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THE GEOLOGY AND GENESIS OF IRON OXIDE-COPPER-GOLD MINERALISATION ASSOCIATED WITH WERNECKE BRECCIA, YUKON, CANADA

VOLUME I

Thesis submitted by
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"It is precisely for this that I love geology. It is infinite and ill defined: like poetry, it immerses itself in mysteries and floats among them without drowning. It does not manage to lay bare the unknown, but it flaps the surrounding veils to and fro; and every so often gleams of light escape and dazzle one's vision."

R. Topfler

Frontispiece: Bonnet Plume River valley, Wernecke Mountains, Yukon, Canada

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ABSTRACT

The large scale Wernecke Breccia system occurs throughout the 13 km-thick Early Proterozoic Wernecke Supergroup (WSG) and is spatially associated with regional-scale faults. Breccia emplacement made use of pre-existing crustal weaknesses and permeable zones; metaevaporitic rocks in the lower WSG may be intimately related to breccia formation. The breccia bodies host vein and disseminated iron oxide-copper-gold ± uranium ± cobalt mineralisation and are associated with extensive sodic and/or potassic metasomatic alteration overprinted by pervasive carbonate alteration. Multiple phases of brecciation, alteration and mineralisation are evident. Six widely spaced breccia bodies that occur in different part of the WSG were examined in this study (i.e. Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor and Olympic). New information includes geological, paragenetic, geochronological, isotopic, fluid inclusion thermometric and compositional data.

Re-Os analyses of molybdenite from a late-stage vein that cross-cuts breccia gave model ages of 1601 ± 6 and 1609 ± 6 Ma. These ages range from older than to within error of the *ca.* 1594.8 ± 4.6 Ma published U-Pb (titanite) date for breccia in the same area. A second molybdenite sample from a late-stage vein gave a Re-Os model age of 1648 ± 5.97 Ma. This date is considered analytically sound but the significance of it is not clear as it is believed to cut the *ca.* 1595 Ma breccia. Step heating ^{40}Ar-^{39}Ar analyses carried out on muscovite from Wernecke Breccia matrix, a syn-breccia vein and two late-stage veins yielded dates of 1178.0 ± 6.1, 1135.0 ± 5.5, 1052 ± 10 and 996.7 ± 8 Ma respectively. These dates are significantly younger than the minimum age (*ca.* 1380 Ma) of Wernecke Breccia indicated by cross-cutting relationships and must have been reset. Samples submitted for U-Pb and Pb-Pb analyses gave discordant results that cannot be used to constrain the age of Wernecke Breccia or Wernecke Supergroup.

Fluids that formed Wernecke Breccia were hot (185-350 ^{o}C), saline (24-42 wt. % NaCl eq.) $NaCl-CaCl_2$ brines. Isotopic compositions for hydrothermal minerals range from: $\delta^{13}C_{carbonate} \approx -7$ to +1 ‰ (PDB), $\delta^{18}O_{carbonate} \approx -2$ and 20 ‰ (SMOW), $\delta^{34}S_{pyrite/chalcopyrite} \approx -13$ to +14 ‰ (CDT) and $\delta^{34}S_{barite} \approx 7$ to 18 ‰. Calculated $\delta^{18}O_{fluid} \approx$ -8 to +14 ‰ The isotopic compositions indicate fluids were likely derived from formation/metamorphic water mixed with variable amounts of organic water ± evolved

meteoric and/or evolved seawater. Metals and sulphur were probably derived from host strata and fluids circulated via tectonic (and/or gravity) processes. Magmatic waters are considered less likely as a fluid source because the isotopic data do not have a magmatic signature and mafic to igneous rocks spatially associated with the breccia are significantly older (i.e. *ca*. 1710 vs. 1600 Ma) thus ruling out a genetic connection. This suggests IOCG mineralisation can occur in non-magmatic environments and a division of the broad IOCG class into magmatic and non-magmatic end-members, with hybrid types in between, is suggested that reflects the involvement of magmatic and non-magmatic fluids. Wernecke Breccia and Redbank are representative of non-magmatic end-members, Lightning Creek is a magmatic end-member and hybrid types include Ernest Henry and Olympic Dam.

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TABLE OF CONTENTS - VOLUME I

SECTION A: Regional-scale Proterozoic iron oxide-copper-gold-mineralised breccia systems: examples from the Wernecke Mountains, Yukon, Canada.

SECTION B: Constraints on the age of Wernecke breccia and associated iron oxide copper-gold mineralisation: new Ar-Ar, U-Pb, Pb-Pb and Re-Os dates.

SECTION C: Stable isotope (C,O,S,D) and fluid inclusion constraints on the origin of Wernecke Breccia and associated iron oxide-copper-gold mineralisation

SECTION D: Wernecke Breccias, Yukon, Canada: an example of a non-magmatic end-member IOCG system and implications for IOCG genesis and classification

TABLE OF CONTENTS - VOLUME II

LIST OF FIGURES

SECTION A

9. Igor area showing location of ‘Igor ridge’, fold axial trace in ‘Igor creek’, Wernecke Breccia, Quartet Group and Cambrian strata.
10. Examples of alteration and mineralization: a) grey sodic-altered breccia, b) red potassic-altered breccia, c) biotite-scapolite alteration in FLG metasiltstone cut by albite veins, cross-cut by calcite vein, d) potassic-altered Quartet Group metasiltstone cut by dolomite-chalcopyrite vein, e) ankerite-magnetite vein cutting FLG, Slab prospect, f) massive chalcopyrite-pyrite vein cutting FLG, Slab prospect, g) clast of massive sulphide in breccia, Slab prospect, h) calcite-chalcopyrite vein cutting FLG, Slab prospect and i) chalcopyrite forming matrix to breccia, Hoover prospect.
11. Composition of a) feldspar and b) carbonate in Wernecke samples, based on microprobe data. Analyses were obtained with the James Cook University electron microprobe in wavelength dispersive mode, at an accelerating voltage of 15 kV and a current of 20 nA.
12. Examples of mineralization and deformation: a) massive magnetite-coarsely crystalline hematite-ankerite-quartz vein, Slats-Frosty area, b) euhedral magnetite replaced by hematite (rl), Igor prospect, c) euhedral magnetite overgrown by pyrite (rl), Igor prospect, d) hematite overgrown by pyrite with chalcopyrite filling fractures (rl), Olympic prospect, e) clasts of foliated metasiltstone in Wernecke Breccia, f) foliated FLG cut by ankerite vein, both have been kinked during Racklan deformation and g) flattened clasts in foliated breccia. rl = reflected light.
13. Simplified paragenesis for prospects in the Wernecke Mountains: a) Slab, b) Hoover, c) Slats-Frosty, d) Slats-Wallbanger, e) Igor and f) Olympic prospects. NB: paragenetic stages apply only to a specific area, e.g. Slab stage 3 ≠ Hoover stage 3 ≠ Igor stage 3.
14. Cartoon depicting evolution of Wernecke Basin and structural controls on location of Wernecke Breccia. Stage I: deposition of FLG. Stage II: deposition of Quartet Group and GLG. Stage III: emplacement of Wernecke Breccia into pre-existing weak zones in deformed and metamorphosed WSG. Breccia bodies are widespread at the top of FLG in metaevaporite-bearing stratigraphy. See text for detailed description of stages.

SECTION B

1. Location of study area, distribution of WSG and Wernecke Breccia (modified from Thorkelson, 2000) plus simplified geology map of the study area (for details see Thorkelson, 2000 and Thorkelson *et al.*, 2002). Legend shows approximate stratigraphic position of IOCG prospects included in this study.
2. Time stratigraphic column of the Wernecke area showing major depositional, intrusive and deformational events (modified from Thorkelson, 2000).
3. Release spectra for ^{40}Ar-^{39}Ar ages for muscovite from four Wernecke Breccia-related samples.
4. ^{207}Pb/^{235}U versus ^{206}Pb/^{238}U plot for three fractions of titanite from a Wernecke Breccia-related vein. Error for Pb/Pb age is quoted at the 2σ level.
5. ^{206}Pb/^{204}Pb versus ^{207}Pb/^{204}Pb plot for 5 analyses of a sample of GLG dolostone.

SECTION C

1. Location of study area, distribution of WSG and Wernecke Breccia plus location of breccia-associated IOCG prospects included in this study (modified from Thorkelson, 2000).
2. Simplified geology map of the study area (for details see Thorkelson, 2000 and Thorkelson *et al.*, 2002). Legend shows approximate stratigraphic position of IOCG prospects included in this study. Slats-F = Slats-Frosty, Slats-W = Slats Wallbanger.
3. Simplified paragenesis for prospects in the Wernecke Mountains: a) Slab, b) Hoover, c) Slats-Frosty, d) Slats-Wallbanger, e) Igor and f) Olympic prospects. NB: paragenetic stages apply only to a specific area, e.g. Slab stage 3 ≠ Hoover stage 3 ≠ Igor stage 3.
4. Photographs of fluid inclusion samples: Slab area - a) large crystals of quartz, calcite and fluorite locally make up the matrix of Wernecke Breccia, b) fluid inclusions in quartz crystal, c) fluid inclusions containing L + V + H and d) fluid inclusion containing L + V + H + h. L=liquid, V=vapour, H=halite, h=hematite; Hoover area – e) metasedimentary clast in Wernecke Breccia, f) trails of secondary fluid inclusions parallel to fractures within quartz in the clast; g) closer view of inclusions in f; Slats-Frosty area - h) ferroan dolomite-pyrite-fluorite vein cutting hematite-altered metasiltstone; the fluorite contains pseudosecondary fluid inclusions, i) fluorite in h), note fluid inclusion trails parallel to fractures that do not

extend beyond the fluorite crystal, j) and k) liquid + vapour fluid inclusions in fluorite; Igor area – l) Wernecke Breccia sample with large barite and magnetite crystals, m) barite crystal in area shown in l); and Olympic area – n) Wernecke Breccia sample used for fluid inclusion analysis, o) quartz grain from breccia matrix with fluid inclusions in the outer rim, p) ferroan dolomite-quartz-chalcopyrite±pyrite vein and q) dolomite with primary fluid inclusions.

5. Summary of fluid inclusion data for samples from the Wernecke Mountains area. Th = final homogenisation temperature; NaCl eq wt % = equivalent weight % NaCl. See Table 1 caption for details of salinity calculations.
6. Comparison of fluid composition for inclusions from the Slab, Hoover, Slats-Frosty, Igor and Olympic areas using the $NaCl$-$CaCl_2$-H_2O system.
7. $\delta^{18}O$ versus $\delta^{13}C$ results for samples of hydrothermal carbonate from the Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor and Olympic areas. Also shown are results for samples of host WSG limestone/dolostone from FLG, Quartet Group and GLG.
8. Overall $\delta^{18}O$ versus $\delta^{13}C$ results for carbonate samples from the Wernecke Mountains. Also shown are fields for common large earth reservoirs that are important in hydrothermal systems. Fields are from Taylor (1974), Sheppard (1977), Graham and Harman (1983), and Hoefs (1987) as compiled in Rollinson (1993). Mean values for Paleoproterozoic carbonates are from Shields and Veizer (2002). FLG = Fairchild Lake Group, Quartet Group = Quartet Group, GLG = Gillespie Lake Group.
9. $\delta^{34}S$ results for samples from the Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor and Olympic areas. See text for discussion.
10. Overall $\delta^{34}S$ results for samples from the Wernecke Mountains. Also shown are fields for common large earth reservoirs that are important in hydrothermal systems. Fields are from Chambers (1982), Kerridge *et al.* (1983) and Chaussidon *et al.* (1989) as compiled in Rollinson (1993).
11. Plot of calculated $\delta^{18}O_{water}$ versus δD_{water} values for mineral separates of biotite, muscovite and actinolite from Wernecke samples. See Appendix VII for sample descriptions. $\delta^{18}O_{water}$ values for biotite and muscovite were calculated using the fractionation equations of Zheng (1993). δD_{water} values for biotite and muscovite were calculated using the fractionation equations of Suzuoki and Epstein (1976).

$\delta^{18}O_{water}$ and δD_{water} values for actinolite were calculated using the fractionation equations of Zheng (1993) and Graham *et al.* (1984) respectively for Tremolite. Magmatic water and formation waters fields are from Taylor (1974). Meteoric water line is from Epstein *et al.* (1965) and Epstein (1970). The metamorphic waters field is from values in Taylor (1974) and Sheppard (1981) as compiled by Rollinson (1993). The fields for felsic magma and high temperature volcanic vapour are from Taylor (1992) and Giggenbach (1992) as shown in Hedenquist *et al.* (1998). Composition of ancient seawater from Sheppard (1986). Isotopic trends are given for: 1) seawater undergoing evaporation (Knauth and Beeunas, 1986), 2) meteoric waters undergoing exchange with ^{18}O in minerals, 3) evaporation of meteoric water and 4) isotopic compositions of Salton Sea and Lanzarote geothermal waters compared to their local meteoric waters (Sheppard, 1986). Black bars beneath the main figure are calculated $\delta^{18}O_{water}$ values for calcite, dolomite and siderite from the Slab, Hoover and Igor areas using the fractionation factors of Zheng (1999).

12. Examples of isochores from the Slab and Igor prospects. Diagrams were constructed using the programme Flinc-Calc that is based on the equations of Zhang and Fratz(1987) and Brown (1998).
13. Log fO_2 – log fS_2 plot for Slab area fluids. The following equations were used to define the mineral stability fields. Log K values were calculated for temperature = 300 °C and pressure = 2500 bars using the programme "The Geochemists Workbench"® release 4.0.2 (GWB uses information from many sources, these are compiled for e.g. in Johnson *et al.*, 1992).

Reaction used	**Equation**	**Log K**
Pyrite-Magnetite:	$3\ FeS_2 + 2\ O_2(g) = Fe_3O_4 + 3\ S_2(g)$	-4.6
Pyrite-Hematite:	$4\ FeS_2 + 3\ O_2(g) = 2\ Fe_2O_3 + 4\ S_2\ (g)$	33.88
Pyrrhotite-Magnetite:	$6\ FeS + 4\ O_2(g) = 2\ Fe_3O_4 + 3\ S_2$	55.34
Bornite-Chalcopyrite:	$Cu_5FeS_4 + 4\ FeS_2 = 5\ CuFeS_2 + S_2$	83.64
Graphite-CO2(g):	$C + O_2(g) = CO_2(g)$	-6.93
Calcite-gypsum:	$2\ CaCO_3 + S_2(g) + 3\ O_2(g) + 4\ H_2O = 2\ CaSO_4 + 2\ CO_2(g)$	36.13

14. Plots of pH versus log fO_2 for the Slab area. a) using a medium value for log aH_2S of -2.6, b) using a low value for log aH_2S of -3.23 and c) using a high value for log aH_2S of -1.97. The positions of $\delta S^{34}{}_{i}$ contours are also shown in a); Numbers in boxes on contours are $\delta S^{34}{}_{pyrite}$ values calculated using $\delta S^{34}{}_{\Sigma S} = 0$ ‰(right side) and

$\delta S^{34}_{\Sigma S}$ = 18 ‰(left side). The shaded oval shows approximate fluid conditions at Slab. The position of sulphur isotope contours were calculated using the method of Ohmoto (1972) and the following conditions: temperature = 300 °C, pressure = 2.5 kb, ionic strength = 3.2 (based on fluid inclusion data). Molality of species was calculated using the programme "The Geochemists Workbench"® release 4.0.2; the following species were most abundant.

Species	**Molality**	**Mole Fraction**
$NaSO_4^-$	0.6985	0.497
$CaSO_4$(aq)	0.3741	0.266
KSO_4^-	0.165	0.117
SO_4^{--}	0.1623	0.115
H_2S(aq)	2.51E-03	0.002
HSO_4^-	1.54E-03	0.001
HS^-	1.30E-03	0.001

SECTION D

1. Location of selected IOCG districts. Modified from Hitzman (2000).
2. Location of Wernecke belt, distribution of Wernecke Breccia and location of breccia-associated IOCG prospects included in this study (modified from Thorkelson, 2000) plus simplified bedrock geology map of the study area (for details see Thorkelson, 2000 and Thorkelson *et al.*, 2002, 2003). Legend shows approximate stratigraphic position of IOCG prospects studied.
3. Typical examples of Wernecke Breccia and associated IOCG mineralisation and alteration: **a)** grey sodic-altered breccia, **b)** red potassic-altered breccia, **c)** breccia with abundant clasts of earlier breccia, Slab area, **d)** photomicrograph of Wernecke Breccia matrix (crossed polars) made up dominantly of sedimentary rock fragments, carbonate, feldspar, lesser quartz and minor hematite and magnetite, **e)** calcite-chalcopyrite vein cutting FLG, Slab prospect, **f)** massive chalcopyrite-pyrite vein cutting FLG, Slab prospect, **g)** chalcopyrite forming matrix to breccia, Hoover prospect, **h)** massive magnetite-coarsely crystalline hematite-ankerite-quartz vein, Slats-Frosty area and **i)** photomicrograph of hematite overgrown by pyrite with chalcopyrite filling fractures (reflected light), Olympic prospect.
4. Fluid temperature and salinity for selected IOCG deposits and prospects. References and abbreviations as in Table 1.

5. Measured sulphur isotope compositions for mineralisation and calculated oxygen isotope compositions for mineralising fluid from selected IOCG deposits. Legend shows mineral(s) used for sulphur isotope analysis and mineral used to calculated oxygen isotopic composition of fluid. References and abbreviations as in Table 1. * Values calculated from actinolite are 5.4 if temperature of 200 ^{o}C is used or 9.5 if temperature of 450 ^{o}C is used (see text for details). For δ^{18}O values for magmatic, metamorphic, formation and meteoric waters and modern seawater – *cf.* Sheppard, 1986; Rollinson, 1993. For δ^{34}S values for mantle and evaporite sources –*cf.* Ohmoto and Goldhaber, 1997.
6. Suggested classification of IOCG systems into magmatic and non-magmatic end-members with hybrid IOCG systems in between. Placement in the classification indicates the degree of involvement of magmatic and/or non-magmatic fluids in the formation of the IOCG system. Placement is also affected by the environment of formation of the IOCG system, i.e., magmatic or non-magmatic, which is determined by whether or not there is a **temporal** association with igneous rocks. At the non-magmatic end of the IOCG spectrum it is possible for an IOCG system to have formed from non-magmatic fluids but to have a temporal relationship with a magmatic system, e.g. Salton Sea. See text for discussion and references.
7. Schematic model showing examples of hydrothermal alteration and mineralisation produced by the circulation of non-magmatic fluids (adapted from Barton and Johnson, 2000). Voluminous metal-depleted, sodic(-calcic) ± shallow K feldspar-hematite alteration forms in inflow zones and along the fluid pathway(s) (*cf.* Barton and Johnson, 1996, 2000). Fluids are heated and as they rise and cool they produce intense sodic and/or potassic alteration (depends on host rock composition) plus overprinting and shallow hydrolytic alteration. Metals are leached along the flow path and precipitate due to cooling and/or fluid mixing.

LIST OF TABLES

SECTION A

2. Representative results of scapolite and biotite microprobe analyses. % meionite = 100 (Ca+Mg+Fe+Mn+Ti) / (Na+K+Ca+Mg+Fe+Mn+Ti); marialite = 100 - meionite (after Deer *et al.*, 1992). Analyses were obtained with the James Cook University JEOL electron microprobe in wavelength dispersive mode, at an accelerating voltage of 15 kV and a current of 20 nA, ZAF corrections were used. [1] average of 70 analyses, [2] average of 10 analyses.
3. Main characteristics of IOCG prospects included in this study, see text for details. In mineralization column: V = vein, D = disseminated, B = forms breccia matrix. [1]Yukon MINFILE (2003) database number. Information from: [2](Thorkelson *et al.*, 2003), [3](Yukon MINFILE, 2003), [4](Stammers, 1995), [5](Eaton & Archer, 1981) and [6](Caulfield, 1994).

SECTION B

1. Summary of published age dates for Wernecke Breccia-related samples, WSG and Slab volcanics.
2. Summary of sample characteristics and new age dates obtained during this study.

SECTION C

1. Summary of fluid inclusion data for samples from the Wernecke Mountains area. Tfm = temperature of first melting, Tm_{ice} = ice melting temperature, Tm_{hh} = hydrohalite melting temperature, Thv = vapour homogenisation temperature, Ths = halite dissolution temperature, Th = final homogenisation temperature. Temperatures in ^{o}C. NaCl eq wt % = equivalent weight % NaCl. NaCl eq wt % values for Slab were approximated using the graphical methods of Vanko *et al.* (1988) and Zwart & Touret (1994). Values for other areas were calculated from Tm_{ice}, $Tm_{hydrohalite}$, Th_{halite} using the programme FlinCalc (J. Cleverley, written communication) which uses information from Zhang and Frantz (1987) and Brown (1998). In the paragenesis column P = primary, S = secondary and PS = pseudo secondary. In the FI (fluid inclusion) Type column L = liquid, V = vapour, H = halite and Op = opaque.
2. Summary of carbon and oxygen isotope results. * mean value for *ca.* 1.8 to 1.7 Ba carbonates is from Shields and Veizer (2002).

3. $\delta^{18}O_{water}$ values calculated from measured $\delta^{18}O$ values of syn-breccia carbonate samples.
4. Summary of sulphur isotope results
5. Summary of hydrogen and oxygen isotope results. Also shown are calculated δD and $\delta^{18}O$ values for co-existing water. #duplicate analysis. See figure 11 caption for calculation details.
6. Estimates of fluid temperature from fluid inclusion and stable isotope data. *All direct fluid inclusion analyses are minimum temperatures, i.e. not trapped during phase separation. ** Using fractionation factors of Sheppard and Schwarcz (1970) and Golyshev *et al.* (1981). *** Using fractionation factors of Ohmoto and Lasaga (1982).
7. Estimates of: 1) thickness of strata overlying the IOCG prospects based on stratigraphic measurements (Delaney, 1981); 2) depth of the prospects based on pressure estimates; 3) pressure from fluid inclusion data; and 4) trapping temperature of fluid (see text for discussion).

SECTION D

1. Size and grade of selected IOCG deposits. References: **Lightning Creek** – Perring *et al.*, 2000; Williams *et al.*, 1999; **Osborne** – Adshead, 1995; Perkins and Wyborn, 1996, 1998; Adshead *et al.*, 1998; Rubenach *et al.*, 2001; **Eloise** – Baker, 1998; Baker and Laing, 1998; Baker *et al.*, 2001; **Olympic Dam** – Roberts and Hudson, 1983, 1984; Creaser, 1989; Reeve *et al.*, 1990; Johnson and Cross, 1991; Oreskes and Einaudi, 1992; Oreskes and Hitzman, 1993; Eldridge and Danti, 1994; Haynes *et al.*, 1995; Reynolds, 2000; **Aitik** – Frietsch *et al.*, 1995, 1997; Carlon, 2000; Wanhainen *et al.*, 2003; **Candelaria** – Ullrich and Clark, 1999; Marschik and Fontboté, 1996, 2001; Marschik *et al.*, 2000; **Salobo** – Requia and Fontboté, 2000; Souza and Vieira, 2000; **Ernest Henry** – Twyerould, 1997; Ryan, 1998; Mark and Crookes, 1999; Mark *et al.*, 2000; Williams *et al.*, in progress; **Wernecke Breccia (Slab)** – Hunt *et al.*, 2004, 2005; **Tennant Creek (West Peko, Eldorado)** – Ahmad *et al.*, 1999; Skirrow and Walshe, 2002; **Redbank** – Orridge and Mason (1975); Knutson *et al.* (1979).
2. Fluid temperature, salinity, composition and source plus precipitation mechanisms for selected IOCG deposits. References and abbreviations as in Table 1.

LIST OF APPENDICES

THESIS INTRODUCTION

This thesis investigates structural, stratigraphic and fluid controls on the location and formation of Wernecke Breccia and associated iron oxide-copper-gold (IOCG) mineralisation. The thesis consists of four sections (A to D) written in journal format that are intended for future publication (Section A has been accepted for publication by Mineralium Deposita). Volume I contains text and Volume II contains figures, tables and appendices. The sections are arranged in a logical progression and each follows on from the previous section. The first three sections present new geological, geochronological and fluid characteristic information for six IOCG prospects in the Wernecke Mountains. The final section compares Wernecke Breccia-related IOCG prospects to other districts and discusses the variability between IOCG systems.

Section A

The first section of the thesis presents a geological introduction to the Wernecke Breccias. This includes a brief summary of major geological units in the region. Descriptions of geology and mineralisation for six IOCG prospects are presented along with evidence for the existence of metaevaporites in Early Proterozoic Wernecke Supergroup strata that are host to the breccia bodies. The section concludes with discussion of the timing and mechanisms of breccia formation.

Section B

The second section focuses on geochronology and constraints on the timing of Wernecke Breccia and associated IOCG mineralisation. New Ar-Ar, U-Pb, Pb-Pb and Re-Os date are presented for samples associated with Wernecke Breccia and mineralisation.

Section C

The third section focuses on fluids that formed Wernecke Breccia and IOCG mineralisation and new fluid inclusion and stable isotope data are presented. The section concludes with a discussion of potential sources for breccia forming and mineralising fluids.

Section D

The final section of the thesis examines the characteristics of IOCG systems and discusses variations between them. Evidence for the non-magmatic genesis of Wernecke Breccia is summarised and modifications to the classification of IOCG deposits that reflect the involvement of magmatic and non-magmatic fluids are suggested. Similarities of end-member magmatic and non-magmatic IOCG types to other types of intrusive- and sediment-related deposits are also discussed.

SECTION A

Regional-scale Proterozoic iron oxide-copper-gold-mineralised breccia systems: examples from the Wernecke Mountains, Yukon, Canada

A.1 Abstract

A large scale Proterozoic breccia system consisting of numerous individual breccia bodies, collectively known as Wernecke Breccia, occurs in north-central Yukon Territory, Canada. Breccias cut Early Proterozoic Wernecke Supergroup sedimentary rocks and occur throughout the approximately 13 km-thick deformed and weakly metamorphosed sequence. Iron oxide-copper-gold ± uranium ± cobalt mineralisation is associated with the breccia bodies and occurs as veins and disseminations within breccia and surrounding rocks and locally forms the breccia matrix. Extensive sodic and potassic metasomatic alteration occurs within and around breccia bodies and is overprinted by pervasive calcite and dolomite/ankerite, and locally siderite, alteration, respectively. Multiple phases of brecciation, alteration and mineralisation are evident. Breccia bodies are spatially associated with regional-scale faults and breccia emplacement made use of pre-existing crustal weaknesses and permeable zones. New evidence indicates the presence of metaevaporitic rocks in lower WSG that may be intimately related to breccia formation. No evidence of breccia-age magmatism has been found to date.

A.2 Introduction

Numerous Proterozoic breccia bodies, collectively known as Wernecke Breccia, occur over large areas in the north-central Yukon Territory of northern Canada (Fig. 1; *cf.* Bell, 1986a,b, Thorkelson, 2000). They are associated with extensive metasomatic alteration and potentially economically significant, but little studied, iron oxide-copper-gold ± uranium ± cobalt (IOCG) mineralisation (*cf.* Bell and Delaney, 1977; Bell, 1978; Archer *et al.*, 1977; Yukon MINFILE, 2003) and represent some of the best preserved examples of Proterozoic-age IOCG mineralisation in North America due to the relatively low grade of metamorphism. The scale of brecciation and alteration is similar to that in other large-scale Proterozoic breccia provinces including those in Australia that host the Ernest Henry (167 Mt @ 1.1% Cu, 0.54 g/t Au; Ryan, 1998) and giant Olympic Dam (2320 Mt @ 1.3% Cu, 0.5 kg/t U_3O_8, 0.5 g/t Au, 2.9 g/t Ag; Reynolds, 2000) deposits. Several authors have drawn connections between the two areas based on similar ages and physical characteristics of the breccias and the possible proximity of ancestral North America to Australia in Proterozoic time (*cf.* Thorkelson *et al.*, 2001a). However, unlike many IOCG districts where brecciation and mineralisation are related to magmatism that acted as a source of fluid(s) (*cf.* Hitzman, 2000; Sillitoe, 2003), and/or provided heat to drive fluid circulation (*cf.* Barton and Johnson, 1996; 2000) a clear relationship with magmatic rocks is not evident for Wernecke Breccias.

New mapping, petrographic and microprobe studies of several Wernecke Breccia-associated IOCG prospects have identified multiple brecciation, alteration and mineralising events. In addition, the studies provide information on structural and stratigraphic features that control the location of breccia and associated alteration and mineralisation and offer clues to the mechanism(s) of large-scale breccia formation.

A.3 Regional geologic setting

Wernecke Breccia bodies and associated IOCG mineralisation occur in Early Proterozoic strata made up of Wernecke Supergroup (WSG), Bonnet Plume River Intrusions (BPRI), and "Slab volcanics[1]" (Figs. 2, 3; *cf.* Gabrielse, 1967; Delaney, 1981, Thorkelson, 2000). The Early Proterozoic rocks are unconformably overlain by Middle Proterozoic Pinguicula Group carbonate and siliciclastic rocks (Fig. 3). The base of WSG is not exposed but is interpreted to sit on ≥ 1.84 Ga crystalline basement that is the westward continuation of the Canadian shield (*cf.* Norris, 1997; Thorkelson, 2000).

A.3.1 Wernecke Supergroup

The Fairchild Lake, Quartet and Gillespie Lake groups make up the WSG and together form an approximately 13 km-thick package of fine-grained marine sedimentary rocks and carbonates (Delaney, 1981) that were deposited pre *ca.* 1710 Ma as two clastic to carbonate grand cycles (*cf.* Thorkelson, 2000). The Fairchild Lake Group (FLG) represents initial subsidence followed by infilling, and the Quartet and Gillespie Lake Groups (GLG) represent subsequent subsidence followed by infilling. The grand cycles may reflect continental rifting and equate to two stages of lithospheric stretching, subsidence and thermal deepening of the basin (Thorkelson, 2000).

Fairchild Lake Group

The Fairchild Lake Group forms the basal part of the WSG and, in the study area, outcrops in a corridor which parallels the Bonnet Plume River valley (Fig. 2; Delaney,

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[1] Many of the names used in this paper are informal at present and are initially shown in quotation marks.

1981; Gordey and Makepeace, 1999; Thorkelson, 2000). It consists of at least 4 km of shallow marine sedimentary rocks that have been divided into five formations named, from base to top, F-1 to F-4 and F-Tr (Fig. 4; Table 1; Delaney, 1981). Contacts between formations are conformable with the exception of F-4 which is considered to be facies equivalent to F-3; the contact between F-3 and F-Tr may, in part, be transitional (Delaney, 1981). The formations are made up of variable amounts of grey-weathering, generally thin-bedded, commonly laminated, siltstone, mudstone, claystone and fine-grained sandstone; minor intercalated carbonate rocks occur in F-2 and F-Tr (*ibid.*).

Metaevaporites

Previously unrecognised metaevaporites were identified in the Slab mountain area in strata that Bell and Delaney (1977) correlate with upper FLG. In general, evaporites dissolve during metamorphism and deformation, however they leave behind indirect evidence that they were once part of the stratigraphy (*cf.* Warren, 1999). This evidence includes metamorphic minerals that are enriched in sodium (e.g. marialitic scapolite and albite) or magnesium (tourmaline and magnesium-rich biotite; *cf.* Kwak, 1977; Warren, 1999). The sodium comes from dissolution of salts such as halite and the magnesium from diagenetic minerals such as dolomite and clays (*cf.* Warren, 1999). However, the presence of a likely metaevaporitic mineral assemblage does not, by itself, indicate an undeniable evaporite protolith because, for example, the NaCl in scapolite could have been derived from an outside source such as hydrothermal or sedimentary brines rather than halite. Further evidence is required from the distribution of the metaevaporitic mineral assemblage. Scapolites derived from hydrothermal or basinal brines tend to form haloes or replacement fronts associated with fluid conduits (Warren, 1999). However, scapolite that is distributed as fine-scale interbedded scapolite-bearing and scapolite free layers is representative of an *in situ* salt precursor (*ibid.*). As evaporite dissolution occurs (during deposition, compaction, and/or during uplift associated with collision and basin inversion) strata overlying or intercalated with the evaporites settle and become fragmented leading to the formation of dissolution breccias. These breccias are commonly preserved in the rock record and are another indication of the former presence of evaporites (*ibid.*). Widespread sodic alteration haloes can also be an indication that evaporites were once part of the stratigraphy. The haloes form in fine-grained sediments adjacent to evaporites during burial, diagenesis and metamorphism due to progressive dissolution of evaporites by circulating fluids (*ibid.*). Resulting fluids are typically saline and chlorine-rich and lead to the growth of metamorphic minerals dominated by sodic phases, thus creating zones of albitisation or scapolitisation as scapolite replaces albite[2].

In the Slab area abundant (< 1 to 5 mm) scapolite occurs in discrete layers < 1 to 30 cm thick, in a sequence of fine-grained metasedimentary rocks at least 100 m thick (Figs. 4c,5). The scapolite is sodium and chlorine-rich with a composition of 69.6 - 77.9 % marialite ($Na_4[Al_3Si_9O_{24}]Cl$) and 22.1 - 30.4 % meionite ($Ca_4[Al_6Si_6O_{24}]CO_3$); Na_2O content ranges from 7.94 - 10.24 weight % (average 9.68 wt.%) and Cl content ranges from 2.90 - 3.55 weight % (average 3.29 wt.%; Table 2). Biotite in this area is magnesium rich with MgO contents from 12.76 – 14.59 weight % (average 13.75 wt. %; Table 2). The occurrence of scapolite within discrete layers and its marialitic composition are consistent with derivation from a protolith that contained abundant halite (*cf.* Warren, 1999). The magnesium-rich nature of biotite associated with the scapolite is also consistent with this interpretation. Probable solution breccia made up of contorted, broken siltstone layers in a carbonate matrix occurs above the scapolite-rich horizon that underlies Slab ridge (Fig. 5). Large zones of sodic metasomatic alteration, typified by albite and marialitic scapolite, occur in fine-grained FLG sedimentary rocks in the Slab area consistent with expected alteration proximal to evaporite-bearing rocks. Taken together, the above indicate *in situ* (halite-facies) evaporites were once present in upper FLG strata.

Quartet Group

The Quartet Group gradationally overlies the FLG and occurs throughout the study area (Fig. 2). It is at least 5 km thick and consists of a monotonous succession of dark grey-weathering clastic rocks that have been divided into basal black carbonaceous shale (Q-1) and conformably overlying, coarsening-upwards, interlayered, shale, siltstone and sandstone (Q-2; Fig. 4 a,d; Table 1; Delaney, 1978, 1981). Q-1 is interpreted to have accumulated in a sediment-starved basin (Delaney, 1981). Q-2 was deposited in a shallow marine (subtidal to intertidal) environment and records a gradually increasing influx of sediment into the basin (*ibid.*).

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[2] 3 Albite + NaCl <—> Marialite i.e. 3 Na(AlSi3O8) + NaCl <—> $Na_4(Al_3Si_9O_{24})Cl$

Gillespie Lake Group

The Gillespie Lake Group gradationally overlies the Quartet Group and forms the upper part of the WSG. It occurs throughout the study area (Fig. 2) and consists of at least 4 km of shallow marine, buff-, orange- and locally grey-weathering dolostone, limestone, claystone, mudstone, and sandstone divided into seven conformable formations named, from base to top, G-Tr and G-2 to G-7 (Fig. 4a; Table 1; Delaney, 1978, 1981). G-Tr consists of siliciclastic sediments that gradually increase in carbonate content up sequence (Delaney, 1981). G-2 to G-4 are made up of fine-grained siliciclastic-dolostone admixtures. G-5 is similar to G-2 to G-4 but contains stromatolitic conglomerate and breccia. G-6 is made up of thick sequences of carbonate and terrigenous-carbonate admixtures. G-7 consists of platformal carbonates some of which are stromatolitic (Fig. 4e).

A.3.2 Igneous rocks

Numerous small Early Proterozoic (*ca*. 1715 to 1705 Ma) igneous bodies, known as Bonnet Plume River Intrusions (BPRI), crosscut the WSG and occur as clasts in Wernecke Breccia (Fig. 3; Thorkelson, 2000; Thorkelson *et al.,* 2001a,b). The intrusions are generally fine- to medium-grained and composed of rift-related tholeiitic diorite or gabbro and lesser syenite and anorthosite (Thorkelson *et al.*, 2001b).

Slab volcanics are preserved as clasts within Wernecke Breccia (*cf.* Thorkelson, 2000). The main exposure is a 160 x 380 m block comprising about 34 steeply dipping, 0.8 to 14 m thick, mafic to intermediate, subaerial flows and minor intercalated volcaniclastic and epiclastic units (*cf.* Laughton *et al.*, 2002; Laughton, 2004). The age of the volcanic rocks is not known. They are geochemically similar to BPRI and may have been comagmatic with them, thus Slab volcanics could be younger than WSG (*cf.* Thorkelson, 2000). Alternatively, Slab volcanics may be part of lower WSG. They have only been observed in Wernecke Breccia emplaced into upper FLG strata in several localities along the Bonnet Plume River valley (Thorkelson, 2000; Laughton *et al.*, 2002; Laughton, 2004; this study). Clasts within Wernecke Breccia appear to be locally derived from proximal host rocks (WSG or BPRI - see below), thus it seems reasonable to assume clasts of Slab volcanics were also locally derived. Therefore, the eruption of Slab volcanics may have been a localized event during lower WSG deposition. The WSG is a dominantly marine succession and Slab volcanics are subaerial however, the presence of metaevaporites within lower WSG rocks indicates very shallow marine to emergent conditions were present at least periodically, thus sub-aerial volcanism could have occurred during deposition of lower WSG strata.

A.4 Metamorphism, deformation, and timing

The Racklan Orogeny produced lower to middle greenschist grade regional metamorphism in WSG rocks (*cf.* Delaney, 1981; Thorkelson, 2000, Brideau *et al.*, 2002). Typical metamorphic mineral assemblages include chlorite-muscovite-chloritoid-quartz±garnet indicating peak metamorphic conditions of 450-550 ^{o}C and approximate pressure of 3-6 kbars (Brideau *et al.*, 2002). Three phases of deformation are ascribed to the Racklan Orogeny (Thorkelson, 2000; Brideau *et al.*, 2002). The first phase occurred during peak metamorphism and produced foliations (ranging from slaty cleavage to schistosity) and north- to east-trending folds (*ibid.*). The second phase produced crenulations and local crenulation cleavage and the third phase led to the formation of kink-bands. During Racklan orogenesis some units, especially those in the upper FLG and Quartet Group, were converted to schist, slate and phyllite in regions of high strain, commonly in the cores, or overturned parts, of tight folds (*cf.* Thorkelson *et al.*, 2003). The age of the Racklan Orogeny is not known, however, it is constrained by the age of Wernecke Breccia that contains clasts of deformed WSG (Fig. 3). In the Slab area, hydrothermal titanite from the matrix of Wernecke Breccia that contains clasts of foliated, kinked metasiltstone returned a U-Pb age of *ca.* 1600 Ma (Thorkelson *et al.,* 2001a), thus, the Racklan Orogeny must be older than *ca.* 1600 Ma.

