



Coeur d'Alene-type Ag-Zn-Pb vein mineralization in the Canadian Appalachians: geological, mineral chemistry, fluid inclusion and isotopic studies of the Sail Pond occurrence, Newfoundland, Canada

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Abstract

Structurally controlled Ag-Zn-Pb deposits are a distinct deposit class that has been suggested as a sub-class of orogenic gold deposits. The Sail Pond occurrence, Newfoundland, represents the first documented occurrence of this type in the Canadian Appalachians. We present the results of detailed geological, mineralogical and geochemical studies from Sail Pond, including mineral chemistry, fluid inclusion and sulphur, lead, carbon and oxygen isotope studies. Mineralization is in structurally controlled quartz- sphalerite-galena-tetrahedrite-tennantite veins in metamorphosed carbonate rocks, with up to 2% Ag in tetrahedrite and tennantite. Geological, isotopic and fluid inclusion studies show that mineralization occurred at high temperatures (>300 °C) and at mesozonal crustal depths (>6 km). Moderate salinity, aqueous-carbonic hydrothermal fluids were generated during the devolatilization of crustal sequences during prograde metamorphism, which leached metals from the nearby crustal rocks. Mineralization precipitated in open space fractures in brittle carbonate host rocks, where metalliferous fluids encountered a suitable sulfate source. Important similarities between typical orogenic gold deposit models and the mineralization at Sail Pond (and other structurally controlled Ag-Zn-Pb deposits in western North America) include the structural control on mineralization, association with regional scale greenschist facies metamorphism in accretionary orogenies and CO₂-rich mineralizing fluids of probable metamorphic origin. This supports the classification of these deposits as Coeur d'Alene-type deposits, a subset of orogenic gold mineral deposits and highlights the exploration potential for this deposit type in the Appalachians and other metamorphosed sedimentary platforms globally.

Keywords Coeur d'Alene-Type deposits · Appalachian · Stable isotopes · Mineral chemistry · Fluid inclusions

Introduction

Structurally hosted Ag-Pb-Zn veins in clastic metasedimentary terranes are a distinct mineral deposit type (Beaudoin and Sangster 1992), and include deposits in the Coeur d'Alene

district, Idaho (Leach et al. 1998; Fleck et al. 2002), and the Keno Hill (Lynch et al. 1990), Kokanee Ranges (Beaudoin and Sangster 1992), and Purcell Belt (Paiement et al. 2012) districts in western Canada. They are typically hosted in greenschist facies sedimentary rocks, with mineralization associated with fluid migration along large-scale crustal structures (Beaudoin and Sangster 1992; Taylor and Hofstra 2025).

Goldfarb et al. (2005) suggested that Ag-Pb-Zn veins in clastic metasedimentary terranes represents a subset of orogenic gold deposits, and Taylor and Hofstra (2025) proposed that the deposits of the Coeur d'Alene district should be referred to as a distinct subset of orogenic gold deposits, termed Coeur d'Alene-type mineral systems. Although the Ag and base metal content of these deposits are much higher than typical orogenic gold deposits (typically 100s to 1000s g/t Ag), the two deposit types share numerous characteristics, including a strong structural control on mineralization,

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association with greenschist facies metamorphism and presence of aqueous-carbonic mineralizing fluids (Beaudoin and Sangster 1992; Goldfarb et al. 2005; Taylor and Hofstra 2025). The high Ag and base metal content of these deposits may be related to higher fluid salinities and/or background enrichment of these metals in the host rocks (Goldfarb et al. 2005; Taylor and Hofstra 2025).

The Sail Pond occurrence is in western Newfoundland, in the Humber Zone of the Canadian Appalachians (Fig. 1). Recent exploration has identified two zones of Ag-Cu-Zn-Pb-Sb±Au veins in metamorphosed carbonates, with a combined strike length of ~9 km. Recent drilling intersected 83 g/t Ag, 0.3% Cu, 1.48% Zn, 0.53% Pb and 0.13% Sb over 6.86 m (Sterling Metals, 2021) and assay

results from channel samples include 499 g/t Ag, 2.66% Cu, 6.80% Pb, 0.27% Zn and 0.33 g/t Au over 1.0 m (South Zone) and 180 g/t Ag, 0.97% Cu, 0.24% Pb and 0.18% Zn over 2.9 m (North Zone) (Piercey 2020). This represents the only known occurrence of structurally controlled Ag-Pb-Zn veins in clastic metasedimentary terranes in the Canadian Appalachians. The Sail Pond occurrence provides an excellent opportunity to study the Ag-Pb-Zn veins in clastic metasedimentary terranes and investigate the relationship between these deposits, orogenic gold deposits, and regional tectonic settings in the Appalachians. In this paper, we present detailed geological, mineralogical and geochemical data from the Sail Pond occurrence and present a genetic model for Ag-Pb-Zn veins in clastic metasedimentary terranes.

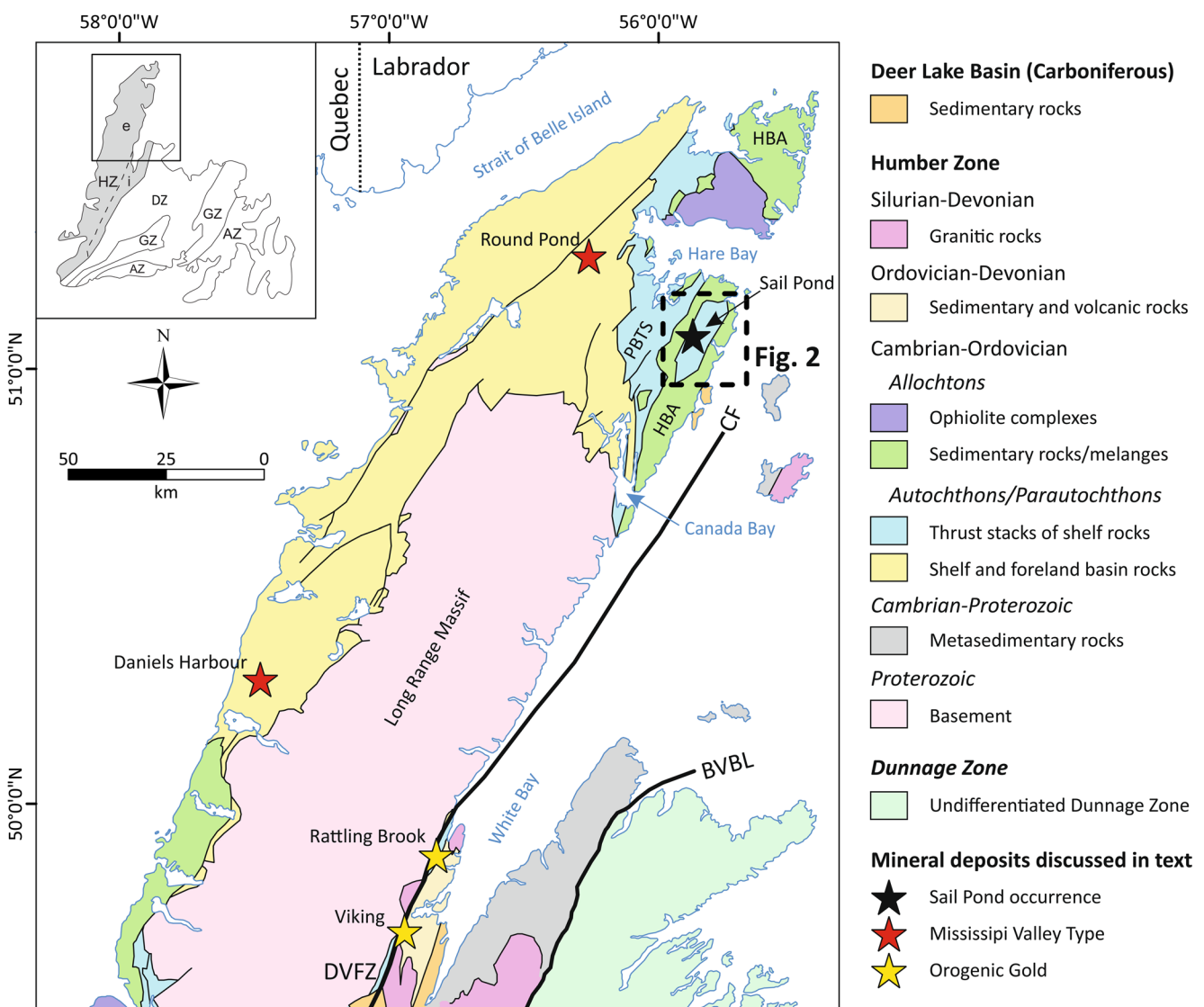


Fig. 1 Simplified geology of the Great Northern Peninsula, western Newfoundland, showing geological terranes and mineral deposits discussed in the text. (modified from Hinchey et al. 2015). AZ=Avalon Zone. BVBL=Baie Verte-Brompton Line. CF=Cabot Fault.

DVFZ=Doucours Valley Fault Zone. DZ=Dunnage Zone. GZ=Gander Zone. HBA=Hare Bay Allochthon. HZ=Humber zone (e=external, I=internal). PBTS=Pistolet Bay Thrust Stack.

Geological setting

The Sail Pond Prospect is in the Humber Zone, the westernmost of five tectonostratigraphic zones in the Canadian Appalachians (Williams 1979). The Humber Zone is subdivided into eastern internal and western external zones based on grade of metamorphism and style of deformation (Fig. 1; Waldron et al. 1998). The external Humber Zone is composed of autochthonous platform rocks, allochthonous deep-water sedimentary rocks, ophiolitic rocks, Grenvillian basement rocks and post-orogenic cover rocks (Fig. 1). These record a complex geological history related to the opening, and subsequent closure, of the Iapetus Ocean, and have been variably affected by multiple deformation events during the Appalachian orogenic cycle (Williams 1979; van Staal and Barr 2012). The internal Humber Zone consists of Precambrian basement rocks overlain by Ediacaran to Ordovician metasedimentary rocks with upper greenschist to amphibolite facies assemblages, reflecting burial during the Taconic and Salinic orogenies (Castonguay et al. 2014; de Wit and Armstrong 2014).

Post-Ordovician cover rocks in the Humber Zone include the Sops Arm Group and the Deer Lake Basin (Fig. 1). The Silurian Sops Arm Group in the White Bay area is bounded by the northeast trending Doucers Valley and Birchy Ridge faults (Sandeman and Dunning 2016), and the Carboniferous Deer Lake Basin formed as a pull apart basin during dextral transtension along the Cabot Fault system (Waldron et al. 1998, 2015). Similar Silurian and Carboniferous sedimentary rocks also crop out ~150–200 km to the northeast of the Deer Lake Basin (Williams and Smith, 1983; Newton 2019). These basins are associated with long-lived northeast trending crustal structures, which were active from at least the Silurian to the Carboniferous (Tuach 1987; Kerr and Knight 2004; Brem 2007).

External Humber Zone

The oldest rocks of the external Humber Zone are Paleoproterozoic, which are interpreted to be part of the Grenville Province (van Staal and Barr 2012). Initial rifting of the Laurentian margin during the opening of the Iapetus Ocean began during the latest Neoproterozoic (~615 Ma) forming the Humber Seaway between Laurentia and the Dashwoods microcontinent (Waldron and van Staal 2001; van Staal and Barr 2012). Early rift-related terrestrial clastic and volcanic rocks are overlain by sedimentary units deposited on a long-lived passive margin to the south of Laurentia (James et al. 1989). Lower to Middle Cambrian clastic rocks of the Labrador Group are overlain by Middle Cambrian to Lower Ordovician carbonate platform rocks of the Port au Port and St. George groups. Deposition of these platform carbonates

was followed by the formation of a marine foreland basin and deposition of shallow- to deep-water carbonates and shales of the Table Head and Goose Tickle groups (James et al. 1989).

