies between 18.00 and 18.40 with the Lachlan Orogen crustal and mantle growth curves after Carr et al. (1995). (b) Plot of $^{207}$Pb/$^{204}$Pb vs $^{206}$Pb/$^{204}$Pb data for the area where $^{206}$Pb/$^{204}$Pb lies between 18.00 and 18.40 with the Lachlan Orogen crustal and mantle growth curves after Carr et al. (1995).

Table 2. Lead isotope data for sulfides from the Phoenix tungsten deposit

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<th>Sample No.</th>
<th>Location</th>
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<th>mGAN</th>
<th>Zone</th>
<th>Sample description</th>
<th>Sample Type</th>
<th>$^{206}$Pb/$^{206}$Pb ratio</th>
<th>$^{207}$Pb/$^{206}$Pb ratio</th>
<th>$^{206}$Pb/$^{204}$Pb ratio</th>
<th>Pb (ppm) or mineral</th>
<th>Q Factor</th>
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<tr>
<td>DUP8</td>
<td>DUP8</td>
<td>671978</td>
<td>6208511</td>
<td>55</td>
<td>Vein quartz with chalcopyrite on fracture plane</td>
<td>Whole rock</td>
<td>2.1050</td>
<td>0.8606</td>
<td>18.146</td>
<td>78</td>
<td>1</td>
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<tr>
<td>DP3N-1</td>
<td>DP3N-1</td>
<td>672094</td>
<td>6208655</td>
<td>55</td>
<td>Altered metavolcanic rock with quartz veins, disseminated chalcopyrite adjacent to vein and within the vein</td>
<td>Whole rock</td>
<td>2.0908</td>
<td>0.8236</td>
<td>19.07</td>
<td>9</td>
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<td>DP3N-1 (a)</td>
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<td>Altered metavolcanic rock with quartz veins, disseminated chalcopyrite adjacent to vein and within the vein</td>
<td>Whole rock</td>
<td>2.0896</td>
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<td>DP3N-1</td>
<td>672094</td>
<td>6208655</td>
<td>55</td>
<td>Altered metavolcanic rock with quartz veins, disseminated chalcopyrite adjacent to vein and within the vein</td>
<td>Whole rock</td>
<td>2.0907</td>
<td>0.8236</td>
<td>19.07</td>
<td>9</td>
<td>4</td>
</tr>
</tbody>
</table>

Notes:
- Q Factor is a measure of the quality of the analysis by the CSIRO Pb-isotope laboratory with Q factor = 1 being high whereas Q factor > 9 is very poor.
- (a) = repeat analysis.
crustal signature. However, they are not shown on Figure 5 as the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are greater than 18.5, which lie outside the range of values associated with initial Pb ratios for the Lachlan Orogen and indicate that these analyses have been modified by the addition of radiogenic lead.

The Pb-isotope data for Phoenix support the interpretation that lead was derived from a low-lead crustal source, such as the poorly exposed granite adjacent to the Phoenix area. Further, little or no lead was sourced from either the host Adaminaby Group or the Hawkins Volcanics because a less-evolved signature should be evident. Little or no lead was sourced from the Ordovician Kenyu Formation as a mantle-derived lead isotope signature should also be evident.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating

$^{40}\text{Ar}/^{39}\text{Ar}$ analysis (Table 3) was carried out on single biotite grains (11 samples) from the alteration selvage of a quartz–biotite–scheelite–sulfide-filled fracture intersected in diamond drillhole DP4N–1A (MGA 671958E 6208779N, zone 55) drilled at the Phoenix mine. The sample (DP4N–1A 133.5 m) is from 133.5 m in DP4N–1A.

Host rock to the mineralised vein (sample DP4N–1A 133.5 m) is a cleaved siltstone to very fine-grained quartz-rich sandstone (quartz dominant with ~1% albite-twinned feldspar grains and 10–15% clay matrix) with minor 1–2 mm thick mudstone beds. The cleavage is defined by a framework of aligned, fine-grained white mica and/or biotite and slightly elongated quartz grains within sandstone and as a penetrative fabric within the mudstone units. Cross-cutting this cleavage, at a low angle, is a 25 mm-wide Stage 3 quartz–biotite–scheelite–sulfide vein that is surrounded by a 2–4 mm wide alteration selvage containing biotite (Photographs 2, 3). Based on the presence of a pre-existing cleavage as well as the absence of significant feldspar grains and quartz crystal fragments, it is suggested that the host unit is part of the Abercrombie Formation rather than being part of the Hawkins Volcanics because rocks of the latter formation are characterised by the abundant feldspar and quartz crystal fragments.

