MAP2007-1

Geology of the Chisel–Anderson lakes area, Snow Lake, Manitoba (NTS areas 63K16SW and west half of 63J13SE)

By
A.H. Bailes and A.G. Galley

GEOSCIENTIFIC MAP

Manitoba

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Winnipeg, 2007

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Bailes, A.H. and Galley, A.G. 2007: Geology of the Chisel–Anderson lakes area, Snow Lake, Manitoba (NTS areas 63K16SW and west half of 63J13SE); Manitoba Science, Technology, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2007-1, scale 1:20 000 plus notes.

NTS grid: 63K16SW, 63J13SE

Keywords: Manitoba, Snow Lake, Paleoproterozoic, Flin Flon Belt, economic geology, boninite, tholeiite, rifting, massive sulphide deposits, hydrothermal alteration, island arcs

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Published by:
Manitoba Science, Technology, Energy and Mines
Manitoba Geological Survey
360-1395 Ellice Avenue
Winnipeg, Manitoba
R3G 3P2 Canada
Telephone: (800) 223-5215 (General Enquiry)
(204) 945-4154 (Publication Sales)
Fax: (204) 945-8427
E-mail: minesinfo@gov.mb.ca
Website: manitoba.ca/minerals

Cover photo: Quartz-feldspar–rich altered rims on pillows in Welch basalt (unit J1), 200 m east of north end of Stroud Lake. This zone of alteration, which characterizes the upper 300–500 m of Welch basalt, has been traced over strike length of more than 20 km.

Abstract

Located at the east end of the exposed Paleoproterozoic Flin Flon Domain, the Chisel–Anderson lakes map area is dominated by 1.84–1.81 Ga fold-thrust–style tectonics and by a southwest-verging allochthon of volcanic rocks (Snow Lake arc assemblage). The area includes a lithologically and structurally diverse sequence of deformed and metamorphosed volcanic, sedimentary and intrusive rocks, and contains 8 of the belt's 27 producing and past-producing base metal mines. These include the currently producing Chisel North Zn-Cu mine and 7 past-producers, plus several unmined deposits, including the newly discovered Lalor deposit. The volcanogenic massive sulphide (VMS) deposits occur in an approximately 6 km thick succession of volcanic rocks (Snow Lake arc assemblage) that records, in its stratigraphy and geochemistry, a temporal evolution in geodynamic setting from a 'primitive arc' (Anderson sequence), to a 'mature arc' (Chisel sequence) and then to an 'arc-rift' (Snow Creek sequence). The VMS-hosting Anderson and Chisel sequences contain evidence of intra-arc rifting. In the Anderson sequence, this includes the combination of boninite, low-Ti basalt and isotopically juvenile rhyolite flows, an association of rock types that has been attributed in both modern and Phanerozoic arcs to high-temperature hydrous melting of refractory mantle sources in an extensional and/or proto-arc environment. Rifting in the Chisel sequence is indicated by the presence of the stratigraphically overlying Snow Creek 'arc-rift' basalt, with mid-ocean ridge basalt (MORB)–like geochemistry. Indirect support for rifting in the Chisel sequence is provided by voluminous volcaniclastic detritus; prominent synvolcanic dike sets; and the local presence of highly fractionated, differentiated magma series. The VMS deposits in both sequences are spatially associated with juvenile rhyolite complexes, synvolcanic intrusive rocks (dominated by two large trondhjemitic bodies) and extensive areas of rocks that underwent hydrothermal alteration.

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Introduction

The Chisel–Anderson lakes map area, located at the east end of the exposed Paleoproterozoic Flin Flon Domain (Figure 1), contains 8 of the belt's 27 producing and past-producing base metal mines (Syme et. al., 1996). These include the currently producing Chisel North Zn-Cu mine and 7 past-producers, plus several unmined deposits, including the newly discovered Lalor deposit, in addition to numerous subeconomic base metal occurrences. The area includes a lithologically and structurally diverse sequence of deformed and metamorphosed volcanic, sedimentary and intrusive rocks. Although most rocks are recrystallized to almandine-amphibolite–facies mineral assemblages, many have recognizable primary features (Figure 2) and are referred to by their original names without metamorphic prefixes. Due to the obliteration of primary textures, rocks in the Chisel–Anderson lakes area that are composed of volcanic fragments are referred to by the general term volcaniclastic (Fisher, 1961) and classified according to the size terminology of Fisher (1966). Fragments are interpreted to have been derived largely by epiclastic processes, with the terms tuff and lapilli tuff not denoting a pyroclastic derivation.

Regional setting

Rocks of the Chisel–Anderson lakes map area are part of the juvenile (internal) Reindeer zone of the Trans-Hudson Orogen (THO), a collision zone formed during the 2.0–1.8 Ga amalgamation of several Archean cratons into the Laurentian supercontinent (Hoffman, 1988). They belong to the Flin Flon Domain, which is a tectonic collage of 1.92–1.88 Ga, predominantly volcanic assemblages juxtaposed during 1.88–1.87 Ga intraoceanic accretion and subsequent 1.84–1.78 Ga terminal collision of the bounding Archean cratons (Lucas et al., 1996). Volcanic rocks in the Flin Flon Domain include oceanic arc, ocean floor (back arc), oceanic island and oceanic plateau assemblages (Syme and Bailes, 1993; Stern et al., 1995a, b). Oceanic arc assemblages include suites of tholeiite, calcalkaline rocks and rare shoshonite and boninite (Stern et al., 1995b), nearly all of which are identical to those forming in modern intraoceanic arcs (e.g., Gill, 1981). All mined volcanogenic massive sulphide (VMS) deposits in the Flin Flon Domain, including those in the map area, are hosted by 1.92–1.87 Ga juvenile oceanic arc assemblages (Syme and Bailes, 1993). Nearly all VMS deposits occur in two of the six fault-bounded, 20–50 km wide, juvenile oceanic arc assemblages in the Flin Flon Domain, specifically those at Flin Flon and Snow Lake (Syme and Bailes, 1993; Syme et al., 1996).

The east end of the Flin Flon Domain (Figure 1) is dominated by 1.84–1.81 Ga fold-thrust–style tectonics (Connors, 1996; Krause and Williams, 1999; Zwanzig, 1999) and is characterized by a series of southwest-verging allochthons of volcanic rocks. The volcanic domains are bounded by thrust faults (e.g., Snow Lake Fault, McLeod Road Fault), separated by intervening slices of younger, ca. 1.84 Ga sedimentary rocks belonging to the Burntwood Group (Zwanzig, 1990; Connors, 1996; David et al., 1996; Bailes and Schledewitz, 1998). They are characterized by layer-parallel foliations and isoclinal to tight folds ($F_{1\text{-}2}$ of Kraus and Williams, 1998). The deformed fold-thrust package is intruded (stitched) by 1.84–1.83 Ga late 'successor arc' granitic plutons, further deformed by northeast-plunging open folds (F_3 of Kraus and Williams, 1998) and recrystallized to lower to middle almandine-amphibolite–facies mineral assemblages during a 1.82–1.81 Ga regional metamorphic episode (Froese and Moore, 1980; David et al., 1996).

Geology of the map area

The map area is underlain by four main components: 1) 1.89–1.88 Ga juvenile oceanic arc basalt, basaltic andesite, dacite and rhyolite, 'arc-rift' basalt and related, mainly subvolcanic, intrusions (units J1–J37) of the Snow Lake arc assemblage; 2) 1.85–1.84 Ga marine turbidite deposits of the Burntwood Group (unit B1); 3) 1.84–1.83 Ga calcalkaline 'successor arc' plutons (units P2–P6); and 4) ca. 1.82–1.80 Ga tectonite units (units W1, W2) and young intrusive rocks (units L1, L2). With the exception of the last group of rocks, all were recrystallized during the 1.82–1.81 Ga regional metamorphic episode.

Snow Lake arc assemblage

The Snow Lake arc assemblage is exposed in a 20 km wide allochthon that contains one of the thickest and best exposed volcanogenic massive sulphide (VMS)–hosting sequences in the Flin Flon Domain. The allochthon is bounded to the north and southwest by 1.85–1.84 Ga Burntwood Group turbiditic sedimentary rocks that are separated to the north by the Snow Lake Fault and to the southwest by an unnamed fault and several crosscutting 1.84–1.83 Ga granitic plutons. The allochthon itself is cored by the $F_{1\text{-}2}$ Anderson Bay antiform that has subsequently been broadly buckled by north-northeast–trending folds, the largest of which is the F_3 Threehouse synform. To the southeast, rocks of the Snow Lake arc assemblage are abruptly truncated by the Berry Creek Fault, a transcurrent dextral strike-slip fault that extends the entire length of the exposed Flin Flon Domain in Manitoba. Most rocks in the map area lie on the north limb of the $F_{1\text{-}2}$ Anderson Bay anticline and, for the most part, young northwards.

The Snow Lake arc assemblage is >6 km thick and has three distinct subdivisions (Figures 3–5). The first and lowermost, the Anderson sequence (units J1–J5), is a bimodal mafic and felsic succession dominated by flows with negligible volcaniclastic rocks that are intruded by a 1.5 km by 22 km, sill-like, trondhjemitic, subvolcanic intrusive complex. In contrast, the overlying Chisel sequence (units J6–J33) is lithologically diverse and displays rapid lateral facies variations and abundant volcaniclastic rocks. The third and uppermost, the Snow Creek sequence, is composed of pillowed basalt (units J34, J35) with no intercalated felsic rocks or volcaniclastic detritus. Together the stratigraphy and geochemistry of these rocks record a temporal evolution in development of the Snow Lake arc assemblage from 'primitive arc' (Anderson sequence) to 'mature arc' (Chisel sequence) to 'arc-rift' (Snow Creek sequence; Bailes and Galley, 1996, 1999).

