Paleoproterozoic gold and its tectonic triggers and traps: Implications from Re-Os sulphide and U-Pb detrital zircon geochronology, Lynn Lake, Manitoba

Christopher J.M. Lawley[1*], David Selby[2], William J. Davis[1], Eric Yang[3], Shuangquan Zhang[4], Simon E. Jackson[1], Duane C. Petts[1], A. Rhys O'Connor[5], and David A. Schneider[5]

[1]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8
[2]Durham University, Science Labs, Lower Mountjoy, South Road, Durham DH1 3LE UK
[3]Manitoba Geological Survey, 365-1395 Ellice Avenue, Winnipeg, Manitoba R3G 3P2
[4]Carleton University, 1125 Colonel By Drive, Ottawa, Ontario K1S 5B6
[5]University of Ottawa, 120 University, Ottawa, Ontario K1N 6N5
*Corresponding author's e-mail: christopher.lawley@canada.ca

ABSTRACT

Orogenic Au deposits require a trigger to transport auriferous fluids from their deep-seated source regions to their depositional site in the mid- to upper-crust. Defining the tectonic trigger of Au ore systems requires precise age constraints. Herein we address that knowledge gap with new Re-Os sulphide (arsenopyrite and pyrite) and U-Pb detrital zircon geochronology results from the Lynn Lake greenstone belt of the Paleoproterozoic Trans-Hudson Orogen, Manitoba, Canada. We document an early-stage of Au-rich veins (ca. 1.82 Ga) that are coeval with syenitic magmatism (1.83–1.82 Ga) and immediately post-date new ages for the opening and closure of synorogenic basins (1.84–1.83 Ga) and late-stage arc magmatism (1.84–1.83 Ga). Together these data point to the importance of stalled subduction and upwelling asthenosphere as possible triggers for fluid release during the earliest stages of continental collision between the Hearne, Superior, and Sask cratons. New dating further documents a second, overprinting generation of auriferous fluids (ca. 1.78 Ga) that post-dates peak metamorphism (ca. 1.81 Ga). These late-stage fluids were driven by a second thermal pulse (ca. 1.78 Ga), which, based on the timing of coeval, crustally derived pegmatitic dykes, may be related to crustal thickening and/or another unrecognized subcrustal heat source. Mineral exploration should focus on the large-scale architecture that is required to focus multiple pulses of auriferous fluids to the same depositional trap over the lifespan of an orogen.

INTRODUCTION

Modern and ancient mountain belts are an important geological setting for giant Au deposits, as demonstrated by the large number of such deposits hosted within accreted terranes along the western Cordillera and older deposits in the Canadian shield (Goldfarb et al., 1998; Groves et al., 1998). The sources of auriferous fluids in accretionary orogens are generally considered to be metamorphic, possibly due to devolatilization of the mafic lower crust and/or sulphidic sediments during the active subduction of oceanic lithosphere (Goldfarb et al., 1998; Goldfarb and Groves, 2015). Because the duration of individual mineralization events is brief (≤0.1 m.y.) relative to the lifespan of an orogen (10–100 m.y.), a trigger is required to drive Au-bearing fluids, likely metamorphic in origin, from their upper mantle and/or crustal sources to their depositional sites in the mid- to upper-crust (Kerrich and Wyman, 1989; Goldfarb et al., 1991; Wyman et al., 2016).

The tectonic trigger(s) for Au-bearing fluid transport in modern accretionary settings include changes in subduction velocity, ridge subduction, slab roll-back, plate reorganization, and/or plume impingement (Goldfarb and Groves, 2015). Many of these triggers are also thought to be important to porphyry and epithermal deposits in modern arc and/or back-arc settings (Sillitoe, 2008; Goldfarb et al., 2014). Collisional orogens are also endowed with orogenic-style Au deposits (Goldfarb et al., 2001), but the tectonic trigger(s) in these settings is less clear because active subduction of oceanic lithosphere may not be directly involved and/or may be occurring 100s to 1000s km away from the Au deposits (Li et al., 2012; Li et al., 2015). Instead, the cessation of subduction, stalling of the subducting slab, lithospheric delamination, plume impingement, and/or far-field subduction effects may trigger the devolatilization of upper mantle to crustal sources in collisional to post-collisional settings (Richards, 2011; Groves and Santosh, 2016).

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Although the tectonic differences and the role active subducting oceanic lithosphere has between accretionary and collisional orogens are well described in the literature, the implications of differing orogenic styles on Au metallogeny are less well understood (Barley and Groves, 1991; Pehrsson et al., 2016). One of the factors contributing to this uncertainty is that Au deposits associated with the early accretionary orogenic stage, if present, were likely reworked and overprinted during later collisional orogenesis (Groves et al., 2016). The reworking of accretionary orogens is proposed as a possible explanation for the Au-rich signature of continent-continent collisions that incorporate early formed oceanic arcs (Leahy et al., 2005). The situation is even more complex along cratonic margins, which were typically reworked during multiple, overprinting orogens spanning billions of years (Carpenter et al., 2005; Davies et al., 2010; Lawley et al., 2013, 2015). Determining the timing of Au mineralization within this complex orogenic collage requires precise age constraints.

Herein we address that knowledge gap with new Re-Os arsenopyrite and pyrite, and U-Pb detrital zircon geochronology for the Paleoproterozoic Lynn Lake greenstone belt (LLGB of Manitoba; Fig. 1–3). These ages provide new constraints on the timing of Au mineralization, which, coupled with new U-Pb detrital zircon ages for synorogenic sedimentary basins, highlight the relationship between changes in tectonic style (e.g. accretionary versus collisional orogenesis) and its relationship to metallogenesis.

