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Abstract
Multiply deformed sediments of the Cambrian Kanmantoo Group, which were metamorphosed to the amphibolite facies, host numerous Cu-Au, Fe-S, and Pb-Zn-Ag-(Cu-Au) deposits, of which the largest Cu-Au deposit is Kanmantoo (34.5 Mt @ 0.6% Cu and 0.1 g/t Au). Mineralization at Kanmantoo is characterized by discordant and pipe-like orebodies (Kavanagh and Emily Star) along with mineralization locally concordant to bedding (Nugent) that is spatially related to meta-exhalative rocks. Previous studies have suggested a syn-sedimentary origin for the Pb-Zn-Ag-(Cu-Au) and Fe-S deposits, whereas the Kanmantoo deposit remains controversial, with syn-sedimentary, metamorphogenic, and post-peak metamorphic models having being applied. The stratiform nature of some parts of the Nugent orebody, and disseminated sulfides locally concordant to bedding along with the recognition of a zone of chalcopyrite-magnetite-rich rocks at the syn-sedimentary Wheal Ellen Pb-Zn-Ag-(Cu-Au) deposit, which shows a metallic mineral assemblage almost identical to the most common assemblage at Kanmantoo, supports a genetic link between the Pb-Zn-Ag-(Cu-Au) deposits and Cu-Au mineralization, and is consistent with a metamorphosed syn-sedimentary model for Cu-Au mineralization. The discordant nature of most orebodies at Kanmantoo is the result of the mineralization having formed as stockwork zones in sub-seafloor pipes. Varying degrees of remobilization were subsequently associated with syn- to post-peak metamorphism. Major and trace element compositions, coupled with principal component analyses, show that the compositions of garnet, biotite, staurolite, chlorite, muscovite, and magnetite in metamorphosed altered rocks spatially associated with sulfide mineralization at Kanmantoo can be distinguished from those, where present, in metamorphosed country rocks (i.e., metapsammites and metapelites). Garnet, chlorite, biotite, and muscovite in quartz garnetite within quartz mica schist associated with the Nugent orebody are elevated in Mn (up to 19.5 wt% MnO – garnet, 2,825 ppm – chlorite, 3,206 ppm – biotite, and 108 ppm – muscovite) and Zn (up to 170 ppm – garnet, 1,602 ppm – chlorite, 1,592 ppm – biotite, and 108 ppm – muscovite) relative to samples in other orebodies and in unmineralized rocks elsewhere in the Kanmantoo Group. Such enrichments in these elements mimic similar enrichments in the same minerals in metamorphosed altered rocks associated with Pb-Zn-Ag-(Cu-Au) deposits in the Kanmantoo Group. Biotite in metamorphosed altered rocks at Kanmantoo contains elevated concentrations of Pb (up to 110 ppm), and, in general, Zn (up to 841 ppm), whereas muscovite is also elevated in Pb, Zn, and Cu (up to 272 ppm Pb, 78 ppm Zn, and 173 ppm Cu). Staurolite in these same rocks contains up to 1.6 wt% ZnO (with one outlier that contains 3.2 wt% ZnO) and is considerably more enriched in Zn than in unaltered country rocks (∼ 0.1 wt% Zn). The trace element enrichments in silicates studied here constitute a potential pathfinder to metamorphosed Cu-Au mineralization in the Kanmantoo Group and emphasize the geochemical and genetic links between the Pb-Zn-Ag-(Cu-Au) and Cu-Au deposits.

Disciplines
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Comments

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The origin of the sediment-hosted Kanmantoo Cu-Au deposit, South Australia: Mineralogical considerations

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ABSTRACT

Multiply deformed sediments of the Cambrian Kanmantoo Group, which were metamorphosed to the amphibolite facies, host numerous Cu-Au, Fe-S, and Pb-Zn-Ag-(Cu-Au) deposits, of which the largest Cu-Au deposit is Kanmantoo (34.5 Mt @ 0.6% Cu and 0.1 g/t Au). Mineralization at Kanmantoo is characterized by discordant and pipe-like orebodies (Kavanagh and Emily Star) along with mineralization locally concordant to bedding (Nugent) that is spatially related to meta-exhalative rocks. Previous studies have suggested a syn-sedimentary origin for the Pb-Zn-Ag-(Cu-Au) and Fe-S deposits, whereas the Kanmantoo deposit remains controversial, with syn-sedimentary, metamorphogenic, and post-peak metamorphic models having being applied. The stratiform nature of some parts of the Nugent orebody, and disseminated sulfides locally concordant to bedding along with the recognition of a zone of chalcopyrite-magnetite-rich rocks at the syn-sedimentary Wheal Ellen Pb-Zn-Ag-(Cu-Au) deposit, which shows a metallic mineral assemblage almost identical to the most common assemblage at Kanmantoo, supports a genetic link between the Pb-Zn-Ag-(Cu-Au) deposits and Cu-Au mineralization, and is consistent with a metamorphosed syn-sedimentary model for Cu-Au mineralization. The discordant nature of most orebodies at Kanmantoo is the result of the mineralization having formed as stockwork zones in sub-seafloor pipes. Varying degrees of remobilization were subsequently associated with syn-to post-peak metamorphism. Major and trace element compositions, coupled with principal component analyses, show that the compositions of garnet, biotite, staurolite, chlorite, muscovite, and magnetite in metamorphosed altered rocks spatially associated with sulfide mineralization at Kanmantoo can be distinguished from those, where present, in metamorphosed country rocks (i.e., metapsammites and metapelites). Garnet, chlorite, biotite, and muscovite in quartz garnetite within quartz mica schist associated with the Nugent orebody are elevated in Mn (up to 19.5 wt% MnO–garnet, 2,825 ppm–chlorite, 3,206 ppm–biotite, and 108 ppm–muscovite) and Zn (up to 170 ppm–garnet, 1,602 ppm–chlorite, 1,592 ppm–biotite, and 108 ppm–muscovite) relative to samples in other orebodies and in unmineralized rocks elsewhere in the Kanmantoo Group. Such enrichments in these elements mimic similar enrichments in the same minerals in metamorphosed altered rocks associated with Pb-Zn-Ag-(Cu-Au) deposits in the Kanmantoo Group. Biotite in metamorphosed altered rocks at Kanmantoo contains elevated concentrations of Pb (up to 110 ppm), and, in general, Zn (up to 841 ppm), whereas muscovite is also elevated in Pb, Zn, and Cu (up to 272 ppm Pb, 78 ppm Zn, and 173 ppm Cu). Staurolite in these same rocks contains up to 1.6 wt% ZnO (with one outlier that contains 3.2 wt% ZnO) and is considerably more enriched in Zn than in unaltered country rocks (~0.1 wt% Zn). The trace element enrichments in silicates studied here constitute a potential pathfinder to metamorphosed Cu-Au mineralization in the Kanmantoo Group and emphasize the geochemical and genetic links between the Pb-Zn-Ag-(Cu-Au) and Cu-Au deposits.

1. Introduction

There is considerable debate in the literature concerning the role of metamorphic processes in the formation and/or modification of pre-existing ore deposits, particularly strata-bound massive sulfide deposits (e.g., Solomon and Groves, 1994; Cartwright and Oliver, 2000; Marshall and Spry, 2000). Disputed issues include the origin of massive sulfide deposits in metamorphic rocks, the source of the metals, and the...
ways to explore for such deposits. Ore deposit models center on whether the base metal mineralization is syn-sedimentary and subsequently metamorphosed, or whether it is syn-deformational and formed during metamorphism (i.e., metamorphic gene). There are many examples in the literature where both models have been applied to the same deposit including the Broken Hill Pb-Zn deposit, Australia (Laing et al., 1978; Parr and Plimer, 1993; White et al., 1996; Large et al., 1996), Cannington Pb-Zn-Ag deposit (Walters and Bailey, 1998; Chapman and Williams, 1998; Roache, 2004), and Dongshengmiao Zn-Pb-Cu deposit, China (Peng et al., 2007; Zhang et al., 2010; Zhong et al., 2015b). In an attempt to decipher the origin of some of these contentious deposits, Marshall and Spry (2000) invoked a probabilistic approach in which 15 guidelines were evaluated. These can be lumped under four major categories: 1. General guidelines (primacy of solid-state remobilization and proportionality of data); 2. Structural and textural guidelines; 3. Petrography and fluid inclusions; and 4. Isotopes and other geochemical tracers.

The Cambrian Kanmantoo Group, South Australia, consists of multiple deformed pelitic and psammitic rocks that were regionally metamorphosed to the amphibolite facies. The Tapanappa Formation of the Kanmantoo Group hosts Cu-Au, Pb-Zn-Ag-(Cu-Au), Fe-S, and minor As, Bi, and W deposits. Although there is consensus that the Fe-S deposits are syn-sedimentary, the origin of the Kanmantoo Cu-Au deposit, is controversial (Seccombe et al., 1985; Oliver et al., 1988; Marshall and Spry, 2000). A syn-sedimentary model for the Pb-Zn-Ag-(Cu-Au) mineralization was favored by Toteff (1999) but a structural model was more recently proposed by Singer (2008) for the Angas Pb-Zn-Ag deposit. The three most widely cited genetic models for the Kanmantoo deposit are: 1. A syn-sedimentary origin in which sulfide mineralization formed by hydrothermal processes at or below the seafloor. In this model, the deposit was subsequently metamorphosed and variably remobilized (M.V. Pollock et al., 2019). 2. A syn-metamorphic model in which sulfide mineralization was introduced during peak metamorphism (Thomson, 1975; Parker, 1986; Solomon and Groves, 1994; Gun, 1998; Oliver et al., 1998; Schiller, 2000; Burt, 2008): 3. A post-metamorphic model where ore-bearing fluids were derived from an orogenic granite late in the Delamerian Orogeny (Foden et al., 1999; Focke et al., 2009; Schmidt Mumm et al., 2009; Tedesco, 2009; Wilson, 2009; Arbon, 2011; Lyons, 2012).

The current contribution involves a petrographic, textual, and major, minor, and trace element mineral study of common rock-forming silicates and oxides (i.e., garnet, biotite, staurolite, chlorite, muscovite, magnetite) spatially associated with Cu-Au mineralization in three (Nugent, Emily Star, Kavanagh) of ten orebodies currently recognized in the Kanmantoo deposit. An aim of this study is to complement the probabilistic approach of Marshall and Spry (2000) by focusing on the mineralogy of the deposit to help understand the origin of the sulfide mineralization. A second aim of the study is to develop a geochemical fingerprint for non-sulfide minerals spatially associated with Cu-Au mineralization at Kanmantoo. Although trace element studies of minerals are increasingly being used as a pathfinder to ore deposits and to discriminate between ore deposit types (e.g., apatite, Belousova et al., 2002; gahnite, O’Brien et al. 2015a,b; magnetite, Dupuis and Beaudoin, 2011; garnet, Spry et al., 2007), this study is among the first to use the compositions of multiple minerals in metamorphic rocks as potential vectors to base metal mineralization.