Anderson sequence

The Anderson sequence is >2.5 km thick and composed of arc tholeiite (unit J1, Welch basalt), boninitic flows (unit J2), isotopically juvenile rhyolite (unit J3; e.g., Daly), local heterolithic mafic volcanic breccia (unit J4), and the sill-like, 1.5 km by 22 km, synvolcanic Sneath intrusive complex (unit J5;

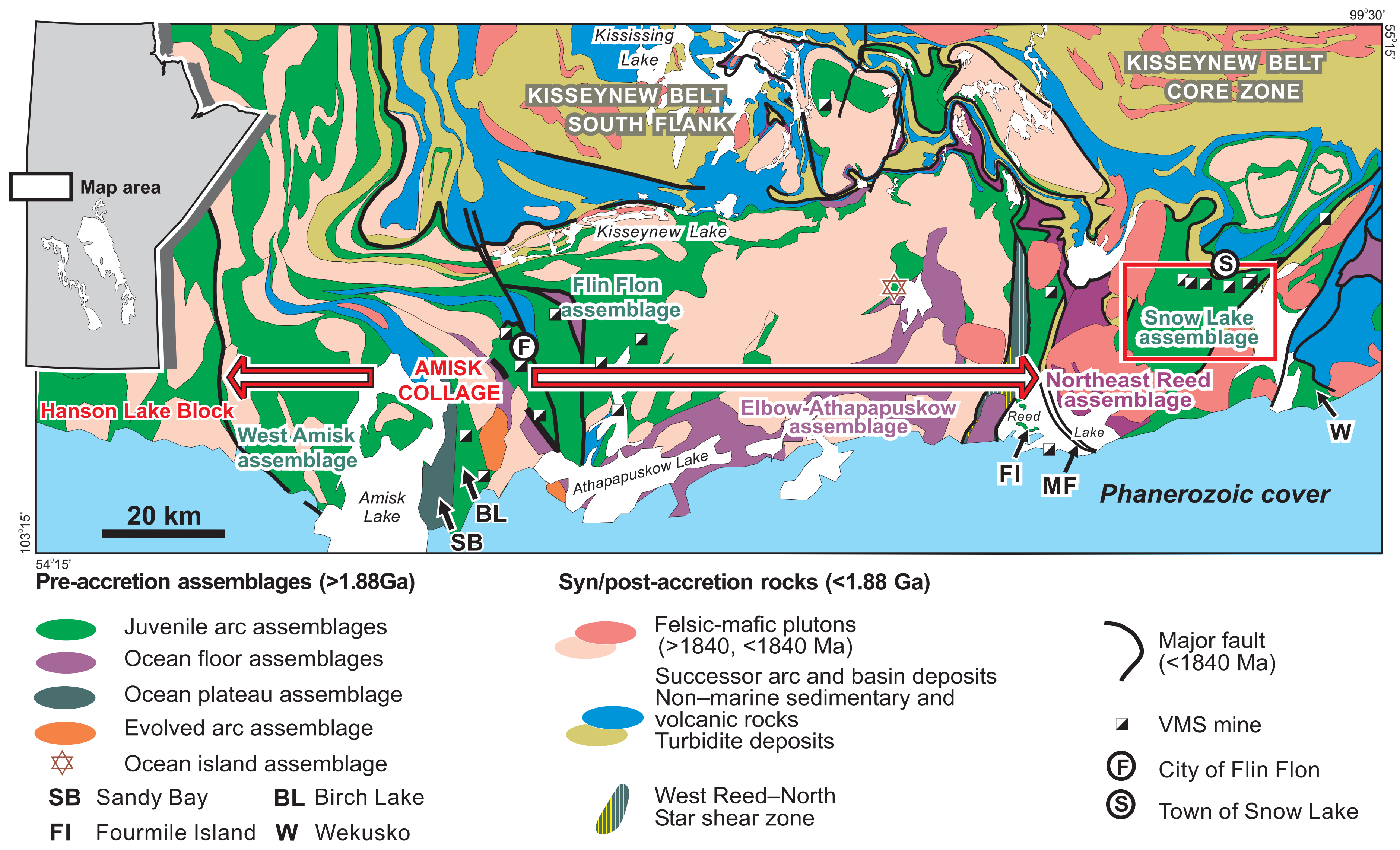

Figure 1: *Simplified geology of the central and eastern portions of the Flin Flon Domain, showing major tectonostratigraphic assemblages and plutons, and locations of mined VMS deposits. Inset shows the location of the Flin Flon Domain (white rectangle) in Manitoba. The Chisel–Anderson lakes map area, which is the rectangle at the east end of the domain that is outlined in red, is characterized by a structural style and rock types that are more comparable to the Kisseynew Domain than those in the Amisk Collage of the central Flin Flon Domain. The map area consists of a series of Kisseynew-type allochthons of 1.89 Ga volcanic and 1.85–1.83 Ga sedimentary rocks, soled by the Morton Lake Fault zone (MF; Syme et al., 1995; Krause and William, 1999).*

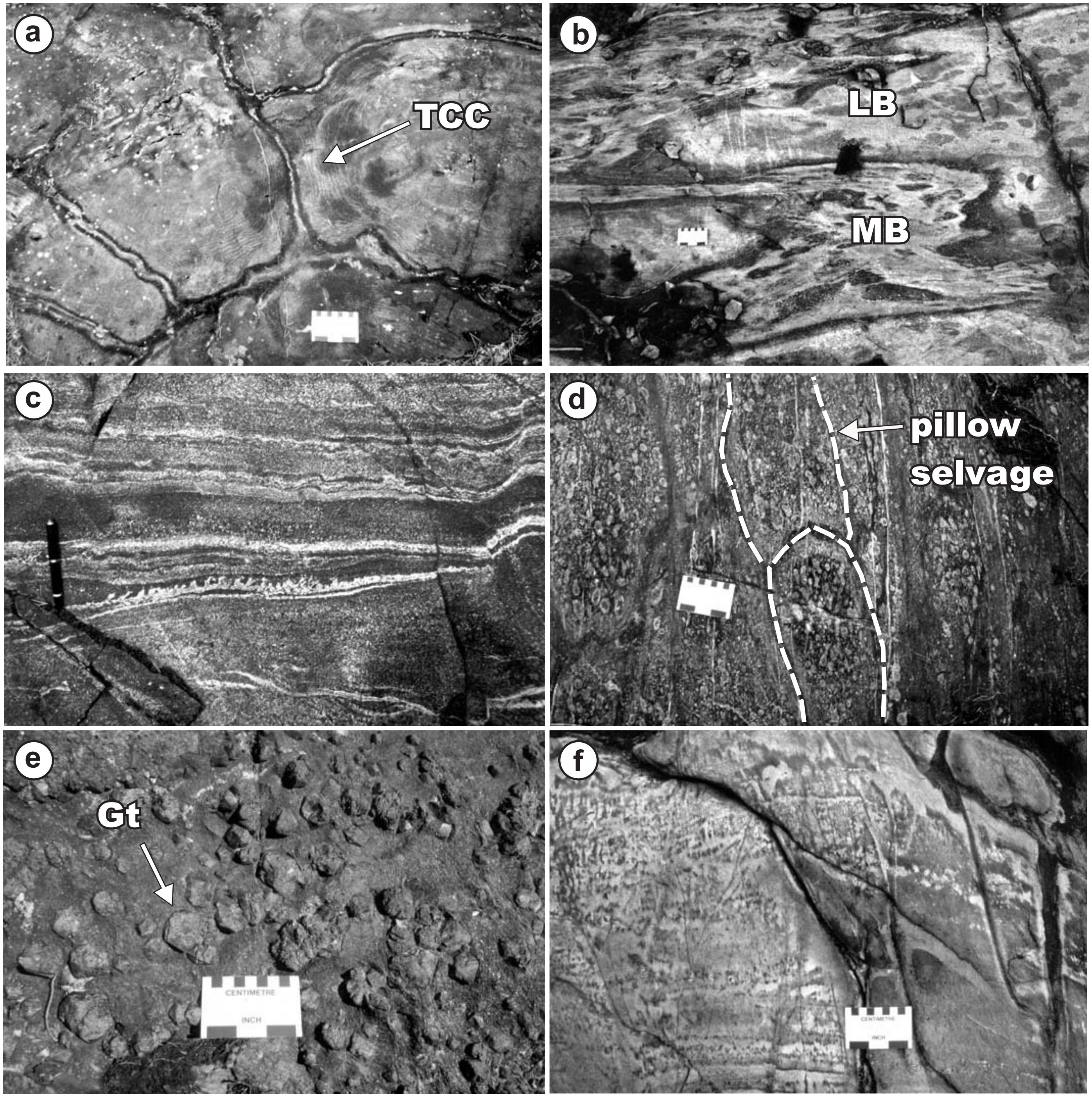

***Figure 2:** Unaltered, well-preserved volcanic rocks (a–c) compared to altered volcanic rocks (d–e) of the Snow Lake arc assemblage: **a)** pillowed, plagioclase-phyric basalt with dark selvages, light interpillow hyaloclastite and albite-filled thermal contraction cracks (TCC), Snell basalt (unit J7); **b)** lobes of massive rhyolite (LB) with intervening microbreccia (MB), Daly rhyolite (unit J3); **c)** thin-bedded mafic tuff with delicate load structures, Threehouse mafic volcaniclastic rocks (unit J19a); **d)** pillowed basalt with prominent mottled to pervasive albitization and silicification (dashed lines outline pillows), Welch basalt (unit J1a); **e)** Fe-Mg–metasomatized mafic breccia completely overprinted by chlorite and prominent porphyroblasts of garnet (Gt) that were produced during the 1.81 Ga regional metamorphic event, Edwards mafic volcaniclastic rocks (unit J8a); **f)** mottled albitization and silicification overprinting mafic wacke, Edwards mafic volcaniclastic rocks (unit J8a).*

Figure 4). Ubiquitous pillowed flows and a paucity of volcaniclastic rocks in the sequence suggest deposition below wave base and at depths that precluded any significant subaerial eruptions. Based on their geochemistry, Stern et al. (1995b) suggested that rocks of the Anderson sequence were formed in a nascent arc tectonic setting.