LOCAL GEOLOGY

The LLGB is hosted within the Paleoproterozoic Trans-Hudson Orogen (1.9–1.8 Ga: Ansdell, 2005; Corrigan et al., 2009). Rhyolitic flows are the oldest dated lithostratigraphic unit and were broadly coeval with mafic to intermediate volcanism (Baldwin et al., 1987; Beaumont-Smith and Böhm, 2002, 2004; Beaumont-Smith et al., 2006). The close structural intercalation of mafic volcanic rocks with disparate geochemical compositions suggests that the Wasekwan Group likely comprises rocks that initially formed in distinct environments but were juxtaposed during later structural reworking (Zwanzig, 1997, 2000; Zwanzig et al., 1999; Beaumont-Smith, 2008). Multiple volcanic rock packages are also demonstrated by distinct rhyolite ages within the northern and southern belts of the LLGB (Beaumont-Smith and Böhm, 2002, 2004). Based on the emplacement of the Pool Lake suite (1.89–1.87 Ga), which intrudes and stiches the different belts, some of these disparate mafic volcanic rock packages must have been structurally intercalated by ca. 1.87 Ga (Baldwin et al., 1987; Beaumont-Smith and Böhm, 2002, 2004; Beaumont-Smith et al., 2006). Mafic to ultramafic intrusions that host the Lynn Lake Ni mine were emplaced into the amalgamated Wasekwan-Pool Lake suite at 1871 ± 2 Ma (i.e. El gabbro: Turek et al., 2000). All three igneous assemblages were subsequently intruded by dykes and plutons of the 1.86–1.85 Ga Wathaman suite (Meyer et al., 1992), which marks the accretion of the LLGB and the Hearne craton to the north at ca. 1.85 Ga (Ansdell, 2005; Corrigan et al., 2005).

The Wasekwan Group is unconformably overlain by conglomerate (e.g. Hughes Lake unconformity; Fig. 3), psammite, arkose, and wacke comprising the Sickle Group (Anderson and Beaumont-Smith, 2001). A similar conglomerate separates the LLGB and large swaths of the Sickle Group and turbidite-like rocks comprising the Burntwood Group (i.e. Kisseynew Basin) to the south (Fig. 2). Previously reported ages suggest that the Burntwood Group extends to older maximum depositional ages than the structurally overlying coarse clastic sedimentary successions (Machado et al., 1999). However, both sedimentary packages are interstratified, suggesting that deposition of coarse clastic and deep-marine sedimentary facies were locally coeval (Zwanzig and Bailes, 2010).

RESULTS

Preliminary Re-Os Arsenopyrite and Pyrite Geochronology

Multiple pyrite and/or arsenopyrite mineral separates were prepared (n = 38) from different vein and alteration types collected at three deposits (MacLellan, Gordon, and Burnt Timber) as part of the current study (Fig. 4). Of these, only 19 samples yielded sufficient Re to attempt Re-Os dating (total analyses = 23) following the methods of Lawley et al. (2013).

Paleoproterozoic arsenopyrite and pyrite results scatter about a ca. 1.8 Ga errorchron (1794 ± 18 Ma; 2σ; MSWD = 159; n = 15; York model 3; initial $^{187}Os/^{188}Os = 1.0 \pm 0.9$). Reproducible replicate analysis of the sample, or multiple samples prepared from the same vein, strongly suggest that the excess data-point scatter is not an analytical artefact. Instead, excess data-point scatter is more likely a product of multiple sulphide generations of slightly to significantly different age.

Two replicate analyses of the two arsenopyrite samples with the highest $^{187}Re/^{187}Os$ ratios (6132–16564; n = 4) yield weighted average Re-model ages at 1824 ± 12 Ma and 1782 ± 16 Ma (n.b. model age uncertainties are reported at 2σ and include uncertainties in the ^{187}Re decay constant of 1.666×10^{-11} yr^{-1} and the initial Os composition = 0.5 ± 0.2; Smoliar et al., 1996). Model age calculations for these two samples is appropriate because nearly all of the measured ^{187}Os ($^{187}Os^{r}$

