Timing of Iron Oxide Cu-Au-(U) Hydrothermal Activity and Nd Isotope Constraints on Metal Sources in the Gawler Craton, South Australia

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Abstract

The eastern Gawler craton hosts Australia's premier uranium-bearing iron oxide copper-gold (IOCG) belt, the >500-km-long Olympic Cu-Au-(U) province. In addition to the Olympic Dam Cu-U-Au and Prominent Hill Cu-Au deposits, numerous barren and weakly mineralized IOCG prospects are present in the province. New geochronological data for hydrothermal minerals combined with constraints from host-rock ages demonstrate that alteration and associated IOCG mineralization formed between ~1570 and ~1600 Ma in three districts of the province. This IOCG hydrothermal activity temporally overlaps with magmatism of the Hiltaba Suite and Gawler Range Volcanics. Titanites in paragenetically early magnetite-bearing alteration in the Olympic Dam and Prominent Hill districts yield U-Pb ion probe ages of 1576 ± 5 and 1567 ± 10 Ma, respectively. Molybdenite in veins crosscutting magnetite-biotite and albitic alteration in the Moonta-Wallaroo district have Re-Os ages of 1574 ± 6 and 1599 ± 6 Ma, respectively. These represent minimum ages of the IOCG alteration assemblages in this district. A muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ age of 1575 ± 11 Ma provides a minimum age of paragenetically later hematitic alteration in the Olympic Dam district.

Neodymium isotope compositions were determined for 44 whole-rock samples from barren and weakly mineralized Cu-Au prospects and host rocks in the Olympic Dam and Prominent Hill districts. The new geochronological framework enables comparison of the Nd isotope data across two IOCG districts at the time of formation of the Olympic Dam deposit (ca. 1590 Ma). Magnetite-rich weakly Cu-mineralized alteration from five prospects yields a relatively narrow range of $\epsilon_{\text{Nd}(1590)}$ values of -5.8 to -4.1 . Both hematite- and magnetite-rich alteration yield generally similar $\epsilon_{\text{Nd}(1590)}$ values that match values from fresh and weakly altered Paleoproterozoic metasedimentary and metagranitic rocks (-6.6 to -3.5) as well as from most felsic Hiltaba Suite intrusions and Gawler Range Volcanics in the eastern Gawler craton (ca. -6 to -4). These data are consistent with crustal sources for REE and, by implication, for associated copper in the barren and weakly mineralized prospects. Mineralization and alteration in these minor IOCG systems can be geochemically discriminated from the giant Olympic Dam deposit, where greater inputs of mantle-derived REE and other ore components are evident.

Introduction

THE 7.7 BILLION METRIC TON (Gt) Olympic Dam Cu-U-Au-Ag deposit is the world's largest uranium deposit and one of the largest copper resources (BHP Billiton, 2007, www.bhp

billiton.com) and is one of the key deposits defining the iron oxide Cu-Au (IOCG) deposit class (Hitzman et al., 1992; Hitzman, 2000). Although this deposit has been the focus of considerable research, its origin and particularly the sources of ore metals, continues to be debated. Based on neodymium isotope data, Johnson and McCulloch (1995) proposed that

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mafic and/or ultramafic rocks or magmas with primitive Nd isotope signatures were important sources of rare earth elements (REE) and Cu in the deposit. Mantle-derived contributions of metals have been suggested for IOCG deposits elsewhere in the world (e.g., Skirrow, 1999; Mathur et al., 2002). However, in the Gawler craton it has not been clear whether such a contribution from mantle-derived rocks or magmas occurred only at the giant Olympic Dam deposit or whether barren and weakly mineralized IOCG hydrothermal systems also show a significant contribution of metals from mantle-derived sources.

Three districts of iron oxide-rich alteration and Cu-Au $\pm$ REE mineralization are recognized in the eastern Gawler craton (Fig. 1; from north to south): the Mount Woods inlier, which hosts the Prominent Hill IOCG deposit (Belperio et al., 2007), the Olympic Dam district, including the recently discovered Carrapateena deposit, and the Moonta-Wallaroo historic mining district. These districts form a >500-km-long IOCG metallogenic belt, termed the Olympic Cu-Au-(U) province (Skirrow et al., 2002, 2006). Previous work at the Olympic Dam deposit established its formation age of ~1590 Ma, broadly synchronous with regional magmatism of the

FIG. 1. Location and simplified geology of the Gawler craton (after Fairclough et al., 2003), showing tectonostratigraphic domains (modified after Parker, 1990), the Olympic Cu-Au province (Skirrow et al., 2002), and major Cu-Au and Au deposits and prospects. The three boxes correspond to Figures 2 to 4 and also outline the major IOCG districts.

Hiltaba Suite granitoids and Gawler Range Volcanics (Johnson and Cross, 1995). Few studies have examined the geology and geochronology of the numerous barren and weakly mineralized IOCG systems, in part due to the almost complete concealment of the province beneath sedimentary basins and regolith. The only previous direct dating of alteration, by Mortimer et al. (1988) at the Acropolis Cu(-Au) prospect (Fig. 2), determined a U-Pb age of 1604 ± 7 Ma for apatite by the TIMS method.

The geology, mineralogy, fluids, and geochemistry of a number of barren and weakly mineralized IOCG prospects within the Olympic Cu-Au-(U) province have been investigated as part of a broader study of the metallogeny of the Gawler craton (see companion papers by Bastrakov et al., 2007; Budd and Skirrow, 2007; Direen and Lyons, 2007; Fraser et al., 2007). In the first part of the present contribution, new geochronological results are presented for barren and weakly mineralized IOCG systems. The findings support the proposal of broadly contemporaneous early Mesoproterozoic IOCG hydrothermal activity along the length of the Olympic Cu-Au-(U) province.

In the second part of this paper, the hypothesis is tested that a mantle-derived contribution of metals is a distinguishing feature of major IOCG deposits in comparison with minor deposits in the Gawler craton. New Sm-Nd whole-rock isotope data are presented for several barren and weakly mineralized IOCG systems in the Olympic Dam district and Mount Woods inlier. The geochronological results provide the basis for interpretation of the sources of REE in these systems and for comparison with Olympic Dam. These and other data from companion studies are used to develop an integrated geologic-geochemical model for the formation of IOCG systems in the Olympic Dam district. We conclude that mantle-derived rocks and/or magmas were more important as sources of REE and ore metals at the Olympic Dam deposit than in barren and weakly mineralized IOCG prospects in the same district. This result has implications for explorers in discriminating major from minor IOCG-forming hydrothermal systems and sheds light on the lithospheric scale of the ore-forming system.

Regional Geologic Setting

The lithostratigraphy and evolution of the Gawler craton have been most recently described by Hand et al. (2007) and Ferris and Schwarz (2003), building on the work of Daly et al. (1998) and earlier studies. Only a summary is presented here, with emphasis on new findings in the eastern Gawler craton. Based on geochronology by Jagodzinski (2005), the Moonta-Wallaroo district and Olympic Dam district are now considered to share similar lithostratigraphy, although they probably represent deeper and shallower crustal levels of exposure, respectively (Skirrow et al., 2002).

The oldest dated units in the Olympic Cu-Au-(U) province are high-grade Archean gneisses in the western and north-western parts of the province, forming an elongate core to the Gawler craton. The Hutchison Group was deposited after ~2000 Ma and most likely before ~1845 Ma on a passive margin to the east of the Archean core (Hand et al., 2007).

Voluminous syntectonic granitoids of the Donington Suite were emplaced between 1850 ± 3.5 and 1860 ± 4 Ma in the

Olympic Dam district and southeastern Gawler craton (Figs. 1, 2A; Jagodzinski, 2005). Red K-feldspar-bearing arkosic metasedimentary rocks were largely derived from, and overlie, the Donington Suite in the Olympic Dam district (Jagodzinski, 2005).

The Wallaroo Group was deposited during the period ~1765 to 1740 Ma over much of the eastern margin of the Gawler craton and is an important host sequence for several of the IOCG alteration systems studied (Figs. 1–3; Jagodzinski, 2005; Hand et al., 2007). The arkoses noted above may form a lower unit, whereas the upper parts of the Wallaroo Group comprise thin-bedded feldspathic to biotite-rich or carbonaceous metasilts, metacarbonate rocks, felsic and mafic volcanic rocks, and minor banded iron formation (BIF).

In the Mount Woods inlier, the depositional ages and stratigraphic affinities of high-grade metasedimentary rocks (BIF, gneiss, schist) hosting some of the IOCG alteration systems in the Mount Woods inlier are poorly constrained and have been presumed to be correlatives of the Hutchison Group (Daly et al., 1998).

The Hiltaba Suite comprises granites, granodiorites, quartz monzonites, quartz syenite, and quartz monzodiorites (Drexel et al., 1993). It is widespread throughout the central and eastern Gawler craton including the Mount Woods inlier (Figs. 1–4), and members of the suite have zircon U-Pb ages ranging from ~1575 to ~1600 Ma (Fanning et al., 1988; Mortimer et al., 1988; Creaser and Cooper, 1993; Johnson and Cross, 1995; Holm, 2005; Jagodzinski, 2005; Hand et al., 2007). In places Hiltaba Suite plutons intrude the extensive Gawler Range Volcanics, a felsic-dominated bimodal sequence (Fig. 1; Drexel et al., 1993). The most precise age for the Gawler Range Volcanics is 1591 ± 3 Ma, from the central Gawler craton (Fanning et al., 1988). Mafic intrusions with ages between 1583 ± 7 and 1596 ± 4 Ma are present in each of the three IOCG districts of the Olympic Cu-Au-(U) province (Zang et al., 2002; Holm, 2005; Jagodzinski, 2005). The Hiltaba Suite and Gawler Range Volcanics are regarded as comagmatic, with both having compositions similar to A-type magmas (Giles, 1988; Creaser and White, 1991; Creaser, 1996). However, recent work has shown that the Hiltaba Suite and Gawler Range Volcanics include both I- and A-type compositions, with a predominance of A-type granites and volcanic rocks in the eastern and central Gawler craton, whereas I types are largely restricted to the central and western parts of the craton (Budd, 2006; Skirrow et al., 2006).

Paleoproterozoic basement and early Mesoproterozoic igneous rocks were unconformably overlain by the Mesoproterozoic Pandurra Formation in much of the eastern Gawler craton (Fanning et al., 1983; Fanning and Link, 2003). The Pandurra Formation was in turn overlain by Neoproterozoic and Cambrian sedimentary and volumetrically minor volcanic successions, deposited in what is known as the Stuart Shelf region.

The tectonothermal evolution of the eastern Gawler craton is not yet well understood. However, new and existing age constraints and recently acquired seismic data (Lyons et al., 2005) permit some generalized comments. For further details, see Hand et al. (2007) and Direen and Lyons (2007). Archean rocks were metamorphosed and deformed during the Sleafordian orogeny at ca. 2450 Ma (Daly et al., 1998). The Cornian (formerly Neill) orogeny at ~1850 Ma (Hand et

FIG. 2. Olympic Dam IOCG district. A. Interpreted basement geology (pre-Pandurra Formation) based on Direen and Lyons (2002), with locations of samples in this study. The position of the seismic reflection line (Lyons et al., 2005) past Olympic Dam is shown as a dashed line and represents the position of the schematic section in Figure 14. B. Distribution of anomalous iron oxide abundances, derived from inversion modeling of potential field data (Williams et al., 2004). Magnetite-rich zones generally correspond to magnetite-K-feldspar-calc-silicate alteration; hematite zones generally correspond to hematite-sericite-chlorite-carbonate alteration.

FIG. 3. Pre-Neoproterozoic basement geology and alteration in the Moonta-Wallaroo Cu-Au district, Yorke peninsula (simplified from Raymond, 2003). Locations of geochronology samples indicated.

al., 2007) was closely associated with emplacement of the Donington Suite granitoids in the eastern Gawler craton. This was followed by the Kimban orogeny at ~1700 to 1740 Ma (Daly et al., 1998; Hand et al., 2007). Deformation and low-to medium-grade metamorphism of the Wallaroo Group is assumed to have occurred during the Kimban orogeny (Conor, 1995; Zang et al., 2002), although there is no direct geochronological evidence for this view. An “anorogenic” setting has been widely cited for the Hiltaba Suite and Gawler Range Volcanics (e.g., Giles, 1988; Drexel et al., 1993). However, this has been challenged over the past decade, and syntectonic features of the Hiltaba Suite intrusions have been increasingly recognized (Creaser, 1996; Ferris et al., 2002; McLean and Betts, 2003; Lyons et al., 2005; Hand et al., 2007).

Post-Hiltaba Suite tectonism affected the northern Gawler craton at ~1540 to 1570 Ma (Daly et al., 1998), followed by shearing and thermal resetting of mica K-Ar and Ar-Ar ages in parts of the western and southeastern Gawler craton during the period ~1450 to 1500 Ma (Webb et al., 1986; Fraser and Lyons, 2006).

IOCG Mineralization and Alteration Setting

Two major IOCG deposits and numerous Cu-Au ± U prospects are known within the Olympic Cu-Au-(U) province.

Table 1 presents summary data on the Olympic Dam and Prominent Hill deposits and on the barren to weakly mineralized hydrothermal systems sampled in the present study. Based on comparison with the major IOCG deposits in conjunction with descriptions of IOCG deposits by Haynes (2000), Hitzman (2000), and Williams et al. (2005), all of the investigated prospects are considered to be related to IOCG-forming hydrothermal systems (see also Skirrow et al., 2002; Bastrakov et al., 2007, for more detailed descriptions of prospects). The key IOCG-related features of the prospects are as follows: (1) hydrothermal alteration is magnetite and/or hematite rich, with mineral assemblages typical of IOCG alteration elsewhere (see next section); although apatite is present as a minor phase in some Fe oxide alteration, large barren magnetite-apatite deposits of the Kiruna type (Hitzman et al., 1992; Hitzman, 2000; Williams et al., 2005) do not appear to be present in the Olympic Cu-Au-(U) province; (2) sulfides are predominantly Cu and Fe bearing, and accessory REE-bearing minerals are common; Au and U are elevated in some prospects, whereas Pb and Zn are generally very low in abundance; (3) alteration and mineralization occur as epigenetic replacements, veins, and breccias, commonly related to faults or shear zones.

Detailed published information on the recently discovered Carrapateena deposit (Fig. 1) is not yet available, but it too

FIG. 4. Pre-Neoproterozoic basement geology of part of the Mount Woods inlier (modified slightly and simplified from Betts et al., 2003), showing Cu-Au prospects, Prominent Hill Cu-Au deposit, and sample locations.

appears to share key features of the major IOCG deposits in the Gawler craton, such as zoned chalcopryrite and bornite mineralization hosted by hematitic breccias (Teck Cominco Ltd. Annual Report 2005, www.teckcominco.com).