Large biotite-rich zones are associated with the scheelite-bearing quartz vein (Photograph 3). Here, biotite occurs as elongate, gently curved zones, up to 0.3 mm long and 0.1 mm wide, within the quartz–scheelite vein and as 0.2 mm long by 0.02 mm wide zones within the adjacent alteration selvage. Also present are thin (0.05 mm wide) gently curved, discontinuous opaque zones that are formed within or on the margins of biotite-rich stringers and the alteration selvage. These are interpreted to be late-stage deformation-related stylolites that formed as a result of removal of quartz by pressure solution resulting in the concentration of immobile iron oxides and hydroxides forming stringers. Minor muscovite is present in some stylolites.

Quartz grains within the vein also display well-developed undulose extinction patterns and deformation bands supporting the interpretation that vein formation predated deformation. Larger quartz grains, invariably consist of subgrains and new grains, indicating post-formation deformation.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Step</th>
<th>Cum.% Ar* (x 10^-16 moles)</th>
<th>± Ar* (x 10^-16 moles)</th>
<th>± Ca/K (Ma)</th>
<th>± % 40Ar*/39Ar</th>
<th>± % 40Ar*/39Ar</th>
<th>± Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite#1</td>
<td>2</td>
<td>0.8153</td>
<td>0.0013</td>
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<td>0.0124</td>
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<td>Biotite#3</td>
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<td>0.0120</td>
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<td>0.022</td>
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<tr>
<td>Biotite#4</td>
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<td>0.0119</td>
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<td>0.002</td>
<td>0.022</td>
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<tr>
<td>Biotite#5</td>
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<tr>
<td>Biotite#6</td>
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<td>0.0110</td>
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<td>Biotite#7</td>
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<tr>
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<td>0.0003</td>
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<tr>
<td>Biotite#9</td>
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<td>0.0003</td>
<td>0.0101</td>
<td>0.002</td>
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<td>0.022</td>
</tr>
<tr>
<td>Biotite#10</td>
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<td>0.0003</td>
<td>0.0101</td>
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<tr>
<td>Biotite#11</td>
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<td>0.0003</td>
<td>0.0101</td>
<td>0.002</td>
<td>0.002</td>
<td>0.022</td>
</tr>
</tbody>
</table>

Notes:
1) Errors are one-sigma uncertainties and exclude uncertainties in the J-value.
2) Data are corrected for system backgrounds, mass discrimination, isotopic interferences, fluence gradients and atmospheric contamination.
3) Interference corrections: (40Ar*/39Ar)\(\text{Ca/K}\) = 2.89 (±0.20) \times 10^6; (36Ar*/37Ar)\(\text{Ca/K}\) = 0.004 (±0.0004).
4) J-value correction: 0.007039 Ma for GA-1550 biotite.
5) Biotite#8 step 1 and biotite#11 step 1 were not recorded as the Daly multiplier hardware trip was triggered due to excess argon release.
6) The decay constants are those of Steiger and Jäger (1977).
7) The dating technique was described in detail by McDougall and Harrison (1999).
Eleven biotite aliquots (2 to 4 grains each) from sample DP4N–1A 133.5 m were separately analysed, with individual grains being step-heated in two or three increments. The analyses yielded a range of apparent ages, from 394.8 ± 1.3 Ma to 408.9 ± 43.8 Ma (Table 3). The weighted mean age for all steps is 401.7 ± 1.5 Ma (21 analyses — 2σ; MSWD = 7.3) (Figure 6, MSWD = mean square weighted deviate). However, this MSWD value is significantly greater than 2.5 and thus is not statistically robust (cf. Baksi 2006). The mean age obtained for selected low temperature (six rejected) and all intermediate and high temperature steps is 403.2 ± 1.1 Ma (15 analyses — 2σ; MSWD = 2.7) (Figure 7). This age (403.2 ± 1.1 Ma) is Early Devonian and is interpreted to represent the time of quartz–biotite–scheelite–sulfide-vein formation and tungsten mineralisation at Phoenix.