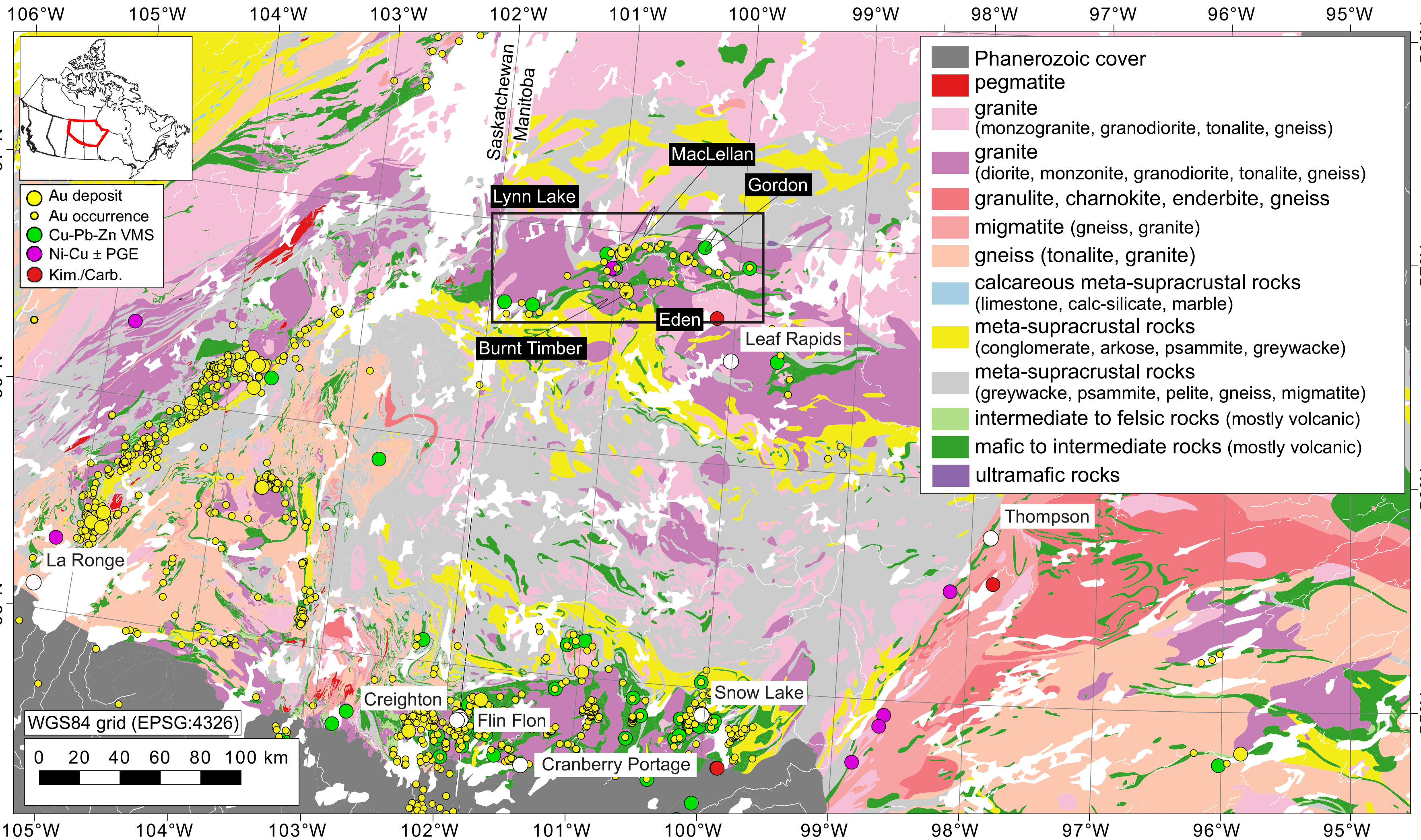


Figure 1. Regional geological map of Saskatchewan and Manitoba (Saskatchewan geological polygons from the Saskatchewan Mining and Petroleum GeoAtlas; Manitoba geological polygons from Manitoba Mineral Resources(2013)). Volcanogenic massive sulphide (VMS; Galley et al., 2007), nickel-platinum group element (Eckstrand and Hulbert, 2007), gold deposits (Dubé and Gosselin, 2007), kimberlite (Kim.; Faure, 2010) and carbonatite (Carb.; Woolley and Kjarsgaard, 2008) are shown for reference. Gold occurrences are also shown for reference (data are commodity-filtered from the provincial mineral occurrence databases).

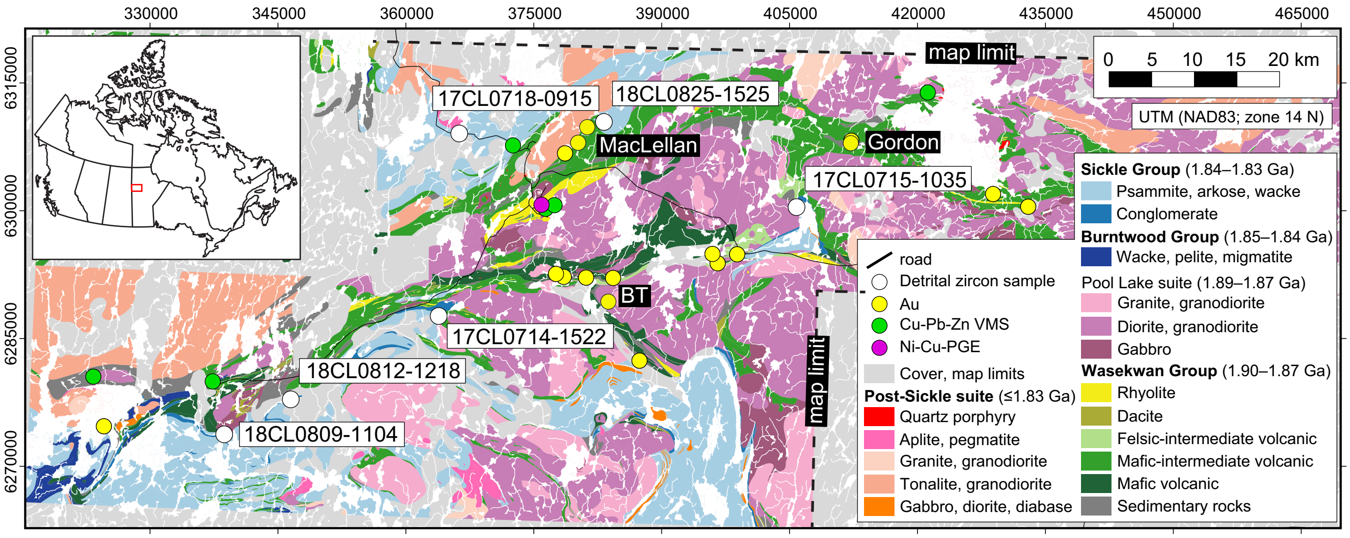


Figure 2. Local geology map of the Lynn Lake greenstone belt (Gilbert et al., 1980; Gilbert, 1993). Detrital zircon sample localities, volcanogenic massive sulphide (VMS; Galley et al., 2007), nickel-platinum group element (Eckstrand and Hulbert, 2007), and gold deposits (Dubé and Gosselin, 2007) are shown for reference.