Allochthonous units of the external Humber Zone consist of Cambrian to Ordovician sedimentary successions and ophiolite complexes that represent the shelf slope and deep-sea floor equivalents of the autochthonous passive margin successions. The Hare Bay allochthon (HBA; Fig. 1) was thrust over the autochthonous Laurentian margin rocks during the Early Ordovician Taconic orogeny (~488 to 473 Ma; Cawood and Williams 1988; Waldron et al. 1998; van Staal and Barr 2012). This thin-skinned thrusting was restricted to upper crustal units (Cawood and Williams 1988; Waldron et al. 1998) and forming of a series of structural slices in the HBA (Williams and Smyth 1983). During the Late Silurian to Early Devonian Acadian orogeny (421–400 Ma; Cawood 1989; van Staal et al. 2009), these thrust faults were reactivated and deep-seated basement-involved normal faults were inverted, forming major thrust nappes (Waldron et al. 1998; White et al., 2019). This thick-skinned thrusting was responsible for the uplift of the autochthonous platform rocks and Grenvillian basement rocks that core the Long Range Mountains. In the Hare Bay and Canada Bay area, thick-skinned thrusting also incorporated sheets of autochthonous units (Port au Port, St. George and Goose Tickle groups), which were imbricated forming the parautochthonous Pistolet Bay thrust stack (Fig. 1; Knight, 1987).

White Arm window anticline

The White Arm window anticline (WAWA) is a northeast trending package of parautochthonous rocks of the Pistolet Bay thrust stack, which is enclosed by structural slices of the HBA (Fig. 2; Stouge and Godfrey 1982; Cawood 1993). The Brent Island Formation consists of crystalline dolostone intercalated with dolostone (Stouge and Godfrey 1982). The overlying Catoche Formation consists dominantly of limestone, which are extensively dolomitized near the top of the formation (Stouge and Godfrey 1982; Knight and James 1987). The Table Head Group consists of limestone interbedded with black shales, and the Goose Tickle Group includes interbedded siltstones, sandstones and shales (Stouge and Godfrey 1982).

Stouge and Godfrey (1982) suggested that the WAWA represented an erosional window, with parautochthonous units exposed beneath the HBA along a folded Taconic thrust contact. However, Cawood (1993) showed that rather than erosion of a simple, folded thrust contact, the WAWA represents a hanging-wall anticline to a west-directed Devonian-age Acadian thrust. The WAWA is subdivided into

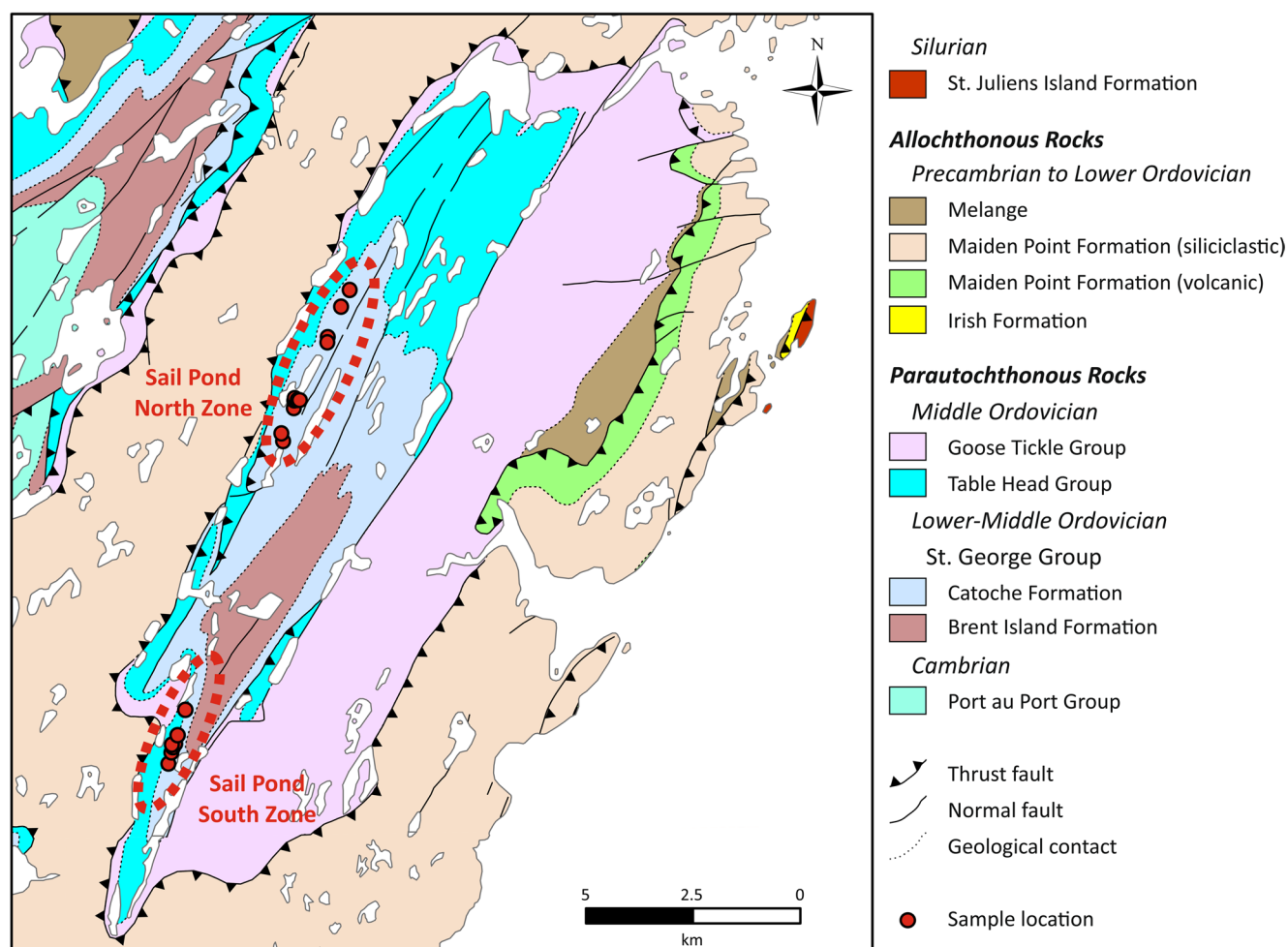


Fig. 2 Simplified geological map of the White Arm window anticline, showing location of North and South zones and sample locations (after Stouge and Godfrey 1982; Knight 1986)

two domains, with predominantly Goose Tickle Group in the east and St. George and Table Head groups in the west (Fig. 2).

The parautochthonous rocks of the WAWA have been metamorphosed to greenschist facies, with limestones and dolostones of the St. George Group commonly metamorphosed to white marbles (Howse 2004), and shales of the Goose Tickle Group locally transformed to grey slate (Stouge and Godfrey 1982). Temperatures of metamorphism in the WAWA have been estimated using the color alteration index (CAI) of conodonts (Stouge 1986; Nowlan and Barnes 1987). The CAI ranges from 5 to 6, which corresponds to temperatures of $>300\text{ }^{\circ}\text{C}$ (Stouge 1986).

Geology of the Sail Pond occurrence

The Sail Pond occurrence consists of two mineralized zones (South Zone and North Zone) on the western margin of the WAWA (Fig. 2). The mineralized zones are stratabound,

northeast trending and dip to the east-southeast, and are parallel to a thrust fault that marks the western boundary of the WAWA (Fig. 2). Trenching has exposed mineralization in both zones - the South Zone is at least 2 km in strike, whereas the North Zone is at least 7 km in strike. The mineralization width is variable, but is up to 200 m wide. Channel sampling identified high grade Ag-Cu-Zn-Pb-Sb $\pm$ Au mineralization, with grades of 499 g/t Ag, 2.66% Cu, 6.80% Pb, 0.27% Zn and 0.33 g/t Au over 1.0 m (South Zone) and 180 g/t Ag, 0.97% Cu, 0.24% Pb and 0.18% Zn over 2.9 m (North Zone) (Piercey 2020). Recent geophysical surveys have identified two other mineralized zones at depth (Heimdall and Heimdall North), and drill testing of these targets have intercepted similar styles of mineralization at depths $>200\text{ m}$ with grades up to 83 g/t Ag, 0.3% Cu, 1.48% Zn, 0.53% Pb and 0.13% Sb over 6.86 m (Sterling Metals Press Release, October 2021).

The geology and style of mineralization in the South and North zones is similar, and they likely represent the same mineralization system. They contain sequences or blocks of

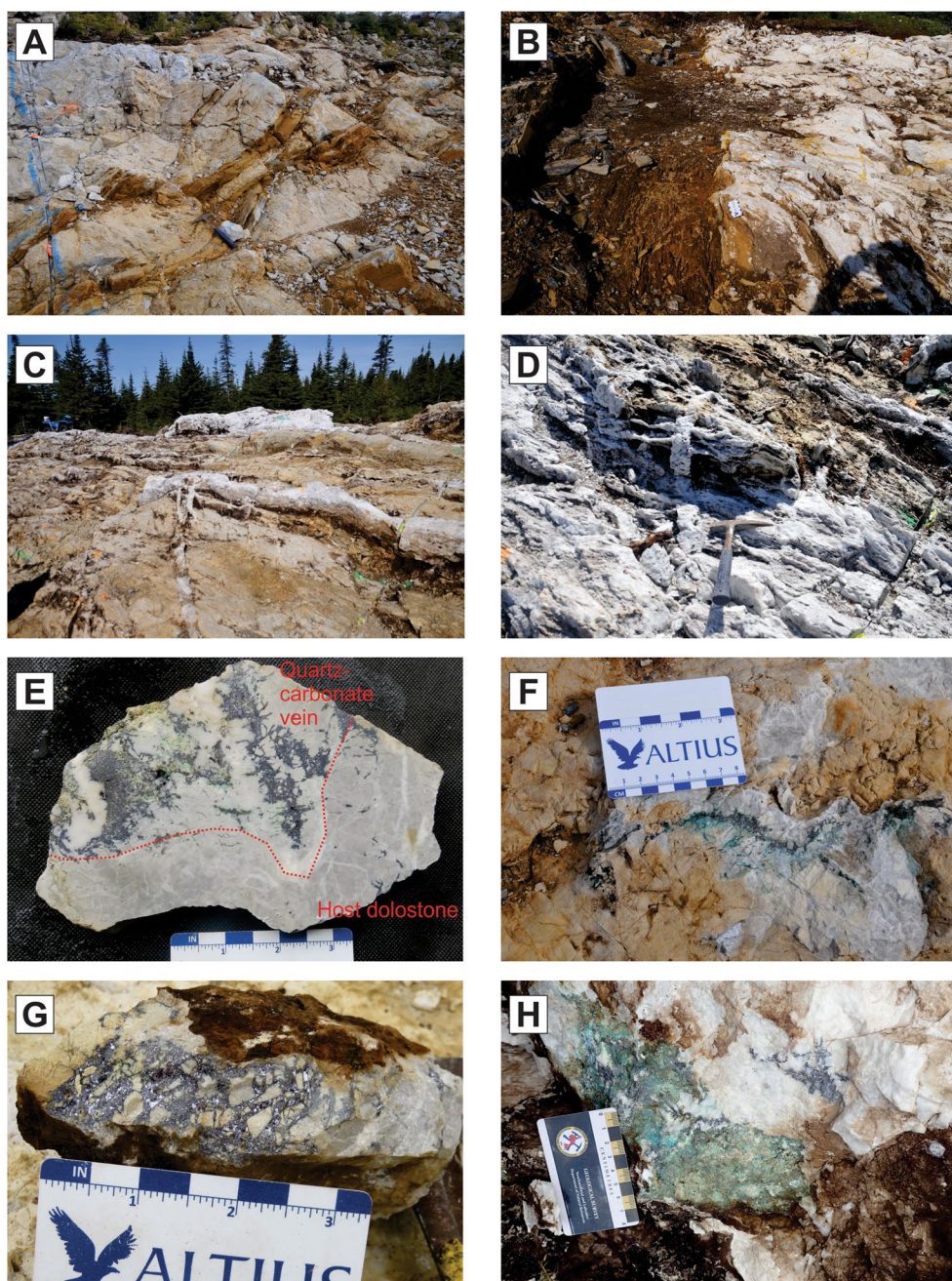
massive cream-colored dolostone of the Catoche Formation, with variable amounts of intercalated marble, limestone and shale (Fig. 3A). The dolostone is overlain by strongly deformed shales of the Table Head or Goose Tickle groups (Fig. 3B). This contact between these units is interpreted as a shear/thrust plane, which can be traced along the western margin of the WAWA (Pilote 2017). The dolostones, which are the host to all known mineralization, are variably faulted, folded and/or boudinaged. Numerous quartz carbonate veins crosscut the dolostone (Fig. 3C) but are not observed in any other lithologies, which indicates that the dolostones had a strong rheological control on mineralization (Pilote 2017).

Quartz veins occur as tension or shear veins in dolostones, and range in thickness from <10 cm to >2 m. Locally these veins constitute up to 30–40% of the volume of exposed rock (Fig. 3D).