Discussion on mineralisation and paragenesis

The 40Ar/39Ar dating of biotite grains from mineralised veins at Phoenix suggest the mineralisation formed at 403.2 ± 1.1 Ma. That date is supported by the available initial ratio lead isotope data which has a crustal signature and an interpreted Early to Middle Devonian lead model age. The timing of mineralisation at Phoenix predates the Middle Devonian Tabberabberan Orogeny, which Glen (2005) suggested was Eifelian in age (397.5 ± 2.7 Ma to 391.8 ± 2.7 Ma using the timescale of Gradstein et al. 2004) for the Eastern Subprovince of the Lachlan Orogen. Thus, although the deposit is within a major structure (Frogmore Fault Zone), it is interpreted to be granite-related. A granite-related origin for the deposit is supported by the presence of early tourmaline in veins and as alteration selvages; a mineral assemblage that includes wolframite, scheelite, bismuth minerals and apatite; a near-zero sulfur isotope signature suggesting that sulfur was sourced from a magmatic reservoir; and the presence of hornfels adjacent to the main vein system (Figure 3).

Gibbons (1960) suggested that the Ballyhooley Granite was the source of the mineralising fluid at Phoenix. However, U–Pb SHRIMP dating of zircons from the Ballyhooley Granite indicates that this pluton is significantly older — having formed at 425 ± 3 Ma (Lennox & Zwingmann 2007). In addition, Wood (2001) noted that the Ballyhooley Granite is not unusually fractionated when compared to other granites forming part of the Wyangala Supersuite, supporting the interpretation that the Ballyhooley Granite was not the source of the mineralising fluids. Wood (2001) observed that fractionated felsic dykes intruded the Ballyhooley Granite and that a second, poorly exposed, deformed muscovite-rich granite outcropped near the mine. Although he suggested that this muscovite-rich granite may be related to the Ballyhooley Granite (as mapped by the Geological Survey of NSW), it seems more likely that this intrusion is significantly younger and is the source of mineralising fluids.

The development of intrusion-related mineralisation at Phoenix in the Early Devonian extends the spatial extent of intrusion-related mineral systems of similar

**Figure 6.** 40Ar/39Ar age data for all heating steps for single biotite grains from 133.5 m depth within drillhole DP4N–1A. Blue boxes indicate low-temperature steps; black boxes show high-temperature (fusion) steps. Error boxes represent two sigma uncertainties. The green line is the mean value for all analyses.

**Figure 7.** 40Ar/39Ar age data for selected low temperature and all high temperature steps for single biotite grains from 133.5 m depth from drillhole DP4N–1A. Blue boxes indicate low-temperature steps; black boxes indicate fusion steps. Error boxes represent two sigma uncertainties. The green line is the mean value for selected grains.
age within the Eastern Subprovince of the Lachlan Orogen. Phoenix is only slightly younger than the Dargues Reef (Figure 1) intrusion-related gold deposit, 11 km south-southwest of Braidwood; of similar age to the Tallawang (Figure 1) magnetite skarn, 42 km north of Mudgee; and only slightly older than the copper–molybdenite mineralisation associated with the Yeoval Complex (Figure 1) 75 km north-northwest of Orange. The timing of mineralisation at those three localities is summarised below.

- McQueen and Perkins (1995) dated alteration-related sericite from Dargues Reef at 411 ± 5 Ma and 406 ± 4 Ma (40K–40Ar dating). The timing of mineralisation at Dargues Reef is supported by U–Pb SHRIMP dating of zircons from the host Braidwood Granodiorite that have crystallisation ages of 410.2 ± 3.1 Ma and 410.8 ± 3.2 Ma (Bodorkos & Simpson 2008).

- The Tallawang magnetite skarn, in the Dungeree Volcanics at the western contact of the Tallawang Granite, has been described by Downes (1999). Barron et al. (1999) suggested that the Tallawang Granite was contemporaneous with other monzodiorite and diorite sills that intrude the Hill End Thrust. These intrusions have been dated at 396 ± 6 Ma and 404 ± 7 Ma (U–Pb SHRIMP dating of zircons from a quartz monzodiorite and from a diorite respectively — Fanning 1997). The Tallawang Granite itself intrudes a rhyolite that Fanning (1997) dated at 415 ± 11 Ma (U–Pb SHRIMP dating of zircons).