Some of the geochronology samples in this study are from the Moonta-Wallaroo Cu-Au district, where ~330,000 t of Cu was mined in the 19th and 20th centuries at an average grade of

5 percent Cu and 1 to 4 g/t Au (Zang et al., 2002). Two main styles of Cu-Au mineralization are present (Table 1). Wallaroo-style mineralization typically occurs as disseminated or quartz vein-hosted chalcopryrite-pyrite $\pm$ pyrrhotite, within strongly foliated biotite-magnetite alteration zones hosted by metasedimentary rocks (Conor, 1995; Zang et al., 2002). In Moonta-style Cu-Au mineralization, chalcopryrite-pyrite-bornite and

TABLE 1. Characteristics of Selected Cu-Au-(U) Deposits and Prospects of the Olympic Cu-Au-(U) Province, Gawler Craton

Deposit/prospect (north to south)	Mineralization (and alteration types present)	Tonnage-grade or intersections	Host-rock geology	Alteration and sulfide mineral paragenetic stages (numbered)	Mineralization textures, style	Age	References
Prominent Hill	Major Cu-Au deposit (Hem-Ser- Chl-Carb)	119 Mt at 1.3% Cu, 0.49 g/t Au	Hem breccias hosted by metasediments, volcanics; mafic, andesitic, felsic dikes	1. Ccp, Bn, Cc, Hem, Ser, Qtz 2. Fl, Ur	Disseminated sulfides in breccia		Belperio (2002); Belperio et al. (2007)
Manxman AL, Jones Dam	Weakly Cu-Au mineralized (Mag- Alb-calc-silicate; Mag-Bt; hematitic)	Manxman: 287 m at 0.23% Cu; Jones Dam: 186 m at 0.13% Cu; short intervals up to 1.7% Cu, 2.5 g/t Au	Mag- and Cal-rich breccias of pelitic metaseds; I-type granitoid, felsic to intermediate gneisses, granulite	1. Alb, Sep, Cpx, Qtz, Act 2. Mag, Bt, Ap, Ttn, REE 3. Po, Ccp, Py, REE- U-Th minerals, Hem, Chl, Cal, Alb, Fl, Kfs, Stp, Brt, Ser	Sulfides as vein fill, disseminated and in breccia matrix; massive Mag and Hem zones		Hampton (1997)
Manxman west (DD86EN26)	Barren (Mag-Alb- calc-silicate)		Fs-Qtz-Bt gneiss; Hiltaba Suite granite dikes	Mag, Alb, Act, Ttn	Rare sulfides	Titanite U-Pb minimum age: 1567 ± 10 Ma; granite zircon U-Pb: 1586 ± 3 Ma	This study; Jagodzinski (2005)
Titan	Weakly Cu mineralized (Mag- Kfs-calc-silicate; Hem-Ser-Chl-Carb)	47 m at 0.27% Cu, 0.06 g/t Au; 5 m at 1.1% Cu, 0.25 g/t Au	Meta-arkose and thin bedded Wallaroo Group; mafic and felsic dikes	1. Bladed Hem 2. Mag, Qtz, Act, Tr, Py, Kfs, Ccp 3. Hem, Carb, Chl	Disseminated and vein sulfides		Tasman Resources (2004 Annual Report, www.tasman resources.com.au)
Olympic Dam	Major Cu-U-Au deposit (Hem-Ser- Chl-Carb)	3.81 Gt at 1.1% Cu, 0.5 g/t Au, 0.4 kg/t U ₃ O ₈	ODBCI hosted by Roxby Downs Granite; multistage hematitic breccias; volcanoclastics, felsic and mafic- ultramafic dikes	1. Mag ± Hem, Chl, Ser, Sd, Py, Ccp, Pt 2. Hem, Ser, Cc, Bn, Py, Brt, Fl, Chl 3. Hem, Qtz, Brt	Disseminated sulfides in hematitic breccia matrix and clasts; also sulfides in veins	~1590–1595 Ma	Reeve et al. (1990), Johnson and Cross (1995), Jagodzinski (2005); WMC Resources (2005)
Torrens	Weakly Cu mineralized (Mag- Kfs-calc-silicate; Hem-Ser-Chl-Carb)		Fine-grained feldspathic metasediments (Wallaroo Group?)	1. Mag, Qtz, Py, Kfs, Ap, Carb 2. Hem, Chl, Ms/Ser, Ccp		Muscovite ⁴⁰ Ar- ³⁹ Ar: 1575 ± 11 Ma	This study
Murdie	Barren (Mag-Alb- Kfs-calc-silicate; weak hematitic)		Fine-grained feldspathic metasediments (Wallaroo Group?)	1. Mag, Alb, Act, Ap, Ttn, Py, Ccp, Carb 2. Kfs 3. Hem, Chl, Cal		Titanite U-Pb: 1567 ± 10 Ma	Paterson and Muir (1986); this study
Oak Dam	Weakly Cu-U mineralized (Mag- bearing; Hem-Ser- Chl-Carb)	7 m at 0.27% Cu and 63 m at 0.32% Cu and 690 ppm U; ~ 560 Mt massive Fe; 216 m at 1.480 ppm La, 1.615 ppm Ce	Donington granitoids and metasediments; cut by Hiltaba Suite? granite dikes, minor diorites, and mafic dikes	1. Mag-Qtz-Ap-Py 2. Colloform Hem 3. Hem after Mag; chalcedony, Qtz-III, Chl, Ser, Muz, Ccp, Py, Sp, Ap, Carb, Ur; 4. Vuggy Qtz ± Brt.	Disseminated sulfides in hematitic breccia matrix and clasts; sulfides in veins; replacement of colloform bands		Davidson and Paterson (1993); Davidson et al. (2007)

TABLE 1. (Cont.)

Deposit/prospect (north to south)	Mineralization (and alteration types present)	Tonnage-grade or intersections	Host-rock geology	Alteration and sulfide mineral paragenetic stages (numbered)	Mineralization textures, style	Age	References
Emmie Bluff	Weakly Cu-Au mineralized (Mag- Kfs-calc-silicate; Hem-Ser-Chl-Carb)	Up to 2.8% Cu, 0.6 g/t Au	Wallaroo Group, Donington granite (north); Gawler Range Volcanics dacite (south)	1. Mag, Cpx, Act, Grt, Qtz, Cal, Kfs, Py, Aln (south) 2. Ccp, Bn, Hem, Chl, Qtz, Cc (north)	Disseminated sulfides in tectonic breccias		Gow (1994a)
Wallaroo	Small Cu-Au deposit (Mag-Bt)		Phyllite, biotite schist (Wallaroo Group)	Qtz, Carb, Tur, Ap, Fl, Amph, Cpx, Mag, Kfs, Bt, Py, Po, Ccp	Replacement and disseminations along shearing		Conor (1995); Raymond et al. (2002)
Moonta (including Wheel Hughes, Poona)	Cu-Au deposit (Qtz-rich, Fe oxide poor variant? of IOCG)	Approx 6.75 Mt at 5% Cu, 1-4 g/t Au (Moonta-Wallaroo combined)	Metarhyolite (Moonta Porphyry in Wallaroo Group)	Qtz, Kfs, Bt, Tur, Chl, Ser, Hem, Mag, Cpx, Py, Ccp, Bn	Fissure veins up to 8 m wide		Both et al. (1993); Conor (1995)
Alford (DDH130)	Weakly Cu mineralized (Mag-Bt; Qtz-rich variant? of Hem-Ser-Chl-Carb)	Up to 0.2% Cu, 3,600 ppm Mo	Feldspathic siltstones, Wallaroo Group	1. Mag, Bt, Alb, Act, Ttn, Tur 2. Qtz, Kfs, Chl, Hem, Mo	Shear-hosted alteration; Qtz- sulfide veinlets along foliation	Molybdenite Re-Os: 1574 ± 6 Ma	Zang et al. (2002); this study
Pridhams (DDH213)	Weakly Cu mineralized (Fe oxide-poor variants? of Mag-Alb-calc- silicate and Hem- Ser-Chl-Carb)	Up to 1% Cu	Calcareous metasediments, Wallaroo Group	Cal, Qtz, Bt, Chl, Ser, Ccp, Py, Mo	Sulfidic veins cutting Scp-bearing metasediments	Molybdenite Re-Os: 1599 ± 6 Ma	Zang et al. (2002); This study

¹ ODBC = Olympic Dam Breccia Complex (Reeve et al., 1990)

Mineral abbreviations (following Kretz, 1983): Act = actinolite, Alb = albite, Amph = amphibole, Ap = apatite, Bn = bornite, Brt = barite, Cal = calcite, Carb = carbonate, Cc = chalcocite, Ccp = chalcopyrite, Chl = chlorite, Cpx = clinopyroxene, Dol = dolomite, Ep = epidote, Fl = fluorite, Gn = galena, Grt = garnet, Hem = hematite, Ill = illite, Kfs = K-feldspar, Mag = magnetite, Mar = marcasite, Mnz = monazite, Mo = molybdenite, Ms = muscovite, Po = pyrrhotite, Pt = pitchblende, Py = pyrite, Qtz = quartz, Scp = scapolite, Sd = siderite, Ser = sericite, Sp = sphalerite, Stp = stibnomelane, Tr = tremolite, Ttn = titanite, Tur = tourmaline, Ur = uraninite

gold occur within veins up to 8 m wide, composed of quartz $\pm$ K-feldspar $\pm$ chlorite $\pm$ hematite $\pm$ sericite $\pm$ tourmaline $\pm$ kaolinite $\pm$ molybdenite (Both et al., 1993; Conor, 1995; Morales Ruano et al., 2002; Zang et al., 2002). The host rocks in the Moonta deposits are variably sheared felsic porphyry within the Wallaroo Group. Iron oxides are much less abundant than in the metasedimentary rock-hosted Wallaroo Cu-Au style. Moonta-style mineralization is observed to have overprinted Wallaroo-style mineralization in the rare cases where both developed at the same locations. The characteristics of Wallaroo-style Cu-Au mineralization are consistent with IOCG mineralization, whereas the quartz-rich Moonta-style may represent a different type of Cu-Au mineralization. Both et al. (1993) and Morales Ruano et al. (2005) suggested a magmatic-hydrothermal origin. Regardless, the relative and absolute timing relationships of Moonta-style mineralization provide useful constraints on the age of IOCG-related hydrothermal processes.

Alteration in the Olympic Cu-Au-(U) province

General characteristics of hydrothermal alteration in the Olympic Cu-Au-(U) province were described by Skirrow et al. (2002). A revised synopsis of these characteristics is given below. Despite most of the observations and sample data being limited to clusters of drill holes at widely separated prospects (Figs. 1–4), four major groupings of alteration assemblages are recognized in the metallogenic province with remarkably consistent mineralogy:

1. Vein and replacement alteration comprising magnetite-albite-calc-silicate has been recognized in all three districts of the Olympic Cu-Au-(U) province. Calcsilicate minerals are generally actinolite or clinopyroxene (diopside). Magnetite-albite-calc-silicate alteration assemblages in places contain minor to trace quantities of quartz, pyrite, apatite, titanite, chalcopryrite, pyrrhotite, scapolite and/or allanite. Rarely preserved coarse specular hematite forms a precursor to magnetite in some alteration systems (e.g., Titan).

2. K-feldspar-bearing magnetite-calc-silicate alteration is known mainly from the Olympic Dam district, where early albite alteration has been almost entirely replaced by K-feldspar (Bastrakov et al., 2007). Other minerals in the magnetite-K-feldspar-calc-silicate alteration are similar to those in magnetite-albite-calc-silicate assemblages.

3. Magnetite-biotite alteration, commonly with minor pyrite-chalcopryrite mineralization and albite, apatite, monazite, and tourmaline, is recognized in the Moonta-Wallaroo district and Mount Woods inlier but not in the Olympic Dam district. In the Manxman and Joes Dam prospects in the Mount Woods inlier, minor titanite, apatite, and REE-bearing phases accompanied the magnetite and biotite alteration (Hampton, 1997).

4. Hematite-sericite-chlorite-carbonate alteration ("hematitic alteration") is associated with pyrite, Cu-Fe sulfides (generally chalcopryrite, with minor bornite in some prospects), barite, fluorite, and U and REE minerals. Hydrothermal white mica is not as widespread as hematite, chlorite, or carbonate; it has been observed at the Emmie Bluff, Torrens, and Titan prospects (Davidson and Paterson, 1993; Gow et al., 1994a; Skirrow et al., 2002; Bastrakov et al., 2007; Davidson et al., 2007).

The magnetite-albite-calc-silicate assemblages are broadly similar to regional Na-Ca alteration in IOCG provinces globally (e.g., Hitzman et al., 1992; Barton and Johnson, 1996; Hitzman, 2000). The magnetite-biotite and magnetite-K-feldspar-calc-silicate assemblages resemble the potassic alteration described in global IOCG provinces, and the hematite-sericite-chlorite-carbonate alteration assemblage can be considered equivalent to the hydrolytic alteration of Hitzman et al. (1992). In the present paper, the alteration assemblages are denoted by their characteristic mineralogy, in preference to the more generalized terms such as potassic or hydrolytic alteration.

The four major alteration assemblages are developed in a wide range of host lithologic units, including metasedimentary rocks of the Wallaroo Group, granitoids of the Donington Suite and Hiltaba Suite, and the Gawler Range Volcanics. Overall, most host rocks are quartzofeldspathic. The dependence of alteration mineralogy on host-rock type noted by Barton and Johnson (1996) appears to be less evident in the Olympic Cu-Au-(U) province than in many other IOCG provinces. Where both hematite- and magnetite-bearing assemblages are observed together at prospects the hematite-sericite-chlorite-carbonate assemblage consistently overprints magnetite-feldspar-calc-silicate and magnetite-biotite assemblages. This relationship was recognized by Gow et al. (1994a) at the Emmie Bluff prospect; our work has confirmed and extended this observation to other parts of the Olympic Cu-Au-(U) province. However, at Olympic Dam, Haynes et al. (1995) described a more complex evolution involving multiple cycles of early magnetite $\pm$ hematite, chlorite, sericite, siderite, pyrite, chalcopryrite, and pitchblende, overprinted by hematite, sericite, chalcocite, bornite, pitchblende, barite, fluorite, and chlorite. In some cases hematitic alteration developed where there is no evidence for earlier magnetite-bearing alteration, for example, in the ore-hosting breccias at the Prominent Hill deposit (Belperio, 2002; Belperio et al., 2007). In general terms, however, hematite-sericite-chlorite-carbonate alteration assemblages in the barren and weakly mineralized prospects resemble Cu-Au $\pm$ U mineralized assemblages at the Olympic Dam and Prominent Hill deposits (Reeve et al., 1990; Belperio, 2002).

Relative timing relationships between the magnetite-feldspar-calc-silicate and magnetite-biotite assemblages are unclear because both assemblages rarely occur in the same locality. At the Manxman A1 prospect in the Mount Woods inlier (Fig. 4) the magnetite-biotite assemblage is interpreted to have overprinted the magnetite-albite-calc-silicate assemblage (Hampton, 1997). Regional-scale magnetite-rich alteration (including both magnetite-albite-calc-silicate and magnetite-biotite types) in the Moonta-Wallaroo district is shown in Figure 3. Magnetite-biotite assemblages developed in brittle-ductile regional shear zones, some of which crosscut Hiltaba granites. On the other hand, magnetite-albite-calc-silicate assemblages show a close spatial association with the contact zones of Hiltaba Suite granites as well as showing mutually crosscutting relationships with these granites (Conor, 1995; Raymond et al., 2002). These relationships are taken to indicate that magnetite-biotite alteration may have at least partly postdated magnetite-albite-calc-silicate alteration.

Geochronology of Alteration and Mineralization

To better constrain the timing of IOCG alteration and mineralization in the Olympic Cu-Au-(U) province, molybdenite was dated by the Re-Os method, hydrothermal titanite was dated by the U-Pb SHRIMP method, and biotite and hydrothermal white mica were dated by the ^{40}Ar - ^{39}Ar method.