Cross-cutting relationships indicate both the BPRI and the Slab volcanics are older than Wernecke Breccia, i.e. the breccia contains clasts of these lithologies (Thorkelson, 2000). The difference in age between the igneous rocks and the breccia is poorly constrained due to the unknown age of the Slab volcanics and minimal reliable age information for Wernecke Breccia. However, available data for the breccias indicates they are significantly younger than BPRI, i.e. *ca.* 1600 Ma versus *ca.* 1710 Ma (Thorkelson, 2000; Thorkelson *et al.*, 2001a,b). The timing of magmatism relative to Racklan deformation remains uncertain as there are no documented outcrops of BPRI or Slab volcanics that contain Racklan deformation fabrics (*ibid.*). Both the BPRI and the Slab volcanics have been affected by low grade metamorphism - for example in the BPRI, pyroxene has altered to chlorite or actinolite and plagioclase is replaced by sericite, albite, potassium feldspar or scapolite (Thorkelson *et al.*, 2001b) and in the Slab volcanics plagioclase is commonly replaced by scapolite (Thorkelson, 2000; Laughton *et al.*, 2002; Laughton, 2004). However, some (or all) of this alteration may be due to metasomatism associated with Wernecke Breccia emplacement.

Several post-Racklan orogenic events also affected the study area (Fig. 3). The Middle Proterozoic Corn Creek Orogeny produced west- to southwest-verging folds and thrust faults mainly in the Pinguicula and Hematite Creek groups (Fig. 3; Thorkelson, 2000). The Late Proterozoic Hayhook extensional event produced generally west-trending normal faults some of which truncate folds and faults related to the Corn Creek Orogeny (*ibid.*). Contractional deformation associated with the Cretaceous to Early Tertiary Laramide Orogeny produced large-wavelength, west-northwest-trending folds and reverse faults in WSG and younger strata (Fig. 2; Norris, 1997; Thorkelson *et al.*, 2003). Laramide-age structures are cut by Tertiary normal faults (Thorkelson 2000).

A.5 Geology of Wernecke Breccia

Six widely spaced IOCG prospects were examined in this study (Figs. 1,2): Slab, Hoover, Igor, Slats-Frosty, Slats-Wallbanger and Olympic. The prospects were chosen based on: differences in metasomatic alteration [M. Stammers pers comm. (2000); R. Gorton pers comm. (2000)]; stratigraphic location within host WSG strata (Eaton and Archer, 1981; Thorkelson, 2000); and the availability of diamond drill core. Mapping was carried out at 1:5,000-scale at Slab, Hoover and Igor (Hunt *et al.*, 2002; Thorkelson *et al.*, 2002, 2003); reconnaissance mapping was carried out at the remaining prospects. Diamond drill core from each of the prospects was logged in detail[3] and samples collected for petrographic, microprobe, stable isotope and fluid inclusion analyses. Mapping and core logging were carried out in order to characterize the structural and stratigraphic settings of the breccias and associated alteration and mineralisation, and to provide paragenetic information. This paper presents the results of field, petrographic and microprobe studies. Stable isotope and fluid inclusion data is presented in Section C; a summary is given here. The main characteristics of the prospects are summarised in

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[3] The following number of diamond drill holes were logged in detail at Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor and Olympic respectively: 5, 2, 2, 3, 2 and 3. An additional 8 holes were examined in less detail at Slab.

Table 3 and described below beginning with a general description of Wernecke Breccia based on combined observations from the prospects studied.

Wernecke Breccia varies from clast to matrix supported (Fig. 6a,b; *cf*. Bell, 1986a,b; Thorkelson, 2000; Hunt *et al*, 2002). Clasts are generally sub-angular to sub-rounded and range from < 1 cm to metres to several hundred metres across (e.g. Slab Fig.7). Clasts are locally derived, dominantly from proximal WSG strata, however, clasts of BPRI (e.g. at Slab, Hoover, Olympic) and Slab volcanics (e.g. at Slab) are abundant where breccia cuts those lithologies. Early phases of Wernecke Breccia are preserved in some locations as clasts within later breccia (Fig. 6c). Breccia matrix is made up of rock fragments and hydrothermal precipitates consisting mainly of feldspar (albite and/or potassium feldspar), carbonate (calcite, or dolomite/ankerite, locally siderite) and quartz (Fig. 6d). Locally, the breccia matrix contains abundant hematite, magnetite, chalcopyrite (see mineralisation section), biotite, muscovite barite and fluorite and lesser tourmaline and actinolite, and rare titanite and monazite. In some places the matrix is coarsely crystalline and is made up of quartz, calcite and fluorite crystals up to 2 cm across. Locally, biotite, muscovite and magnetite crystals up to 1 cm across occur within finer grained matrix.

Wernecke Breccia complexes vary greatly in shape and size. In plan view they are elliptical, elongate, or irregular in shape. In vertical section they can be discordant or parallel to layering with no or numerous offshoots. The size of breccia bodies varies from a few centimetres to several hundred metres to several kilometres across. Contacts between country rock and breccia vary from sharp to diffuse/gradational (Fig. 6 e,f). Gradational boundaries with the WSG extend for a few centimetres to several tens of metres and the degree of brecciation gradually decreases outwards from strongly disrupted sedimentary rocks to layers that appear simply “pushed apart” to fractured country rock.

A.6 Wernecke Breccia: structural & stratigraphic controls

A.6.1 Regional scale

Wernecke Breccias are spatially associated with regional-scale faults (*cf*. Bell, 1986a,b; Thorkelson, 2000). In the study area, breccia bodies lie on the southwestern edge of the Richardson Fault array (Fig. 1). This is a series of deep-seated, long-lived structures that developed in a dilational tectonic regime and are traceable for approximately 600 km in an area that marks the transition from relatively undeformed rocks of the Northern Interior Platform to those of the Cordilleran Orogenic System (Delaney, 1981; Norris, 1997; Thorkelson, 2000).

A.6.2 Local scale

Breccia emplacement appears to have exploited pre-existing crustal weaknesses (*cf.* Bell, 1986a,b; Thorkelson, 2000) at all scales including the faulted cores of folds, high strain zones, pathways previously used by the BPRI, jointing/fractures and permeable sedimentary layers. Breccia occurs throughout the WSG but is most widespread in the upper FLG (Delaney, 1981).

The Slab prospect is underlain by upper FLG strata (Bell and Delaney, 1977) that are, in part, made up of metaevaporites (Fig. 5; Table 3). It occurs on the eastern limb of a large northwest-trending anticlinal structure proximal to a flexure in the trend of the fold (Fig. 2) and is cut by a northwest-trending zone of schist and phyllite interpreted to a be a high strain zone (Fig. 7; Thorkelson, 2000; Brideau *et al.* 2002). Minor outcrops of BPRI occur within the strain zone. Extensive Wernecke Breccia occurs as large, elongate, irregular-shaped bodies proximal to the high strain zone, as elliptical pipe-like bodies a few metres in diameter within the strain zone and as narrow bodies parallel to layering in metasedimentary rocks. Cross-cutting relationships indicate there are at least three phases of breccia (Hunt *et al.*, 2002). Breccias in this area contain the largest clasts observed with some up to several hundred metres across and it is one of few locations where breccia contains clasts of Slab volcanics (Fig. 7; *cf.* Thorkelson, 2000; Hunt *et al.*, 2002; Laughton *et al.* 2002; Laughton, 2004). A detailed description of part of the Slab area (Slab creek) is given in Brooks *et al.* (2002) and Brooks (2002).

The Hoover prospect is located about 20 km north of Slab and is underlain dominantly by FLG to Quartet Group transitional rocks including a 4 to 10 m-thick regional marker carbonate unit (Figs. 2, 8; Thorkelson *et al.*, 2002, 2003). Like Slab, Hoover occurs on the eastern limb of the antiform whose axis underlies the Bonnet Plume River valley (*ibid.*). Wernecke Breccia occurs in two locations (Fig. 8; *ibid.*): 1) an irregular-shaped breccia body at least 100 m thick occurs at the base of the slope in strongly folded FLG metasiltstone and phyllite and 2) small, discontinuous breccia bodies 1-20 m thick occur about 500 m upslope in Quartet Group at the contact between slate and metasiltstone. Large (up to 12 x 10 m) clasts of BPRI diorite are abundant locally in the lower breccia. Rare clasts of earlier breccia indicate there have been multiple episodes of brecciation.

The Igor prospect, located about 28 km west of Slab, is underlain by Quartet Group metasedimentary rocks (unit Q-2 of Delaney, 1981) that are faulted against and unconformably overlain by Paleozoic clastic sedimentary rocks and limestone (Figs. 2,9; *cf.* Norris, 1997). Quartet Group rocks have been folded into a north-trending anticline. The core of the anticline is coincident with schistose rocks that likely represent a high strain zone. Abundant Wernecke Breccia occurs in the core of the anticline and cross-cutting relationships indicate several phases of breccia are present. In addition, narrow zones of breccia (< 2 m wide) occur in Quartet Group adjacent to the main breccia complex.

Slats-Frosty is located about 15 km west of Slab and is underlain by the FLG and abundant BPRI dykes and sills (Fig. 2; Thorkelson *et al.*, 2002, 2003). It is located on the east side of a large northeast-trending overturned antiform (Thorkelson, 2000; Thorkelson *et al.*, 2002, 2003). Breccia occurs dominantly as numerous discontinuous narrow bands up to 1.5 m thick emplaced parallel to layering in the FLG, and less commonly as narrow zones (up to 0.8 m) parallel to prominent jointing. The Slats-Wallbanger prospect is located approximately 4 km south of Slats-Frosty and is underlain by Quartet Group-GLG transitional strata (units Q-2 and G-Tr of Delaney, 1981) and abundant BPRI (Thorkelson, 2000). Breccia occurs throughout the Wallbanger area as narrow bands (2 m across) and larger irregular bodies proximal to BPRI rocks. The Olympic prospect, located approximately 15 km southeast of Slab, occurs in the upper part of the WSG and is underlain by GLG dolostone and BPRI (Fig. 2; *cf.* Thorkelson, 2000). The dolostone is in part stromatolitic (Fig. 4e, unit G-5 or G-7 of Delaney, 1981) and intraformational dolostone breccia is locally abundant. Wernecke Breccia occurs as irregular-shaped bodies in the Olympic area that locally contain abundant clasts of BPRI anorthosite, from < 1 cm to tens of metres across (Thorkelson, 2000; this study). Sparse clasts of earlier breccia indicate multiple phases of brecciation have occurred.

A.7 Alteration

Extensive metasomatic alteration occurs within Wernecke Breccia and extends into host WSG and/or BPRI rocks for a few metres to tens of metres (*cf.* Thorkelson, 2000, Hunt *et al.* 2002, 2003a,b). There are two major alteration types, sodic and potassic whose abundance appears to depend on host rock lithology. Sodic alteration generally occurs in breccia hosted by lower WSG rocks where there is evidence for metaevaporites. Breccia and WSG rocks affected by sodic alteration are dominantly grey in colour (Fig. 10a) and contain abundant albite and lesser scapolite. This type of alteration is dominant, for example, at Slab and Hoover where feldspar from breccia clasts, breccia matrix, veins and sedimentary rocks is dominantly albite (Fig. 11a); feldspar at Slab is composed of albite$_{96-100}$, orthoclase$_{0-1}$ and anorthite$_{0-4}$ and at Hoover is composed of albite$_{98-100}$, orthoclase$_{0-1}$ and anorthite$_{0-2}$. Minor orthoclase occurs in both areas (Fig. 11a) and has a narrow compositional range of orthoclase$_{86-99}$, albite$_{0-16}$, and anorthite$_{0-14}$.

Potassic alteration is dominant in breccias hosted by fine-grained clastic rocks i.e. in the Quartet Group, in the FLG where no evidence for metaevaporites was observed, and in the GLG where dolostone is interlayered with fine-grained clastic rocks. Rocks affected by potassic alteration are largely pink to red in colour and contain abundant orthoclase ± sericite (Fig 10b). This type of alteration occurs, for example, at Slats-Frosty and Olympic where feldspar is dominantly orthoclase (Fig. 11a); feldspar at Slats-Frosty is composed of orthoclase$_{94-99}$, albite$_{0-6}$, anorthite$_{0-1}$ and at Olympic is composed of orthoclase$_{88-100}$, albite$_{0-9}$, anorthite$_{0-6}$. Metasomatic alteration at Igor is dominantly potassic consisting of orthoclase and sericite; minor albite is present (Fig. 11a). Feldspar at Igor is composed of orthoclase$_{82-96}$, albite$_{3-18}$, anorthite$_{0-1}$ and albite$_{99-100}$, orthoclase$_{0-1}$, anorthite$_{0-1}$.

At all of the prospects examined, sodic or potassic alteration is overprinted by carbonate that forms the dominant phase in breccia matrix and cross-cutting veins (Fig. 10c,d). The composition of the carbonate also varies. Calcite is dominant in breccias with sodic alteration, for example at Slab and Hoover (Fig. 11b); dolomite occurs in minor amounts. Dolomite and ankerite are dominant in potassically-altered breccias, for example at Slats-Frosty and Olympic. The Igor prospect differs from others in the study in that siderite, in addition to dolomite/ankerite, is locally abundant in breccia matrix and narrow cross-cutting veins (Fig. 11b). Igor is also the only prospect studied where barite is abundant. It occurs as veins, lenses or pods (locally coarsely crystalline) up to 3 x 15 m that cross-cut breccia, and as disseminations within breccia matrix and zones of massive magnetite-hematite (Archer, 1980; Eaton and Archer, 1981; this study).

A.8 Mineralisation

Sixty five Wernecke Breccia-associated prospects are known from the Wernecke and Ogilvie Mountains and all are associated with IOCG mineralisation (Fig. 1, Archer and Schmidt, 1978; Yukon MINFILE, 2003). Mineralisation is similar at each prospect and occurs as veins and disseminations in breccia and surrounding rocks and locally as breccia matrix; multiple episodes of mineralisation are evident (Figs. 1, 10,12; *cf.* Archer and Schmidt, 1978; Yukon MINFILE, 2003; Hunt *et al.*, 2002, 2003a; Brooks *et al.*, 2002). IOCG minerals include magnetite, hematite, chalcopyrite and lesser pitchblende, brannerite and cobaltite; gold was not observed but reports with copper in assay results (*cf.* Archer and Schmidt, 1978; Yukon MINFILE, 2003; Brooks, 2002). General descriptions of mineralisation at the prospects studied are given below. The characteristics of each prospect, including main mineral assemblage and resource or best intersection, are summarised in Table 3.

Multiple phases of oxide and sulphide mineralisation occur at Slab which has a resource of 20 million tons of 0.35 % Cu and 0.17 g/t Au (Thorkelson *et al.*, 2003). Magnetite is dominant in the early stages of brecciation and locally occurs as massive ankerite-magnetite veins up to 1 m across (Fig. 10e; Hunt *et al.*, 2002). These are cross-cut by and occur as clasts in later breccia. Lesser amounts of magnetite occur in later paragenetic stages as disseminated fine-grained blebs and euhedral crystals. Hematite, chalcopyrite and pyrite occur throughout the paragenesis but are most abundant during brecciation following ankerite-magnetite alteration when they occur mainly as breccia matrix and syn-breccia veins (Figs. 10f). Breccia locally contains clasts of massive pyrite-chalcopyrite up to 20 cm across indicating multiple phases of sulphide mineralisation (Fig. 10g). All breccia is cut by veins up to 1 m thick composed dominantly of calcite ± quartz-albite-hematite-magnetite-muscovite-biotite-fluorite. These veins locally contain chalcopyrite (Fig. 10h) and pyrite and rarely molybdenite.

Breccia-associated mineralisation at Hoover is similar to that at Slab and occurs along the northeast side of the Bonnet Plume River valley for over 6 km (Fig. 2); best intersections returned 0.32 g/t Au and 0.42 % Cu over 73 m and 3.6 % Cu over 3.1 m

(Thorkelson *et al.*, 2003, Yukon MINFILE, 2003). Mineralisation is dominantly copper-gold with minor uranium and cobalt and occurs: 1) as disseminations in albite-quartz-pyrite-chalcopyrite veinlets/replacement layers within siltstone clasts in the breccia and as rare massive sulphide clasts, 2) as disseminations and blebs in breccia matrix, locally forming the entire matrix (Fig. 10i), 3) as blebs up to 5 cm across and disseminations in calcite-chlorite-muscovite-pyrite-chalcopyrite-hematite ± magnetite, quartz-hematite-pyrite-chalcopyrite and calcite ± chalcopyrite veins that crosscut breccia and WSG and 4) as blebs and disseminations in quartz-chalcopyrite ± feldspar ± muscovite ± hematite veins that are parallel to and crosscut calcareous layers in siltstone. In general mineralisation occurs as large zones of low-grade copper-gold that contain higher grade pockets (Table 3).

At the Igor prospect, mineralisation occurs largely as pods up to 4 x 15 m of massive hematite-magnetite-pyrite-chalcopyrite and minor pitchblende with dolomite, ankerite, siderite, barite, quartz and chlorite (Archer, 1976, 1980; Eaton and Archer, 1981; Eaton, 1982; this study). The pods occur within breccia in east-trending fractures that are approximately orthogonal to the axis of the large anticline in Igor creek, and in narrow bands parallel to foliation in breccia. Within the pods, hematite, pyrite and/or chalcopyrite locally form/?replace breccia matrix. Less abundant mineralisation occurs as disseminations in breccia and Quartet Group rocks and in quartz ± siderite ± chalcopyrite ± hematite ± pyrite ± magnetite veins, barite veins and dolomite veins that cut the massive mineralisation. Drilling produced encouraging results including a 19.7 m intersection of 4.74% Cu, 0.088% U_3O_8 and 325 ppm Co and a separate 10.6 m intersection of 6.14% Cu, 0.89% U_3O_8 and 358 ppm Co (Eaton and Archer, 1981). Cobalt is associated with pyrite and is unevenly distributed between pyrite grains (Eaton and Archer, 1981). In some instances a cobalt-bearing pyrite grain is surrounded by barren pyrite grains. Uranium mineralisation appears to be related to chalcopyrite and occurs mainly as pitchblende: 1) associated with barite and minor chalcopyrite that discontinuously fills fractures/joints in breccia, 2) as disseminations in zones of chalcopyrite-rich massive magnetite-hematite and 3) as rims on breccia clasts and disseminations in breccia matrix (Archer, 1980). Thorium is a minor constituent; U:Th is 47:1. Uranium is also found in rare brannerite that occurs in fractures in metasedimentary rocks adjacent to breccia (Eaton and Archer, 1981).

At the Slats prospects (Frosty and Wallbanger) copper ± gold ± cobalt mineralisation is widely distributed and low grade with sporadic higher grade occurrences (Table 3). In general, at Frosty, chalcopyrite occurs as disseminations in breccia and in veins and fractures. Pods/veins of massive hematite or massive magnetite-coarsely crystalline hematite-ankerite-quartz up to 2 x 4 m occur locally in breccia zones (Fig. 12a); samples returned low values of copper and gold with the best results (1380 ppb Au and 9650 ppm Cu) being returned by a grab sample (Thorkelson *et al.*, 2003). Quartz vein material containing visible gold and brannerite also occurs in the Frosty area (Yukon MINFILE, 2003). Selected samples returned values of 686 to 10,285 g/t Au (*ibid*), however the source of this material has not been found and it is not known if it is related to Wernecke Breccia. Mineralisation at Wallbanger is similar to that in other areas (Table 3); pyrite locally forms the breccia matrix and a pod of massive magnetite occurs at a contact between breccia and shale; the best intersection returned 450 ppb Au, 1115 ppm Cu and 5800 Co over 1 m (Stammers, 1995). IOCG mineralisation at the Olympic prospect is generally low grade and not abundant; the best intersection returned 1593 ppm Cu, 40 ppm Co and 23 ppb Au over 11 m (Table 3; Caulfield, 1994). The observed mineralisation consists of chalcopyrite and pyrite on fractures at siltstone-breccia contacts, euhedral pyrite + chalcopyrite ± erythrite in ankerite-quartz veins in crackle brecciated siltstone, and sparse chalcopyrite porphyroblasts that overprint breccia matrix and clasts.

A.9 Paragenesis

Broad paragenetic sequences were established for the breccia occurrences based on cross-cutting relationships in outcrop and drill core (Fig. 13; Hunt *et al.*, 2002, 2004). Paragenetic stages are unique to each prospect and not equivalent to those of other areas; for example Slab stage 3 ≠ Hoover stage 3 ≠ Olympic stage 3. Each phase in the development of a breccia complex was probably multistage and overlapped other stages (Delaney, 1981). However, there is an overall general trend of: 1) metasomatic alteration (sodic or potassic) that overprints greenschist facies metamorphic mineral assemblages; 2) early stage brecciation accompanied by abundant magnetite ± hematite alteration; 3) main phase of brecciation accompanied by hematite and chalcopyrite-pyrite ± magnetite mineralisation; and 4) syn to post breccia carbonitization (calcite, ankerite/dolomite, siderite) ± pyrite, chalcopyrite, hematite, magnetite. Locally barite veins are abundant during stage 4 (e.g. at Igor). During the main phase of brecciation early euhedral magnetite is commonly replaced by hematite (Fig 12b) and/or pyrite

(Fig. 12c) and, in general, where chalcopyrite is present with other metallic minerals it is usually the latest mineralisation phase (Fig. 12d).

A.10 Discussion

A.10.1 Timing of breccia emplacement: Cross-cutting relationships demonstrate Wernecke Breccia was emplaced syn- to post-Racklan orogenesis, after peak metamorphism. Sodic and potassic metasomatic alteration associated with Wernecke Breccia overprints greenschist facies metamorphic assemblages and all breccia bodies examined contain clasts of foliated metasedimentary rocks (Fig. 12e) thus indicating that Wernecke Breccia was emplaced after commencement of Racklan orogenesis. Syn-deformation breccia emplacement is indicated by foliated breccia at the Igor prospect that occurs as clasts within cross-cutting breccia that does not contain a fabric (Eaton and Archer, 1981) and by kinked breccia-related ankerite-magnetite and magnetite veins at Slab (Fig. 12f). Post-deformation breccia emplacement is demonstrated by the presence of phyllite clasts in the breccia that contain all three phases of deformation, e.g. Slab (Thorkelson, 2000). Locally, Wernecke Breccia contains clasts of foliated metasiltstone and is itself foliated (Fig. 12g), e.g. at Slats-Wallbanger, however, it is not clear if this deformation is related to Racklan orogenesis or to later deformation.

The absolute age of Wernecke Breccia is constrained by a U-Pb age of *ca.* 1600 Ma that was returned by hydrothermal titanite from a sample of breccia matrix at Slab (Thorkelson, 2000; Thorkelson *et al.*, 2001a); this breccia contains foliated, crenulated, kinked clasts of WSG rocks (*ibid*). Additional age dating including Ar-Ar, Re-Os and U-Pb analyses is presented in Section B.

A.10.2 Mechanism of breccia emplacement: Several hypotheses have been suggested for the mechanism of breccia emplacement including mud diapirs (Lane, 1990), phreatomagmatic explosions (Laznicka and Edwards, 1979), diatremes (Tempelman-Kluit, 1981; Bell and Delaney, 1977), modified evaporite diapirs (Bell, 1989) and explosive expansion of volatile-rich fluids associated with deeply buried intrusions (Thorkelson, 2000; Thorkelson *et al.*, 2001a). Recent studies place constraints on these possible mechanisms (*cf.* Thorkelson 2000 for a review). For example: 1) mud diapirism is ruled out because Wernecke Breccia contains clasts of deformed WSG rocks indicating that the sediments were lithified and deformed prior to brecciation; 2) a diatreme origin seems unlikely because breccia clasts are locally derived; 3) age dating shows that breccia is considerably younger than BPRI (*ca.* 1710 versus *ca.* 1600 Ma) making BPRI magmatism an unlikely mechanism of brecciation; and 4) brecciation caused by the explosive expansion of volatile-rich fluids derived from deeply buried intrusive rocks appears unlikely because stable isotope data do not indicate input from magmatic fluids (see below and Section C).

Any model for the formation of Wernecke Breccia must take into account the following observations: 1) the large-scale of the brecciation and metasomatic alteration; 2) multiple phases of cross-cutting breccia and mineralisation; 3) the spatial association of breccia with regional-scale faults and the occurrence of breccia bodies in weak zones such as cores of folds and faults; 4) the spatial association of breccia with BPRI and the age difference between breccia and BPRI; 5) the emplacement of breccia syn- to post-Racklan deformation and after peak metamorphism; 6) the widespread occurrence of breccia in the upper FLG; 7) the presence of metaevaporites in the upper FLG; 8) the preservation of Slab volcanics as clasts in breccia emplaced into the FLG; 9) the derivation of breccia clasts from proximal host rocks; and 10) the distribution of sodic and potassic metasomatic alteration. Any proposed brecciation mechanism must also take into account characteristics of the breccia-forming fluid. Information available to date is provided by fluid inclusion and stable isotope data (Section C) and is summarised below. Syn-breccia fluids were low to moderate temperature (185 – 350 ^{o}C), high salinity (24-42 wt. % NaCl eq.) $NaCl-CaCl_2-H_2O$ brines. Carbon, oxygen and sulphur isotope results for each of the IOCG prospects studied are similar and, in general, do not vary systematically with paragenetic stage: $\delta^{13}C \approx -7$ to +1 ‰ (V-PDB), $\delta^{18}O \approx 9$ to 20 ‰ (V-SMOW), $\delta^{34}S_{sulphide} \approx -12$ to +13 ‰ (CDT) and $\delta^{34}S_{sulphate} \approx 8$ to 17 ‰ (CDT). The fluid composition and isotopic ratios appear to be rock buffered (*ibid*). The $\delta^{13}C$ values of hydrothermal carbonates indicate the carbon was derived in large part from the host WSG. The $\delta^{34}S$ values of hydrothermal pyrite, chalcopyrite and barite point to a seawater (or sediments/evaporites deposited from seawater) source for sulphur. These data combined with limited hydrogen isotope data indicate the source of fluids was likely formation/metamorphic water mixed with variable amounts of low δD (e.g. organic) water ± evolved meteoric and/or evolved seawater (*ibid.*). Fluid inclusion data also enable the depth of breccia emplacement to be estimated. This was calculated for Slab to be 7.4 to 9.0 km (2.4-3.0 kbars; *ibid.*), thus indicating that breccia at Slab was emplaced at considerable depth and not close to surface as previously suggested (*cf.* Thorkelson, *et al.*, 2001a; Laughton, 2004).

The following proposed brecciation mechanism takes into account the above constraints and relates breccia emplacement to evolution of the Wernecke Basin, as shown in Figure 14.

Stage I: Deposition of the lower WSG in a region underlain by attenuated continental crust (Thorkelson, 2000). This is represented by the FLG calcareous sedimentary rocks and minor limestone that make up the first WSG clastic to carbonate grand cycle (Delaney, 1981; Thorkelson, 2000). Evaporites (halite facies) were deposited in (at least) the upper part of the FLG. Slab volcanics may also have been deposited at this time.

Stage II: Subsidence and thermal deepening of the basin due to rifting, and the subsequent deposition of the middle and upper parts of the WSG (Thorkelson, 2000). The transition is marked by an abrupt change from dolostone to carbonaceous shale at the top of the FLG (Delaney, 1981). Quartet Group fine-grained clastic rocks and the overlying GLG dolostone were deposited at this time and form a second clastic to carbonate grand cycle (Delaney, 1981; Thorkelson; 2000). The accumulation of thick shallow-water carbonates in the upper part of the WSG indicates protracted subsidence of the basin during this time; abrupt facies changes within the GLG indicate normal faulting was coincident with deposition (Thorkelson, 2000). The suite of BPRI were emplaced *ca.* 1710 Ma, after deposition of the WSG. However, their timing with respect to deformation is poorly constrained and they could be pre- or syn-Racklan orogenesis (Thorkelson, 2000; Thorkelson *et al.*, 2001b).

Stage III: Racklan Orogenesis caused greenschist facies metamorphism and three phases of deformation in WSG rocks (Thorkelson, 2000; Brideau *et al.*, 2002). Peak metamorphic conditions (450-550 ^{o}C and 3-6 kbars), folds and foliations were produced during the first phase. The second and third phases produced crenulations/crenulation cleavage and kink bands respectively (*ibid.*). Cross-cutting relationships indicate that Wernecke Breccias were produced syn- to post-deformation, but post peak metamorphism. The length of time occupied by Racklan orogenesis is poorly constrained. If the *ca.* 1710 Ma. BPRI were emplaced before orogenesis then deformation and breccia emplacement could have occurred over a geologically short period of time prior to (close to) *ca.*1600 Ma. However, if BPRI were emplaced syn-orogenesis this implies a period of deformation and breccia production that spanned over 100 Ma years (i.e. *ca.* 1710-1600 Ma).

During stages I and II, as the WSG sediments were undergoing diagenesis and compaction, evaporites within the sedimentary pile would have begun to dissolve. Dissolution would have caused disruption of intercalated and overlying sediments and led to the formation of solution breccias (Fig. 5), and strata overlying evaporites would have been extensively tilted and faulted (*cf.* Warren, 1999). Temperatures would have become elevated in deeper parts of the basin due to the increasing weight of overlying sediments as deposition continued. For example, assuming a geothermal gradient of 25 ^{o}C/km (average geothermal gradient of the earth, *cf.* Raymond (1995) and a surface temperature of 25 ^{o}C the temperature at the top of the FLG (~9 km depth) would have been 250^{o}C (Fig. 14). Greenschist facies metamorphism with temperatures of at least 450 ^{o}C was superimposed on the above scenario along with contractional deformation (Brideau *et al.*, 2002). Wernecke Breccia was therefore produced within WSG rocks that had already been affected by syn-depositional faulting, likely contained zones of disrupted and brecciated strata associated with evaporites, had been intruded by BPRI, had reached elevated temperatures due to burial and metamorphism and had been affected by at least one phase of deformation.

The brecciation mechanism that produced Wernecke Breccia was probably not violently explosive, i.e. not diatreme- or volcanic-like. Clasts are derived from proximal host strata and do not show evidence of transport. Clasts of country rocks are commonly angular to subangular and locally, have clear jigsaw-fit texture (e.g. Fig. 10i); this is especially apparent where breccia cross-cuts BPRI or well-laminated sedimentary rocks. Subrounded to rounded clasts (e.g. Figs. 6c, 10a) and intense zones of alteration also occur, generally in areas where cross-cutting relationships demonstrate that multiple brecciation events have occurred. The amount of fluid was probably not voluminous, as evidenced by the rock buffered nature of alteration and isotopic ratios, except in areas of cross-cutting breccia which likely mark zones of more abundant fluid flow.

This evidence best fits a scenario in which breccias were created by the relatively non-violent expansion of over-pressured basinal fluids. These fluids were locally focussed along permeable pathways, such as faults or shear zones, which led to multiple brecciation events in the same location as pressure repeatedly built up and was released. Basinal fluid would likely have become over-pressured during burial and compaction of sediments, especially in deeper parts of the basin where plastic deformation would be likely to decrease permeability and close porosity leading to an increase in pore fluid pressure (McCaig *et al.*, 2000). It seems probable that pressure within basinal fluids would have been released during initial phases of Racklan deformation. However, in situ evaporites were still present during metamorphism and could have acted as a local seal to fluid flow. In addition, at least at Slab, breccia was emplaced at considerable depth (7-9 km) and strata may have been deforming plastically which would cause over-pressuring of the pore fluid if no permeable pathways such as faults were available. Periodic release of pressure, perhaps by breaking of seals on permeable pathways during continuing deformation, would lead to rapid expansion of fluid and the formation of breccia, particularly in weak, fractured zones such as the cores of folds. In order to maintain repeated brecciation events fluid would have to be added to the system. This may explain why Wernecke Breccia is spatially related to the large-scale Richardson Fault array. These crustal-scale faults could have allowed fluids to enter the basin and recharge the system. The occurrence of widespread breccia bodies in upper FLG strata compared to the rest of the WSG (Delaney, 1981) may be due to the presence of metaevaporites in this part of the stratigraphy. Dissolution of evaporites during diagenesis, compaction and metamorphism would have caused disruption of intercalated and overlying sediments and led to the formation of solution breccias (Warren, 1999) thus providing widespread weak zones that could be utilized by over pressured fluids to produce brecciation. Wernecke Breccias that occur at this stratigraphic level contain the largest clasts observed (100’s of metres across) and this also may be related to the presence of evaporites. Large blocks could initially have formed during evaporite dissolution due to foundering of strata overlying evaporite layers. These blocks could then have become entrained in Wernecke Breccia as fluid utilized the pre-existing weak, and probably permeable zone during breccia emplacement.

A.11 Summary

In summary, zones of Wernecke Breccia occur over a large part of north-central Yukon in areas of regional-scale faulting. They occur throughout host WSG strata but are most widespread in lower WSG where metaevaporites make up part of the stratigraphy. WSG strata and breccia proximal to metaevaporites have been affected by dominantly sodic metasomatism. WSG rocks and breccias distal to evaporites have undergone mainly potassic metasomatism. Sodic and potassic alteration are overprinted by carbonate alteration. IOCG mineralisation is associated with the breccia bodies and occurs as disseminations, veins and breccia matrix. Cross-cutting relationships indicate multiple alteration, brecciation and mineralising events. Breccia emplacement occurred in weak zones of the crust and was probably caused by the expansion of over pressured basinal fluids. Fluids that formed the breccias were low to moderate temperature, high salinity NaCl-$CaCl_2$-H_2O brines likely derived from dominantly basinal sources. There is no evidence for the input of magmatic fluids and the high salinity of the fluid is interpreted to be due to interaction with (halite facies) evaporites (Hunt *et al.*.2004; Section C). Thus, the fluids involved are similar to those proposed by Barton and Johnson (1996, 2000) for the formation of some IOCG deposits.

Wernecke Breccias were emplaced into a thick, dominantly marine sedimentary sequence that was likely deposited in an intra-continental rift basin setting (*cf.* Thorkelson, 2000). Other IOCG districts that formed in a similar tectonic environment and also have large-scale brecciation and metasomatic alteration include those of the Cloncurry district in Australia and the Lufilian arc in central Africa (Hitzman, 2000). Little is known about the IOCG mineralisation in central Africa however the Cloncurry district, has received much exploration attention and several deposits have been discovered including the large Ernest Henry (167 Mt @ 1.1% Cu, 0.54 g/t Au) deposit (*cf.* Ryan, 1998; Williams and Skirrow, 2000; Mark *et al.*, 2000). Breccia-associated IOCG prospects in the Wernecke Mountains contain evidence for multiple cross-cutting brecciation and mineralising events (*cf.* Yukon MINFILE, 2003) but they have received little exploration to date and their full potential remains unknown.

A.12 References:

Archer AR (1976) Report on soil geochemistry, geology and radiometric survey Igor 1-26 claims. Yukon Assessment Report 090083, 8pp.

Archer AR (1980) Drill report Igor 1-26 claims. Yukon Assessment Report 090562, 12pp.

Archer AR and Schmidt U (1978) Mineralised breccias of Early Proterozoic age, Bonnet Plume River District, Yukon Territory. CIM Bulletin, 71:53-58.

Archer A, Bell RT, Delaney GD and Godwin CI (1977).Mineralised breccias of Wernecke Mountains Yukon. Geological Association of Canada Program with Abstracts, 2:5.

Barton MD and Johnson DA (2000) Alternative brine sources for Fe-Oxide(-Cu-Au) systems: implications for hydrothermal alteration and metals. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, Volume 1, PGC Publishing, Adelaide, pp. 43-60.

Barton MD and Johnson DA (1996) Evaporitic-source model for igneous-related Fe oxide-(REE-Cu-Au-U) mineralization. Geology vol. 24, 3:259-262.

Bell RT (1989) A conceptual model for development of megabreccias and associated mineral deposits in Wernecke Mountains, Canada, Copperbelt, Zaire, and Flinders Range, Australia. In: Uranium resources and geology of North America: proceedings of a technical committee meeting, organized by the Internal Atomic Energy Agency held in Saskatoon, Canada, 1987, pp.149-169.

Bell RT (1986a) Megabreccias in northeastern Wernecke Mountains, Yukon Territory. Current Research, Part A, Geological Survey of Canada, Paper 86-1A, pp 375-384.

Bell RT (1986b) Geological map of north-eastern Wernecke Mountains, Yukon Territory. Geological Survey of Canada, Open File 1027.

Bell RT (1978) Breccias and uranium mineralisation in the Wernecke Mountains, Yukon Territory – a progress report. Current Research, Part A, Geological Survey of Canada, Paper 78-1A, pp317-322.

Bell RT and Delaney GD (1977) Geology of some uranium occurrences in Yukon Territory. Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, pp 33-37.

*Brideau M-A, Thorkelson DJ, Godin, L and Laughton JR (2002) Paleoproterozoic deformation of the Racklan Orogeny, Slats Creek (106D/16) and Fairchild Lake (106C/13) map areas, Wernecke Mountains, Yukon. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 65-72.

Brooks M (2002) Alteration, brecciation and Fe oxide-Cu (-Au) mineralisation at Slab creek, Yukon Territory, Canada. Honours thesis, James Cook University, Australia, 113p.

*Brooks M, Baker T and Hunt J (2002) Alteration zonation, veining and mineralisation associated with the Wernecke Breccias at Slab creek, Yukon Territory, Canada. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 249-258.

Caulfield DA (1994) Olympic property diamond drilling logs. Yukon Assessment Report 093222, 40pp.

Deer WA, Howie RA and Zussman J (1992) An introduction to rock forming minerals (second edition), Longman Scientific & Technical, 696pp.

*Delaney GD (1978) A progress report on stratigraphic investigations of the lowermost succession of Proterozoic rocks, northern Wernecke Mountains, Yukon Territory. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File Report EGS 1978-10, 12pp.

Delaney GD (1981) The Mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory; In: Proterozoic Basins of Canada, Geological Survey of Canada, Paper 81-10, pp 1-23.

Eastoe CJ, Long A and Knauth LP (1999) Stable chlorine isotopes in the Palo Duro Basin, Texas: evidence for preservation of Permian evaporite brines. Geochimica Cosmochimica Acta 63:1375-1382.

Eaton WD (1982) Wernecke Joint Venture diamond drill report Igor 1-26 claims. Yukon Assessment Report 091445, 17pp.

Eaton WD and Archer AR (1981) Wernecke Joint Venture drill report Igor 1-26 claims. Yukon Assessment Report 090756, 13pp.

Gabrielse H (1967) Tectonic evolution of the northern Canadian Cordillera. Canadian Journal of Earth Sciences, 4:271-298.

*Gordey SP and Makepeace AJ (1999) Yukon digital geology. , Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File 1999-1 (D). Also available as Geological Survey of Canada Open File D3826.

Hitzman MW (2000) Iron oxide-Cu-Au deposits: what, where, when, and why? In: Porter TM (ed) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 9-25.

*Hunt JA, Laughton JR, Brideau M-A, Thorkelson DJ, Brookes ML and Baker T (2002). New mapping around the Slab iron oxide-copper-gold occurrence, Wernecke Mountains. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 125-138.