2. Regional geology

The Kanmantoo Cu-Au deposit occurs in the Tapanappa Formation of the Kanmantoo Group within the Adelaide Fold Belt (Fig. 1; Sprigg and Campana, 1953; Seccombe et al., 1985). The Kanmantoo Group is a structurally thickened package (~7–8 km thick; Haines et al., 2001) of metamorphosed pelitic and psammitic rocks that were deposited as high-density sediment gravity flows within an extensional back-arc basin (i.e., Kanmantoo Trough) along the Paleo-Pacific margin of Gondwana (Flöttmann et al., 1998; Gun, 1998; Jago et al., 2003). Development of the Kanmantoo Trough commenced in the early Cambrian due to extensional tectonism along the southeast margin of the Mount Lofty Ranges. The fault-controlled basin was filled with clastic and minor carbonate sediments derived from the Ross Orogen and deposited in a shallow- to deep-marine environment (Parker, 1986; Flöttmann et al., 1998; Totteff, 1999; Haines et al., 2001).

The stratigraphic sequence of the Kanmantoo Group has been variously defined by Forbes (1957), Thomson and Horowitz (1962), Mirams (1962), Daily and Milnes (1971, 1972, 1973), Dyson et al. (1996), and Jago et al. (2003). Utilizing the most recent nomenclature of Jago et al. (2003), the Kanmantoo Group unconformably overlies the Early Cambrian Normanville Group, which consists of platform carbonates, sandstone, siltstone, shale, and minor volcanic rocks (Flöttmann et al., 1998; Gun, 1998). The Kanmantoo Group is divided into the Keynes and Bollaparudda subgroups with the latter hosting the base metal deposits and largely consisting of a package of deep sea sediments (Flöttmann et al., 1998). Within the Bollaparudda subgroup, the Talisker Formation is a metamorphosed sandstone, with subordinate siltstone and calc-silicate rocks (e.g., Coalinga Member), which hosts the Mt. Torrens Pb-Zn-Ag and Talisker Pb-Ag deposits, and the overlying Nairne Pyrite Member with sulfide-enriched, pyritic meta-siltstone lenses that hosts Fe-S deposits (e.g., Brukunga deposit). The overlying Tapanappa Formation is the primary host to Cu-Au (e.g., Kanmantoo, Bremer, South Hill) and Pb-Zn-Ag-(Cu-Au) deposits (e.g., Angas, Wheal Allen, Aclaire, Strathalbyn, Scotts Creek) and consists predominantly of metamorphosed sandstone and greywacke with intercalated siltstone and pyritic mudstone (Seccombe et al., 1985; Spry et al., 1988; Both, 1990; Toteff, 1999; Gun, 1998). A discontinuous unit of quartz-biotite-garnet-andalusite ± staurolite rock, which extends intermittently for over ~30 km from about 10 km north of Kanmantoo toward Strathalbyn, hosts many of the Cu-Au and Pb-Zn-Ag-(Cu-Au) deposits and is considered to be a regional stratiform alteration zone that was subsequently metamorphosed (Seccombe et al., 1985).

A U-Pb age of zircon in the Sellick Hill tuff near the top of the Normanville Group yields a maximum age for the Kanmantoo Group of 522 ± 2 Ma (Jenkins et al., 2002). The minimum age (514 ± 3 Ma) of the Kanmantoo Group was obtained by Foden et al. (2006) using zircon U-Pb ages in the earliest syn-deformational magmatic intrusion (i.e., Rathjen Gneiss). Based on these dates, it is apparent that the Kanmantoo Group was rapidly deposited over 8 ± 5 Myr. Igneous activity related to the Kanmantoo Group includes syn-tectonic I- and S-type granites (e.g., Rathjen Gneiss, Palmer Granite, Encounter Bay Granite; Milnes et al., 1977; Foden et al., 2002b), and MORB-like intrusions (Chen and Liu, 1996) followed by late-Delamerian E-MORB tholeiite dikes (Foden et al., 2002a) and post-Delamerian (490–480 Ma) tholeiite dikes, plutons, and A-type granites (Turner et al., 1992; Turner, 1996; Foden et al., 1999).

The onset of the Delamerian Orogeny in the Late Cambrian (514 ± 3 Ma; Foden et al., 2006) terminated sedimentation in the Kanmantoo Trough and included three primary deformation events (Offer and Fleming, 1968; Fleming and White, 1984; Parker, 1986; Both, 1990; Mancktelow, 1990; Preiss, 1995; Oliver et al., 1998). Based on the structural and magmatic history, Fleming and White (1984) suggested compression spanned approximately 35 Myr, whereas later U-Pb and Rb-Sr ages by Foden et al. (2006) proposed a shorter period (~24 Myr) of orogenesis from 514 to 490 Ma during which time uplift, cooling, and extension related to post-tectonic magmatism commenced. The orogeny is characterized by a metamorphic gradient that reflects higher temperatures in the north and east, ranging from greenschist facies (~350–400 °C, biotite zone) to upper amphibolite facies (~650–700 °C, fibroite/stilliminate zone) with local zones of migmatite near the Palmer Granite (Fleming and White, 1984; Mancktelow, 1990; Sandiford et al., 1990; Dymoke and Sandiford, 1992; Oliver et al., 1998;
Based on a Sm-Nd age of monazite, xenotime, and apatite in migmatite near the Palmer Granite, Hammerli et al. (2014) concluded that peak metamorphism occurred at 512 ± 22 Ma. Proposed heat sources during metamorphism include crustal extension due to mantle upwelling and syn-tectonic magmatism (Sandiford et al., 1990; Chen and Liu, 1996). Orogenesis was proceeded by post-tectonic (490–480 Ma) bimodal magmatism comprising A-type granite intrusions southeast of the Adelaide Fold Belt and mafic intrusions at Black Hill (Milnes et al., 1977; Turner et al., 1992; Turner, 1996; Foden et al., 1999).

The first deformation event (D₁) was also the most penetrative, producing a regional slaty cleavage (S₁) and major upright folds (F₁). The second deformation event (D₂) was accompanied by peak metamorphic conditions and overprinted D₁ structures with a weak crenulation (S₂) of D₁ slaty cleavage, and weak folds (F₂, e.g., Kanmantoo syncline; Talbot, 1964; Preiss, 1987; Mancktelow, 1990; Belperio et al., 1998). The final deformation event (D₃) appears as open folds (F₃), faults, and a weak crenulation (S₃) localized in the eastern Mount Lofty Range (Offiler and Fleming, 1968; Mancktelow, 1990). One of the major structural features of the Kanmantoo Group in the eastern Mount Lofty
Range is the southerly plunging Kanmantoo syncline, which formed as a result of west-vergent compression during D2 (Thomson and Horowitz, 1962; Mancktelow, 1990; Preiss, 1995; Belperio et al., 1998; Totell, 1999; Schiller, 2000). The hinge of the syncline trends north–south subparallel to the major Kanmantoo and Bremer Faults to the east (Fig. 1).

3. Sampling, analytical methods, and data classification

3.1. Sample collection and analytical methods

Approximately 150 polished thin sections were prepared from samples collected from diamond drill cores KTDD149, KTDD54, KTDD34, KTDD141, and KTDD125 from the Kavanagh, Emily Star, and Nugent orebodies (Fig. 2). Major element compositions (2490 analyses) of garnet, biotite, staurolite, chlorite, muscovite, and magnetite were obtained using a JEOL JXA-8900 Electron Probe Microanalyzer at the University of Minnesota. Analyses were conducted using a 15 kV accelerating voltage with a 20nA beam current, a 1–2 μm spot size, and mineral standards including hornblende (Si, Al, Mg, Ca), ilmenite (Ti, Fe), albite (Al, Na), spessartine (Al, Mn), pyrope (Si, Mg), K-feldspar (K), garnetite (Zn, Al), tugtupite (Cl), and apatite (F). Chlorite and biotite are reported as $X_{Fe}$, which is the molar proportion of total iron relative to iron and magnesium (i.e., $X_{Fe} = \text{Fe}/(\text{Fe} + \text{Mg})$).

Trace element concentrations (1123 analyses) were measured at the U.S. Geological Survey in Denver, Colorado using laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS), comprising a Photon Machines Analyte LA system (excimer 193 nm) coupled to a PerkinElmer DRC-e ICP-MS. Depending on the grain size and presence of inclusions, minerals were ablated using spot sizes of 30, 40, 65, 85, 110, and 135 μm. Analyses were externally calibrated using the basalt glass microbeam reference standard GSD-1G (USGS), and internally standardized using values of Al (garnet, biotite, staurolite, chlorite, muscovite) and Fe (magnetite) from electron microprobe analyses. GSD-1G was analyzed three times during a 10-h day to correct for instrument sensitivity drift. Concentrations of elements were determined using off-line calculations, following the procedures of Longerich et al. (1996). Signals were screened for microinclusions before further data processing (Nadoll and Koenig, 2011). Trace element concentrations

![Fig. 2. Geologic map of the Kanmantoo deposit area, showing the location of meta-exhalites, orebodies, and drill core locations from this study. 1. KTDD54, Kavanagh. 2. KTDD149, Kavanagh. 3. KTDD141, Nugent. 4. KTDD125, Nugent. 5. KTDD34, Emily Star (modified after Totell, 1999).](https://example.com/fig2.png)
that are typically above detection limits for all minerals include $^{24}\text{Mg}$, $^{27}\text{Al}$, $^{28}\text{Si}$, $^{47}\text{Ti}$, $^{51}\text{V}$, $^{52}\text{Cr}$, $^{55}\text{Mn}$, $^{57}\text{Fe}$, $^{59}\text{Co}$, $^{60}\text{Ni}$, $^{66}\text{Zn}$, and $^{71}\text{Ga}$, whereas other elements above detection limits are mineral specific.

### 3.2. Data classification

Trace element data were pre-treated following the method of Croghan and Egeghy (2003) such that LA-ICP-MS results that include up to 40% censored data were substituted with the detection limit divided by the square root of 2. To unconstrain the data used for multivariate statistics and more easily recognize patterns in the dataset, a centered log-ratio transformation was applied using CoDaPack version 2.0.15 (Murithi, 2015). Using a multivariate statistical approach, trace element data for garnet and biotite were discriminated through principal component analysis (PCA) on JMP Pro version 11.0.0. Principal component analysis is a statistical method that extracts the dominant sources of variation (i.e., principal components are the eigenvectors of a variance–covariance matrix) in a multivariate dataset, thereby distinguishing trends in large datasets (Davis, 2002). In the present study, the following elements were analyzed: Li, Mg, Ca, Sc, Ti, V, Cr, Mn, Fe, Co, Ni, Zn, Ga, Ge, Sr, Y, Sn, Cs, Ti, and Pb, although the elements Ni, Sr, Sn, Cs, Ti, and Pb were generally below detection limits and not discussed further.