Anderson supracrustal rocks

Welch basalt (unit J1): These pillowed to massive aphyric (subunit J1a), sparsely porphyritic (subunit J1b), plagioclase (±pyroxene)–phyric (subunit J1c) and local pyroxene (±plagioclase)–phyric (subunit J1f) flows display a crude upward change in composition from basalt to andesite. They

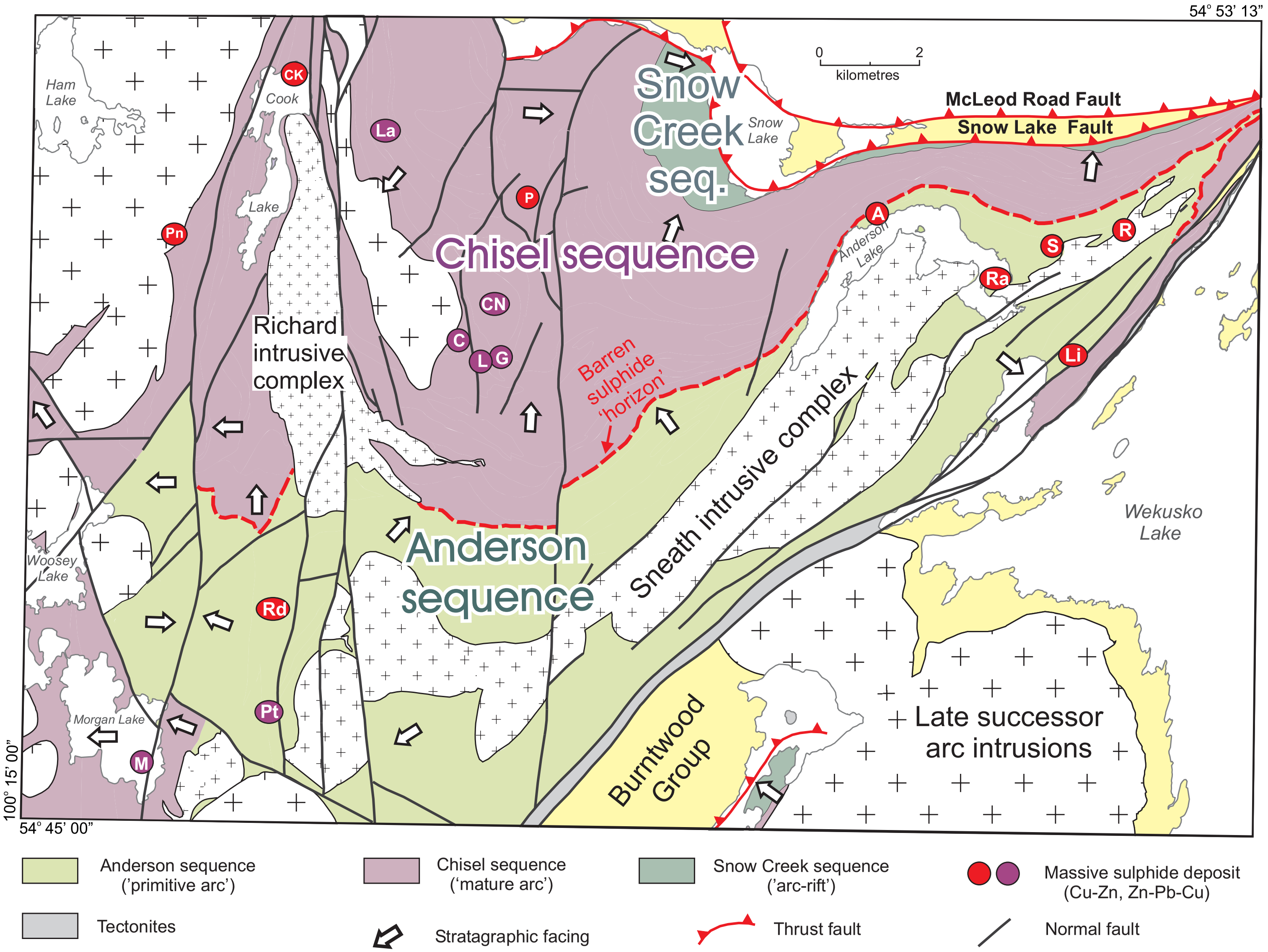

Figure 3: *Plan view of the 1.89 Ga Anderson ('primitive arc'), Chisel ('mature arc') and Snow Creek ('arc-rift') sequences of the Snow Lake arc assemblage. The area covered is identical to the map but has been extended slightly north in order to show the location of the Cook Lake VMS deposit. The two large synvolcanic intrusive complexes, Sneath and Richard, belong to the Anderson and Chisel sequences, respectively. The ca. 1.84–1.83 Ga sedimentary Burntwood Group rocks are composed of detritus derived from volcanic and intrusive rocks of the Flin Flon Domain. Arrows show facing directions of supracrustal rocks. The late successor arc intrusive rocks truncate the ca. 1.84 Ga thrust faults (teeth) and are themselves cut by ca. 1.78 Ga normal faults (solid lines). Filled circles and ovals (red, Cu-Zn; purple, Zn-Cu) show distribution of VMS deposits: A, Anderson Lake; C, Chisel Lake; CK, Cook Lake (located just north of map boundary); CN, Chisel North; G, Ghost Lake; La, Lalor; Li, Linda; L, Lost Lake; M, Morgan Lake; Pt, Pot Lake; P, Photo Lake; Pn, Pen; Rd, Raindrop; R, Rod; Ra, Ram; S, Stall Lake. The 'Foot-Mud horizon' (thick red dashed line) is a pyritic, fine-grained volcaniclastic unit located at the contact between the Anderson ('primitive arc') and Chisel ('mature arc') sequences.*