Hunt JA, Baker T and Thorkelson DJ (2003a) Basin-scale breccia processes associated with proterozoic iron oxide-copper-gold mineralisation: an example from the Wernecke Mountains, Canada. In: The Geological Society 2003 Fermor Flagship Meeting, World Class Mineral Deposits and Earth Evolution, extended abstracts. Applied Earth Science, Transactions of the Institutions of Mining and Metallurgy: Section B, volume 112, number 2, pp 204-206.

Hunt JA and Baker T (2003b) Wernecke Breccia-associated iron oxide-Cu-Au mineralization, Yukon, Canada. In: Eliopoulos DG *et al.* (eds) Mineral Exploration and Sustainable Development, Proceedings of the seventh biennial SGA meeting, Athens, Greece, volume 2, pp 985-987.

Hunt JA, Baker T, Davidson G, Fallick AE, and Thorkelson DJ (2004) Origin of Wernecke Breccia: results of fluid inclusion and stable isotope analyses. Abstract, GSA Convention, Hobart, Tasmania, February 2004.

Kwak TAP (1977) Scapolite compositional change in metamorphic gradient and its bearing on the identification of metaevaporite sequences. Geological Magazine, 114:343-354.

Lane RA (1990) Geologic setting and petrology of the Proterozoic Ogilvie Mountains Breccia of the Coal Creek inlier, southern Ogilvie Mountains, Yukon Territory. Unpublished MSc thesis, University of British Columbia, Vancouver, Canada, 223p.

Laughton JR (2004) The Proterozoic Slab volcanics of northern Yukon, Canada: megaclasts of a volcanic succession in Proterozoic Wernecke Breccia, and implications for the evolution of northwestern Laurentia. PhD Thesis, Simon Fraser University, Burnaby, British Columbia, Canada.

*Laughton JR, Thorkelson DJ, Brideau M-A and Hunt JA (2002). Paleoproterozoic volcanism and plutonism in the Wernecke Mountains, Yukon. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 139-145.

Laznicka P and Edwards RJ, 1979. Dolores Creek, Yukon – a disseminated copper mineralization in sodic metasomatites. Economic Geology, 74:1352-1370.

Mark G, Oliver NHS, Williams PJ, Valenta RK and Crookes RA (2000) The evolution of the Ernest Henry Fe-oxide-(Cu-Au) hydrothermal system. In: Porter TM (ed) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 123-136.

McCaig AM, Tritlla J and Banks DA (2000) Fluid mixing and recycling during Pyrenean thrusting: evidence from fluid inclusion halogen ratios. Geochimica Cosmochimica Acta 64:19:3395-3412.

Norris DK (1997) Geology and mineral and hydrocarbon potential of northern Yukon Territory and northwestern district of Mackenzie. Geological Survey of Canada, Bulletin 422, 401pp.

Raymond, J. (1995) Petrology: The study of igneous, metamorphic and sedimentary rocks. McGraw-Hill Higher Education, New York, 742p.

Reynolds LJ (2000) Geology of the Olympic Dam Cu-U-Au-Ag-REE deposit. In: Porter TM (ed) Hydrothermal Iron Oxide-Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 93-104.

Ryan (1998) Ernest Henry copper-gold deposit. In: Australasian Institute of Mining and Metallurgy Monograph 22, pp. 759-768.

Sillitoe RH (2003) Iron oxide copper-gold deposits: an Andean view. Mineralium Deposita, 38:787-812.

Stammers MA (1995) 1995 drilling report on the Slats mineral claims. Yukon Assessment Report 093436, 19pp.

*Tempelman-Kluit DJ (1981) Nor, summary of assessment work and description of mineral properties. In: Yukon Geology and Exploration, 1979-1980, Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, pp. 300-301.

*Thorkelson DJ (2000) Geology and mineral occurrences of the Slats Creek, Fairchild Lake and “Dolores Creek” areas, Wernecke Mountains, Yukon Territory. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 10, 73pp.

*Thorkelson DJ, Laughton JR, Hunt JA and Baker T (2003) Geology and mineral occurrences of the Quartet Lakes map area (NTS 106E/1), Wernecke and Mackenzie mountains, Yukon. In: Emond DS and Lewis LL (eds) Yukon Exploration and Geology 2002, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 223-239.

Thorkelson DJ, Laughton JR and Hunt JA (2002) Geological map of Quartet Lakes area (106E/1), Wernecke Mountains, Yukon (1:50,000 scale). Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Geoscience map 2002-2.

Thorkelson DJ, Mortensen JK, Davidson GJ, Creaser RA, Perez WA and Abbott JG (2001a) Early Mesoproterozoic intrusive breccias in Yukon, Canada: The role of hydrothermal systems in reconstructions of North America and Australia. Precambrian Research 111:31-55.

Thorkelson DJ, Mortensen JK, Creaser RA, Davidson GJ and Abbott JG (2001b) Early Proterozoic magmatism in Yukon, Canada: constraints on the evolution of northwestern Laurentia. Canadian Journal of Earth Science 38:1479-1494.

Warren J (1999) Evaporites, their evolution and economics. Blackwell Science Ltd, 438p.

Williams, PJ and Skirrow RG (2000) Overview of iron oxide-copper-gold deposits in the Curnamona Province and Cloncurry District (Eastern Mount Isa block), Australia. In: Porter TM (ed) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 105-122.

*Yukon MINFILE, 2003. Database of Yukon mineral occurrences. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, CD-ROM (2).

* These items can be downloaded from the Yukon Geological Survey website www.geology.gov.yk.ca

SECTION B

Constraints on the age of Wernecke Breccia and associated iron oxide-copper-gold mineralisation: new Ar-Ar, U-Pb, Pb-Pb and Re-Os dates

B.1 Abstract

Muscovite, molybdenite and titanite from Wernecke Breccia matrix and/or associated veins were dated using Ar-Ar, Re-Os and U-Pb analytical techniques in order to better constrain the age of Wernecke Breccia and associated IOCG mineralisation. Dolostone from the upper unit of the Wernecke Supergroup (Gillespie Lake Group) was dated using the Pb-Pb method in an attempt to obtain an absolute age for the supergroup. Step heating ^{40}Ar-^{39}Ar analyses carried out on muscovite from Wernecke Breccia matrix, a syn-breccia vein and two late-stage veins yielded dates of 1178.0 ± 6.1, 1135.0 ± 5.5, 1052 ± 10 and 996.7 ± 8 Ma respectively. These dates are significantly younger than the minimum age (*ca.* 1380 Ma) of Wernecke Breccia indicated by cross-cutting relationships and must have been reset. Resetting is interpreted to be due to thermal heating associated with igneous events at *ca.* 1380 Ma and 1270 Ma and burial beneath Middle Proterozoic strata. The Ar-Ar dates probably reflect cooling ages that were set during uplift, some of which is attributed to the Corn Creek Orogeny. Samples submitted for U-Pb and Pb-Pb analyses gave discordant results that cannot be used to constrain the age of Wernecke Breccia or Wernecke Supergroup. Re-Os analyses of molybdenite from a late-stage vein that cross-cuts breccia gave model ages of 1601 ± 6 and 1609 ± 6 Ma. These ages range from older than to within error of the *ca.* 1594.8 ± 4.6 Ma published U-Pb (titanite) date for breccia in the same area. The late-stage vein also cross-cuts mineralisation, thus, the Re-Os dates also indicate that at least some IOCG mineralisation is ≤ *ca.* 1600 Ma. A second molybdenite sample from a late-stage vein gave a Re-Os model age of 1648 ± 5.97 Ma. This date is considered analytically sound but the significance of it is not clear as it is believed to cut the *ca.* 1595 Ma breccia. The age of breccia emplacement is still considered to be best reflected by the published U-Pb date of *ca.* 1595 Ma and is supported by the *ca.* 1600 Ma Re-Os dates from this study. This is considered a reasonable age for breccia emplacement as it is at least consistent with the broad age range (*ca.* 1710-1380 Ma) defined by cross-cutting relationships. However, in detail the age of the breccia and IOCG mineralisation remains poorly constrained; this may be a reflection of the multi-phase nature of the breccia events.

B.2 Introduction

Numerous breccia bodies, collectively known as Wernecke Breccia, are located in the northern Yukon Territory, Canada (Fig. 1; *cf.* Bell, 1986a,b, Thorkelson, 2000). Extensive iron oxide-copper-gold ± uranium ± cobalt (IOCG) mineralisation occurs within the breccia zones and adjacent host rocks and has been the focus of periodic mineral exploration over the past 50 years (*cf.* Yukon MINFILE, 2003). The Wernecke Breccias are similar to breccia-associated IOCG occurrences elsewhere in the world and they have been compared to those in the Cloncurry and Gawler districts of Australia which host the large Ernest Henry and Olympic Dam deposits (*cf.* Bell and Delaney, 1977; Bell, 1978; Archer *et al.*, 1977; Hitzman *et al.*, 1992). The similarity of Wernecke Breccias and their host Early Proterozoic strata to those of comparable age in Australia has been used in arguments to support plate reconstructions for the Proterozoic (*cf.* Thorkelson *et al.*, 2001a,b). One of the ties used in this argument is the parallel between the *ca.* 1595 Ma age proposed for Wernecke Breccia (Thorkelson *et al.*, 2001a) and the *ca.* 1590 Ma age (Reeve *et al.*, 1990; Johnson and Cross, 1995) of the Olympic Dam deposit. However, the age of Wernecke Breccia is not well constrained. Cross-cutting relationships indicate an age between *ca.* 1710 and 1380 Ma and a U-Pb (titanite) date indicates breccia was emplaced into the lower part of the host Wernecke Supergroup (WSG) *ca.*1595 Ma (Thorkelson, 2000; Thorkelson *et al.*, 2001a,b). No reliable age dates are available for breccia bodies hosted in the upper parts of the WSG and the age of the IOCG mineralisation is not known.

This paper reports new geochronological data for Wernecke Breccia and associated IOCG mineralisation. Ar-Ar dates were obtained from: a) muscovite in the matrix of Wernecke Breccia which is hosted in the middle part of WSG; b) muscovite associated with the main stage of mineralisation; and c) muscovite associated with late-stage carbonate veins and alteration that contain minor mineralisation. U-Pb (titanite) and Re-Os (molybdenite) dates were also obtained for the late-stage veins. A Pb-Pb date was obtained from dolostone in the Gillespie Lake Group (GLG) which forms the upper part of the WSG.

B.3 Geologic framework

Wernecke Breccia occurs in areas underlain by Early Proterozoic rocks (Fig. 1) made up of the WSG, Bonnet Plume River Intrusions (BPRI), and “Slab volcanics[1]” (*cf.* Gabrielse, 1967; Delaney, 1981; Bell, 1986b; Thorkelson, 2000). The WSG is an approximately 13 km-thick package of marine sedimentary rocks that has been divided into the Fairchild Lake, Quartet and Gillespie Lake Groups (Figs. 1 & 2; *cf.* Delaney, 1981; Thorkelson, 2000). WSG strata are cut by *ca.* 1710 Ma mafic to intermediate dykes and sills that are part of the BPRI suite (Fig. 2; Thorkelson, 2000; Thorkelson *et al.*, 2001b). Intermediate composition Slab volcanics also occur locally in this region but their timing is uncertain as they have been observed only as clasts within Wernecke Breccia (*cf.* Thorkelson, 2000; Thorkelson *et al.*, 2001b). Strata of the WSG were metamorphosed to greenschist facies and multiply deformed during the Racklan Orogeny (*cf.* Thorkelson, 2000; Brideau *et al.*, 2002). The orogeny is Early Proterozoic but its exact age is unknown. Multiply deformed clasts occur in the breccia body that was dated at *ca.* 1595 Ma, therefore the age of the Racklan Orogeny is constrained to pre (to syn) *ca.* 1595 Ma (Thorkelson, 2000; Thorkelson *et al.*, 2001a,b).

Bodies of Wernecke Breccia, from 0.1 to 10 km^2 in size, occur throughout the WSG and are most abundant in the lower part of the stratigraphy (*cf.* Delaney, 1981; Lane, 1990). The breccia is made up largely of clasts of the supergroup in a matrix of rock flour and hydrothermal precipitates (*cf.* Hunt *et al.*, 2005). Where the breccia bodies cross-cut BPRI or Slab volcanics they contain abundant clasts of these lithologies (*cf.* Thorkelson, 2000).

At least 65 bodies of Wernecke Breccia are known and all are associated with IOCG mineralisation (*cf.* Yukon MINFILE, 2003). Mineralisation occurs as disseminations and veins within the breccia and surrounding WSG rocks and is largely made up of early magnetite and/or hematite and later chalcopyrite and pyrite (*cf.* Brookes *et al.*, 2002; Hunt *et al.*, 2002, 2005; Thorkelson *et al.*, 2003; Yukon MINFILE, 2003). Minor uranium (pitchblende, brannerite) and cobalt (cobaltian pyrite, erythrite) mineralisation occurs locally. Gangue is dominantly composed of carbonate (calcite, dolomite, siderite), quartz, albite and K-feldspar with lesser biotite, muscovite, chlorite and fluorite (*cf.* Brookes *et al.*, 2002; Hunt *et al.*, 2002, 2005). Extensive sodic or potassic metasomatic alteration is spatially associated with the breccia bodies and cross-cutting relationships demonstrate that multiple phases of brecciation and mineralisation have occurred. The alteration and breccia zones are cut by late-stage carbonate veins that contain minor pyrite and chalcopyrite and trace molybdenite (Hunt *et al.*, 2002, 2005).

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[1] Many of the names used in this paper are informal and are initially shown in quotation marks.

B.3.1 Age constraints on Wernecke Breccia

Cross-cutting relationships

Cross-cutting relationships provide constraints on the age of Wernecke Breccia and limit it to between 1710 and 1380 Ma (*cf.* Thorkelson, 2000; Thorkelson *et al.* 2001a,b). Breccia zones cross-cut WSG, BPRI and Slab volcanics. Attempts at dating the WSG and Slab volcanics have not been successful and their ages are poorly constrained (*cf.* Thorkelson, 2000). However, BPRI have been dated via the U-Pb (zircon) method; ages of *ca*. 1710 Ma were obtained (Thorkelson *et al.*, 2001b). Wernecke Breccia bodies locally contain abundant clasts of Bonnet Plume River Intrusions and must therefore be younger than *ca*. 1710 Ma (Thorkelson, 2000; Thorkelson *et al.*, 2001b). The WSG, BPRI, Slab volcanics and Wernecke Breccia are unconformably overlain by the Pinguicula Group (Fig. 2; *cf.* Thorkelson, 2000). The age of the Pinguicula Group is poorly defined however, the base of this group is cut by a diorite stock that yielded a U-Pb (zircon) date of *ca*. 1380 Ma (Thorkelson, 2000). Hence, Wernecke Breccia must be older than 1380 Ma.

Radiometric dates

A U-Pb date of 1594.8 ± 4.6 Ma was obtained from hydrothermal titanite in the matrix of Wernecke Breccia hosted by the Fairchild Lake Group (FLG) in the lower part of the WSG at the Slab prospect (Thorkelson, 2000; Thorkelson *et al.*, 2001a). This date is considered to be a reliable estimate of the age of breccia emplacement at this stratigraphic level (Fig. 1; Thorkelson, 2000; Thorkelson *et al.*, 2001a). Archer *et al.* (1977) report a K-Ar age of 1500 Ma for biotite from a breccia complex at Quartet Mountain which cuts strata at a similar stratigraphic level (Fig. 1). There are a limited number of other published radiometric dates for Wernecke Breccia (Table 1) but most do not appear to represent the time of breccia emplacement. Archer and Schmidt (1978) and Archer *et al.* (1986) report U-Pb dates ranging from approximately 1250 to 400 Ma determined from U-bearing minerals (brannerite, pitchblende) and whole rock samples. These dates are highly discordant which renders them difficult to interpret and they are considered unreliable estimates of the ages of brecciation and hydrothermal activity (Parrish and Bell, 1987). A U-Pb date of 1270 ± 40 Ma was calculated for monazite from a sample of brannerite-hematite-monazite from the Nor breccia body (Parrish and

Bell, 1987). The uranium concentrations in the monazite were unusually low (*ibid*) and it is not clear what effect this had on the analysis. In addition, the Nor property is located about 130 km north of the main belt of Wernecke Breccia occurrences (Fig. 1) and its relationship to the majority of Wernecke Breccia is not known.

B.4 Radiometric Dating

Samples were selected for geochronology (Table 2) with the aims of dating: 1) emplacement of Wernecke Breccia (Ar-Ar), 2) the deposition of main stage mineralisation (Ar-Ar), 3) the formation of late-stage carbonate alteration that hosts minor mineralisation (Ar-Ar, U-Pb, Re-Os) and 4) upper WSG strata (Pb-Pb).

B.4.1 $^{40}Ar/^{39}Ar$ (Muscovite)

Samples for Ar-Ar analysis were chosen from the Igor and Slab prospects (Fig. 1) because these two areas host rocks that contain abundant muscovite. At the Igor prospect hydrothermal muscovite occurs in the matrix of Wernecke Breccia that was emplaced into the Quartet Group, the middle part of the WSG. Thus, dating the muscovite would provide an age for breccia emplacement into middle WSG strata. At the Slab prospect muscovite occurs in the selvages of a 10 cm-thick vein of massive pyrite-chalcopyrite that cross-cuts FLG strata adjacent to Wernecke Breccia (Section A, Fig. 10f). Dating this muscovite would constrain the age of mineralisation. Muscovite also occurs in late-stage calcite-quartz-albite-molybdenite-chalcopyrite veins at the Slab prospect (e.g. Section A, Fig. 10h). Dating this muscovite would provide an age for the late-stage carbonate alteration.

Ar-Ar analysis was carried out at the Geochronology Section of the Geological Survey of Canada in Ottawa. Dating was done using the $^{40}Ar/^{39}Ar$ step heating technique. Analytical protocols and mineral sample preparation methods are provided in Appendix II. Plateau ages are presented for three of the samples and an inverse isochron diagram for the remaining one (M. Villeneuve, pers. comm., 2004).

Muscovite from Wernecke Breccia matrix at the Igor prospect (sample JH02-12-1) yielded a somewhat noisy spectrum. However, degassing did produce an internally consistent plateau that gives an age of 1078 ± 6.1 Ma (Fig. 3a, Table 2). Excess noise is likely the result of slight under-irradiation of the sample which resulted in increased sensitivity to applied corrections (M. Villeneuve, pers. comm., 2004). Muscovite from the selvages of the pyrite-chalcopyrite vein at the Slab prospect (JH02-14-1a) produced a well defined multi-step plateau that gives an age determination of 1135 ± 5.5 Ma (Fig. 3b, Table 2). Muscovite from a late-stage calcite-quartz-albite-molybdenite-chalcopyrite vein that cross-cuts breccia at the Slab prospect (sample SB94-1-345.1 m) yielded a complex Ar-Ar spectrum (Fig. 3c, Table 2). Low temperature steps suggest Ar-loss and possible degassing of a contaminating phase (M. Villeneuve, pers. comm., 2004). Mid-temperature steps produced a weakly defined plateau and an age of 996.7 ± 8.0 Ma was interpreted from these steps (Fig. 3c). The fusion step appears to give an older age and may be a second contaminating phase containing excess ^{40}Ar. Muscovite from a second late-stage vein at the Slab prospect (sample JH01-27-3D, Table 2) also yielded complex results and the plot shown in Figure 3d represents mid-temperature steps from replicate analyses on two aliquots with a clear excess ^{40}Ar component (^{40}Ar/^{36}Ar = 4244 ± 894, MSWD = 3.5). However, the sample appears to give a reasonable and reproducible age of 1052 ± 10 Ma from the key heating steps (M. Villeneuve, pers. comm., 2004). Low temperature steps for this sample appear to have variable Ar-loss. The highest temperature steps (of both aliquots) showed distinctly older ages suggesting contamination by a phase that does not degas until high temperatures (M. Villeneuve, pers. comm., 2004).

B.4.2 U-Pb (Titanite)

A calcite-biotite-titanite vein at the Slab prospect (drill core sample SB94-1-6.9 m; Fig. 1, Table 2) was chosen for U-Pb dating. The vein cuts biotite-altered FLG metasiltstone proximal to Wernecke Breccia and is probably related to late-stage carbonate alteration of the breccia. Thus, dating hydrothermal titanite in the vein will provide a date for the carbonate alteration and place a constraint of the timing of breccia formation.

U-Pb dating of the sample was carried out in the Geochronology Laboratory at The University of British Columbia by isotope dilution thermal ionization mass spectrometry. Sample preparation methods and analytical protocols followed those outlined in Friedman *et al.* (2001) and Mortensen *et al.* (1995), as summarised in Appendix II. Three titanite fractions were analysed and all gave discordant results, which may be due to Pb loss (Fig. 4; J.K. Mortenson pers. comm., 2004). Fraction T1 is from the coarsest, turbid, straw coloured grains. Results from this fraction have relatively poor precision due to very low U (~2 ppm) and Pb (~0.4 ppm) contents.

Fractions T2 and T3 consist of yellow, clear subhedral grains. T2 is coarser than T3 and has a U content of 8 ppm. Fraction T3 has 30 ppm U and gave the most precise results. A ^{207}Pb/^{206}Pb date of 1113 ± 5 Ma was calculated for this fraction and is considered a minimum age for the titanite (J.K. Mortenson pers. comm., 2004).

B.4.3 Pb-Pb (Carbonate)

A sample of stromatolitic GLG dolostone was chosen for Pb-Pb analysis to try and better constrain the age of the WSG. Pb-Pb analysis was carried out at the University of Melbourne using the sample preparation and mass spectrometry procedures outlined in Woodhead and Hergt (1997), as summarised in Appendix II.

Five samples from a hand specimen of dolostone (sample JH02-21-1, Table 2; Section A, Fig. 4e) were analysed and the results are shown in Figure 5. Two dissolution residues were also analysed, however, the residue data do not appear to be in equilibrium with the samples, i.e. they do not fall on the same isochron, and may be a detrital component (J Woodhead, pers. comm., 2003). They have been discarded from the calculations. A conventional isochron calculation gives an age of 1368 ± 160 Ma.

B.4.4 Re-Os (Molybdenite)

Two molybdenum-bearing late stage calcite-quartz-albite-chalcopyrite-muscovite veins that cross-cut breccia at the Slab prospect were chosen for Re-Os analysis (samples SB94-1-345.1 m & JH01-27-3D; Fig. 1). Dates from the molybdenite would provide an age for the carbonate alteration and an upper limit on the timing of breccia formation (Table 2). These samples were also submitted for Ar-Ar analysis (see Ar-Ar section above).

Re-Os isotope analyses were carried out in collaboration with Robert Creaser and Dave Selby at the Radiogenic Isotope Facility of the Department of Earth and Atmospheric Sciences, University of Alberta, Canada. Molybdenite was analysed using analytical procedures for chemical separation of Re and Os modified from methods described by Shirey and Walker (1995), Cohen and Waters (1996) and Birck *et al.* (1997) as described by Selby and Creaser (2001; Appendix II). Re and Os isotope ratios were measured using negative thermal ionization mass spectrometry (Creaser *et al.*, 1991; Völkening *et al.*, 1991) on a Micromass Sector 54 mass spectrometer using Faraday collectors.

Two analyses of molybdenite in sample SB94-1-345.1 m yielded model ages of 1601 ± 6 Ma and 1609 ± 6 Ma (R. Creaser, pers. comm, 2003). Analysis of molybdenite in sample JH01-27-3D gave a date of 1648.1 ± 5.97 Ma (R Creaser, pers. comm., 2002).

B.5 Discussion

The timing of breccia emplacement into the lower WSG (FLG) is constrained by a U-Pb date of 1594.8 ± 4.6 Ma yielded by titanite obtained from the matrix of breccia at the Slab prospect (Thorkelson, 2000; Thorkelson *et al.*, 2001a). Molybdenite from late-stage calcite veins, also from the Slab prospect, yielded Re-Os model ages of 1609 ± 6 and 1601 ± 6 Ma. The veins cross-cut breccia and were expected to have ages younger than the *ca*. 1595 Ma date, instead they range from older than to within error of the above U-Pb date. One explanation for this discrepancy may be the multiphase nature of the breccia. It is possible that the breccia body that was observed to be cut by the late-stage veins is an older phase of breccia and not the same as the breccia body that was dated using U-Pb techniques. The Re-Os dates also provide constraints on the age of IOCG mineralisation associated with Wernecke Breccia. The main phase of mineralisation occurs within breccia and is cross-cut by the late-stage calcite veins, thus, the mineralisation must be ≤ *ca*. 1600 Ma. However, like the breccia, the mineralisation is multi-phase and there may be more than one age of mineralisation. A sample of molybdenite from a second late-stage vein that cross-cuts breccia gave a Re-Os model age of 1648 ± 5.97 Ma. This date is considered analytically sound (D Selby, pers. comm., 2003) but the significance of it is not clear as the host vein is believed to cut the breccia that was dated at *ca*. 1595 Ma. As mentioned above, one explanation for this discrepancy may be the multiphase nature of the breccia, i.e. the breccia body that was observed to be cut by the late-stage veins may be an older phase of breccia.

As mentioned earlier, cross-cutting relationships constrain the age of Wernecke Breccia to lie between *ca.* 1710 and *ca.* 1380 Ma (Thorkelson, 2000). Breccia-related muscovite samples in this study yielded dates of 1178.0 ± 6.1, 1135.0 ± 5.5, 1052 ± 10 and 996.7 ± 8.0 Ma (Table 2), all of which are considerably younger than 1380 Ma (Table 2). The muscovite-bearing rocks are paragenetically well constrained by cross-cutting relationships and are clearly related to Wernecke Breccia, thus the Ar-Ar dates must have been reset. These results are similar to reset ^{40}Ar/^{39}Ar dates of 980 ± 4 and

788 ± 8 Ma yielded by the step heating of white mica from schist in the Early Proterozoic FLG that is cross-cut by Wernecke Breccia (Thorkelson, 2000; Thorkelson *et al.*, in review).

Resetting of the isotope systematics is interpreted to result from reheating of the rocks by igneous events and burial of the strata during the deposition of Middle Proterozoic rocks (Thorkelson, 2000; Thorkelson *et al.*, in review). The geological history of the Wernecke area is incompletely known and several gaps of 200 to 300 Ma exist for which no rock record has been observed (Fig. 2). There are documented igneous events at *ca.* 1380 and *ca.*1270 Ma (*cf.* Thorkelson, 2000; Thorkelson *et al.*, 2001a,b) which could have been responsible for resetting the Ar isotopic systematics. The event at *ca.* 1380 produced the Hart River mafic intrusions 25 to 50 km south of the study area (*cf.* Green, 1972; Abbott, 1997). Rutile in pre *ca.* 1595 Ma Slab volcanics yielded a U-Pb age of *ca.* 1383 Ma and is considered to have crystallised at this time (Thorkelson, 2000; Thorkelson *et al.*, 2001a). Igneous activity at *ca.* 1270 Ma produced the mafic to intermediate composition Bear River dykes in the Wernecke area (*ibid.*) These dykes may be part of the large *ca.* 1270 Ma igneous event that includes the Mackenzie dyke swarm, the Muskox intrusion and the Coppermine lavas in the Northwest Territories and adjacent parts of the Canadian Shield (*cf.* Thorkelson *et al.*, 2001a; in review). Increased temperatures caused by the burial of WSG and Wernecke Breccia beneath Middle Proterozoic strata could also have disturbed the isotope systematics (*ibid*). Middle Proterozoic strata include the approximately 3 km-thick Pinguicula Group made up of marine sedimentary rocks and the up to 1 km-thick carbonate and clastic rocks of the Hematite Creek Group (Fig. 2).

The Ar-Ar dates of *ca.* 980 ± 4 and 788 ± 8 Ma calculated for white mica in FLG were interpreted to reflect cooling ages that were set during uplift caused by the Corn Creek orogeny (<1000 & > 750 Ma), a contractional deformational event characterised by west- to southwest-verging thrust faults (Fig. 2; Thorkelson, 2000; Thorkelson *et al.*, 2001a; in review). Ar-Ar dates of *ca.* 1180 to 1000 Ma calculated for muscovite from breccia related samples in this study are also interpreted to reflect cooling ages. Two of the dates (996.7 ± 8 and 1052 ± 10 Ma) are close to the older age limit of the Corn Creek orogeny and may reflect uplift related to this deformation event. The other two dates are considerably older (1135.0 ± 5.5 and 1178.0 ± 6.1 Ma) and do not correlate to a known orogenic event. However, there is some evidence of uplift during this period in the Wernecke area (*cf.* Thorkelson, 2000) and they may reflect a more localised deformational event or possibly a more significant, but presently unrecognised event.

Lead isotopic analyses were carried out on a breccia-related vein sample and on GLG dolostone. The calculated minimum U-Pb age for titanite from a breccia-related late-stage vein at the Slab prospect (Fig. 4) is 1113 ± 5 Ma. This date, like the Ar-Ar (muscovite) dates discussed above, is considerably younger than the minimum age of 1380 Ma indicated for Wernecke Breccia. The results for the titanite fractions were all discordant and the calculated age is based on the most precise fraction (T3 in Fig. 4). It is considered a minimum age only as this fraction is quite discordant and has likely undergone Pb loss (J.K. Mortensen pers. comm., 2004). Therefore, this date cannot be used to constrain the age of Wernecke Breccia. A conventional isochron calculated for Pb-Pb data yielded by the sample of GLG stromatolitic dolostone gives a date of 1368 ± 160 Ma (Fig. 5). However, BPRI dykes which cut WSG have reported U-Pb (zircon) dates of 1705.9 ± 0.7, 1709.4 ± 1.4, 1711.1 ± 5.1 and 1713.6 ± 12.7 Ma (Thorkelson *et al.*, 2001a) therefore, 1368 ± 160 Ma cannot represent the age of deposition of GLG and may reflect a later diagenetic/metamorphic event.

B.6 Conclusions

New Re-Os (molybdenite) analyses from a late-stage calcite vein that cross-cuts Wernecke Breccia gave model ages of *ca.* 1601 and 1609 Ma. These ages are slightly older than to within error of the published U-Pb (titanite) date of *ca.* 1595 Ma for breccia emplacement and may be a reflection of the multi-phase nature of breccia emplacement. The dates do confirm that at least some breccia and IOCG mineralisation was emplaced at or before *ca.* 1600 Ma.

Step heating ^{40}Ar-^{39}Ar analyses carried out on muscovite from Wernecke Breccia matrix and associated veins yielded dates of *ca.* 1178 to 997 Ma. These dates are significantly younger than the lower limit of the age range (*ca.* 1710-1380 Ma) for Wernecke Breccia indicated by cross cutting relationships, and must have been reset. The Ar-Ar dates are similar to those calculated for white mica in Early Proterozoic FLG that have also been reset. Resetting is interpreted to be due to thermal heating associated with igneous events at *ca.* 1380 Ma and 1270 Ma that produced the Hart River sills and Bear River intrusions respectively (*cf.* Thorkelson *et al.*, in review). Burial of WSG and Wernecke Breccia by Middle Proterozoic strata may also have added to the thermal heating. The Ar-Ar dates probably reflect cooling ages that were set during uplift, some of which is attributed to the Corn Creek Orogeny. Samples submitted for U-Pb and Pb-Pb analyses gave discordant results that cannot be used to constrain the age of Wernecke Breccia or WSG.

This study was unable to further constrain the age of Wernecke Breccia and the age of breccia emplacement is still considered to be best reflected by the published U-Pb date of *ca*. 1595 Ma (Thorkelson, 2001a). Support for this date was provided by the *ca*.1600 Ma Re-Os dates obtained in this study. This is believed to be a reasonable age for breccia emplacement as it is at least consistent with the broad age range (*ca*. 1710-1380 Ma) defined by cross-cutting relationships. The exact age of the breccia and IOCG mineralisation remains poorly constrained and may be a reflection of the multiple nature of the breccia events.

B.7 References:

*Abbott JG (1997) Geology of the upper Hart River area, eastern Ogilvie Mountains, Yukon Territory (116A/10, 116A/11). Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 9, 73pp.

Archer A and Schmidt U (1978) Mineralised breccias of early Proterozoic age, Bonnet Plume River district, Yukon Territory, Canadian Mining and Metallurgy Bulletin 71:53-58.

Archer A, Bell RT, Delaney GD and Godwin CI (1977).Mineralized breccias of Wernecke Mountains Yukon. Geological Association of Canada Program with Abstracts, 2:5.

Archer A, Bell RT and Thorpe RI (1986). Age relationships from U-Th-Pb isotope studies of uranium mineralisation in Wernecke Breccias, Yukon Territory. Current Research, Part A, Geological Survey of Canada Paper 86-1A:385-391.

Bell RT (1986a) Megabreccias in northeastern Wernecke Mountains, Yukon Territory. Current Research, Part A, Geological Survey of Canada, Paper 86-1A, pp 375-384.

Bell RT (1986b) Geological map of north-eastern Wernecke Mountains, Yukon Territory. Geological Survey of Canada, Open File 1027.

Bell RT (1978) Breccias and uranium mineralisation in the Wernecke Mountains, Yukon Territory – a progress report. Current Research, Part A, Geological Survey of Canada, Paper 78-1A, pp317-322.

Bell RT and Delaney GD (1977) Geology of some uranium occurrences in Yukon Territory. Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, pp 33-37.

Birck JL, Roy Barman M and Capmas F (1997) Re-Os isotopic measurements at the femtomole level in natural samples. Geostandards Newsletter – Journal of Geostandards and Geoanalysis, 20:19-27.

*Brideau M-A, Thorkelson DJ, Godin, L and Laughton JR (2002) Paleoproterozoic deformation of the Racklan Orogeny, Slats Creek (106D/16) and Fairchild Lake (106C/13) map areas, Wernecke Mountains, Yukon. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 65-72.

*Brooks M, Baker T and Hunt J (2002) Alteration zonation, veining and mineralisation associated with the Wernecke Breccias at Slab creek, Yukon Territory, Canada. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 249-258.

Cohen As and Waters FG (1996) Separation of osmium from geological materials by solvent extraction for analysis by thermal ionization mass spectrometry. Analytica Chimica Acta 332:269-275.

Creaser RA, Papanastassiou DA and Wasserburg GJ (1991) Negative thermal ion mass spectrometry of osmium, rhenium and iridium. Geochimica et Cosmochimica Acta 55:397-404.

Delaney GD (1981) The Mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory; In: Proterozoic Basins of Canada, Geological Survey of Canada, Paper 81-10, pp 1-23.

Gabrielse H (1967) Tectonic evolution of the northern Canadian Cordillera. Canadian Journal of Earth Sciences, 4:271-298.

Green H (1972) Geology of Nash Creek, Larsen Creek and Dawson Creek map areas, Yukon Territory. Geological Survey of Canada, Memoir 364, 157pp.

*Hunt JA, Laughton JR, Brideau M-A, Thorkelson DJ, Brookes ML and Baker T (2002) New mapping around the Slab iron oxide-copper-gold occurrence, Wernecke Mountains. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 125-138.

Hunt, JA, Baker, T and Thorkelson, DJ (2005, in press). Regional-scale Proterozoic IOCG-mineralised breccia systems: examples from the Wernecke Mountains, Yukon, Canada. Mineralium Deposita, X:x:-x.

Johnson JP and Cross KC (1995) U-Pb geochronological constraints on the genesis of the Olympic Dam Cu-U-Au-Ag deposit, South Australia. Economic Geology 90:1046-1063.

Lane RA (1990) Geologic setting and petrology of the Proterozoic Ogilvie Mountains breccia of the Coal Creek Inlier, southern Ogilvie Mountains, Yukon Territory. MSc. Thesis, University of British Columbia, pp. 223.

Mortensen JK, Ghosh D and Ferri F (1995) U-Pb age constraints of intrusive rocks associated with copper-gold porphyry deposits in the Canadian Cordillera. In: Schroeter TG (ed.) Porphyry deposits of the northwestern Cordillera of North America. Canadian Institute of Mining and Metallurgy, Special volume 46:142-158.

Parrish RR and Bell RT (1987) Age of the Nor breccia pipe, Wernecke Supergroup, Yukon Territory. In: Radiogenic Age and Isotopic Studies: Report 1. Geological Survey of Canada, Paper 87-2, pp. 39-42.

Reeve JS, Cross KC, Smith RN and Oreskes N (1990) Olympic Dam copper-uranium-gold-silver deposit. In: Hughes FE (ed.) Geology of the Mineral Deposits of Australian and Papua New Guinea, Australasian Institute of Mining and Metallurgy, Monograph 14, pp. 1009-1035.

Selby D and Creaser RA (2001) Re-Os geochronology and systematics in molybdenite from the Endako porphyry molybdenum deposit, British Columbia, Canada. Economic Geology 96:197-204.

Shirey SB and Walker RJ (1995) Carius tube digestion for low blank Re-Os analysis. Analytical Chemistry 67:2136-2141.

*Thorkelson DJ (2000) Geology and mineral occurrences of the Slats Creek, Fairchild Lake and “Dolores Creek” areas, Wernecke Mountains, Yukon Territory. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 10, 73pp.

Thorkelson DJ, Mortensen JK, Davidson GJ, Creaser RA, Perez WA and Abbott JG (2001a) Early Mesoproterozoic intrusive breccias in Yukon, Canada: the role of hydrothermal systems in reconstructions of North America and Australia. Precambrian Research 111:31-55.

Thorkelson DJ, Mortensen JK, Creaser RA, Davidson GJ and Abbott JG (2001b) Early Proterozoic magmatism in Yukon, Canada: constraints on the evolution of northwestern Laurentia. Canadian Journal of Earth Science 38:1479-1494.

Thorkelson DJ, Laughton JR and Hunt JA (2002) Geological map of Quartet Lakes area (106E/1), Wernecke Mountains, Yukon (1:50,000 scale). Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Geoscience map 2002-2.

*Thorkelson DJ, Laughton JR, Hunt JA and Baker T (2003) Geology and mineral occurrences of the Quartet Lakes map area (NTS 106E/1), Wernecke and Mackenzie mountains, Yukon. In: Emond DS and Lewis LL (eds) Yukon Exploration and Geology 2002, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 223-239.

Thorkelson DJ, Abbott JG, Mortensen JK, Creaser RA, Villeneuve ME, McNicoll VJ and Layer PW (in review). Early and Middle Proterozoic evolution of Yukon, Canada. Canadian Journal of Earth Science: SNORCLE Special Issue.

Völkening J, Walczyk T and Heumann KG (1991) Osmium isotope ratio determinations by negative thermal ion mass spectrometry. International Journal of Mass Spectrometry Ion Processes 105:147-159.

Woodhead JD and Hergt JM (1997) Application of the “double spike” technique to Pb-isotope geochronology. Chemical Geology (Isotope Geoscience):138:311-321.

Woodhead JD, Hergt JM and Simonson BM (1998) Isotopic dating of an Archean bolide impact horizon, Hamersley basin, Western Australia. Geology, 26:1:47-50.

Yukon MINFILE, 2003. Database of Yukon mineral occurrences. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, CD-ROM (2).