Structural geology

Structural studies at the Sail Pond property have identified a complex history of deformation and associated vein emplacement, which are likely related to thrusting during the Taconic and Acadian orogenies (Pilote 2017; Lebrun

Fig. 3 (A) Deformed and boudinaged altered dolostone with siltstone and mudstone layers. (B) Contact between altered dolostone and sheared shale/phyllite. (C) Conjugate quartz veins, which are mutually cross cutting and generally trending 060° and 120°. (D) Cross cutting quartz veins constituting ~30–40% of volume of exposed rock. (E) Tetrahedrite-tennantite-galena-sphalerite mineralization at margins of quartz-carbonate veins. (F) Pod of high-grade mineralization (tetrahedrite-tennantite-galena) in a shear vein, variably weathered to malachite. (G) Tetrahedrite-tennantite-galena-sphalerite mineralization in open space filling (clot) in quartz veins. (H) Supergene alteration of mineralization with malachite, azurite and chalcocite-covellite-bornite after tetrahedrite-tennantite



2021; Hrabi 2022, 2023). These studies highlight the strong structural control on mineralization, with the distribution of mineralization controlled by rheological contrasts between more ductile marble and shale and more competent dolostone (Pilote 2017; Hrabi 2023).

The first phase of deformation (D1) is a steeply east-southeast dipping foliation (S1) and is associated with development of open to isoclinal, southwest plunging folds (F1) in the more competent lithologies (sandstone and dolostone) and development of boudinages in less competent lithologies (Pilote 2017; Lebrun 2021). This was followed D2, which developed minor crenulations on D1 foliations (Pilote 2017). The third stage of deformation (D3) overprints earlier D1-D2 deformation and was interpreted to be due to east-northeast directed shortening and involved varying responses and orientations related to local rheology of the rocks (Lebrun 2021; Hrabi 2022, 2023).

Three main quartz vein generations have been recorded at Sail Pond, with mineralized veins confined to dolomitic units proximal to sheared/boudined contacts with marble and/or shale that have experienced brittle fracturing. The earliest generation of barren quartz veins (V1) form north-northeast trending veins parallel to regional S1 foliations and are interpreted to have formed during northwest shortening (Lebrun 2021). V1 veins are folded, brecciated and crosscut by two generations of steeply dipping mineralized quartz veins (V2A and V2B). The mineralized veins occur as shear, extension, breccia, and tension gash, locally occurring as conjugate sets with Riedel-type geometries, both of which contain mineralization in varying amounts. V2A veins are typically west-northwest trending, and are variably mineralized, while V2B veins are typically east-northeast trending, host the bulk of the mineralization at Sail Pond and are associated with mineralized breccia zones (Fig. 4B, Lebrun 2021).

The timing of deformation and mineralization at Sail Pond is unknown. Knight (1987) and Cawood (1993) showed that deformation in the parautochthonous sequences, such as the WAWA, was related to the Acadian orogeny, with the WAWA representing a hanging wall anticline to an Acadian thrust sheet (Cawood 1993). Therefore, it is likely that all deformation at Sail Pond postdates Taconic thrusting and is related to the emplacement of the WAWA during westward directed Acadian thrusting and later normal and strike slip deformation during the latter stages of the Acadian or Neoacadian orogenies.

Petrography and paragenesis of mineralization

Ag-Cu-Zn-Pb-Sb mineralization at the Sail Pond occurrence is generally within or spatially associated with V2A and V2B veins, occurring as open-space infilling (clots),

disseminations, and vein-parallel massive bands or veinlets (Fig. 3E-G). Mineralization also occurs within the matrix of dolostone breccias, possibly as a solution breccia matrix replacement (Fig. 3E). Although all sulphides and sulphosalts are intergrown, which indicate coeval precipitation, spatial relationship between minerals have been used to determine mineral paragenesis for the Sail Pond occurrence (Fig. 6).

The host dolostones consist primarily of granular carbonates (dolomite >> calcite), with local mixed clastic laminae that contain clay minerals (e.g., muscovite, chlorite) and pyrite (Fig. 5A and B). The pyrite in these units has a rounded nature that is variably recrystallized to subhedral shapes, but is interpreted to be pre-mineralization framboidal pyrite of biogenic origin as it is exclusively dispersed in the clastic laminae (Fig. 5B; e.g., Wilkin and Barnes 1997). Alteration is not extensively developed, generally being restricted to 10s of centimeters of vein margins, with carbonates recrystallized to coarse grained dolomite and calcite, minor muscovite/illite alteration, and pyrite in the clay-rich pods that are recrystallized to euhedral pyrite crystals (Fig. 5A and B).

Mineralized veins consist predominantly of granular to coarse quartz with undulatory extinction and minor coarse dolomite (+/-calcite). The primary sulfide-sulfosalt assemblage consists of sphalerite, galena and tetrahedrite-tennantite with minor pyrite, including pre- and syn-mineralization pyrite (Fig. 6). In some cases, pre-mineralization pyrite is found as inclusions within galena and sphalerite, whereas in other cases pyrite forms large grains surrounded by galena or as grains along the edge of sphalerite-galena- tetrahedrite-tennantite (Fig. 5C) or in the host rock but proximal to mineralization (Fig. 5I), suggesting syn-mineralization pyrite that co-precipitated with the sphalerite-galena-tetrahedrite-tennantite assemblage. Sphalerite, galena, and tetrahedrite form complex intergrown grains that are coarse and interpreted to have precipitated together during the main phase of mineralization (Fig. 5D-H). Later generations of galena have been recorded in fractures in sphalerite and pyrite, as well as interstitial to recrystallized pyrite in the host dolostones (Fig. 5C and H), suggesting local remobilization post-mineralization.

In many locations the mineralized veins have surface Cu staining, with abundant malachite, azurite, chalcocite, and other Cu-phases (Fig. 3F-H). This reflects partial degradation of the primary sulfide-sulfosalt assemblage into secondary, supergene minerals (Fig. 6), with tetrahedrite-tennantite the most highly affected phase. The degree of supergene replacement of primary tetrahedrite-tennantite is variable, ranging from incipient replacement along fractures to complete replacement by masses of chalcocite-covellite-bornite

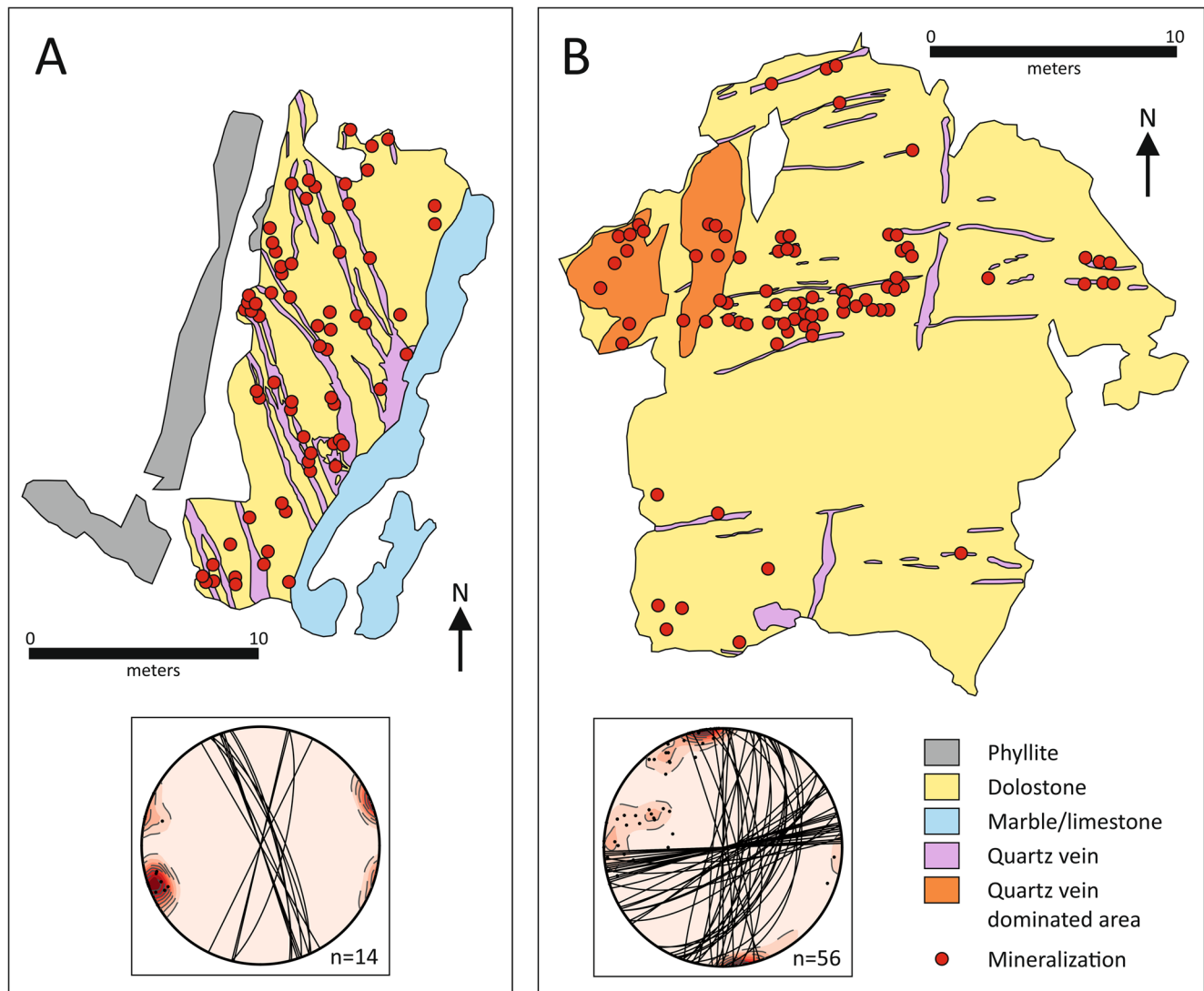


Fig. 4 Trench maps from the Sail Pond occurrence illustrating the structural control on veins and associated mineralization (stereonet projection of measured quartz veins from Pilote 2017). **(A)** Trench map of SRTR-03 from the South Zone. The distribution of veins suggests local rheological control with strain taken up by the shale/phyllite and marble units, whereas the dolostone broke leading to north-

west trending mineralized veins (V2A?). **(B)** Trench map of SRTR-10 from the North Zone with barren, north-trending V1 veins, mineralized west-trending V2A veins and local massive V2B vein domains with mineralized breccias. Dots are pole to plane (heat map representing density of poles) and great circles represent lineation features (e.g., fold axes, stretching lineation, and intersection lineation)

(Fig. 5J). The chalcocite-covellite-bornite assemblage is itself replaced or converted to malachite (Fig. 5K), and needles of Pb-arsenate minerals (mimetite) are also found in this supergene assemblage (Fig. 5L).

Sampling and analytical methods

Nineteen samples were collected from the Sail Pond deposit, 11 samples from the North Zone and 8 samples from the South Zone. Samples were selected to represent the different styles of mineralization (e.g., vein-hosted, breccia-hosted) and abundance of mineralization, as well as samples of

unmineralized dolostone. A full description of the field samples and subsequent analytical work undertaken thereon are presented in Electronic Supplementary Material 1 (ESM1).

Thin sections were prepared and analyzed using standard transmitted and reflected light petrography. A subset of seven samples were further imaged on a JEOL 7110 F field-emission gun-scanning electron microscope (FEG-SEM) equipped with a backscattered electron (BSE) detector and Thermo energy dispersive spectrometry (EDS) system with a silicon drift detector. Operating conditions for backscatter imaging were 15.0 kV accelerating voltage and a 12–13 mm working distance.