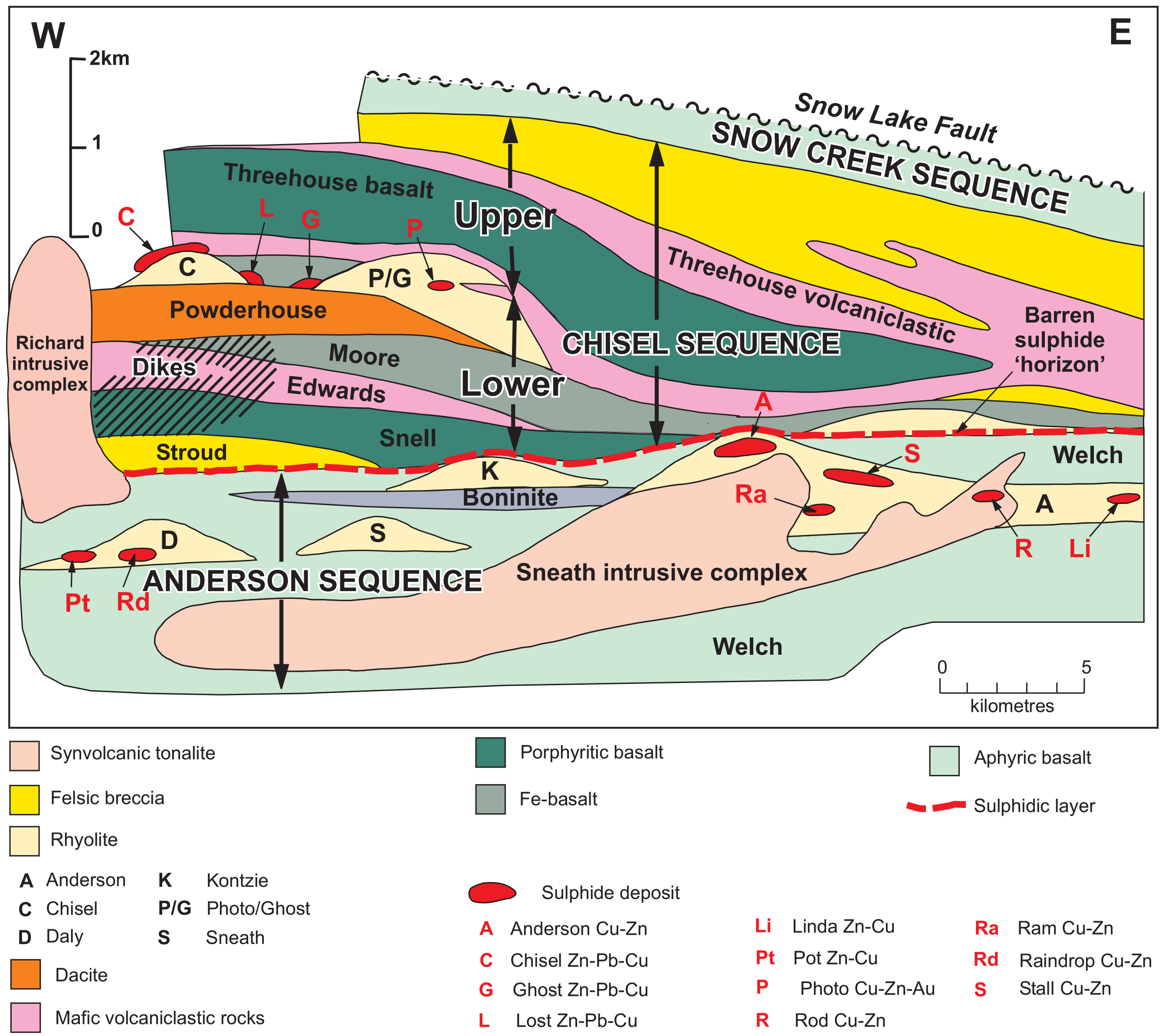

Figure 4: *Schematic stratigraphic section showing setting of VMS deposits in the eastern two-thirds of the ca. 1.89 Ga Snow Lake arc assemblage. Note the spatial association of sulphide deposits with rhyolite complexes (names of rhyolite complexes indicated by letters. The Sneath and Richard plutons are composite intrusions that include several phases, with younger phases (not shown separately in this diagram) intruding VMS-hosting rhyolite units and alteration zones. For example, a late phase of the Sneath intrusion has intruded and enveloped the Rod VMS deposit (R). The U-Pb zircon age for the Sneath Lake pluton is poorly constrained (1886 +17/–9 Ma; David et al., 1996) and only indicates the subvolcanic nature of the pluton and not an actual age relative to the younger Stroud breccia (1892 ±3 Ma; David et al., 1996) and subvolcanic Richard pluton (1889 +8/–6 Ma; Bailes et al., 1991).*

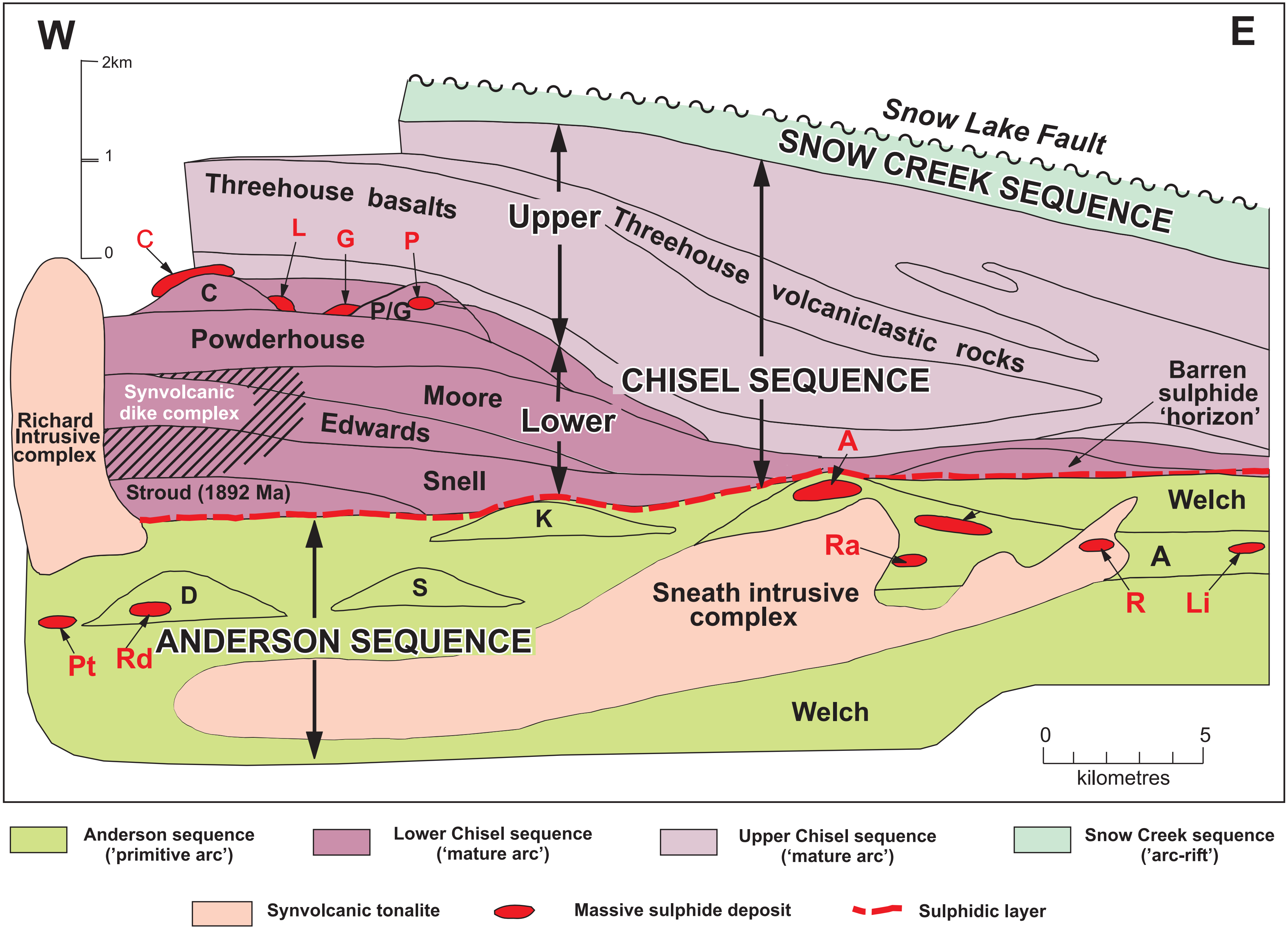

Figure 5: *Division of the rocks into their three main subdivisions: Anderson sequence ('primitive arc'), Chisel sequence ('mature arc'), Snow Creek ('arc-rift'). All notations on this figure are identical to those used on Figure 4.*

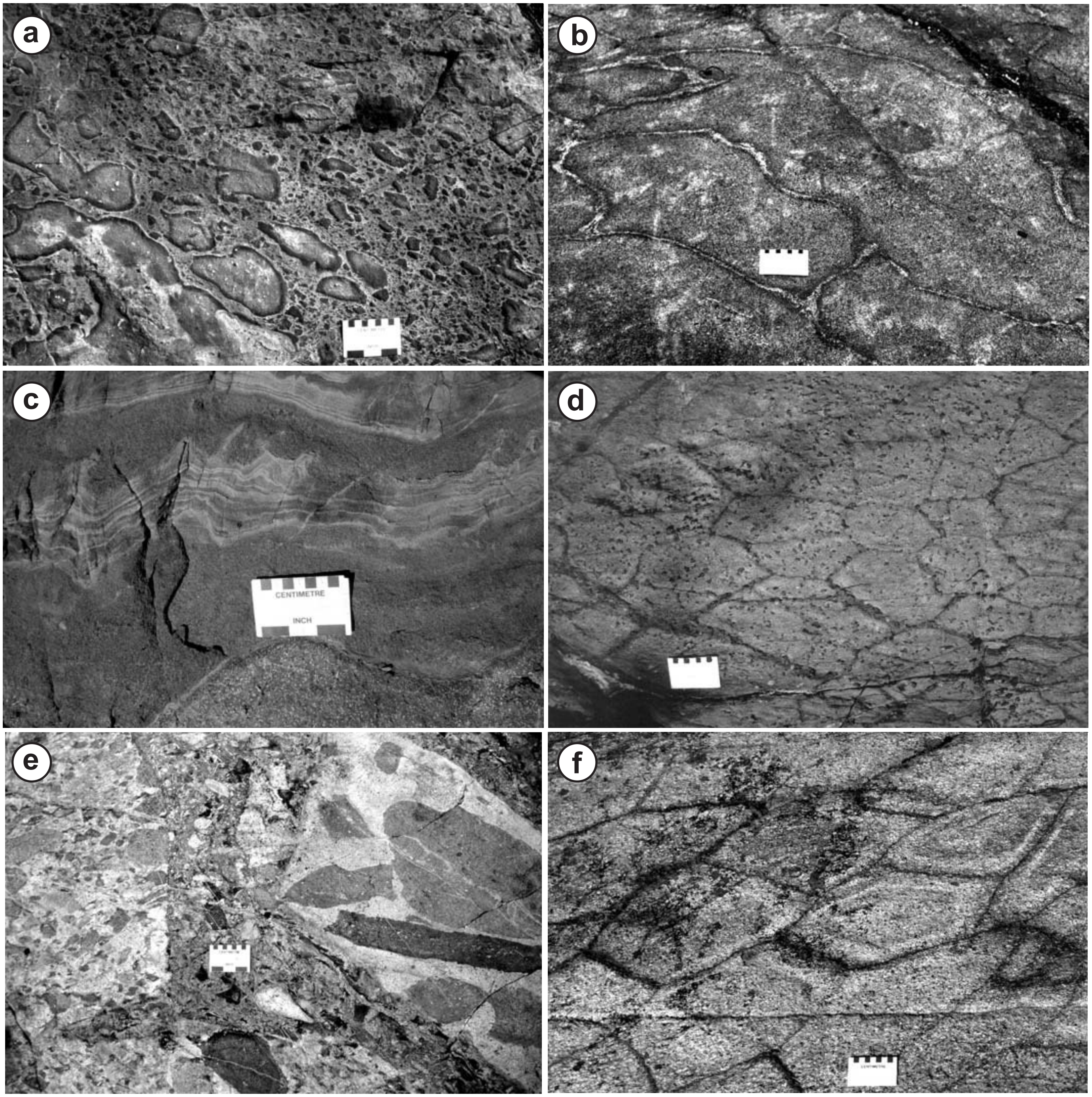

Figure 6: *Anderson sequence ('primitive arc') volcanic and intrusive rocks:* ***a)*** *amoeboid pillow breccia, Welch basalt (subunit J1a); note dark selvages (former glassy rims) around fragments);* ***b)*** *draped imbricated pillows in pyroxene (amphibole-replaced)-plagioclase–phyric basalt, Welch boninite (subunit J2b); base of scale card parallels flow contact;* ***c)*** *graded bedding and load structures in interflow mafic tuff overlying Welch boninite flow shown in photo B; note top of flow directly beneath scale card;* ***d)*** *polygonally jointed massive Daly rhyolite (subunit J3b);* ***e)*** *coarse intrusion breccia, Sneath intrusive complex (subunit J5g);* ***f)*** *polygonally jointed tonalite 30 m from chilled margin of Sneath intrusive complex (subunit J5a).*