### 4. Local geology

#### 4.1. Lithology

The Kanmantoo Cu-Au deposit is spatially associated with three major rock types: quartz-mica schist (QMS), a quartz-biotite-garnet-andalusite ± chlorite ± staurolite ± magnetite rock, and biotite-garnet-chlorite rock, the last of which is the immediate host to much of the sulfide mineralization. Minor rock types include exalhites (e.g., quartz garnetite, garnet-quartz-cummingongite schist, plagioclase rock, banded iron formation), garnite-bearing mica schist, pyritic schist, and calc-silicate rocks (Verwoerd and Cleghorn, 1975; Seccombe et al., 1985; Oliver et al., 1998; Toteff, 1999; Schiller, 2000).

Quartz-mica schist consists principally of fine-grained quartz, biotite, muscovite, and feldspar with lesser garnet, chlorite, magnetite, ilmenite, and sulfides (Fig. 3A; Spry et al., 1988). Whole rock compositions obtained by Spry (1976) and Drexel (1978) show that QMS contains 30–125 ppm Cu, 13–26 ppm Pb, and 68–175 ppm Zn. Relict bedding is defined by alternating bands of quartzofeldspathic and micaeous schists, which are thought to represent original psammitic and pelitic rocks (Schiller, 2000).

The quartz-biotite-garnet-andalusite ± chlorite ± staurolite ± magnetite and garnet-biotite-chlorite rocks are interpreted to be metamorphosed hydrothermally altered rocks. Whole rock analysis by Oliver et al. (1998) and Schiller (2000) show that these rocks involved a loss of K, Ca, Na, and a gain of Fe, Mg, Si, and S during the alteration process. One common variant, garnet-andalusite-biotite schist (GABS), occurs as a series of lenses that run north–south through the Kanmantoo deposit for approximately 10 km in a N-S direction (Schiller, 2000; Wilson, 2009). Common minerals include garnet, andalusite, biotite, and quartz with minor staurolite, muscovite, chlorite, magnetite, ilmenite, rutile, and sulfides (Fig. 3B-D). Bulk rock compositions of GABS show that they contain 28–165 ppm Cu, 32–38 ppm Pb, and 71–113 ppm Zn (Spry, 1976; Drexel, 1978). Banding, which Wilson (2009) interpreted as relict bedding, is marked by coarse-grained layers of andalusite that grade into a finer-grained assemblage of andalusite-biotite-quartz. Andalusite porphyroblasts are typically 5 mm in length but can range up to 3 cm, and occur within a well-developed schistose biotite-quartz matrix.

Garnet-andalusite-biotite-staurolite schist (GABS) is a common variety of GABS that is enriched in staurolite and chlorite, and depleted in quartz
(Fig. 3E). It contains garnet, andalusite, biotite, staurolite, and chlorite with lesser quartz, sulfides, and magnetite. A weak foliation is defined by coarse-grained biotite and muscovite. Where staurolite is the dominant mineral, it occurs with biotite as elongate grains up to 1.5 cm across. Staurolite-poor zones are characterized by finer-grained staurolite that occurs within a biotite-garnet-chlorite matrix (Toteff, 1999; Schiller, 2000).

Biotite-garnet-chlorite schist (BGCS) occurs within GABS and GABSS and immediately envelopes Cu-Au mineralization in the Kavanagh and Emily Star orebodies (Fig. 3F and G). The typical mineral assemblage consists of biotite, garnet, chlorite, quartz, and minor pyrrhotite, chlorapatite, and magnetite. However, Spry et al. (1988) also noted other assemblages in the altered rock package that include quartz-chlorite-garnet-biotite ± chalcopyrite, staurolite-biotite-garnet ± pyrrhotite ± chlorapatite, staurolite-biotite-garnet-chlorite ± chlorapatite, and staurolite-chlorite-magnetite ± pyrrhotite ± chalcopyrite.

Exhalative rocks, using the terminology of Ridler (1971) and Parr and Plimer (1993), were recognized by Toteff (1999) in QMS and GABS within 3 km of the Kanmantoo mine (Fig. 3H and I). The most prominent exhalative horizon occurs within and along strike from the Nugent orebody in a north–south trending direction. In the Nugent orebody, exhalative rocks are manifested as a 10-m thick spessartine garnet-quartz-rock (i.e., quartz garnetite) that consists of garnet, muscovite, quartz, biotite, and lesser pyrite, pyrrhotite, and chlorapatite. Northeast and east of the Kavanagh orebody an exhalative unit consisting of banded iron formation, up to 19 m wide, contains quartz, magnetite, cummingtonite, and minor garnet, which extends intermittently for ~1.8 km, is the most prominent exhalative unit in the Kanmantoo Group (Smith, 1998; Toteff, 1999). Between the Nugent orebody and the banded iron formation is an outcrop of garnite-bearing muscovite schist. Other exhalative rocks identified by Toteff (1999) include magnetite-rich andalusite schist and albite. At the Scotts Creek Pb-Zn-Fe prospect, 2.5 km SW of the Kanmantoo mine, sul-

5. Results

5.1. Mineral compositions

Analyses of silicates (garnet, biotite, staurolite, chlorite, and muscovite) and oxides (magnetite) were obtained from 104 samples from five drill holes intersecting the Kavanagh, Emily Star, and Nugent orebodies. Representative major, minor, and trace element compositions of garnet, biotite, staurolite, chlorite, muscovite, and magnetite in garnet-andalusite-biotite schist (GABS), garnet-andalusite-biotite-staurolite schist (GABSS), and biotite-garnet-chlorite schist (BGCS) at Kanmantoo are given in Table 1. Supplementary files of major element and trace element data collected for this study are available in electronic supplement that accompanies this paper.

Major element compositions of garnet, biotite, staurolite, and muscovite in unmineralized altered and unaltered schist from the Kanmantoo Group were derived from the studies of Bollenhagen (1993), Smith (1998), and Schiller (2000) estimated a peak metamorphic pressure of 3–4 kbar with peak temperatures of 530–565 °C and 480–510 °C at the Kanmantoo deposit based on phase equilibria and mineral geothermometry, respectively. Note that although Oliver et al. (1998) placed garnet growth syn- or post-peak metamorphism, Schiller (2000) suggested that garnet grew syn- to post-D1 but pre-peak metamorphism, which is associated with D2.

4.3. Mineralization

Copper mineralization occurs as stockwork vein selvages and pipe-like podiform lenses dominated by chalcopyrite with lesser amounts of pyrrhotite, magnetite, pyrite, ilmenite, marcasite, sphalerite, cubanite, chalcocite, covellite, pentlandite, mackinawite, cobaltite, molybdenite, wolframite, bismuthinite, native bismuth, laitakarite, Se-bearing galena, native gold, and native silver (Fig. 3J–L; Secombe et al., 1985; Faulkner, 1996; Gum, 1998; Oliver et al., 1998; Schiller, 2000; Abbott et al., 2005; Arbon, 2011). Sulfide lodes are commonly discordant to relict bedding in the Kavanagh orebody and are enclosed in lenses of GABS and GBCS that steeply plunge 80° northeast (Secombe et al., 1985; Flöttmann et al., 1998; Gum, 1998; Toteff, 1999). Sulfides also occur within S2-parallel veins, shear, and fault zones, and as massive pockets, bands, and disseminations parallel to relict bedding (Preiss, 1995; Gum, 1998; Oliver et al., 1998; Abbott et al., 2005). Although sulfides occur in lenses of GABS, Schiller (2000) found that lodes associated with BGCS contain the highest copper grades. The stratiform nature of the Nugent orebody can be observed in drill core where sulfides parallel alternating layers of garnet and quartz, and because of the discordancy of sulfides to bedding in adjacent drill holes (David Rawlins, personal communication, September 2017). Unfortunately, we were unable to observe sulfide mineralization in the Nugent pit because of a wall collapse that covered mineralization prior to the current study.

Chalcopyrite in the Kanmantoo deposit occurs as disseminated grains, generally up to 4 mm in length, intergrown with pyrrhotite and secondary pyrite, and also as massive bands and veins. Sulfide ± quartz veins are typically cm- to m-scale in length and range from 1 to 15 mm across (Schiller, 2000). Chalcopyrite grains may also contain inclusions of pyrrhotite, magnetite, sphalerite, secondary pyrite, native bismuth, bismuthinite, galena, and native gold. Pyrrhotite occurs as abundant anhedral grains. Magnetite is commonly associated with chlorite, and occurs as euhedral to subhedral grains up to 0.5 mm in diameter. Three generations of pyrite are noted: pre-metamorphic colloform pyrite with inclusions of magnetite, ilmenite, sphalerite, and chalcopyrite, late syn-metamorphic euhedral pyrite in veins with chalcopyrite, and supergene botryoidal melnikovite pyrite locally intergrown with marcasite that is likely after pyrrhotite along with late cavity-filling supergene pyrite (Secombe et al., 1985). Schiller (2000) and Arbon (2011) observed two generations of chlorite, an early Fe-poor (Mg-rich) chlorite and a later Fe-rich chlorite hosting magnetite and bismuth minerals.
Representative compositions of garnet, biotite, staurolite, chlorite, and muscovite.

Table 1

<table>
<thead>
<tr>
<th>Element</th>
<th>Garnet</th>
<th>Biotite</th>
<th>Staurolite</th>
<th>Chlorite</th>
<th>Muscovite</th>
</tr>
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<td>wt%</td>
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<td></td>
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<tr>
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<td>100.27</td>
<td>100.02</td>
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Number of atoms in formulae (garnet oxygen basis 12, staurolite 23, biotite 12, chlorite 38, muscovite 11, magnetite 4)

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<th>Element</th>
<th>Garnet</th>
<th>Biotite</th>
<th>Staurolite</th>
<th>Chlorite</th>
<th>Muscovite</th>
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<td>0.075</td>
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<tr>
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<td>–</td>
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</tr>
<tr>
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<td>3.58</td>
<td>11.36</td>
<td>11.49</td>
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<td>14.704</td>
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Table continued on next page...
Hammerli et al. (2016), and compared with the compositions of silicates and oxides in mineralized altered rocks that were obtained here. A limited number of trace element data for biotite, muscovite, and garnet in unmineralized country rock schist are given in Hammerli et al. (2016) and are also compared with data obtained for these minerals in the present study. Note that end-member garnet compositions shown in the supplementary file take into account Fe in both the tetrahedral (Fe$^{2+}$) and octahedral (Fe$^{3+}$) sites, although the Fe content in the latter site is minor to nonexistent in nearly all analyses.

### 5.1. Garnet

End-member compositions of garnet from GABS, GABSS, BGCS, quartz-mica schist (QMS), and quartz garnetite are plotted relative to stratigraphy and assay values of Cu and Au in the five drill holes studied here (Fig. 4), along with a ternary diagram (Fig. 5), and bimodal plots of Ca vs. Mn (Fig. 6A) and Fe vs. Mn (Fig. 6B). Figs. 5 and 6A and B also show the compositions of garnet from unmineralized altered and unaltered schist from Bollenhagen (1993), Smith (1998), and Hammerli et al. (2016). Garnet in these schists shows the following range of compositions: 0.25–2.3 wt% CaO, 1.6–7.4 wt% MnO, and 29.4–39.0 wt% FeO. Garnet in GABS, GABSS, and BGCS from the Kavanagh and Emily Star orebodies and drill hole KTDD125 through the Nugent orebody are generally more Fe-rich (31.1–42.0 wt% FeO), and are markedly different in composition from garnet in GABS and quartz-garnetite in drill hole KTDD14 that intersects the northern end of the Nugent orebody (17.8–33.0 wt% FeO, 5.07–19.5 wt% MnO, 1.38–5.01 wt% CaO; Table 2).