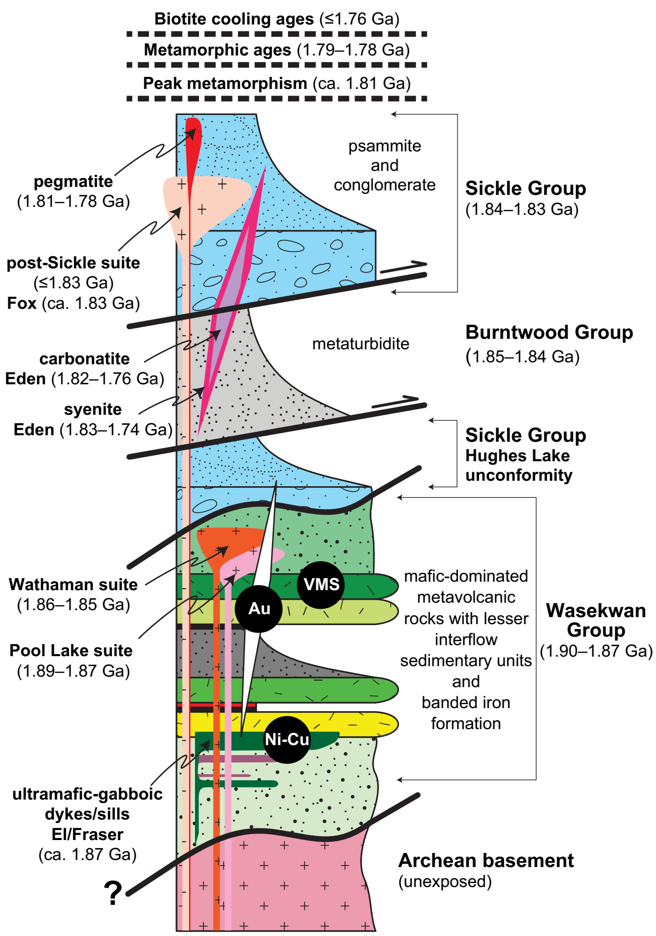


Figure 3. Schematic stratigraphic column summarizing the main supracrustal groups and plutonic suites comprising the Lynn Lake greenstone belt. Ages are summarized from several sources (*see* text for further details). Abbreviations: BIF = banded iron formation, VMS = volcanogenic massive sulphides.

= 99%) is due to radiogenic decay of ^{187}Re (Stein et al., 2000). The older ca. 1.82 Ga arsenopyrite sample was taken from a pre- to syn-D_2 vein at MacLellan; whereas the younger ca. 1.78 Ga arsenopyrite sample was collected from the sulphidized metavolcanic host rock at the same deposit. Two replicate analyses of one moderately radiogenic pyrite sample (^{187}Re/^{188}Os = 694–1005; ^{187}Os^{r} = 98%) represent the best available timing estimate for sulphide replacement at Gordon (1838 ± 28 Ma). A regression through a subset of reproducible arsenopyrite and pyrite samples from both deposits yield a York Model 1 Re-Os isochron age of 1827 ± 9 Ma (MSWD = 0.4; n = 5; initial ^{187}Os/^{188}Os = 0.46 ± 0.03).

The remaining arsenopyrite and pyrite analyses scatter to lower ^{187}Re/^{188}Os ratios (57–2418), but are still moderately radiogenic (^{187}Os^{r} = 57–99%) and yield a range of Re-Os model ages (1980–764 Ma). A subset of the older (≥1.83 Ga) pyrite vein analyses from MacLellan yield a York Model 1 Re-Os isochron age of 1865 ± 21 Ma (MSWD = 0.04; n = 4; initial ^{187}Os/^{188}Os = 0.48 ± 0.06). Two of these three samples were of pyritized biotite schist (Fig. 4e,f), which may suggest that the oldest Re-Os model ages reflect early sulphide deposition that predates regional metamorphism. However, the remaining pyrite sample within this potential group was sampled adjacent to vein pyrite samples that yield Re-Os model ages older than the host rock, suggesting that some of the older Re-Os model ages may also reflect isotopic disturbance. A subset of anomalously young samples, including the two samples from Burnt Timber, yield model ages that cluster around ca. 1.3 and 0.8 Ga. The scatter of anomalously young Re-Os results may reflect disturbance and/or variable resetting to produce geologically mean-

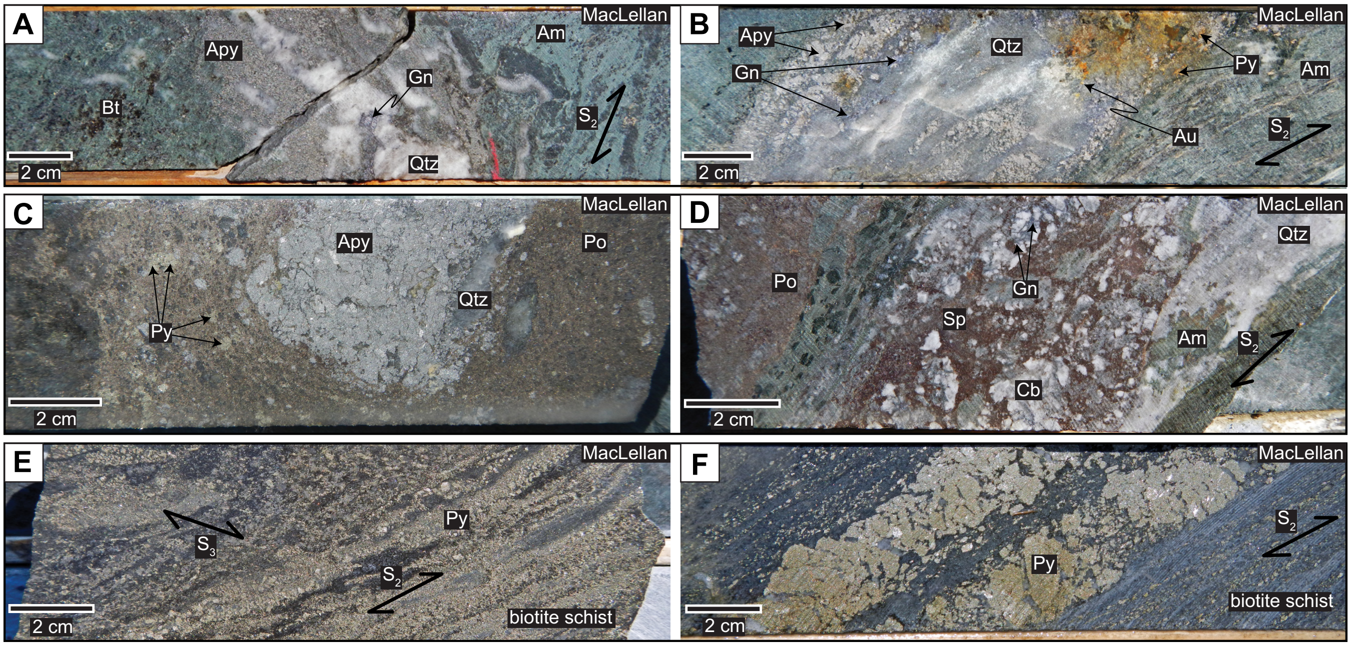


Figure 4. Photographs of drill core from ore zones of the MacLellan Au deposit: **a and b)** folded and/or transposed pre- to syn-D_2 quartz-arsenopyrite veins that cut metavolcanic rocks (amphibolite); **c)** sulphide replacement ore zone with coarse arsenopyrite; **d)** transposed and brecciated base metal vein cutting metavolcanic rocks (amphibolite); **e and f)** sulphide replacement with biotite schist. Pyrite defines S_2 and S_3 fabrics, suggesting post-D_2 fluids and/or remobilization (e). Coarsening of arsenopyrite (c) and pyrite (f) likely occurred during metamorphism. Abbreviations: Am = amphibole, Apy = arsenopyrite, Bt = biotite, Cb = carbonate, Gn = galena, Po = pyrrhotite, Py = pyrite, Qtz = quartz, and Sp = sphalerite.