are intercalated with lenses of monolithic breccia and pillow fragment breccia (subunits J1d, e), amoeboid breccia (Figure 6a) and heterolithic volcanic breccia (unit J4). Monolithic mafic breccia and pillow fragment breccia (subunits J1d, e) are only widespread north and south of Epp Lake, where they likely record an event of synvolcanic subsidence, slumping and breakup of volcanic flows from nearby fault escarpments. The upper 500 m of the unit, which is capped by a 2–4 m thick, base-metal–poor, sulphide-rich 'tuffite', consists of prominently silicified, albitized and epidotized flows (Bailes and Galley, 1996).

The Welch basalt is slightly to strongly depleted in rare earth elements (REE) and high field strength elements (HFSE; e.g., Zr, Y, Ti, Hf) relative to mid-ocean-ridge basalt (MORB). It displays higher MORB-normalized contents of Th relative to Nb, and Th/Nb ratios >0.1, features characteristic of subduction-related magmas formed within oceanic arc tectonic settings (Gill, 1981; Tarney et al., 1981). Low (La/Yb)n and

Th/Nb ratios, and low REE, Th, Ti, Zr, and Y contents are features that Stern et al. (1995b) have suggested may be due to derivation from a depleted, refractory mantle source that was residual after extraction of MORB or back-arc basin basalt.

Boninite and boninite-like flows (unit J2): This unit, which is exposed between Stroud Lake and the south end of Anderson Lake, is up to 500 m thick and consists of aphyric, high Ca-boninite (subunit J2a) and pyroxene (amphibole replaced)-plagioclase–phyric mafic flows (subunit J2b) that display boninite-like characteristics (e.g., higher MgO, Ni and Cr values than typical Welch basalt). Local crosslamination in interflow mafic lapilli tuff and imbrication of pillows (Figure 6b) in flows indicate that deposition was from the west (present co-ordinates). The high-Ca boninite, low-Ti arc tholeiite basalt and isotopically juvenile felsic rhyolite in the 'primitive arc' is a rock association that has been interpreted, from both modern (Crawford et al., 1981; Beccaluva and Serri, 1988) and Phanerozoic (Swinden, 1996; Lapierre et al., 1985) rocks, to have been produced by high-temperature, hydrous partial melting of refractory mantle sources in proto-arc environments during arc extension. According to Crawford et al. (1989) and Tatsumi and Maruyama (1989), boninitic and refractory magmas are a reflection of zones of extremely high heat flow because their formation requires high heat of fusion. This rock association has been linked with VMS deposits (Lapierre et al., 1985; Swinden, 1996), which may reflect a common genetic affiliation with zones of high heat flow, rifting and attendant increase in fluid circulation and geothermal activity (Bailes and Galley, 1999).

Rhyolite and rhyodacite (unit J3): The Anderson sequence contains four discrete rhyolite flow-dome complexes, known from west to east as the Daly, Sneath, Kontzie and Anderson–Stall. Individual flow domes are composed of quartz and quartz-plagioclase–phyric (subunit J3a) and aphyric to sparsely porphyritic (subunit J3b) rhyolite, with rare heterolithic breccia (subunit J3c) and recrystallized quartzo-feldspathic (subunit J3d) equivalents. Flows are typically massive and lobe-hyaloclastite (Figure 4b) types. The Anderson–Stall rhyolite complex is the largest, extending approximately 8 km in strike length and 2 km in thickness. The rhyolite complexes, which punctuate the ongoing Welch basalt volcanism, have been suggested by Stern et al. (1995b) and Bailes and Galley (1999) to possibly be a product of melting at the base of the lithosphere, with the associated basalt derived by hydrous melting of underlying refractory mantle.

Heterolithic volcanic breccia (unit J4): Volumetrically minor lenses of heterolithic volcanic breccia (subunits J4a–c) and mafic tuff and lapilli tuff (subunit J4d) occur sporadically within Welch basalt flows (unit J1). The breccia is mainly mafic (subunit J4a) in composition but may include a felsic component (subunit J4b) and, in some instances, is dominated by felsic detritus (subunit J4c). The largest lens of breccia is a 200–300 m thick unit of matrix-supported heterolithic mafic breccia (subunit J4a) that directly overlies boninite-like mafic flows (subunit J2b) and has a strike length of over 3 km. It is composed of a mixture of aphyric and plagioclase-phyric mafic clasts with rare aphyric felsic clasts. The breccia locally displays moderate alteration, with fragments bleached and silicified and the matrix overprinted by acicular amphibole porphyroblasts produced in the altered rocks during ca. 1.81 Ga almandine-amphibolite–facies regional metamorphism.

Anderson intrusive rocks

The Anderson sequence supracrustal rocks are cut by a number of different intrusions. Most of them are of uncertain age and genetic affinity, many of them likely being feeders to the younger Chisel and Snow Creek sequences, as well as post–1.84 Ga 'successor arc' magmatism. The Sneath intrusive complex (unit J5) is the only unit that is clearly synvolcanic to the Anderson sequence volcanism.

Sneath intrusive complex (unit J5): This semiconformable intrusive complex displays many of the features expected of a high-level synvolcanic intrusion: absence of a metamorphic halo, multiple phases, bodies of intrusion breccia (Figure 6c), textural variation, porphyritic character, columnar jointing (Figure 6f), internal alteration (including disseminated chalcopyrite), and over-printing by all the kinematic and metamorphic events recorded in the host Anderson sequence ('primitive arc') volcanic rocks. A synvolcanic age is further supported by a U-Pb zircon age of 1886 +17/–9 Ma for the quartz megacrystic tonalite (subunit J5b; Bailes et al., 1991) and 1892 ±4 Ma for the overlying Stroud breccia (unit J6; David et al., 1996). Fragments of the distinctive Sneath quartz megacrystic tonalite (subunits J5b, c) in overlying Anderson–Stall rhyolite (subunit J3c) and Stroud breccia (unit J6) are conclusive evidence that the intrusion is synvolcanic. The intrusive complex is composed of at least seven texturally distinct tonalite bodies that are separated by intrusive contacts. The largest individual body, located south of Anderson Lake, is 1.5 km by 5 km, semiconformable and composed of quartz-megacrystic tonalite (subunit J5b). From west to east, this body cuts up the Anderson sequence, at a shallow angle, through approximately 1 km of volcanic stratigraphy, to within 0.3 km of the top of Anderson sequence at its eastern end. A quartz-feldspar porphyry phase (subunit J5c), previously interpreted to be extrusive (Coats et al., 1970; Studer, 1982; Walford and Franklin, 1982; Zaleski, 1989), completely envelops the Rod VMS deposit and intrudes hangingwall portions of the Welch basalt (unit J1) and Anderson–Stall rhyolite (unit J3) east of Stall Lake. This suggests a range in age for the different phases of the Sneath intrusive complex, with some phases postdating massive sulphide deposition. The Sneath intrusive complex (unit J5) has similar trace, rare earth element and εNd values, as do spatially associated rhyolite (unit J3), further supporting the synvolcanic origin of the Sneath Lake intrusive complex.

Chisel sequence

The transition from the Anderson to the Chisel sequence is characterized by an abrupt change from a chemically bimodal sequence dominated by subaqueous mafic flows to a chemically diverse sequence dominated by units of heterolithic volcaniclastic detritus. It is also marked by distinctive changes in geochemical character of associated mafic flows. The Chisel mafic flows typically have higher Th/Nb, Th/Yb and La/Yb ratios, and lower εNd values (0 to +2.7) than do those in the underlying Anderson sequence (+3; Stern et al., 1992; Bailes and Galley, 1999). Stern et al. (1992) suggested that the Chisel magmas were likely affected by one or more of the following:

within-plate enrichment, derivation from a more fertile mantle source, lower average extents of melting at greater depths and contamination from older crustal fragments. Intracrustal contamination during Chisel magma genesis is supported by the presence of an inherited zircon population with ages of 2650–2824 Ma in the 1892 Ma Stroud breccia (unit J6; David et al., 1996).