A principal component analysis (PCA, n = 384) of major, minor, and trace elements that are typically above detection limits (i.e., Li, Mg, Ca, Sc, Ti, V, Cr, Mn, Fe, Co, Zn, Ga, Ge, and Y) was undertaken to further identify compositional variations for garnet in altered rocks associated with the Kavanagh, Emily Star and Nugent orebodies (Fig. 7A–C). The scores of the first two principal components are represented along the x- and y-axes on the bivariate score plot (Fig. 7A). Scores for garnet from the Kavanagh orebody and drill hole KTDD125 from the Nugent orebody generally overlap, and can be distinguished from garnet in altered rocks in drill hole KTDD141 (Nugent) that are offset to the left in Fig. 7A. Garnet in Emily Star possesses compositions that transition between these two clusters of data. Principal component 1 (PCA1) accounts for 39.9% of variance and reveals a negative correlation between Ca, Ti, Mn, Zn, Ge, and Y, and Li, Mg, Sc, V, Cr, Fe, Co, and Ga, whereas PCA2 accounts for 11.9% of variance and shows a negative correlation between Li, Sc, V, Mn, Fe, Ga, Ge, and Y, and Mg, Ca, Ti, Cr, Co, and Zn (Fig. 7C).

Major and trace element bivariate plots of garnet (in ppm) are presented in terms of Zn vs. Mn (Fig. 5A), Zn vs. Gd (Fig. 5B), Eu vs. Gd (Fig. 5C), and Co vs. V (Fig. 5D). The plots of Zn vs. Mn, and Zn vs. Gd reflect the patterns shown in the PCA in that the composition of garnet in drill hole KTDD141 (Nugent) are distinguished, primarily due to the enrichment in Zn for garnet in altered rocks from Kavanagh and KTDD125 (Nugent), with the compositions of garnet from Emily Star forming a transition between these two clusters of data. Set apart from the clustered samples in Fig. 5A, garnet in Nugent (KTDD141) occurs as an isolated group of data characterized by elevated Zn (up to 170 ppm) and Mn (up to 19.5 wt% MnO). For comparison, garnet analyzed from pelitic and psammitic rocks in the Tapanappa Formation, metamorphosed to amphibolite facies ~25 km northeast of Kanmantoo, contains 13.6–16.8 ppm Zn and 1.6–7.4 wt% MnO (Bollenhagen, 1993; Smith, 1998; Hammerli et al., 2016). Enrichment in Ge (0.23–27.8 ppm) and Gd (~10.1–66.0 ppm) in garnet from both drill holes intersecting the Nugent orebody can be distinguished from those associated with the Kavanagh and Emily Star orebodies, which generally have Ge values < 10 ppm and Gd values < 24 ppm (6.1–16.9 ppm Gd in Hammerli et al., 2016; Fig. 5C). Additionally, garnet in Nugent (KTDD141) is depleted in Co (< 18 ppm) relative to garnet in altered rocks from the Kavanagh and Emily Star orebodies (4.77–93.5 ppm Co; Fig. 5D). In a plot of Co vs. V (Fig. 5D), the trace element compositions of garnet from the two Kavanagh drill holes overlap. Garnet in altered rocks associated with drill hole KTDD149 are generally more enriched in Co (up to 78.5 ppm), whereas those from KTDD54 commonly have V concentrations < 20 ppm. Those of Hammerli et al. (2016) range from 32.1 to 43.9 ppm V.

Rare earth element (REE) patterns for garnet normalized to chondrite values of McDonough and Sun (1995) show enrichment in heavy REEs (HREE) relative to light REEs (LREE; Fig. 9A). Enrichment of Gd (0.23–27.8 ppm) and Gd (~10.1–66.0 ppm) in garnet from both drill holes intersecting the Nugent orebody can be distinguished from those associated with the Kavanagh and Emily Star orebodies, which generally have Ge values < 10 ppm and Gd values < 24 ppm (6.1–16.9 ppm Gd in Hammerli et al., 2016; Fig. 5C). Additionally, garnet in Nugent (KTDD141) is depleted in Co (< 18 ppm) relative to garnet in altered rocks from the Kavanagh and Emily Star orebodies (4.77–93.5 ppm Co; Fig. 5D). In a plot of Co vs. V (Fig. 5D), the trace element compositions of garnet from the two Kavanagh drill holes overlap. Garnet in altered rocks associated with drill hole KTDD149 are generally more enriched in Co (up to 78.5 ppm), whereas those from KTDD54 commonly have V concentrations < 20 ppm. Those of Hammerli et al. (2016) range from 32.1 to 43.9 ppm V.

### 5.1.2. Biotite

Biotite from QMS, GABS, GABSS, and BGCS in the Kavanagh, Emily Star, and Nugent orebodies show a compositional range with values of 13.8–25.9 wt% FeO, 6.39–13.2 wt% MnO (Fig. 10A), intermediate XFe ratios of 0.38 to 0.67, and Al$^3+$ of 2.59 to 2.81 (Fig. 10B). Biotite with the lowest XFe ratios are those in the Nugent orebody (0.38 to 0.52) that...
were sampled from drill hole KTDD141. Although limited in number, biotite analyzed by Bollenhagen (1993) and Hammerli et al. (2016) from country rock mica schists in the Kanmantoo Group have similar $X_{Fe}$ ratios to those in rocks spatially associated with the Kanmantoo orebodies, but have a wider range of Al$^{IV}$ values (2.37 to 2.99). In general, biotite in unmineralized schist is enriched in Mn and depleted in Al relative to those associated with Cu-Au mineralization.

Representative trace element compositions of biotite are shown in
observed for the PCA for garnet, samples of biotite in QMS from KTD141 (Nugent) plot separately and can be distinguished from the other samples analyzed here primarily due to the high score value of Zn.

5.1.3. Staurolite
Staurolite in QMS, GABS, GABSS, and BGCS are Fe-rich (12.1–17.2 wt% FeO, $X_{Fe} = 0.38–0.69$). Of the 62 samples of staurolite analyzed here, 61 contain up to 1.60 wt% ZnO, with one outlier sample (KTD125–2) that contains up to 3.20 wt% ZnO (Fig. 11A). The major element compositions are similar to those reported previously from the Kanmantoo deposit by Schiller (2000) that contain up to 1.32 wt% ZnO. Trace element analyses show that staurolite also contains elevated concentrations of Mn (46–2070 ppm), Co (23.4–344 ppm), and V (14.9–285 ppm; Table 1). Bivariate plots (in terms of ppm) are shown for Zn vs. Mn (Fig. 11B) and Co vs. V (Fig. 11C). Although drill holes KTD149 and KTD54 intersect the Kavanagh orebody, Zn and Co values are markedly different in that most samples from KTD149 contain up to 16,230 ppm Zn and 344 ppm Co, whereas samples from drill hole KTD54 contain < 9,622 ppm Zn and < 121 ppm Co (Fig. 11C).

5.1.4. Chlorite
As was recognized previously by Schiller (2000) and Arbon (2011) and verified here, chlorite consists of an early Mg-rich variety ($X_{Fe} = 0.71–0.95$; Fig. 12A). Although both generations are associated with sulfide mineralization, the Fe-rich chlorite is spatially associated with magnetite and Bi mineralization. Chlorite from GABS, GABSS, and BGCS in the Kavanagh, Emily Star, and Nugent (KTD141) orebodies cluster toward the epidotoid end-member, whereas those from KTD125 (Nugent) occur over a broader compositional range. Magnesium-rich chlorite in drill hole KTD141 (Nugent) has a narrow compositional range of 15.8–18.3 wt% MgO and 19.1–22.3 wt% FeO, whereas Mg-rich chlorite in Kavanagh and Nugent (KTD125) ranges from 11.5 to 14 wt% MgO and 24.4–29.8 wt% FeO. Those from Emily Star are characterized by less Mg (6.28–10.8 wt% MgO) and more Fe (29.3–35.4 wt% FeO) relative to chlorite from Nugent and Kavanagh. Conversely, Fe-rich chlorite from Kavanagh (KTD54) and Nugent (KTD125) are more Fe-rich (24.6–40.7 wt% FeO) and Mg-poor (3.40–13.9 wt% MgO; Fig. 12B).

Other than the major elements (Mg, Fe, Al, and Si), chlorite is enriched in Mn, Co, Zn, Ba, and Pb (Table 1). Mg-rich chlorite from KTD141 (Nugent), contains > 1780 ppm Mn and 725 to 1600 ppm Zn (Fig. 12C), whereas Fe-rich chlorite from Kavanagh and Nugent (KTD125) contains 153–628 ppm Mn and 106–202 ppm Zn. It is noted

---

**Table 1.** Bivariate plots (in ppm) for Zn vs. Mn, and Zn vs. Pb are shown in Figs. 10C and D, respectively. In these plots, biotite in KTD141 is distinguished from biotite in other drill holes due to an enrichment in both Mn and Zn (> 1070 ppm Mn and > 410 ppm Zn). Relative to biotite compositions in unaltered country rock schist obtained by Hammerli et al. (2016), which contain 23.6–1592 ppm), biotite in most samples from the Kanmantoo deposit are elevated in Zn (413–1592 ppm). Of the 62 samples of staurolite analyzed here, 61 contain up to 1.60 wt% ZnO, with one outlier sample (KTD125–2) that contains up to 3.20 wt% ZnO (Fig. 11A). The major element compositions are similar to those reported previously from the Kanmantoo deposit by Schiller (2000) that contain up to 1.32 wt% ZnO. Trace element analyses show that staurolite also contains elevated concentrations of Mn (46–2070 ppm), Co (23.4–344 ppm), and V (14.9–285 ppm; Table 1). Bivariate plots (in terms of ppm) are shown for Zn vs. Mn (Fig. 11B) and Co vs. V (Fig. 11C). Although drill holes KTD149 and KTD54 intersect the Kavanagh orebody, Zn and Co values are markedly different in that most samples from KTD149 contain up to 16,230 ppm Zn and 344 ppm Co, whereas samples from drill hole KTD54 contain < 9,622 ppm Zn and < 121 ppm Co (Fig. 11C).

---

**Fig. 5.** Ternary plot of average garnet composition for individual samples from the Kanmantoo deposit and country rock schist (Bollenhagen, 1993; Smith, 1998; Hammerli et al., 2016) shown by an ellipse. The compositional field of garnet associated with Pb-Zn-Ag-(Cu-Au) deposits is shown as a separate ellipse.