ingless model ages, or possibly reflect reworking of the Trans-Hudson Orogen during the assembly and breakup of Rodinia (McDannell et al., 2018). As a result, the reproducible Re-Os model ages for the two highly radiogenic samples from MacLellan represent the preferred timing for sulphidation in the LLGB (ca. 1.82 and 1.78 Ga).

Preliminary U-Pb Detrital Zircon Geochronology

Six samples for detrital zircon analyses were collected across the LLGB as part of the current study (Fig. 2). Zircon mineral separates were prepared and analyzed using the sensitive high mass-resolution ion microprobe (SHRIMP) at the Geological Survey of Canada following the methods of Stern (1997).

Three of the samples represent Sickle Group conglomerate, which at the Hughes Lake area unconformably overlie the mafic volcanic host rocks (i.e. Wasekwan Group) of the Au deposits (Fig. 2, 3). Two samples, comprising psammite and pebble conglomerate, were collected from what is thought to be higher stratigraphic levels of the Sickle Group. One additional sample was collected from what was previously mapped as the Ralph Lake conglomerate. The youngest reproducible zircon grain from the entire data set is 1836 ± 15 Ma, which is interpreted to represent the maximum depositional age for the Sickle Group. The youngest detrital zircon grains from each of the other Sickle Group samples overlap within analytical uncertainty at 2σ (≤1.84 Ga); whereas the youngest grain from the Ralph Lake conglomerate is older at 1860 ± 5 Ma. The minimum depositional age for the Sickle Group is constrained by a suite of intermediate (i.e. tonalite, diorite, granodiorite, and quartz diorite) to felsic (i.e. granite) intrusions and plutons that are thought to intrude the Sickle Group (i.e. post-Sickle suite; Fig. 2, 3; Beaumont-Smith and Böhm, 2002, 2004). The Fox mine tonalite is thought to represent one of these suspected post-Sickle intrusions and is dated at 1831 ± 4 Ma (Turek et al., 2000). If correct, deposition of the Sickle Group therefore likely occurred during the earliest stage of the ca. 1.83 Ga collision between the composite Hearne-LLGB with the Superior craton to the southeast (Fig. 1, 5, 6). The implications of synorogenic basin closure to the Au endowment of the LLGB are discussed below.

DISCUSSION

New U-Pb detrital zircon geochronology from multiple quartzite and conglomerate localities, together with the age of the so-called post-Sickle intrusive suite (e.g. Fox mine Tonalite, 1831 ± 4 Ma: Turek et al., 2000; Dunphy Lakes tonalite 1829 ± 2 Ma: Beaumont-Smith, 2003), constrain the depositional timing of the Sickle Group to 1836–1831 Ma in the LLGB (Fig. 2, 3, 5, 6). The interpreted Sickle Group depositional age is coeval with similar siliciclastic rocks comprising the Missi

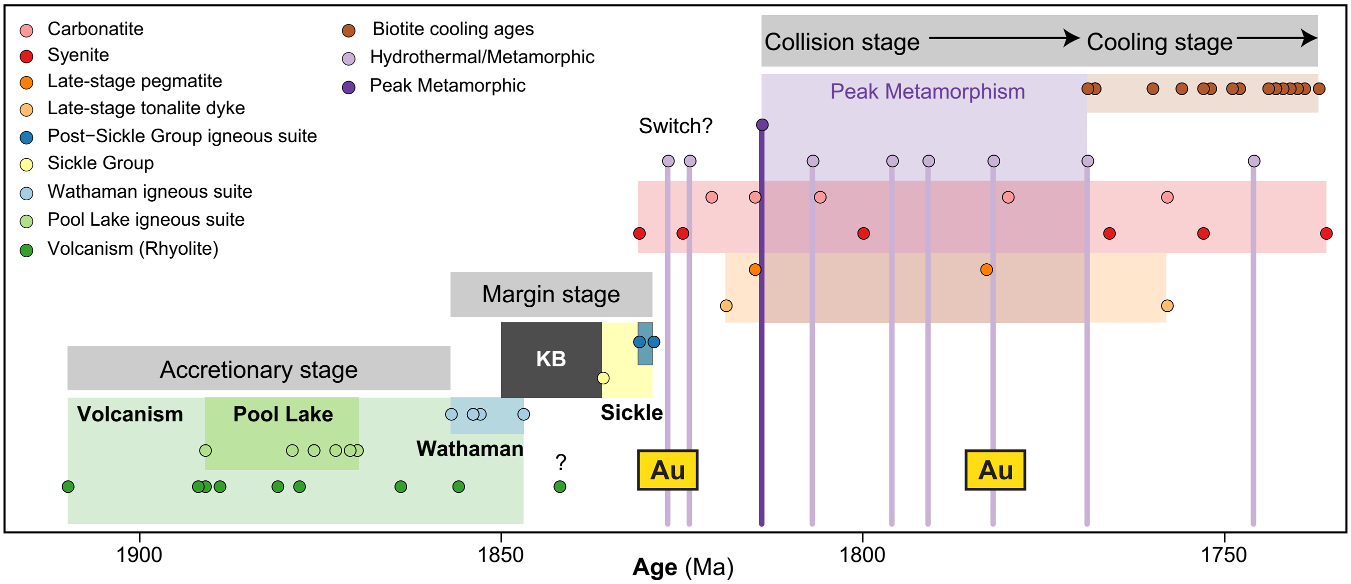
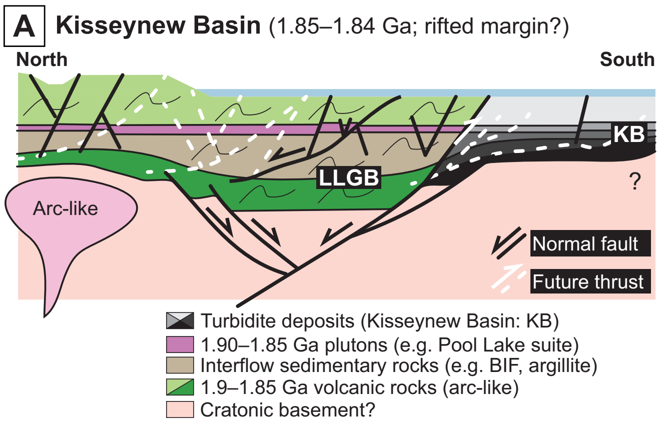