Chisel sequence units (Figure 7) are typically thin (compared to the Anderson sequence) and discontinuous, display rapid lateral facies variations and include a large volume of volcaniclastic deposits. The high proportion of volcaniclastic rocks (40–50%) and local unconformities (Bailes and Simms, 1994) in the Chisel section suggest that considerable topographic relief was developed, possibly as a result of intra-arc

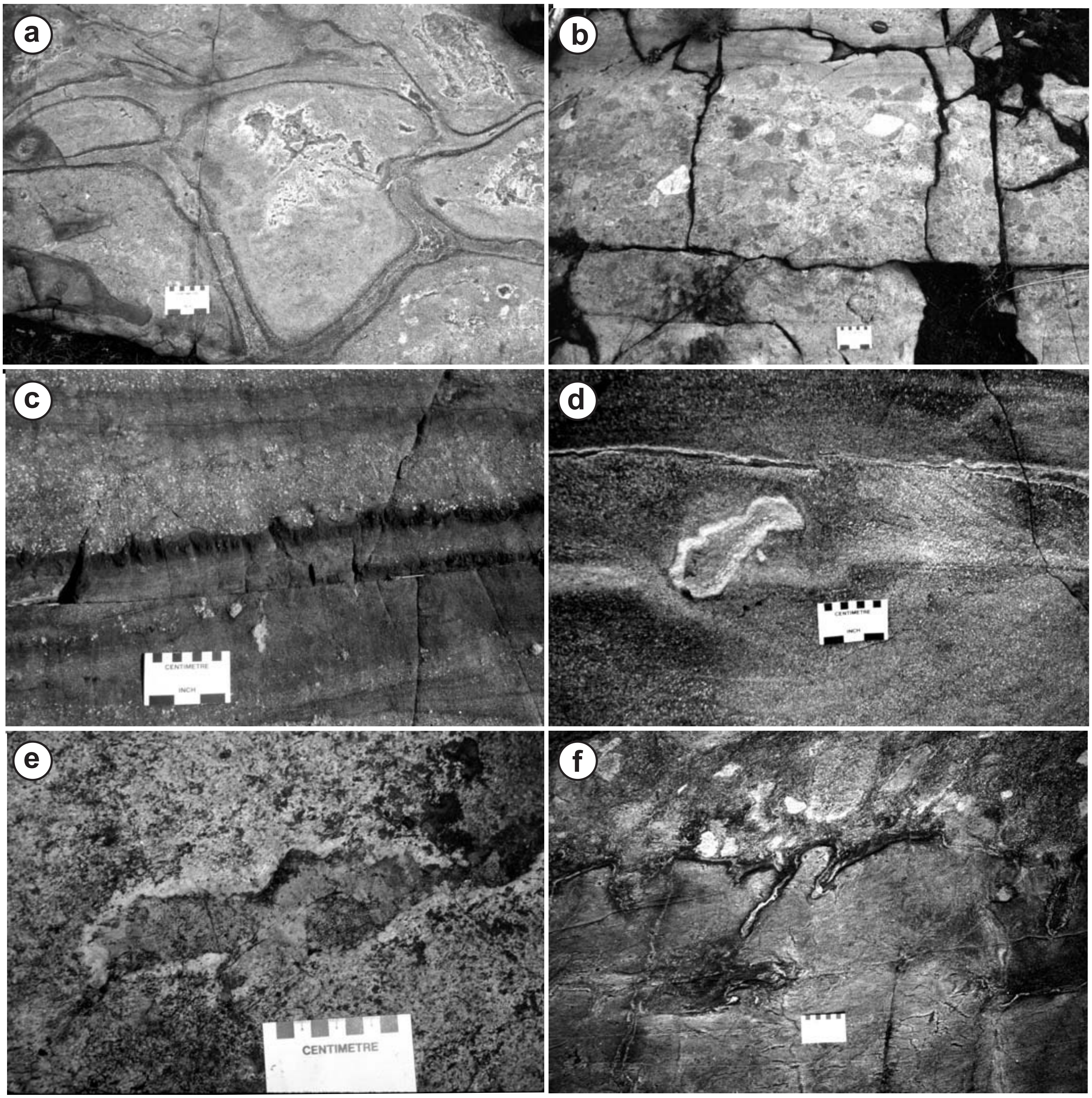

***Figure 7:** Chisel sequence ('mature arc') volcanic and intrusive rocks: **a)** pillowed, plagioclase-phyric Snell basalt with dark selvages, light interpillow hyaloclastite and irregular white domains of albitization/silicification (unit J7); **b)** heterolithic mafic breccia bed deposited by debris flow, showing reverse size-grading at base (scale card), normal size-grading in centre and a laminated mafic wacke top (lens cap), Edwards mafic volcaniclastic rocks (unit J8a); **c)** bed of light-coloured, normally size-graded mafic wacke underlain by dark-weathering mafic mudstone with delicate lode structures, Edwards mafic volcaniclastic rocks (unit J8b); **d)** potential bomb sag beneath block in Threehouse mafic tuff (unit J19a); note crossbedding on 'lee side' of clast; **e)** quartz-rimmed, epidote-filled miarolitic cavitiy in Richard Lake felsic intrusive complex (unit J18 a); **f)** amoeboid margin on synvolcanic mafic dike (unit J15) intruding Edward heterolithic mafic breccia (unit J8a); note dark-weathering chilled selvage.*

extension, rifting and caldera development.

The Chisel sequence, >3 km thick, is subdivided into lower (units J6–J18) and upper (units J19–J24) parts. The contact between these subdivisions coincides with a hiatus in volcanism during which chemical sedimentation occurred, accompanied by deposition of the Chisel, Lost, Ghost, North Chisel and, perhaps, Lalor Zn-Cu VMS deposits. Chisel sequence rocks of uncertain stratigraphic position are undivided with respect to the lower and upper parts, and have been combined in units J25–J33. These undivided units are most abundant west of the Kobar Lake Fault, where distinctive stratigraphic markers and a lower number of geochemical analyses do not permit unambiguous identification of units.

The Chisel sequence is cut by a wide variety of intrusive rocks, ranging from dikes less than a metre across to large plutons several kilometres in diameter. Most of these intrusions are demonstrably synvolcanic, and many can be clearly related to specific extrusive equivalents. Two groups of synvolcanic intrusions are recognized: one of these (units J14–J18) feeds Lower Chisel volcanism and the other (units J23, J24) is responsible for Upper Chisel volcanism. The abundant (locally up to 50% by volume) synvolcanic dikes support the hypothesis that rifting and caldera development accompanied accumulation of the Chisel sequence.

Lower Chisel rocks

Lower Chisel supracrustal rocks

The Lower Chisel supracrustal rocks contain two discrete components. The first consists of felsic and mafic volcaniclastic rocks (units J6, J8), which are derived by mass wasting of a source terrane located to the west (current co-ordinates), and a geochemically distinctive basalt flow (unit J7) that was derived from more mature and evolved arc magmatism than that responsible for the underlying Anderson sequence. The second component (units J9–J13) is a suite of differentiated basalt, andesite, dacite and rhyolite flows that are geochemically distinct. The Zn-Cu VMS deposits are spatially associated with rhyolite flow-dome complexes (unit J12) lying atop this latter succession. Ubiquitous pillows in the mafic flows, lobe and tongue facies in rhyolite flows, and graded bedding in mafic volcaniclastic rocks indicate that the Lower Chisel rocks were deposited subaqueously.

<u>Stroud breccia (unit J6):</u> This 0–400 m thick unit consists of well-bedded, heterolithic to monolithic felsic breccia, tuff and lapilli tuff (subunit J6a), containing beds of coarsely porphyritic felsic breccia (subunit J6b) up to 18 m thick, and locally intercalated with sulphide-bearing intermediate to mafic volcanic tuff and lapilli tuff (subunits J6c, d). Felsic breccia beds are organized in a manner consistent with deposition by subaqueous debris flows (Cas and Wright, 1987). Most of this unit lies directly upon Anderson sequence volcanic rocks, although rare beds of this felsic breccia occur intercalated with the Snell basalt (unit J7) higher in the section. Fragments from the felsic breccia are geochemically and texturally indistinguishable from Anderson sequence rhyolite (unit J3) and the Sneath intrusive complex tonalite (unit J5), suggesting an unconformable lower contact with, and derivation from, uplifted portions of the underlying Anderson sequence. Zircons in a sample taken from a 1 m thick felsic tuff and lapilli tuff bed yielded two zircon populations (David et al., 1996). One suite defined a discordia line with an upper intercept of 1892 ±3 Ma, interpreted to be the age of crystallization of felsic material forming the felsic tuff and lapilli tuff. The second population yielded inversely discordant Pb-Pb ages of 2652 and 2674 Ma, and Pb-Pb discordant ages of 2715, 2823 and 2691 Ma that David et al. (1996) interpreted as inherited.

<u>Snell basalt (unit J7):</u> Up to 500 m of porphyritic mafic flows (unit J7) overlie, and are locally intercalated with, the Stroud felsic volcaniclastic rocks (unit J6). These mafic flows are interpreted to have accumulated rapidly from a different magma source than that supplying the underlying basalt (unit J1) of the Anderson sequence. They display amongst the lowest ε_{Nd} values of any mafic flows in the Flin Flon Domain, a feature interpreted by Stern et al. (1992) to be a consequence of interaction with much older, possibly Archean, lithosphere. The presence of inherited Archean zircons in the Stroud breccia (David et al., 1996) is consistent with this interpretation. Intercalation of Snell mafic flows with Stroud felsic breccia suggests that magma extrusion occurred in an environment with topographic relief (caldron subsidence and/or rifting?) and contemporaneous erosion of portions of the previously deposited Anderson sequence felsic complexes.