* Can be downloaded from www.geology.gov.yk.ca

SECTION C

Stable isotope (C,O,S,H) and fluid inclusion constraints on the origin of Wernecke Breccia and associated iron oxide-copper-gold mineralisation

C.1 Abstract

Numerous Proterozoic breccia bodies, collectively known as Wernecke Breccia, occur in the north-central Yukon Territory, Canada. The breccias cut Early Proterozoic Wernecke Supergroup (WSG) sedimentary rocks, Bonnet Plume River Intrusions and Slab volcanics. The breccia zones are associated with widespread albite and/or potassium feldspar alteration and are cross-cut by carbonate veins and alteration. Iron oxide (Cu-U-Au-REE) (IOCG) mineralisation occurs within breccia zones and adjacent rocks as disseminations and veins. Multiple phases of brecciation and alteration are evident. Six widely spaced IOCG prospects (Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor & Olympic) that occur in different parts of WSG were studied. Fluid inclusion data indicate syn-breccia fluids were high salinity (24-42 wt. % NaCl eq.) $NaCl\text{-}CaCl_2\text{-}H_2O$ brines with low to moderate temperatures (185 – 350°C). Fluid composition appears to echo that of the host strata, e.g. at the Slab prospect which is hosted by evaporite-bearing Fairchild Lake Group (FLG) strata the fluids are dominated by Na and at the Olympic prospect which is hosted by Gillespie Lake Group (GLG) dolostone the fluids are dominated by Ca. Estimates of fluid pressure determined from fluid inclusion data vary from 0.4 to 2.4 kb and are in reasonable agreement with those based on the approximate thickness of overlying strata. Carbon and oxygen isotope data for hydrothermal carbonates from the prospects range from $\delta^{13}C \approx -11$ to +1.5 ‰ (PDB) and $\delta^{18}O \approx -2$ to 20 ‰ (V-SMOW). Results for host WSG limestone/dolostone range from: FLG $\delta^{13}C \approx -2$ to 0.5 ‰, $\delta^{18}O \approx 12$ to 14 ‰; Quartet Group $\delta^{13}C \approx -2$ to -0.3 ‰, $\delta^{18}O \approx 14$ to 16 ‰; and GLG $\delta^{13}C \approx -2$ to 1.6 ‰, $\delta^{18}O \approx 16$ to 25 ‰. Sulphur isotope date for hydrothermal sulphides (pyrite and chalcopyrite) and barite from the prospects vary from $\delta^{34}S_{sulphide} \approx -12$ to +13 ‰(CD T) and $\delta^{34}S_{sulphate} \approx 8$ to 17 ‰. Deuterium and oxygen isotope data for syn-breccia hydrothermal biotite, muscovite and actinolite range from $\delta D = -141$ to -18 ‰(V -SMOW) and $\delta^{18}O \approx 7$ to 12 ‰ (V-SMOW). $\delta^{18}O$ values for hydrothermal carbonates appear to reflect those of the host limestone/dolostone. $\delta^{13}C$ values indicate the carbon in hydrothermal carbonates was derived in large part from the host WSG. The $\delta^{34}S$ values of hydrothermal pyrite, chalcopyrite and barite point to a largely seawater (or sediments/evaporites deposited from seawater) source for sulphur. These data combined with the limited deuterium isotope data indicate the source of fluids was likely formation/metamorphic water mixed with variable amounts of organic water ± evolved meteoric and/or evolved

seawater. Fluid flow may have been driven by tectonic and/or gravity processes and metals may have been scavenged from host strata.

C.2 Introduction

Numerous Proterozoic breccia bodies, collectively known as Wernecke Breccia, occur over large areas in the north-central Yukon Territory, Canada (Fig. 1; *cf.* Bell, 1986a,b, Thorkelson, 2000). They are associated with extensive metasomatic alteration and significant, but little studied, iron oxide-copper-gold ± uranium ± cobalt (IOCG) mineralisation (*cf.* Bell and Delaney, 1977; Bell, 1978; Archer *et al.*, 1977; Yukon MINFILE, 2003) that due to the low grade of metamorphism represents some of the best preserved examples of Proterozoic-age IOCG mineralisation in North America. The scale of brecciation and alteration is similar to that in other large scale Proterozoic breccia provinces, including those in Australia that host the Ernest Henry (167 Mt @ 1.1% Cu, 0.54 g/t Au; Ryan, 1998) and giant Olympic Dam (2320 Mt @ 1.3% Cu, 0.4g/t U_3O_8, 0.5 g/t Au, 2.9 g/t Ag; Reynolds, 2000) deposits. However, although many IOCG districts display brecciation and mineralisation that is intimately related to magmatism (*cf.* Hitzman, 2000; Sillitoe, 2003), a clear relationship with magmatic rocks is not evident in the Wernecke Mountains.

One of the key aims of this study was to determine if magmatic fluids were responsible for forming IOCG mineralisation and related breccias in the Wernecke Mountains. Six IOCG prospects (Slab, Hoover, Slats-Frosty, Slats-Wallbanger, Igor & Olympic) were selected for analysis to test this hypothesis. The prospects were chosen based on their varied stratigraphic locations within host Wernecke Supergroup (WSG) strata, their association with extensive sodic or potassic metasomatic alteration, and the accessibility of drill core for study (Fig. 2). This paper presents the results of stable isotope (C, O, S and D) and fluid inclusion analyses for the six prospects and discusses how these results constrain possible fluid sources.

C.3 Regional geologic setting

Bodies of Proterozoic Wernecke Breccia, from 0.1 to 10 km^2 in size, occur over large areas of the north-central Yukon Territory that are underlain by Early Proterozoic rocks made up of WSG, Bonnet Plume River Intrusions, and "Slab volcanics[1]" (Figs. 1, 2; *cf.* Gabrielse, 1967; Delaney, 1981; Bell, 1986b; Thorkelson, 2000). The WSG is a thick package of marine sedimentary rocks that have been divided, from base to top, into the Fairchild Lake, Quartet and Gillespie Lake Groups (Fig. 2; *cf.* Delaney, 1981; Thorkelson, 2000). The Fairchild Lake Group (FLG) is at least 4 km thick and made up dominantly of fine-grained sandstone and siltstone, and minor limestone (Delaney, 1981; Thorkelson, 2000). These rocks are overlain by the Quartet Group, an approximately 5 km-thick package of carbonaceous to calcareous fine-grained siltstone and sandstone. Transitionally overlying the Quartet Group, and forming the upper part of the WSG, is the approximately 4 km-thick Gillespie Lake Group (GLG) made up largely of dolostone. Mafic to intermediate dykes and sills of the Bonnet Plume River Intrusions (BPRI), *ca.* 1710 Ma., cut WSG (Thorkelson, 2000; Thorkelson *et al.*, 2001a). Intermediate composition Slab volcanics were erupted prior to emplacement of Wernecke Breccia; however, they are preserved only as clasts within the breccia and it is not clear if they are a local unit within WSG or a separate unit that overlies it (Thorkelson, 2000; Section A). Host rocks to Wernecke Breccia underwent deformation and greenschist facies metamorphism during the Early Proterozoic Racklan Orogeny (Thorkelson, 2000).

Wernecke Breccia is confined to areas of WSG and is made up largely of clasts of the supergroup in a matrix of rock flour and hydrothermal precipitates; locally the breccia contains abundant clasts of BPRI and Slab volcanics (Thorkelson, 2000; Hunt *et al.*, 2002, 2005; Section A). Iron oxide-copper-gold (IOCG) mineralisation occurs as disseminations and veins within the breccia and surrounding WSG rocks (*cf.* Yukon MINFILE, 2003; Hunt *et al.*, 2002, 2005). Extensive sodic and potassic metasomatic alteration is spatially associated with the breccias (*cf.* Yukon MINFILE, 2003; Thorkelson, 2000; Section A). Multiple phases of brecciation and mineralisation are evident (*cf.* Hunt *et al.*, 2002, 2005). Cross-cutting relationships indicate syn- to post-deformational emplacement of the breccias (*cf.* Thorkelson, 2000; Hunt *et al.*, 2005).

Wernecke Breccia bodies examined in this study were emplaced into various parts of the WSG stratigraphy (Fig. 2). In the Slab area, breccia occurs in the upper part of the FLG within a sequence of calcareous metasiltstone, minor limestone and probable metaevaporites, cut by BPRI; numerous and large clasts of Slab volcanics occur locally

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[1] Many of the names used in this paper are informal and are initially shown in quotation marks.

in the breccia (*cf.* Delaney, 1981; Thorkelson, 2000; Hunt *et al.*, 2005). In the Hoover area, breccia bodies are in calcareous metasiltstone and carbonaceous shale/slate at the transition from FLG to Quartet Group; minor BPRI dykes cut the WSG stratigraphy (*ibid.*). In the Slats region, breccia occurs in two areas known informally as 'Frosty' and 'Wallbanger'. Wallbanger is located about 4 km south-southeast of Frosty (Fig. 2). Breccia occurs within upper FLG calcareous metasiltstone and phyllite in the Frosty area and within transitional Quartet to GLG interlayered calcareous metasiltstone, shale and dolostone in the Wallbanger area (*ibid*). Abundant BPRI dykes cut the WSG in both areas. At Igor, breccia occurs within calcareous to carbonaceous metasiltstone and shale of the Quartet Group (*cf.* Norris, 1997). Breccias in the Olympic area were emplaced into locally stromatolitic GLG dolostone in the upper part of the WSG (*cf.* Thorkelson, 2000). In this area minor BPRI diorite dykes cut the GLG and abundant clasts of BPRI anorthosite occur locally in the breccia (*ibid.*). The age of brecciation is considered by Thorkelson (2000) and Thorkelson *et al.* (2001b) to be *ca*. 1595 Ma based on a U-Pb date of 1595 +8/-5 Ma for hydrothermal titanite from the matrix of breccia in the Slab area.

Each phase in the development of a breccia complex was probably multistage and overlapped other stages (Delaney, 1981). However, broad paragenetic sequences, summarized in Figure 3, were established for the breccia occurrences based on cross-cutting relationships. Paragenetic stages are unique to each prospect and not equivalent to those of other areas; for example Slab stage 3 ≠ Hoover stage 3 ≠ Olympic stage 3. In general there is an overall trend of: 1) metasomatic alteration (sodic or potassic) that overprints greenschist facies metamorphic mineral assemblages; 2) early stage brecciation accompanied by abundant magnetite ± hematite alteration; 3) main phase of brecciation accompanied by hematite and chalcopyrite-pyrite ± magnetite mineralisation; and 4) syn to post breccia carbonitization (calcite, ankerite/dolomite, siderite) ± pyrite, chalcopyrite, hematite, magnetite. Locally barite veins are abundant during stage 4 (e.g. at Igor).

C.4 Fluid inclusion studies

Very limited published fluid inclusion data (Hitzman *et al.*, 1992 – Igor prospect) exist for Wernecke Breccia and a preliminary study was undertaken in order to evaluate fluid processes and conditions during the formation of breccia and mineralisation. Further work on Wernecke Breccia fluid inclusions is underway (*cf.* Gillen *et al.*, 2004) at James Cook University, Australia.

Fluid inclusion petrography was carried out on quartz-, calcite-, fluorite- or barite-bearing samples from Wernecke breccia or associated veins at each prospect to identify well preserved inclusions that likely represent syn-breccia fluid. Fluid inclusion paragenesis was established using the criteria for primary, pseudosecondary, and secondary inclusions outlined in Roedder (1984). Three types of inclusions were identified on the basis of phases present at room temperature: 1) liquid plus vapour ± opaque (L+V); 2) L+V plus halite (L+V+H); and 3) L+V+H plus opaque. Fluid inclusion microthermometry was performed on a Linkam, gas-flow heating/freezing stage at James Cook University. The stage was regularly calibrated using synthetic fluid inclusion standards having known phase transitions at -56.6^{o}C and 0.0^{o}C. During calibrations, the temperatures indicated by the thermocouple were within 0.2^{o}C of the standards. Paired heating-freezing microthermometry was carried out on the same inclusions. In each sample all freezing experiments were carried out before the sample was heated. Results of the fluid inclusion analyses are summarized in Table 1 and Figures 4, 5 and 6; complete results are in Appendix III.

C.4.1 Results

Slab: A 1.5 cm-diameter euhedral quartz crystal from the matrix of Wernecke Breccia was selected for microthermometry in the Slab area (Fig. 4a). The quartz contains clusters of fluid inclusions that are probably primary and therefore syn-breccia, i.e. Slab paragenetic stage 3 (Fig. 3a). At room temperature the inclusions contain liquid (70-90%), vapour (5-10%) and halite (Fig. 4b,c). Rarely, they contain a red mineral that is probably hematite (Fig. 4d). During freezing experiments the formation of brown ice was observed in all inclusions and initial melting temperatures (Tfm) were below -50 ^{o}C (Table 1). Final ice melting temperatures (Tm_{ice}) ranged from -50 to -36 ^{o}C. In heating experiments homogenization occurred via halite dissolution (Th = Ths) between 226 and 245 ^{o}C; homogenization of the vapour bubble (Thv) occurred between 154 and 175 ^{o}C (Fig. 5, Table 1). The estimated salinity of fluid in the inclusions is about 42 weight % NaCl equivalent and the Na:Ca ratio is approximately 1.4 (Table 1, Figs. 5, 6).

Hoover: A metasedimentary clast from Wernecke Breccia that contains 1 mm diameter quartz grains was selected for microthermometry in the Hoover area (Fig. 4e). The quartz contains trails of secondary fluid inclusions that probably relate to syn-breccia fluids, i.e. Hoover paragenetic stage 3 (Figs. 3b, 4f,g). At room temperature most of the inclusions contain liquid (80-90%) and vapour (10-20%); one inclusion contains liquid (90-95%), vapour (5-10%) and halite. In freezing experiments the formation of ice was observed in all inclusions, Tfm was below -50 °C and Tm_{ice} ranged from -34 to -25 °C (Table 1). Hydrohalite ($NaCl.2H_2O$) was observed in two L+V inclusions and melted at -1 and +1 °C (Tm_{hh} Table 1). Na:Ca ratios for these two inclusions are 0.4 and 0.7 (Table 1, Figs. 5, 6). Estimates of salinity for L+V inclusions range from 26 to 32 weight % NaCl equivalent (Table 1). The salinity of the L+V+H inclusion is estimated to be 38 weight % NaCl equivalent and the Na:Ca ratio is approximately 1.2 (Table 1, Figs. 5, 6). During heating experiments L+V inclusions homogenized via vapour bubble disappearance between 153 and 172 °C (Fig. 5, Table 1). In the L+V+H inclusion final homogenization occurred via halite dissolution at 188 °C (Fig. 5, Table 1).

Slats-Frosty: A syn-breccia ferroan dolomite-fluorite-pyrite-chalcopyrite vein was selected for microthermometry in the Slats-Frosty area (Fig. 4h,i). Fluorite within the vein contains fracture-parallel trails of pseudo-secondary fluid inclusions that probably relate to syn-breccia fluid, i.e. Slats-Frosty paragenetic stage 3 (Figs. 3c, 4j). At room temperature the inclusions contain liquid (90-98%) and vapour (2-10%; Fig. 4k) ± solid. Solids are halite, hematite and black opaques. The formation of brown ice was observed in all fluid inclusions during freezing, Tfm ranged from -103 to -72 °C and Tm_{ice} was between -35 and -23 °C (Table 1). Hydrohalite was observed in two L+V inclusions and in most L+V+solid inclusions; Tm_{hh} ranged from -22 to 3 °C. Salinity estimates are between 24 and 32 weight % NaCl equivalent and Na:Ca ratios vary from 0.8 to 1.6 (Table 1, Figs. 5, 6). Daughter phases homogenized at higher temperatures than vapour in all inclusions (i.e. Th = Ths); however, homogenization temperatures were not obtained for the solid phases due to cracking of the polished section during heating and subsequent mass decrepitation of the fluid inclusions. Homogenization of the vapour bubble occurred between 68 and 160 °C (Table 1).

Igor: Barite from Wernecke Breccia matrix was selected for microthermometry in the Igor area (Figs. 4l, m). The barite contains clusters of small fluid inclusions that are probably primary and therefore syn-breccia, i.e. Igor paragenetic stage 3 (Fig. 3e). At room temperature the inclusions contain liquid (90-95%) plus vapour (5-10%). Phase changes were difficult to observe in this sample due to the small size of inclusions and the cloudiness of the barite. Only one initial melting temperature of -77 °C was recorded

(Table 1). Tm_{ice} ranged from -54 to -50 ^{o}C. Hydrohalite was observed in one inclusion and melted at -26 ^{o}C. This inclusion has an estimated salinity of 34 weight % NaCl equivalent and a Na:Ca ratio of 0.1 (Table 1, Figs. 5, 6). Thv was between 220 and 250 ^{o}C, somewhat higher than the range of Thv values (70 to 200 ^{o}C) reported by Gillen *et al.* (2004) for quartz from Wernecke Breccia matrix in the Igor area but similar to the upper range of homogenization temperatures (80 to 300 ^{o}C) for inclusions in dolomite, calcite, fluorite and quartz in magnetite-rich assemblages reported by Hitzman *et al.* (1992).

Olympic: Two samples were selected for microthermometry from the Olympic area: a zoned quartz crystal from the matrix of Wernecke Breccia (Fig. 4n,o) and dolomite from a ferroan dolomite-quartz-chalcopyrite±pyrite vein that cross-cuts Wernecke Breccia (Fig. 4p,q). Phase changes were difficult to observe in both samples due to the small size of inclusions. The outer rim of the zoned quartz contains small clusters of fluid inclusions that are probably primary and therefore syn-breccia, i.e. Olympic paragenetic stage 2 (Figs. 3f, 4o). At room temperature the inclusions contain liquid (~80%) plus vapour (~20%). During freezing experiments Tm_{ice} ranged from -29 to -26 ^{o}C; salinity is estimated to be 26 to 28 weight % NaCl equivalent (Table 1). On heating, homogenization occurred via vapour bubble disappearance between 158 and 170 ^{o}C (Table 1, Fig. 5). Dolomite within the vein, in the second sample, contains small clusters of inclusions that are probably primary and therefore representative of post-breccia fluids, i.e. Olympic paragenetic stage 5 (Fig. 3f). At room temperature the inclusions contain liquid (70 to 98%) plus vapour (2 to 30%). During freezing experiments Tm_{ice} ranged from -56 to -25 ^{o}C; salinity is estimated to be 26 to 35 weight % NaCl equivalent (Table 1, Fig. 5). Hydrohalite was observed in one inclusion and melted at -13 ^{o}C. This inclusion has a Na:Ca ratio of 0.1 (Table 1, Fig. 5, 6). Homogenization occurred via vapour bubble disappearance with Thv between 188 and 223 ^{o}C.

C.4.2 Summary of fluid inclusion results

Fluid inclusion data collected during this study is limited but gives an indication of fluid conditions at the time of Wernecke Breccia formation. In the Slab area quartz from breccia matrix contains primary fluid inclusions that indicate syn-breccia fluids were high salinity (~ 42 eq. wt. % NaCl) and Na-dominant with minimum temperatures of

226 to 245°C (based on Th which is < trapping temperature). Quartz from the Hoover prospect contains secondary fluid inclusions that indicate syn-breccia fluids were moderate salinity (26-38 eq. wt. % NaCl), varied from Na-dominant to Ca-dominant and had temperatures of at least 153 to 188°C. At the Slats-Frosty prospect fluorite in a syn-breccia vein contains pseudo-secondary fluid inclusions that indicate syn-breccia fluids were moderate salinity (24-32 eq. wt. % NaCl), ranged from Na-dominant to equal amounts of Na and Ca and had minimum temperatures of 112 to >160°C. Barite from breccia matrix at the Igor prospect contains primary fluid inclusions that indicate syn-breccia fluids had minimum temperatures of 220 to 250°C, were moderately saline (34 eq. wt. % NaCl) and Ca-dominant (based on limited data). Quartz from breccia matrix at the Olympic prospect contains primary fluid inclusions that indicate syn-breccia fluids were moderately saline (26-28 eq. wt. % NaCl) and at least 158 to 170°C.

C.5 Isotope chemistry (C, O, S & H)

Mineral separates from Wernecke Breccia and associated veins were analyzed for carbon-oxygen (94), sulphur (49) and hydrogen (14) isotopes in order to characterize the fluids that produced the breccia and associated IOCG mineralisation. Samples were chosen from various paragenetic stages (Fig. 3) at each prospect in order to document changes in the fluids that may have occurred through time and to compare fluids from prospects hosted in different parts of WSG stratigraphy. In addition, twenty three samples of limestone/dolostone and three samples of carbonaceous shale were analyzed for carbon-oxygen and carbon isotopes, respectively, to provide regional host rock background values. Results are summarized in Tables 2 to 4 and presented in Figures 7 to 11; complete results are in Appendix IV

C.5.1 Methods: carbon – oxygen isotopes

Carbonates were extracted by crushing and handpicking under a binocular microscope and identified using GADDS (general area detector diffraction system) at the James Cook University Advanced Analytical Centre. Whole rock samples were crushed and powdered. Isotopic analyses of mineral separates and whole rock samples were carried out at the University of Tasmania Central Science Laboratory using a modification of the method of McRea (1950). Samples were reacted with H_3PO_4 at 50°C for a 24 hour period, and then CO_2 was separated in a chemical separation line. Gas was analysed on a Micromass Optima Stable Isotope mass spectrometer. Results were corrected for machine error and are expressed relative to Vienna Standard Mean Ocean Water (V-SMOW) and PeeDee belemnite (PDB). Previous analyses of the Biggenden calcite, used as the international standard, have a standard error of ± 0.06 ‰ for $\delta^{13}C$ and ± 0.1 ‰ for $\delta^{18}O$.

C.5.2 Methods: sulphur isotopes

Sulphides and sulphates were extracted by crushing and handpicking under a binocular microscope after the preparation of cut slabs and inspection of sulphide/sulphate phases in thin section for contaminant inclusions. Isotopic analyses were carried out at the University of Tasmania Central Science Laboratory using the method of Robinson and Kusabe (1978). Powdered samples were reacted with cuprous oxide at high temperatures, and then SO_2 was separated from H_2O, CO_2 and non-condensable gases in a chemical separation line. Gas was analysed on a Sira VG Series 2 mass spectrometer. Results were corrected for machine error and are expressed relative to Cañon Diablo Troilite (CDT). Results were calibrated against a secondary gas standard, and results were reproducible to within 0.2 ‰

C.5.3 Methods: hydrogen – oxygen isotopes

Muscovite, biotite and actinolite were extracted by crushing and handpicking under a binocular microscope after the preparation of polished thin sections and petrographic and microprobe inspection of the phases for alteration and/or contaminant inclusions. Fourteen mineral separates from three of the prospects (Slab, Hoover & Igor) were submitted for hydrogen and oxygen isotope analysis. Unfortunately the other study areas did not yield suitable material for analysis. Analyses were carried out at the Scottish Universities Environmental Research Centre, East Kilbride, Scotland. Oxygen was extracted using a laser fluorination system based on that of Sharp (1990), converted to CO_2, and analysed on a VG PRISM III mass spectrometer. Hydrogen was extracted by the method described in Fallick *et al.* (1993) except that hot chromium rather than uranium was used (see Donnelly *et al.*, 2001). Reproducibility for isotopically homogeneous material is ± 5 ‰ (at 1σ) and NBS 30 gives $\delta D = -65$ ‰ All data are reported relative to V-SMOW.

No indications of alteration were found during petrographic and microprobe analyses of the samples. However, two biotite samples gave abnormally high yields of hydrogen and may be chloritized; in addition, three samples of muscovite failed to generate sufficient hydrogen for isotopic assay. These five samples have been omitted from the following discussion.

C.5.4 Results: carbon – oxygen isotopes

WSG: Calcite from limestone layers in FLG returned values of $\delta^{18}O$ = 11.8 to 14.4 ‰ and $\delta^{13}C$ = -2.0 to 0.4 ‰ (Table 2; Fig. 7). Ankerite and ferroan dolomite from a dolostone layer in Quartet Group and from clasts of Quartet Group dolomitic siltstone and dolostone in Wernecke Breccia returned values similar to those of FLG with $\delta^{18}O$ and $\delta^{13}C$ values of 13.9 to 15.9 ‰and -1.9 to -0.3 ‰respectively. Dolomite from dolostone layers, stromatolitic dolostone and intraformational breccia within GLG returned $\delta^{13}C$ values similar to those of FLG and Quartet Group (-1.9 to 1.6 ‰) but significantly higher $\delta^{18}O$ values (16.4 to 24.6 ‰). Whole rock samples of Quartet Group carbonaceous shale returned $\delta^{13}C$ values of -26.7 to -20.8 ‰, consistent with an organic matter origin.

Slab: Samples of ankerite-magnetite alteration, veins and breccia that formed early in the brecciation process at Slab, i.e. Slab paragenetic stage 2 (Fig. 3a), returned $\delta^{13}C$ and $\delta^{18}O$ values of -3.7 to -1.2 ‰ and 10.9 to 14.8 ‰ respectively (Table 2; Fig. 7a). Samples from the main brecciation phase, i.e. Slab stage 3, returned $\delta^{18}O$ values between -2.1 and 15.3 ‰ and $\delta^{13}C$ values of -3.7 to 1.5 ‰ Results for samples from veins that cross-cut breccia, i.e. Slab stage 4, overlap those of stage 3 to some degree and range from $\delta^{18}O$ = 10.1 to 14.6 ‰and $\delta^{13}C$ = -2.7 to 1.24 ‰(Fig. 7 a).

One sample only (SB97-9-119.9 m, Appendix IV) contains coexisting calcite and dolomite. It returned $\delta^{18}O$ values of 13.0 and 14.1 ‰respectively. If isotopic equilibrium was attained then the difference in $\delta^{18}O$ values between the two types of carbonate indicates a fluid temperature of approximately 300°C (using the fractionation factors of Sheppard and Schwartz, (1970) or Golyshev *et al.* (1981).

Hoover: Syn-breccia calcite from the Hoover prospect, i.e. Hoover paragenetic stage 3 (Fig. 3b), returned $\delta^{18}O$ and $\delta^{13}C$ values of 9.4 to 13.5 ‰and -6.8 to -2.3 ‰ respectively (Table 2). The lowest $\delta^{13}C$ values were returned by samples from veins and breccia that cut Quartet Group carbonaceous slate/shale (Fig. 7b).

Slats-Frosty: Samples of syn- and syn to post-breccia carbonate from the Slats-Frosty area, i.e. paragenetic stages 2, 3 and 4 (Fig. 3c), returned $\delta^{18}O$ values between 13.7 and 20.1 ‰ with most samples falling between 13.7 and 15.6 ‰(Fig. 7 c; Table 2). $\delta^{13}C$ values range from -4.4 to -0.2 ‰ with the lowest values returned by veins that cross-cut carbonaceous shale (Fig. 7c).

Slats-Wallbanger: Syn- and syn to post-breccia carbonate from the Slats-Wallbanger area, i.e. paragenetic stages 5 and 6 (Fig. 3d), returned a narrow range of $\delta^{18}O$ values (14.1 to 16.8 ‰) and a wide range of $\delta^{13}C$ values (-10.6 to -2.2 ‰; Table 2). All but two $\delta^{13}C$ values are less than -4 ‰. The highest $\delta^{13}C$ values were returned by veins cutting interlayered metasiltstone and dolostone at the base of GLG (Fig. 7d). The lowest $\delta^{13}C$ values (-10.5, -10.6 ‰) were returned by stage 6 veins that cut a massive magnetite ± ankerite vein.

Igor: Syn- and syn to post-breccia siderite and ferroan dolomite from the Igor prospect, i.e. Igor paragenetic stages 3 and 4 (Fig. 3e) returned $\delta^{13}C$ values of -6.1 to -1.4 ‰ and $\delta^{18}O$ values of 14.6 to 20.0 ‰ (Table 2; Fig. 7e).

Olympic: Syn-breccia, syn to post-breccia and post-breccia carbonate from the Olympic area, i.e. paragenetic stages 1, 2, 3 and 5 (Fig. 3f), returned a fairly narrow range of oxygen and carbon isotope values with $\delta^{18}O$ between 14.8 to 18.7 ‰ and $\delta^{13}C$ from -5.1 to 0.4 ‰ (Table 2; Fig. 7f). One sample (OY94-4-38.90) contains coexisting dolomite and ankerite that returned $\delta^{18}O$ values of 14.8 and 15.6 ‰respectively.

Summary and discussion of carbon – oxygen isotope results

Mean carbon and oxygen isotope values for Paleoproterozoic carbonates, comparable in age to WSG, are $\delta^{18}O \approx 18$ to 22 ‰ and $\delta^{13}C \approx -2$ to 2 ‰ (*cf.* Shields and Veizer, 2002). Samples of WSG have $\delta^{13}C$ values similar to the mean values but variable $\delta^{18}O$ values (Fig. 8). The lower and middle parts of WSG, i.e. FLG and Quartet Group, returned $\delta^{18}O$ values lower than the mean values (~ 12-16 vs. 18-22 ‰; Table 2). GLG, which forms the upper part of WSG, returned $\delta^{18}O$ values similar to the mean values (~ 16-25 vs. 18-22 ‰; Table 2). Thus, although the number of samples collected from the WSG is not large, the results indicate that the lower and middle parts of the supergroup are significantly depleted in $\delta^{18}O$ compared to typical Proterozoic marine carbonate. Samples were collected from areas outside those with visible Wernecke Breccia-related alteration, hence this variation likely reflects either a meteoric water-dominated diagenetic history of pre-GLG strata, or post-WSG carbonate alteration that preferentially affected the lower and middle WSG. Coincidentally, most mineralized Wernecke Breccia sites occur in the lower WSG, i.e. within the most $\delta^{18}O$ depleted stratigraphy.

In general $\delta^{18}O$ results for the prospects studied reflect those of the host strata. For example, the Hoover prospect which is located at the top of the FLG returned $\delta^{18}O$ values of 9.4 to 13.5 ‰ similar to those of the FLG (11.8-14.4 ‰). The Slats-Wallbanger prospect is hosted by strata at the top of the Quartet Group and $\delta^{18}O$ results (14.1 to 16.8 ‰) are comparable to those of the host rocks (13.9 to 15.9 ‰). Likewise, the Olympic prospect returned $\delta^{18}O$ results (14.8-18.7 ‰) similar to host GLG rocks (16.4-24.6 ‰). At Slab, most oxygen isotope results are similar to those of the host FLG, however some samples returned low $\delta^{18}O$ values. The three lowest $\delta^{18}O$ values (-2.1, 2.7, 3.0 ‰) are from breccia zones (Slab stage 3) that are made up of smaller and more rounded clasts than those in other breccia bodies, and appear to have been the focus of abundant fluid flow. Hence, the low $\delta^{18}O$ values are likely a reflection of interaction with large amounts of low $\delta^{18}O$ fluid such as meteoric and/or formation water.

Carbon isotope ratios also appear to be influenced by host rock type. For example, at Hoover, samples from breccia and veins hosted in carbonaceous shale returned values of $\delta^{13}C = -6.8$ to -4.5 ‰while those h osted in calcareous metasiltstone returned higher values of $\delta^{13}C = -3.9$ to -2.3 ‰(Fig. 7b). Similar results are seen at Slats-Frosty and Slats-Wallbanger where $\delta^{13}C$ results for samples from veins in carbonaceous shale or metasiltstone are lower than those from samples in calcareous metasiltstone or dolostone (Figs. 7c,d).

In general, carbon and oxygen isotope ratios for the prospects studied do not vary systematically with paragenetic stage. An exception to this is the Slab prospect where early ankerite-magnetite alteration (Slab stage 2) forms a distinct group with $\delta^{18}O = 10.9$ to 14.8 ‰and $\delta^{13}C = -3.7$ to -1.2 ‰ (Fig. 7a; Table 2). Some of the spread in $\delta^{18}O$ values seen at the prospects and in the WSG samples may be due to the expected fractionation of oxygen isotopes between different types of carbonate (*cf.* Rye and Ohmoto, 1974; Valley *et al.*, 1986; Zheng, 1999). For example, carbonate in a sample from the Olympic prospect (sample OY94-4-38.90 in Appendix IV) is composed of dolomite ($\delta^{18}O = 14.8$ ‰) and ankerite ($\delta^{18}O = 15.6$ ‰); The measured $\delta^{18}O$ difference, $\Delta_{\text{dolomite-ankerite}} = 0.8$, is close to the expected value of 0.97 calculated for equilibrium conditions using the fractionation factors of Zheng (1999), suggesting it may be due to oxygen isotope fractionation.

Measured oxygen isotope ratios of syn-breccia carbonates (stage 3 for Slab, Hoover, Slats-Frosty and Igor and stage 2 for Olympic) were used in conjunction with the fractionation factors of Zheng (1999) for calcite-H_2O, dolomite-H_2O, ankerite-H_2O and siderite-H_2O along with estimates of temperature for syn-breccia fluid: Slab = 300°C, Hoover = 285°C, Slats-Frosty = 235°C, Igor = 350°C and Olympic = 185°C (see section C.6.1 for temperature determination) to calculate $\delta^{18}O_{water}$ values for the prospects. Resulting (syn-breccia) $\delta^{18}O_{water}$ values range from: Slab = -7.9 to 9.5 ‰; Hoover = 3.0 to 7.2 ‰; Slats -Frosty = 5.4 to 11.6 ‰; Igor = 9.4 to 14. 7 ‰ and Olympic = 5.7 to 7.8 ‰ (Table 3; shown for comparison on Fig. 11).

C.5.5 Results: sulphur isotopes

Slab: Pre to syn-breccia, syn-breccia and syn to post-breccia sulphide samples from the Slab area, i.e. Slab paragenetic stages 1 to 4 (Fig. 3a), returned overlapping $\delta^{34}S$ values between -11.5 and 7.1 ‰ (Table 4; Fig. 9a). Samples from the same paragenetic stage show a wide range of values. For example, two samples of chalcopyrite from a stage 3 vein returned $\delta^{34}S$ values of -11.5 and 2.4 ‰ (labelled 1 in Fig. 9a) and samples of pyrite from stage 4 veins returned values of $\delta^{34}S$ = -11.0 and 3.8 ‰ (labelled 2 in Fig. 9a).

Hoover: Syn-breccia and syn to post-breccia sulphide samples from the Hoover prospect, i.e. Hoover paragenetic stages 3 and 4 (Fig. 3b), returned a wide range of $\delta^{34}S$ values between -12.4 and 13.4 ‰ (Table 4; Fig. 9b) with no systematic variation with paragenetic stage. Though, based on the limited data, there is variation in $\delta^{34}S$ with host rock type. Samples from veins and breccia hosted in calcareous metasiltstone returned $\delta^{34}S$ = -12.4 to -1.6 ‰ and samples from those hosted in carbonaceous shale/slate returned $\delta^{34}S$ = 4.6 to 13.4 ‰ (Fig. 9b). However, unlike the C-O data, sulphur isotopic ratios in breccia-related samples cannot be compared to background diagenetic sulphide compositions as this aspect was not investigated during the study.

Slats: Syn-breccia and syn to post-breccia sulphides from the Slats-Wallbanger area, i.e. paragenetic stages 5 and 6 (Fig. 3d), returned $\delta^{34}S$ values between -6.8 and -1.7 ‰ (Table 4; Fig. 9c). The highest $\delta^{34}S$ value (-1.7 ‰) is from a stage 6 ankerite-chalcopyrite-quartz vein that cuts massive magnetite (labelled 1 in Fig. 9c). The remaining values are from stage 5 veins that cut calcareous siltstone at the base of GLG. Pyrite from a hematite-pyrite-quartz vein (paragenetic stage unknown) cutting FLG phyllite at Slats-Frosty returned a $\delta^{34}S$ value of 4.2 ‰ (Fig. 9a).

Igor: Syn-breccia and syn to post-breccia sulphides from the Igor prospect, i.e. Igor paragenetic stages 3 and 4 (Fig. 3e), returned $\delta^{34}S$ values between -8.4 and 4.8 ‰ (Table 4; Fig. 9d). Igor is the only prospect in the study that contains significant amounts of sulphate, as stage 4 barite. Samples of the barite returned $\delta^{34}S$ values between 7.7 and 17.1 ‰. Three samples of co-existing sulphate and sulphide were analyzed and gave the following results: 1) pyrite from the selvage of a barite vein returned $\delta^{34}S$ = 3.3 ‰; $\delta^{34}S_{barite}$ = 7.7 ‰ (labelled 1 in Fig. 9d); 2) chalcopyrite and barite from a hematite-magnetite-chalcopyrite-barite vein/pod returned $\delta^{34}S$ values of -7.9 and 17.1 ‰ respectively (labelled 2 in Fig. 9d); and 3) chalcopyrite and barite from a siderite-chalcopyrite-pyrite-barite vein returned $\delta^{34}S$ values of -8.4 and 8.7 ‰ respectively (labelled 3 in Fig. 9d). Samples 2 and 3 give reasonable temperatures of 243 and 356°C based on the equations[2] of Ohmoto and Lasaga (1982). Sample 1 gives a temperature of 1026°C which is clearly unreasonable indicating this barite-pyrite pair are not in equilibrium (*cf.* Ohmoto and Goldhaber, 1997).

Olympic: Abundant sulphides were not observed in the Olympic area and only three samples were analysed. Syn-breccia (Olympic paragenetic stage 2, Fig 3f) and syn to post-breccia (stage 3) chalcopyrite returned $\delta^{34}S$ = -10.8 ‰ and 5.3 ‰ respectively (Table 4; Fig. 9e). Pyrite in Wernecke Breccia proximal to a diorite dyke, returned a $\delta^{34}S$ value of -0.5 ‰.

Summary and discussion of sulphur isotope results

Sulphur isotope results for sulphide samples from the Wernecke Mountains show a wide range in values from approximately $\delta^{34}S$ = -12 to 13 ‰ with a broad peak between -7 to 0 ‰ (Figs. 9 & 10). In general, the isotopic ratios do not vary systematically with paragenetic stage and can vary greatly even within a single stage. This is illustrated dramatically at Slab where chalcopyrite samples collected a few centimetres apart in a vein returned values of $\delta^{34}S$ = -11.5 and +2.4 ‰

Mechanisms to account for variations in the isotopic composition of sulphide and sulphate minerals include: 1) changes in the temperature of the fluid; 2) changes in

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[2] $1000 \ln\alpha_{\text{sulphate-pyrite}} = 6.463 \times 10^6/T^2 + 0.56$; $1000 \ln\alpha_{\text{sulphate-chalcopyrite}} = 6.513 \times 10^6/T^2 + 0.56$

the redox state of the fluid; 3) changes in the environments where variations in the SO_4^{-2}/H_2S take place; and 4) multiple sources of sulphur (*cf.* Ohmoto and Rye, 1979; Ohmoto and Goldhaber, 1997). If the dominant species in a hydrothermal fluid is H_2S temperature variations cause little change in $\delta^{34}S$ (Ohmoto and Goldhaber, 1997). However, if the fluid contains significant amounts of SO_4^{-2} and H_2S (or other oxidised and reduced species) a change in temperature can cause large variations in the $\delta^{34}S$ values of minerals precipitating from the fluid (*ibid*).