Figure 5. Compilation of previously reported and new ages for the Lynn Lake greenstone belt. Volcanic and plutonic U-Pb ages are from Baldwin et al. (1987), Turek et al. (2000), Beaumont-Smith and Böhm (2002, 2004), Beaumont-Smith et al. (2006), and Lawley et al. (2019). Inferred peak metamorphic zircon age from Beaumont-Smith and Böhm (2002, 2004). The scatter of syenite and carbonatite ages are from Elliott (2009) and may reflect a combination of crystallization ages and metamorphic resetting. Sickle Group (U-Pb detrital zircon maximum depositional age) and hydrothermal/metamorphic ages (Re-Os sulphide and U-Pb monazite and xenotime) were collected as part of the current study. Biotite cooling ages are from O'Connor et al. (2019).

and MacLellan groups south and west of Lynn Lake, respectively (Fig. 1; Ansdell et al., 1995; Machado et al., 1999; Maxeiner and Demmans, 2000; Zwanzig, 2000). South of the LLGB, synorogenic basins (i.e. Missi Group) share a similar tectonic evolution from rapid deposition, wide-spread intrusion of arc-like magmatism at 1.83–1.82 Ga (Ansdell and Norman, 1995; Norman et al., 1995; Connors et al., 1999; Machado et al., 1999; Zwanzig, 2000; Hollings and Ansdell, 2002), folding, burial, and peak metamorphism (ca. 1.81 Ga; Fig. 6). Ansdell and Norman (1995) suggested that the youngest of these intrusions (1.83–1.82 Ga) marks a period of intra-arc rifting prior to peak metamorphism. Extension and/or asthenospheric upwelling prior to the major phase of crustal thickening is partially supported by syenitic intrusions and mafic to ultramafic intrusions within the Kisseynew Basin (e.g. 1.83–1.82 Ga Touchbourne suite; Machado et al., 1999; Martins et al., 2011, 2012) and similar rocks east of Lynn Lake (ca. 1.83–1.82 Ga syenite within the Eden Lake complex; Fig. 1, 3, 5, 6; Chakhmouradian et al., 2008; Elliott, 2009). Potassic intrusions and volcanic rocks (i.e. Christopher Island Formation), which are exposed across much of the northern Trans-Hudson Orogen in Nunavut, occurred at broadly the same time (Ansdell, 2005). If correct, this brief synorogenic extensional period must have occurred immediately prior to crustal thickening, which ultimately led to the intrusion of crustally derived peraluminous granite at ca. 1816–1814 Ma (Kraus and Menard, 1997; Whalen et al. 1999; Zwanzig and Bailes, 2010) and pegmatite dykes (1815 ± 3 Ma; Fig. 5, 6; Beaumont-Smith and Böhm, 2002).

The timing of hydrothermal activity, metamorphism, and deformation are constrained by U-Pb monazite and xenotime ages (1.83–1.75 Ga; Lawley et al., 2019) and new Re-Os arsenopyrite ages reported herein (ca. 1.82 and 1.78 Ga; Fig. 5). The oldest, garnet-hosted xenotime provide a maximum age estimate for the peak metamorphic mineral assemblage at ≤1827 Ma, which is coeval with the oldest Re-Os model age of the highly radiogenic arsenopyrite sample at MacLellan (1824 ± 12 Ma). Both ages demonstrate that hydrothermal activity occurred within a few million years after deposition of the Sickle Group (1836–1831 Ma; Fig. 5, 6). One of the dated xenotime crystals occurs within the same veinlet as native Au (Lawley et al., 2019), which, coupled with close spatial association between arsenopyrite and ore zones at MacLellan, suggests that this earliest hydrothermal event was Au-bearing (Fig. 4a–c). The timing of early auriferous fluids marks an important transitional period in the development of the Trans-Hudson Orogen, which was triggered by the terminal collision between the composite Hearne-LLGB with the Superior and/or Sask cratons to the south (Fig. 5; Ansdell, 2005). Early-stage Au veins thus postdate the accretion of volcanic arcs (i.e. Wasekwan Group) to the Hearne cratonic margin by ca. 30 m.y. (Fig. 5). This early phase of auriferous fluids also predates most estimates for peak metamorphism within this part of the Trans-Hudson Orogen (ca. 1.81 Ga; Schneider et al., 2007), but are coeval with some estimates for an early-