Analytical methods

Re-Os analysis: Methods used for molybdenite analysis are described in detail by Selby and Creaser (2001, 2004). The ^{187}Re and ^{187}Os concentrations in molybdenite were determined at the University of Alberta by isotope dilution mass spectrometry using Carius-tube, solvent extraction, anion chromatography, and negative thermal ionization mass spectrometry techniques. Isotopic analysis was conducted using a Micromass Sector 54 mass spectrometer by Faraday collector. Total procedural blanks for Re and Os were less than 5 and 2 pg (~ 25 fg ^{187}Os), respectively. These procedural blanks are insignificant in comparison to the Re and Os concentrations in molybdenite. The Chinese molybdenite powder HLP-5 (Markey et al., 1998), which is used as an inhouse control sample by AIRIE, Colorado State University, is also routinely analyzed at the University of Alberta. For this control sample we determined an average Re-Os date of 220.5 ± 0.5 Ma ($\pm 0.45\%$ 2σ , $n = 17$). This Re-Os date is identical, within uncertainties, to that reported by Markey et al. (1998) of 221.0 ± 1 Ma ($\pm 0.4\%$ 1σ , $n = 19$) determined by alkaline fusion, and two Carius tube analyses of 219.8 ± 0.7 and 221.0 ± 0.8 Ma (H. Stein, pers. commun., 2000). The age uncertainties in the results reported in the present study are quoted at 2σ level and include all known analytical uncertainty but do not include a ~ 0.31 percent uncertainty in the decay constant of ^{187}Re .

U-Pb SHRIMP methods: Titanites were analyzed in situ using the SHRIMP RG at the Research School of Earth Sciences, Australian National University, in polished thin sections. Selected areas of polished thin sections were cast into epoxy disks together with chips of the BLR-1 reference titanite (provided by John Aleinikoff, USGS, Denver). Reflected and transmitted light photomicrographs and backscattered electron (BSE) images were prepared for all areas of titanite intended for analysis. These were used to target specific areas within the titanite grains, although in most cases the transmitted light images were a useful guide as to the less versus more altered nature of the titanite.

Each SHRIMP analysis consisted of seven scans through the mass range. The data have been reduced in a manner similar to that described by Williams (1998, and references therein), using the SQUID Excel Macro of Ludwig (2000). The Pb/U ratios for the titanite have been normalized relative to a value of 0.1769 for the $^{206}\text{Pb}/^{238}\text{U}$ ratio of the BLR-1 reference zircons, equivalent to an age of 1050 Ma (John Aleinikoff, pers. commun., USGS, Denver). Uncertainties given for individual analyses (ratios and ages) are at the 1σ level, however, the uncertainties in calculated weighted mean ages are reported with 95 percent confidence limits. The "Wetherill" concordia plots, relative probability plots with stacked histograms and weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age calculations, were calculated using ISOPLOT/EX (Ludwig, 2001).

^{40}Ar - ^{39}Ar dating methods: Mineral separates were prepared at Geoscience Australia by standard crushing, sieving, and heavy liquid methods, followed by handpicking of grains. Samples were irradiated together with the fluence monitor standard 77-600 hornblende for 24 d in the X33 or X34 position at the HIFAR facility, Lucas Heights, Australia. The irradiation canister included a 0.2-mm-thick Cd sleeve to shield material for thermal neutrons and was inverted three times during the irradiation period to minimize the effect of the large neutron fluence gradients at the HIFAR facility. Isotopic analyses of both fluence monitors and unknowns were carried out at the Research School of Earth Sciences, Australian National University, via heating in a resistance furnace attached to a VG-1200 mass spectrometer. Mass discrimination was monitored by repeat measurements of atmospheric argon and averaged ($0.99847 \pm 0.3\%$ /amu) over the course of the analyses. Analysis of K glass situated at each end of the irradiation can revealed no significant departure of the $(^{40}\text{Ar}/^{39}\text{Ar})_{\text{K}}$ correction factor from the established value for HIFAR, thus correction for K- and Ca-derived nucleogenic isotopes was made via the following correction factors: $(^{36}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} = 3.5 \times 10^{-4}$, $(^{39}\text{Ar}/^{37}\text{Ar})_{\text{Ca}} = 7.86 \times 10^{-4}$, and $(^{40}\text{Ar}/^{39}\text{Ar})_{\text{K}} = 2.7 \times 10^{-2}$, following Tetley et al. (1980). Data processing was via the inhouse Macintosh software Noble and did not include correction for any chlorine-related nucleogenic ^{36}Ar . Ages were calculated assuming the age of 77-600 hornblende as 418.3 Ma (Spell and McDougall, 2003). Plateau ages have been calculated using the statistical criteria of Isoplot (Ludwig, 2001). Uncertainties on plateau ages shown in Table 4, and in the figures and discussed in the text are quoted at the 2σ level and include uncertainties in the J parameter, estimated at 0.5 percent (1σ), and propagated by the method of Karner and Renne (1998). The uncertainties of the plateau ages do not include uncertainty in the age of 77-600 hornblende or decay constant uncertainties.

Results

Molybdenite Re-Os ages: Two samples of molybdenite were dated from the Moonta-Wallaroo Cu-Au district, from the Alford prospect (DDH130) and the Pridhams prospect (DDH213; Fig. 3, Table 2).

Drill hole DDH130 intersected part of a ~ 4 -km-long zone of shear-controlled magnetite-rich alteration along the southern contact of a composite Hiltaba Suite intrusion (Fig. 3). Magnetite-biotite-altered Wallaroo Group metasedimentary rocks are cut by veins of quartz, chlorite, K-feldspar, and minor hematite, chalcopyrite, pyrite, tourmaline, and molybdenite (Fig. 5A-C). Sample DDH130 241.7 contains one such quartz-pyrite vein with a few vol percent disseminated fine-grained molybdenite. Earlier biotite, albite, and magnetite in altered wall rock adjacent to the vein are replaced by chlorite, K-feldspar, and hematite, respectively. The earlier alteration is IOCG-related magnetite-biotite alteration, whereas the overprinting quartz-sulfide veins and wall-rock alteration are similar in mineralogy to those of Moonta-style mineralization or may represent a variant of hematite-sericite-chlorite-carbonate alteration. The molybdenite yields Re-Os ages of 1574 ± 6 and 1577 ± 6 Ma (the latter is a repeat analysis 6 months after the initial analysis). These dates are interpreted as minimum ages of IOCG-related magnetite-biotite alteration and

TABLE 2. Molybdenite Re-Os Dating Results for the Moonta-Wallaroo District

Drill hole, depth (m)	Host rock	Mineralogy, texture	Relationship to IOCG alteration	Sample (mg)	^{187}Os (ppb)	Re (ppm)	Age (Ma)	$\pm 2\sigma$
DDH130, 241.7	Wallaroo Group, fine-grained albite-quartz-biotite-K-feldspar	Quartz-pyrite vein with minor disseminated fine molybdenite (~130 mg concentrate)	Quartz-K-feldspar-chlorite-pyrite-chalcopyrite-molybdenite veins overprint magnetite-biotite and albite alteration	72	1,072	64.20	1574	6
DDH130, 241.7 repeat	Wallaroo Group, fine-grained albite-quartz-biotite-K-feldspar	Quartz-pyrite vein with minor disseminated fine molybdenite (~130 mg concentrate)	Quartz-K-feldspar-chlorite-pyrite-chalcopyrite-molybdenite veins overprint magnetite-biotite and albite alteration	56	1,080	66.44	1577	6
DDH213, 826.0	Wallaroo Group, bedded calcareous metasedimentary rocks	Calcite-pyrite vein with minor molybdenite, cutting assemblage of scapolite, biotite, carbonate, albite, actinolite, epidote-clinozoisite	Calcite-pyrite-chalcopyrite-chlorite veins with biotite alteration aureole are a variant of magnetite-biotite alteration	101	87.36	5.147	1599	6

represent the age of quartz-sulfide veining and chloritic alteration at the Alford prospect.

Drill hole DDH213 in the eastern part of the Moonta-Wallaroo district (Fig. 3) intersected altered feldspathic and carbonate-bearing metasedimentary rocks of the Wallaroo Group. Parts of the drill hole contain an early assemblage of albite, actinolite, epidote-clinozoisite, and recrystallized scapolite-bearing spots up to 3 mm in diameter (Figs. 5D, 6A-B). This assemblage, which may be an oxide-free variant of magnetite-albite-calc-silicate alteration, is cut by veins of calcite-quartz-chlorite-sericite-chalcopyrite-pyrite-molybdenite (Figs. 5D, 6A). Based on their mineralogy, these veins are interpreted to represent a variant of either hematite-sericite-chlorite-carbonate alteration or Moonta-style mineralization. Biotite-rich alteration aureoles adjacent to some of the veins overprint the scapolite spots but appear to predate the sulfidic veins. The lack of hematite and magnetite may reflect the relatively low iron content of the host metasedimentary strata. Minor molybdenite occurring in a calcite-pyrite vein at a depth of 826.0 m, similar to the calcite veins illustrated in Figure 5D, yields an Re-Os age of 1599 ± 6 Ma (Table 2). This date is interpreted to represent the age of hematite-sericite-chlorite-carbonate alteration or Moonta-style veining and, importantly, is also a minimum age of the albitic Na-Ca alteration at the Pridhams prospect.

In summary, the two molybdenite Re-Os ages indicate that IOCG-related magnetite-biotite alteration formed prior to 1574 ± 6 Ma in the Alford area, whereas Na-Ca alteration at Pridhams predated the hematite-sericite-chlorite-carbonate alteration or Moonta-style veins dated at 1599 ± 6 Ma.

Titanite SHRIMP U-Pb ages: Titanite was dated from the Murdie Murdie prospect (drill hole MRD1) in the eastern part of the Olympic Dam district (Fig. 2) and in the southern Mount Woods inlier, ~5 km northwest of the Manxman prospect (Fig. 4, drill hole DD86EN26). The results are summarized in Table 3, and analytical data are presented in Appendix Tables A1-A3.

At the Murdie Murdie prospect, two samples were obtained ~17 m apart in drill hole MRD1, where medium- to coarse-grained hydrothermal magnetite, K-feldspar, apatite, pyrite, carbonate, actinolite, titanate, and quartz replaced fine-grained albite-quartz-rich metasiltstone units interpreted as the Wallaroo Group (Fig. 7A; see Bastrakov et al., 2007, for further details). Relatively fresh, subhedral to euhedral titanite crystals in both samples (821- and 838-m depth) are intergrown with subhedral to euhedral magnetite and K-feldspar (Fig. 7C). No reaction rims or alteration along titanite-magnetite or titanite-feldspar contacts are evident. The textures suggest that the titanite grew in equilibrium with the enclosing minerals. Titanites in both samples have oscillatory zoning or have mottled internal structure under BSE imaging (Fig. 7B, D). Twenty areas have been analyzed in six large grains in sample MRD1 821. The areas analyzed have low to moderate U concentrations (~60–125 ppm) and high Th/U ratios (≥ 1.3 , most ≥ 2.6). The zoned internal structure, U concentration, and Th/U ratios are similar to titanites formed during igneous processes (Aleinikoff et al., 2002). However, we interpret the titanites in sample MRD1 as hydrothermal in origin because (1) they are intergrown with magnetite and other minerals of hydrothermal replacement origin, (2) the host rock

FIG. 5. Selection of drill core samples used in the Re-Os and ^{40}Ar - ^{39}Ar geochronology study, Moonta-Wallaroo district. A. Alford prospect, DDH130. Metasedimentary rock cut by magnetite (Mag) vein with albite- (Alb) and biotite-rich (Bt) alteration aureole, cut by K-feldspar veins. B. Alford prospect, DDH130. Quartz-pyrite (Qtz, Py) vein cutting magnetite-biotite-albite (Mag, Bt, Alb)-altered metasedimentary rock; albite and biotite are partly replaced by K-feldspar (Kfs) and chlorite (Chl) adjacent to the quartz vein. C. Alford prospect, DDH130. Tightly folded albite veins within strongly foliated biotite-pyrite (Bt-Py)-altered metasedimentary rock. Dated molybdenite (1574 ± 6 Ma) occurs in a quartz-pyrite vein similar to that in the middle piece of core. D. Moonta-Wallaroo district, Pridhams prospect, DDH213, sample depth 884.2 m. Thin calcite-chalcopryrite (Cal-Ccp) veins with minor pyrite (Py), chlorite (Chl), and sericite (Ser) cutting bedded fine-grained scapolite-rich metasedimentary rocks. Note gray spots of recrystallized and albite-calcite-altered scapolite (Scp). Dated molybdenite (1599 ± 6 Ma) from 826.0-m depth occurs in a calcite-pyrite vein similar to those illustrated. The thick black line marks the position of a thin section, and the small box (thin dotted line) outlines the location of the photomicrograph in Figure 6A.

is metasedimentary, not igneous, and (3) the titanites are coarser than the grain size of the precursor rock and are therefore unlikely to have been inherited from this metasiltstone.

The analyzed areas are relatively low in common Pb (mostly $\leq 0.5\%$ ^{206}Pb) yielding relatively well constrained radiogenic $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and ages. On the Wetherill concordia plot (Fig. 8), the analyses are all within analytical uncertainty of concordia, although the data in Appendix Table A1 show that there is some slight reverse discordance on the order of 1 to 5 percent. A possible interpretation is that the chemical composition of these titanites is different from that of the reference titanite (BLR) used to calibrate the U/Pb ratios, thereby leading to a slight bias in the U/Pb ratios. This should have no effect on the measured and calculated radiogenic $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and ages. A relative probability plot (Fig. 8) shows a simple bell-shaped distribution, and all 20 analyses have no excess scatter about the weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1574 ± 6 Ma (MSWD = 1.11).

In sample MRD1 838, 16 areas of titanite have been analyzed in seven large grains (App. Table A2). Four of the areas

analyzed are significantly enriched in common Pb (~ 8 – 18% ^{206}Pb) and this is attributed to Pb situated in cracks. Extensive cracking of the titanite can be seen in Figure 7C–D. These four areas with high common Pb yield ages with large uncertainties and have not been considered in any of the age calculations. The remaining analyses have low to moderate U concentrations (~ 110 – 205 ppm), similar to the U levels in MRD1 821 titanites, although the analysis of grain 6 indicated only ~ 2 ppm U. The Th/U ratios are quite variable, although within a particular grain there is some consistency; for example, the three analyses of grain 7 have Th/U ratios of between 0.2 and 0.4, whereas others tend to be significantly higher (≥ 1.0 , most ≥ 3.0). The titanite in this sample may have been affected by secondary processes, giving rise to variable high common Pb and some radiogenic Pb loss. The relative probability plot (Fig. 9) shows a skew in the $^{207}\text{Pb}/^{206}\text{Pb}$ ages with a more prominent ~ 1560 to 1580 Ma group and a less prominent ~ 1600 Ma subgroup. Excluding the four analyses with high common lead and the analysis with only ~ 2 ppm U, a weighted mean of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages for 11 of the areas

FIG. 6. Photomicrographs of drill core samples from the Moonta-Wallaroo and Olympic Dam districts. A. Moonta-Wallaroo district, Pridhams prospect, DDH213, sample depth 884.2 m. Photomicrograph of small box in Figure 5D, showing chalcopyrite-calcite-chlorite-sericite (Ccp, Cal, Chl, Ser) vein at right, cutting metasedimentary rock with albite-calcite-altered scapolite spots (Scp). The vein has a biotite-rich (Bt) alteration aureole. Plane-polarized light. B. Moonta-Wallaroo district, DDH7, sample depth 371.9 m. Photomicrograph of fresh scapolite (Scp) spots in banded biotite-rich (Bt) metasedimentary rock. Plane-polarized light. C. Olympic Dam district, Torrens prospect, TD2, sample depth 751.0 m. Muscovite (Ms) dated by the Ar-Ar method, intergrown with chlorite (Chl) and almost completely martitized magnetite (mar) and pyrite. Plane-polarized light.