In open systems, the ratio of oxidised to reduced sulphur species can be changed by reactions with wall rocks, for example by reaction with Fe^{+2}-or Fe^{+3}-bearing minerals such as magnetite and hematite, or by precipitation of sulphide or sulphate minerals (*ibid*). Oxidation of the fluid (i.e. an increase in the ratio of oxidised to reduced species) will cause a decrease in the $\delta^{34}S$ values of individual species in the fluid (if equilibrium between oxidised and reduced species is maintained) and reduction may cause an increase (*ibid*). Reduction of fluid can also occur by reaction with organic matter and causes increases in the $\delta^{34}S$ values of H_2S and SO_4^{-2} and subsequently in the $\delta^{34}S$ values of minerals precipitated from this fluid (*ibid*). Indications of this mechanism are seen at the Hoover prospect where veins and breccia hosted in carbonaceous shale returned higher $\delta^{34}S$ values than those hosted in calcareous metasiltstone (Fig. 9b).

The residence time of sulphur species in a particular environment (in addition to differences in pH, fO_2 and fS_2 conditions in the fluid) can affect the degree of isotopic equilibrium attained by sulphate and sulphide (Ohmoto and Rye, 1979). It takes a certain amount of time at given fluid conditions to attain equilibrium between SO_4^{-2} and H_2S (*ibid*). If a change occurs in the SO_4^{-2}/H_2O ratio of the fluid at the site of mineral deposition and minerals are precipitated before the SO_4^{-2} and H_2S have time to attain equilibrium the sulphide and sulphide minerals will reflect this isotopic disequilibrium (*ibid*). Limited isotopic data available for coexisting minerals at the IOCG prospects studied, e.g. pyrite-barite and chalcopyrite-barite at Igor (see above) indicates equilibrium and non-equilibrium conditions were present.

Hydrothermal fluids may acquire sulphur from more than one source by one or more mechanisms and the proportions of sulphur from different sources may vary during the history of ore formation as the plumbing system evolves (*ibid*). Possible sources of sulphur include: 1) seawater; 2) sulphur leached from host sedimentary rocks which may include biogenic and non-biogenic sulphur; 3) sulphur leached from igneous rocks; and 4) magmatic fluid. At Igor, one barite sample has a $\delta^{34}S$ value (17.1 ‰ – see above) similar to that of Proterozoic seawater (18 ‰ – Strauss, 1993) suggesting possible derivation of sulphur from the seawater. At Hoover, some sulphur may have been leached from biogenic pyrite associated with organic matter because sulphide hosted in carbonaceous shale has different $\delta^{34}S$ values than sulphide hosted in metasiltstone (Fig. 9b). At Olympic, a sample of pyrite from Wernecke Breccia that is adjacent to a BPRI dyke has a $\delta^{34}S$ value of 0 ‰suggesting some sulphur may have been derived locally from igneous rocks.

Any, or a combination, of the above mechanisms can cause large differences in the observed sulphur isotopic ratios in an ore deposit. For example, if a fluid acquires aqueous SO_4^{-2} via dissolution of anhydrite that was precipitated from seawater (e.g. $\delta^{34}S$ = 20 ‰) sulphate minerals that precipitate from this fluid may have $\delta^{34}S$ values similar to that of the SO_4^{-2} (i.e. ~ 20 ‰). If the SO_4^{-2} is reduced to H_2S the value of $\delta^{34}S_{H_2S}$ will vary depending on the degree of sulphate reduction – for instance, reduction by interaction with Fe^{+2}-bearing minerals at 250°C would give $\delta^{34}S_{H_2S}$ values between -5 and +20 ‰(*ibid*).

The above explanations for heterogeneous sulphur isotope response will be revisited in the overall discussion of fluid properties (section C.6).

C.5.6 Results: hydrogen – oxygen isotopes

Samples of syn-breccia biotite, muscovite and actinolite were analysed for hydrogen and oxygen isotopes. Biotite from Wernecke Breccia matrix in the Slab area returned δD values of -115 and -84 ‰and $\delta^{18}O$ values of 8.3 and 9.5 ‰ (Table 5). Muscovite from breccia matrix returned δD = -45 ‰ and $\delta^{18}O$ = 10.6 ‰. Biotite from a vein in a breccia clast in the Slab area returned δD and $\delta^{18}O$ values of -141 and 6.7 ‰ respectively. Muscovite in veins cutting altered FLG proximal to Wernecke Breccia in the Slab area returned δD values of -54 and -21 ‰and $\delta^{18}O$ values of 11.1 and 11.6 ‰; actinolite from a vein returned δD = -22 and -18 ‰ and $\delta^{18}O$ = 11.0 ‰. Biotite from Wernecke Breccia matrix at the Hoover prospect returned δD and $\delta^{18}O$ values of -119 and 7.8 ‰ respectively (Table 5). Muscovite from Wernecke Breccia matrix in the Igor area returned δD and $\delta^{18}O$ values of -55 and 9.9 ‰ respectively (Table 5).

Hydrogen and oxygen isotope values for water co-existing with syn-breccia muscovite, biotite and actinolite were calculated using the measured hydrogen and oxygen isotopic ratios, fractionation factors for muscovite-H_2O, biotite-H_2O and actinolite-H_2O (or tremolite-H_2O) and estimates of temperature (Slab = 300°C, Hoover = 285°C, Igor = 350°C – see section C.6.1 for temperature determination). Calculated $\delta^{18}O_{water}$ values for the three prospects fall within a fairly narrow range from 8 to 11 ‰; however, there is a wide spread in calculated δD_{water} values (Table 5, Fig. 11). δD_{water} values calculated from biotite and muscovite range from -73 to -16 ‰and -7 to +27 ‰ respectively; δD_{water} values calculated from actinolite are 0 and 4 ‰

Discussion of hydrogen – oxygen isotope results

Variations in δD are not well understood and may involve: 1) mixing of different waters; 2) evaporation; 3) magma degassing; 4) interaction with younger fluids; 5) exchange with hydrous minerals; and 6) exchange with H_2S of H_2 (Sheppard, 1986; Ohmoto, 1986; Hedenquist *et al.*, 1998). Low and/or variable δD values in the Wernecke samples could be due to any of these, however 1 and 2 seem to be most likely. Organic water derived from carbonaceous material in host WSG rocks would have low δD values (Fig. 11) and could be present in variable amounts in the fluid. Evaporation of meteoric water (or seawater) would cause a shift to higher δD values (*cf.* Sheppard, 1986). The presence of evaporites in the lower part of WSG (Hunt *et al.*, 2005) indicates arid conditions existed at least during the deposition of FLG, however it is not known how long these conditions persisted. Muscovite in Wernecke Breccia and associated veins demonstrates Ar-Ar re-equilibration (see Section B), most likely due to re-heating, and it is not clear what effect, if any, this would have had on the hydrogen isotopic ratios.

$\delta^{18}O$ values calculated for water co-existing with biotite, muscovite and actinolite in Wernecke samples are similar to those calculated for water co-existing with hydrothermal carbonate at the six prospects (Fig. 11). This could be due to their precipitation from fluid with the same oxygen isotope signature or may be an indication that the fluid(s) underwent isotopic exchange with host WSG sedimentary rocks (*cf.* Shepperd, 1986; Ohmoto, 1986). It is not possible to differentiate between the two scenarios, however, isotopic information from the Slab prospect suggests fluid-rock interaction may have been significant. At Slab, limestone layers in FLG have $\delta^{18}O$ values 6 to 10 ‰lower than expected for Proterozoic rocks of similar age (Fig. 8; Table 2), suggesting interaction with a fluid that had a low $\delta^{18}O$ value, such as meteoric, sea or formational water.

C.6 Discussion: fluid characteristics

C.6.1 Constraints on fluid P-T-X

Pressure and temperature

Isochores constructed from fluid inclusion microthermometry data (salinity, Tm_{ice}, Thv and Ths) were used in conjunction with **limited** independent estimates of temperature for Slab and Igor (Table 6) to determine fluid pressure using the methods of Shepherd and Rankin (1998). Independent temperature estimates for Slab and Igor are based on isotopic data, assuming isotopic equilibrium was attained. At Slab a co-existing dolomite-calcite pair indicates a temperature of approximately 300°C (Section C.5.4). At Igor two barite-chalcopyrite pairs indicate temperatures of approximately 240 and 350°C (Section C.5.5). The higher temperature was used to estimate pressure because fluid inclusion data indicate minimum temperatures at Igor were 250°C (this study) to 300°C (Hitzman, *et al.*, 1992). Results for the Slab area range from 2420 to 3000 bars which, assuming lithostatic confining pressure and a rock density of 2.7 g/cm^3 (Shepherd *et al.*, 1985), correspond to depths of approximately 9.1 to 11.3 km (Fig. 12; Table 7). The results are similar to upper estimates for the combined thickness of Quartet Group and GLG (7.4 to 9 km) that overlie the Slab prospect (Fig. 2) based on stratigraphic sections measured by Delaney (1981) that take into account present day exposure and deformation. Results for Igor range from 1500 to 1900 bars (Fig. 12), corresponding to depths of approximately 5.7 to 7.2 km. This depth range is in agreement with the > 4 km estimated thickness for Quartet and GLG strata that overlie the Igor prospect.

The reasonable agreement between stratigraphic thickness obtained from estimates of pressure (this study) and from field mapping (Delaney, 1981; Thorkelson, 2000) allow approximate fluid inclusion trapping temperatures to be calculated for the Hoover, Slats-Frosty and Olympic areas. The Hoover and Slats-Frosty prospects are overlain by approximately 7 to 9 km of strata and the Olympic prospect by 0.4 to 1.5 km (based on stratigraphic sections of Delaney, 1981). Using the upper limit of these estimates on isochore diagrams gave average trapping temperatures of 285, 235 and 185 °C for syn-breccia fluids at Hoover, Slats-Frosty and Olympic respectively (Table 7). The upper estimates of measured thickness were used as these gave the best agreement with fluid inclusion data at Slab and Igor.

Composition

Fluid associated with the formation of Wernecke Breccia is a H_2O-NaCl-$CaCl_2$ plus iron system, based on: 1) the occurrence of halite and hematite daughter phases and 2) the presence of $CaCl_2$ in the fluid as indicated by the formation of brown ice during freezing and initial melting temperatures below -50 °C (*cf.* Potter *et al.*, 1978, Vanko *et al.*, 1988; Zwart and Touret, 1994).

The composition of syn-breccia fluid and the isotopic ratios of minerals precipitated from it appear to be controlled, to some degree, by the composition of the host rocks. Na-rich, high salinity fluids occur at the Slab prospect which is hosted in upper FLG rocks that contain metaevaporite (halite facies; Figs. 5, 6; Table 1; Hunt *et al.*, 2005). Ca-rich, lower salinity fluids occur at the Olympic prospect which occurs in GLG dolostone. Fluids of varied composition and salinity occur between these two end member types in prospects that are hosted by the Quartet Group. Lithological control on isotopic ratios is indicated by $\delta^{18}O$ values for the prospects, i.e. they are similar to those of the host WSG strata (Fig. 8). Carbon isotope results are also indicative of lithological control in that samples from veins hosted by carbonaceous shale have lower $\delta^{13}C$ values than those hosted by calcareous siltstone (Fig. 7b,c).

The composition of alteration that is associated with Wernecke breccia also suggests extensive fluid-rock interaction. Alteration extends for metres to hundreds of metres beyond the breccias and varies from sodic- to potassic- to calcic-dominant with changes in lithology from FLG to Quartet Group to GLG (Section A; Hunt *et al.,* 2005). However, at each prospect, syn to post-breccia alteration and veining are dominated by carbonate. The type of carbonate varies with host lithology. Calcite is the dominant phase at Slab and Hoover; dolomite and ankerite are dominant at Slats and Olympic (*ibid*). Igor is somewhat different in that the late carbonate phase includes siderite, as well as dolomite and ankerite, and is accompanied by abundant barite.

C.6.2 Constraints on fluid sources

Possible sources for fluids that formed Wernecke Breccia and associated IOCG mineralisation are: meteoric water, seawater, water derived from diagenesis and compaction of basinal sediments (including evaporites), water derived during metamorphism of WSG, and magmatic water. Tracers for the origin of water are provided by its D/H and $^{18}O/^{16}O$ ratios (Sheppard, 1986). Calculated values for water co-existing with syn-breccia muscovite, biotite and actinolite in Wernecke Breccia show wide scatter, especially in δD (-73 to +27 ‰; Fig. 11) and thus do not uniquely constrain the fluid source. The fluid may have been a mixture of waters from a number of sources and several possibilities are outlined below, based on the limited data available.

Evolved seawater: The δD and $\delta^{18}O$ values of seawater are initially increased by evaporation as shown in Figure 11 (Knauth and Beeunas, 1986). Seawater evolved in this way could account for the high δD_{water} values calculated from muscovite in Wernecke Breccia samples and represents one of the few ways such values could be formed (assuming the δD values for muscovite have not been affected by younger events – see Section C.5.6). Some of the high $\delta^{18}O$ values could also be due to exchange between seawater and ^{18}O-rich minerals such as carbonates (Sheppard, 1986). Mixing of this evolved seawater with variable amounts of low δD water (e.g. organic water) could produce the remainder of the depleted δD_{water} values seen.

Evolved meteoric water: The evaporation of meteoric water can also lead to higher δD and $\delta^{18}O$ values and its interaction with ^{18}O-rich minerals in the host rocks can lead to marked enrichment in ^{18}O relative to local meteoric water (Sheppard, 1986; Fig. 11). It is typical for these meteoric-hydrothermal waters to acquire high $\delta^{18}O$ values as demonstrated by geothermal waters in the Salton Sea and Lanzarote areas (Fig. 11). Therefore, meteoric water that has undergone evaporation and/or water-rock isotope exchange is a possible source of Wernecke Breccia fluids; the range in δD values could be due to mixing with variable amounts of low δD water and/or evolved seawater.

Several syn-breccia calcite samples that returned highly depleted $\delta^{18}O$ values (Fig. 7a) may be giving a window into local meteoric waters at the time of breccia formation. Calculated $\delta^{18}O_{water}$ values for these samples range from approximately -8 to -3 ‰(Fig. 11). The $\delta^{18}O_{water}$ values correspond to calculated δD values of -54 to -14 ‰[3] assuming ancient and modern meteoric waters are comparable and that there were no local affects on isotopic compositions by geographic parameters (e.g. temperature, latitude, altitude; Sheppard, 1986). By analogy with present-day meteoric water patterns

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[3] $\delta D = 8\delta^{18}O + 10$ (*cf.* Sheppard, 1986)

these values would require temperate (to arctic) latitudes (Sheppard, 1986). Paleogeographic reconstructions for the Proterozoic are ambiguous (*cf.* Moores, 1991; Karlstom *et al.*, 1999; Pisarevsky *et al.*, 2003) and it is not possible to determine if these values are reasonable for meteoric water of Proterozoic age. However, in general, the high salinity of Wernecke Breccia fluids is not consistent with a dominantly meteoric water source for the fluids.

Formation water/metamorphic water: Calculated δD_{water} and $\delta^{18}O_{water}$ values for Wernecke Breccia samples fall in the overlapping area between high temperature formation waters (these predominate at the heavy $\delta^{18}O$ end of this field) and metamorphic waters (Fig. 11). Thus, an origin as evolved formation water would be consistent with most of the calculated isotopic ratios.

Magmatic waters: Several of the samples have δD–$\delta^{18}O$ fluid values that fall within the magmatic fields on Figure 11. Magmatic waters are considered less likely as a fluid source than other sources because there are no magmatic rocks of appropriate age in the Wernecke area, but devolatilization of magmas beneath the WSG is geologically feasible. The breccias are spatially associated with BPRI however the *ca.* 1600 Ma age of Wernecke Breccia rules out a genetic link between BPRI magmatism (*ca.* 1710 Ma) and breccia/mineralizing fluids (Thorkelson *et al.*, 2001a,b). The age of the Slab volcanics is unknown however, they occur as clasts within Wernecke Breccia and so must also be older than the breccia (Thorkelson, 2000). High temperature volcanic vapour is also considered a less likely fluid source because no vapour-rich fluid inclusions were observed in the breccia samples; however, this could simply reflect trapping conditions.

Constraints from carbon and oxygen isotope data

Dissolution of limestone/dolostone and oxidation of organic matter are possible sources of carbon for hydrothermally precipitated carbonates (*cf.* Ohmoto and Rye, 1979; Giuliani *et al.*, 2000) associated with Wernecke Breccia. Magmatic waters as a source are considered less likely for reasons already discussed. Limestone/dolostone can provide CO_2 to hydrothermal fluids through dissolution and decarbonation reactions (Ohmoto and Rye, 1979). At high temperatures (> ~ 200°C) dissolution reactions should produce CO_2 that is isotopically similar to that of the original carbonate (*ibid*), i.e. $\delta^{13}C_{CO_2}$ produced by dissolution ≈ $\delta^{13}C$ of the limestone/dolostone (*ibid*). Decarbonation reactions may produce $\delta^{13}C$ enriched CO_2 (*ibid*). Organic components in sediments typically have low δ^{13}C values (approx. -35 to -10 ‰) and oxidation of this material produces CO_2 with a similar isotopic composition (*cf.* Ohmoto and Rye, 1979; Ganor *et al.*, 1994). Thus, carbonates with δ^{13}C values approaching those of limestone/dolostone would precipitate from fluids containing carbon derived from limestone dissolution and carbonates with low δ^{13}C values would precipitate from fluids containing CO_2 derived from the oxidation of organic matter (*ibid*).

δ^{13}C values for carbonaceous shale in the Quartet Group range from -26.7 to -20.8 ‰. Calcite precipitated from the oxidation of this organic matter would have δ^{13}C values of -16.9 to -11.0 ‰ at Slab (300 °C), -16.5 to -10.6 ‰ at Hoover (285 °C), -14.9 to -9.0 ‰ at Slats-Frosty (235 °C), -18.1 to -12.2 ‰ at Igor (350 °C) and -12.9 to -7.0 ‰ at Olympic (185 °C) based on the fractionation factors of Chacko *et al.* (1991) for calcite-graphite; data are not available for other carbonate compositions at these temperatures. Measured δ^{13}C values for carbonates are higher than these values at all of the prospects (Table 2), thus organic matter is unlikely to have been the main source of carbon for the hydrothermal precipitates. However, low δ^{13}C values returned by veins that cut carbonaceous shale at the Hoover and Slats-Frosty prospects may have formed from fluid that contained minor amounts of carbon derived from organic matter in addition to carbon from limestone/dolostone sources (see below).

Carbon isotopic ratios for limestone/dolostone of the WSG range from -2.0 to 1.6 ‰(Table 2). Calculated δ^{13}C values for carbonate precipitated from a 200 to 300°C fluid that contained carbon derived by dissolution of WSG limestone/dolostone range from -4.0 to 1.4 ‰for calcite and -3.3 to -1.2 ‰for dolomite – based on $\delta^{13}C_{\text{carbonate mineral}} = \delta^{13}C_{CO_2} + \Delta_{\text{carbonate mineral} - CO_2}$, which is valid for temperatures ≤ 300°C, and fractionation factors for calcite-CO_2 and dolomite-CO_2 (Ohmoto and Rye, 1979). δ^{13}C values for hydrothermal carbonates from the prospects studied (Table 2) are similar to the calculated values suggesting dissolution of carbonates was the main source of carbon in the fluid. The influence of WSG strata on the isotopic composition of the fluid is also reflected in the δ^{18}O results for the hydrothermal carbonates, i.e. they are similar to those of the host rocks (Figs. 7 & 8).

Constraints from sulphur isotopes

Sulphur in hydrothermally precipitated sulphide and sulphate minerals associated with Wernecke Breccia could have been derived from several sources including: seawater, host sedimentary and igneous rocks and magmatic fluids. There is a wide spread in sulphur isotopic ratios for the prospects studied and no sulphur reservoir is clearly defined (Fig. 10). Constraints on possible sulphur sources can be obtained from sulphur isotope contours on a log fO_2 versus pH diagram using the method of Ohmoto (1972). This approach was used at the Slab prospect as it had been examined in more detail than the other prospects in the study. Values of $\delta^{34}S$ were calculated for pyrite that would have precipitated from the fluid at Slab assuming: 1) a seawater/sedimentary source and 2) a magmatic source for sulphur. No published sulphur isotope contours were available for fluid conditions similar to those at Slab (i.e. highly saline, 300 ^{o}C fluids at 2.5 kb) and new diagrams were constructed as described below.

Oxygen and sulphur fugacity conditions were estimated by examining the mineral assemblages present. At Slab, magnetite (Slab stage 2) is replaced by hematite, but continues to occur in minor amounts in later paragenetic stages (Fig. 3a; Hunt *et al.*, 2005). The hematite occurs with pyrite and both are overprinted by chalcopyrite (Slab stages 3 & 4); pyrrhotite and bornite are not present. This assemblage indicates the oxygen and sulphur fugacity of the fluids changed through time but remained close to the magnetite-hematite-pyrite triple junction. Construction of an oxygen fugacity versus sulphur fugacity diagram at pressure (2.5 kb) and temperature (300 ^{o}C) appropriate for Slab fluids shows that in order for the above mineral assemblage to occur log fO_2 and log fS_2 values are limited to approximately -31 to -30 and -10 to -7 respectively (Fig. 13). The presence of calcite and/or ankerite, and the absence of graphite or anhydrite, with the above mineral assemblage puts constraints on possible CO_2 fugacity values and limits them to log $fCO_2 > -1$ and < 2 (Fig. 13).

pH versus log fO_2 diagrams (Fig. 14) were constructed for activities of H_2S determined from the range of sulphur fugacity values identified above. Sulphur isotope contours were constructed, as outlined in Ohmoto (1972), for the middle of this range and are shown on Fig. 14a. Expected pyrite $\delta^{34}S$ values were calculated for the contours assuming the initial source of sulphur was: 1) magmatic with $\delta^{34}S = 0$ ‰(*cf.* Ohmoto and Goldhaber, 1997) and 2) Paleoproterozoic seawater with $\delta^{34}S = 18$ ‰(Strauss, 1993). Calculated $\delta^{34}S_{pyrite}$ values that fall within fluid conditions at Slab (i.e. near the triple junction) range from about -22 to -15 ‰ for a magmatic source of sulphur and from -3.4 to +3.6 ‰ for a seawater source of sulphur (Fig. 14a). Measured sulphur isotope values for pyrite at Slab range from -5.8 to +7.1 ‰, similar to those calculated for a seawater source of sulphur. Thus, in the Slab area sulphur isotope results indicate that the source of sulphur was likely non-magmatic and may have been derived from seawater or sediments/evaporite deposits precipitated from that seawater as these would have a similar isotopic composition to the seawater (*cf.* Strauss, 1997).

The wide range of $\delta^{34}S$ values at Slab may be due to small changes in fO_2 and/or pH because fluid conditions occur where the $\delta^{34}S$ contours are close together (Fig. 14a), hence sulphide minerals precipitated from the fluid may exhibit a large variation in $\delta^{34}S$ values that reflect only a slight variation in fO_2 or pH in the hydrothermal fluids (Ohmoto, 1972). Variability in $\delta^{34}S$ values from the Hoover, Slats, Igor and Olympic prospects may also be due to changes in fO_2 or pH in the hydrothermal fluids however, further data is required to determine this.

Constraints from composition

Fluid inclusion data indicate Wernecke Breccia fluids are high salinity (24-42 wt. % NaCl eq.) $NaCl-CaCl_2-H_2O$ brines. This is consistent with their derivation from an evaporite-bearing sedimentary sequence deposited at a continental margin (*cf.* Yardley and Graham, 2002). Thus, derivation of the fluid from the shallow marine, halite-bearing WSG is plausible.

C.7 Conclusions

Fluid inclusion analyses of Wernecke Breccia samples allowed the estimation of PTX characteristics for the breccia forming-IOCG mineralizing fluids. Estimates of fluid pressure determined from fluid inclusion data are in reasonable agreement with those based on the thickness of overlying stratigraphy and vary from approximately 0.4 to 2.4 kb. Pressure adjusted fluid temperatures range from about 185 to 350 ^{o}C (Table 7). Mineral assemblages and cross-cutting relationships observed at all of the IOCG prospects indicate fluid conditions were close to the magnetite-hematite-pyrite triple junction during brecciation. The fluids are high salinity (24-42 wt. % NaCl equiv.) $NaCl-CaCl_2-H_2O$ brines with compositions that appear to reflect those of the host strata. For example, at the Slab prospect which is hosted by upper FLG that contains metaevaporites (metahalite; Hunt *et al.*, 2005) the fluids are dominated by Na. Fluids at the Olympic prospect which is hosted by GLG dolostone are dominated by Ca.

Hydrogen isotope data indicate the source of fluids that formed Wernecke Breccia and associated IOCG mineralisation was most likely formational/metamorphic water mixed with variable amounts of low δD (e.g. organic) water ± evolved meteoric water and/or evolved seawater (Fig. 11). The $\delta^{13}C$ values of hydrothermal carbonates indicate the carbon was derived in large part from the host WSG (Fig.7; Table 2). The $\delta^{34}S$ values of hydrothermal pyrite, chalcopyrite and barite point to seawater (or sediments/evaporites deposited from seawater) as a likely source for much of the sulphur (Fig. 14) with possible additional sources from the leaching of biogenic pyrite and/or sulphides in local igneous rocks (BPRI and/or Slab volcanics). The high salinity of the fluid is consistent with derivation from an evaporite-bearing sedimentary sequence deposited at a continental margin (Yardley and Graham, 2002).

Taken together these results suggest fluid that formed Wernecke Breccia and associated IOCG mineralisation was dominantly formation-metamorphic water. This is similar to the evaporitic-source model proposed by Barton and Johnson (1996, 2000). Wernecke Breccias have the features indicative of an "evaporitic component", i.e. direct association with an evaporite-bearing basin, voluminous sodic alteration and geochemical data that point to evaporitic rather than magmatic fluids. In Barton and Johnson's (1996, 2000) model evaporitic sources provide the chloride necessary for the transport of metals that are derived from igneous rocks. The source of metals in Wernecke Breccia fluids is unknown but it may have been derived from the host rocks. Mafic to intermediate BPRI and/or the Slab volcanics locally contain disseminated chalcopyrite and/or malachite and could have acted as a source of copper. Published geochemical analyses for BPRI and Slab volcanics indicate Cu contents of 2.7 to 2124 ppm (n = 20) and 10.2 to 11.6 ppm (n = 2) respectively (Thorkelson, 2000). However, the samples analysed were collected from locations proximal to Wernecke Breccia and the results may not represent background values. Geochemical analyses reported for unmineralized samples of WSG (Goodfellow, 1979; Lane, 1990; Thorkelson, 2000) indicate that, at least in part, the FLG and Quartet Group contain elevated levels of Cu, U and Co. Values range from - FLG: Cu = 3-890 ppm (n = 15, avg. = 75 ppm), U = 0.7-40 ppm (n = 10, avg. = 8 ppm) and Co = 1-24 ppm (n = 15, avg. = 12 ppm); and Quartet Group: Cu = 5-1230 ppm (n = 24, avg. = 72 ppm), U = 1.2-18.4 ppm (n = 19, avg. = 4) and Co = 3-27 ppm (n = 24, avg. = 11 ppm). Background values in the GLG are lower than those in the FLG and Quartet Group (*ibid*) and range from: Cu = 5-22 ppm (n = 14, avg. = 9 ppm), U = 0.6-5 ppm (n = 14, avg. = 2 ppm) and Co = 2-11 ppm (n = 14, avg. = 5 ppm). Thus, fluid could have leached metals from the FLG and Quartet Group rocks.

The Barton and Johnson (1996, 2000) model invokes magmatism as the source of heat to drive fluid circulation and generate the required high temperatures. This is problematic in the Wernecke Mountains where no intrusive rocks of appropriate age are known. Previous authors (*cf.* Thorkelson, 2000 for a review) have suggested buried intrusive as the source of heat. Another possible solution that would form fluids with the required high temperatures is to use a simple burial model. An average geothermal gradient of 25-30 ^{o}C/km (*cf.* Raymond, 1995, 2000) and a surface temperature of 25 ^{o}C would produce temperatures of 250 to 295 ^{o}C at depths of 7 to 9 km. The geothermal gradient may have been higher than average in the Wernecke Mountains area because it is postulated to have been a rifting/extensional environment during deposition of WSG and emplacement of BPRI (Thorkelson, 2000; Thorkelson *et al.*, 2001a). Wernecke Breccias were formed syn- to post-deformation (Thorkelson, 2000; Hunt *et al.*, 2005), thus fluid circulation could have been driven by tectonic processes.

Constrained by these new geochemical parameters, the genesis of Wernecke Breccia and associated IOCG mineralisation appears independent of a magmatic cycle and is likely related to temporal evolution of the basin. The circulation of high salinity brines, formed in part from dissolution of evaporites, being derived by deformation and mobilizing mineralizing components from the enclosing rocks. The emerging hypothesis is that periodic (tectonic-stratigraphic) over-pressuring of the fluids led to repeated brecciation and mineral precipitation within hydraulic breccias, vein networks and disseminations. Fluids that formed Wernecke Breccia are perhaps therefore similar to high temperature oil field brines, or to brines that formed emerald deposits in Columbia, the salinity of which is due to dissolution of evaporites (*cf.* Banks *et al.*, 1995; Cheilletz and Giuliani, 1996; Giuliani *et al.*, 2000). Wernecke Breccia-related fluids may also be similar to mineralizing brine described by Knutson *et al.* (1979) from the breccia-hosted Redbank copper deposits in the Northern Territory of Australia. The brine is interpreted to have been derived from sedimentary units and possibly seawater based on C and S isotopic data that indicate their source was the host sedimentary and igneous rocks (*ibid*). The circulation of this non-magmatic brine through fractured igneous and sedimentary rocks led to the precipitation of copper mineralisation within permeable breccia zones (*ibid*). However, the Redbank deposit differs from Wernecke Breccia in that the breccia pipes are interpreted to be the result of the explosive release of fluids from magma with mineralisation precipitated later from circulating non-magmatic fluids (*ibid*).

If, as is indicated by the new fluid data, Wernecke Breccia-related IOCG mineralisation is independent of magmatism this has implications in the search for IOCG deposits on a broader scale. Traditionally, exploration for IOCG deposits has focussed on areas where brecciation is temporally related to magmatism. However, the new information obtained from Wernecke Breccia suggests IOCG mineralisation may also be found in thick sedimentary sequences (thick enough to produce high temperature basinal fluids) that contain sources of metals (host sedimentary of igneous rocks), a source of chloride (evaporites) for metal transport and traps (e.g. breccia zones) for metal precipitation, plus a mechanism to drive fluid flow (e.g. tectonics, gravity).

C.8 References:

Archer A, Bell RT, Delaney GD and Godwin CI (1977).Mineralized breccias of Wernecke Mountains Yukon. Geological Association of Canada Program with Abstracts, 2:5.

Banks DA, Yardley BWD, Cheilletz A, Giuliani G and Rueda F (1995) Chemistry and source of high temperature brines in the Colombian emerald deposits. In: Pasava J, Kribek B and Karel Z (eds.) Mineral Deposits: from their origin to their environmental impacts. Proceedings of the third biennial SGA meeting, Prague. AA Balkema, Rotterdam, pp. 557-560.

Barton MD and Johnson DA (2000) Alternative brine sources for Fe-Oxide(-Cu-Au) systems: implications for hydrothermal alteration and metals. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, Volume 1, PGC Publishing, Adelaide, pp. 43-60.

Barton MD and Johnson DA (1996) Evaporitic-source model for igneous-related Fe oxide-(REE-Cu-Au-U) mineralisation. Geology 24:3:259-262.

Bell RT (1986a) Megabreccias in northeastern Wernecke Mountains, Yukon Territory. Current Research, Part A, Geological Survey of Canada, Paper 86-1A, pp 375-384.

Bell RT (1986b) Geological map of north-eastern Wernecke Mountains, Yukon Territory. Geological Survey of Canada, Open File 1027.

Bell RT (1978) Breccias and uranium mineralisation in the Wernecke Mountains, Yukon Territory – a progress report. Current Research, Part A, Geological Survey of Canada, Paper 78-1A, pp317-322.

Bell RT and Delaney GD (1977) Geology of some uranium occurrences in Yukon Territory. Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, pp 33-37.

Bethke C (2000) The Geochemist's Workbench software package. University of Illinois, Urbana -Champaign.

Brown PE (1998) Fluid inclusion modeling for hydrothermal systems. In: Richards JP and Larson PB (eds.) Techniques in Hydrothermal Ore Deposits Geology, Reviews in Economic Geology 10, pp 151-171.

Chacko, T., Mayeda, T.K., Clayton, R.N. & Goldsmith, J.R. (1991). Oxygen and carbon isotope fractionations between CO2 and calcite, Geochimica et Cosmochimica Acta 55: 2867-2882.

Chambers LA (1982) Sulphur isotope study of a modern intertidal environment and the interpretation of ancient sulphides. Geochimica et Cosmochimica Acta 46: 721-728.

Chaussidon M, Albarede F and Sheppard SMF (1989) Sulphur isotope variations in the mantle from ion microprobe analysis of micro-sulphide inclusions. Earth and Planetary Science Letters 92: 144-156.

Cheilletz A and Giuliani G (1996) The genesis of Colombian emeralds: a restatement. Mineralium Deposita 31:359-364.

Delaney GD (1981) The Mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory; In: Proterozoic Basins of Canada, Geological Survey of Canada, Paper 81-10, pp 1-23.

Donnelly T, Waldron S, Tait A, Dougans J and Bearhop S (2001) Hydrogen isotope analysis of natural abundance and deuterium-enriched waters by reduction over chromium on-line to a dynamic dual inlet isotope-ratio mass spectrometer. Rapid Communications in Mass Spectrometry 15:1297-1303.

Epstein S (1970) Antarctic ice sheet: stable isotope analysis of Byrd Station coves and interhemispheric climactic implications. Science 168: 570-572.

Epstein S, Sharp RP and Gow AJ (1965) Six-year record of hydrogen and oxygen isotope variation in South Pole fur. Journal of Geophysical Research 70: 1809-1814.

Fallick AE, Macaulay CI and Haszeldine RS (1993) Implications of linearly correlated oxygen and hydrogen isotopic compositions for kaolinite and illite in the magnus sandstone, North Sea. Clays and Clay Minerals 41:2:184-190.

Gabrielse H (1967) Tectonic evolution of the northern Canadian Cordillera. Canadian Journal of Earth Sciences, 4:271-298.

Ganor J, Matthews A and Schliestedt M (1994) Post-metamorphic low $\delta^{13}C$ in the Cycladic complex (Greece) and their implications for modelling fluid infiltration processes using carbon isotope compositions. European Journal of Mineralogy 6:365-379.

Giggenbach WF (1992) Magma degassing and mineral deposition in hydrothermal systems along convergent plate boundaries. Economic Geology 87: 1927-1944.

Gillen D, Baker T and Hunt JA (2004) Fluid inclusion studies in Fe-oxide Cu-Au deposits, Wernecke Mountains. In: McPhie J and McGoldrick P (eds), 17th Australian Geological Convention, Hobart, Dynamic Earth: Past, Present and Future, Geological Society of Australia, Abstracts 73, pp 82.

Giuliani G, France-Lanford C, Cheilletz A, Coget P, Branquet Y and Laumomnier B (2000) Sulfate reduction by organic matter in Colombian Emerald deposits: chemical and stable isotope (C,O,H) evidence. Economic Geology 95: 1129-1153.

Graham CM and Harmon RS (1983) Stable isotope evidence on the nature of crust-mantle interactions. In: Hawkesworth CJ and Norry MJ (eds) Continental Basalts and Mantle Xenoliths. Shiva, Nantwich, pp. 20-45.

Graham CM, Harmon RS and Sheppard SMF (1984) Experimental hydrogen isotope exchange between amphibole and water. American Mineralogist 69:128-138.

Goodfellow WD (1979) Geochemistry of copper, lead and zinc mineralisation in Proterozoic rocks near Gillespie Lake, Yukon. Geological Survey of Canada Paper 79-1A: 333-348.

Golyshev SI, Padalko NL & Pechenkin SA (1981) Fractionation of stable oxygen and carbon isotopes in carbonate systems. Geochemistry International 18:85-99.

Hedenquist JW, Arribas A(Jr) and Reynolds TJ (1998) Evolution of an intrusion-centred hydrothermal system: far southwast Lepanto porphyry and epithermal Cu-Au deposits, Philippines. Economic Geology 93:4:373-404.

Hitzman MW, Oreskes N and Einaudi MT (1992) Geological characteristics and tectonic setting of Proterozoic iron oxide (Cu-U-Au-REE) deposits. Precambrian Research 58:241-287.

Hitzman MW (2000) Iron oxide-Cu-Au deposits: what, where, when, and why? In: Porter TM (ed) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 9-25.

Hoefs J (1987) Stable Isotope Geochemistry. 3rd edition, Spriner-Verlag, Berlin.

Hunt JA, Laughton JR, Brideau M-A, Thorkelson DJ, Brookes ML and Baker T (2002) New mapping around the Slab iron oxide-copper-gold occurrence, Wernecke Mountains. In: Emond DS, Weston LH and Lewis LL (eds) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 125-138.

Hunt, JA, Baker, T and Thorkelson, DJ (2005, in press). Regional-scale Proterozoic IOCG-mineralised breccia systems: examples from the Wernecke Mountains, Yukon, Canada. Mineralium Deposita, vol. X, no. x, p. x-x.

Johnson JW, Oelkers EH and Helgeson HC (1992) SUPCRT92: a software package for calculating the standard molal thermodynamic properties of minerals, gases, aqueous species, and reactions from 1 to 5000 bars and 0 to 1000 degrees C. Computers and Geosciences 18: 7: 899-947.

Karlstrom KE, Williams ML, McLelland J, Geissman JW and Ahall K-I (1999) Refining Rodinia: geologic evidence for the Australia-western U.S. connection in the Proterozoic. GSA Today 9:10:1-7.

Kerridge JF, Haymon RM and Kastner M (1983) Sulphur isotope systematics at the 21°N site, East Pacific Rise. Earth and Planetary Science Letters 66:91-100.

Knauth LP and Beeunas MA (1986) Isotope geochemistry of fluid inclusions in Permian halite with implications for the isotopic history of ocean water and the origin of saline formation waters. Geochimica et Cosmochimica Acta 50:419-433.

Knutson J, Fergurson J, Roberts WMB and Donnelly TH (1979) Petrogenesis of copper-bearing breccia pipes, Redbank, Northern Territory, Australia. Economic Geology 74:4:814-826.

Lane RA (1990) Geologic setting and petrology of the Proterozoic Ogilvie Mountains breccia of the Coal Creek Inlier, southern Ogilvie Mountains, Yukon Territory. MSc. Thesis, University of British Columbia, pp. 223.

McCrea T (1950) The isotopic chemistry of carbonates and a paleotemperature scale. Journal of Chemical Physics 18:849-857.

Moores EM (1991) Southwest U.S.-east Antarctic (SWEAT) connection: a hypothesis. Geology 19:425-428.