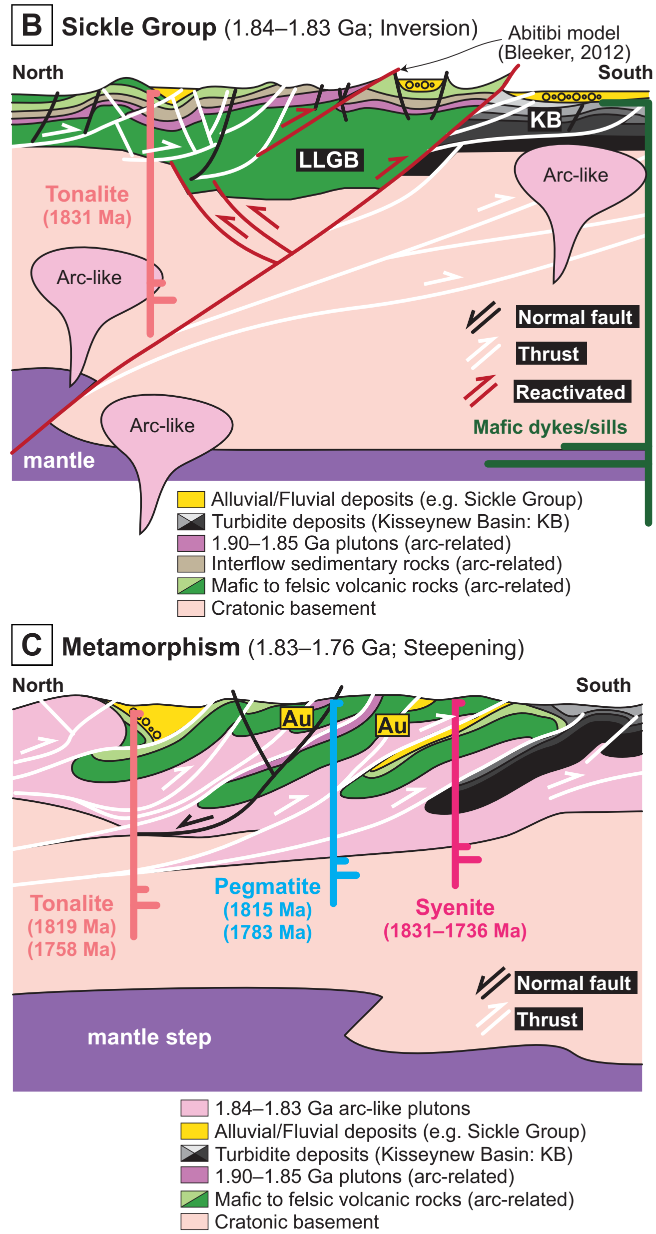


Figure 6. Schematic diagram showing basin development and inversion (inverted basin geometry *adapted from* Scisciani et al., 2014). The Wathaman batholith stitches the southern Hearne cratonic margin with the Trans-Hudson Orogen, suggesting that the Lynn Lake greenstone belt (LLGB) had collided by ca. 1.85 Ga. If correct, deep-water sediments of the Kisseynew Basin (KB) would have been deposited during a period of extension at 1.85–1.84 Ga (Ansdell, 2005; Zwanzig and Bailes, 2010). Structural inversion of this extensional basin was coeval with the youngest generation of arc-like granitic plutons and deposition of the fluvial/alluvial Sickle Group (1.84–1.83 Ga), which were deposited on top of the older arc LLGB basement (1.90–1.85 Ga). In the Abitibi greenstone belt (Ontario-Quebec), Timiskaming-type basins mark reactivated extensional faults that focus younger auriferous fluids (Cameron, 1993; Bleeker, 2012). Collision of the amalgamated Hearne-LLGB cratonic margin with the Superior and/or Sask cratons was coeval with peak metamorphism and steepening and folding of the inverted basin structural architecture. The earliest auriferous fluids dated herein immediately post-date the tectonic switch to basin inversion, but are coeval with syenite and small, late-stage dykes (tonalite and pegmatite). Abbreviation: BIF = banded iron formation.

stage of high-temperature, low-pressure metamorphism south of the LLGB, early folding and syenitic intrusions (e.g. 1.83–1.82 Ga; Fig. 5; Ansdell and Norman, 1995; Norman et al., 1995; Connors et al., 1999; Machado et al., 1999; Elliott, 2009; Martins et al., 2011).

Younger phosphate and sulphide ages suggest that hydrothermal activity and metamorphism continued for at least 80 m.y. after the onset of continent-continent collision (Fig. 5; Lawley et al., 2019). Arsenopyrite dated at 1782 ± 16 Ma based on the current study is broadly equivalent with individual hydrothermal monazite and xenotime ages that range from 1.81 to 1.77 Ga (Lawley et al., 2019). Monazite and xenotime at the margins of pyrite within the hydrothermal MacLellan host rocks, which, along with lesser sulphide-hosted phosphate inclusions, suggest that hydrothermal fluids overprinted auriferous veins during multiple tectonothermal events within this time interval (Fig. 4c,e,f). The punctuated hydrothermal history closely matches the clustering of metamorphic ages determined by monazite and titanite dating within unmineralized rocks peripheral to the LLGB (Schneider et al., 2007; Couëslan et al., 2013). Hydrothermal and/or metamorphic phosphate ages in the LLGB are also consistent with a broad period of high-temperature and low- to medium-pressure metamorphism across the Trans-Hudson Orogen, including peak metamorphic conditions during crustal thickening and on-going continent-continent collision at ca. 1.81 Ga (Beaumont-Smith and Böhm, 2002; Ansdell, 2005; Schneider et al., 2007; Couëslan et al., 2013). Ultimately, crustal thickening led to the production of pegmatite dykes (≤1.81 Ga) and other crustally derived melts (e.g. 1816–1814 Ma peraluminous granite; Kraus and Menard, 1997; Whalen et al., 1999; Zwanzig and Bailes, 2010) during the same period (Fig. 6). Because sulphide replacement represents the dominant ore style in some parts of the MacLellan deposit, it is possible that hydrothermal fluids at ca. 1.78 Ga introduced new Au into the system (Fig. 5). The MacLellan deposit thus likely records multiple, overprinting Au-bearing events. A similar multi-stage genesis at this deposit was previously based on the relative chronology of veins and contrasting hydrothermal alteration assemblages (Samson and Gagnon, 1995; Samson et al., 1999; Hastie et al., 2018).