**Fig. 6.** Compositions of garnet from the Kanmantoo deposit and country rock schist (noted by an ellipse; Bollenhagen, 1993; Smith, 1998; Hammerli et al., 2016). A. Ca (apfu) vs. Mn (apfu). B. Fe (apfu) vs. Mn (apfu).
here that chlorite in Emily Star is characterized by an enrichment in Cu (up to 593 ppm) relative to all other samples that typically contain <10 ppm Cu. The reason for this is unclear, although it is possible that the laser ablation analysis contained an inclusion of chalcopyrite.

5.1.5. Muscovite

Trace element compositions of muscovite from QMS, GABS, and BGCS that are generally above detection limits include Na (up to 593 ppm), relative to all other samples that typically contain <10 ppm Cu. The reason for this is unclear, although it is possible that the laser ablation analysis contained an inclusion of chalcopyrite.

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<th>MnO wt%</th>
<th>MgO wt%</th>
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<td>Gd ppm ppm</td>
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Compositions of garnet, biotite, and muscovite are shown for unmineralized metapelites and metapsammites in the Kanmantoo Group compared with compositions characteristic of those from QMS and quartz garnetite from KTDD141 (Nugent) and those from GABS, GABSS, and BGCS from KTDD54, KTDD149, KTDD34, and KTDD125 (Kavanagh, Emily Star, Nugent); Major element composition of muscovite in unmineralized metapelites and metapsammites from the Kanmantoo Group compared to those in (2); Chlorite compositions at the Kanmantoo deposit only for (1) those in QMS and quartz garnetite from KTDD141 (Nugent) compared to those in (2), and late Fe-rich chlorite from Nugent and Kavanagh.

Table 2

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<td>MgO wt%</td>
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<tr>
<td>CaO wt%</td>
<td>0.00–0.03</td>
<td>0.04–0.04</td>
</tr>
<tr>
<td>ZnO wt%</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>V ppm</td>
<td>15.8–116</td>
<td>40.9–370</td>
</tr>
<tr>
<td>Co ppm</td>
<td>0.97–2.28</td>
<td>0.37–7.74</td>
</tr>
<tr>
<td>Cu ppm</td>
<td>3.86–14.42</td>
<td>0.25–173</td>
</tr>
<tr>
<td>Zn ppm</td>
<td>77.6–108</td>
<td>0.51–77.6</td>
</tr>
<tr>
<td>Ge ppm ppm</td>
<td>0.13–1.53</td>
<td>0.09–5.00</td>
</tr>
<tr>
<td>Gd ppm ppm</td>
<td>–0.07</td>
<td>0.06–1.17</td>
</tr>
<tr>
<td>Pd ppm ppm</td>
<td>102–118</td>
<td>0.02–272</td>
</tr>
</tbody>
</table>

5.1.6. Magnetite

Magnetite is intergrown with chlorite, chalcopyrite, and pyrrhotite in GABS, GABB, and BGCS. Concentrations above detection limits were observed for Al (1,375–7,070 ppm), Ca (50–12,000 ppm), Ti (265–4,520 ppm), V (291–5,085 ppm), Cr (1.0–1,285 ppm), Cu (0.01–23.200 ppm), Zn (0.10–12,715 ppm), and Ga (3.3–252 ppm; Table 1). A bivariate plot (in ppm) of Ni vs. Cr shows that relative to magnetite compositions obtained by Tedesco (2009), magnetite analyzed here from Emily Star is elevated in Ni (up to 407 ppm), whereas some samples from Kavanagh are elevated in Cr (up to 1,285 ppm; Fig. 14A). A plot of Sn/Ga (ppm) vs. Al/Co (ppm) was developed by Singoyi et al. (2006) to assess provenance of hydrothermal magnetite from skarns, iron oxide-copper-gold, Broken Hill-type Pb-Zn-Ag, and volcanic-hosted massive sulfide (VHMS) deposits. Although there is no designated field for SEDEX deposits, it should be noted here that most magnetite from the Kanmantoo deposit plots within the VHMS field (Fig. 14B).

6. Discussion

6.1. The origin of the Kanmantoo deposit

In order to assess the origin of the Kanmantoo deposit, the following characteristics need to be considered in light of the pre-metamorphic syn-sedimentary (e.g., Seccombe et al., 1985; Toteff, 1999), metamorphic (e.g., Oliver et al., 1998; Schiller, 2000), and post-metamorphic models (e.g., Foden et al., 1999; Tedesco, 2009; Wilson, 2009; Arbon, 2011; Lyons, 2012) that have been proposed: 1) structural and textural relationships, 2) petrological, 3) mineral chemistry and sulfide assemblages, 4) stable isotopes, 5) composition and conditions of the ore-forming fluids, and 6) source of metals.
6.1.1. Structural and textural relationships

Most orebodies within the Kanmantoo Cu-Au deposit are variably dipping and typically subparallel to the S2 schistosity (Seccombe et al., 1985; Oliver et al., 1998; Schiller, 2000), although Wilson (2009) noted the relationship between mineralization and late-stage, east-dipping extensional faults. Mineralization in the Kavanagh orebody dips to the east and roughly parallels north-northeast trending shear zones, fractures, and faults in a series of randomly oriented sulfide-bearing veins and lenses (Verwoerd and Clegborn, 1975; Seccombe et al., 1985; Wilson, 2009). Mineralization in Emily Star locally parallels bedding and wraps around a south-plunging S2 fold, whereas ore lenses in the Nugent orebody trend southwest-northeast and parallel relict bedding (S0). The stockwork nature of the Kavanagh and Emily Star orebodies reinforces the concept proposed by Seccombe et al. (1985) that sulfide
mineralization formed in the feeder zone of a SEDEX deposit, which is observed elsewhere at, for example, the Rammelsberg Cu-Zn-Pb deposit, Germany (Leach et al., 2005). Locally, sulfide mineralization parallels bedding (Lindqvist, 1969; Wilson, 2009, Fig. 4, p. 55). The parallelism of the Nugent orebody to S0 and its discordancy to S2, however, is difficult to reconcile with syn-deformational (i.e., metamorphogenic or post-peak metamorphic) models. Post-orogenic sulfide quartz veins described by Tedesco (2009) likely resulted from the remobilization of syn-sedimentary sulfides.

An annealed metamorphic fabric and deformation twinning in chalcopyrite and pyrrhotite, the presence of an assumed pre-metamorphic alteration halo enriched in Fe and Mg, and depleted in Ca and Na (Seccombe et al., 1985; Spry et al., 1988; Belperio et al., 1998; Toteff, 1999), the presence of chalcopyrite and pyrrhotite in S1 inclusion trails in andalusite porphyroblasts (a D1 mineral), and colloform pyrite with magnetite-ilmenite-sphalerite-chalcopyrite inclusions suggests the presence of Cu ore prior to peak metamorphism (D2; Seccombe et al., 1985; Marshall and Gilligan, 1987; Spry et al., 1988; Schiller, 2000). Parker (1986) proposed the spatial association of Cu-Au mineralization to a northwest-trending shear zone, which may be a manifestation of a basement fault that reactivated during formation of the Kanmantoo Trough and was a potentially important control on mineralization.

6.1.2. Petrological

The immediate host to much of the Cu ore at Kanmantoo is biotite-garnet-chlorite schist enveloped in garnet-andalusite-biotite schist (an Fe-Mg-rich, Na-Ca-poor unit; Toteff, 1999). The garnet-andalusite-biotite and sulfide-enveloping biotite-garnet-chlorite rocks were considered by Seccombe et al. (1985), Faulkner (1996), and Toteff (1999) to be the manifestation of a metamorphosed alteration halo, whereas Gum (1998) and Oliver et al. (1998) suggested that they are a product of syn-metamorphic metasomatism. The spatial association of a garnet-andalusite-biotite rock is not unique to the Kanmantoo deposit area. This rock extends intermittently for > 30 km in a unit that passes through the Scotts Creek Pb-Zn prospect, the South Hill Cu-Au prospect, adjacent to the Wheal Ellen Pb-Zn-Ag-(Cu-Au) deposit and St. Ives prospects, and envelopes the Angas Pb-Zn-Ag deposit. The precursor to GABS was considered by Schiller (2000) to be a mixture of quartz, illite, chlorite, iron oxides and hydroxides, and minor plagioclase. The

![Fig. 8. Compositions of garnet from the Kanmantoo deposit and country rock schist (outlined in A, B; Hammerli et al., 2016) A. Zn (ppm) vs. Mn (ppm x 10³). B. Zn (ppm) vs. Gd (ppm). C. Ge (ppm) vs. Gd (ppm). D. Co (ppm) vs. V (ppm).](image-url)
metamorphic minerals are similar to those in aluminous stratabound hydrothermal alteration zones associated with metamorphosed base metal sulfide deposits elsewhere in the world (e.g., LaRonde Penna, Dubé et al., 2007; Snow Lake, Bailes et al., 2016). There is a spatial association of meta-exhalative rocks with Pb-Zn-Ag-(Cu-Au) and Cu-Au mineralization at the Strathalbyn, Wheal Ellen, Scotts Creek, Aclare, and Angas Pb-Zn-Ag-(Cu-Au) deposits, and the Kanmantoo Cu-Au deposit (Fig. 1; Toteff, 1999). At Kanmantoo, banded iron formation (quartz ± magnetite ± cummingtonite ± rock) occurs over a distance of > 1 km north of the deposit and along strike from gahnite-bearing rocks (Toteff, 1999). A stratigraphically equivalent unit of quartz garnetite rock occurs further south within and along strike from the Nugent orebody. Toteff (1999) also identified numerous outcrops of iron formation, quartz-garnet-cummingtonite rocks, and plagioclase rocks in the vicinity of the Kanmantoo deposit. Similar rocks also occur in and along strike from the Scotts Creek (quartz garnetite), Angas (quartz-garnet rock, quartz-gahnite rock), St. Ives (gahnite-bearing GABS), and Wheal Ellen deposits (iron formation, gahnite-rich rocks). All of these exhalative rocks occur at the same stratigraphic level within the Tapanappa Formation (the so-called Angas Garnet Member of Gum, 1998). Similar rocks have been described elsewhere in the world spatially associated with metamorphosed exhalative/inhalative massive sulfide deposits (i.e., Broken Hill-type, VHMS, and SEDEX; Spry, 2000; Spry et al., 2000; Slack, 2012; Steadman and Spry, 2015). Given that such exhalites are generally associated with metamorphosed syn-sedimentary deposits, it would be remarkable that the same types of exhalative units spatially associated with metamorphosed Pb-Zn-Ag-(Cu-Au) SEDEX deposits are also spatially associated with the Kanmantoo deposit if it has a syn-deformational origin (Marshall and Spry, 2000). Not only are the exhalative units associated with the Kanmantoo deposit the most extensive in the Kanmantoo Group, they also occur in and along strike from the Nugent orebody that is conformable to bedding.