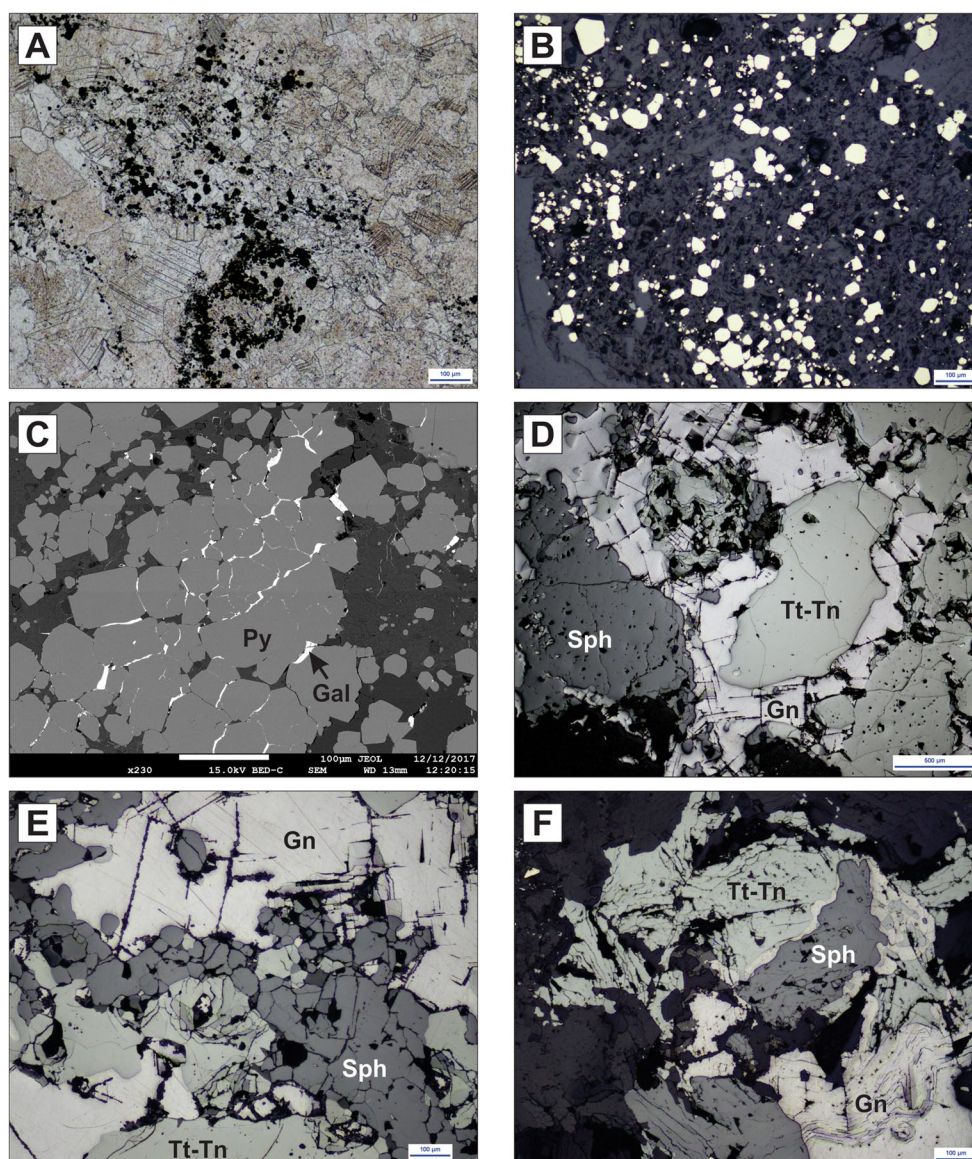


Fig. 5 Photomicrographs (plane polarized light (ppl) and reflected light (rl)) and back scatter electron-scanning electron microscope (BSE-SEM) images. (A) Host dolostone with medium-grained dolomite and pod of illite/muscovite and pyrite (ppl). (B) Euhedral to subhedral pyrite restricted to clastic layer, likely representing recrystallized framboidal pyrite (rl). (C) BSE-SEM image of recrystallized pyrite in clastic layer, with late interstitial galena. (D) Sphalerite and tetrahedrite-tennantite grains with interstitial galena (rl). (E) Galena and tetrahedrite-tennantite grains with interstitial sphalerite (rl). (F) Complex intergrowth of sphalerite, galena and tetrahedrite-tennantite (rl). (G) Sphalerite rimmed by pyrite and tetrahedrite-tennantite (rl).

Based on textural and mineralogical relationships a further sub-set of samples were selected for determination of major and minor elements in sulfides (pyrite-galena-galena-tetrahedrite) and carbonates (calcite-dolomite) by electron probe microanalysis (EPMA). Sulfur isotopes were determined in galena by in situ secondary ion mass spectrometry (SIMS) and in sphalerite by conventional

(H) Fractured sphalerite grain with late stage pyrite and galena filling fractures (BSE-SEM). (I) Pyrite at margin of galena-tetrahedrite-tennantite mineralization. Note bright galena inclusions in pyrite grain, including small micro inclusion in pyrite marked by red arrow (BSE-SEM). (J) Advanced supergene weathering with relict tetrahedrite-tennantite in a broader mass of chalcocite-covellite-bornite (rl). (K) Malachite mass intergrown with quartz and calcite and containing relicts of tetrahedrite-tennantite and chalcocite-covellite-bornite (ppl). (L) Zoned grain of tetrahedrite-tennantite with a rim of chalcocite-covellite-bornite, needles of Pb-arsenate (mimetite), and a final rim of malachite (BSE-SEM)

isotope ratio mass spectrometry (IRMS) with all samples being reported as $\delta^{34}\text{S}$ relative to the Vienna Canon Diablo Troilite (VCDT) international standard. Lead isotopes were determined in galena using in situ SIMS. Carbon and oxygen isotope analyses were undertaken on host rock dolomite and mineralization stage dolomite and calcite and were obtained via microdrill and conventional analysis

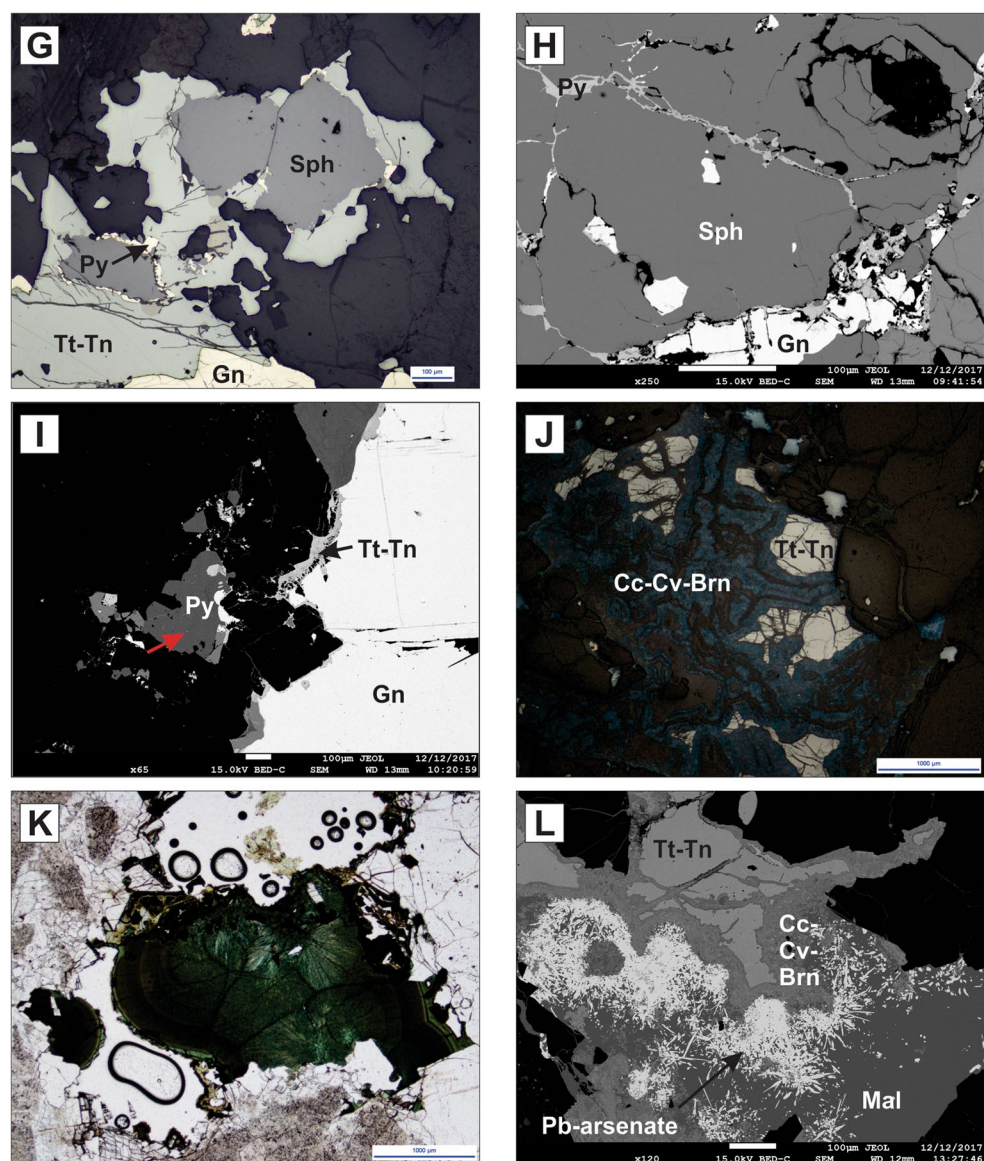


Fig. 5 (continued)

by IRMS; all results are presented as $\delta^{13}\text{C}$ relative to the Vienna Pee Dee belemnite (VPDB) international standard and $\delta^{18}\text{O}$ relative to the Vienna Standard Mean Ocean Water (VSMOW) international standard. Full details of analytical conditions for EMPA and isotopic measurements are provided in ESM2.

Double polished fluid inclusion wafers ($\sim 100\ \mu\text{m}$ thick) were prepared from representative mineralized samples, and petrographic descriptions of fluid inclusions were carried out using the fluid inclusion assemblage (FIA) method (Goldstein 2003). Microthermometric fluid-inclusion analyses at Memorial University of Newfoundland used a Linkam THMSG600 heating–freezing stage, calibrated using synthetic H_2O and $\text{CO}_2\text{-H}_2\text{O}$ fluid inclusion standards. The following phase measurements were recorded: melting

temperature of carbonic phases ($T_m\ \text{CO}_2$), last melting temperature of ice ($T_m\ \text{ice}$), clathrate melting ($T_m\ \text{clath}$), homogenization temperature of carbonic phases ($T_h\ \text{CO}_2$) and total homogenization temperature ($T_h\ \text{tot}$), as well as the mode of homogenization (to liquid, vapor or critical phases). If inclusions decrepitated prior to total homogenization this temperature was also recorded. The volumetric proportion of the carbonic phase was visually estimated after $T_h\ \text{CO}_2$ (vol. CO_2 /vol. total in percentage). For aqueous-carbonic fluid inclusions these microthermometric measurements were used to calculate the volume fraction of CO_2 (X_{CO_2}), fluid salinity (eq. wt% NaCl), bulk fluid density (in g/cm^3) and minimum trapping pressures at homogenization or decrepitation were calculated using the model of Steele-MacInnis (2018) for the $\text{H}_2\text{O}-\text{CO}_2\text{-NaCl}$ system.