TABLE 3. Summary of Titanite U-Pb SHRIMP Dating Results

Drill hole and depth (m)	District; prospect	Host rock	Alteration type	Sample mineralogy, texture	n	Age (Ma)	MSWD ¹
MRD1 821	Olympic Dam district; Murdie Murdie	Wallaroo(?) Group metasilstone(?)	Magnetite-K-feldspar-actinolite	Euhedral, clear, orange titanite up to 2-mm intergrown with magnetite and K-feldspar	20	1574.0 ± 6.2	1.11
MRD1 838	Olympic Dam district; Murdie Murdie	Wallaroo(?) Group metasilstone(?)	Magnetite-K-feldspar-actinolite	Euhedral, clear, orange titanite up to 2-mm intergrown with magnetite and K-feldspar	11	1580.1 ± 9.3	0.98
DD86EN26 166.95–167.3	Mount Woods inlier; Manxman West	Fine-grained feldspar-quartz-biotite gneiss and meta-igneous rock	Magnetite-albite	Veinlets of magnetite and partially leucoxene-altered, euhedral titanite up to 5-mm length, in albite alteration	18	1567 ± 9.6	0.56

¹ MSWD = mean standard weighted deviation

FIG. 7. Core samples photographs and backscattered electron (BSE) images of thin sections of selected titanite geochronology samples. A. Olympic Dam district, drill hole MRD1, sample depth 821 m. Core sample showing less altered albitic metasedimentary rock (sed) replaced by coarse hydrothermal magnetite (Mag) and K-feldspar (Kfs). B. Drill hole MRD1, sample depth 821 m, BSE image of titanite (Ttn) crystal intergrown with hydrothermal K-feldspar (Kfs), showing distinct domains with different BSE response. Several SHRIMP analysis spots are indicated by the circles. C. Drill hole MRD1, sample depth 838 m. Photomicrograph (transmitted light) showing coarse titanite (Ttn) intergrown with K-feldspar (Kfs) and magnetite (Mag). D. BSE image of same area as C, with titanite outlined and several SHRIMP analysis spots indicated by circles. E. Mount Woods inlier, Manxman West prospect, drill hole DD86EN26, sample depth 167.4 m. Core sample showing coarse titanite (Ttn) crystals (replaced by cream anatase, Ant) associated with hydrothermal magnetite (Mag) alteration of albitized meta-igneous rock. F. Drill hole DD86EN26, sample depth 166.95 to 167.3 m. Photomicrograph (transmitted light) showing relict clearer titanite (Ttn) intergrown with magnetite (Mag). Most titanite was replaced by anatase and other alteration products (Ant). SHRIMP analysis spots were in clearer areas of titanite. G. BSE image of same area as in (F).

FIG. 8. U-Pb results for titanite in sample MRD1 821, determined by in situ SHRIMP analysis. Wetherill concordia plots, relative probability plots with stacked histograms and weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages were calculated using ISOPLOT/EX (Ludwig, 2001).

analyzed yields an age of 1580 ± 9 Ma (MSWD = 0.98) with no excess scatter.

Given that samples MRD1 821 and MRD1 838 are from the same drill hole, the same lithology, and exhibit similar textures and parageneses, the U-Pb titanite analyses can be pooled, and treated as a single sample. This results in an age of 1576 ± 5 Ma (MSWD = 1.07).

Drill hole DD86EN26 at the barren Manxman West prospect in the Mount Woods inlier (Fig. 4) intersected fine- to medium-grained feldspar-quartz-biotite gneisses, intruded

FIG. 9. U-Pb results for titanite in sample MRD1 838, determined by in situ SHRIMP analysis. Wetherill concordia plots, relative probability plots with stacked histograms and weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages were calculated using ISOPLOT/EX (Ludwig, 2001). Unshaded ellipse represents data not included in the age calculation (see text).

by granite dikes. Both rock types were extensively veined and replaced by magnetite-albite $\pm$ actinolite alteration (albitic CAM). Titanite is most abundant within pink albite-altered domains that also contain disseminations and veinlets of magnetite. It also occurs within magnetite-actinolite veins. Sulfides are rare in this drill hole. At around 167-m depth in drill hole DD86EN26, partly anatase-altered coarse euhedral titanite is intimately associated with disseminated and veinlet magnetite alteration and associated albite (Fig. 7E-G). Twenty-two spots in seven less altered, clearer areas of the coarse titanite grains were analyzed and yielded surprisingly well-constrained $^{207}\text{Pb}/^{206}\text{Pb}$ ages. A number of the analyses are extremely reversely discordant (analyses 1.1, 3.1, 3.2, and 6.3; App. Table A3) and this too is most likely due to the unusual chemistry of these areas giving rise to an anomalous U-Pb ratio calibration relative to the BLR-1 reference titanite. The anomalous analyses have been excluded from any age calculations and interpretations. The relative probability plot (Fig. 10) shows a dominant group of $^{207}\text{Pb}/^{206}\text{Pb}$ ages at ~ 1560 Ma with minor scattered analyses around 1500 Ma and younger and one slightly older analysis at about 1630 Ma. The younger analyses may be due to radiogenic Pb loss; the older anomaly may be a consequence of incorrect assumptions with respect to the initial common Pb used to correct the measured ratio. A weighted mean of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages for 18 of the areas analyzed has no excess scatter at 1567 ± 10 Ma. For reasons given below, this date is interpreted as a minimum age of hydrothermal titanite associated with magnetite, albite, and actinolite alteration at the Manxman West prospect.

Mica ^{40}Ar - ^{39}Ar ages: Step-heating experiments were conducted on one muscovite sample from IOCG mineralization and two biotite samples from igneous rocks in the Olympic Dam district. The results for the micas are summarized in Table 4 and presented in Appendix Tables A4 to A6 and Figure 11.

FIG. 10. U-Pb results for titanite in sample DD86EN26, determined by in situ SHRIMP analysis. Wetherill concordia plots, relative probability plots with stacked histograms, and weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages were calculated using ISOPLOT/EX (Ludwig, 2001).

TABLE 4. Summary of $^{40}\text{Ar}/^{39}\text{Ar}$ Dating Results for the Olympic Dam District

Sample no., Drill hole, depth (m)	Prospect	Lithology	Mineral	Plateau age (Ma)	Comments	Interpretation
TD2 751.0	Torrens	Metasomatite of hematite-chlorite-muscovite-altered magnetite and pyrite	Muscovite (hydrothermal)	1575 $\pm$ 11	MSWD ¹ = 0.67, ~90% of gas	Alteration age; represents hematite-sericite-chlorite-carbonate alteration
2000366054 HUD1, 481.2-481.3	Hunter Hill	Dacite-Gawler Range Volcanics	Biotite (igneous)	No plateau	Six steps, comprising ~75% of gas give weighted mean age of 1593 $\pm$ 14 Ma, MSWD = 1.4	Extrusive age; biotite occurs as primary magmatic phenocrysts
2000366032, IDD3, 447.8-448.0	Island Dam	Mylonitic granite- Donington Suite	Biotite (igneous/ metamorphic)	1593 $\pm$ 12	MSWD = 1.70, ~80% of gas	Cooling or isotopic resetting age

¹ MSWD = mean standard weighted deviation

At the Torrens prospect (Fig. 2), hosted by intensely metasomatized fine-grained metasedimentary rocks, the abundance of white mica correlates with the intensity of hematitic replacement of preexisting magnetite (Bastrakov et al., 2007). White mica is associated with chalcopyrite, pyrite, chlorite, and carbonate (hematite-sericite-chlorite-carbonate alteration), a pattern repeated elsewhere in the Olympic Cu-Au (U) province. Relatively coarse grained (up to ~200 μm) hydrothermal muscovite from drill hole TD2 (Fig. 6C) yielded an $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum in which ~90 percent of the gas is released in 10 consecutive steps with concordant ages, defining a plateau age of 1575 $\pm$ 11 Ma (Fig. 11A). This age cannot be an inherited age, since the dated white mica is part of the hematite-sericite-chlorite-carbonate alteration assemblage and not part of the preexisting mineralogy. The hematite-sericite-chlorite-carbonate alteration is therefore constrained to have occurred no later than 1575 $\pm$ 11 Ma. The possibility remains that argon isotopes were reset at ~1575 Ma by the extensive Hiltaba-Gawler Range Volcanics magmatism, allowing hematite-sericite-chlorite-carbonate alteration to significantly predate the measured $^{40}\text{Ar}/^{39}\text{Ar}$ age. However, we consider this unlikely as the measured $^{40}\text{Ar}/^{39}\text{Ar}$ age is within uncertainty of the U-Pb titanite ages reported above for the magnetite-K-feldspar-calc-silicate alteration assemblage in the nearby MRD1 drill hole (Fig. 2; Bastrakov et al., 2007). Given the textural observations of hematite-sericite-chlorite-carbonate alteration overprinting magnetite-K-feldspar-calc-silicate alteration, the titanite age from the magnetite-bearing alteration represents a maximum age constraint on the hematite-sericite-chlorite-carbonate alteration in the district. In combination, the U-Pb titanite and $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite ages constrain the magnetite-bearing and hematitic styles of alteration to within analytical uncertainty of each other at ~1575 Ma. The interpretation of the $^{40}\text{Ar}/^{39}\text{Ar}$ date as an alteration age is further supported by fluid inclusion microthermometry and oxygen isotope geothermometry, which point to formation temperatures of ~150° to 300°C for hematite-sericite-chlorite-carbonate assemblages in the Olympic Dam district (Oreskes and Einaudi, 1992; Haynes et al., 1995; Bastrakov et al., 2002, 2007). These temperatures are below the commonly accepted argon closure temperature for white mica of ~350° to 400°C (e.g., Robbins, 1972; Purdy and Jager, 1976; Hames and Bowring, 1994; Kirschner et al., 1996; McDougall and Harrison, 1999). White mica grown under such conditions is likely to quantitatively retain radiogenic argon since crystallization and, therefore, preserve a crystallization age rather than a closure age.

Very shallow crustal levels of formation, perhaps even epithermal environments, have been suggested for some of the hydrothermal activity in the Olympic Dam deposit (Reeve et al., 1990) and Oak Dam prospect (Davidson and Paterson, 1993). The widespread preservation of Gawler Range Volcanics also indicates shallow crustal levels across most if not all the Olympic Dam district by ~1590 Ma. At such shallow crustal levels any thermal disturbance is likely to be relatively short-lived and localized. This inference is supported by an $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 1593 $\pm$ 12 Ma (2 σ) for igneous biotite from a sample of Donington Suite granite (U-Pb zircon age 1860 $\pm$ 4 Ma, Jagodzinski, 2005) in drill hole IDD3 approximately 15 km west of the Torrens prospect (sample

FIG. 11. ^{40}Ar - ^{39}Ar age spectra for muscovite and biotite in samples from the Olympic Dam district. The steps included in the calculation of plateau ages are indicated by the arrows above the age spectra.

2000366032, Table 4; Figs. 2, 10B). The $^{40}\text{Ar}/^{39}\text{Ar}$ age is interpreted as the timing of resetting of the igneous biotite during Hiltaba-Gawler Range Volcanics magmatism. Approximately 50 km farther southwest of the Emmie Bluff area, dacite from the Gawler Range Volcanics (drill hole HUD1, sample 2000366054, Table 4; Figs. 2, 10C) contains igneous biotite phenocrysts that yield an $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum that does not satisfy a strict “plateau age” definition. However, with the exception of two steps, representing ~10 percent of the gas released, the final ~85 percent of gas yields ages between 1590 and 1608 Ma, with a weighted mean age of 1593 ± 14 Ma (2σ), indistinguishable from U-Pb zircon ages for the Gawler Range Volcanics (Fanning et al., 1988; Creaser and Cooper, 1993).

Geochronological framework for IOCG hydrothermal activity

The available geochronological data for the Olympic Cu-Au-(U) province are summarized in Figure 12. Although the new results for hydrothermal minerals are reconnaissance in nature, for the first time there are sufficient data to show the timing of hydrothermal activity in relationship to other regional events in the eastern Gawler craton. In broad terms, the ages of hydrothermal minerals in IOCG alteration assemblages in the Olympic Cu-Au-(U) province overlap with the span of zircon U-Pb ages for Hiltaba Suite felsic and mafic magmatism, as discussed for each of the three IOCG districts below. It should be noted that in comparing results from different isotopic systems the error bars shown in Figure 12 represent only the analytical uncertainties and do not include uncertainties in decay constants. For the K-Ar and Re-Os systems these may each amount to ± 1 percent (Begemann et al., 2001).

In the Moonta-Wallaroo district, the minimum age of magnetite-biotite alteration based on an Re-Os molybdenite age constraint (1575 ± 6 Ma) is consistent with the U-Pb (SHRIMP) ages of hydrothermal titanite and monazite from magnetite-biotite alteration (~1570–1620 Ma; Raymond et al., 2002; Zang et al., 2002; Skirrow et al., 2006). The titanite and monazite ages also overlap with the U-Pb zircon ages of Hiltaba Suite granites and gabbroic intrusions in the Moonta-Wallaroo district, the most precise of which range from 1598 ± 7 to 1575 ± 7 Ma (Fanning in Connor, 1995; Zang et al.,

2002). In contrast, the K-Ar ages of hydrothermal and metamorphic biotite and hornblende from the district are significantly younger (~1425–1545 Ma; Webb et al., 1986; Fig. 12), indicating thermal disturbance after IOCG alteration and emplacement of the Hiltaba Suite.

In the Mount Woods inlier, the U-Pb age of 1567 ± 10 Ma for hydrothermal titanite associated with magnetite in DD86EN26 is interpreted as a minimum age for magnetite-albite-calc-silicate alteration. This is based on the observation that the southern Mount Woods inlier experienced at least local medium- to high-grade thermal metamorphism between ~1576 and ~1587 Ma (Fig. 12; Betts et al., 2003; Holm, 2005; Jagodzinski, 2005). For example, in one sample of recrystallized leucogabbro from drill hole PKN1 (Fig. 4) both magmatic and metamorphic ages have been recorded in the same zircons (cores: 1587 ± 4 Ma; rims: 1576 ± 7 Ma; Jagodzinski, 2005). As the titanite age is within error of the age of metamorphism, it is interpreted as a cooling age, indicating that the titanite crystallized at or before 1567 ± 10 Ma. The age of magnetite alteration is also constrained by the observation that a Hiltaba Suite granite dike from the same drill hole as the titanite sample has mutually crosscutting relationships with hydrothermal magnetite alteration. Igneous zircons in the dike yield an age of 1586 ± 3 Ma (SHRIMP U-Pb, Jagodzinski, 2005), which is interpreted to also represent the timing of magnetite alteration in this area.