Norris DK (1997) Geology and mineral and hydrocarbon potential of northern Yukon Territory and northwestern district of Mackenzie. Geological Survey of Canada, Bulletin 422, 401pp.

Ohmoto H (1972) Systematics of sulfur and carbon isotopes in hydrothermal ore deposits. Economic Geology 67:5:551-578.

Ohmoto H (1986) Stable isotope geochemistry of ore deposits. In: Valley JW, Taylor HP Jr and O'Neil JR (eds.) Stable Isotopes in High Temperature Geological Processes, Reviews in Mineralogy, volume 16, Mineralogical Society of America, pp. 491-559.

Ohmoto H and Goldhaber MB (1997) Sulphur and carbon isotopes. In: Barnes HL (ed.) Geochemistry of hydrothermal ore deposits. John Wiley and Sons, New York, USA, pp. 517-612.

Ohmoto H and Lasaga AC (1982) Kinetics of reactions between aqueous sulfates and sulfides in hydrothermal systems. Geochimica et Cosmochimica Acta 46:1727-1745.

Ohmoto H and Rye RO (1979) Isotopes of sulfur and carbon. In: Barnes HL (ed) Geochemistry of hydrothermal ore deposits. John Wiley & Sons, pp. 509-567.

Pisarevsky SA, Wingate MTD and Harris LB (2003) Late Mesoproterozoic (*ca.* 1.2 Ga) palaeomagnetism of the Albany-Fraser orogen: no pre-Rodinia Australia-Laurentia connection. Geophysics Journal International, 155:F6-F11.

Pitzer KS (1991) Activity coefficients in electrolyte solutions. Chapter 3, CRC Press, Boca Raton, Florida.

Potter RW, Clynne MA and Brown DL (1978) Freezing point depression of aqueous NaCl solutions. Economic Geology 73:284-285.

Raymond, J. (2000) Mantle of the earth. In: Sigurdsson H (ed) Encyclopedia of Volcanoes, Academic Press, San Diego, USA, pp 41-54.

Raymond, J. (1995) Petrology: The study of igneous, metamorphic and sedimentary rocks. McGraw-Hill Higher Education, New York, 742p.

Reynolds LJ (2000) Geology of the Olympic Dam Cu-U-Au-Ag-REE deposit. In: Porter TM (ed) Hydrothermal Iron Oxide-Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 93-104.

Robinson BW and Kusabe M (1975) Quantitative preparation of SO_2 for $^{34}S/^{32}S$ analyses, from sulphides by combustion with cuprous oxide. Analytical Chemistry 47:1179-1181.

Roedder E (1984) Fluid inclusions. Rev Mineral 12:1-644.

Rollinson H (1993) Using geochemical data: evaluation, presentation, interpretation. Longman Group, UK, 352p.

Ryan (1998) Ernest Henry copper-gold deposit. In: Australasian Institute of Mining and Metallurgy Monograph 22, pp. 759-768.

Rye RO and Ohmoto H (1974) Sulfur and carbon isotopes and ore genesis: a review. Economic Geology 69:826-842.

Sharp ZD (1990) A laser-based microanalytical method for the in situ determination of oxygen isotope ratios in silicates and oxides. Geochimica et Cosmochimica Acta 54:1353-1357.

Shepherd TJ and Rankin AH (1998) Fluid inclusion techniques of analysis. In: Richards JP and Larsen PB (eds.) Techniques in Hydrothermal Ore Deposits Geology, Reviews in Economic Geology Volume 10, Society of Economic Geologists, pp. 125-150.

Shepherd TJ, Rankin AH and Alderton DHM (1985) A practical guide to fluid inclusion studies. Blackie & Son Limited, Glasgow, 200pp.

Sheppard SMF (1977) The Cornubian batholith, SW England: D/H and $^{18}O/^{16}O$ studies of kaolinite and other alteration minerals. Journal of the Geological Society 133: 573-591.

Sheppard SMF (1981) Stable isotope geochemistry of fluids. In: Rickard DT and Wickman FE (eds) Chemistry and Geochemistry of Solutions at High Temperatures and Pressures. Physical Chemistry of the Earth 13/14: 419-445.

Sheppard SMF (1986) Characterization and isotopic variations in natural waters. In: Valley JW, Taylor HP Jr and O'Neil JR (eds.) Stable Isotopes in High Temperature Geological Processes, Reviews in Mineralogy, volume 16, Mineralogical Society of America, pp. 165-183.

Sheppard SMF and Schwarcz HP (1970) Fractionation of carbon and oxygen isotopes and magnesium between coexisting metamorphic calcite and dolomite. Contributions to Mineralogy and Petrology 26:161-198.

Shields G and Veizer J (2002) Precambrian marine carbonate isotope database: version 1.1. Geochemistry, Geophysics Geosystems, 3:6, 12pp. An electronic journal of the earth sciences published by AGU and the Geochemical Society

Sillitoe RH (2003) Iron oxide copper-gold deposits: an Andean view. Mineralium Deposita 38:787-812.

Strauss H (1997) The isotopic composition of sedimentary sulphur through time. Palaeogeography, Palaeoclimatology, Palaeoecology 132: 97-118.

Strauss H (1993) The sulphur isotope record of Precambrian sulfates: new data and a critical evaluation of the existing record. Precambrian Research 63:225-246.

Suzuoki T and Epstein S (1976) Hydrogen isotope fractionation between OH-bearing minerals and water. Geochimica et Cosmochimica Acta 40:1229-1240.

Taylor BE (1992) Degassing of H_2O from rhyolite magma during eruption and shallow intrusion, and the isotopic composition of magmatic water in hydrothermal systems. Japan Geological Survey Report 279: 190-194.

Taylor HP (1974) The application of oxygen and hydrogen isotope studies to problems of hydrothermal alteration and ore deposition. Economic Geology 69:843-883.

Thorkelson DJ (2000) Geology and mineral occurrences of the Slats Creek, Fairchild Lake and "Dolores Creek" areas, Wernecke Mountains, Yukon Territory. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 10, 73pp.

Thorkelson DJ, Laughton JR and Hunt JA (2002) Geological map of Quartet Lakes area (106E/1), Wernecke Mountains, Yukon (1:50,000 scale). Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Geoscience map 2002-2.

Thorkelson DJ, Mortensen JK, Creaser RA, Davidson GJ and Abbott JG (2001a) Early Proterozoic magmatism in Yukon, Canada: constraints on the evolution of northwestern Laurentia. Canadian Journal of Earth Science 38:1479-1494.

Thorkelson DJ, Mortensen JK, Davidson GJ, Creaser RA, Perez WA and Abbott JG (2001b) Early Mesoproterozoic intrusive breccias in Yukon, Canada: The role of hydrothermal systems in reconstructions of North America and Australia. Precambrian Research 111:31-55.

Valley JW, Taylor HP Jr. and O'Neil JR (1986) Stable isotopes in high temperature geological processes. Reviews in Mineralogy Volume 16, Mineralogical Society of America, 570p.

Vanko DA, Bodnar RJ and Sterner SM (1988) Synthetic fluid inclusions: VIII. Vapour-saturated haltite solubility in part of the system $NaCl-CaCl_2-H_2O$, with application to fluid inclusions from oceanic hydrothermal systems. Geochimica et Cosmochimica Acta 52:2451-2456.

Yardley BWD and Graham JT (2002) The origins of salinity in metamorphic fluids. Geofluids 2:249-256.

Yukon MINFILE, 2003. Database of Yukon mineral occurrences. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, CD-ROM (2).

Zhang Y-G and Fratz JD (1987) Determination of the compositional limits of immiscibility in H (sub 2) O-CO (sub 2) -CaCl (sub 2) fluids to 600 degrees C and 2.0 kbar using synthetic fluid inclusions. AGU 1987 fall meeting, EOS Transactions, American Geophysical Union 68, 44:1539.

Zheng YF (1993) Calculation of oxygen isotope fractionation in hydroxyl-bearing silicates. Earth and Planetary Science Letters 120:247-263.

Zheng YF (1999) Oxygen isotope fractionation in carbonate and sulfate minerals. Geochemical Journal 33:109-126.

Zwart EW and Touret JLR (1994) Melting behaviour and composition of aqueous fluid inclusions in fluorite and calcite: applications within the system H_2O-$CaCl_2$-NaCl. European Journal of Mineralogy 6:773-786.

SECTION D

Wernecke Breccias, Yukon, Canada: an example of a non-magmatic end-member IOCG system and implications for IOCG genesis and classification

D.1 Abstract

Genetic models for iron oxide-copper-gold (IOCG) systems typically invoke formation from magmatic or hybrid magmatic – non-magmatic fluids. New data from Wernecke Breccia indicates IOCG systems can also form in non-magmatic environments. Bodies of Wernecke Breccia are hosted by Early Proterozoic sedimentary rocks of the Wernecke Supergroup (WSG). The location of the breccias is structurally controlled on a regional and local scale. They occur in weak zones and likely formed during the expansion of over-pressured fluids. IOCG mineralisation occurs as veins and disseminations within the breccia bodies and surrounding WSG and is largely made up of early magnetite and/or hematite and later chalcopyrite and pyrite. The breccia bodies are associated with extensive sodic or potassic metasomatic alteration and they are cut by late-stage carbonate, or rarely, barite veins. Cross-cutting relationships indicate multiple phases of brecciation, alteration and mineralisation have occurred. The breccias were formed by low to moderate temperature (185-350 ^{o}C), moderate to high salinity (24-42 wt % NaCl eq.) brines that were likely derived from basinal/metamorphic fluids. Magmatic waters are considered less likely as a fluid source because isotopic data for the breccias do not have a magmatic signature and spatially associated mafic to igneous rocks are significantly older (i.e. *ca*. 1710 vs. 1600 Ma) thus ruling out a genetic connection. Metals and sulphur were probably derived from host strata and fluids circulated via tectonic (and/or gravity) processes.

Modifications to the definition of IOCG systems are proposed that reflect the degree of involvement of magmatic and/or non-magmatic fluids. A division into magmatic, non-magmatic and hybrid types is suggested. Typical magmatic end-member IOCG deposits include Lightning Creek, Osborne and Eloise. Hybrid examples include Ernest Henry, Olympic Dam and Aitik. Wernecke Breccia and Redbank are representative of non-magmatic end members. End-member magmatic deposits such as Lightning Creek have similarities to some porphyry deposits and non-magmatic end-members share characteristic with some sediment hosted Cu deposits (e.g. Tsumeb, Mount Isa) suggesting the range of IOCG deposits may form a link between intrusive- and sedimentary-related deposits.

D.2 Introduction

The discovery of the giant Olympic Dam deposit (Fig. 1, Table 1) in the 1970's and the ensuing increase in exploration and research aimed at the discovery of similar deposits led to proposals for a new deposit class: iron oxide (Cu-U-Au-REE) deposits (*cf.* Meyer, 1988; Einaudi and Oreskes, 1990; Gandhi and Bell, 1990). A description of general characteristics, was first published in Hitzman *et al.* (1992), based largely on studies of Olympic Dam and prospects in the Stuart Shelf (Australia), the Kiruna district (Sweden), the southeast Missouri iron district (USA) and the Wernecke Breccias (Canada). The definition was refined in Hitzman (2000) and "Kiruna-type" deposits were no longer included, instead magnetite-apatite deposits and iron oxide-Cu-Au (IOCG) deposits were considered to form end members of a continuum.

The IOCG end of the spectrum is loosely defined and incorporates deposits worldwide of Archean to Tertiary age that are linked primarily by the presence of abundant Ti-poor (< 2% TiO_2) magnetite and/or hematite and extensive alteration, particularly Fe-, Na- and K-metasomatism (Hitzman *et al.*, 1992; Hitzman, 2000; Sillitoe, 2003). However, there are also noteworthy differences both between deposits within a district and between districts and this has led to a number of genetic models. Some authors propose formation from dominantly magmatic fluids (*cf.* Hitzman *et al.*, 1992; Pollard *et al.*, 1998; Wyborn, 1998; Skirrow, 1999; Perring *et al.*, 2000; Pollard, 2000, 2001). Others propose magmatic fluid as the main source of metals (and sulphur) but invoke mixing with other fluids, usually cooler, less saline and more oxidising, as the mechanism responsible for metal precipitation (*cf.* Mark *et al.*, 2000; Baker, 1998; Reeve *et al.*, 1990; Marschik and Fontboté, 2001); the contribution made to the metal budget by these additional fluids is not clear. Several authors suggest non-magmatic fluids are *necessary* for the formation of IOCG mineralisation (*cf.* Haynes *et al.*, 1995, Haynes, 2000; Barton and Johnson, 1996, 2000; Barton *et al.*, 2000; Hitzman, 2000) and that magmatism is important primarily as a source of heat to drive hydrothermal convection (Barton and Johnson, 1996; 2000).

New information from Wernecke Breccia-associated IOCG prospects suggests the above spectrum of magmatic and hybrid magmatic–non-magmatic models for IOCG deposits should be expanded to accommodate systems formed in an environment unrelated to magmatism, i.e. IOCG mineralisation formed from non-magmatic fluid circulated by non-magmatic processes. This paper presents a summary of new information for Wernecke Breccia and explores variations in IOCG systems. To reflect this variability modifications to the definition of IOCG systems are proposed that accommodate the degree of involvement of magmatic and/or non-magmatic fluids and it is suggested that IOCG systems be divided into non-magmatic and magmatic end-members with hybrid IOCG types in between. Brief descriptions are given of the characteristics of typical end-member IOCG deposits and of those that best fit a hybrid IOCG magmatic – non-magmatic model. Such changes will enable the characteristics of individual types to be better delineated, thus providing more narrowly defined attributes to be used in mineral exploration models.

The range of IOCG systems may form a link between intrusive- and sedimentary-related deposits. Those systems dominated by magmatic and magmatic – non-magmatic fluids have been compared to porphyry deposits (*cf.* Pollard, 2000; Barton and Johnson, 2000; Sillitoe, 2003) and the two deposit types considered to form a continuum of intrusive-related deposits (*cf.* Pollard, 2000; Sillitoe, 2003). IOCG systems dominated by highly saline non-magmatic fluids have similarities with some sediment-hosted Cu deposits, e.g. Mount Isa, Zambian Copper Belt. Thus, IOCG systems may provide valuable information that can be used in the study of, and in exploration for, other types of mineral deposits.

D.3 Wernecke Breccia – Canada

Proterozoic Wernecke Breccia is found in areas of the north-central Yukon Territory that are underlain by the Early Proterozoic Wernecke Supergroup (WSG; Fig. 2). The WSG is an approximately 13 km-thick package of marine fine-grained sandstone, siltstone, dolostone and minor limestone (*cf.* Gabrielse, 1967; Delaney, 1981; Bell, 1986b; Thorkelson, 2000) that was metamorphosed to greenschist facies and multiply deformed during the Proterozoic Racklan Orogeny (*cf.* Thorkelson, 2000; Brideau *et al.*, 2002). Bodies of Wernecke Breccia cross-cut the WSG and are made up largely of clasts of the supergroup in a matrix of rock flour and hydrothermal precipitates (Fig. 3; Table 2). At least 65 bodies of Wernecke Breccia are known and all are associated with iron oxide-Cu(±U±Au±Co) mineralisation, including the Slab (Table 1), Hoover, Slats, Igor and Olympic prospects that were examined in this study (Fig. 2). Mineralisation occurs as disseminations and veins within Wernecke Breccia and surrounding WSG rocks and is largely made up of early magnetite and/or hematite and later chalcopyrite and pyrite (Fig. 3; *cf.* Brookes *et al.*, 2002; Hunt *et al.*, 2002, 2005; Thorkelson *et al.*, 2003; Yukon MINFILE, 2003). Gold-bearing phases were not observed but gold reports with copper in assay results (*cf.* Yukon MINFILE, 2003). Minor uranium (pitchblende, brannerite) and cobalt (cobaltian pyrite, erythrite) mineralisation occurs locally. Gangue is dominantly composed of carbonate (calcite, dolomite, siderite), quartz, albite and K-feldspar with lesser biotite, muscovite, chlorite and fluorite and locally includes minor rutile, epidote (some is allanite), apatite, tourmaline and monazite (*cf.* Brookes *et al.*, 2002; Hunt *et al.*, 2002, 2005). Extensive sodic or potassic metasomatic alteration is spatially associated with the breccia bodies and they are cut by late-stage carbonate veins dominantly composed of calcite and/or dolomite/ankerite that contain minor pyrite and chalcopyrite and trace molybdenite (Fig. 3; Table 3; Hunt *et al.*, 2002, 2005). At the Igor prospect late stage veins are composed dominantly of barite ± siderite.

The breccia bodies are associated with faults on a regional (Richardson Fault array) and local scale (Fig. 2; *cf.* Bell, 1978, 1986a,b; Bell and Delaney, 1977; Thorkelson, 2000) and were emplaced into weak zones such as fault and shear zones, fold axes and lithological contacts during the expansion of over-pressured fluids (Hunt *et al.*, 2005; Section A). Wernecke Breccia occurs throughout the WSG but is most abundant in the lower part of the stratigraphy (Delaney, 1981; Lane, 1990) where there is a transition from calcareous sedimentary strata that contain halite facies meta-evaporites to overlying carbonaceous shale (Hunt *et al.*, 2005; Section A). Fluid pressure calculated from fluid inclusion data for breccias at this stratigraphic level ranges from approximately 2.5 to 3 kbars indicting depth of emplacement of about 7.5 to 9.0 km, roughly equivalent to the thickness of overlying WSG strata (see section C for details).

Cross-cutting relationships indicate multiple phases of brecciation and mineralisation occurred after peak metamorphism and syn- to post-deformation (*cf.* Thorkelson, 2000; Brideau *et al.*, 2002; Hunt *et al.*, 2005). The best estimate for the age of Wernecke Breccia comes from a U-Pb (titanite) date of *ca.* 1595 Ma from the matrix of breccia (Slab prospect) emplaced into the FLG, the lowermost unit of the WSG (Thorkelson, 2000; Thorkelson *et al.*, 2001a). This date is supported by *ca.* 1600 Ma Re-Os dates (R. Creaser and D. Selby pers. comm, 2003, 2004) obtained during this study from molybdenite in late-stage carbonate veins at the Slab prospect (see Section B for details).

All Wernecke Breccia prospects examined in this study are spatially associated with Bonnet Plume River intrusions (BPRI). The BPRI are mafic to intermediate in composition and generally form narrow (< 1 to 5 m) dykes and sills and small stocks throughout the study area (Thorkelson, 2000; Thorkelson *et al.*, 2001b). Clasts of the intrusive rocks are locally abundant in Wernecke Breccia and they are thus, clearly older than the breccia bodies (*ibid*). U-Pb (zircon) dating of four BPRI bodies shows they are *ca*. 1710 Ma (Thorkelson *et al.*, 2001b) and thus rules out a genetic link between BPRI magmatism and breccia/mineralising fluids. Clasts of Slab volcanics are also locally abundant in Wernecke Breccia. Attempts at dating the Slab volcanics have been unsuccessful and their age remains poorly constrained. Cross-cutting relationships show they are older than Wernecke Breccia (Thorkelson, 2000; Thorkelson *et al.*, 2001a; Laughton, 2004; Laughton *et al.*, in review) but the difference in age between the timing of breccia emplacement and volcanism is unknown. However, Slab volcanics have been observed only locally (*ibid.*) and thus are unlikely to have been responsible for brecciation that is observed over hundreds of kilometres (Fig. 2).

The coincidence of Wernecke Breccia and BPRI (and Slab volcanics) appears to be a consequence of the use of the same fluid pathways. The coincident location of some *ca*. 1270 Ma dykes of the Bear River suite indicates these fluid pathways were important for hundreds of millions of years. In addition, ductility contrasts between BPRI and host WSG sedimentary rocks may have produced dilational zones during folding that focussed hydrothermal fluids leading to the preferential development of breccia in these sites, similar to the mechanism proposed by Bell *et al.* (1988) for the Mount Isa deposit.

Fluids that formed Wernecke Breccia and the associated mineralisation were low to moderate temperature (185-350 ^{o}C), moderate to high salinity (24-42 wt. % NaCl eq.) NaCl-$CaCl_2$ brines (Fig. 4; Table 4; Hunt *et al.*, 2004; see Section C for details). Carbon isotopic compositions for hydrothermal carbonates at the prospects studied range from: $\delta^{13}C_{carbonate} \approx$ -7 to +1 ‰ (V-PDB) and $\delta^{18}O_{carbonate}$ values are between -2 and 20 ‰ (V-SMOW). Calculated $\delta^{18}O_{fluid}$ values derived from carbonate range from approximately -8 to +14 ‰ (Fig. 5). Sulphur isotope values for hydrothermal pyrite, chalcopyrite and barite range from: $\delta^{34}S_{pyrite/chalcopyrite} \approx$ -13 to +14 ‰ (CDT) and $\delta^{34}S_{barite} \approx$ 7 to 18 ‰ (CDT; Fig. 5). The $\delta^{18}O$ values for hydrothermal carbonates generally reflect those of the host WSG strata and $\delta^{13}C$ values indicate that carbon was derived in large part from the WSG. The $\delta^{34}S$ values of pyrite, chalcopyrite and barite point to a seawater (or sediments/evaporites deposited from seawater) as a likely source for much of the sulphur with possible additional sulphur from the leaching of biogenic pyrite and/or sulphides in local igneous rocks (Section C). The carbon, oxygen and sulphur isotopic compositions combined with limited hydrogen isotope data indicate fluids were likely derived from formation/metamorphic water mixed with variable amounts of low δD (e.g. organic) water ± evolved meteoric and/or evolved seawater (Section C). The high salinity of the fluid is consistent with derivation from an evaporite-bearing sedimentary sequence (*cf.* Yardley and Graham, 2002). Magmatic waters are considered less likely as a fluid source because the isotopic data do not have a magmatic signature and there are no igneous rocks of appropriate age.

The isotopic signature of Wernecke Breccia fluids and the lack of an obvious intrusive heat source suggest fluid circulation and the high temperatures reached by the fluid(s) occurred via mechanisms other than those related to magmatic heat flow. Temperatures would have increased within the basin during intrusion of the BPRI and throughout prograde metamorphism. The presence of marialitic scapolite in metahalite layers in the lower part of the WSG indicates temperatures reached at least 400 ^{o}C (*cf.* Kwak, 1977) in the deep part of the basin during metamorphism. However, metamorphism occurred prior to brecciation and metamorphic/basinal fluids may have cooled considerably before breccia emplacement and mineralisation occurred. Nevertheless, fluid temperatures in at least the deeper part of the WSG would have been elevated due to the thickness of overlying sediments. A simple burial model would produce fluids with the required high temperatures. An average geothermal gradient of 25-30 ^{o}C (*cf.* Raymond, 1995, 2000) and a surface temperature of 25^{o}C would produce temperatures of 250-295 ^{o}C at depths of 7 to 9 km. The geothermal gradient may have been higher than average in the Wernecke Mountains area because it is postulated to have been a rifting/extensional environment during deposition of the WSG and emplacement of the BPRI (Thorkelson, 2000; Thorkelson *et al.*, 2001b). Cross-cutting relationships indicate Wernecke Breccias were formed syn- to post-deformation (Thorkelson, 2000; Hunt *et al.*, 2005), thus fluid circulation could have been driven by tectonic (and/or gravity) processes (*cf.* Torgerson, 1990; Garven *et al.*, 2001). Another possible mechanism to provide high temperature fluids and drive circulation is the method proposed by Deming (1992) whereby heat is released periodically from continental crust by the onset of free convection in orogenic zones. In this mechanism orogeny increases the permeability of the crust and causes the deep, transitory circulation of fluids through the upper crust setting up free convection cells. The free convection cells supply a tectonically/gravity-driven flow system periodically with heat (and possibly metals) as they form, release heat from the crust and then die out. Thus the influx of heat is transitory but may be large.

The source of metals that formed IOCG mineralisation associated with Wernecke Breccia is unknown but may have been the host strata. BPRI and Slab volcanics locally contain disseminated chalcopyrite and/or malachite and could have acted as sources of copper. Published geochemical analyses indicate Cu contents of 2.7-2124 ppm and 10.2-11.6 ppm respectively for the BPRI and Slab volcanics (Thorkelson, 2000), however, the samples were collected proximal to Wernecke Breccia and the results may not represent background values. Analyses of reportedly unmineralised samples of WSG (Goodfellow, 1979; Lane, 1990; Thorkelson, 2000) indicate the FLG and Quartet Group contain elevated levels of Cu (up to 1230 ppm), U (up to 40 ppm) and Co (up to 27 ppm; see Section C for details) and thus could have acted as a source of metals.

D.4. IOCG systems

IOCG deposits are highly variable and at present the classification scheme is very broad (*cf.* Hitzman *et al.*, 1992; Hitzman, 2000). A single genetic model is not able to explain all the variations seen within this broad class of deposits and the genesis of IOCG mineralisation is variably interpreted to be related to magmatic (*cf.* Hitzman *et al.*, 1992; Baker, 1998; Rotherham *et al.*, 1998; Pollard, 2000; Sillitoe, 2003), non-magmatic (Section C) or hybrid mixtures (*cf.* Barton and Johnson, 1996, 2000; Haynes, 2000; Hitzman, 2000) of fluids. Narrower definitions of IOCG systems would be useful for mineral exploration and research purposes and could be facilitated by dividing up the broad spectrum of deposits. A division into magmatic and non-magmatic end-member IOCG systems with hybrid IOCG systems in between is suggested to reflect the various genetic models (Fig. 6). The proposed division is based on new data for Wernecke Breccia and on existing studies in other areas that include temperature, salinity, fluid composition and stable isotope data. It is intended to reflect the degree of involvement of magmatic fluids and of non-magmatic fluids such as formation and metamorphic waters, (evolved) seawater and/or meteoric waters. The environment of formation, i.e.

magmatic or non-magmatic, is also taken into account in this classification scheme as shown in the lower part of figure 6.

D.4.1 End-member non-magmatic IOCG systems

New geological, fluid inclusion and stable isotope data indicate IOCG mineralisation associated with Wernecke Breccia formed from highly saline non-magmatic fluid (Section C). Ore fluids were low to moderate temperature brines derived from basinal or metamorphic fluids, (evolved) seawater and/or meteoric water that were circulated by tectonic, gravity and/or density related processes. Sulphur isotope values are more consistent with an evaporitic rather than a magmatic source for sulphur. Metals may have been leached from host strata, transported as chloride complexes and precipitated due to changes in temperature, pressure and fluid-rock interaction when fluid expanded forming breccia and possibly subsequently as fluid flowed through the high permeability breccia zones and mixed with meteoric and/or seawater.

Other possible non-magmatic end-member systems include deposits in the Tennant Creek Inlier and Redbank areas of the Northern Territory of Australia (Fig. 1). Tennant Creek mineralisation is interpreted by some authors to have formed from metamorphic/ formation water (*cf.* Large, 1975; Skirrow and Walshe, 2002) and at Redbank isotopic data indicate carbonate and sulphide minerals related to copper mineralisation were deposited from non-magmatic fluids (Knutson *et al.*, 1979). Although there is a spatial relationship with igneous rocks in both areas no genetic connection has been demonstrated.

The Salton Sea geothermal system, where iron oxide-Cu-REE-Co-Au mineralisation is precipitated from evaporite-derived brines may be a modern analogue of a non-magmatic system (*cf.* McKibben and Hardie, 1997; Barton and Johnson, 2000). However, it occurs in a magmatic environment and fluid is interpreted to be circulated by igneous-driven convection (*ibid*). Thus, it cannot be considered an end-member non-magmatic IOCG system, but would lie between hybrid IOCG systems and end-member non-magmatic IOCG systems.

Tennant Creek - Iron oxide-associated Au(-Cu-Bi) mineralisation in the Tennant Creek area, e.g. the West Peko and Eldorado deposits (Table 1), occurs within *ca.* 1860 Ma (*cf.* Compston, 1995) weakly metamorphosed greywacke, siltstone and shale of the

Warramunga Formation (Table 2; *cf.* Large, 1975; Skirrow and Walshe, 2002). Magnetite (and/or hematite)-chlorite-quartz ± talc ± carbonate ironstone bodies occur within the sedimentary rocks in dilatent zones that were generated by shearing and folding during the *ca.* 1860-1840 Ma Barramundi orogeny (*cf.* Skirrow and Walshe, 2002). Au(-Cu-Bi) mineralisation overprinted selected ironstones during later deformation (at or before *ca.* 1830-1825 Ma) and appears to be structurally controlled (*cf.* Wedekind and Love, 1990; Huston *et al.*, 1993; Skirrow and Walshe, 2002). For example, the West Peko deposit is proximal to the axis of a regional-scale syncline and the Eldorado deposit is located within parasitic folds on the limb of a regional-scale anticline (Skirrow and Walshe, 2002).

Mineralisation ranges from reduced to oxidised and occurs as: 1) magnetite-pyrrhotite-pyrite Cu-rich deposits; 2) magnetite-dominant Au-Bi deposits and 3) hematite-dominant Au deposits (*ibid*). At West Peko the central parts of two magnetite-hematite-quartz-chlorite ironstone bodies were overprinted by sulphides (pyrrhotite, chalcopyrite ± pyrite), gold, chlorite ± stilpnomelane ± minnesotaite ± talc ± siderite and ± calcite resulting in lenses and zones of (reduced) chalcopyrite-gold-bismuthinite mineralisation (Table 3; *ibid*). At the Eldorado deposit hematite-magnetite-quartz ironstone occurs as discontinuous, discordant bodies proximal to hematitic shale and siltstone units of the Warramunga Formation (*ibid*). Au-Bi (oxidised) mineralisation occurs near a change in the trend of the ironstones and occurs mainly in stringer and breccia zones that broadly correlate with areas of pervasive silicification and zones of higher magnetite/(hematite + martite; *ibid*).

The ironstone bodies formed from hot (350-400 ^{o}C) NaCl + $CaCl_2$ brines (probably formation waters) at pressures of approximately 2.5 to 5 kbars (Fig. 4; Wedekind and Love, 1990; Skirrow and Walshe, 2002). Au-Cu-Bi mineralisation at the West Peko deposit is interpreted to have formed from pyrrhotite ± magnetite-stable, 300-340 ^{o}C, weakly acidic, sulphur- and N_2±CH_4-rich, low salinity (3-10 wt% NaCl eq.) hybrid metamorphic–formation water(s) as it reacted with pre-existing magnetite ± hematite-rich ironstone (Fig. 4; Table 4; Skirrow and Walshe, 2002). Fluid flowed along shear and fault zones that tapped reduced rock packages (that may have included mafic and felsic rocks) beneath the oxidised Warramunga Formation (Skirrow, 1999; Skirrow and Walshe, 2002). The deposition of gold and native bismuth likely occurred due to desulphidation ± oxidation of the reduced fluid as it reacted with ironstone. Desulphidation of the fluid was likely driven by precipitation of chalcopyrite, pyrite, pyrrhotite, and Bi sulphides or sulphosalts in response to increased pH of the fluids (*ibid*). Eldorado mineralisation is interpreted to have formed from intermediate fO_2, low to moderate salinity ore fluid (a variant of the reducing fluid that formed West Peko mineralisation) that mixed with hematite-stable Ca-Na-Cl brine in the presence of pre-existing magnetite-rich ironstone (Skirrow and Walshe, 2002). Fluid mixing likely caused an increase in copper solubility leading to undersaturation of copper minerals and the deposition of gold-bismuth mineralisation without copper phases (*ibid*).

The mechanism of fluid circulation and the source of heat and metals for the Tennant Creek deposits are not clear (*cf.* Large, 1975; Wedekind and Love, 1990; Skirrow and Walshe, 2002). The West Peko deposit occurs several hundred metres above a 300-500 m-thick stratiform quartz-potassium feldspar-plagioclase porphyry and geophysical data for the Eldorado area indicate granite occurs nearby (Skirrow and Walshe, 2002), however the age of the intrusive rocks is not well constrained, hence their role as a potential heat or metal source is not known. The deposits are associated with regional-scale structures and Au(-Cu-Bi) mineralisation occurred during deformation (*cf.* Wedekind and Love, 1990; Skirrow and Walshe, 2002) suggesting fluid flow may have been driven by tectonic and/or gravity processes. Metals may have been derived from the host sedimentary strata and/or igneous rocks as fluid flowed along shear and fault zones.

Redbank – IOCG prospects in the Redbank area (Fig. 1) are hosted by sedimentary and volcanic rocks of the Paleoproterozoic Tawallah Group (< 1780 to 1720 Ma) in the McArthur Basin (Knutson *et al.*, 1979; Jackson and Southgate, 2000; Betts *et al.*, 2003). Mineralisation occurs within breccia pipes located along a series of basement faults (Orridge and Mason, 1975). Approximately 50 pipes have been recognised and significant copper mineralisation occurs in several, e.g. Bluff and Sandy Flat (Table 1; Knutson *et al.*, 1979). Chalcopyrite and minor covellite and chalcocite occur as breccia infill and as veins and disseminations in breccia clasts and host rocks; sulphur isotope values for chalcopyrite range from $\delta^{34}S \approx 4$ to 16 ‰ (*ibid*). Brecciation was accompanied by pervasive potassium and iron metasomatism and abundant K-feldspar, hematite and chlorite occurs in breccia and host strata (*ibid*). Mineralisation is associated with carbonate alteration and isotopic values for hydrothermal dolomite range from $\delta^{13}C = -4.6$ to +3.2 ‰and $\delta^{18}O = 13.1$ to 22.6 ‰, indicating a fluid temperature of 150-300°C (*ibid*). The carbon and oxygen isotopic compositions reflect those of the host strata and fluid related to the copper mineralisation was probably derived from the sedimentary units (*ibid*). Sulphur was likely derived from marine sulphate and dissolved sedimentary and magmatic sulphides; copper may have been leached from host sedimentary and igneous strata (*ibid*).

D.4.2 End-member magmatic IOCG systems

IOCG deposits at the magmatic end of the spectrum formed from ore fluids with a dominantly magmatic origin and are typified by deposits such as Lightning Creek, Osborne and Eloise (Tables 1 to 4; Fig. 1). The deposits are generally temporally associated with igneous rocks, e.g. Lightning Creek (*cf.* Perring *et al.*, 2000). However, exposed, coeval igneous rocks are not always present and the presence of buried intrusions may be inferred from geophysical data, for example at Eloise (Baker, 1998). Fluid was moderate to high temperature, saline (Fig. 4; Table 4), probably a source of metals, and circulated by heat from magmatism (*cf.* Adshead *et al.*, 1998; Pollard, 2000, 2001). Isotopic values indicate sulphur and fluids were derived from a magmatic source (Fig 5).

Lightning Creek, Osborne and Eloise: regional setting – The Lightning Creek, Osborne and Eloise deposits are located in the Cloncurry district of Australia in the eastern part of the Mount Isa inlier (Fig. 1; *cf.* Adshead *et al.*, 1998; Williams and Skirrow, 2000). They occur within an intracratonic basin (Table 2) that consists of metamorphosed pelitic, psammitic, calcareous and evaporitic sedimentary rocks and felsic and basic volcanic rocks (*cf.* Beardsmore *et al.*, 1988; Blake and Stewart, 1992) that have been divided into three superbasin sequences: Leichhardt (*ca.* 1800-1750 Ma), Calvert (*ca.* 1730-1690 Ma) and Isa (*ca.* 1670-1575 Ma; *cf.* Southgate *et al.*, 2000; Scott *et al.*, 2000). Deformation and metamorphism of the host rocks occurred during the Diamantina (*ca.* 1600 Ma) and Isan orogenies (*ca.* 1550-1500 Ma; *cf.* O'Dea *et al.*, 1997; Betts *et al.*, 1998; Laing, 1998; MacCready *et al.*, 1998) and significant granitoid igneous activity occurred *ca.* 1550 Ma and *ca.* 1540-1500 Ma (Page and Sun, 1998; Pollard *et al.*, 1998; Wyborn, 1998). Intense, widespread sodic-calcic alteration is temporally and spatially associated with emplacement of the intrusions and is concentrated along regional-scale, deep-seated structures such as the Cloncurry fault (*cf.* De Jong and Williams, 1995; Mark and De Jong, 1996).

Lightning Creek – The Lightning Creek prospect is hosted by a suite of *ca.*1540-1500 Ma (Page and Sun, 1998) intrusive rocks (Table 2; Perring *et al.*, 2000). The suite is largely made up of porphyritic quartz monzodiorite (that contains enclaves of quartz diorite) intruded by porphyritic monzogranite and fine-grained alkali feldspar granite (*ibid*). High temperature (> 500 ^{o}C) waters with a composition similar to primary magmatic fluid caused widespread sodic-calcic alteration of the host rocks and removed K, Fe, Cl and Cu (Fig. 4; Tables 3, 4; *ibid*).

Quartzofeldspathic and aplitic sills that are interpreted to be late-stage differentiates cut the altered intrusive rocks (Perring *et al.*, 2000). The sills crystallised at temperatures > 500^{o}C and pressures > 1.5 kbars under hydrous conditions and episodically released a fluid phase (*ibid*). The released fluid underwent phase separation that resulted in CO_2-rich vapour and a hypersaline (33-55 wt. % NaCl eq.) aqueous phase (Williams *et al.*, 1999; *ibid*). The hypersaline fluid contained significant amounts of Fe (~ 10 wt %) and Cu (~1wt %) plus Na, Ca, K, Cl and Ba and led to the formation of Ca-Fe ± Na alteration (pyroxene-albite ± magnetite) within the sills and quartz-magnetite ± clinopyroxene ± albite ± pyrite veins outside the sills (Williams *et al.*, 1999; Perring *et al.*, 2000). Calculated $\delta^{18}O_{water}$ values for quartz in the quartz-magnetite veins range from +7.8 to +9.9 ‰ and sulphur isotope values for pyrite range from -4.9 to -3.3 ‰, compatible with a magmatic fluid source (Fig. 5; *ibid*).

Narrow, steeply-dipping calcite ± chlorite ± pyrite veins cut the quartz-magnetite veins and contain minor Cu(-Au) mineralisation. The carbonate veins likely formed from cooler (< 200 ^{o}C), lower salinity (15-28 wt % NaCl eq.) fluid that may have been produced by mixing meteoric water with magmatic hydrothermal fluid (Fig. 4; Table 4). Sulphur isotope values for pyrite and chalcopyrite in the carbonate veins are -3.2 and -9.7 ‰respectively (Fig. 5; *ibid*).

Osborne – The Osborne deposit occurs within the Isa Superbasin in the Soldiers Cap Group of Cover Sequence 3 (*ca.* 1670-1600 Ma; Table 2; *cf.* Beardsmore *et al.*, 1988). Host rocks include feldspathic psammite (± pelite) and banded ironstone (magnetite-quartz-apatite) cut by tholeiitic amphibolite dykes and a body of ultra mafic rock (*cf.* Adshead *et al.*, 1998). Host rocks were metamorphosed and deformed *ca.* 1595 Ma (Perkins and Wyborn, 1998; Gauthier *et al.*, 2001; Rubenach *et al.*, 2001). Na-Ca-Fe metasomatism accompanied this deformation and produced albitites and metasomatic ironstones (Rubenach *et al.*, 2001). Pegmatite dykes were emplaced late syn- to immediately post-deformation (*ibid*) and pre-date and cross-cut the mineralisation (Adshead *et al.*, 1998). A second generation of albitisation and calc-silicate alteration (also dated at *ca*. 1595 Ma) replaced country rocks adjacent to the pegmatite dykes and the Osborne orebodies (Table 3; *cf.* Rubenach *et al.*, 2001).