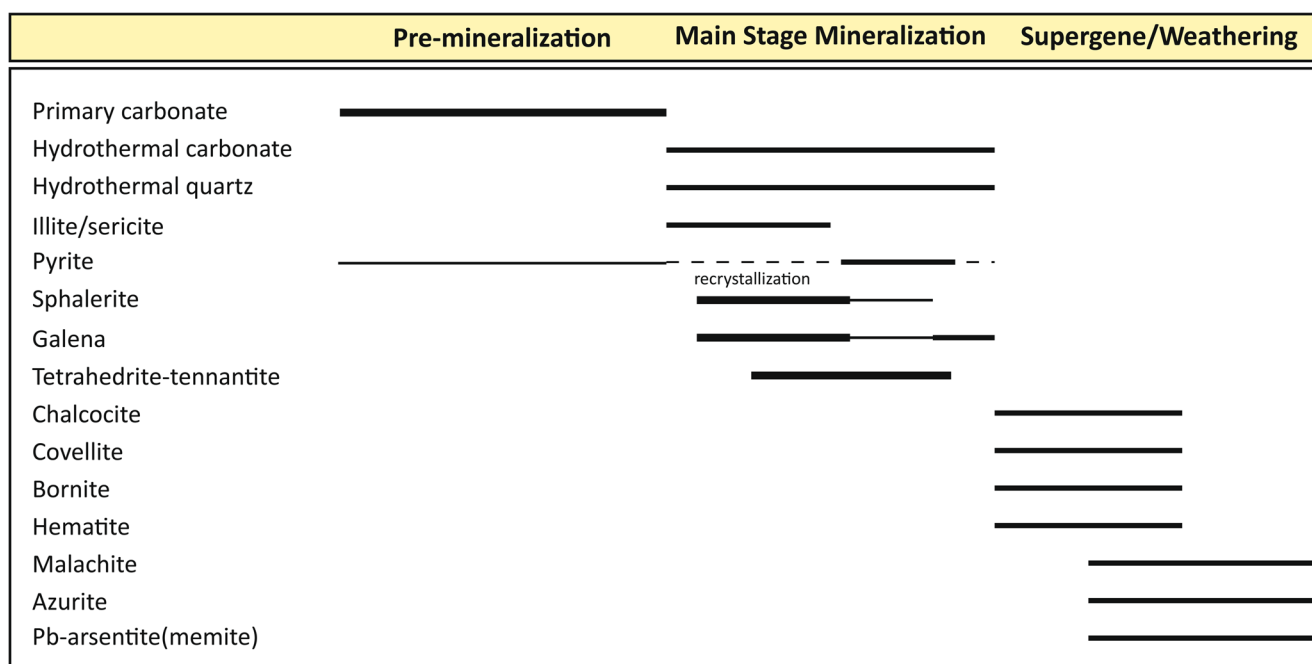


Fig. 6 Paragenetic chart for primary and secondary hydrothermal phases, sulfides-sulfosalts, and supergene phases

Results

Electron microprobe analysis

Recrystallized framboidal pyrite in the host dolomites ($n=7$) and pyrite associated with the main mineralization stage ($n=23$) were analyzed by EPMA. All grains are dominated by Fe and S and are stoichiometric (ESM3, ESM4) with low ($<0.5\%$) concentrations of other elements, less some diagenetic and mineralization-associated pyrite that has elevated Co (up to 0.149%) and As (up to 1.877%) (Fig. 7a; ESM3). Some grains also high Zn, Pb and Cu contents, due to micro- or nano-inclusions of sphalerite, galena and Cu-bearing minerals within the pyrite grains or along their edges (Fig. 5C and I).

Sphalerite grains were analyzed in five samples ($n=20$), with data collected from isolated grains and sphalerite intergrown with galena and tetrahedrite-tennantite. No variation is observed between samples, with all grains having high Zn and S, low Fe contents (<0.358 wt%) and minor Pb (ESM3; ESM4). Sphalerite grains also have elevated Cd contents (0.925%) compared to other mineral phases (Fig. 7b) and are higher than Cd in sphalerite from MVT deposits, but similar to vein-hosted deposits in carbonates (e.g., Schwartz 2000).

Galena grains were analyzed in five samples ($n=20$) and are dominated by Pb and S with minor Ag reported from two samples (0.024 to 0.036 wt%)(ESM3).

Complex sulfides/sulfosalts are dominated by tetrahedrite ($n=10$) and tennantite ($n=28$) and were analyzed in five samples. Analyses were taken on isolated grains, intergrowths with sphalerite and galena, and grains that exhibited incipient supergene mineral formation. Tetrahedrite and tennantite form a solid solution but were separated from one another using the atomic per formula unit (APFU) Sb/(Sb+As) ratio with tetrahedrite having values >0.5 and tennantite values <0.5 (Fig. 7c). They have been classified using the nomenclature of Biagioni et al. (2020) as Zn-bearing (i.e., tetrahedrite (Zn) and tennantite (Zn)). Notably, tennantite exceeds tetrahedrite in abundance, and tetrahedrite is restricted to the South Zone (Fig. 7c). However, it is uncertain if this is a real difference between the two zones or a function of sampling bias. Ag-enrichment occurs both tetrahedrite and tennantite, while overlapping Ag enrichment tends towards higher values in tetrahedrite (Figs 7c-d).

Carbonate minerals are dominated by calcite (CaCO_3) and dolomite ($\text{Ca, Mg}(\text{CO}_3)_2$) and are subdivided into granular carbonates from the host rock and coarse hydrothermal carbonates from proximal to or within veins (ESM5, ESM6). There are no differences in chemical compositions between host rock carbonates and those associated with veins and vein margins, most having near stoichiometric values of mol% MgCO_3 and CaCO_3 . Fe contents are typically >1 wt% with low concentrations of Mn, Ba and Sr (Table 2).

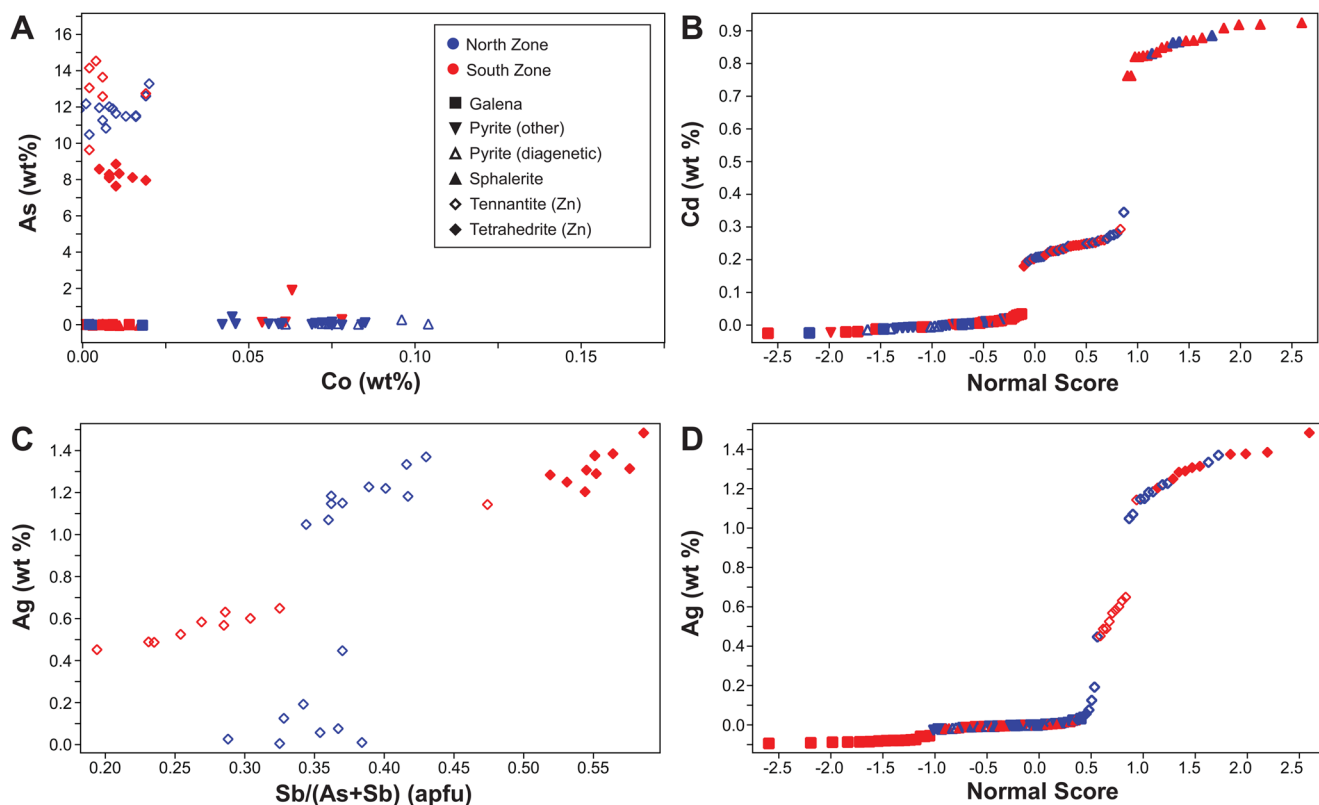


Fig. 7 Microprobe data for sulfide and sulfosalt phases. **(A)** Co-As distribution within the Sail Pond sulfides. Notably the pyrite grains have higher Co compared to other phases but generally low As, whereas the tetrahedrite and tennantite have elevated As. **(B)** Probability plot of Cd distribution with notable Cd enrichment in sphalerite and to a lesser extent tetrahedrite and tennantite **(C)** Ag-Sb/(As + Sb) for the

tetrahedrite-tennantite series minerals and **(D)** probability plot of Ag in all mineral phases showing Ag-enrichment in tennantite and tetrahedrite. The x-axis in **(B)** and **(D)** is the N-score, which are Z-scores from a standard normal distribution where the mean is 0 and the standard deviation is 1. The more positive (and more negative) values indicate more anomalous values relative to the mean

Lead isotope analysis of galena

Nine in situ SIMS lead isotope analyses were obtained from galena. Analyses were conducted on large (>3 mm) grains of galena from the main stage of mineralization (Fig. 6), either occurring as complex intergrowths with sphalerite and tetrahedrite-tennantite (sample 12819) or as isolated grains in quartz-carbonate veins (samples 12821 and

12827). $^{206}\text{Pb}/^{204}\text{Pb}$ values range from 17.705 to 17.949, $^{207}\text{Pb}/^{204}\text{Pb}$ from 15.546 to 15.762 and $^{208}\text{Pb}/^{204}\text{Pb}$ from 37.839 to 38.160 (Table 1). Internal variations in isotopic compositions in individual galena grains are generally small with overlapping values (except for one analysis of $^{208}\text{Pb}/^{204}\text{Pb}$ in sample 12819). The data does show a significant variation in $^{207}\text{Pb}/^{204}\text{Pb}$ between samples with the samples clustering between the young upper crust towards

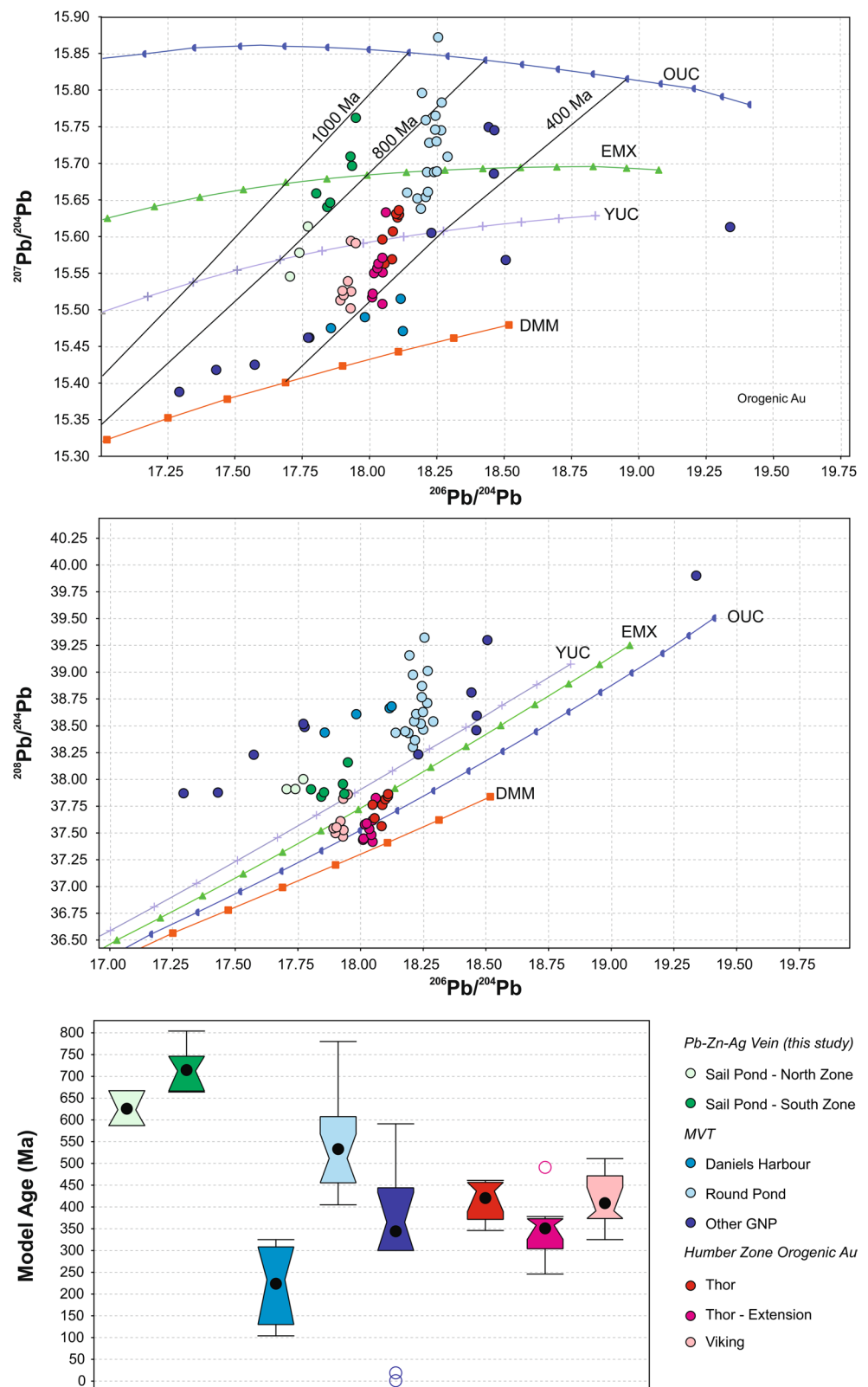
Table 1 Lead isotope composition, internal precision (1 σ (%)) and μ -values (after Stacey and Kramers, 1975) of galena from the Sail Pond occurrence