6.1.3. Mineral chemistry and sulfide assemblages

Major element compositions of garnet, staurolite, and minor chlorite obtained here from samples of GABS, GABSS, and BGCS from the Kavanagh (KTDD54, KTDD149), Emily Star (KTDD34), and Nugent (KTDD125 only) orebodies, show that they are generally Fe-rich as was previously recognized by Schiller (2000), Wilson (2009), and Arbon (2011), whereas biotite has an intermediate composition (XFe = 0.38–0.67). Garnet associated with the Kavanagh, Emily Star, and Nugent (KTDD125 only) orebodies

![Diagram A](image_url1)

![Diagram B](image_url2)

![Diagram C](image_url3)

**Fig. 9.** Chondrite-normalized rare earth element patterns for garnet. A. KTDD054 and KTDD149, Kavanagh. B. KTDD34, Emily Star. C. KTDD141 and KTDD125, Nugent.
Fig. 10. Compositions of biotite from the Kanmantoo deposit and country rock schist (outlined; Bollenhagen, 1993; Hammerli et al., 2016) and principal component analysis (n = 217) using 10 elements, Sc, V, Co, Ni, Zn, Sr, Sn, Cs, Tl, Pb. A. Fe (apfu) vs. Mg (apfu). B. AlIV (apfu) vs. Fe/(Fe + Mg). C. Zn (ppm) vs. Mn (ppm). D. Zn (ppm) vs. Pb (ppm). E. Score plot of the first two principal components with percent variance for component 1 and 2 on the X and Y axis, respectively. F. Loading plot showing the vector representation of data projected onto the score plot for each element.
contains higher concentrations of FeO, lower amounts of MnO and about the same CaO contents as garnet in QMS and quartz garnetite in Nugent drill hole KTDD141. Not only are there differences in major element compositions between garnet from QMS and quartz garnetite, and those associated with GABS, GABSS, and BGCS in the other orebodies, but there are also differences in trace element compositions for both garnet and biotite as shown in the respective PCAs and associated loading plots (Fig. 7A, B and 10E, F). In terms of trace element compositions, garnet and biotite in QMS and quartz garnetite from KTDD141 (Nugent) contain elevated Zn (up to

Fig. 11. Compositions of staurolite. A. Histogram in terms of wt.% ZnO. B. Zn (ppm) vs. Mn (ppm). C. Co (ppm) vs. V (ppm). Not shown in 11A is a single outlier for sample KTDD125-2 that contains 3.2 wt% ZnO, which is an average of two analyses.

Fig. 12. Compositions of chlorite. A. Chlorite end-member compositions, including Fe-rich chlorite at Kanmantoo from Schiller (2000). B. FeO (wt.%) vs. MgO (wt.%) showing a negative linear relationship. C. Zn (ppm) vs. Mn (ppm).
170 ppm in garnet and 1,592 ppm in biotite) relative to these minerals in KTDD125 (Nugent), KTDD54 (Kavanagh), KTDD149 (Kavanagh), and KTDD34 (Emily Star). However, what is notable is that major and trace element compositions of garnet and biotite are similar to the compositions of these minerals spatially associated with the Wheal Ellen, Angas, and Scotts Creek Pb-Zn-Ag-(Cu-Au) occurrences in the Kanmantoo Group (100–252 ppm Zn in garnet and up to 1,200 ppm Zn in biotite; Tott et al. 2016). Garnet associated with these Pb-Zn-Ag-(Cu-Au) deposits contains 1.02–6.31 wt% CaO and 8.96–24.05 wt% MnO (Tottie, 1999), and is similar in composition to garnet in QMS and quartz garnite from KTDD141 (Nugent).

The primary metallic mineral assemblage in the Kanmantoo deposit consists of chalcopyrite, pyrrhotite, and magnetite with lesser amounts of primary and secondary pyrite (Seccombe et al., 1985), whereas that associated with Pb-Zn-Ag-(Cu-Au) deposits mostly contains sphalerite, galena, and primary pyrite, with lesser amounts of chalcopyrite, pyrrhotite, and magnetite. However, there are two exceptions among the Pb-Zn-Ag-(Cu-Au) deposits where zones of Cu form part of the deposit (Strathalbyn and Wheal Ellen). For example, Duncan et al. (1971) reported Pb-Zn and Cu zones along strike from each other at the Wheal Ellen deposit with Cu mineralization occurring in veins. Recent drilling by Hillgrove Resources at Wheal Ellen identified new zones of Cu-rich mineralization (1 m @ 2.55% Cu and 1 m Cu @ 1.74% Cu; http://www.hillgroveresources.com.au/article/SA_Exploration/Wheal_Ellen_Project?section_id=41). Petrographic studies of one of these zones show that typical massive sphalerite-galena-pyrite ± pyrrhotite ± chalcopyrite mineralization grades into the assemblage sphalerite-chalcopyrite-pyrite-magnetite ± pyrrhotite (Fig. 17A and B). The presence of magnetite along with chalcopyrite, which has not been described previously in ore from Wheal Ellen, resembles the chalcopyrite-magnetite assemblages seen at Kanmantoo, further suggesting a genetic link between the Cu-Au and Pb-Zn-Ag-(Cu-Au) deposits. Utilizing the ore fluid composition proposed by Seccombe et al. (1985) for formation of the Kanmantoo deposit, which was based on the stability of the primary assemblage pyrrhotite, chalcopyrite, magnetite, and minor pyrite, the bulk sulfur isotopic composition of the aqueous species in a slightly acid, low salinity, hydrothermal fluid associated with sediment-hosted and VHMS deposits, and the solubility of Cu as a copper chloride complex, a plot of log\(f_O_2\) versus temperature superimposed onto the system Fe-S-O, which also shows the solubilities of chalcopyrite and sphalerite (Large, 1977), reveals that a change in temperature of \(\sim 25^\circ C\) could account for the metallic mineral
assemblages at Kanmantoo and Wheal Ellen (Fig. 16). For the conditions shown in Fig. 16, the Cu-rich assemblage would form at ~265 °C (Kanmantoo), while the Cu-Zn-rich assemblage could conceivably form around 240 °C (Wheal Ellen), at approximately the same fO2 conditions.

6.1.4. Stable isotopes

Sulfur isotope studies of Cu-Au, Pb-Zn-Ag-(Cu-Au), and Fe-S deposits in the Kanmantoo Group were undertaken by Jensen and Whittle (1969), Seccombe et al. (1985), Gum (1998), and Lyons (2012), Seccombe et al. (1985) proposed a dual sulfur source of Cambrian seawater (δ34S = −17‰) and biogenic sulfur from the Nairne Pyrite Member (δ34S = −5‰). Isotopic values at the Kanmantoo Cu-Au deposit (δ34S = +8 ± 2.3‰) are generally heavier than those associated with the other Cu-Au (Bremer, South Hill) and Pb-Zn-Ag-(Cu-Au) occurrences (δ34S = −4.7 to +6.5‰). Seccombe et al. (1985) attributed the variability in the isotopic compositions for the various deposits as being the result of different degrees of mixing of sulfur from these two sources, and to variations in temperature, pH, δ34S, total S, ionic content, or fO2 during ore deposition. Both Oliver et al. (1998) and Marshall and Spry (2000) advocated that the sulfur isotope compositions of sulfides do not constrain a model of formation for the Kanmantoo deposit and could allow for both syn-sedimentary and syn-deformational origins. Oliver et al. (1998) preferred a felsic igneous rock source for sulfur at the Kanmantoo deposit, but the closest outcropping felsic body (Monarto granite) occurs 13 km to the east of the Kanmantoo deposit (Schiller, 2000). An intrusive body below the Kanmantoo deposit, is yet to be reported in the literature.

As with sulfur isotopes, ore models based on oxygen isotopes are not well constrained. Oliver et al. (1998) advocated a 500 m zone of locally depleted whole-rock oxygen isotope values (δ18O = +6.4 to +9.3‰, relative to V-SMOW) within and adjacent to the Kanmantoo deposit that formed as a result of localized fluid flow derived from a crystallizing magma. However, Marshall and Spry (2000) suggested the isotopic anomaly could just as easily result from hydrothermal fluid flow related to the formation of massive sulfide deposits via syn-sedimentary hydrothermal exhalations (e.g., Nesbitt, 1996; Brauhart et al., 2000; Lerouge et al., 2001).

6.1.5. Composition and conditions of the ore-forming fluids

The syn-sedimentary model does not rely upon fluid inclusion data, as any primary fluid inclusions would have been erased during metamorphism. The lack of syn-depositional volcanics requires an internally derived heat source, such as an unusually high geothermal gradient (common in sedimentary exhalative deposits; Leach et al., 2005) in the sediments below the Tapanappa Formation, or a thinning lithosphere (Seccombe et al., 1985; Both, 1990). Due to the rapid rate of sedimentation in the Kanmantoo Trough (8 ± 5 Ma; Foden et al., 2006), and based on the location of the Nairne Pyrite Member, hydrothermal convective cells sourced from formation water and seawater circulated to at least 3–4 km depth and were subsequently concentrated along faults (Seccombe et al., 1985).

Oliver et al. (1998) suggested a two-phase fluid flow syn-metamorphic model for which the fluid was a mixture of metamorphic and magmatic components. Based on oxygen and sulfur isotopes and a hypothesized increase in metamorphic grade approaching the Kanmantoo deposit, Oliver et al. (1998) proposed the first phase involved the replacement of regional synorogenic granitic bodies that drove lateral and up-temperature flow of metamorphic fluids. The second phase localized ore fluids at the site of deposition, with the magmatic fluid component being derived from an unidentified granite body directly beneath the deposit. Regardless of the presence or absence of such an intrusive synorogenic body, or even a syn-sedimentary intrusive rock, which may have been responsible for the formation of the Fe-rich alteration assemblage spatially associated with the Kanmantoo deposit, such a single intrusion would not account for the regional stratabound alteration zone in the Tapanappa Formation.
The peak metamorphic model relies on fluid inclusion data in primary fluid inclusions formed during metamorphism in Cu-bearing quartz veins and titaniferous-in-quartz geothermometry (Focke et al., 2009; Schmidt Mumm et al., 2009; Tedesco, 2009). Tedesco (2009) showed they had high salinity fluid inclusions (26–32 NaCl eq. wt.%) that formed at 372 ± 2.7 °C. He attributed the high salinities to a magmatic input, despite the fact that fluids in sedimentary exhalative systems range from 10 to 30 wt% NaCl (Leach et al., 2004; Polito et al., 2006). We consider that the lower temperatures reflect the remobilization of pre-metamorphic Cu during post-peak metamorphism, as based on U-Th-Pb dates of Wilson (2009) on monazite in garnet in unmineralized rocks (492 ± 15 Ma) and sulfate-bearing rocks (492 ± 9 Ma).