Magnetite-pyrite-chalcopyrite mineralisation occurs in zones of silicification mainly along contacts between ironstone and psammite (*cf.* Adshead *et al.*, 1998). The mineralisation has high copper and gold grades and contains anomalous concentrations of cobalt, molybdenum, silver, selenium, bismuth, mercury, tellurium, tin, fluorine and chlorine (Table 1; Adshead, 1995). Disseminated mineralisation occurs within the banded ironstone units and is associated with secondary hematite-pyrite-magnetite. One ore body (3E) is not associated with ironstone and is hosted by pegmatite, psammite, metatholeiite and schist (Adshead *et al.*, 1998). This ore body contains abundant pyrrhotite and pyrite-magnetite (Adshead, 1995). Cu:Au ratios vary across the deposit and pyrrhotite-bearing associations have higher ratios than hematite-stable alteration. Gangue minerals include quartz, calcite, chlorite, muscovite, magnetite, pyrite and/or pyrrhotite, iron-cobalt sulphides, apatite, molybdenite and tourmaline plus bismuth sulphides and sulphosalts (Adshead, 1995).

Primary fluid inclusions in pre-mineralisation silica flooding contain high temperature (>450 ^{o}C), chemically complex (Na, Fe, Ca, K, Mn, Ba, CO_2, CH_4) hypersaline (60-70 wt % salts) brines or $CO_2 \pm CH_4$ vapours (Fig. 4; Table 4; Adshead, 1995; Adshead *et al.*, 1998). Fluid inclusions related to mineralisation contain lower temperature (~300 ^{o}C), aqueous, less saline (20-37 wt % salts) fluids (*ibid*). Fluids are interpreted to have a magmatic and/or retrograde metamorphic origin (*ibid*). Sulphur isotope values for chalcopyrite range from -3 to +2.5 ‰consistent with a magmatic source for sulphur (Fig. 5; *cf.* Rotherham *et al.*, 1998). Calculated oxygen isotope values for water in equilibrium with magnetite range from +5 to +12 ‰, overlapping the magmatic and metamorphic fluids fields (Fig. 5; *cf.* Rotherham *et al.*, 1998). Gold and copper were likely carried as chlorocomplexes and precipitated due to decreases in temperature and salinity and an increase in pH of the fluid (Adshead, 1995; Adshead *et al.*, 1998).

Eloise – The Eloise deposit is hosted within altered and deformed meta-arkose, quartz-biotite schist and amphibolite of the Soldiers Cap Group (Table 2; Cover Sequence 3;

Baker, 1996, 1998). No granitic rocks have been found in the Eloise area however, a large gravity low 10 km east of the deposit may be a pluton related to the Williams batholith (Baker, 1998). The deposit occurs proximal to a bend in a regional shear zone (Baker, 1996) and metasomatism and mineralisation are interpreted to have occurred synchronous with ductile-brittle deformation during the waning stages of the Isan Orogeny and emplacement of the Williams and Naraku batholiths (*ca.* 1540-1490 Ma; Baker and Laing, 1998).

Pyrrhotite-chalcopyrite mineralisation (Table 1; *ca.* 1530 Ma; Baker *et al.*, 2001) occurs in veins, silicified zones and massive sulphide bodies, within highly strained, hornblende-biotite-altered, metasedimentary rocks proximal to and within the Eloise shear zone (*cf.* Baker, 1998). Gold is associated with chalcopyrite. The mineralised system is zoned, i.e. minor magnetite-pyrite occurs south of the pyrrhotite-chalcopyrite mineralisation and pyrrhotite occurs to the northwest (*ibid*). Gangue minerals include quartz, calcite, actinolite/hornblende, chlorite, biotite, muscovite and potassium feldspar. Host rocks were affected by pervasive albitisation and ± hornblende ± biotite ± quartz veins and alteration prior to mineralisation (Table 3). Post mineralisation veins are dominantly composed of potassium feldspar, siderite-hematite or chlorite and are overprinted by silicification and chlorite-calcite veins and late pyrite-quartz ± adularia veins (*ibid*).

Fluid that formed the pervasive pre-mineralisation albite alteration was high temperature (400-500 ^{o}C) ultrasaline brine (Fig. 4; Table 4; Baker, 1996; 1998). Pre-mineralisation hornblende-biotite assemblages that cross-cut the albite alteration also formed from high temperature (450-600 ^{o}C), ultrasaline (32-68 wt% total salts) brines. Fluid that formed mineralisation was lower temperature (200-450 ^{o}C) and lower salinity (30-47 wt% total salts). Calculated oxygen isotope values for water in equilibrium with pre-mineralisation assemblages and with mineralisation overlap those of typical magmatic and metamorphic fluids (Fig. 5; Baker *et al.*, 2001). Sulphur isotope values for pyrite, chalcopyrite and pyrrhotite range from 0.0 to 2.3 ‰ consistent with a magmatic source for sulphur (Fig. 5; *ibid*). The source of fluid is interpreted to be magmatic based on the high salinity of fluid inclusions and on sulphur isotope data (and to a lesser degree oxygen and hydrogen isotope data) that have a magmatic signature (*ibid*). Metals were likely carried as chlorocomplexes with H_2S as the dominant sulphur species. Ore deposition is interpreted to have been predominantly controlled by decreasing fluid temperature and the sulphidation of early Fe-rich alteration (*ibid*).

D.4.3 Hybrid magmatic – non-magmatic IOCG systems

Hybrid magmatic – non-magmatic IOCG systems formed from fluids with a significant non-magmatic component (*cf.* Barton and Johnson, 1996, 2000). They include the large deposits at Olympic Dam (*cf.* Haynes *et al.*, 1995), Aitik (*cf.* Wanhainen *et al.*, 2003), Candelaria (*cf.* Ullrich and Clark, 1999), Salobo (*cf.* Souza and Vieira, 2000) and Ernest Henry (*cf.* Mark *et al.*, 2000; Fig. 1; Table 1). Hybrid IOCG systems are generally spatially and/or temporally associated with igneous rocks (Table 2) and fluid circulation is inferred to have been driven by heat from magmatism; the magmatic fluid may be a source of metals and/or sulphur (*cf.* Barton and Johnson, 1996; 2000). Isotopic and geochemical data indicate input from non-magmatic fluids such as basinal brines or meteoric water (*cf. ibid*). Non-magmatic fluids are commonly highly saline and may be evaporite-derived (*cf. ibid*, Yardley and Graham, 2002). For example, highly saline surficial water is interpreted to have mixed with magmatic water during the formation of the Olympic Dam deposit (*cf.* Reeve *et al.*, 1990; Haynes *et al.*, 1995) and mixing with evaporite-derived brines is invoked at Aitik (Wanhainen *et al.*, 2003).

Olympic Dam – The Olympic Dam deposit (Table 1) occurs in the eastern part of the Gawler Craton (Fig.1), a region largely underlain by *ca.* 1845 Ma deformed granite (Donington Suite) unconformably overlain by metasedimentary and metaigneous units of the Wallaroo Group (*cf.* Creaser, 1995; Ferris *et al.*, 2002). Dominantly felsic lavas and ignimbrites, and minor mafic lavas, of the Gawler Range Volcanics overlie the Wallaroo Group and are intruded by/coeval with (*ca.* 1595-1575 Ma) Hiltaba Suite K-feldspar dominant granite (to granodiorite; *cf.* Ferris *et al.*, 2002). The deposit formed at shallow depth during the late stages of intrusion of voluminous, felsic melts *ca.* 1590 Ma (Table 2; *cf.* Roberts and Hudson, 1983; Reeve *et al.*, 1990; Oreskes and Einaudi, 1990; Cross *et al.*, 1993; Oreskes and Hitzman, 1993; Johnson and Cross, 1995; Haynes *et al.*, 1995; Reynolds, 2000; Skirrow *et al.*, 2002).

Olympic Dam is hosted by fractured, hematite-sericite-altered granite over an ~ 7 x 3 km area and is associated with a zone of dilation related to a regional-scale fault zone (*ibid*). The deposit consists of multi-stage hematitic breccias around a core of barren hematite-quartz breccias (*ibid*). Mineralisation was broadly contemporaneous with brecciation (Oreskes and Einaudi, 1990; Reeve *et al.*, 1990) and occurs dominantly within breccia matrix as disseminated Cu-Fe sulphides and uraninite (*cf.* Roberts and

Hudson, 1983; Oreskes and Hitzman, 1993; Reynolds, 2000). Multiple mineralising episodes have occurred and each has the following general paragenetic assemblages (*cf.* Oreskes and Einaudi, 1990; Haynes *et al.*, 1995): a) magnetite (± hematite), chlorite, sericite, siderite and minor pyrite, chalcopyrite and uraninite, b) hematite, sericite, chalcocite, bornite, pitchblende, barite, fluorite and chlorite, and c) hematite, or hematite + quartz ± barite (Table 3). Assemblages a to c overlap in time and space and there is a transition from chalcopyrite-bearing assemblages to bornite (± chalcocite)-bearing assemblages about 100-300 m below the top of the deposit that locally follows upflow zones (*ibid*).

The repeated brecciation and alteration events that occurred during the formation of the Olympic Dam deposit created complex zoning and structural patterns that make it difficult for any one study area to be representative of the whole deposit. This has led to significant differences in detailed published accounts of the geology and to a number of proposed models for ore formation. Most models indicate formation from a combination of magmatic and non-magmatic fluids (*cf.* Roberts and Hudson, 1983; Oreskes and Einaudi, 1990, 1992; Reeve *et al.*, 1990; Haynes *et al.*, 1995), but at least one model (Johnson and Cross, 1991) suggests fluids were dominantly magmatic. Interaction between hot, saline, relatively reduced metal-bearing brine and cooler, oxidised saline meteoric waters was proposed by Reeve *et al.*, (1990). Haynes *et al.*, (1995) proposed a model involving the repeated mixing of hot, relatively oxidized, iron-rich, F, Ba and CO_2-bearing hydrothermal fluid (magmatic or deeply circulated meteoric water derived from a felsic volcanic or granitic source) with cooler, highly oxidized meteoric or connate fluid that contained Cu, U, Au, Ag and sulphate derived from the interaction of saline lacustrine (or ground) water with mafic volcanic rocks. Oreskes and Einaudi (1990, 1992) suggested fluids that deposited early magnetite were hot (~400 °C), had oxygen isotope characteristics ($\delta^{18}O_{water}$ = 8-10 ‰) similar to those of primary magmatic fluid or deeply circulating fluid that equilibrated with metamorphic basement, were possibly highly saline (31-42 wt % NaCl eq.) and carried Fe, Cu and other metals (Figs. 4, 5; Table 4). They suggest the hot fluids were followed by cooler (200-?360 °C), lower salinity fluids with low $\delta^{18}O_{water}$ values (-2.5 to +4.5 ‰) that may have been derived from surficial fluids (evolved meteoric, closed-basin, ground and/or seawater). Johnson and Cross (1991) presented a model that involves two fluids based on Sm-Nd isotopic evidence: 1) fluid in isotopic equilibrium with Hiltaba Suite granites led to the precipitation of early magnetite and 2) fluid derived from an ascending volatile phase exsolved from mafic/ultramafic magma and enriched in Cu and REE led to the precipitation of copper mineralisation. Reynolds (2000) suggests ore fluids and metals have magmatic sources and Knutson *et al.* (1992) report that Mesoproterozoic basalts are a possible source of copper.

Sulphur isotope data indicate mean δ^{34}S values of: pyrite = -6 ‰, chalcopyrite = -7 ‰, bornite = -7 ‰, chalcocite = -10 ‰ and barite = 11 ‰ (Fig. 5; Eldridge and Danti, 1994). Differences between sulphur isotope values for different minerals are broadly compatible with a temperature of formation of approximately 300 °C (*ibid*). Bimodal results for sulphur isotope data in an area with elevated gold values suggest boiling may have been important in gold deposition (*ibid*).

Aitik - The Aitik deposit (Table 1) is located in northern Fennoscandia (Fig. 1) in a 40 x 5 km NNW-trending belt of Cu(-Au) mineralisation (*cf.* Frietsch *et al.*, 1997, Carlon, 2000). The belt is spatially associated with a regional-scale fault/shear zone and is underlain by terrestrial and marine intermediate to felsic metavolcanic rocks of the *ca.* 1910-1880 Ma Porphyry Group (Table 2; *cf.* Skiöld and Cliff, 1984; Skiöld, 1987; *ibid*). Cu(-Au) mineralising events occurred within the belt *ca.* 1870 and 1770 Ma and the genesis of the mineralisation is interpreted to be related to large-scale Na-Cl-rich fluid fluxes synchronous with granitic igneous activity *ca.* 1890-1860 Ma and *ca.* 1830-1770 Ma (*cf.* Skiöld, 1987; Frietsch *et al.*, 1997, Carlon, 2000).

Aitik is hosted by intermediate composition volcanic and clastic sedimentary rocks that have been metamorphosed to amphibolite facies; *ca.* 1870 Ma porphyritic quartz monzodiorite occurs in the footwall (*cf.* Frietsch *et al.*, 1997; Wanhainen *et al.*, 2003). Mineralisation is controlled by a NNW-trending fault zone and exhibits multiple phases of brittle-ductile deformation (*ibid*; Carlon, 2000). Chalcopyrite and pyrite and minor magnetite, pyrrhotite, bornite, chalcocite and molybdenite occur as disseminations and stringers in garnet-bearing biotite schist and gneiss and muscovite schist (*cf.* Frietsch *et al.*, 1997; Wanhainen and Martinsson, 1999; Carlon, 2000; Wanhainen *et al.*, 2003). Stockwork quartz-chalcopyrite-pyrite veins extend into the underlying quartz monzodiorite (Wanhainen *et al.*, 2003). Barite-magnetite-actinolite-chalcopyrite-pyrite veins occur locally. Native gold and gold alloys occur with chalcopyrite and pyrite and are intergrown with non-sulphide minerals (*ibid*). Gangue minerals include quartz, barite, fluorite, calcite, tourmaline, scapolite and apatite. Pervasive sericitisation, scapolitization and tourmalinization have affected host rocks

(Table 3; *cf.* Frietsch *et al.*, 1997). Mineralisation is associated with extensive biotite and sericite alteration (Wanhainen *et al.*, 2003). K-feldspar and epidote alteration occurs along fault zones that bound the ore zone (*ibid*).

Chalcopyrite mineralisation was deposited prior to bornite from moderate temperature (140-373 ^{o}C), highly saline (31-37 eq. wt% NaCl) fluids (Fig. 4; Table 4; Wanhainen *et al.*, 2003). Bornite mineralisation was deposited from moderate temperature (100-222 ^{o}C), less saline (18-27 eq. wt% NaCl + $CaCl_2$) fluids (*ibid*). Sulphur isotope values for pyrite and chalcopyrite range from -3.4 to +3.3 ‰and from +6.7 to +13.8 ‰for barite (Fig. 5; Frietsch *et al.*, 1995).

The Aitik deposit has been interpreted to be genetically related to hydrothermal fluids exsolved from the underlying syntectonic quartz monzodiorite (Monro, 1988) with a contribution to the fluid from evaporitic sources (Wanhainen *et al.*, 2003). Based on isotopic data (Fig. 5) sulphur is interpreted to have been derived from a magmatic source, either from magmatic fluid or by dissolution of igneous sulphides (Frietsch *et al.*, 1995). The high calcium content (in addition to Na) of fluid inclusions related to the main chalcopyrite mineralisation is interpreted to indicate at least some fluid was derived from evaporitic sources (Wanhainen *et al.,* 2003). Highly saline hydrothermal fluids derived from an evaporitic source have also been invoked to explain pervasive scapolite and albite alteration that is widespread in the region (Frietsch *et al.,* 1997).

Candelaria (Punta del Cobre Belt) – An approximately 5 x 20 km belt of IOCG and iron deposits, that includes Candelaria (Table 1) and deposits within the Punta del Cobre district, occurs along the eastern margin of the coastal batholith, near Copiapó, Chile (Fig. 1; *cf.* Marschik and Fontboté, 2001). The deposits are located east of, but proximal to, the Atacama fault zone a >1000 km-long, arc-parallel, strike-slip fault system (*cf.* Scheuber and Andriessen, 1990). The area is underlain by an Early Cretaceous continental arc and marine back-arc basin terrane made up of andesitic volcanic flows and volcaniclastic rocks overlain by limestones and minor evaporites (Table 2; *cf.* Ullrich and Clark, 1999; Marschik *et al.*, 2000; Marschik and Fontboté, 2001). Significant iron oxide-Cu-Au(-Zn-Ag) mineralisation, including the large Candelaria deposit, occurs at the intersection of northwest-trending brittle faults with the contact between massive relatively impermeable volcanic rocks and overlying porous and permeable volcaniclastic rocks (*cf.* Marschik *et al.*, 2000; Marschik and Fontboté, 1996, 2001). Marschik *et al.* (2000) and Marschik and Fontboté (2001) suggest ore-forming fluids were focussed along faults and formed sub-vertical orebodies in low permeability strata and lens-shaped concordant bodies in permeable horizons.

Copper mineralisation occurs as veins and hydrothermal breccia infill, or overprints massive magnetite replacement bodies (*cf.* Marschik *et al.*, 2000; Marschik and Fontboté, 1996, 2001). Ore consists mainly of magnetite and/or hematite, chalcopyrite and pyrite, locally abundant pyrrhotite and sphalerite and trace amounts of molybdenite and arsenopyrite. Gold occurs as inclusion in chalcopyrite, in fractures within pyrite and as Hg-Au-Ag alloy. Malachite, chrysocolla, chalcocite and covellite occur in poorly developed supergene oxidation and enrichment zones (Sillitoe and Clark, 1969). Gangue consists mainly of quartz and anhydrite at Candelaria and calcite and/or quartz in the Punta del Cobre district (*cf.* Marschik *et al.*, 2000; Marschik and Fontboté, 1996, 2001).

Alteration and mineralisation in the Punta del Cobre belt is multiphase with numerous cross-cutting relationships that can be divided into six general stages (*cf.* Ullrich and Clark, 1999; Marschik and Fontboté, 1996, 2001). Stage 1: large-scale pervasive albitization. Stage 2: intense biotite-quartz-magnetite ± K-feldspar alteration associated with emplacement of large magnetite-quartz bodies; coeval specular hematite formed in dilational zones but is replaced by later magnetite. Stage 3: Ca-Na metasomatism (calcic amphibole, albite) ± K-feldspar associated with the main ore stage (Cu-Au); sodic scapolite occurs in sedimentary strata (limestones and metaevaporites of the Abundancia Formation) that overlie the volcanic rocks which host mineralisation. Stage 4: epidote-chlorite alteration. Stage 5: hematite-calcite-chalcopyrite. Stage 6: anhydrite and calcite-chlorite alteration (Table 3).

Mineralisation was broadly contemporaneous with regional uplift and emplacement of calc-alkaline diorite to quartz monzonite of the *ca.* 119-97 Ma Copiapó Batholith (Ullrich and Clark, 1999; Marschik and Fontboté, 2001) and likely occurred at 2 to 3 km depth (Marschik and Fontboté, 2001). Intrusion of the batholith created an extensive contact metamorphic aureole and peak contact metamorphism occurred between stages 2 and 3 (Marschik and Fontboté, 2001). The temperature of oxide mineralisation (stage 2) is interpreted to be 500-600 ^{o}C (*cf.* Marschik *et al.*, 2000). Fluid inclusions in quartz associated with main stage Cu-Au mineralisation are hypersaline and CO_2-rich with homogenisation temperatures from ~ 330 to 440 ^{o}C (Fig. 4; Table 4; Ullrich and Clark, 1999; Marschik *et al.*, 2000). Fluid inclusions in late stage calcite have homogenisation temperatures of ≤ 236 °C. Sulphur isotopes values for main ore stage (stage 3) sulphides at Candelaria range from -1.3 to + 5.7 ‰and from 14.5 to 17.5 ‰for anhydrite associated with the chalcopyrite (Fig. 5; Ullrich and Clark, 1999; Marschik and Fontboté, 2001). δ^{34}S values for late stage chalcopyrite at Candelaria range from +0.8 to +7.2 ‰(*ibid*). Oxygen isotope values for fluid in equilibrium with ore stage quartz at Candelaria range from approximately 6 to 9 ‰(Fi g. 5; *cf.* Marschik and Fontboté, 2001). $\delta^{18}O_{fluid}$ values in equilibrium with late stage calcite range from -5.4 to +1.3, +4.6 to +7.7 and -2.8 to + 4.7 respectively for the Candelaria, Carola and Santos/Socavón Rampa IOCG deposits in the Punta del Cobre district (*ibid*).

Magmatic and non-magmatic fluids are interpreted to have been involved in ore formation at Candelaria and deposits in the Punta del Cobre district (*cf.* Ullrich and Clark, 1999; Marschik *et al.*, 2000; Marschik and Fontboté, 2001). A magmatic fluid contribution to the hydrothermal system is based on calculated fluid oxygen isotope values, the presence of hypersaline, CO_2-rich and saline fluid inclusions, the oxidised nature of the ore fluid as indicated by early formed hematite, coeval ages for mineralisation and intrusive activity, similar Pb isotope values for sulphides, volcanic and intrusive rocks, and similar initial ^{187}Os/^{188}Os ratios for magnetite, sulphides and magmatic magnetite (*cf.* Marschik *et al.*, 2000; Marschik and Fontboté, 2001). The presence of non-magmatic fluid (basinal brines, formation or meteoric waters) is inferred from the low calculated δ^{18}O values for fluid in equilibrium with calcite (*ibid*) and from sulphur isotopic data that indicate early, reduced, Cu-depositing fluids with near magmatic S compositions ($\delta^{34}S_{fluid}$ = -1.3 to +5.7 ‰in early stage 3) w ere replaced by more oxidised and probably evaporite-sourced brines ($\delta^{34}S_{fluid}$ = 11.7 to 16.8 ‰in late stage 3; $\delta^{34}S_{fluid}$ = 13 to 20.2 ‰in stage 5 ; Ullrich and Clark (1999). Thus, suggesting the main ore forming event may have been dominated by the mixing of sulphur-bearing fluids from magmatic and evaporitic sources (*ibid*).

Salobo – The Carajás metallogenic province in Brazil (Fig. 1) is host to Salobo (Table 1) and a number of other Fe oxide-Cu-Au (± Mo,Ag,U,REE) deposits including Igarapé Bahia, Alemão, Cristalino, Sossego and Águas Claras (*cf.* Tazava and Olivera, 2000). The deposits occur within the Carajás Basin which is underlain by Late Archaean (*ca.* 2750 Ma) volcano-sedimentary rocks of the Itacaiúnas Supergroup unconformably overlain by Archaean (*ca.* 2680 Ma) marine siliciclastic rocks of the Águas Claras Formation or Rio Fresco Group, all of which were likely deposited in an extensional continental environment (*cf.* DOCEGO, 1988; Lindenmayer, 1990; Machado *et al.*, 1991; Pinheiro, 1997). Rocks within the basin were intruded by *ca.* 2750 Ma mafic-ultramafic rocks (Luanga complex), *ca.* 2740 Ma granitoids and diorites (Plaquê Suite), *ca.* 2570 Ma granite (Old Salobo granite/Estrela granitoid complex) and *ca.* 1880 Ma granitoids (Carajás Suite; (*cf.* Machado *et al.*, 1991; Barros *et al.*, 1997). The IOCG deposits are enriched in LREE, Co, Ni, Pb, Zn, As, Bi, W and U, are associated with abundant magnetite and/or hematite and are structurally controlled (*cf.* Requia and Fontboté, 2000; Souza and Vieira, 2000; Tazava and de Olivera, 2000, Ronzê *et al.*, 2000).

The Salobo deposit is situated at the western termination of the Cinzento Transcurrent System and is located within a lens of Itacaiúnas Supergroup bounded by basement gneiss (Table 2; *cf.* Pinheiro, 1997; Souza and Vieira, 2000). The deposit is hosted by metagreywackes that contains lenses or layers of amphibolite (*cf.* Requia and Fontboté, 2000) and are cut by *ca.* 2573 (Old Salobo Granite) and *ca.* 1880 Ma (Young Salobo Granite) granitoids and *ca.* 550-560 Ma diabase dykes (*cf.* Machado *et al.*, 1991; Souza and Vieira, 2000). Cross-cutting relationships demonstrate mineralisation at Salobo is post metamorphic and isotopic dating, based on Re-Os and Pb-Pb analyses of sulphides and magnetite, indicates an age of *ca.* 2500 Ma (*cf.* Machado *et al.*, 1991; Souza and Vieira, 2000), roughly coeval with emplacement of Old Salobo Granite.

Mineralisation occurs along a NNW-trending shear zone (*ibid*) and consists dominantly of steeply-dipping lenses of: 1) abundant early magnetite with minor hematite and local graphite and 2) less abundant, later, chalcopyrite, bornite and chalcocite (Requia and Fontboté, 2000). Copper sulphides also occur in veins cutting iron-rich rocks. Accessory minerals include hematite, molybdenite, ilmenite, uraninite, digenite and covellite; native gold occurs as inclusions in cobaltite, safflorite and copper sulphides, or interstitial to magnetite and chalcopyrite grains (*cf.* Machado *et al.*, 1991; Souza and Vieira, 2000).

Alteration around the deposit includes early Na-metasomatism (Na-plagioclase) overprinted by extensive K-metasomatism (K-feldspar, biotite; Requia and Fontboté, 2000). Intense K-metasomatism is spatially associated with the main ore zone (Table 3; Requia and Fontboté, 2000). Propylitic alteration overprints earlier phases and is characterized by the infiltration of Ca-bearing fluids. Veinlets with quartz, stilpnomelane, fluorite, allanite, chalcopyrite, molybdenite, cobaltite and gold formed during propylitic alteration (Lindenmayer, 1990; Souza and Vieira, 2000).

Chloritization associated with brittle shearing accompanied the propylitic alteration (Lindenmayer, 1990; Souza and Vieira, 2000).

The interaction of a mixture of magmatic and connate brines with a more oxidized fluid is suggested as an ore forming mechanism (*cf.* Requia and Fontboté, 2000). Sulphur isotope values of 0.2 to 1.6 ‰suggest a dominantly magmatic source for sulphur (Fig. 5 ; *ibid*).

Ernest Henry – The Ernest Henry deposit (Table 1), like the magmatic end-member Lightning Creek, Osborne and Eloise deposits, occurs in the Cloncurry district of Australia (Fig. 1). The deposit is hosted by strata of Cover Sequence 2 that include *ca.* 1740 Ma plagioclase-phyric volcanic rocks intercalated with siliciclastic, calc-silicate-rich (scapolite-bearing) and graphitic metasedimentary rocks; *ca.* 1660 Ma metadiorite also occurs in this area (Table 2; *cf.* Mark *et al.*, 2000). Host rocks have been affected by: 1) Na-Ca alteration characterised by hematite-bearing albite and 2) disseminated biotite-magnetite alteration and garnet-K feldspar-biotite alteration (Ryan, 1998; Mark *et al.*, 2000). Mineralisation (*ca.* 1510-1500 Ma) was roughly coeval with emplacement of the Williams and Naraku batholiths *ca.* 1540-1500 Ma (Blake *et al.*, 1990; Ryan, 1998; Mark *et al.*, 2000) and is associated with extensive K-feldspar-hematite alteration (Table 3; Ryan, 1998; Mark *et al.*, 2000). Post ore alteration is dominated by carbonate (Mark *et al.*, 2000).

The location of the orebody is structurally controlled and occurs within volcanic rocks that were brecciated during reverse fault movement along bounding shear zones (Mark *et al.*, 2000). Breccia infill is made up dominantly of a magnetite-carbonate-sulphide assemblage but also contains biotite, K-feldspar, hematite, garnet, barite, fluorite and quartz (Ryan, 1998; Mark *et al.*, 2000). Primary and supergene mineralisation occur. Chalcopyrite and pyrite are the dominant sulphide minerals in the primary zone and there is a strong correlation between copper and gold (*ibid*). The supergene zone extends to 150 m depth and contains chalcocite, bornite, secondary chalcopyrite and native copper (*ibid*). In this zone gold is largely decoupled from copper, magnetite has been oxidised to hematite and secondary calcite and siderite are ubiquitous. Anomalous amounts of cobalt, molybdenum, uranium, REE, arsenic, fluorite and barium are associated with the copper mineralisation in both zones (*ibid*).

Ore fluids were high temperature (~ 400-450 ^{o}C) and high salinity (>26 wt % NaCl eq; Fig. 4; Table 4; Mark *et al.*, 2000). Stable isotope data are compatible with a significant magmatic contribution to ore-forming fluid(s) and/or sulphur ($\delta^{34}S_{chalcopyrite}$ = -1 to 4 ‰; $\delta^{18}O_{fluid}$ = 8-11 ‰; Fig. 5; *ibid*). However, the wide range of elements enriched in Ernest Henry ore (Table 1) suggest involvement of more than one fluid (*ibid*). Cu and Au (+ Fe, Ba) were probably carried as chloride complexes and pyrite and chalcopyrite (plus magnetite and barite) were likely deposited via a mechanism(s) that involved fluid mixing and/or cooling during brecciation (*ibid*).

D.5 Discussion

IOCG deposits do not appear to form in any one specific geological setting (Tables 1 to 4 and references therein). They are found in a variety of environments that are able to provide basic ingredients for formation, i.e. hot, saline (oxidized) fluid(s), permeable flow paths (e.g. faults, shear zones), source(s) of metals and sulphur (either in original fluid or leached from rocks en route), mechanism(s) to drive fluid flow (e.g. heat, tectonics, gravity, density gradients), and means to precipitate ore minerals (e.g. fluid mixing, cooling, changes in pH and/or fO_2). Thus, unlike other types of mineral deposits which form in a restricted range of geological environments (e.g. volcanic-associated massive sulphide, porphyry copper, skarn, Mississippi Valley-type) IOCG deposits can form in a variety of environments ranging from those containing abundant igneous rocks (e.g. Lightning Creek, Olympic Dam) to those with large volumes of igneous and sedimentary rocks (e.g. Cloncurry district) to those dominated by sedimentary rocks (e.g. Wernecke Breccia). The type of alteration and mineralisation in any IOCG district or at any one deposit will be affected by, for instance, the abundance of fluid of various types (e.g. magmatic, metamorphic, basinal, meteoric), the composition of host strata (e.g. igneous rocks, evaporites), the degree of interaction between fluid and rocks along the fluid pathway (e.g. function of fluid temperature, pH, fO_2) and the degree of permeability of the host rocks (e.g. presence of major structures to act as fluid paths, permeable lithologies; *cf.* Williams and Blake, 1993; Barton and Johnson, 1996, 2000). Thus, IOCG systems can be viewed as forming a spectrum of deposits ranging from those dominated by magmatic environments and fluids to those where non-magmatic fluids dominate.

D.5.1 Comparison of IOCG systems

Tectonic setting – The tectonic and structural settings at the time of mineralisation are poorly understood for most IOCG districts, however, in general this deposit type commonly occurs in extensional settings (*cf.* Hitzman *et al.*, 1992). For example, Candelaria, Aitik and possibly Olympic Dam were emplaced within arc-back arc systems above subducting slabs (*cf.* Ullrich and Clark, 1999; Marschik *et al.*, 2000; Marschik and Fontboté, 2001; Gorbatschev and Bogdanova, 1993; Ferris *et al.*, 2002) and deposits within the Carajás region likely formed in a continental rift environment (*cf.* DOCEGO, 1988; Lindenmayer, 1990; Machado *et al.*, 1991; Pinheiro, 1997). Wernecke Breccia-related IOCG mineralisation may also have formed in an extensional environment. WSG sediments that host Wernecke Breccia were likely deposited in a rift-related basin (Thorkelson, 2000). Mafic to intermediate sills and dykes of the BPRI that cut the WSG have major and trace element chemistry compatible with a rift setting (Thorkelson, 2000; Thorkelson *et al.*, 2001) as do the Slab volcanics (Laughton, 2004). However, further work is required to determine if an extensional setting still existed at the time of emplacement of Wernecke Breccia.

The deposits are largely coeval with tectonic activity, are spatially related to regional-scale fault and/or shear zones and also exhibit structural control on a smaller scale (Table 2 and references therein). For example, the Ernest Henry and Olympic Dam deposits occur in dilational zones (*cf.* Ryan, 1998; Mark *et al.*, 2000; Roberts and Hudson, 1983; Reeve *et al.*, 1990; Oreskes and Einaudi, 1990; Cross *et al.*, 1993; Oreskes and Hitzman, 1993; Johnson and Cross, 1995; Haynes *et al.*, 1995; Reynolds, 2000; Skirrow *et al.*, 2002), the ore zone at Candelaria is located at the intersection of shear/fault zones with a lithological contact between relatively impermeable rocks and overlying permeable strata (*cf.* Marschik *et al.*, 2000; Marschik and Fontboté, 1996, 2001) and Wernecke Breccia mineralisation occurs in structurally weak, permeable zones such as fractured fold hinges (Section A).

Host rocks – IOCG systems are not restricted to particular types of host rocks and can occur in intrusive, volcanic, sedimentary and ironstone hosts (Table 2 and references therein). Magmatic and hybrid IOCG systems are temporally and, generally, spatially related to intrusive rocks (*ibid*). Non-magmatic IOCG systems are not temporally related to igneous rocks but may be spatially related, e.g. Wernecke Breccia. A single area may host various types of IOCG systems. For example the Cloncurry district is host to magmatic (Lightning Creek, Eloise, Osborne) and hybrid (Ernest Henry) IOCG systems (Tables 1 to 4 and references therein), reflecting different fluid pathways and sources present in the area. The Tennant Creek Inlier hosts non-magmatic IOCG systems (*cf.* Skirrow and Walshe, 2002). The presence of evaporites in the host rock package appears to be important (*cf.* Haynes *et al.*, 1995; Barton and Johnson, 1996, 2000; Hitzman, 2000). For example, Wernecke Breccia is most abundant proximal to a metahalite horizon (Section A). New Br/Cl data for the Cloncurry area suggests a variable but significant contribution of evaporitic halite to the fluid locally (Williams *et al.*, in progress).

Alteration – Most IOCG systems are associated with regional, pervasive alteration (Table 3 and references therein). For example, extensive sodic-calcic alteration occurs in the Cloncurry district, widespread sodic alteration occurs in the Aitik and Candelaria regions, and Salobo, Wernecke Breccia and Redbank are associated with abundant sodic and or potassic alteration (*ibid*). Extensive hematite-sericite-chlorite alteration occurs in the Olympic Dam area (*cf.* Reynolds, 2000). In general, late alteration in IOCG systems is dominated by carbonate (Table 3 and references therein).

Fluid composition, temperature and salinity – IOCG deposits were formed by moderate to high temperature, complex brines largely made up of H_2O-NaCl-$CaCl_2$ with variable amounts of K, Mn, Fe, Sr, Ba, Si, Zn, CH_4, CO_2 and/or Cu (*cf.* Table 4 and references therein; Barton and Johnson, 1996, 2000; Williams *et al.*, in progress). The fluids were varied and different brine compositions occur within a deposit, between deposits in a given district, and between districts (*ibid*). The sulphur content of the brines was typically low and oxygen fugacity was variable (*ibid*).

In general, early-stage alteration in magmatic IOCG systems formed from high temperature, high salinity fluids (Fig. 4). For example, pre-mineralisation quartz-magnetite veins at Lightning Creek formed from fluids with a temperature $> 500\ ^{o}C$ and a salinity of 33 to 55 weight % (NaCl eq.) and early silica flooding at Osborne was produced by $> 450\ ^{o}C$, highly saline (60-70 wt % NaCl eq) fluid (Tables 1 to 4, Figs. 4 & 5 and references therein). Early phases in some hybrid IOCG systems also formed from fluids with high temperatures and salinities, e.g. pre-mineralisation biotite-quartz-magnetite alteration at Candelaria and early magnetite at Olympic Dam (*ibid*). Fluid temperature and salinity generally decrease towards later stages and differences between IOCG types are less apparent. For example, mineralisation stage fluid temperature varies from < 200 to 450, 100 to 450 and 150 to 350 ^{o}C and salinity ranges from 15 to

47, 18 to 37 and 3 to 42 weight % (NaCl eq.) respectively for the magmatic, hybrid and non-magmatic IOCG systems listed in Table 1(*ibid*).

Oxygen isotopic composition of fluid – Calculated $\delta^{18}O_{fluid}$ values for magmatic IOCG systems generally overlap values for magmatic fluid (5-10 ‰; Fig. 5 and references therein). For example, $\delta^{18}O_{fluid}$ = 7 to 10, 5 to 12 and 5 to 10 ‰ respectively for Lightning Creek, Osborne and Eloise (*ibid*). Fluids related to hybrid IOCG systems show a wider range of isotopic compositions and commonly vary temporally from early magmatic values to those reflecting the later input of other fluids (Fig. 5). For example, $\delta^{18}O_{fluid}$ = 8 to 10 ‰ and -2 to +5 ‰ for fluid in equi librium with early magnetite and later hematite, respectively, at Olympic Dam (Oreskes and Hitzman, 1993). At Candelaria $\delta^{18}O_{fluid}$ = 6 to 9 ‰ for ore stage mineralisation and $\delta^{18}O_{fluid}$ = -5 to +2 ‰ for later mineralisation (Marschik and Fontboté, 2001). Non-magmatic IOCG systems have a wide range of fluid oxygen isotopic compositions, for example $\delta^{18}O_{fluid}$ = 0 to +16 and -7 to +14 respectively for Redbank and Wernecke Breccia (Knutson *et al.*, 1979; Section C).

Sulphur isotopes – Sulphur isotope values for magmatic IOCG systems are generally close to those of mantle-derived sulphur (Fig. 5), e.g. $\delta^{34}S$ = 0 to 2.3 ‰ at the Eloise deposit (*cf.* Baker, 1998). Some hybrid IOCG systems also have sulphur isotopic compositions close to those of mantle-derived sulphur but others show a wider range. Non-magmatic IOCG systems have a wide range of sulphur isotope values, for example $\delta^{34}S$ values for chalcopyrite and pyrite from Wernecke Breccia range from -12 to +14 ‰ (Section C) and $\delta^{34}S$ values for Redbank vary from -1 to +16 ‰. In general, the large hybrid IOCG systems in Figure 5 do not show as large a variation in sulphur isotopic compositions as the non-magmatic IOCG systems, although Eldridge and Danti (1994) report one value of -47 ‰ from Olympic Dam. However, at least one small hybrid IOCG system has a wide range of sulphur isotope compositions, i.e. pyrite and chalcopyrite from Monakoff, which is interpreted to have formed from the mixing of granite-derived and metaevaporite-derived fluids returned $\delta^{34}S$ values of -10 to +12 ‰ (*cf.* Davidson *et al.*, 2002).