Sample	Zone	²⁰⁶ Pb/ ²⁰⁴ Pb	1σ (%)	²⁰⁷ Pb/ ²⁰⁴ Pb	1σ (%)	²⁰⁸ Pb/ ²⁰⁴ Pb	1σ (%)	μ	Model Age
South Zone									
12819	South	17.949	0.189	15.762	0.238	38.160	0.284	10.57	801.72
		17.935	0.087	15.696	0.115	37.866	0.141	10.27	700.62
		17.930	0.087	15.709	0.140	37.958	0.145	10.34	729.41
12821	South	17.802	0.167	15.659	0.208	37.908	0.236	10.15	730.43
		17.854	0.067	15.646	0.143	37.880	0.208	10.06	666.41
		17.843	0.085	15.640	0.104	37.839	0.118	10.03	663.44
North Zone									
12827	North	17.771	0.094	15.614	0.101	38.003	0.127	9.93	666.70
		17.705	0.115	15.546	0.131	37.909	0.156	9.64	589.05
		17.740	0.111	15.578	0.120	37.911	0.127	9.79	627.82

the old upper crust curves of Kramers and Tolstikhin (1997) indicative of derivation from sources with variable μ values (9.64–10.57) (Fig. 8; Table 1). The North Zone tends to

have lower $^{207}\text{Pb}/^{204}\text{Pb}$ values at a given $^{206}\text{Pb}/^{204}\text{Pb}$ compared to South Zone samples (Fig. 8; Table 1). When compared to Pb isotopic values of other Humber Zone-hosted

Fig. 8 Pb isotope data for galena from the Sail Pond occurrence and MVT and orogenic Au mineralization from the Humber Zone, Newfoundland Appalachians. (A) Uranogenic plot of $^{207}\text{Pb}/^{204}\text{Pb}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ with isochrons. (B) Thorogenic plot of $^{208}\text{Pb}/^{204}\text{Pb}$ – $^{206}\text{Pb}/^{204}\text{Pb}$. (C) Box and whisker plot of Stacey and Kramers (1975) Pb model ages. Reservoir data in A and B from Kramers and Tolstikhin (1997): OUC – old upper crust, YUC – young upper crust, DMM – depleted mid-ocean ridge basalt (MORB) mantle, and EMX – erosion mix. Data from other areas from Lane (1990), Conliffe et al. (2018), Swinden et al. (1988) and Minnett (2012)



deposits, they are displaced towards lower $^{206}\text{Pb}/^{204}\text{Pb}$ on a distinctive isochron indicative of Pb from older sources. This is supported by Stacey-Kramers model ages (Stacey and Kramers 1975) that range from 589 to 802 Ma and reflect the age of the source of Pb, not the age of mineralization, which overlap the Round Pond occurrences, but are generally older than most other galena hosted within Humber Zone rocks (Fig. 8C). In $^{208}\text{Pb}/^{204}\text{Pb}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ space the Sail Pond samples have relatively restricted $^{208}\text{Pb}/^{204}\text{Pb}$ values, and lie above all upper crustal curves, suggesting derivation of Pb from sources with higher Th/U than normal upper crustal sources (e.g., Zartman and Doe 1981; Kramers and Tolstikhin, 1997). The $^{208}\text{Pb}/^{204}\text{Pb}$ values also overlap with some of the Pb isotope values from Humber Zone hosted mineralization, including the Viking-Thor orogenic Au occurrences (Fig. 8).

Sulfur isotope analysis

Nine in situ SIMS sulfur isotope analyses of galena were obtained from three samples previously analyzed for lead isotopes. The $\delta^{34}\text{S}$ values range from +4.6 to +12.7‰ (Table 2; Fig. 9), with $\delta^{34}\text{S}$ in individual grains varying by +0.9 to +3.4‰.

Bulk sulfur isotope data was also obtained from sphalerite in two samples, representing complex intergrowths of sulphides (sample 12820) and isolated sphalerite grains in quartz-carbonate veins (sample 12828). A total of three

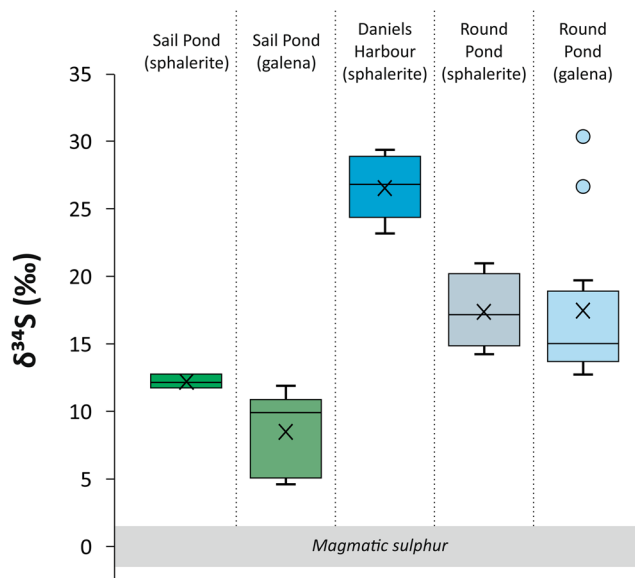


Fig. 9 Box and whisker plot of $\delta^{34}\text{S}$ variation in sphalerite and galena from the Sail Pond occurrence. X marks mean value, line marks median. Also included are $\delta^{34}\text{S}$ data from the Daniels Harbour (Lane 1990; authors unpublished data) and Round Pond (Conliffe et al. 2018) MVT deposits. Colors used same as in Fig. 8

analyses were carried out, with the $\delta^{34}\text{S}$ values ranging from +11.7 to +12.7‰ (Table 2; Fig. 9).

Samples 12819 and 12820 came from the same location and based on the petrographic setting of intergrown sphalerite and galena in both samples it is likely that sphalerite

Table 2 Stable isotope composition of sulfides and carbonates from the Sail Pond occurrence

Sample	Zone	Galena	Sphalerite	Dolomite		Calcite		CO ₂ (350°C)
		δ ³⁴ S	δ ³⁴ S	δ ¹³ C _{VPDB}	δ ¹⁸ O _{VSMOW}	δ ¹³ C _{VPDB}	δ ¹⁸ O _{VSMOW}	δ ¹³ C _{VPDB}
Host carbonates								
12816	South			-0.6	21.3			
12822	South			-0.5	20.8			
Main stage mineralization								
12816	South			-0.5	21.2			1.3
12819	South	11.8				-0.3	20.8	2.1
		10.9						
		8.1						
12820	South		12.7					
			12.1					
12821	South	5.2						
		4.9						
		4.6						
12822	South					-3.7	19.0	-1.3
12823	North			-1.3	21.1			0.5
12827	North	9.9						
		10.0						
		10.8						
12828	North		11.7	1.6	22.3			3.4
				0.7	22.3			2.5
12829	North			0.8	23.0			2.6
12836D	North			0.6	21.4			2.4

and galena precipitated at the same time (Fig. 6), at or close to isotopic equilibrium. Sphalerite generally has higher $\delta^{34}\text{S}$ than galena (average $\sim 2.2\text{‰}$ higher $\delta^{34}\text{S}$), which yields temperatures of isotopic equilibration of 306.8 °C (according to the equations of Li and Liu 2006). Although care must be taken interpreting this temperature, it is consistent with CAI temperatures of $>300\text{ °C}$ from the Sail Pond area (Stouge 1986), and fluid inclusion results from this study.

Carbon and oxygen isotope analysis

Carbon and oxygen isotopic data was collected from seven samples, representing early dolomite ($n=2$) in the host carbonates and coarse calcite ($n=2$) and dolomite ($n=6$) from mineralized veins. The $\delta^{13}\text{C}$ values of unaltered, fine-grained dolomite from the host rock range from -0.5 to -0.6‰ , with $\delta^{18}\text{O}$ of 20.8 to 21.3‰ (Table 2), within the range of previously reported values from unaltered early dolomite in the Catoche Formation elsewhere in the Great Northern Peninsula (Fig. 10; Azomani et al. 2013; Conliffe et al. 2018). Hydrothermal dolomite and calcite from main stage mineralization have $\delta^{13}\text{C}$ values that range from -3.7 to 1.6‰ and $\delta^{18}\text{O}$ values of 19.0 to 22.3‰ . Except for one sample of hydrothermal calcite, all hydrothermal dolomite and calcite samples have $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$ values equal to or above the range of Catoche Formation dolomites elsewhere on the Great Northern Peninsula (Fig. 10; Azomani et al. 2013; Conliffe et al. 2018). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of hydrothermal calcite and dolomite also show a strong positive correlation ($R^2=0.81$; Fig. 10).

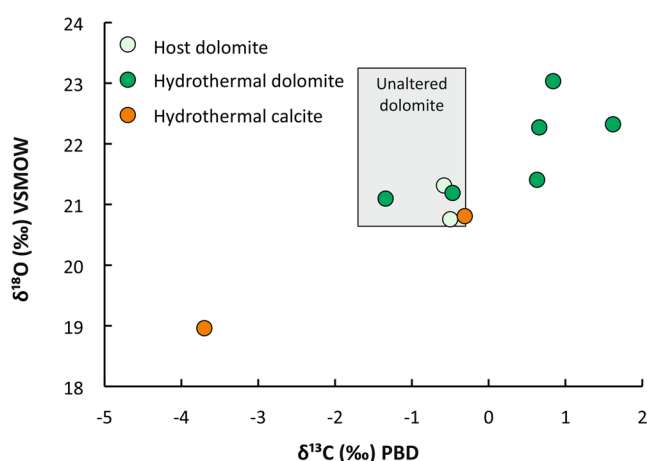


Fig. 10 Bivariate plot of $\delta^{13}\text{C}_{\text{VPDB}}$ and $\delta^{18}\text{O}_{\text{VSMOW}}$ data from host dolomite and hydrothermal dolomite and calcite from the Sail Pond occurrence. Also included is field for least altered dolomite in the Catoche Formation (from Azomani et al. 2013; Conliffe et al. 2018)

Fluid inclusion petrography and microthermometry

Fluid inclusion studies were carried out on hydrothermal quartz and sphalerite from mineralized veins from both the South and North zones. Five samples were selected for detailed microthermometric analysis based on initial petrographic screening, including mineral suitability, spatial relationships to primary ore sphalerite-galena-tetrahedrite-tennantite, and identification of FIA with no evidence of post entrapment modification reflecting they were primary inclusions from mineralizing fluids.