<u>Edwards mafic volcaniclastic rocks (unit J8):</u> This unit, up to 500 m thick, coarsens upward and consists of fine-grained, thin-bedded, mafic volcaniclastic sedimentary rocks (subunit J8b) overlain by coarse, thick-bedded mafic breccia (subunit J8a). East of Edwards Lake, the heterolithic mafic breccia (subunit J8a) is separated from the underlying mafic volcanic tuff and lapilli tuff (subunit J8b) by a rapidly gradational boundary. The boundary is marked by an increase in heterogeneity of clast population, in addition to an abrupt increase in maximum clast dimensions (Skirrow, 1987). Organization of breccia beds is consistent with deposition by subaqueous debris flows (Bailes, 1987; Skirrow, 1987). A local source for the mafic breccia is indicated by the presence of very thick beds composed of boulder-sized clasts. A source from the west (current co-ordinates) is suggested by an increase in thickness of the unit and a coarsening of contained breccia to the west.

<u>Caboose andesite (unit J9):</u> This 0–300 m thick unit of andesite flows, which varies from massive to pillowed, contains 2–5% amphibole and biotite phenocrysts (2–10 mm) and up to 5% quartz amygdules (3–10 mm). Altered portions of the flows are overprinted by porphyroblasts of actinolite. The stratigraphically underlying volcanic rocks are intruded by hornblende-phyric diorite (unit J14) that is chemically and texturally similar to, and most likely represents subvolcanic feeders for, the Caboose andesite flows (unit J9).

<u>Moore basalt (unit J10):</u> This Fe-rich basalt unit, up to 1000 m thick, displays marked lateral variations in thickness and flow facies. Where thickest, 3 km east of Chisel Lake, it consists mainly of pillowed aphyric basalt (subunits J10a, b). South of Chisel Lake, it is less than 350 m thick and composed of a less proximal mixture of amoeboid pillow breccia (subunit J10c). Southwest and west of Chisel Lake, it is composed of pillow-fragment and heterolithic mafic breccia (subunits J10d, e). This east to west change in unit J10 is interpreted to reflect a transition from more vent-proximal to more distal downslope

equivalents. Moore pillowed flows are characterized by abundant vesicles, including many 2–3 cm long radial pipe vesicles. The Moore basalt (unit J10) is the first unit of a comagmatic, chemically distinctive, mafic to felsic cycle that is characterized by high incompatible element and elevated light REE abundances (Bailes and Galley, 1999). The remaining units in this cycle are the overlying Powderhouse dacite (unit J11) and the Chisel–Ghost–Photo rhyolite complexes (unit J12). Stern et al. (1995b) suggested that high HFSE and light REE abundances of the Moore basalt may indicate involvement of an enriched mantle source component in their generation. Plank and Langmuir (1988) and Woodhead and Johnson (1993) pointed out that incompatible trace element enrichments in arc volcanic rocks, such as those shown by the Moore basalt, may reflect eruption through thicker lithosphere as a consequence of lower average extents of melting due to the shorter length of the adiabatic melting column.

Powderhouse dacite (unit J11): A 100–250 m thick unit of dacite tuff and lapilli tuff overlies the Moore basalt and forms the stratigraphic footwall to the Ghost, Lost, Chisel and (part of) Chisel North Zn-rich massive sulphide deposits. Amygdaloidal dacite flows (subunit J11d) are also locally intercalated with the Moore basalt. A prominent, footwall dacite dike complex (unit J16) is chemically identical to, and a feeder system for, Powderhouse dacite volcanism. The Powderhouse dacite tuff and lapilli tuff (subunit J11a) are characterized by 5–15% tablet-shaped plagioclase phenocrysts (0.5–2 mm) and glomerocrysts and, more rarely, quartz phenocrysts. The dacite tuff is generally massive and featureless, only rarely displays fragments and is only occasionally well bedded. Heterolithic, coarse debris-flow breccia (subunit J11b), composed of mixed felsic and mafic detritus, forms a unit up to 30 m thick near the base of the Powderhouse dacite. Well-bedded dacite tuff and lapilli tuff (subunit J11c) form a unit up to 20 m thick at the top of the Powderhouse formation.

Chisel-Ghost-Photo-Cook rhyolite (unit J12): All of these rhyolite flow-dome complexes display La/Yb greater than 1.2 and contain spatially associated Zn-Cu VMS deposits. They vary considerably in size from small (e.g., Chisel flow-dome complex) to large (e.g., Photo and Cook complexes). The least deformed and metamorphosed (Chisel and Ghost) are clearly composed of massive rhyolite and lobe-hyaloclastite rhyolite flows (subunits J12a–c). Those exposed near the northern boundary of the map area (Cook and Photo) have been recrystallized during regional metamorphism to quartzofeldspathic gneiss (subunit J12d). The rhyolite complexes locally contain monolithic, probably autoclastic, fragmental rhyolite (subunit J11c) and flows with gas cavities and large quartz amygdules (subunit J12e). The latter, which are most abundant north of Ghost and Chisel lakes, commonly display weak to moderate Fe-Mg metasomatism. The rhyolite of unit J12 displays the elevated light REE pattern typical of the Chisel sequence, but its ε_{Nd} value (+3.3) is higher than those of the underlying Moore basalt (+2.5) and Powderhouse dacite (+2.6 to 3.05) but identical to that of the Richard tonalite (+3.35; Stern et al., 1992).

Balloch basalt and basaltic andesite (unit J13): Massive aphyric basalt and basaltic andesite flows (unit J13) form a southwest-facing unit, measuring 300–1000 m in width and over 7 km in length, that is truncated to the southeast by a fault. A 10 m wide massive flow directly overlying the Ghost Lake VMS deposit has also been included in unit J13, but its relationship to main unit is uncertain. Rare pillows and amoeboid breccia indicate that the Balloch Lake basalt was deposited subaqueously, but the presence of abundant large amygdules and local intercalation with scoria lapilli tuff suggest a moderately shallow-water environment. The mafic flows display subalkaline tholeiitic chemistry and the characteristic elevated light REE typical of Chisel sequence volcanism. The Balloch basalt is intercalated with rhyolite flows (subunit J12b).

Lower Chisel intrusive rocks

Lower Chisel intrusive rocks (units J14–J18) are particularly abundant stratigraphically below the massive-sulphide deposits of the Chisel area, where they locally form up to 50% by volume of the footwall rocks. The intrusions, which were possibly part of a vent facies or eruptive centre, include small plugs and stocks of hornblende-phyric diorite (unit J14), a suite of texturally variable gabbro and diorite dikes and plugs (unit J15), a voluminous suite of dacite dikes (unit J16), dikes and stocks of quartz and quartz-feldspar porphyry (unit J17) and a large subvolcanic pluton (unit J18). Dike margins of Lower Chisel intrusive rocks in volcaniclastic strata are commonly amoeboid (Figure 7f), consistent with intrusion into unconsolidated, water-saturated detritus. Altered rocks are spatially associated with many of these synvolcanic bodies, and many of the intrusions are themselves hydrothermally altered, indicating that they were likely involved in generating and sustaining hydrothermal fluid flow.

Hornblende-phyric diorite (unit J14): Small plugs of hornblende-phyric diorite (subunit J14a) cut both the Snell basalt (unit J7) and the Edwards mafic volcaniclastic rocks (unit J8), but are not found in younger volcanic rocks. The diorite is texturally and compositionally similar to the Caboose andesite (unit J9), and may be an intrusive equivalent. A 100 m by 300 m body of the hornblende-phyric diorite, 1.3 km west of Edwards Lake, is crosscut by dikes and sills of dacite (unit J16), and is also present as xenoliths in larger bodies of unit J16 dacite.

Gabbro, diorite and quartz diorite (unit J15): Dikes and small plugs of this unit are common south of Chisel Lake. They vary in grain size from fine to coarse, in texture from aphyric to porphyritic and in composition from gabbro to quartz diorite. Some of these intrusions are texturally similar to the younger mafic intrusions of unit J37 but are distinguished from the latter by being crosscut by unit J16 dacite dikes. Intrusions of unit J15 south of Chisel Lake can locally be demonstrated to be geochemically similar to, and the intrusive equivalent of, the Moore basalt (unit J10).

Dacite dike complex (unit J16): The Chisel Lake area is underlain by a dacite sill and dike swarm (unit J16) that is distinguished by its high incompatible and light REE abundances. The intrusions are texturally and chemically identical to the overlying Powderhouse dacite (unit J11) and are clearly part of a subvolcanic feeder system for this unit. The intrusions are restricted in occurrence to rocks stratigraphically below the Powderhouse dacite and are spatially related to a large zone of silicification and chlorite-amphibole alteration. Galley (2003) interpreted this dacite dike complex to be the first phase of the

Richard intrusive complex (unit J18).

Quartz porphyry, quartz-plagioclase porphyry (unit J17): Quartz porphyry (subunit J17a) and quartz-plagioclase porphyry (subunit J17b) vary in size from small dikes a few metres wide to stocks up to 400–500 m in diameter. South of Chisel Lake, many of these intrusions cut across dacite dikes (unit J16) and are, in turn, cut by the Richard Lake tonalite (unit J18). At Photo Lake, small stocks of unit J17 are truncated (unconformably?) at their contact with Upper Chisel mafic volcaniclastic rocks of unit J19. Most of these intrusions are interpreted to be subvolcanic feeders to overlying rhyolite flows of unit J12, and the similarity of their chemical analyses to those of overlying rhyolite flows supports this interpretation. However, some intrusions are younger, as they intrude Upper Chisel volcaniclastic rocks (e.g., unit J19) in the Anderson Lake area and unit J30 in the Morgan and Woosey lakes area.