In the Olympic Dam district, the timing of magnetite-K-feldspar-calc-silicate alteration is constrained as follows. In contrast to the Mount Woods inlier, no post-Hiltaba-aged regional thermal events capable of resetting the titanite ages in drill hole MRD1 are known in the Olympic Dam district, as evident from the preservation of ~1593 Ma ^{40}Ar - ^{39}Ar ages for biotites in Gawler Range Volcanics and Donington Suite granite samples, noted above. Fluid inclusion data and oxygen isotope geothermometry indicate temperatures of ~400° to 500°C during formation of magnetite-K-feldspar-calc-silicate assemblages in sample MRD1 and elsewhere in the Olympic Dam district (Bastrakov et al., 2007). These are below the closure temperature estimated for the U-Pb system in titanite (~700°C; Cherniak, 1993; Pidgeon et al., 1996; Zhang and Schärer, 1996). Therefore, we interpret the pooled titanite age of 1576 ± 5 Ma from the Murdie Murdie prospect as recording the age of magnetite-K-feldspar-calc-silicate alteration.

FIG. 12. Summary diagram of the geochronology of the Olympic Cu-Au province, subdivided by district and arranged generally according to decreasing age and sample type. The gray band represents the age range of Hiltaba Suite and Gawler Range Volcanics in the Gawler craton (1575–1600 Ma). ODBC = Olympic Dam Breccia Complex. Error bars are 2σ . Samples that have been analyzed by the Sm-Nd method in this study are labelled (e.g., SGD4, PKN1) and correspond to samples in Table 5. Sources of published data are Webb et al. (1986), Mortimer et al. (1988), Creaser and Cooper (1993), Johnson (1993), Fanning *in* Connor (1995), Johnson and Cross (1995), Daly et al. (1998), Raymond et al. (2002), Zang et al. (2002), Holm (2005), Jagodzinski (2005).

If magnetite and hematitic Cu-U-Au mineralization at Olympic Dam formed contemporaneously with the Olympic Dam Breccia Complex at ~1590 to 1595 Ma, as previously proposed (Reeve et al., 1990; Johnson and Cross, 1995; Jagodzinski, 2005), then our new titanite and muscovite dating results (1576 ± 5 and 1575 ± 11 Ma, respectively) and paragenetic relationships imply a possible later hydrothermal magnetite alteration event followed by hematitic alteration in the Olympic Dam district. This hydrothermal activity may have been up to 10 to 20 m.y. later than the formation of the Olympic Dam Breccia Complex. Additional data are required to test this hypothesis. In relationship to the timing of the Hiltaba Suite and mafic magmatism across the Olympic Dam district (as indicated by the available high precision U-Pb zircon ages of 1588 ± 4 to 1596 ± 4 Ma: Creaser and Cooper, 1993; Johnson and Cross, 1995; Jagodzinski, 2005; Fig. 12),

the pooled titanite age indicates magnetite-K-feldspar-calc-silicate alteration was either coeval with the final stages of magmatism or postdated the magmatism by up to several million years.

Sm-Nd Isotope Tracing of REE Sources

The effectiveness of Sm-Nd isotopes as tracers in REE-rich iron oxide hydrothermal systems has previously been demonstrated by Johnson and McCulloch (1995) and Gleason et al. (2000). As noted by Ridley and Diamond (2000) the isotopic compositions of hydrothermal precipitates will reflect the integrated effects of prior isotopic exchange between rock and fluid where this has occurred along the fluid flow paths. In the case of the Sm-Nd isotope system, mantle-derived rocks generally show primitive initial Nd isotope compositions (i.e., relatively high $\epsilon_{Nd(initial)}$ values), whereas crustally derived

rocks tend to have more evolved isotopic compositions (more negative $\epsilon_{\text{Nd}}(\text{initial})$ values) and lower Nd contents. Whole-rock Sm-Nd isotope compositions were therefore determined for samples of hydrothermal alteration and mineralization in the Olympic Cu-Au-(U) province to provide insights on the relative contributions of REE from mantle-derived versus crustal rocks encountered by the IOCG-forming hydrothermal fluids along their flow paths. The geochronological constraints on the age of IOCG alteration and mineralization presented above are an essential foundation for comparing the Nd isotope compositions of mineralization and potential source rocks.

Analytical methods

Whole-rock samples were spiked before dissolution with a mixed ^{149}Sm - ^{150}Nd tracer. They were dissolved overnight using HF and HNO_3 , evaporated to dryness, then followed by oven dissolution in fresh HF and HNO_3 for 4 d at 160°C . Samples were evaporated to dryness and converted to chloride form with 6M HCl. They were again placed in bombs overnight with 6M HCl to ensure complete equilibration. Sm and Nd were separated by conventional ion exchange methods. Nd isotope ratios were measured on a Finnigan MAT 262 multicollector, and Sm isotope ratios on a Finnigan MAT 261 single collector instrument at the University of Adelaide. During the period of this study, the La Jolla Nd isotope standard gave $^{143}\text{Nd}/^{144}\text{Nd} = 0.511845 \pm 16$ (2σ , $n = 9$), and laboratory blanks averaged <160 pg Sm and <320 pg Nd.

Sm-Nd isotope results

Representative samples of three of the major groups of IOCG alteration and mineralization assemblages and host lithologic units were analyzed for their Sm-Nd isotope composition. The total of 44 samples includes 35 host rocks, magnetite- and hematite-rich samples, and Cu-Au mineralization from barren and weakly mineralized prospects in the Olympic Dam district (Bill's Lookout, Titan, Emmie Bluff, Oak Dam, and Murdie Murdie) and in the Mount Woods inlier (Manxman A1, Manxman West). To examine regional mafic rocks as potential metal sources in IOCG systems of the Olympic Dam district and the Mount Woods inlier, a range of mafic intrusive rocks also was analyzed. These included gabbros from Bill's Lookout prospect (BLD drill holes, Fig. 2) which contain 1764 ± 12 Ma zircons (Johnson, 1993), mafic intrusions of Hiltaba Suite age dated by Jagodzinski (2005) from the Snake Gully area near Olympic Dam (drill hole SGD4, Fig. 2), and Hiltaba Suite gabbros from the Peculiar Knob area of the Mount Woods inlier (PKN1, Fig. 4). In addition, eight samples of variably hydrothermally altered Hiltaba Suite granitoids were analyzed, from drill holes in the Burgoyne Batholith a few to 30 km from the Olympic Dam deposit. In these samples hematite and sericite are the principal alteration minerals but original igneous textures are preserved (Creaser, 1989). Results are presented in Table 5 and Figures 13 and 14.

All ϵ_{Nd} values discussed below are calculated at 1590 Ma, based on the age of mineralization at the Olympic Dam deposit (Johnson and Cross, 1995). This allows direct comparison of results with those for the deposit. Since the geochronological data for hydrothermal minerals presented herein can

FIG. 13. Neodymium isotope compositions (ϵ_{Nd} calculated at 1590 Ma) of whole-rock samples vs. copper content (A) and SiO_2 content (B). Hiltaba Suite granitoid samples from this and previous studies (Creaser, 1989, 1995; Johnson and McCulloch, 1995) are mostly distal from the Olympic Dam deposit (see Table 4, Fig. 2). "Altered granitoid" field in (A) includes sericitized and chloritized granite breccia from the Olympic Dam deposit and Na-depleted regional granitoid rocks. Data for Olympic Dam mineralization and magnetite breccia are from Johnson and McCulloch (1995).

be interpreted to indicate that at least some IOCG alteration occurred up to ~ 1575 Ma, data were also recalculated for this age. However this yields ϵ_{Nd} values with negligible difference to those calculated at 1590 Ma (approx 0.1 units of ϵ_{Nd}), and consequently the interpretations of the Sm-Nd isotope data are unchanged.

Neodymium isotope data from variably altered Hiltaba Suite granitoid rocks from the Olympic Dam area are presented in Figure 13. Two least altered samples have initial ϵ_{Nd} values of -4.3 and -4.8 , indistinguishable from previous results of least altered samples from Creaser (1995, 1996). Variably altered granites, as indicated by sericite-hematite alteration of igneous feldspar (Creaser, 1989) have higher ϵ_{Nd} values. Figure 13A shows this Nd isotope signature is roughly correlated to Cu concentration. This result reinforces the data of Johnson and McCulloch (1995), who first noted this correlation in highly altered granitic breccias in the Olympic Dam deposit (Fig. 13A). The available data support the interpretation that Nd and Cu were cotransported in the same hydrothermal fluids.

The ϵ_{Nd} values of regional mafic rocks range from -0.8 to -6.0 . At the Bill's Lookout prospect in the Olympic Dam district, ϵ_{Nd} values for four samples from drill hole BLD1 range from -0.8 to -2.1 . A mafic dike of Hiltaba Suite age from drill hole SGD4 (Jagodzinski, 2005), to the northeast of Olympic

TABLE 5. Sm-Nd Isotope Data for Whole-Rock Samples of IOCG Mineralization and Regional Host Rocks

Sample no.	DDH	Depth (m)	Description	Unit/Suite	Alteration	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}^1$	$\epsilon_{\text{Nd}}(1590 \text{ Ma})^2$
Mount Woods inlier										
Peculiar Knob (PKN) and Manxman prospects										
2002360028A	PKN1	324.1-326.7	Micro-gabbro	Hiltaba-aged(?)	Least altered	6.9	36.2	0.1147	0.511518 ± 9	-5.2
2002360028B	PKN1	326.7-328.8	Leuco-gabbro	Hiltaba-aged ³	Least altered	10.8	57.0	0.1141	0.511470 ± 9	-6.0
2001368015F	DD86EN26	276.05-276.7	Foliated granitic dike	Hiltaba Suite ³	Least altered	9.3	45.7	0.1234	0.511564 ± 7	-6.0
2001368017E	DD89EN61	327.3-328.2	Coarse-grained granite	Hiltaba Suite(?)	Least altered	8.9	58.6	0.0921	0.511254 ± 8	-5.7
2001368017O	DD89EN61	420-420.5	Mag. massive	Paleoproterozoic(?)	Mag-Alb-calc	1.0	7.4	0.0844	0.511250 ± 7	-4.2
2001368017D	DD89EN61	322.10	Mag-Py-Cep vein	Paleoproterozoic(?)	Mag-Alb-calc	9.5	63.2	0.0911	0.511319 ± 10	-4.2
Olympic Dam district										
Bill's Lookout intrusions										
2001368001O	BLD1	693.8-694.5	Gabbro	1764 ± 12 Ma ⁴	Least altered	3.2	14.1	0.1353	0.511892 ± 7	-2.1
2001368001P	BLD1	696.3-696.9	Gabbro	1764 ± 12 Ma ⁴	Least altered	2.7	11.7	0.1374	0.511918 ± 7	-2.0
2001368001Q	BLD1	699.1-700.1	Dolerite		Least altered	4.9	19.2	0.1540	0.512152 ± 8	-0.8
2001368001T	BLD1	718.3-718.7	Gabbro	1764 ± 12 Ma ⁴	Least altered	4.7	18.3	0.1539	0.512150 ± 7	-0.8
Titan Prospect										
TI002 753.63	TI002	753.63	Mafic dike	Hiltaba-aged(?)	Least altered	12.0	72.6	0.0997	0.511493 ± 9	-2.6
TI002 656.19	TI002	656.19	Felsic dike	Gawler Range Volcanics(?)	Least altered	2.5	12.1	0.1226	0.511539 ± 8	-6.4
TI002 368005	BD1	897.4-897.7	Mag-calc-Py	Lower Wallaroo Gp(?)	Mag-Kfs-calc	2.0	14.3	0.0841	0.511200 ± 10	-5.1
TI002 744	TI002	744	Hem breccia	Lower Wallaroo Gp(?)	HSCC	16.1	85.3	0.1139	0.511504 ± 8	-5.3
Emmie Bluff prospect										
2001368023A	SAE7	1090.5-1090.9	Trachytic lava	Gawler Range Volcanics(?)	Weak Mag-Kfs-calc	11.3	75.9	0.0900	0.511306 ± 9	-4.3
2001367000D	SAE4	1152.7-1153.1	Granite	Donington Suite ⁵	Least altered	3.2	14.4	0.1357	0.511736 ± 8	-5.2
2001367000F	SAE4	1162.7-1163.1	Altered granite	Donington Suite ⁵	Weak HSCC	8.1	43.8	0.1118	0.511482 ± 8	-5.3
2001368007S	SAE6	1013.80	Altered metasediment	Upper Wallaroo Gp(?)	Weak HSCC	9.8	52.1	0.1132	0.511485 ± 7	-5.5
2001368023B	SAE7	1112.2-1112.4	Mag-calc-silicate	Upper Wallaroo Gp(?)	Mag-Kfs-calc	2.4	18.6	0.0794	0.511133 ± 7	-5.5
2001368007U	SAE6	1022.30	Massive laminated Mag	Upper Wallaroo Gp(?)	Mag-Kfs-calc	1.9	9.9	0.1147	0.511533 ± 7	-4.9
2001368007X	SAE6	1029.00	Mag-Py-Cep	Upper Wallaroo Gp(?)	Mag-Kfs-calc	2.5	12.7	0.1209	0.511555 ± 10	-5.7
2001368007Y	SAE6	1029.00	Mag-Py-Cep	Upper Wallaroo Gp(?)	Mag-Kfs-calc	2.5	12.6	0.1212	0.511555 ± 11	-5.8
2001368007E	SAE6	959.70	Late Qtz-Py	Upper Wallaroo Gp(?)	HSCC	9.7	58.4	0.1003	0.511389 ± 9	-4.8
2001368007G	SAE6	970.50	Late Hem-Qtz	Upper Wallaroo Gp(?)	HSCC	0.8	1.8	0.2752	0.512994 ± 10	-9.1
Oak Dam prospect										
GD23A	AD1	740.5	Altered granite	Donington Suite(?)		12.1	124.7	0.0588	0.511013 ± 8	-4.3
GD23B	AD1	742.3	Altered granite	Donington Suite(?)		11.3	81.4	0.0838	0.511180 ± 8	-5.5
GD23C	AD1	744.1	Altered granite	Donington Suite(?)		10.9	102.2	0.0647	0.511080 ± 8	-3.5
GD52	AD1	711.5	Mag alteration	Donington Suite(?)	Mag-Kfs-calc	68.2	606.5	0.0680	0.511070 ± 9	-4.4
GD53	AD1	718.1	Mag alteration	Donington Suite(?)	Mag-Kfs-calc	26.2	189.5	0.0836	0.511250 ± 9	-4.1
GD46	AD1	648.8	Py-Hem ironstone	Donington Suite(?)	HSCC	5.5	27.3	0.1226	0.511577 ± 10	-5.6
GD45	AD1	640.3	Hem ironstone	Donington Suite(?)	HSCC	5.6	26.9	0.1248	0.511630 ± 8	-5.1
GD44a	AD1	629.7	Hem breccia	Donington Suite(?)	HSCC	38.0	272.7	0.0843	0.511295 ± 9	-3.3
Murdie prospect										
2001368002E	MRD1	828.00	Altered metasediment	Wallaroo Gp(?)	Weak Mag-Kfs-calc	3.0	17.3	0.1030	0.511454 ± 9	-4.0
2001368002G	MRD1	834.30	Mag	Wallaroo Gp(?)	Mag-Kfs-calc	1.7	8.9	0.1159	0.511568 ± 7	-4.4
2001368025Z	SAR9	1114.85-1115.4	Fresh metasediment	Wallaroo Gp(?)	Least altered	1.8	7.9	0.1358	0.511667 ± 10	-6.6
Hiltaba Suite granitoids and Gawler Range Volcanics (Olympic Dam district)										
20003660039	OFD1	456.5-457.3	Granitoid; 55.6% SiO ₂	Hiltaba Suite	Least altered; 211 ppm Cu	7.1	41.6	0.1029	0.511398 ± 7	-5.1
20003660047	OFD3	481-484.1	Granitoid; 57.7% SiO ₂	Hiltaba Suite	Altered; 986 ppm Cu	9.7	64.3	0.0912	0.511453 ± 7	-1.6
9339	MA-55		Granitoid; 67.1% SiO ₂	Hiltaba Suite	Na-depleted; 2070 ppm Cu	10.2	83.1	0.0740	0.511176 ± 9	-3.5

TABLE 5. (Cont.)