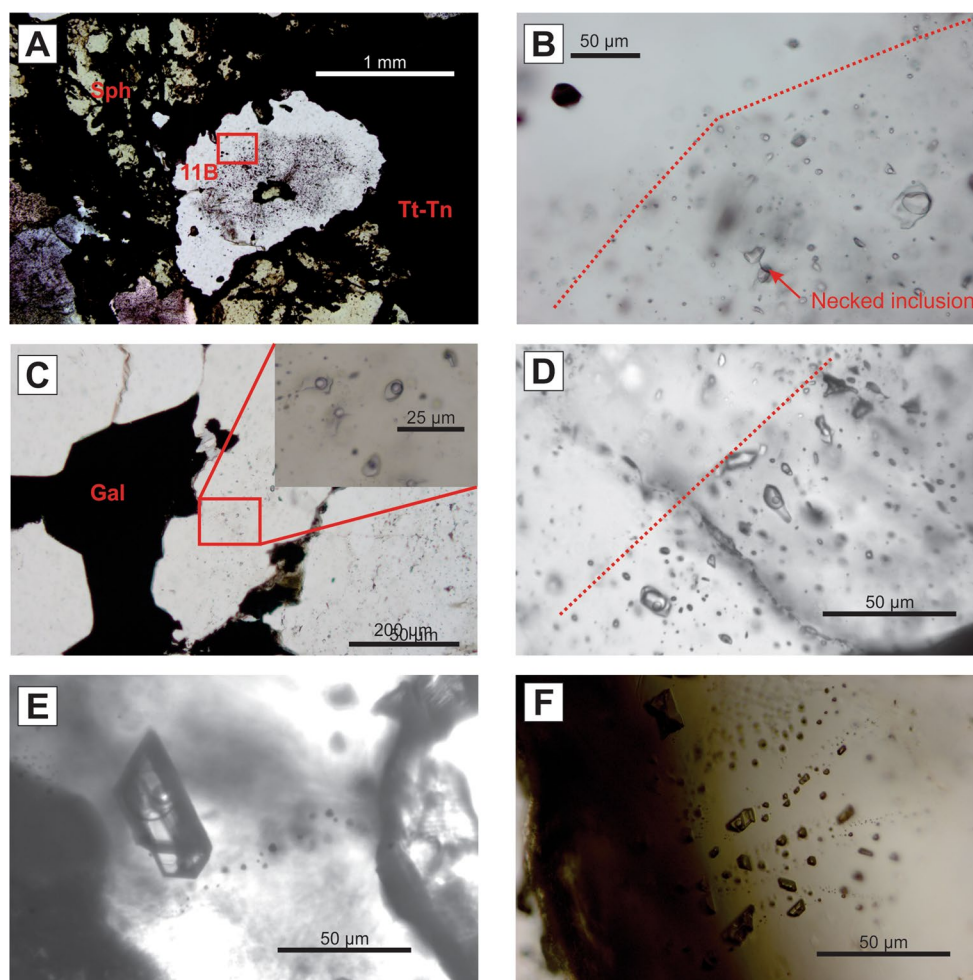
Quartz grains have a high density of fluid inclusions, making identifying primary and secondary FIA difficult. Care was taken to only analyze fluid inclusions in primary FIA in the center of quartz grains with no evidence of secondary fracturing (Fig. 11C), as isolated FIA in clear rims of quartz grains, or rarely as trails of inclusions that are parallel to growth zones in clear quartz crystals (Fig. 11D). In addition, some inclusions show evidence of post entrapment modifications, such as necking down and leaking (Fig. 11B), and microthermometric measurements were not obtained from these inclusions.

Fluid inclusions in sphalerite are rare, and where present occur in clusters close to grain boundaries or as large, isolated inclusions (up to $200\text{ }\mu\text{m}$; Fig. 11E, F). The petrographic and microthermometric characteristics of fluid inclusions hosted in quartz and sphalerite are similar, and they are considered a single fluid inclusion population that represent the fluid present during mineralization.

All inclusions contain aqueous-carbonic fluids, and upon cooling below 0 °C , the carbonic phase in all inclusions separated into a liquid and vapor phase ($\text{L}_{\text{H}_2\text{O}}\text{-L}_{\text{CO}_2}\text{-V}_{\text{CO}_2}$). Inclusions within the same FIA can have highly variable $\text{H}_2\text{O-CO}_2$ ratios, with the volume fraction of the carbonic phase (ϕ_{CO_2}) approximately ranging from 30 to 85%. ϕ_{CO_2} was estimated using a visual estimate of the proportion of phases and some range in ϕ_{CO_2} would be expected based on the calculation of a 3D volume using 2D images (Bakker and Diamond 2006). Some FIA where a significant variation in ϕ_{CO_2} ($>15\%$) show evidence post-entrapment modification of individual fluid inclusions (Fig. 11B), and no microthermometric data was collected from these FIA. In FIA with no evidence for post entrapment modification the coexistence of carbonic and aqueous dominated inclusions suggests at least intermittent trapping of heterogeneous fluids (effervescence).

Microthermometric data was collected from 75 fluid inclusions in 19 FIA (see ESM7). This data was only collected from FIA where the ϕ_{CO_2} varied by $<15\%$, indicating that these FIA represent trapping of a single homogeneous fluid. $T_{\text{m CO}_2}$ was recorded between -58.2 and -56.5 °C , at or below the melting point of pure CO_2 (-56.6 °C),

Fig. 11 (A) High density of fluid inclusions in core of quartz grain, with no evidence of crosscutting secondary fluid inclusion trails. (B) Close up of inclusions in core of quartz vein shown A), showing high abundance of inclusions in zone parallel to growth zone in quartz (marked by red line). Note that inclusions in this FIA show evidence of post-entrapment modification (e.g. necked inclusions) and therefore no microthermometric data was collected. (C) Isolated group of inclusions in core of quartz grain, with no evidence of post-entrapment modification. Inset showing three-phase aqueous-carbonic inclusions. (D) Primary FIA with elongate, three-phase aqueous-carbonic fluid inclusions in trail parallel to growth zone in quartz crystal, with long axis of inclusions perpendicular to the growth zone. (E) Large three-phase aqueous-carbonic fluid inclusions in sphalerite, with negative crystal shape. (F) Sphalerite-hosted FIA with aqueous-carbonic inclusions



indicating the presence of other dissolved gases in these fluids (e.g. CH_4 , N_2). $T_{\text{h CO}_2}$ occurred to the liquid phase over a wide temperature range (3.5 to 26.5 °C), indicating highly variable CO_2 densities. Combined with ϕ_{CO_2} , $T_{\text{h CO}_2}$ equate to CO_2 concentrations of ~13 to 47% (Steele-MacInnis 2018). $T_{\text{m clath}}$ was observed prior to homogenization of the carbonic phase between 3.9 and 6.8 °C, which was used to calculate fluid salinities of 6.1 to 10.7 eq. wt% NaCl.

Most inclusions decrepitated between 200 and 250 °C, due to the high internal pressures of the aqueous-carbonic fluids, a phenomenon typical in CO_2 -bearing fluid inclusions in orogenic gold deposits (e.g. Tuba et al. 2021). $T_{\text{h tot}}$ was recorded in 20 fluid inclusions, with inclusions in samples from the South Zone having $T_{\text{h tot}}$ to the liquid and vapor phases (rare critical homogenization) between 304 and 368 °C. Fluid inclusions from a single sample from the North Zone have lower $T_{\text{h tot}}$ (263–285 °C), with homogenization only recorded to the liquid phase.

The bulk composition of aqueous-carbonic fluids was used to calculate the pressure at homogenization or at the temperature at which the inclusion decrepitated, which represents the minimum trapping pressures of these inclusions

(Steele-MacInnis 2018). These calculated pressures likely have a high degree of uncertainty caused by visual estimates of ϕ_{CO_2} (Bakker and Diamond, 2006) and possible presence of other dissolved gasses (e.g. CH_4 , H_2S). Minimum pressures were calculated from 43 inclusions and range from 1.9 to 2.2 kbars in the North Zone and 1.7 to 2.9 kbars in the South Zone (see ESM7). Assuming a density of 2.65 g/cm³ for the overlying rock column, these minimum pressure estimates correspond to minimum paleodepths of between 6.7 and 11.2 km under lithostatic pressure conditions. This suggests that mineralization at Sail Pond occurred at mesozonal rather than epizonal crustal depths.

Discussion

Structural control and timing of mineralization

The distribution of mineralized veins at Sail Pond is structurally and lithologically controlled with Ag-Cu-Zn-Pb-Sb mineralization spatially associated with brittle fractures in the dolostones within 100 m of dolostone-shale contacts.

This indicates a strong rheological control on mineralization, with the brittle dolostones of the St George Group interpreted to have been more conducive to open space fracturing than the surrounding shales of the Table Head or Goose Tickle groups, which underwent strong ductile deformation. A similar lithological control on the distribution of mineralized veins has been observed in other Ag-Pb-Zn deposits in clastic metasedimentary terranes. For example, at Keno Hill mineralized veins are preferentially distributed in quartzite, which was interpreted to have been more susceptible to brittle fracturing than the surrounding schists (Lynch 1989; Cathro 2006).

There is also a strong regional control on Sail Pond mineralization with all occurrences being located on the western margin of the WAWA in the hanging wall of a west directed Acadian thrust (Cawood 1993). This Acadian thrust is potentially associated with movement on larger scale crustal structures, such as the Cabot Fault, a major fault system in the Canadian Appalachians that extends along the eastern margin of the Great Northern Peninsula (Fig. 1). The Cabot Fault system has a prolonged tectonic history, and is estimated to have accommodated up to 250 km of dextral strike-slip from the Late Ordovician into the Carboniferous (Waldron et al. 2015; van Staal and Zagorevski, 2020). This major crustal structure could have functioned as a conduit for the transport of metal-bearing fluids, with mineralization occurring in second or third order structural splays from the main fault system (similar to models for orogenic gold mineralization; Goldfarb et al. 2005).

Although the exact timing of mineralization at Sail Pond is not well constrained, structural evidence indicates that veins were emplaced during this Acadian thrusting event, which occurred between 421 and 400 Ma in western Newfoundland (van Staal et al. 2009; van Staal and Barr 2012). Structurally controlled orogenic gold mineralization is reported along the eastern margin of the Great Northern Peninsula (Fig. 1), which is associated with second and third order structures off the Douciers Valley Fault Zone, part of the Cabot Fault system (Sandeman and Dunning 2016). At the Rattling Brook orogenic gold deposit in White Bay, to the south of the Sail Pond occurrence (Fig. 1), mineralization associated with the Douciers Valley Fault Zone has been dated between 415 and 409 Ma (Kerr and van Breeman 2007). Minnett (2012) reported ^{40}Ar - ^{39}Ar thermochronology from the nearby Viking deposit, indicating that the main stage of gold mineralization occurred at ca. 409 Ma. These data show that hydrothermal fluids were generated during peak metamorphism along the Cabot Fault zone during the Acadian (421–400 Ma). Further work is required to fully constrain the timing of mineralization at Sail Pond, including ^{40}Ar - ^{39}Ar thermochronology of muscovite/illite alteration of the host rocks to the mineralization. However,

current regional arguments suggest that it may be part of a broader regional-scale orogenic mineralization event that formed both orogenic Au and Pb-Zn-Ag vein mineralization during Acadian orogenic activity.

Genetic model for the formation of the Sail Pond occurrence

The geological, mineral chemistry, isotopic and fluid inclusion characteristics from Ag-Cu-Zn-Pb-Sb rich veins at the Sail Pond occurrence are markedly different from nearby carbonate hosted Zn-Pb deposits (Lane 1990; Conliffe et al. 2018, 2023) and highlight the need to develop an alternative genetic model for the Sail Pond occurrences. Fluid inclusions in quartz and sphalerite show mineralization is associated with an influx of high temperature ($>300\text{ }^{\circ}\text{C}$), moderate salinity (6.1 to 10.6 eq. wt% NaCl) aqueous-carbonic fluids. Mineralizing fluids at Sail Pond are similar to mineralizing fluids typically reported from mesozonal orogenic gold deposits globally (Ridley and Diamond 2000; Bodnar et al. 2014; Goldfarb and Groves 2015), as well as aqueous-carbonic fluids recorded in other structurally hosted Ag-Pb-Zn veins in clastic metasedimentary terranes such as Keno Hill (Lynch et al. 1990) and Coeur d'Alene (Leach et al. 1988).

Aqueous-carbonic fluids in orogenic gold systems are interpreted to represent fluids produced during devolatilization of metasedimentary (+ other) rocks during prograde greenschist facies metamorphism (Phillips and Powell 2010; Gaboury 2019). A similar origin for aqueous-carbonic fluids at Sail Pond is envisioned, with devolatilization of sedimentary rocks during the Acadian orogeny when they were metamorphosed to at least greenschist facies ($>300\text{ }^{\circ}\text{C}$; Stouge 1986). Calculated minimum trapping temperatures from fluid inclusions indicates that mineralization occurred at mesozonal crustal depths ($>6.7\text{ km}$), typical of orogenic gold deposits. There is no evidence for significant magmatic activity in the Humber Zone of western Newfoundland during the Acadian orogeny, and therefore a magmatic component of mineralization fluids is considered unlikely.