The youngest xenotime occurs with pyrite and is also aligned with the main, ductile deformation fabric (D_2) at the microscale, suggesting that deformation reactivated earlier structures at and/or after ca. 1746 Ma. Late-stage reactivation of early structures significantly postdates Re-Os arsenopyrite ages, which is consistent with folded and transposed auriferous veins at each of the studied deposits (Fig. 4). Metamorphism continued until approximately the same time based on new U-Pb apatite (1.75–1.70 Ga; O'Connor et al., 2019) and Ar-Ar biotite cooling ages (1.77–1.74 Ga; O'Connor et al., 2019). Continued cooling ultimately led to resetting of K-Ar biotite and muscovite ages to between 1.76 and 1.61 Ga (Fig. 5; Moore et al., 1960; Lowdon et al., 1963; Turek, 1967; O'Connor et al., 2019). It is possible that younger and undated hydrothermal events led to minor remobilization of Au into late-stage structures during cooling (Beaumont-Smith et al., 2000). However, pre- to syn-D_2 faults are the dominant structural control for Au mineralization across the LLGB (Anderson and Beaumont-Smith, 2001; Beaumont-Smith and Böhm, 2002). The ages and sequence of events described above are based on preliminary ages that will be refined in a future publication that is currently in preparation.

IMPLICATIONS FOR EXPLORATION

The timing of Au mineralization has important implication for ore system models and mineral exploration targeting. Giant orogenic Au deposits in the southern Abitibi greenstone belt (Ontario and Quebec) are interpreted to have formed late during the main deformational event (e.g. Canadian Malartic at 2664 ± 11 Ma: De Souza et al., 2015), which was coeval with or slightly pre-dating peak metamorphism (e.g. 2657.5 ± 4.4 Ma: Piette-Lauzière et al., 2019). This peak Abitibi Au event also occurred within a few million years after the closure of the synorogenic sedimentary basins (i.e. ca. 2669 Ma Timiskaming-type assemblage: Bleeker, 2012) and ca. 2680–2670 Ma alkaline magmatism. The spatial relationship between synorogenic basins and fertile fault segments have long been used as an exploration vector in the Abitibi (Cameron 1993, Hodgson, 1993; Robert et al., 2005; Bleeker, 2012).

The similar timing relationships between Au deposition, sedimentation, metamorphism, and alkaline magmatism documented for the LLGB herein points to a particular orogenic stage that is common to multiple Au districts through space and time (Fig. 5, 6; Kerrich and Wyman, 1989). Although the dated syenitic intrusions are not observed within the Au-deposit stratigraphy in the LLGB (Fig. 1) and are thus unlikely to represent the source of auriferous fluids, their timing, coupled with closure of the synorogenic basins, marks an important shift in the tectonic evolution of the Trans-Hudson Orogen (Fig. 5, 6). Ansdell (2005) suggested that alkalic, potassic, and carbonatitic magmatism and volcanism at this time was related to upwelling asthenosphere, which may have re-melted metasomatized portions of the lithospheric mantle after the termination of subduction. We demonstrate that the earliest auriferous fluids in the LLGB were coeval with this tectonic switch and were possibly driven by the heat from the upwelling asthenosphere (Fig. 5). In the Abitibi, upwelling asthenosphere was coeval with extensional basins as demonstrated by the deposition of deep-water sedimentary facies that are coeval with alkaline magmatism within the Timiskaming assemblage. Ansdell and Norman (1995) also attributed syenitic magmatism and synorogenic basins to a period of extension immediately predating continent-continent collision within the Trans-Hudson Orogen. Maximum depositional ages for deep-water facies within the Kisseynew Basin (≤1845 and ≤1842 Ga: Machado et al., 1999) are consistent with opening of synorogenic extensional basins at broadly the same time. However, deep-water facies are not recognized within the Sickle and Missi groups. Instead, these nominally younger synorogenic basins developed during an early phase of thrusting and folding (Fig. 6; Zwanzig, 1997), which obscures the link, if any, between extensional faulting and Au (cf. Bleeker, 2012). Sickle Group rocks also host very few Au occurrences and none of the major Au deposits in the LLGB (Fig. 2), which is unlike the mineralized Timiskaming-style basins in the Abitibi (e.g. Bleeker, 2012). In fact, synorogenic basins in the central Kisseynew Basin are almost entirely devoid of Au occurrences (Fig. 1). Because these basins were also metamorphosed and deformed by the same generations of faults that host Au in the LLGB, we speculate that their poor endowment is likely due to the scarcity of appropriate depositional traps and/or fluid focusing within the older greenstone belts. More fertile volcano-sedimentary panels may still underlie synorogenic basins in some cases, but such primary stratigraphic relationships have since been reworked during multiple stages of folding and faulting (Fig. 6).

From an exploration perspective, the distribution of synorogenic basins and syenitic intrusions is likely to demarcate the large-scale, lithospheric architecture that may have been important for focusing auriferous fluids from their deep crustal and/or upper mantle source regions. Critically, this earliest phase of Au vein-hosted mineralization is overprinted by a later generation(s) of Au-rich veins, suggesting that this large-scale architecture was also important for focusing younger auriferous fluids within the evolving orogen (Fig. 5). The tectonic trigger for the younger, post-peak metamorphic fluids is less well understood, but is coeval with a second thermal pulse, as demonstrated by coeval prograde metamorphic monazite ages elsewhere in the Trans-Hudson Orogen (Schneider et al., 2007; Couëslan et al., 2013) and crustally derived pegmatitic dykes (Chiarenzelli et al., 1998; Bickford et al., 2005). The isotopic composition (Pb-Pb and Sm-Nd) of these pegmatitic dykes suggests that they may have remelted the Archean basement (Bickford et al., 2005). Crustal anatexis may have been triggered by significant crustal thickening (White, 2005) and/or some other as yet poorly defined subcrustal heat source. Together the data point to the importance of multiple thermal pulses (e.g. asthenospheric upwelling; crustal anatexis) as the main drivers of deep-seated fluids. The tectonic triggers for auriferous fluids were coeval with these thermal pulses, including the switch to the main collisional orogenic stage of the Trans-Hudson Orogen (Fig. 5). Ascending auriferous fluids were focused along large-scale structures before being deposited in a variety of geochemical and structural traps within the older accretionary arcs (Fig. 1, 2). Mineral exploration targeting should thus focus on identifying the large-scale architecture exploited by these fluids whilst also considering the most favourable depositional environments for Au within the host greenstone belts.

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