Sample no.	DDH	Depth (m)	Description	Unit/Suite	Alteration	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}^1$	$\epsilon_{\text{Nd}}(1590 \text{ Ma})^2$
9349	HRD2	445.1-447.2	Granitoid; 57.7% SiO_2	Hiltaba Suite	Least alt; 23 ppm Cu	13.0	69.1	0.1141	0.511561 ± 7	-4.2
9356	BLD2	827.2-829.6	Granitoid; 59.5% SiO_2	Hiltaba Suite	Na-altered?; 235 ppm Cu	10.1	53.6	0.1140	0.511550 ± 8	-4.4
9395	WRD1	494-495.8	Felsic extrusion; 62.1% SiO_2	Gawler Range Volcanics	Na-depleted; 381 ppm Cu	8.9	66.5	0.0807	0.511289 ± 8	-2.7
9403	OFD2	406.4-406.7	Granitoid; 54.0% SiO_2	Hiltaba Suite	Least altered; 98 ppm Cu	9.6	53.8	0.1075	0.511509 ± 7	-3.9
9410	RD693	679.5-679.65	Granitoid; 73.3% SiO_2	Hiltaba Suite	Na-depleted; 2210 ppm Cu	5.9	37.1	0.0965	0.511132 ± 8	-9.0
2000366005	SGD4	405-445	Mafic dike; 44.5% SiO_2	Hiltaba-aged ³	Least altered; 191 ppm Cu	5.3	24.6	0.1295	0.511818 ± 8	-2.3

Notes: Abbreviations: HSCC = hematite-sericite-chlorite-carbonate, Alb = albite, calc = calc-silicate, Cep = chalcopyrite, Hem = hematite, Kfs = K-feldspar, Mag = magnetite, Py = pyrite, Qtz = quartz; PKN samples courtesy Minotaur Exploration Ltd., SGD sample courtesy WMC Resources, Titan T1002 samples courtesy Tasman Resources Ltd.; Hiltaba granite samples courtesy Rob Creaser/La Trobe University/WMC Resources/Liz Jagodzinski, Oak Dam samples courtesy Garry Davidson/WMC Resources

¹ Measured values, normalised to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, standard error given in 2σ

² $\epsilon_{\text{Nd}}(1590)$ calculated at the age of the Olympic Dam deposit (Johnson and McCulloch, 1995), using $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1966$ as present day chondritic uniform reservoir values

³ Jagodzinski (2005)

⁴ Johnson (1993)

⁵ O. Holm, writ. commun. (2004)

FIG. 14. Frequency histograms for Nd isotope compositions (n = number of samples), calculated at 1590 Ma, for barren and weakly mineralized IOCG prospects in the Mount Woods inlier (A) and Olympic Dam district (B-D), compared to the Olympic Dam deposit (E). Data for (A)-(D) from this study; data for Olympic Dam from Johnson and McCulloch (1995). The vertical gray bar spans the range of $\epsilon_{\text{Nd}}(1590)$ measured on least altered granitoids (see also Fig. 13) and also encompasses the majority of data from magnetite-bearing and hematitic alteration at the barren and weakly mineralized prospects. In contrast, $\epsilon_{\text{Nd}}(1590)$ values for Cu-rich hematitic mineralization at Olympic Dam (labelled hematitic in E) lie to the right of this range, interpreted to result from an Nd component derived from mafic and/or ultramafic dikes, as shown by the arrow in (E). Abbreviations: Alb = albite, calc = calc-silicate, HSCC = hematite-sericite-chlorite-carbonate, Kfs = K-feldspar, Mag = magnetite.

Dam, gave ϵ_{Nd} of -2.3 . Significantly lower values were recorded for gabbroic rocks in the Peculiar Knob area of the Mount Woods inlier (Fig. 4), with two samples from a Hiltaba-age comagmatic suite returning ϵ_{Nd} values of -5.2 and -6.0 . In contrast to the mafic rocks analyzed from the Olympic Dam district, these values are close to the compositions of two Hiltaba-age granitic intrusions at the Manxman prospect in the southern Mount Woods inlier (-5.7 , -6.0 ; Table 5, Fig. 4). A distinctly crustal contribution of REE is therefore evident in both felsic and mafic intrusions of Hiltaba Suite age in the Mount Woods inlier.

Data for IOCG prospects (Fig. 14) were obtained from the Mount Woods inlier (Manxman A1 and Manxman West; Fig. 4) and in the Olympic Dam district from the Titan, Emmie Bluff, Oak Dam, and Murdie Murdie prospects (Fig. 2). Two magnetite-rich mineralized samples from drill hole DD89EN61 at Manxman A1 have an ϵ_{Nd} of -4.2 , slightly less negative than the host granitic lithologic units.

At the Titan prospect, samples of magnetite-K-feldspar-calc-silicate alteration with weak chalcopyrite mineralization (typically 0.1–0.2 wt % Cu; Tasman Resources NL, Annual Report 2004, www.tasmanresources.com.au) and later barren hematitic breccia (2001368005 and TI002 744) have indistinguishable ϵ_{Nd} values of -5.1 and -5.3 . An undeformed felsic porphyry dike of inferred Gawler Range Volcanics affinity that is cut by magnetite-K-feldspar-calc-silicate alteration yielded a value of -6.4 , whereas an unaltered fine-grained mafic dike that is spatially and temporally associated with the hematitic breccia in drill hole TI002 yielded a significantly higher value of -2.6 .

Four samples of host rocks from the Emmie Bluff prospect, including weakly altered trachytic lava, granitic and metasedimentary rocks, yield ϵ_{Nd} values between -5.5 and -4.3 . All five samples of mineralization and alteration from the Emmie Bluff prospect, including both magnetite-K-feldspar-calc-silicate and hematite-sericite-chlorite-carbonate alteration, have ϵ_{Nd} values between -5.8 and -4.8 , with the exception of sample 2001368007G, a massive magnetite sample almost completely replaced by bladed and specular hematite and silica. Its ϵ_{Nd} value is -9.1 , and it also has the highest measured $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratio. This anomalous measured ratio and the high Sm/Nd ratio suggest that the REE have been strongly fractionated, resulting in relative LREE depletion. Three samples of host granitoid from the Oak Dam prospect are variably altered, and ϵ_{Nd} values range between -3.5 and -5.5 . Five examples of both magnetite- and hematite-bearing alteration from Oak Dam also were analyzed. These samples yielded the highest REE concentrations from any prospect, with Nd up to 607 ppm in one magnetite-altered sample (GD52). Values of ϵ_{Nd} range from -5.6 to -3.3 . Two samples of host metasedimentary rock and massive magnetite-K-feldspar-calc-silicate alteration from Murdie Murdie prospect have ϵ_{Nd} values of -4.0 and -4.4 , respectively.

Discussion

Sources of REE in IOCG hydrothermal systems

Differences in the Nd isotope compositions of hematitic ores compared to less mineralized granite host rock led Johnson and McCulloch (1995) to conclude that the ore signature

at Olympic Dam must have involved contributions from mantle-derived source rocks or magmas with a significantly more primitive Nd isotope composition (i.e., more positive ϵ_{Nd}) than the granitic host rock. Based on two-component mass-balance calculations, Johnson and McCulloch (1995) argued that at least 30 percent of the Nd in the deposit was sourced from mantle-derived magmas or igneous rocks such as the altered alkaline mafic and/or ultramafic dikes in the deposit, with the remainder contributed from crustal source(s) such as the host granite. It was suggested that a 13 percent source rock fraction of mafic and/or ultramafic rocks can account for ~50 percent of the Cu contained in the deposit, if Nd and Cu were leached from the same source rocks. This mass-balance calculation would not be applicable, however, if metals were introduced directly by magmatic volatiles (Johnson and McCulloch, 1995). In contrast, magnetite-rich breccia and a magnetite mineral separate at Olympic Dam have lower $\epsilon_{\text{Nd}(1590)}$ values of -4 to -4.9 , which Johnson and McCulloch (1995) interpreted as resulting from fluids in equilibrium with the host granite.

The conclusion that the REE were derived in part from a source more primitive than the granitic host rock is further supported by the new data from variably altered Hiltaba Suite granitoid rocks in the vicinity of the Olympic Dam deposit (Table 5, Fig. 13). Copper concentrations in the variably altered granites ranges from 15 to 2,100 ppm Cu (Creaser, 1989), and minor sulfides are visible in some samples (Creaser, 1995, 1996). The positive correlation between Cu and initial ϵ_{Nd} values for the variably altered granites is shown in Figure 13A, along with the previous data from Johnson and McCulloch (1995) and Creaser (1995, 1996). An increase in ϵ_{Nd} values in the granitoids could be attributed to mixing of mafic magma of high ϵ_{Nd} with felsic magma of lower ϵ_{Nd} , yet the plot of SiO_2 versus initial ϵ_{Nd} (Fig. 13B) shows that there is no clear correlation between the two. We therefore suggest that the relationship between Cu content in the granites and ϵ_{Nd} is further evidence of a hydrothermal contribution of REE from a primitive source, even in weakly altered granites many kilometres from the deposit, and justifies the use of the Nd isotope system as a tracer of IOCG fluids in the Olympic Cu-Au-(U) province.

Figure 14 presents our $\epsilon_{\text{Nd}(1590)}$ data in comparison with previously published data for the Olympic Dam deposit. A striking result is that without exception the studied prospects have lower $\epsilon_{\text{Nd}(1590)}$ values than those for copper ores in hematitic breccias at Olympic Dam. This applies to both magnetite- and hematite-rich samples from the barren and weakly mineralized prospects. The differences in Nd isotope compositions between the prospects and hematite-copper ores at Olympic Dam are attributed to different sources of REE. At Emmie Bluff, the $\epsilon_{\text{Nd}(1590)}$ values reach a low of -9 , whereas at Oak Dam they range up to -3.5 . As summarized in Figure 13, $\epsilon_{\text{Nd}(1590)}$ values for four barren and weakly mineralized prospects of less than about -4 strongly suggest that Nd was derived dominantly from crustal sources such as the Hiltaba Suite granites and/or Gawler Range Volcanics (shown as a gray band in Fig. 14), and/or from Wallaroo Group metasedimentary rocks. Unlike the Olympic Dam hematite-Cu ore signature, which has distinctly more positive $\epsilon_{\text{Nd}(1590)}$ values than magnetite in the deposit (Johnson and McCulloch,

1995), there is no compelling evidence for significant input of Nd from primitive sources in these four prospects. The new data imply that hematitic alteration of previously magnetite-altered rocks at Emmie Bluff and Oak Dam have lowered, or have not significantly shifted, the $\epsilon_{\text{Nd}(1590)}$ values. Indeed the lowest $\epsilon_{\text{Nd}(1590)}$ value of -9 is for a sample rich in hydrothermal specular hematite at Emmie Bluff, requiring isotopic exchange between the hematite-forming fluids and a source with strongly negative $\epsilon_{\text{Nd}(1590)}$ signature such as Archean or Early Paleoproterozoic rocks.

In the Nd isotope study of Gleason et al. (2000) the isotopic compositions of REE-rich magnetite-apatite deposits in four districts in North America were compared with their coeval igneous host rocks and regional rocks. It was concluded that the primary sources of REE in the Fe deposits were the coeval igneous host rocks. Our results for barren and weakly mineralized IOCG hydrothermal systems of the Gawler craton also can be interpreted to indicate that the REE were sourced dominantly from local host rocks (Fig. 14). However, the $\epsilon_{\text{Nd}(1590)}$ compositions of the broadly coeval igneous rock sources (i.e., Hiltaba Suite, Gawler Range Volcanics) are similar to those of the metasedimentary and metaigneous host rocks, so we cannot discriminate between these possible crustal sources of REE.

A mantle-derived contribution of REE in iron oxide Au-Cu-Bi deposits of the Tennant Creek district, Australia, was suggested by Skirrow (1999) on the basis of Nd isotope data. Barren magnetite-rich deposits show strongly evolved Nd isotope signatures, indicating crustal sources of REE, whereas Au- and Cu-Au-rich mineralization has more positive ϵ_{Nd} compositions, consistent with REE contributions from relatively primitive sources.

Mathur et al. (2002) utilized the Re-Os isotope system to compare Chilean IOCG systems with barren magnetite-apatite deposits. The isotopic compositions of the magnetite-apatite deposits indicate a radiogenic (dominantly crustal) source of Os, whereas a mixed mantle-crustal isotopic signature is evident in the major IOCG deposits.

Clearly there are parallels between our findings and these other studies of metal sources in IOCG hydrothermal systems. Additional radiogenic isotope data for IOCG provinces, including major deposits and barren systems, are required to further test the generality of mantle-derived contributions of metals to major IOCG deposits.

Constraints on fluid and sulfur sources from stable isotopes

Although the Sm-Nd isotope data are useful in discriminating mantle from crustal sources of REE, other methods are required to unravel the roles of different crustal sources in the formation of the IOCG systems. Bastrakov et al. (2002, 2007) use O, H, and S isotope and fluid inclusion geochemical data to place constraints on the sources of fluids related to the magnetite-bearing and hematite-sericite-chlorite-carbonate alteration and on the sources of sulfur in the Olympic Dam district. Brines responsible for magnetite-K-feldspar-calc-silicate alteration were hypersaline, $>400^\circ\text{C}$ and carried >500 ppm copper, whereas fluids related to hematite-sericite-chlorite-carbonate alteration had lower salinity and temperatures of 200° to 300°C with low copper contents in studied prospects. The results for the magnetite-related brines point

to extensive geochemical and isotopic (Nd, O, H, S, Br/Cl) exchange with Wallaroo Group or other metasedimentary rocks, resulting in properties not unlike a high-temperature variant of basinal or crystalline basement brines (Bastrakov et al., 2002, 2007). Nevertheless, the data for magnetite-K-feldspar-calc-silicate alteration are permissive of parental fluids of ultimately magmatic origin that have lost some or most of their original (igneous) geochemical and isotopic character through reequilibration with metasedimentary rocks. Based on more limited isotopic and geochemical data for fluids related to hematite-sericite-chlorite-carbonate alteration, the evidence is equivocal for direct involvement of meteoric waters in formation of hematitic assemblages in most barren and weakly mineralized prospects of the Olympic Dam district (Bastrakov et al., 2007). Deeply sourced or deeply circulated fluids appear to have been responsible for both magnetite-K-feldspar-calc-silicate and hematite-sericite-chlorite-carbonate alteration in some prospects (e.g., Titan, Emmie Bluff), whereas meteoric waters may have been important in other prospects (e.g., Torrens).