Carbon and oxygen isotopes of carbonates in hydrothermal ore systems are commonly used to characterize mineralizing fluids (e.g. McCuaig and Kerrich 1998; Bierlein et al. 2004; Lüders et al. 2015). However, the C and O isotopic signature can be influenced by interaction with host rocks, and isotopic fractionation during mineralization controlled by temperature, fluid chemistry (Eh, pH) and relative proportions of CO_2 and CH_4 ; therefore, interpretation of this data is often difficult or ambiguous (Goldfarb and Groves 2015). Despite these challenges, C isotope data from hydrothermal carbonates at Sail Pond shows a clear shift to more positive $\delta^{13}\text{C}$ values compared to the host dolostones (Fig. 10), suggesting that carbon was not sourced from decarbonization of

organic-rich metasedimentary rocks during metamorphism, which would result in hydrothermal carbonates with more negative ^{13}C values (Kontak and Kerrich 1997; Lüders et al. 2015; Conliffe et al. 2023). An alternative source for carbon may be decarbonization of the carbonate host rocks during metamorphism, which at temperatures $>300\text{ }^{\circ}\text{C}$ would be associated with strong carbon isotope fractionation and an increase of the $\delta^{13}\text{C}_{\text{CO}_2}$ values by about 2.5 to 3‰ relative to the parental carbonate source (Lüders et al. 2015).

The mineralizing fluids at Sail Pond have relatively high salinities when compared to fluids commonly associated with orogenic gold systems (3–7 eq. wt% NaCl; Ridley and Diamond 2000; Goldfarb and Groves 2015). The continental platform setting of many of the host rocks likely facilitated production of more saline metamorphic brines during greenschist facies metamorphism, compared to accretionary prism siliciclastic rocks and oceanic rocks more typical of Phanerozoic orogenic gold belts (Yardley and Graham 2002; Mortensen et al. 2022). The importance of fluid salinities in transporting base metals has been recognized from several lower temperature base metal deposits (Cooke et al. 2000; Hitzman et al. 2010; Huston et al. 2010; Wilkinson 2014) and has been suggested as a common characteristic of structurally hosted Ag-Pb-Zn veins in clastic metasedimentary terranes (Goldfarb et al. 2005; Taylor and Hofstra 2025). In orogenic gold systems, metals are thought to predominantly be transported as bisulfide complexes (Williams Jones et al., 2009), but the presence of low gold grades ($<336\text{ ppb Au}$; Piercey 2020) in Sail Pond mineralization suggests either the absence of gold or insufficient bisulfide to transport $\text{Au}(\text{HS})_2^-$. The low concentration of base metals in typical orogenic gold systems has been attributed to the low salinity of mineralizing fluids and the lack of chloride complexes to carry base metals (Phillips and Powell 2010; Yardley and Cleverly, 2015), and modelling of metamorphic devolatilization of pelitic rocks has shown only a small portion of base metals are liberated by the same processes associated with gold release (Zhong et al. 2015). The fluids from Sail Pond have higher salinities than typical orogenic fluids, however, indicating that the Sail Pond occurrence may have had sufficient salinity to allow both the leaching of Zn, Pb, Cu and Ag along the fluid pathway from footwall rocks, and subsequent transportation of metals in solution as chloride complexes (Yardley 2005). This is also consistent with results from polymetallic vein deposits at Coeur d'Alene, which have been interpreted to have been deposited from moderate to high salinity mineralizing fluids in polymetallic vein deposits at Coeur d'Alene (Taylor and Hofstra 2025).

Lead isotope data from galena at Sail Pond indicates a crustal source of these metals, consistent with leaching of

metals during fluid flow. The high degree in scatter in Pb isotope data observed at Sail Pond is typical of crustal-scale fault-associated (CSF) Phanerozoic orogenic gold deposits, which reflect the derivation of metals from multiple external sources during fluid transport along crustal-scale fault zones (Mortensen et al. 2022). Furthermore, the inter-grain variability present in the different samples suggest at least some local control on the Pb distribution in the mineralization, and that the Pb isotope signature of the Humber Zone and associated rocks was heterogeneous at the time of formation of mineralization. This is not surprising given the range of juvenile rocks (e.g., Cambrian rift rocks, obducted Cambro-Ordovician ophiolitic rocks) and varying ages of evolved rocks (e.g., Proterozoic Grenvillian basement and Cambro-Ordovician sedimentary cover) present in the region. Despite these inter-grain variations, there is significant overlap with Pb isotope data from other MVT and orogenic gold deposits on the Great Northern Peninsula (Fig. 8), suggesting that the metals in these deposits were sourced from similar crustal rocks but precipitated in structurally controlled veins at Sail Pond.

The metal-bearing fluids are interpreted to have been transported along crustal scale structures (e.g., thrusts in the WAMA), with mineralization occurring where these fluids encountered a suitable sulfur source and physiochemical conditions to precipitate sulfides. In situ and bulk sulfur isotopic data from sulfides at Sail Pond yield $\delta^{34}\text{S}$ values of 4.6 to 12.7‰ (Table 2; Fig. 9), higher than would be expected from a mantle or igneous sulfur source (typically $0\pm 5\text{‰}$, Seal, 2006), and indicate sulfur was not sourced from leaching of igneous or ophiolitic rocks or from magmatic activity. Given the high mineralization temperature ($>300\text{ }^{\circ}\text{C}$) and positive $\delta^{34}\text{S}$ values, bacteriogenic sulfate reduction is not considered a viable sulfate reduction mechanism at Sail Pond. Therefore, thermochemical sulfate reduction (TSR)-derived sulfur leached from the host rocks is the most likely sulphur source at Sail Pond, as has been suggested for nearby MVT-style Zn-Pb deposits (Conliffe et al. 2018).

Given the spatial associations with splays off of regional fault systems, broadly similar local to regional structural styles of mineralization, and requirement for Pb from older Humber Zone rocks, it is possible that the mineralization at Sail Pond was part of a regional post-peak metamorphic, Acadian regional-scale hydrothermal fluid flow system that involved the deposition of both orogenic Au (e.g., White Bay) and Ag-Zn-Pb (e.g., Sail Pond). This hypothetical origin requires further testing, but our results are consistent with regional models for orogenic Au deposits where deep crustal regions have Au mineralization, whereas at shallower levels of the crust there are structurally controlled Ag-Zn-Pb veins (e.g., Goldfarb et al. 2005).

Global comparisons and exploration implications

The geological and geochemical characteristics of mineralization at Sail Pond are highly unusual in the Appalachians and is unlike other mineral deposits in the Humber Zone. Carbonate hosted Zn-Pb mineralization formed at low temperatures (<180 °C) from the influx of metalliferous basinal brines sourced from the surrounding sedimentary sequences, at low metamorphic grades (sub-greenschist) in relatively undeformed carbonate platforms (Lane 1990; Conliffe et al. 2018). Vein hosted Zn-Cu-Pb-Ag mineralization on the Gaspé Peninsula, Quebec is interpreted as being related to IOCG mineralization (Simard et al. 2006) or epithermal in nature with a late orogenic fluid overprint (D'hulst 2007), and the abundance of felsic and mafic magmatism suggests a link between magmatism and mineralization.

In contrast, mineralization at Sail Pond shares more characteristics with orogenic gold deposits as defined by Goldfarb et al. (2005). These include a strong structural control on mineralization in second and third order structures on regional scale faults, association with greenschist facies metamorphism, and aqueous-carbonic mineralizing fluids with similar characteristics to orogenic Au deposit fluids. Moreover, there is evidence for the generation of potentially coeval orogenic Au fluids and associated deposits along the similar regional faults in the Humber Zone during the Acadian orogeny (e.g. Viking and Rattling Brook deposits along the DVFZ; Fig. 1).

On a global scale, mineralization at Sail Pond is most similar to structurally controlled Ag-Zn-Pb±Sb, Cu, Au mineralization in western North America at Keno Hill, Coeur d'Alene, the Kokanee Range and the Purcell Basin (Leach et al. 1988; Lynch 1989; Lynch et al. 1990; Beaudoin and Sangster 1992; Beaudoin et al. 1992; Fleck et al. 2002; Païement et al. 2012; Taylor and Hofstra 2025). Beaudoin and Sangster (1992) noted that these deposits are in clastic metasedimentary rocks that have undergone at least greenschist facies metamorphism, most are close to major crustal faults and are not associated with coeval felsic magmatism. Mineralization occurs in quartz-carbonate veins that are preferentially located in brittle lithologies that were susceptible to open fracturing during deformation (e.g. Keno Hill quartzite - Lynch 1989; Nelson Batholith - Beaudoin et al. 1992). Fluid inclusion studies from Keno Hill, Coeur d'Alene and Kokanee show that the mineralizing fluids were aqueous-carbonic fluids trapped at high temperature (250–350 °C) and at relatively deep crustal depths (>5 km; Leach et al. 1988; Lynch 1990; Beaudoin et al. 1992); evidence for fluid boiling and/or effervescence during mineralization is also common. Taylor and Hofstra (2025) also noted that mineralizing fluids in the Coeur d'Alene region are more saline than typical orogenic fluids,

and that chloride was an important ligand in metal transport. These fluids are interpreted to have been released during prograde metamorphism in accretionary orogenic settings with sulfur and metals being sourced from the surrounding metasedimentary rocks (Leach et al. 1988; Beaudoin and Sangster 1992; Beaudoin 1997; Païement et al. 2012; Hofstra et al. 2013). These features are all consistent with the geological setting, fluid inclusion data, and Pb-S isotope geochemistry of mineralization at Sail Pond.

Taylor and Hofstra (2025) suggested that the polymetallic vein deposits at the Coeur d'Alene district should be classified as a separate subset of orogenic gold deposits, due to the similarities between these deposit types. We further propose that the mineralization at Sail Pond, as well as that in other areas of western North America (Keno Hill, Kokanee and Purcell), could potentially be classified along with the Coeur d'Alene deposits as Coeur d'Alene-type deposits. This would represent a subset of orogenic gold deposits or broader orogenic-style mineralization due to the similarities in structural setting and fluid characteristics between Ag-Zn-Pb vein and orogenic gold deposits. In particular, the Sail Pond occurrence can be classified as a subset of the CSF subtype of Phanerozoic orogenic gold deposits proposed by Mortensen et al. (2022). They form in similar tectonic settings during prograde metamorphism in an accretionary environment, with mineralization forming close to the brittle-ductile boundary due to rapid depressurization of metal bearing fluids during tectonic activity (as shown by evidence for intermittent fluid effervescence during mineralization at Sail Pond). Although these deposits may have anomalous Au values (up to 1 g/t), they are characterized by their unusually high base metal and Ag contents. This may be due to pre-enrichment of the host sedimentary basins in base-metals and Ag prior to orogenesis, as they are often spatially associated with economic base-metal deposits (e.g. the world class Sullivan deposit in the Purcell Basin and the Daniels Harbour deposit in western Newfoundland). The main difference between Ag-Zn-Pb and orogenic Au systems are the higher fluid salinities of the mineralizing fluids in Ag-Zn-Pb systems, where fluid inclusions suggest these fluids would have been able to transport significantly higher concentrations of these metals as chloride complexes compared to typical low-salinity orogenic gold fluid.

Conclusions

The Sail Pond occurrence represents structurally controlled epigenetic vein hosted mineralization, similar to vein hosted Ag-Cu- Zn-Pb-Sb mineralization in western North America (e.g. Keno Hill, Coeur d'Alene, Kokanee and Païement areas). Geological, isotopic and fluid inclusion studies show

that mineralization at Sail Pond occurred at high temperatures (> 300 °C) and at mesozonal crustal depths (8–12 km), with fluids generated during the devolatilization of crustal sequences during prograde metamorphism. These moderate salinity (up to 7 eq. wt% NaCl) aqueous-carbonic fluids leached metals from the nearby crustal rocks and the mineralized veins formed in open space fractures in brittle host rocks, where metalliferous fluids encountered a suitable sulfate source.

The mineralization at Sail Pond and in western North America share many similarities with orogenic gold deposits, including: (1) location of deposits in second and third order structures related to major crustal faults; (2) regional metamorphism and deformation to at least greenschist facies, typically in major accretionary orogenies; and (3) CO₂-rich mineralizing fluids of probable metamorphic origin. Following the work of Taylor and Hofstra (2025), we propose that these deposits should be classified as Coeur d'Alene-type deposits, which represent a subset of orogenic gold deposits. The higher base metal and silver content of fluids related to the source rocks and the higher salinity of the mineralizing fluids. The mineralization at Sail Pond represents the first description of this deposit type outside of western North America and highlights the exploration potential for this deposit type in the Appalachians and other metamorphosed sedimentary platforms globally.

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Declarations

Conflict of interest The authors have no competing interests to declare that are relevant to the content of this article.

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