6.1.6. Source of metals

Recent studies by Pitcairn et al. (2006), Yardley and Cleverley (2015), Zhong et al. (2015a), Hammerli et al. (2016), and Finch and Tomkins (2017) have evaluated the role in which metals are released to metamorphic fluids as a result of dehydration and desulfidation during prograde metamorphism. Studies of the Otago and Alpine schist, New Zealand by Pitcairn et al. (2006), and of metasediments in a metamorphic aureole surrounding the Ballachulish igneous complex, Scotland by Finch and Tomkins (2017), show that even though prograde metamorphism may release metals including Au, As, Bi, Sb, and W to a metamorphic fluid to conceivably form orogenic gold deposits, Cu, Zn, and Pb are not released to metamorphic fluids to the same extent. Although not specifically discussing the formation of Pb-Zn-Ag-(Cu-Au) or Cu-Au deposits in the Kanmantoo Group, Hammerli et al. (2015, 2016) noted that the Pb and Zn concentrations in staurolite-absent metasedimentary rocks and in biotite in the eastern part of the Kanmantoo Group decreased at higher metamorphic grades, the implication being that Pb and Zn could be released during dehydration to form Pb-Zn deposits. However, it should be noted that Cu is not released during prograde metamorphism, implying that such a mechanism was unlikely to have been responsible for the formation of the Kanmantoo deposit.

Alternatively, Oliver et al. (1998) proposed a late syn-tectonic model whereby Cu was released from a hidden syn-orogenic granite as it crystallized below the Kanmantoo deposit. In proposing a post-orogenic granitic source of metals for the Kanmantoo deposit, Arbon (2011) considered that Bi was a diagnostic indicator of such a source, in part because of the general association of Bi with magmatic ore deposits. However, it should be noted that Bi can occur in accessory phases in sedimentary exhalative deposits where it has been found, for example, in association with the metamorphosed sediment-hosted Roca de Ponent Cu deposit, Spain (Canet et al., 2003; Haldar, 2013).

Schiller (2000) suggested that MORB-like meta-dolerite dikes, plugs, and sills may have been a source of heat and possibly metals. However, the mafic rocks are younger than the metasedimentary rocks of the Kanmantoo Group, as indicated by a U-Pb zircon age of 510 ± 2 Ma, which suggests that some of the mafic rocks formed post-peak metamorphism (Chen and Liu, 1996). This late age negates the possibility that mafic rocks were a heat source for syn-sedimentary Cu-Au mineralization.

In considering the potential for the presence of SEDEX Pb-Zn deposits in the Kanmantoo Group other than those already identified, Gum (1998) did a simple mass balance calculation by considering the ~1 km thick Backstairs Passage Formation at the base of the Kanmantoo Group as a source of metals with 14 ppm Pb and 2 ppm Zn being leached from the sediments. The amount of metals leached was based upon the difference between the average metal content of sandstone and the median values of base metals in the Backstairs Passage Formation. A total of ~28 Mt Pb and 4 Mt Zn were determined from this calculation. Because he did not consider metals being leached from the Carrickalinga Head Formation, members of the Normanville Group, or pyritic units of the Talisker Formation, the amount of Pb-Zn derived from sediments based on his calculation is consequently a minimum. Using the same assumptions made by Gum (1998) for his calculation of Pb and Zn and considering a depletion of 17 ppm Cu in the Backstairs Passage Formation, his data would yield ~35 Mt of Cu. Such a preliminary calculation demonstrates that there should be more than enough base metals to generate all the known base metal deposits in the Kanmantoo Group from exhalative processes, including the Kanmantoo Cu deposit.

Based largely upon the results of their sulfur isotope study, Seccombe et al. (1985) suggested that the Kanmantoo deposit was formed from a hydrothermal fluid that circulated to the base of the Kanmantoo Group and interacted with the isotopically light pyritic units, including the Nairne Pyrite Member. They also suggested that these pyritic units could have been a source of metals. Although the Cu, Pb, and Zn values of pyritic schists vary markedly in the Kanmantoo Group, they are anomalous in these metals relative to other rock types. Schiller (2000) analyzed 27 pyritic schists originally collected by George (1967) and found median values of 111 ppm Cu, 3,970 ppm Zn, and 3,520 ppm Pb. These metals are present in chalcopyrite, sphalerite, and galena along with a variety of other less abundant sulfides and sulfur sulfosalts (Nenke, 1972), but they also occur as trace elements in the lattice of pyrite and pyrrhotite. Although evaporites are not known in the Kanmantoo Group, a source of Cl to carry Cu, Pb, and Zn as chloride complexes could have been derived from the precursors to scapolite, which is common in the Talisker and Backstairs Passage Formation, or the precursor to Cl-bearing apatite, which was reported by Hammerli et al. (2016) as a detrital mineral in the Kanmantoo Group. Although it may be possible to generate sufficient Cu to form a Cu-Au deposit the size of Kanmantoo by dehydration and desulfidation during prograde metamorphism associated with the conversion of pyrite to pyrrhotite as proposed by Finch and Tomkins (2017), there is no textural evidence to show that pyrrhotite formed by the breakdown of pyrite in the Kanmantoo Group. Moreover, Hammerli et al. (2015, 2016) showed no evidence to suggest that Cu was released from pelitic and psammitic rocks during prograde metamorphism from greenschist to upper amphibolite facies.

6.2. Preferred model of formation for the Kanmantoo Cu-Au deposit

The mineralogical and geochemical data obtained in the present study support the metamorphic syn-sedimentary model of Seccombe et al. (1985), Toteff (1999), and Marshall and Spry (2000). Sediments in the Kanmantoo Group were deposited rapidly in the Kanmantoo Trough over a short 8 ± 5 Ma period (522 ± 2 to 514 ± 3 Ma, Jenkins et al., 2002; Foden et al., 2006). Extension in the trough, accompanied by a thinning lithosphere, generated an anomalously steep geothermal gradient, forming convective hydrothermal fluids (300–350 °C, reducing brine) to depths of at least 4 km (Seccombe et al., 1985). These circulating fluids leached Cu, Pb, Zn, Ag, Au, Bi, Cl, and S from the basal sediments in the Kanmantoo Group (i.e., Normanville Group, Carrickalinga Head Formation, Backstairs Passage Formation, and particularly the Talisker Formation), and then rose along syn-sedimentary growth faults, which include the reactivated prominent Nairne and Bremer faults, during basin development. The fluid likely migrated via a process of seismic pumping as advocated by Seccombe et al. (1985) and Toteff (1999). Active syn-sedimentary faulting in the Kanmantoo Trough, as evidenced by abrupt variations in the thickness of sediments in the Kanmantoo Group, likely provided the conduits necessary for ore emplacement (Thomson, 1975; Parker, 1986; Toteff, 1990, 1999). Fluids focused along these growth faults and then permeated the Tapanappa Formation to produce a regional stratabound hydrothermal zone consisting of quartz, illite, chloride, iron oxides and hydroxides, and minor plagioclase (Schiller, 2000). At the Kanmantoo deposit, relatively sulfur-poor fluids, which produced the dominant metallic mineralogy of chalcopyrite-pyrrhotite-magnetite, formed several orebodies including Kavanagh and Emily Star in a subsurface, stockwork feeder zone. Locally, more S-rich fluids produced minor pyrite in the deposit (Fig. 17). Overall, the hydrothermal fluids were enriched in Fe, Mg, and Cu (± S), and depleted in Na and Ca (e.g., Schiller, 2000). Higher up in the stockwork zone, the hydrothermal system continued to
vent Cu-bearing fluids that produced the stratiform Nugent deposit. Exhalations of Fe-Mn ± Zn-rich hydrothermal fluid at the seawater-rock interface, produced an extensive (> 1 km long) horizon of chemical sediments along strike from the Nugent orebody. These fluids carried the chemical components to form the precursors to banded iron formations (i.e., precursor to quartz garnetite and meta-BIF) and hydrothermal sediments (i.e., precursor to quartz garnetite and meta-BIF) are also shown.

Fig. 17. Schematic diagram of the proposed genetic model of the Kanmantoo deposit showing the spatial relationship of Kavanagh and Emily Star to Nugent in the subsurface feeder zone, and the spatial relationship of the Kanmantoo Cu-Au deposit to the Strathalbyn and Wheal Ellen Pb-Zn-Ag-(Cu-Au) deposits. The pre-metamorphic hydrothermal alteration zone (now manifested as GABS(S)) and hydrothermal sediments (i.e., precursor to quartz garnetite and meta-BIF) are also shown.

It is proposed here that the Cu orebodies at Kanmantoo are syn-sedimentary and formed in an exhalative system in a sub-seafloor environment. Deposits that formed within the hydrothermal vent associated with hydrothermally altered rocks (i.e., GABS, GABSS, BGCS) include the Kavanagh and Emily Star orebodies. Part of the Nugent orebody where it is intersected by drill hole KTDD125 is also associated with these altered rocks. However, where the Nugent orebody is intersected by drill hole KTDD141, the spatially associated rocks consist mostly of QMS intercalated with minor quartz garnetite.

The only trace element study conducted on silicates in the Tapanappa Formation was that of Hammerli et al. (2016) on garnet, biotite, feldspar, muscovite, and titanite in unmineralized metapelitic and metapsammites rocks unrelated to sulfide mineralization who determined their concentration in rocks at different metamorphic grades (greenschist to upper amphibolite facies). Hammerli et al. (2016) determined the composition of 30 trace elements (Li, V, As, Cu, Pb, Zn, Rb, Sr, Y, Zr, Nb, Mo, Nb, Sb, Cs, Ba, Hf, Ta, Th, and U, as well as 11 REEs). For the base metals (Cu, Pb, Zn), Pb in biotite in the altered rocks at Kanmantoo (9.3–110 ppm) is elevated relative to biotite (up to 11.4 ppm Pb) analyzed by Hammerli et al. (2016) in schists that were subjected to peak metamorphic temperatures of approximately 475°C to 600°C (i.e. conditions similar to those at the Kanmantoo deposit), as it was Zn in 49% of biotite analyses (Table 3). Values of Mn in biotite associated with these rocks is approximately the same as those reported by Hammerli et al. (2016), whereas biotite in quartz garnetite rocks associated with the Nugent deposit contains up to 3,206 ppm Mn. The concentration of Cu in biotite in unmineralized schists and altered rocks associated with the Kanmantoo deposit are indistinguishable from those in unmineralized schists analyzed by Hammerli et al. (2016). In contrast, approximately two thirds of Cu (64%), Pb (69%), and Zn (72%) values for muscovite analyzed here are greater than the values analyzed by Hammerli et al. (2016) in unmineralized rocks. The Cu concentration of garnet in the unmineralized rocks of Hammerli et al. (2016) was below detection limits, but there seems to be no distinction between the Pb and Zn values of garnet in altered rocks in Kanmantoo versus the unmineralized rocks analyzed by Hammerli et al. (2016). In contrast, the Mn and Zn values of garnet in quartz garnetite in the Nugent Deposit meralized by hydrothermal fluids.
orebody are enriched (up to 19.5 wt% MnO, 170 ppm Zn) relative to garnet analyzed by Hammerli et al. (2016). Moreover, the high Mn content of garnet in quartz garnetite resembles that of garnet spatially associated with Pb-Zn-Ag-(Cu-Au) deposits in the Kavanagh Group (Tott et al., 2016), and massive sulfide deposits in other metamorphic terranes where it has been shown to be an important tool in exploration (e.g., Spry et al., 2000; Avrill; 2001; Heimann et al., 2009).