Deposit size and grade – Mineralisation occurs dominantly as disseminations, veins and breccia infill in all IOCG types, however, the size of deposits varies widely (Table 1 and references therein). In general, based on those listed in Table 1, magmatic IOCG deposits tend to be small and higher grade. For example, Eloise has a resource of 3.2 million tonnes of 5.8 % Cu, 1.5 g/t Au and 19 g/t Ag (*cf.* Baker and Laing, 1998).

Hybrid IOCG deposits are generally low grade but have the potential to be very large, for example Olympic Dam is reported to have had a pre-mining resource of 2320 million tonnes of 1.3 % Cu, 0.5 g/t Au, 2.9 g/t Ag and 0.4 kg/t U_3O_8 (*cf.* Reeve *et al.*, 1990). Non-magmatic IOCG deposits can be small to medium in size and high or low grade (Table 1). Most IOCG systems contain a variety of associated elements including cobalt, molybdenum, REE and bismuth (Table 1 and references therein). Eloise and Olympic Dam are reported to contain elevated levels of Ni which may be the reflection of primitive magmatic contribution (*cf.* Baker *et al.*, 2001; Reynolds, 2000).

Mechanisms that would cause the precipitation of mineralisation include: fluid mixing, possibly with less saline, oxidised, and/or sulphur-bearing fluid (*cf.* Table 4 and references therein; Barton and Johnson, 1996, 2000; Williams *et al.*, in progress); CO_2 unmixing (e.g. Osborne; *cf.* Adshead, 1995; Mustard *et al.*, 2003); and the sulphidation of pre-existing iron silicates (e.g. Eloise; Baker, 1998). The presence of abundant fluid (i.e. magmatic + other sources) and a magmatic source of heat ± metals ± sulphur (*cf.* Barton and Johnson, 2000) may account for the large tonnage potential of hybrid systems (Fig. 7). During fluid circulation Fe and base metals could be leached from host rocks along the fluid pathway. For example, in the Cloncurry district K, Fe and Cu were leached from host rocks during widespread sodic-calcic alteration creating a fluid with the potential to contribute to the overall metal budget of the district (*cf.* Williams and Blake, 1993; De Jong and Williams, 1995; Adshead *et al.*, 1998; Baker, 1998; Rotherham *et al.*, 1998; Mark *et al.*, 1999; Perring *et al.*, 2000). Metals and/or sulphur could also be supplied by magmatic fluid, for example PIXE analyses of fluid inclusions from the Lightning Creek deposit indicates a copper content of up to 1 weight % (Perring *et al.* 2000; Williams *et al.* in progress). In magmatic environments fluid flow tends to be focussed into permeable zones and fluid generally cools rapidly as it migrates away from the heat source (*cf.* Barton and Johnson, 2000). The rapid cooling causes precipitation of sulphides and the focussed flow concentrates precipitation in a relatively limited area. The amount of sulphides precipitated would be limited largely by the sulphur content of the fluid (*ibid*). In end member magmatic IOCG systems fluid is derived only from plutons and may be less abundant leading to smaller more focussed deposits. In non-magmatic environments fluid flow may not be focussed and widely disseminated sulphides may result, rather than concentrated sulphide deposits (*ibid*). Metal concentrations may also be lower because there is no contribution from magmatic fluid. In all cases metals present in the fluid will be lost if the sulphur content is low and/or if there is no favourable trap (*ibid*).

D.5.2 Relationship to other deposit types

Non-magmatic end-member IOCG systems have potential links to other sediment-hosted copper systems, e.g. Tsumeb and Mount Isa. Wernecke Breccia, Tsumeb and Mount Isa all occur in rift-related settings (*cf.* Thorkelson, 2000; Frimmel *et al.*, 1996; O'Dea *et al.*, 1997; Betts *et al.*, 1998). All are hosted by thick sedimentary packages that contain(ed) evaporite horizons and they are not temporally associated with igneous intrusions (*cf.* Thorkelson *et al.*, 2001b; Frimmel *et al.*, 1996; Heinrich *et al.*, 1995). Mineralisation is structurally controlled and at least partly synchronous with tectonic activity (*cf.* Section A; Frimmel *et al.*, 1996; Perkins, 1984; Swager, 1985). In each region the onset of deformation may have been a driving force for the movement of hydrothermal fluid. All were formed from hot, high salinity brines that indicate the involvement of evaporites and are likely evolved basinal/metamorphic waters (*cf.* Section C; Hughes, 1987; Chetty and Frimmel, 2000; Heinrich *et al.*, 1993). Sulphur isotopic compositions for sulphides have a wide range of values in each area: -12 to +14; -8 to +26 and +8 to +21 ‰ respectively for Wernecke Breccia, Tsumeb and Mount Isa, indicating a dominantly non-magmatic source for sulphur (Section C; Hughes, 1987; Andrew *et al.*, 1989). Host-strata and mafic rocks are likely sources of copper in each area (*cf.* Section C; Chetty and Frimmel, 2000; Heinrich *et al.*, 1995).

End-member magmatic IOCG deposits are similar to some porphyry Cu(-Au) deposits (*cf.* Pollard, 2000; Sillitoe, 2003). For example Lightning Creek is associated with K-rich, magnetite-series metaluminous intrusive rocks that are similar to intrusive rocks associated with Cu-Au porphyry systems (Pollard *et al.*, 1998; Perring *et al.*, 2000). Pollard (2000) suggests the development of the two styles of mineralisation is related to the composition of the granitoids and the evolution of the hydrothermal fluids, i.e. boiling of magmatic-derived H_2O-salt fluids can lead to porphyry deposits and unmixing of magmatic-derived H_2O-CO_2-salts fluids can lead to IOCG deposits. The variable CO_2 content of the magmas is suggested as a key factor in the evolution of the mineralising system.

D.6 Conclusions

Wernecke Breccia-associated IOCG mineralisation was formed by low to moderate temperature (185-350 ^{o}C), moderate to high salinity (24-42 wt % NaCl eq.) brines that were likely derived from basinal/metamorphic fluids. Magmatic waters are considered less likely as a fluid source because isotopic data for the breccias do not have a magmatic signature and spatially associated mafic to igneous rocks are significantly older (i.e. *ca*. 1710 vs. 1600 Ma) thus ruling out a genetic connection. Metals and sulphur were probably derived from host strata and fluids circulated via tectonic (and/or gravity) processes. This suggests that IOCG systems can form in non-magmatic environments that are able to provide basic ingredients for formation, i.e. hot, saline (oxidized) fluid(s), permeable flow paths (e.g. faults, shear zones), source(s) of metals and sulphur (either in original fluid or leached from rocks en route), mechanism(s) to drive fluid flow (e.g. heat, tectonics, gravity, density gradients), and means to precipitate ore minerals (e.g. fluid mixing, cooling, changes in pH and/or fO_2).

A division of the broad class of IOCG deposits in to end-member magmatic and non-magmatic IOCG systems with hybrid IOCG systems in between would allow their attributes to be more narrowly defined and these could be used in exploration models. For example, large low grade deposits, e.g. Olympic Dam, Aitik, Ernest Henry, are hybrid IOCG systems. Magmatic end-member IOCG deposits are generally smaller but can be higher grade e.g. Osborne, Eloise. Non-magmatic end-member IOCG deposits are small to medium in size and low to high grade (Table 1).

End-member magmatic IOCG deposits such as Lightning Creek have similarities to some porphyry deposits (*cf.* Pollard, 2000). Non-magmatic end-member IOCG deposits share characteristic with some sediment hosted Cu deposits, e.g. Tsumeb, Mount Isa (& Zambian Copper Belt?). This suggests that the range of IOCG deposits may form a link between intrusive- and sedimentary-related deposits, thus, information obtained through the study of IOCG deposits may also be useful in the search for other types of deposits.

D.7 References:

Adshead ND (1995) Geology, alteration and geochemistry of the Osborne Cu-Au deposit, Cloncurry district, NW Queensland, Australia. PhD Thesis, James Cook University, Australia.

Adshead ND, Voulgaris P and Muscio VN (1998) In: Berkman DA and Mackenzie DH (eds.) Geology of Australian and Papua New Guinean Mineral Deposits, The Australasian Institute of Mining and Metallurgy, Melbourne, pp. 793-799.

Ahmad M, Wygralak AS and Ferenczi PA (1999) Gold deposits of the Northern Territory: Northern Territory Geological Survey Report 11, 95p.

Andrew AS, Heinrich CA, Wilkins RWT and Patterson DJ (1989) Sulphur isotope systematics of copper ore formation at Mount Isa, Queensland. Economic Geology 84:1614-1626.

Baker T (1996) The geology and genesis of the Eloise Cu-Au deposit, Cloncurry district, NW Queensland, Australia. PhD Thesis, James Cook University, Australia.

Baker T (1998) Alteration, mineralization, and fluid evolution at the Eloise Cu-Au deposit, Cloncurry district, northwest Queensland, Australia. Economic Geology 93:1213-1236.

Baker T and Laing WP (1998) Eloise Cu-Au deposit, East Mount Isa block: structural environment and structural controls on ore. Australian Journal of Earth Sciences 45:429-444.

Baker T, Perkins C, Blake KL and Williams PJ (2001) Radiogenic and stable isotope constraints on the genesis of the Eloise Cu-Au deposit, Cloncurry district, northwest Queensland. Economic Geology 96:4:723-742.

Barros CE de M, Dall'Agnoll R, Barbey P and Boullier, AM (1997) Geochemistry of the Estrela Granite complex, Carajás region, Brazil: and example of an Archaean A-type granitoid. Journal of South American Earth Sciences 10:321-330.

Barton MD and Johnson DA (2000) Alternative brine sources for Fe-Oxide(-Cu-Au) systems: implications for hydrothermal alteration and metals. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 43-60.

Barton MD and Johnson DA (1996) Evaporitic-source model for igneous-related Fe oxide-(REE-Cu-Au-U) mineralisation. Geology 24, 3:259-262.

Barton MD, Johnson DA and Zurcher L (2000) Phanerozoic iron oxide (-REE-Cu-Au-U) systems in southwestern North America and their origins. In: Roberts MD and Fairclough MUSCOVITE (eds), Fe-oxide-Cu-Au Deposits: A Discussion of Critical Issues and Current Developments. EGRU contribution 58, James Cook University, pp 5-11.

Beardsmore TJ, Newbery SP and Laing WP (1988) The Maronan Supergroup: an inferred early volcanosedimentary rift sequence in the Mount Isa Inlier, and its implications for ensialic rifting in the Middle Proterozoic of northwest Queensland. Precambrian Research 40/41:487-507.

Bell RT (1978) Breccias and uranium mineralisation in the Wernecke Mountains, Yukon Territory – a progress report. Current Research, Part A, Geological Survey of Canada, Paper 78-1A, pp317-322.

Bell RT (1986a) Megabreccias in northeastern Wernecke Mountains, Yukon Territory. Current Research, Part A, Geological Survey of Canada, Paper 86-1A, pp 375-384.

Bell RT (1986b) Geological map of north-eastern Wernecke Mountains, Yukon Territory. Geological Survey of Canada, Open File 1027.

Bell TH, Perkins WG and Swager CP (1988) Structural controls on development and localization of syntectonic copper mineralisation at Mount Isa, Queensland. Economic Geology: 83:1:69-85.

Bell RT and Delaney GD (1977) Geology of some uranium occurrences in Yukon Territory. Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, pp 33-37.

Betts PG, Lister GS and O'Dea MG (1998) Asymmetric extension of the Middle Proterozoic lithosphere, Mount Isa Inlier, Queensland, Australia. Tectonophysics 296:293-316.

Betts PG, Giles D and Lister GS (2003) Tectonic environment of shale-hosted massive sulphide Pb-Zn-Ag deposits of Proterozoic northeastern Australia. Economic Geology 98: 3: 557-576.

Blake DH and Stewart AJ (1992) Stratigraphic and tectonic framework, Mount Isa Inlier. In: Stewart AJ and Blake DH (eds.) Detailed studies of the Mount Isa Inlier, Bureau of Mineral Resources Bulletin 243, pp 1-11.

Blake DH, Etheridge MA, Page RW, Stewart AJ, Williams PR and Wyborn LAI (1990) Mount Isa Inlier – regional geology and mineralisation. In: Hughes FE (ed.) Geology of the mineral deposits of Australia and Papua New Guinea, Australasian Institute of Mining and Metallurgy: Melbourne, pp 915-925.

Brideau M-A, Thorkelson DJ, Godin, L and Laughton JR (2002) Paleoproterozoic deformation of the Racklan Orogeny, Slats Creek (106D/16) and Fairchild Lake (106C/13) map areas, Wernecke Mountains, Yukon. In: Emond DS, Weston LH and Lewis LL (eds.) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 65-72.

Brooks M, Baker T and Hunt J (2002) Alteration zonation, veining and mineralisation associated with the Wernecke Breccias at Slab creek, Yukon Territory, Canada. In: Emond DS, Weston LH and Lewis LL (eds.) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 249-258.

Carlon, CJ (2000) Iron oxide systems and base metal mineralisation in northern Sweden. . In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing Adelaide, pp 283-296.

Chetty D and Frimmel HE (2000) The role of evaporites in the genesis of base metal sulphide mineralisation in the Northern Platform of the Pan-African Damara Belt, Namibia: geochemical and fluid inclusion evidence from carbonate wall rock alteration. Mineralium Deposita 35:364-376.

Compston DM (1995) Time constraints on the evolution of the Tennant Creek block, northern Australia. Precambrian Research 71:107-129.

Creaser RA (1995) Neodymium isotopic constraints for the origin of Mesoproterozoic felsic magmatism, Gawler Craton, South Australia. Canadian Journal of Earth Science 32:460-471.

Cross KC, Daly SJ and Flint RB (1993) Olympic Dam deposit. In: Drexel JF, Preiss WV and Parker AJ (eds.) The geology of South Australia. Geological Survey of South Australia, Bulletin 54, volume 1, pp 132-138.

Davidson GJ, Davis BK and Garner A (2002) Structural and geochemical constraints on the emplacement of the Monakoff oxide Cu-Au (-Co-U-REE-Ag-Zn-Pb) deposit, Mt Isa Inlier, Australia. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 2, PGC Publishing Adelaide, pp 49-75.

De Jong G and Williams PJ (1995) Evolution of metasomatic features during exhumation of mid crustal Proterozoic rocks in the vicinity of the Cloncurry fault, NW Queensland. Australian Journal of Earth Sciences 42: 281-290.

Delaney GD (1981) The Mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory; In: Proterozoic Basins of Canada, Geological Survey of Canada, Paper 81-10, pp. 1-23.

Deming D (1992) Catastrophic release of heat and fluid flow in the continental crust. Geology 20:83-86.

DOCEGEO, (1988) Revisão litoestratigráfica da Província Mineral de Carajás. In: SBG, Congresso Brasileiro de Geologia, 35, Belém, Anexo aos Anais, pp. 11-56.

Einaudi MT and Oreskes N (1990) Progress towards an occurrence model for Proterozoic iron oxide (Cu, U, REE, Au) deposits – a comparison between the ore provinces of South Australia and SE Missouri. In: Pratt WP and Sims P (eds.) The midcontinent: Permissive terrane for an Olympic Dam deposit? US Geological Survey Bulletin, 1392: 58-69.

Eldridge CS and Danti K (1994) Low sulphur isotope ratios; high gold values – a closer look at the Olympic Dam deposit via SHRIMP. Geological Society of America Abstracts with Programs, 1994 Annual Meeting, pp. A-498 – A-499.

Ferris GM, Schwarz MP and Heithersay P (2002) The geological framework, distribution and controls of Fe oxide Cu-Au mineralisation in the Gawler Craton, South Australia – Part 1 – Geological and tectonic framework. In: Porter TM (ed.), Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 2, PGC Publishing, Adelaide, pp. 9-31.

Frietsch R, Billström K and Perdahl JA (1995) Sulphur isotopes in Lower Proterozoic iron and sulphide ores in northern Sweden. Mineralium Deposita 30: 275-284.

Frietsch R, Tuisku P, Martinsson O and Perdahl, J (1997) Early Proterozoic Cu(-Au) and Fe ore deposits associated with regional Na-Cl metasomatism in northern Fennoscandia. Ore Geology Reviews 12: 1-34.

Frimmel HE, Deane JG and Chadwick PJ (1996) Pan-African tectonism and the genesis of base metal sulphide deposits in the northern foreland of the Damara orogen, Namibia. Society of Economic Geologists, Special Publication 4:204-217.

Gabrielse H (1967) Tectonic evolution of the northern Canadian Cordillera. Canadian Journal of Earth Sciences, 4:271-298.

Gandhi SS and Bell RT (1990) Metallogenic concepts to aid exploration for the giant Olympic Dam-type deposits and their derivatives. 8th IAGOD Symposium, Program with Abstracts, Ottawa, August 1990, Canada, pp. A7.

Garven G, Bull SW and Large RR (2001) Hydrothermal fluid flow models of stratiform ore genesis in the McArthur Basin, Northern Territory, Australia. Geofluids 1:4:289-311.

Gauthier L, Hall G, Stein H and Schaltegger U (2001) The Osborne deposit, Cloncurry district: a 1595 Ma Cu-Au skarn deposit. In: Williams PJ (ed.) 2001: A hydrothermal odyssey. Extended Conference Abstracts, Economic Geology Research Unit, James Cook University, Contribution 59, pp. 58 -59.

Goodfellow WD (1979) Geochemistry of copper, lead and zinc mineralization in Proterozoic rocks near Gillespie Lake, Yukon. Geological Survey of Canada Paper 79-1A: 333-348.

Gorbatshev R and Bogdanova S (1993) Frontiers in the Baltic Shield. Precambrian Research 64:3-21.

Haynes DW (2000) Iron oxide copper (-gold) deposits: their position in the ore deposit spectrum and modes of origin. In: Porter TM (ed.) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 71-90.

Haynes DW, Cross KC, Bills RT and Reed MH (1995) Olympic Dam ore genesis: a fluid-mixing model. Economic Geology 90:281-307.

Heinrich CA, Bain JHC, Fardy JJ and Waring C (1993) Bromine/chlorine geochemistry of hydrothermal brines associated with Proterozoic metasediment-hosted copper mineralisation at Mount Isa, northern Australia. Geochimica Cosmochimica Acta 57:2991-3000.

Heinrich CA, Bain JHC, Mernagh TP, Wyborn LAI, Andrew AS and Waring C (1995) Fluid and mass transfer during metabasalt alteration and copper mineralisation at Mount Isa, Australia. Economic Geology 90:705-730.

Hitzman MW (2000) Iron oxide-Cu-Au deposits: what, where, when, and why? In: Porter TM (ed.) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 9-25.

Hitzman MW, Oreskes N and Einaudi MT (1992) Geological characteristics and tectonic setting of Proterozoic iron oxide (Cu-U-Au-REE) deposits. Precambrian Research 58:241-287.

Hughes MJ (1987) The Tsumeb ore body, Namibia, and related dolostone-hosted base metal ore deposits of Central Africa. PhD Thesis, University of Witwatersrand, Johannesburg, 448pp. University Microfilms International, Ann Arbor, Michigan, USA.

Hunt JA, Laughton JR, Brideau M-A, Thorkelson DJ, Brookes ML and Baker T (2002) New mapping around the Slab iron oxide-copper-gold occurrence, Wernecke Mountains. In: Emond DS, Weston LH and Lewis LL (eds.) Yukon Exploration and Geology 2001, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 125-138.

Hunt JA, Baker T, Davidson G, Fallick AE, and Thorkelson DJ (2004) Origin of Wernecke Breccia: results of fluid inclusion and stable isotope analyses. Abstract, GSA Convention, Hobart, Tasmania, February 2004.

Hunt JA, Baker, T and Thorkelson, DJ (2005, in review). Regional-scale Proterozoic IOCG-mineralised breccia systems: examples from the Wernecke Mountains, Yukon, Canada. Mineralium Deposita, X:x:x-x.

Huston DL, Bolger C and Cozens G (1993) A comparison of mineral deposits at the Gecko and White Devil deposits: implications for ore genesis in the Tennant Creek district, Northern Territory, Australia. Economic Geology 88: 5: 1198-1225.

Jackson MJ and Southgate PN (2000) Evolution of three unconformity-bounded sand-carbonate successions in the McArthur River region of Northern Australia: the Lawn, Wide and Doom Supersequences in a proximal part of the Isa Superbasin. Australian Journal of Earth Sciences 47: 3: 625-636.

Johnson JP and Cross KC (1991) Geochronological and Sm-Nd constraints on the genesis of the Olympic Dam Cu-U-Au-Ag deposit, South Australia. In: Pagel M and Leroy JL (eds.) Source, Transport and Deposition of Metals: Proceedings, 25 Years' SGA Anniversary Meeting, Nancy, Aug. 30 – Sept. 3, 1991, AA Balkema, Rotterdam, The Netherlands, pp 395-400.

Johnson JP and Cross KC (1995) U-Pb geochronological constraints on the genesis of the Olympic Dam Cu-U-Au-Ag deposit, South Australia. Economic Geology 90:1046-1063.

Knutson J, Ferguson J, Roberts WMB, Donnelly TH and Lambert, IB (1979). Petrogenesis of the copper-bearing breccia pipes, Redbank, Northern Territory, Australia. Economic Geology 74:4:814-826.

Knutson J, Donnely TH, Eadington PJ and Tonkin DG (1992) Hydrothermal alteration of Middle Proterozoic basalts, Stuart shelf, South Australia – a possible source for copper mineralization. Economic Geology 87:1054-1077.

Kwak TAP (1977) Scapolite compositional change in metamorphic gradient and its bearing on the identification of metaevaporite sequences. Geological Magazine, 114:343-354.

Laing WP (1998) Structural-metasomatic environment of the East Mt Isa Block base metal-gold province. Australian Journal of Earth Sciences 45:413-428.

Large RR (1975) Zonation of hydrothermal minerals at the Juno mine, Tennant Creek goldfield, central Australia. Economic Geology 70: 1387-1413.

Lane RA (1990) Geologic setting and petrology of the Proterozoic Ogilvie Mountains breccia of the Coal Creek Inlier, southern Ogilvie Mountains, Yukon Territory. MSc. Thesis, University of British Columbia, pp. 223.

Laughton JR (2004) The Proterozoic Slab volcanics of northern Yukon, Canada: megaclasts of a volcanic succession in Proterozoic Wernecke Breccia, and implications for the evolution of northwestern Laurentia. PhD Thesis, Simon Fraser University, Burnaby, British Columbia, Canada.

Laughton JR, Thorkelson DJ, Brideau M-A, Hunt JA and Marshall DD (in review) Early Proterozoic orogeny and exhumation of Wernecke Supergroup revealed by vent facies of Wernecke Breccia, Yukon, Canada. Canadian Journal of Earth Sciences, xx: x-x.

Lindenmayer ZG (1990) Salobo sequence, Carajás, Brazil. PhD Thesis, University of Western Ontario, Canada, 406p.

MacCready T, Goleby BR, Goncharov A, Drummond BJ and Lister GS (1998) A framework of overprinting orogens based on the interpretation of the Mount Isa deep seismic transect. Economic Geology 93:1422-1434.

Machado N, Lindenmayer ZG, Krogh TE and Lindenmayer D (1991) U-Pb geochronology of Archean magmatism and basement reactivation in the Carajás area, Amazon Shield, Brazil. Precambrian Research 49:329-354.

Mark G and De Jong G (1996) Synchronous granitoid emplacement and episodic sodic-calcic alteration in the Cloncurry district: styles, timing and metallogenic significance. Economic Geology Research Unit, James Cook University, Townsville, Queensland, Australia, Contribution 55, pp 77-80.

Mark G and Crookes RA (1999) A hydrothermal origin for the Ernest Henry Fe oxide-(Cu-Au) deposit, Cloncurry district, Australia. In: Stanley CJ *et al.* (eds.) Mineral Deposits: Processes to Processing. Balkema, Rotterdam pp. 235-238.

Mark G, Oliver NHS, Williams PJ, Valenta RK and Crookes RA (2000) The evolution of the Ernest Henry Fe-oxide-(Cu-Au) hydrothermal system. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp 123-136.

Marschik R and Fontboté L (1996) Copper(-iron) mineralisation and superposition of alteration events in the Punta del Cobre belt, northern Chile. In: Camus F, Sillitoe RH, Peterson R (eds.) Andean Copper Deposits: New Discoveries, Mineralisation, Styles and Metallogeny. Society of Economic Geologists Special Publication #5, pp. 171-190.

Marschik R and Fontboté L (2001) The Candelaria – Punta del Cobre iron oxide Cu-Au (-Zn-Ag) deposits, Chile. Economic Geology 96:1799-1826.

Marschik R, Leveille RA and Martin W (2000) La Candelaria and the Punta del Cobre district, Chile: Early Cretaceous iron oxide Cu-Au(-Zn-Ag) mineralisation. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp 163-175.

McKibben MA and Hardie LA (1997) Ore-forming brines in active continental rifts. In: Barnes HL (ed.) Geochemistry of Hydrothermal Ore Deposits. John Wiley and Sons, pp. 877-936.

Meyer C (1988) Proterozoic ore-forming habitats – tectonic and chemical transitions. In: Kisvarsanyi G and Grant SK (eds.) Tectonic controls of ore deposits and vertical and horizontal extent of ore systems. Conference Proceedings, University of Missouri, Rolla, October 6-8, 1987, pp. 217-235.

Monro D (1988) The geology and genesis of the Aitik copper-gold deposit, Arctic Sweden. PhD Thesis, Cardiff College, University of Wales, U.K.

Mustard R, Baker T, Williams P, Ulrich T, Mernagh T, Ryan CG, VanAchterbergh E and Adshead N (2003) Cu-rich brines at the Osborn and Starra deposits: implications for immiscibility in Fe-oxide Cu-Au systems. Applied Earth Science Transactions of the Institute of Mining and Metallurgy: Section B: 112: B189-B191.

O'Dea MG, Lister GS and MacCready T (1997) Geodynamic evolution of the Proterozoic Mt Isa Terrain. In: Burg J-P and Ford M (eds.) Orogeny Through Time. Geological Society of London Special Publication 121 pp 99-122.

Ohmoto H and Goldhaber MB (1997) Sulphur and carbon isotopes. In: Barnes HL (ed.) Geochemistry of hydrothermal ore deposits. John Wiley and Sons, New York, USA, pp. 517-612.

Oreskes N and Einaudi MT (1990) Origin of REE-enriched hematite breccias at the Olympic Dam Cu-U-Au-Ag deposit, Roxby Downs, South Australia. Economic Geology 85:1-28.

Oreskes N and Einaudi MT (1992) Origin of hydrothermal fluids at Olympic Dam: preliminary evidence from fluid inclusions and stable isotopes. Economic Geology 87:1:64-88.

Oreskes N and Hitzman MW (1993) A model for the origin of Olympic Dam-type deposits. In: Kirkham RV, Sinclair WD, Thorpe RI and Duke JM (eds.) Mineral Deposit Modeling. Geological Association of Canada, Special Paper 40, pp 615-633.

Orridge GR and Mason AAC (1975) Redbank copper deposits. In: Knight CL (ed.) Economic Geology of Australia and Papua New Guinea, 1, Metals. Australian Institute of Mining and Metallurgy Monograph 5 pp. 339-343.

Page RW and Sun S-S (1998) Aspects of geochronology and crustal evolution in the Eastern fold belt, Mount Isa inlier. Australian Journal of Earth Sciences 45:343-361.

Perkins WG (1984) Mount Isa silica dolomite and copper orebodies: the result of a syntectonic hydrothermal system. Economic Geology 79:601-637.

Perkins C and Wyborn L (1996) The age of Cu-Au mineralisation, Cloncurry district, Mount Isa Inlier, as determined by $^{40}Ar/^{39}Ar$ dating. Australian Geological survey Organisation Research Newsletter, 25:8-9.

Perkins C and Wyborn L (1998) Age of Cu-Au mineralisation, Cloncurry district, Mount Isa Inlier, as determined by $^{40}Ar/^{39}Ar$ dating. Australian Journal of Earth Sciences 45:233-246.

Perring CS, Pollard PJ, Dong G, Nunn AJ and Blake KL (2000) The Lightening Creek sill complex, Cloncurry district, northwest Queensland: a source of fluids for Fe oxide Cu-Au mineralisation and sodic-calcic alteration. Economic Geology 95: 1067-1090.

Pinheiro RVL (1997) Reactivation history of the Carajás and Cinzento strike-slip systems, Amazon, Brazil. PhD Thesis, University of Durham, England, 408p.

Pollard PJ (2001) Sodic(-calcic) alteration in Fe oxide-Cu-Au districts: an origin via unmixing of magmatic-derived H_2O-CO_2-NaCl±$CaCl_2$-KCl fluids. Mineralium Deposita 36:93-100.

Pollard PJ (2000) Evidence of a magmatic fluid and metal source for Fe oxide Cu-Au mineralisation. In: Porter TM (ed.) Hydrothermal Iron Oxide-Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 27-41.

Pollard PJ, Mark G and Mitchell L (1998) Geochemistry of post-1540 Ma granites in the Cloncurry district. Economic Geology 93:1330-1344.

Raymond, J. (2000) Mantle of the earth. In: Sigurdsson H (ed.) Encyclopaedia of Volcanoes, Academic Press, San Diego, USA, pp 41-54.

Raymond, J. (1995) Petrology: The study of igneous, metamorphic and sedimentary rocks. McGraw-Hill Higher Education, New York, 742p.

Reeve JS, Cross KC, Smith RN and Oreskes N (1990) Olympic Dam copper-uranium-gold-silver deposit. In: Hughes FE (ed.) Geology of the Mineral Deposits of Australian and Papua New Guinea, Australasian Institute of Mining and Metallurgy, Monograph 14, pp. 1009-1035.

Requia K and Fontboté L (2000) The Salobo iron oxide-copper-gold deposit, Carajás, northern Brazil. In: Porter TM (ed.) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 225-236.

Reynolds LJ (2000) Geology of the Olympic Dam Cu-U-Au-Ag-REE deposit. In: Porter TM (ed.) Hydrothermal Iron Oxide-Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 93-104.

Roberts DE and Hudson GRT (1983) The Olympic Dam copper-uranium-gold deposit, Roxby Downs, South Australia. Economic Geology 78:5:799-822.

Roberts DE and Hudson GRT (1984) The Olympic Dam copper-uranium-gold deposit, Roxby Downs, South Australia – A reply. Economic Geology 79:1944-1945.

Rollinson H (1993) Using geochemical data: evaluation, presentation, interpretation. Longman Group, UK, 352p.

Ronzê PC, Soares ADV, Giovanni S and Berreira CF (2000) Alemão copper-gold (U-REE) deposit, Carajás, Brazil. In: Porter TM (ed.) Hydrothermal Iron Oxide-Copper-Gold & Related Deposits: A Global Perspective, volume 1, PGC Publishing, Adelaide, pp. 191-202.

Rotherham JF, Blake KL, Cartwright I and Williams PJ (1998) Stable isotope evidence for the origin of the Mesoproterozoic Starra Cu-Au deposit, Cloncurry district, northwest Queensland. Economic Geology, 93:1435-1449.

Rubenach M, Adshead N, Oliver N, Tullemans F, Esser D and Stein H (2001) The Osborne Cu-Au deposit: geochronology and genesis of mineralisation in relation to host albitites and ironstones. In: Williams PJ (ed.) 2001: A hydrothermal odyssey. Extended Conference Abstracts, Economic Geology Research Unit, James Cook University, Contribution 59, pp. 172-173.

Ryan (1998) Ernest Henry copper-gold deposit. In: Berkman DA and Mackenzie DH (eds.) Geology of Australian and Papua New Guinean Mineral Deposits, Australasian Institute of Mining and Metallurgy: Melbourne, Monograph 22, pp. 759-768.

Scheuber E and Andriessen PAM (1990) The kinematic and geodynamic significance of the Atacama Fault Zone, northern Chile. Journal of Structural Geology 12:243-257.

Scott DL, Rawlings DJ, Page RW, Tarlowski CZ, Idnurm M, Jackson MJ and Southgate PN (2000) Basement framework and geodynamic evolution of the Paleoproterozoic superbasins of north-central Australia: an integrated review of geochemical, geochronological and geophysical data. Australian Journal of Earth Sciences 47:341-380.

Sheppard SMF (1986) Characterization and isotopic variations in natural waters. In: Valley JW, Taylor HP Jr and O'Neil JR (eds.) Stable Isotopes in High Temperature Geological Processes, Reviews in Mineralogy, volume 16, Mineralogical Society of America, pp. 165-183.

Sillitoe RH (2003) Iron oxide copper-gold deposits: an Andean view. Mineralium Deposita, 38:787-812.

Sillitoe RH and Clark AH (1969) Copper and copper iron sulphides as the initial products of supergene oxidation, Copiapó mining district, northern Chile. American Mineralogist 54:1684-1710.

Skiöld, T (1987) Aspects of Proterozoic geochronology on northern Sweden. Precambrian Research, 35: 161-167.

Skiöld, T and Cliff RA (1984) Sm-Nd and U-Pb dating of Early Proterozoic mafic-felsic volcanism in northernmost Sweden. Precambrian Research, 26: 1-13.

Skirrow RG (1999) Proterozoic Cu-Au-Fe mineral systems in Australia: filtering key components in exploration models. In: Stanley CJ *et al.* (eds.) Mineral Deposits: Processes to Processing. Balkema, Rotterdam pp. 1361-1364.

Skirrow RG and Walshe JL (2002) Reduced and oxidized Au-Cu-Bi iron oxide deposits of the Tennant Creek Inlier, Australia: an integrated geologic and chemical model. Economic Geology 97:1167-1202.

Southgate PN, Bradshaw BE, Domagala J, Jackson MJ, Idurnam M, Krassay AA, Page RW, Sami TT, Scott DL, Lindsay JF, McConachie BA and Tarlowski C (2000) Chronostratigraphic basin framework for Paleoproterozoic rocks (1730-1575 Ma) in northern Australia and implications for base-metal mineralisation. Australian Journal of Earth Sciences 47:3:461-483.

Souza, LH and Vieira EA (2000) Salobo 3 Alpha Deposit: geology and mineralisation. In: Porter TM (ed.) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 213-224.

Swager CP (1985) Syndeformational carbonate-replacement model for the copper mineralisation at Mount Isa, Queensland: a microstructural study. Economic Geology 80:107-125.

Tazava E and de Olivera CG (2000) The Igarape Bahia Au-Cu(-REE-U) deposit, Carajás mineral province, northern Brazil. In: Porter TM (ed.) Hydrothermal iron oxide copper-gold & related deposits: a global perspective, volume 1, pp. 203-212.

Thorkelson DJ (2000) Geology and mineral occurrences of the Slats Creek, Fairchild Lake and "Dolores Creek" areas, Wernecke Mountains, Yukon Territory. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 10, 73pp.

Thorkelson DJ, Mortensen JK, Davidson GJ, Creaser RA, Perez WA and Abbott JG (2001a) Early Mesoproterozoic intrusive breccias in Yukon, Canada: The role of hydrothermal systems in reconstructions of North America and Australia. Precambrian Research 111:31-55.

Thorkelson DJ, Mortensen JK, Creaser RA, Davidson GJ and Abbott JG (2001b) Early Proterozoic magmatism in Yukon, Canada: constraints on the evolution of northwestern Laurentia. Canadian Journal of Earth Science 38:1479-1494.

Thorkelson DJ, Laughton JR and Hunt JA (2002) Geological map of Quartet Lakes area (106E/1), Wernecke Mountains, Yukon (1:50,000 scale). Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Geoscience map 2002-2.

Thorkelson DJ, Laughton JR, Hunt JA and Baker T (2003) Geology and mineral occurrences of the Quartet Lakes map area (NTS 106E/1), Wernecke and Mackenzie mountains, Yukon. In: Emond DS and Lewis LL (eds.) Yukon Exploration and Geology 2002, Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, pp 223-239.

Torgersen T (1990) Crustal-scale fluid transport, magnitude and mechanisms. EOS (Transactions, American Geophysical Union) 71:1,4,13.

Twyerould SC (1997) The geology and genesis of the Ernest Henry Fe-Cu-Au deposit, northwest Queensland, Australia. University of Oregon, PhD Thesis, 494 p.

Ullrich TD and Clark AH (1999) The Candelaria copper-gold deposit, Region III, Chile: paragenesis, geochronology and fluid composition. In: Stanley CJ *et al.* (eds.) Mineral deposits: Processes to Processing, volume 1. Proceedings of the 5th Biennial SGA Meeting and 10th Quadrennial IAGOD Symposium, London, UK, AA Balkema, Rotterdam, pp 201-204.

Wanhainen C and Martinsson O (1999) Geochemical characteristics of host rocks to the Aitik Cu-Au-Ag deposit, Gellivare area, northern Sweden. Proceedings of the fifth biennial SGA meeting and the tenth quadrennial IAGOD meeting, London, 22-25 August 1999, Extended Abstracts, pp. 1443-1446.

Wanhainen C, Broman C and Martinsson O (2003) The Aitik Cu-Au-Ag deposit in northern Sweden: a product of high salinity fluids. Mineralium Deposita 38: 715-726.

Wedekind MR and Love RJ (1990) Warrego gold-copper-bismuth deposit. In: Hughes FE (ed) Geology of the Mineral Deposits of Australia and Papua New Guinea, The Australian Institute of Mining and Metallurgy, Melbourne, pp 839-843.

Williams PJ and Blake KL (1993) Alteration in the Cloncurry district: Roles of recognition and interpretation on exploration for Cu-Au and Pb-Zn-Ag deposits. Economic Geology Research Unit Contribution 49, James Cook University, Australia, 75pp.

Williams PJ and Skirrow RG (2000) Overview of iron oxide-copper-gold deposits in the Curnamona Province and Cloncurry District (eastern Mount Isa block), Australia. In: Porter TM (ed.) Hydrothermal Iron Oxide Copper-Gold & Related Deposits: A Global Perspective, volume. 1, PGC Publishing, Adelaide, pp 105-122.

Williams PJ, Dong G, Pollard PJ and Perring CS (1999) Fluid inclusion geochemistry of Cloncurry (Fe)-Cu-Au deposits. In: Stanley CJ *et al.* (eds.) Mineral deposits: processes to processing. Rotterdam, Balkema, pp 111-114.

Williams PJ, Broman C, Guoyi D, Mark G, Martinsson O, Mernagh TP, Pollard PJ, Ryan CG and Win TT (in progress) PIXIE characterization of fluid inclusion brines from Proterozoic Fe oxide-bearing Cu-Au deposits, Norbotten (Sweden) and the Cloncurry district (NW Queensland), 50 pp.

Wyborn LA (1998) Younger *ca.* 1500 Ma granites of the Williams and Naraku batholiths, Cloncurry district, eastern Mt Isa inlier: geochemistry, origin, metallogenic significance and exploration indicators. Australian Journal of Earth Sciences 45:397-411.

Yardley BWD and Graham JT (2002) The origins of salinity in metamorphic fluids. Geofluids, 2:249-256.

Yukon MINFILE, 2003. Database of Yukon mineral occurrences. Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, CD-ROM (2).