In contrast, the oxygen isotope data for the Olympic Dam deposit clearly indicate the presence of fluids of relatively low $\delta^{18}\text{O}$ ($>1\text{‰}$) during hematite development (Oreskes and Einaudi, 1992). These fluid isotopic compositions can be explained by influx of meteoric waters that either mixed with isotopically "heavy" fluids of deep-seated origin (Haynes et al., 1995) or postdated such deeply sourced fluids (Einaudi and Oreskes, 1992). Alternatively, the low $\delta^{18}\text{O}_{\text{fluid}}$ values are consistent with down-temperature reequilibration of fluids with initially high $\delta^{18}\text{O}$ with granitic or felsic volcanic rock at low water/rock ratios (Bastrakov et al., 2007). Moreover, sulfide minerals at Olympic Dam have average $\delta^{34}\text{S}$ values (-6 to -10‰ , Eldridge and Danti, 1994) lower than most prospects, with the exceptions of Oak Dam and Manxman A1 (Bastrakov et al., 2007). These relatively low values may be the product of either (1) relatively oxidized fluids ($\Sigma\text{SO}_4^{2-} \approx \Sigma\text{H}_2\text{S}$) carrying sulfur leached from igneous rocks or contributed directly from magmas, or (2) very oxidized fluids ($\Sigma\text{SO}_4^{2-} \gg \Sigma\text{H}_2\text{S}$) with relatively high $\delta^{34}\text{S}_{\text{SS}}$ values (e.g., 13‰ , Bastrakov et al., 2007). Possible sources of "heavy" sulfur for the second scenario are metasedimentary rocks, sulfate-bearing evaporitic brines, or magmatic-hydrothermal sulfate. Notwithstanding these alternative sources of fluids and sulfur, we conclude that the fluid evolution during hematite-sericite-chlorite-carbonate alteration and mineralization at the Olympic Dam deposit differed from that at most of the barren and weakly mineralized prospects in the district. Oak Dam arguably shows the most similarities to Olympic Dam in terms of its Nd and S isotope characteristics.

An integrated hydrothermal model of the Olympic Dam district

Synthesis of the Nd isotope and geochronology results with the available fluid inclusion and stable isotope data provides the basis for an integrated model of IOCG hydrothermal systems in the central part of the Olympic Cu-Au-(U) province. This is illustrated schematically in Figure 15 for the Olympic Dam district, with the geology reconstructed for ~ 1570 to 1595 Ma. The model shows a near-surface paleoenvironment for the Olympic Dam deposit at ~ 1590 Ma, as proposed by

FIG. 15. Schematic crustal cross section of iron oxide Cu-Au hydrothermal systems at ~1570 to 1595 Ma in the Olympic Dam district of the Olympic Cu-Au province. The location of the section is shown in Figure 2. The reconstructed geology is based on interpretation of seismic reflection data (Lyons et al., 2005) and recent geochronology (e.g., Jagodzinski, 2005; this study). Mafic and/or ultramafic dikes observed at Olympic Dam (Reeve et al., 1990) are not imaged in the seismic data and hence are illustrated as subvertical bodies. Schematic fluid pathways are inferred from known and interpreted alteration distribution (e.g., Fig. 2B), geochemical data (this study; Bastrakov et al., 2007), and geophysical studies (Williams et al., 2004; Dieren and Lyons, 2007). The volcanic and/or diatreme setting for the Olympic Dam deposit is based on Reeve et al. (1990).

Reeve et al. (1990). Volcanic edifices of the Gawler Range Volcanics are depicted at this time because the geochronological results for Olympic Dam (Johnson and Cross, 1995; Jagodzinski, 2005) and elsewhere in the district (this study) indicate that IOCG hydrothermal activity temporally overlapped and probably outlasted this volcanism. Creaser (1996) determined emplacement depths of 4 to 8 km for parts of the Burgoyne batholith, based on Al hornblende geobarometry and P_{H_2O} estimates. Given that ages of Hiltaba Suite granites and Gawler Range Volcanics are indistinguishable at the 2σ level of the uncertainties (Mortimer et al., 1988; Creaser and Cooper, 1993; Johnson and Cross, 1995; Jagodzinski, 2005), we infer that at least parts of the batholith were rapidly exhumed and unroofed prior to or during volcanism and coeval formation of the Olympic Dam Breccia Complex. In Figure 15 the Burgoyne batholith is therefore shown at a shallow position during Olympic Dam Breccia Complex formation rather than at its initial emplacement at depth.

The model draws upon the crustal architecture interpreted from seismic reflection data recently acquired in the vicinity of Olympic Dam (Lyons et al., 2005). The Burgoyne batholith is the only Hiltaba Suite intrusive complex to be clearly imaged in the north-south seismic section, and there is no evidence for a caldera complex (cf. Gow et al., 1994b). Paleoproterozoic units (Walleroo Group, Donington Suite metagranitoids, and inferred Hutchison Group) overlie a distinctive mid-crustal reflector at around 10-km depth, which may be partly mafic in composition (Lyons et al., 2005). The ages of this part of the crust and the underlying middle to lower crust are unknown. Both middle and upper crust are cut by

northerly dipping reflective features in the seismic data that are interpreted as major fault complexes. Structural geometries indicate post-Walleroo Group thrusting on some of the upper crustal major faults (Lyons et al., 2005; O. Holm, unpubl. geochronological data). The Wirrda Well, Acropolis, Torrens, and several other IOCG prospects lie within or close to one such complex, the Andamooka fault zone (Figs. 2, 14). Based on these spatial relationships together with geophysical inversion modeling of the three-dimensional distribution of hematite and magnetite alteration (Fig. 2B; Williams et al., 2004), we suggest that the Andamooka fault zone and associated northwest-trending splays were part of the plumbing systems for IOCG fluid flow in the period 1595 to 1570 Ma. These structures are imaged in seismic data to depths of ~10 km, although it is not clear whether any movement at 1595 to 1570 Ma was reverse or normal and/or strike-slip.

Magnetite-K-feldspar-calc-silicate alteration in prospects of the Olympic Dam district involved high-temperature brines with significant inputs of O, H, Br, and Cl from non-magmatic sources such as the metasedimentary rocks of the Wallaroo Group. On the other hand, Nd (and Cu?) in magnetite-K-feldspar-calc-silicate alteration was sourced mainly from crustal felsic igneous rocks or magmas and/or Wallaroo Group, and input of S was dominantly from felsic or mafic igneous rocks or magmas. This scenario is consistent with high-temperature brines having reacted extensively with both metasedimentary and felsic igneous rocks known in the district, including the Wallaroo Group, the Donington Suite granitoids, Gawler Range Volcanics, and the Burgoyne batholith. In Figure 15, the pathways of the

magnetite-forming brine are therefore shown within the mixed meta-igneous-metasedimentary Paleoproterozoic rock package, although the pathways also could have involved the Mesoproterozoic felsic igneous rocks at the highest crustal levels.

The ultimate source(s) of the magnetite-forming brines is enigmatic. If the parental fluids were magmatic-hydrothermal in origin, as permitted by the stable and radiogenic isotope data, their isotopic and geochemical signatures were heavily disguised through reaction with the regional host-rock packages. Magmatic-hydrothermal derivation could account for the high temperatures of the brines at shallow crustal levels and their high salinities. However, no intrusions of Hiltaba or Gawler Range Volcanics age have been identified proximal to the magnetite-K-feldspar-calc-silicate alteration zones examined in this study, with the exception of the Titan prospect, and none are evident in seismic data (Lyons et al., 2005). Either the sources of such fluids were deeply buried magmas that are not recognized in seismic data or other geophysical data or, alternatively and more likely, they were nonmagmatic.

Available Nd and oxygen compositions and fluid inclusion properties of magnetite-bearing alteration in the Olympic Dam deposit (Oreskes and Einaudi, 1992; Johnson and McCulloch, 1995) are similar to those of magnetite-K-feldspar-calc-silicate alteration in barren and weakly mineralized prospects elsewhere in the Olympic Cu-Au-(U) province. It is suggested that the regional processes leading to magnetite alteration were probably similar at Olympic Dam and at the barren and weakly mineralized prospects of the Olympic Cu-Au-(U) province (see arrows representing magnetite-forming brine in Fig. 15).

Hematite-sericite-chlorite-carbonate and magnetite-feldspar-calc-silicate alteration in some of the barren and weakly mineralized prospects involved similar crustal sources of Nd, S, O, and possibly H, suggesting the fluids responsible for hematite-sericite-chlorite-carbonate alteration in some cases may have evolved from the brines that formed magnetite-feldspar-calc-silicate alteration and utilized similar flow paths. Involvement of surficial waters appears to have been minimal except in some systems hosted by Gawler Range Volcanics (e.g., Mt. Gunson area) and at the Torrens and perhaps Oak Dam prospects. However, the Olympic Dam deposit is distinguished from the hematite-sericite-chlorite-carbonate alteration in barren and weakly mineralized prospects in a number of important geological and geochemical characteristics. First, the Olympic Dam Breccia Complex was an active intrusive and probably extrusive center during breccia formation and alteration, involving felsic and mafic and/or ultramafic dikes and multiple overprinting phreatomagmatic and hydrothermal brecciation events (Reeve et al., 1990; Johnson and Cross, 1995). Second, the Nd isotope data (Johnson and McCulloch, 1995; this study) suggest that Nd in the Olympic Dam copper ores has a different, more primitive, signature than in minor copper mineralization in outlying barren and weakly mineralized prospects in the region, reflecting greater input of Nd and probably Cu from mafic and/or ultramafic rocks and/or magmas. This is illustrated schematically in Figure 15 with arrows of hematite-forming waters interacting with both mantle-sourced mafic-ultramafic intrusions and Hiltaba Suite granite. Such fluids at the Olympic Dam deposit would have carried greater concentrations of both cop-

per and uranium (leached from granite?) than hematite-related fluids in barren and weakly mineralized prospects that lacked either mafic and/or ultramafic or Hiltaba Suite granite sources of these metals. Finally, surficial waters carrying dissolved sulfate (Haynes et al., 1995) may have been more significant at the Olympic Dam deposit than in barren and weakly mineralized prospects (Fig. 15). This could explain the differences in $\delta^{34}\text{S}$ values of sulfides and very oxidized depositional conditions at the giant deposit in comparison to the prospects (Bastrakov et al., 2007).

Conclusions

Dating of hydrothermal minerals in conjunction with relative timing constraints at a number of barren and weakly mineralized prospects in three districts in the eastern Gawler craton demonstrates that IOCG hydrothermal systems were active during the period ~1575 to 1600 Ma. These results support the existence of a >500-km-long IOCG metallogenic belt, the Olympic Cu-Au-(U) province (Skirrow et al., 2002). The geochronological and geological results demonstrate there is a regional spatial and temporal association between the IOCG hydrothermal systems and felsic and mafic magmatism of the Hiltaba Suite and Gawler Range Volcanics. Mantle-derived mafic and/or ultramafic rocks and/or magmas were more important as source(s) of REE and Cu in the Olympic Dam deposit than in barren and weakly mineralized IOCG systems in the Olympic Dam district and the Mount Woods inlier.

For the Olympic Dam district, at least, the neodymium isotope signature of mineralization may provide a means of discriminating between major IOCG deposits and minor mineralization. Such information could assist exploration decision making where, for example, few drill holes have tested deeply buried alteration and explorers need to assess whether the hydrothermal alteration is part of a large and fertile IOCG ore-forming system.

Mafic and/or ultramafic rocks or magmas may represent both a source of ore components and a manifestation of the crustal- to lithospheric-scale thermal processes that drove large IOCG hydrothermal systems. The presence of mafic intrusions of Hiltaba Suite age in each of the three main IOCG districts of the Olympic Cu-Au-(U) province could be taken to indicate that each district has potential for one or more major IOCG deposits, provided that such mineralization was preserved through any subsequent exhumation. The Olympic Dam, Prominent Hill, and Carrapateena deposits may be the first three of several major deposits awaiting discovery in the Olympic Cu-Au-(U) province.

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APPENDIX

TABLE A1. Summary of SHRIMP U-Pb Titanite Results for Sample MRD1 821

Area spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Radiogenic Ratios						Age (Ma)					
							²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	±	ρ	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	% Disc	
1.1	76	206	2.71	18	0.000035	0.06	0.2790	0.0048	3.751	0.070	0.0975	0.0007	0.928	1587	24	1577	13	-1
1.2	106	337	3.18	27	0.000042	0.07	0.2959	0.0047	3.992	0.068	0.0978	0.0006	0.938	1671	24	1583	11	-6
1.3	96	305	3.18	24	0.000068	0.11	0.2876	0.0047	3.825	0.067	0.0965	0.0006	0.933	1629	24	1557	12	-5
1.4	124	436	3.52	31	0.000089	0.14	0.2927	0.0046	3.937	0.066	0.0976	0.0006	0.926	1655	23	1578	12	-5
2.1	69	147	2.14	17	0.000029	0.05	0.2829	0.0050	3.814	0.072	0.0978	0.0007	0.930	1606	25	1582	13	-2
2.2	66	182	2.74	16	0.000173	0.28	0.2715	0.0050	3.655	0.075	0.0977	0.0009	0.892	1548	25	1580	17	2
2.3	71	96	1.34	18	-	<0.01	0.2879	0.0049	3.852	0.070	0.0970	0.0006	0.934	1631	24	1568	12	-4
2.4	67	178	2.66	16	0.000029	0.05	0.2795	0.0050	3.689	0.071	0.0957	0.0007	0.928	1589	25	1543	13	-3
3.1	70	194	2.76	17	0.000202	0.32	0.2753	0.0048	3.681	0.075	0.0970	0.0010	0.861	1568	24	1567	19	0
3.2	97	334	3.44	23	0.000280	0.45	0.2736	0.0044	3.737	0.078	0.0991	0.0013	0.772	1559	22	1607	25	3
3.3	77	264	3.43	18	0.000313	0.50	0.2776	0.0051	3.762	0.083	0.0983	0.0012	0.828	1580	26	1591	23	1
3.4	89	264	2.97	21	0.000117	0.19	0.2685	0.0045	3.597	0.071	0.0972	0.0010	0.847	1533	23	1571	20	2
5.1	79	211	2.66	19	0.000056	0.09	0.2831	0.0048	3.788	0.069	0.0970	0.0007	0.930	1607	24	1568	13	-2
7.1	157	604	3.85	38	0.000072	0.11	0.2851	0.0041	3.870	0.060	0.0985	0.0005	0.937	1617	21	1595	10	-1
8.1	62	180	2.92	15	0.000070	0.11	0.2869	0.0053	3.854	0.077	0.0974	0.0008	0.918	1626	26	1575	15	-3
8.2	61	171	2.81	15	0.000249	0.40	0.2874	0.0052	3.823	0.086	0.0965	0.0013	0.805	1628	26	1557	25	-5
8.3	118	418	3.55	29	0.000045	0.07	0.2903	0.0044	3.871	0.062	0.0967	0.0006	0.935	1643	22	1561	11	-5
1.5	65	236	3.61	16	0.000056	0.09	0.2895	0.0051	3.871	0.073	0.0970	0.0007	0.930	1639	25	1567	13	-5
1.6	106	168	1.58	27	0.000294	0.47	0.2907	0.0046	3.960	0.073	0.0988	0.0009	0.858	1645	23	1602	18	-3
1.7	121	442	3.65	30	0.000789	1.26	0.2859	0.0043	3.847	0.091	0.0976	0.0018	0.633	1621	22	1578	34	-3