- Mineralisation associated with the Yeoval Complex has been described by Downes (1999). This complex is part of the Boggy Plain Supersuite of Wyborn et al. (1987). U–Pb SHRIMP dating of zircons from unnamed phases of the Yeoval Complex gave a range of ages — including 397 ± 4 Ma, 398 ± 4 Ma and 399 ± 4 Ma (Black 1998, cited in Wyborn et al. 1999).

The Phoenix mineralisation is significantly younger than a number of other mineralised zones associated with plutons of the Wyangala Batholith. Those older deposits include the large Browns Creek gold–copper skarn (28 km south-southeast of Orange (Figure 1) — adjacent to the Carcoar Granodiorite); the Dryburgh greisen-type tungsten mineralisation (15 km northeast of the Phoenix mine (Figure 1) — possible source is the Reids Flat Granite) and the Rye Park tungsten–tin–molybdenum skarn (28 km south of the Phoenix mine — source is the Rye Park Granite).

Kovacs (2000) dated the Carcoar Granodiorite at 430.4 ± 4.7 Ma and the ‘Mine Dyke’ group, within the Browns Creek mine area, at 430.0 ± 5.4 Ma and 432.3 ± 4.9 Ma (U–Th–Pb dating of zircons). Based on that dating, Kovacs (2000) suggested that Browns Creek formed at 431 ± 3 Ma. Additional dating by Lennox et al. (2005) indicated that the Carcoar Granite may be slightly older (434.4 ± 5.5 Ma — U–Pb SHRIMP dating of zircons). However, all dates are within error and indicate that Browns Creek formed between 429 Ma and 435 Ma. Lennox and Zwingmann (2007) dated the Reids Flat Granite at 425 ± 4 Ma (U–Pb SHRIMP dating of zircons), while the timing of mineralisation at the Rye Park skarn is less well-constrained. The Rye Park Granite intrudes units of the Hawkins Volcanics that Thomas and Pogson (in prep.) suggest is of probable Wenlock age. The skarn itself has a lead model age of 427 ± 11 Ma (Downes 2009), supporting the interpretation that the Rye Park skarn is of probable middle Silurian age.

The identification of several intrusion-related mineralising events associated with plutons of the Wyangala Batholith extends the potential for additional intrusion-related mineralisation within the Eastern Subprovince of the Lachlan Orogen.

Conclusions
The 403.2 ± 1.1 Ma age for biotite in mineralised veins at the Phoenix tungsten deposit indicates that the impact of the early Devonian thermal and metallogenic event in the Eastern Subprovince of the Lachlan Orogen is more widespread than previously recognised. In addition, the identification of several metallogenic events associated with plutons of the Wyangala Batholith highlights the mineral potential of that terrane and suggests that further exploration is warranted for a range of commodities.

The ‘ore’ and alteration mineralogy, together with the sulfur isotope and lead isotope data, for Phoenix, are consistent with the deposit being formed by a granite-associated magmatic hydrothermal system. The causative pluton for these ore-forming fluids has yet to be identified. There is potential for further mineralisation associated with that pluton (and other plutons).

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We wish to thank Rad Flossman and Joanne Wilde (both from the University of New South Wales) for preparing thin sections and polished thin sections; Phil Secombe, Richard Bale (both from the University of Newcastle) and Anita Andrew (CSIRO) for the sulfur isotope analyses; and Graeme Carr (CSIRO) for the lead isotope analyses. Stan Szczepanski of the University of Melbourne is acknowledged for assistance with the 40Ar/39Ar analyses. Statistical calculations were carried out using the ISOPLOT software package of Ken Ludwig. Paul Meszaros is thanked for allowing access to confidential information. Thanks are also extended to Owen Thomas who reviewed an earlier draft of this paper, and to Kierran Maher (peer reviewer).
Quarterly notes

Future papers:

‘Review of Cambrian and Ordovician stratigraphy in New South Wales’ by I.G. Percival, C.D. Quinn & R.A. Glen

‘Early Permian fossils of the Dalwood Group near Paterson, New South Wales’ by N.S. Meakin, P.A. Flitcroft and L. Sherwin

Correction

QN134

References


Published papers 2010

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