The Zn content of staurolite in metamorphosed massive sulfide deposits is commonly enriched (e.g., up to 9 wt% ZnO, Bleikvassli Zn-Pb-(Cu) deposit, Norway, Rosenberg et al., 1998; up to 6.9 wt% ZnO, LaRonde Penna Zn-Au-Ag-Cu-Pb deposit, Dubé et al., 2007) and has been shown to be a good indication of proximity to sulfides (e.g., Spry, 2000; Spry et al., 2000). Toteff (1999) noted that staurolite in “normal” country rock in the Kavanagh Group was depleted in Zn (∼0.1 wt% ZnO) relative to staurolite in GABB (up to 1.32 wt% ZnO; Schiller, 2000), the latter of which is consistent with up to 1.60 wt% ZnO obtained in the same rock type herein (with a single outlier of 3.20 wt% ZnO). For comparison, up to 5.6 wt% ZnO occurs in staurolite from the Angas Pb-Zn-Ag deposit (Tott et al., 2016).

Chondrite-normalized rare earth element plots for garnet show depleted LREEs and elevated heavy REEs (See Fig. 9A-C). This pattern likely reflects the uptake of HREEs as a substitute for Al³⁺, which depends on the metamorphic fluid temperature, fO₂, detrital input, REE content of the metamorphic fluid, metamorphic fluid-rock ratio, and fractionation between the premetamorphic hydrothermal fluid and precursor minerals (Bau, 1991; Otamendi et al., 2002; Heimann et al., 2009; Moore et al., 2013; White, 2013). The presence of positive Eu anomalies in garnet has been used in the past as an indicator of proximity to sulfide mineralization in metamorphic terranes (Spry et al., 2000, 2015; Rozendaal and Stalder, 2001; Heimann et al., 2009). Positive Eu anomalies reflect reduced, higher temperatures (>250°C) of the premetamorphic hydrothermal fluids in equilibrium with precursor materials that may have formed proximal to ore emplacement whereas negative Eu anomalies reflect an increased detrital component that is carried in an oxidized, lower temperature (<250°C) hydrothermal fluid. The negative anomalies in garnet may also be associated with sulfide mineralization more distal to a vent system or proximal to a vent where the hydrothermal system is small. At the giant Broken Hill Pb-Zn-Ag deposit, where garnet-rich rocks are spatially associated with massive sulfide orebodies (up to 130 m wide), garnet in garnetite exhibits positive Eu anomalies proximal to the vent system whereas negative Eu anomalies are found in garnet in stratigraphically equivalent rocks distal to the Broken Hill vent system, in some quartz garnetite within ore, or proximal to very small Broken Hill-type deposits where the ratio of hydrothermal to detrital components is small (Heimann et al., 2009). Similarly, positive anomalies were reported by Spry et al. (2015) for garnet in the small metamorphosed massive Dammuergert Zn-Pb-Ag deposit in the Stollberg ore field, Sweden. This positive anomaly for garnet in altered rocks occurs over a distance of approximately 100 m on either side of the deposit. Further away from the sulfide mineralization the Eu anomaly in garnet is negative. At Kavanagh, garnet in GABS, GABSS, and BGCS generally shows a negative Eu anomaly, although ~25% show weakly positive anomalies. Assuming the ore fluid was hot (>250°C) and reducing (Seccombe et al., 1985) at Kavanagh, the most likely cause for the negative Eu anomalies in garnet is a high ratio of detrital to hydrothermal components in the precursors to GABS, GABSS, and BGCS.

Approximately 95% of chlorine at Kavanango (Fig. 12A) has an intermediate composition (X_{Fe} = ~0.5–0.7, so-called Fe-poor chlorine of Schiller, 2000) and is associated with Cu-Au ore, whereas minor Bi mineralization is linked to the Fe-rich variety of chlorine (X_{Fe} = ~ > 0.7, Arbon, 2011), where it replaced the rims of garnet and staurolite. Magnesium-rich chlorine from Nugent in drill hole KD1D141 contains considerably higher Mn (>1,780 ppm) and Zn (400–1,600 ppm Zn, Fig. 12C) than the more Fe-rich chlorine from diamond drill hole KD1D125 (Nugent) and the Kavanagh orebody (153–628 ppm Mn and 106–202 ppm Zn).

Magnetite has been extensively used as an exploration guide for metamorphosed Zn-Pb-Ag deposits, iron oxide-copper–gold deposits, volcanogenic massive sulfide deposits, porphyry deposits, Fe-Cu skarn deposits, Ni-Cu-PGE deposits, Kiruna-type deposits, and banded iron formations (Dupuis and Beaudoin, 2011; Nadoll et al., 2012, 2014). The presence of magnetite alone is indicative of proximity to Cu mineralization at Kavanango as it can occur in high-grade ore in the assemblage chlorite-magnetite, or it occurs in BiF along strike from sulfide mineralization. The trace element compositions obtained by Tedesco (2009) from the Kavanagh orebody (drill holes KD1D053 and KD1D117) are similar to those obtained here from ten samples from Kavanagh (drill hole KD1D054). However, the Ni contents of magnetite

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**Table 3**

Concentrations of Cu, Pb, and Zn in garnet, biotite, and muscovite from QMS, GABS, GABSS, BGCS, and quartz garnetite from the Kanmantoo deposit compared to those from unmineralized altered and unaltered rocks from Hamnerli et al. (2016).

<table>
<thead>
<tr>
<th>Garnet</th>
<th>Cu (ppm)</th>
<th>Pb (ppm)</th>
<th>Zn (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Unmineralized</td>
<td>6</td>
<td>n.d.</td>
<td>0.02–0.17</td>
</tr>
<tr>
<td>This study</td>
<td>375</td>
<td>0.03–7.65</td>
<td>0.02–3.75</td>
</tr>
<tr>
<td>2</td>
<td>–</td>
<td>48%</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>–</td>
<td>29%: 17% Kavanagh, 9.5% Nugent, 2.5% Emily Star</td>
<td>49%: 20.5% Kavanagh, 23% Nugent; 5.5% Emily Star</td>
</tr>
<tr>
<td>4</td>
<td>–</td>
<td>23%: 13% Kavanagh, 10% Nugent (KTDD125 only)</td>
<td>41%: – 13.6% Kavanagh, – 13.6% Emily Star, – 13.6% Nugent (KTDD125 only)</td>
</tr>
</tbody>
</table>

**Biotite**

1: Unmineralized 15 1.54–374 0.93–11.4 23.6–214
This study 224 0.04–24,330 9.27–168 21.46–1952
2 43% | 50% | | |
3 | 7% Emily star only | 99.6%: All samples | 49.1%: All samples |
4 | 50%: All samples | – | 0.9%: Kavanagh (KTDD054) only |

**Muscovite**

1: Unmineralized 8 < 2.85 3.88–23.9 3.82–76.6
This study 39 0.25–173 0.01–272 0.51–108
2 36% | | 20% |
3 | 64%: All samples | 69%: 46% Emily star, 18% Nugent (KTDD141 only), 5% Kavanagh | 72%: 44% Emily star, 18% Nugent (KTDD141 only), 10% Kavanagh |
4 | – | 15%: Kavanagh only | 8%: Kavanagh only |

1. Concentrations in samples from unmineralized altered and unaltered country rock from Hamnerli et al. (2016), peak metamorphic T of rocks from ~475°C to 600°C; 2. Percentage of samples from this study that fall within the same range as those in 1; 3. Percentage of samples from this study that are enriched relative to those in 1; 4. Percentage of samples from this study that are depleted relative to those in 1. Proportion of samples from each orebody that make up the percentages in 3 and 4 are shown. No data: n.d.
7. Conclusions

The Kanmantoo Cu-Au deposit is the likely product of a syn-sedimentary sub-seafloor hydrothermal exhalative system in which the Kavanagh and Emily Star orebodies formed in a stockwork feeder zone associated with hydrothermally altered rocks (i.e., the precursor to GABS, GABSS, BGCS), which was subsequently metamorphosed. The Nugent orebody formed proximal to a hydrothermal vent on the seafloor in an environment similar to that responsible for the formation of syn-sedimentary Pb-Zn-(Cu-Au) deposits in the Tapanappa Formation of the Kanmantoo group as evidenced by stratiform mineralization, spatially associated meta-exhalatives, and similar enrichments of Mn and Zn in silicates from QMS and quartz garnetite in the Nugent orebody and QMS in the Pb-Zn-Ag-(Cu-Au) deposits. There is a genetic link between the two deposit types that is supported by the stratigraphic equivalence of the Kanmantoo deposit and stratiform Pb-Zn-Ag-(Cu-Au) deposits. The presence of a chalcocite-pyrite-magnetite zone at the Wheal Ellen Pb-Zn-Ag-(Cu-Au) deposit further supports this genetic link as it shows that the two end-member types of deposits (Cu-Au and Pb-Zn-Ag-(Cu-Au)) may form from similar hydrothermal fluids. The sub-parallel nature of the Kavanagh (± Emily Star) orebody to syn-deformational features (i.e., S2, east-dipping extensional faults, north-northeast trending shear zones, fractures, and faults) is attributed to its discordancy to bedding in a premetamorphic hydrothermal vent with subsequent in situ remobilization of the Cu ore during the Delamerian Orogeny.

Exploration efforts on the regional scale in the Kanmantoo Group should focus on garnet-andalusite-biotite ± chlorite ± staurolite ± magnetite and biotite-garnet-chlorite rocks in the Tapanappa Formation of the Kanmantoo Group since they are known to be spatially associated with base metal mineralization and represent metamorphosed stratabound hydrothermal alteration zones. In the search for Cu-Au deposits on the local scale, garnet, biotite, staurolite, and muscovite should at least be analyzed for Pb, Zn, Cu, and Mn, since biotite is elevated in Pb and generally in Zn, whereas muscovite is likely to be elevated in Pb, Zn, and Cu. Staurolite is Zn-rich and garnet is Mn-rich in quartz garnetite compared to unmineralized country rocks and constitute excellent exploration guides, as does the presence of magnetite (regardless of its trace element composition). Although positive Eu anomalies have been used in the past to indicate proximity to hydrothermal ore deposits in metamorphic terranes elsewhere (e.g., Broken Hill Pb-Zn-Ag, Heimann et al., 2009; Dammberget Pb-Zn-Ag, Spy et al., 2015), the paucity of positive Eu anomalies in garnet from Kanmantoo suggests that this feature is not a good exploration guide to Cu-Au deposits in the Kanmantoo Group. From a practical standpoint, given the resistate nature of staurolite, garnet, and magnetite, these minerals can be sampled from stream sediments, soil, or outcrop, whereas biotite, chlorite, and muscovite should be extracted from altered metamorphic rocks from outcrop or subcrop.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.oregeorev.2018.02.017.

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