Notes: Uncertainties given at the 1σ level; f_{206} % denotes the percentage of ^{206}Pb that is common Pb; correction for common Pb made using the measured $^{204}\text{Pb}/^{206}\text{Pb}$ ratio; for % Disc., 0% denotes a concordant analysis

TABLE A2. Summary of SHRIMP U-Pb Titanite Results for Sample MRD1 838

Area spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	Radiogenic Ratios						Age (Ma)					
							²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	±	ρ	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	% Disc	
1.1	117	368	3.15	29.0	0.006258	10.07	0.2600	0.0046	3.427	0.478	0.0956	0.0132	0.127	1490	24	1540	260	3
1.2	125	436	3.50	31.1	0.000716	1.15	0.2872	0.0047	3.837	0.114	0.0969	0.0024	0.544	1628	23	1565	47	-4
1.3	124	427	3.44	30.3	0.000751	1.20	0.2815	0.0046	3.802	0.093	0.0980	0.0018	0.669	1599	23	1585	34	-1
1.4	120	418	3.48	36.3	0.011391	18.37	0.2872	0.0063	3.752	1.033	0.0948	0.0260	0.079	1627	31	1523	517	-7
2.1	123	428	3.48	30.8	0.001535	2.46	0.2843	0.0042	3.802	0.136	0.0970	0.0032	0.414	1613	21	1567	61	-3
2.2	111	385	3.48	27.4	0.000015	0.02	0.2882	0.0045	3.898	0.065	0.0981	0.0006	0.940	1632	23	1588	11	-3
2.3	119	382	3.21	31.4	0.005184	8.42	0.2811	0.0047	3.555	0.444	0.0917	0.0113	0.135	1597	24	1462	235	-9
3.1	118	143	1.21	29.1	-	<0.01	0.2879	0.0049	3.921	0.074	0.0988	0.0008	0.903	1631	24	1601	15	-2
3.2	157	177	1.13	38.4	0.000302	0.48	0.2838	0.0042	3.806	0.067	0.0973	0.0009	0.844	1610	21	1573	18	-2
4.1	139	176	1.27	33.5	0.000244	0.39	0.2802	0.0043	3.798	0.077	0.0983	0.0013	0.762	1592	22	1592	25	0
4.2	146	118	0.80	31.0	0.006591	10.61	0.2205	0.0040	2.900	0.435	0.0954	0.0142	0.122	1285	21	1535	280	16
5.1	131	441	3.37	29.9	0.001037	1.66	0.2619	0.0040	3.513	0.103	0.0973	0.0024	0.520	1499	20	1573	47	5
6.1	2	13	5.71	0.5	0.000989	1.59	0.2471	0.0210	3.272	0.347	0.0960	0.0061	0.803	1424	109	1548	119	8
7.1	134	30	0.23	31.9	0.000013	0.02	0.2766	0.0042	3.696	0.060	0.0969	0.0005	0.942	1574	21	1566	10	-1
7.2	182	41	0.23	45.0	0.000270	0.43	0.2861	0.0042	3.919	0.068	0.0993	0.0009	0.845	1622	21	1612	17	-1
7.3	205	81	0.39	49.7	0.000034	0.05	0.2822	0.0041	3.780	0.058	0.0971	0.0005	0.947	1603	21	1570	9	-2

Notes: Uncertainties given at the 1σ level; f_{206} % denotes the percentage of ^{206}Pb that is common Pb; correction for common Pb made using the measured $^{204}\text{Pb}/^{206}\text{Pb}$ ratio; for % Disc., 0% denotes a concordant analysis

TABLE A3. Summary of SHRIMP U-Pb Titanite Results for Sample DD86EN26 368

Area spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	<i>f</i> ₂₀₆ %	Radiogenic Ratios						Age (Ma)					
							²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	±	<i>ρ</i>	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	% Disc	
1.1	135	489	3.62	59	0.000438	0.71	0.5023	0.0079	6.499	0.122	0.0938	0.0010	0.835	2624	34	1505	19	-74
1.2	132	483	3.67	33	0.000134	0.21	0.2904	0.0045	3.892	0.067	0.0972	0.0007	0.900	1644	23	1571	14	-5
2.1	151	439	2.91	33	0.000217	0.35	0.2565	0.0041	3.403	0.060	0.0962	0.0007	0.898	1472	21	1552	15	5
2.2	165	463	2.81	39	0.000479	0.78	0.2727	0.0044	3.473	0.078	0.0924	0.0014	0.721	1554	22	1475	30	-5
2.3	156	564	3.62	42	0.000607	0.97	0.3104	0.0047	4.127	0.084	0.0964	0.0013	0.750	1742	23	1556	25	-12
3.1	92	109	1.18	30	0.000379	0.61	0.3771	0.0076	5.076	0.132	0.0976	0.0016	0.776	2063	36	1579	31	-31
3.2	125	338	2.71	41	0.000505	0.81	0.3767	0.0066	4.860	0.105	0.0936	0.0012	0.816	2061	31	1500	24	-37
3.3	135	616	4.57	34	0.000226	0.36	0.2933	0.0051	3.946	0.077	0.0976	0.0009	0.894	1658	26	1578	16	-5
4.1	50	97	1.92	12	0.000380	0.61	0.2697	0.0071	3.562	0.118	0.0958	0.0019	0.799	1539	36	1544	37	0
4.2	47	135	2.86	11	0.000286	0.46	0.2731	0.0071	3.780	0.127	0.1004	0.0021	0.777	1557	36	1631	39	5
5.1	105	352	3.36	26	0.000251	0.40	0.2858	0.0056	3.844	0.095	0.0975	0.0015	0.793	1621	28	1578	28	-3
5.2	75	180	2.40	17	0.000659	1.06	0.2608	0.0061	3.470	0.144	0.0965	0.0033	0.566	1494	31	1557	64	4
6.1	98	283	2.89	24	0.000206	0.33	0.2847	0.0058	3.819	0.093	0.0973	0.0013	0.835	1615	29	1573	25	-3
6.2	101	298	2.94	25	0.000101	0.16	0.2883	0.0058	3.859	0.088	0.0971	0.0010	0.884	1633	29	1569	20	-4
6.3	118	335	2.83	42	0.000408	0.66	0.4118	0.0074	5.438	0.116	0.0958	0.0011	0.844	2223	34	1544	22	-44
6.4	115	324	2.83	28	0.000146	0.23	0.2844	0.0054	3.788	0.079	0.0966	0.0008	0.918	1614	27	1559	15	-3
6.5	157	431	2.75	40	0.000309	0.50	0.2931	0.0049	3.882	0.079	0.0961	0.0011	0.823	1657	25	1549	22	-7
6.6	121	368	3.04	29	0.000220	0.35	0.2758	0.0052	3.689	0.076	0.0970	0.0008	0.915	1570	26	1568	16	0
7.1	137	374	2.72	34	0.000408	0.66	0.2888	0.0056	3.811	0.103	0.0957	0.0018	0.725	1636	28	1542	35	-6
7.1	118	330	2.80	29	0.000201	0.32	0.2812	0.0060	3.776	0.093	0.0974	0.0012	0.864	1598	30	1575	23	-1
7.3	105	298	2.84	26	0.000153	0.24	0.2890	0.0061	3.933	0.092	0.0987	0.0010	0.909	1637	31	1600	18	-2
7.4	101	294	2.91	25	0.000196	0.31	0.2910	0.0063	3.950	0.106	0.0984	0.0016	0.805	1647	31	1595	30	-3

Notes: Uncertainties given at the 1 σ level; *f*₂₀₆ % denotes the percentage of ²⁰⁶Pb that is common Pb; correction for common Pb made using the measured ²⁰⁴Pb/²⁰⁶Pb ratio; for % Disc., 0% denotes a concordant analysis

TABLE A4. $^{40}\text{Ar}/^{39}\text{Ar}$ Isotope Data, Sample TD2, Muscovite ($J = 0.010608$)

Temp (°C)	^{36}Ar mol	^{37}Ar mol	^{39}Ar mol	^{40}Ar mol	$^{40}\text{Ar}^{\circ}$ (%)	$^{40}\text{Ar}^{\circ}/^{39}\text{Ar}_K$	^{39}Ar (%)	Age (Ma) ± 1 s.d.	Ca/K
650	1.89E-16	2.55E-16	6.36E-15	6.05E-13	90.8	86.33	1.0	1173 ± 8	13
700	5.98E-17	7.12E-17	5.68E-15	7.97E-13	97.8	137.35	1.9	1622 ± 16	42
760	6.61E-17	7.12E-17	1.32E-14	1.80E-12	98.9	135.39	3.9	1606 ± 6	97
800	2.73E-17	7.72E-17	1.26E-14	1.69E-12	99.5	133.19	5.8	1589 ± 5	86
840	4.60E-17	5.10E-17	2.79E-14	3.76E-12	99.6	134.46	10.1	1599 ± 6	287
880	1.44E-16	2.48E-16	9.20E-14	1.22E-11	99.6	132.30	24.3	1582 ± 5	196
900	4.40E-17	4.01E-15	4.04E-14	5.32E-12	99.7	131.30	30.6	1574 ± 8	5
925	1.48E-17	7.17E-15	3.95E-14	5.19E-12	99.9	131.40	36.7	1575 ± 6	3
950	5.22E-17	1.60E-16	3.54E-14	4.63E-12	99.6	130.46	42.1	1567 ± 6	116
980	1.14E-17	1.60E-16	3.87E-14	5.09E-12	99.9	131.55	48.1	1576 ± 5	127
1010	6.34E-17	1.34E-15	4.25E-14	5.58E-12	99.6	130.86	54.7	1570 ± 5	17
1040	3.00E-17	1.61E-16	3.18E-14	4.20E-12	99.8	131.90	59.6	1579 ± 7	104
1080	1.17E-16	3.55E-16	1.10E-13	1.44E-11	99.7	131.26	76.5	1574 ± 4	163
1150	1.61E-16	1.16E-15	9.88E-14	1.31E-11	99.6	131.75	91.7	1578 ± 6	45
1220	1.36E-16	8.71E-15	5.14E-14	6.79E-12	99.4	131.42	99.7	1575 ± 6	3
1350	5.77E-17	3.26E-16	2.12E-15	2.79E-13	93.9	123.83	100.0	1513 ± 53	3
Total	1.22E-15	2.43E-14	6.48E-13	8.55E-11		131.33		1574 ± 6	

TABLE A5. $^{40}\text{Ar}/^{39}\text{Ar}$ Isotope Data, Sample 2000366032, Biotite ($J = 0.010651$)

Temp (°C)	^{36}Ar mol	^{37}Ar mol	^{39}Ar mol	^{40}Ar mol	$^{40}\text{Ar}^{\circ}$ (%)	$^{40}\text{Ar}^{\circ}/^{39}\text{Ar}_K$	^{39}Ar (%)	Age (Ma) ± 1 s.d.	K/Ca
650	8.21E-15	2.88E-17	1.62E-14	3.87E-12	37.3	89.17	4.1	1205 ± 35	295.9
700	9.10E-15	1.62E-15	3.89E-14	7.74E-12	65.2	129.91	14.0	1567 ± 12	12.6
730	4.55E-16	1.80E-16	4.79E-14	6.53E-12	97.9	133.57	26.2	1596 ± 5	140.4
760	1.77E-16	3.05E-16	6.35E-14	8.49E-12	99.4	132.80	42.3	1590 ± 8	109.6
800	1.85E-16	1.43E-14	1.09E-13	1.46E-11	99.6	133.43	69.9	1595 ± 5	4.0
830	1.87E-16	7.99E-15	5.95E-14	8.05E-12	99.3	134.20	85.0	1601 ± 8	3.9
860	1.62E-17	5.84E-15	1.15E-14	1.52E-12	99.7	131.70	88.0	1581 ± 14	1.0
900	6.41E-17	4.90E-15	8.70E-15	1.15E-12	98.4	129.55	90.2	1564 ± 10	0.9
940	5.97E-17	1.38E-16	1.22E-14	1.64E-12	98.9	132.68	93.3	1589 ± 11	46.7
980	9.41E-17	6.91E-16	1.30E-14	1.83E-12	98.5	138.50	96.6	1635 ± 9	9.9
1050	5.46E-17	6.17E-16	1.10E-14	1.47E-12	98.9	131.95	99.4	1583 ± 10	9.3
1150	5.84E-17	1.38E-16	2.20E-15	2.18E-13	92.1	91.01	99.9	1223 ± 38	8.4
1350	1.16E-16	1.38E-16	3.22E-16	4.90E-14	30.3	46.15	100.0	721 ± 435	1.2
Total	1.88E-14	3.69E-14	3.94E-13	5.71E-11		130.95		1575 ± 9	

TABLE A6. $^{40}\text{Ar}/^{39}\text{Ar}$ Isotope Data, Sample 2003366054, Biotite ($J = 0.010608$)

Temp (°C)	^{36}Ar mol	^{37}Ar mol	^{39}Ar mol	^{40}Ar mol	% $^{40}\text{Ar}^{\circ}$	$^{40}\text{Ar}^{\circ}/^{39}\text{Ar}_K$	^{39}Ar (%)	Age (Ma) ± 1 s.d.	K/Ca
600	4.29E-16	4.01E-15	5.66E-16	1.43E-13	11.4	28.88	0.4	482 ± 104	0.1
630	2.45E-16	4.53E-15	2.02E-15	1.05E-13	31.4	16.33	1.7	288 ± 21	0.2
700	1.40E-16	3.48E-15	1.73E-15	1.10E-13	62.9	40.08	2.9	639 ± 17	0.3
740	1.56E-16	1.90E-15	1.42E-15	1.72E-13	73.4	89.08	3.8	1200 ± 22	0.4
770	1.23E-16	6.69E-16	2.16E-15	2.85E-13	87.3	114.94	5.3	1438 ± 22	1.7
800	1.93E-16	5.95E-17	3.99E-15	5.49E-13	89.6	123.12	8.0	1507 ± 19	35.3
840	1.13E-16	5.95E-17	1.03E-14	1.35E-12	97.5	127.74	14.9	1545 ± 7	90.9
880	6.36E-17	8.16E-16	1.64E-14	2.20E-12	99.1	133.25	25.9	1590 ± 7	10.5
920	5.09E-18	5.96E-17	1.66E-14	2.20E-12	99.9	132.41	37.0	1583 ± 6	146.8
960	3.20E-17	4.90E-15	1.12E-14	1.49E-12	99.4	132.76	44.5	1586 ± 11	1.2
990	3.12E-17	2.40E-15	5.64E-15	7.62E-13	98.8	133.62	48.3	1592 ± 9	1.2
1020	6.43E-17	5.97E-17	5.10E-15	7.41E-13	97.4	141.60	51.7	1655 ± 11	44.8
1060	8.00E-17	4.19E-15	9.75E-15	1.36E-12	98.3	137.16	58.2	1620 ± 9	1.2
1120	1.29E-16	5.07E-15	3.47E-14	4.71E-12	99.2	134.62	81.5	1600 ± 5	3.6
1300	2.68E-16	4.01E-14	2.76E-14	3.78E-12	98	134.49	100.0	1599 ± 6	0.4
Total	2.07E-15	7.23E-14	1.49E-13	2.00E-11		129.83		1562 ± 8	