Richard intrusive complex (unit J18): This suite of leucotonalite to mesotonalite forms a 1.7 km by 7.3 km intrusion that transects and structurally stitches together Anderson and Lower Chisel sequences. It cuts across dacite (unit J16) and quartz porphyry (unit J17) dikes as well as several zones of hydrothermally altered rocks. The synvolcanic nature of the Richard intrusive complex is supported by a U-Pb zircon date of 1889 +8/–6 Ma (Bailes et al., 1991) and a date of 1892 ±3 Ma (David et al., 1996) on the Stroud breccia (unit J6). Also consistent with its synvolcanic age are 1) the presence of miarolitic cavities (Figure 7e), which indicate shallow depth of emplacement (Galley, 2003); 2) geochemical similarity of the Richard tonalite and the overlying ore-hosting Chisel and Ghost rhyolite flows (unit J12; Bailes and Galley, 1996, 1999); 3) an increase in intensity of regional-scale semiconformable alteration zones in the Lower Chisel sequence close to the Richard intrusive complex (Bailes and Galley, 1996); and 4) internal vein-breccia systems, which suggest epizonal intrusion (Galley, 2003). The intrusion of the synvolcanic Richard intrusive complex across the boundary between the Anderson and Lower Chisel rocks is important because it demonstrates that the Anderson and Chisel sequences belong to a structurally coherent stratigraphic succession.

Upper Chisel rocks

Upper Chisel supracrustal rocks

The Lower Chisel flow-dome complexes (unit J12) at Chisel and Potten (Lost) lakes are directly overlain by 1–2 m of sulphide-bearing, cherty sedimentary rocks that grade upwards into approximately 50 m of well-bedded, mafic turbidite tuff and lapilli tuff (subunit J19a), and several hundred metres of poorly bedded, scoria-rich, mafic tuff-breccia and heterolithic mafic breccia (subunits J19b, c). Contacts between these units are structurally intact and rapidly gradational. The Threehouse mafic volcaniclastic rocks at Chisel Lake most likely accumulated in a topographic depression or basin, shed from an adjacent shallow-water to subaerial mafic edifice.

The simple hangingwall stratigraphy above the Chisel and Ghost flow-dome complexes breaks down at a regional scale and is replaced by, and intercalated with, a more complex hangingwall succession that includes a variety of mafic flows (unit J20), heterolithic breccia (unit J21) and rhyolite flows and breccia (unit J22). Bailes and Simms (1994) noted an angular relationship between Lower and Upper Chisel stratigraphy near Photo Lake and suggested that this may reflect an unconformable contact, with substantial portions of the Lower Chisel stratigraphy eroded prior to deposition of the Upper Chisel rocks.

Threehouse mafic volcaniclastic rocks (unit J19): This 250–350 m thick unit consists of a basal 30–100 m of well-bedded, fine-grained, mafic volcaniclastic rocks (subunit J19a) overlain by 250–300 m of nonbedded to crudely bedded, scoria-rich, mafic lapilli tuff and tuff-breccia (subunit J19b). The basal mafic volcaniclastic rocks (subunit J19a) display excellent graded bedding; scour channels; load structures; A, AB and ABE Bouma bed zonation; and local trough cross-bedding, pebble lags, scoria clasts and bomb sags (Figure 7d; Bailes 1990). They were clearly deposited from turbulent subaqueous density currents in a shallow-water to possibly subaerial environment (Galley and Bailes, 2002). Upward coarsening of the Threehouse mafic volcaniclastic rocks, which is recorded by a gradual increase in both grain size and bed thickness, and increasing heterogeneity of clast populations (unit 19c), indicate a likely increase in topographic relief and complexity of the source terrane. Rare preservation of scoria clasts, abundant prominently amygdaloidal basalt clasts in coarser beds, and limited compositional and textural range of detritus all suggest that the source may have been a volcanic construct, possibly in part erupted under shallow-water to subaerial conditions. Upward coarsening of the unit may reflect growth of this volcanic construct.

Threehouse basalt and basaltic andesite (unit J20): Pillowed basalt and basaltic andesite flows dominate the Upper Chisel strata northeast of Threehouse Lake. They include both plagioclase (subunit J20a)– and pyroxene-plagioclase (subunit J20b)–phyric flows that are texturally identical to clasts preserved in along-strike and intercalated breccia beds (subunits J19b, c). The flows display the elevated light REE pattern typical of the Chisel sequence, and the low Nb, Zr and Hf values on N-MORB–normalized incompatible element diagrams that characterize magmas derived by subduction processes (Sun and McDonough, 1989). They have lower overall contents of incompatible trace elements compared to highly fractionated Lower Chisel volcanic rocks (units J10, J11 and J12) directly underlying the VMS deposits in the Chisel area.

Heterolithic mafic volcanic breccia (unit J21): Northwest of Anderson Lake, the Threehouse mafic breccia (subunit J19b) is overlain by 200–300 m of crudely layered, thick-bedded, heterolithic mafic breccia (unit J21). The breccia contains scoriaceous mafic detritus similar to that contained in the underlying Threehouse formation (units J19 and J20), as well as abundant aphyric and porphyritic, nonvesicular to weakly vesicular basalt fragments. Also contained in the breccia are fragments of distinctive mafic dike rocks that are elsewhere observed to cut the underlying Threehouse breccia. In localities where the heterolithic mafic breccia (unit J21) contains abundant scoria fragments, its contact with the underlying Threehouse formation breccia (unit J19) is poorly constrained. A plagioclase-phyric mafic breccia (subunit J21a) with rare felsic fragments forms a discontinuous unit, up to 80 m thick, at the base of unit J21. To the northwest, the strongly heterolithic mafic breccia (unit J21) is rapidly transitional into a weakly heterolithic pyroxene- and pyroxene-plagioclase–phyric mafic breccia (subunit J21b). Subunit J21b locally contains tonalite clasts that are texturally similar to, and may record unroofing of, the Richard intrusive complex (unit J18). To the east, the heterolithic mafic breccia (unit J21) is intercalated with, and gradually replaced by, felsic volcanic rocks (unit J22), undivided mafic tuff and lapilli tuff (subunit J29c), and porphyritic mafic flows (unit J27). This is interpreted as an along-strike facies change, as no indications of reversals in bed topping directions or isoclinal folding have been observed. Southeast-facing, heterolithic mafic breccia (unit J21) between Bartlett and Berry Creek faults and on the south limb of the Anderson Bay F_{1-2} anticline is dominated by plagioclase-phyric mafic volcanic clasts with less common aphyric mafic volcanic clasts and rare aphyric and porphyritic felsic volcanic fragments. The lower third of this unit contains up to 10% quartz-megacrystic felsic clasts similar to the underlying felsic volcanic rock (subunit J21a); quartz-megacrystic tonalite clasts in this unit are indistinguishable from a phase in the Sneath intrusive complex (unit J5).

Rhyolite and rhyodacite (unit J22): A 200 m thick unit of felsic volcanic rocks, composed of porphyritic (subunit J20a), aphyric (subunit J20b) and local fragmental (subunit J20c) varieties, directly overlies the heterolithic mafic breccia (unit J21) north of Anderson Lake. To the east, this unit is more prominent and includes intercalations in the stratigraphically underlying heterolithic mafic breccia (unit J21) and the Threehouse mafic volcaniclastic rocks (unit J19). Rocks of unit J22 are typically massive and partly recrystallized to fine-grained felsic rocks. Local remnant quartz phenocrysts attest to the felsic volcanic origin of many of these recrystallized rocks. The felsic volcanic rocks of unit J22 trace to the northeast into quartzofeldspathic metamorphic rocks (subunit J22d), which contain ubiquitous fine-grained biotite and scattered garnet porphyroblasts.

Upper Chisel intrusive rocks

These intrusions include pyroxenite, melagabbro and gabbro (unit J23), and plagioclase- and plagioclase-pyroxene–phyric gabbro (unit J24). The textural, compositional and chemical variation in these intrusions is comparable to that displayed by the Threehouse pillowed basalt and basaltic andesite flows (unit J20). They are interpreted to be subvolcanic feeders for Upper Chisel volcanism. Consistent with this interpretation is their shallow emplacement, indicated by scattered (<<1%) 1–3 mm quartz amygdules, local prominent zones of quartz amygdules, and occasional pepperitic and amoeboid contacts with Threehouse mafic volcaniclastic rocks (unit J19).

Pyroxenite, melagabbro and gabbro (unit J23): Pyroxene-rich mafic dikes are common throughout the Chisel sequence. Large, irregular, mappable bodies are most common in Threehouse mafic volcaniclastic rocks (unit J19), where they likely represent shallow subvolcanic magma chambers that fed the overlying volcanic rocks. A large, 100 m wide by >600 m long sill in Lower Chisel rocks, 2 km southwest of Chisel Lake, is fractionated and zoned from a basal pyroxenite to a melagabbro centre (subunit J23a) and a gabbro top (subunit J24b). Intrusions of unit J23 into Lower Chisel rocks (units J6–J18) cut across altered varieties of these units, thus supporting the synvolcanic nature of the alteration.

Plagioclase- and plagioclase-pyroxene–phyric gabbro and diorite (unit J24): Texturally distinctive plagioclase- and plagioclase-pyroxene–porphyritic gabbro (unit J24) forms highly irregular plugs and stocks east and north of Chisel Lake and northwest of Photo Lake. Where these intrusions are emplaced into Lower Chisel rocks, they typically occur as small stocks and narrow crosscutting dikes. In contrast, they typically 'inflate' into large, irregular, sill-like bodies up to 100 m thick and 1.5 km long in the overlying, and apparently unconsolidated, Threehouse mafic volcaniclastic rocks (unit J19).

Undivided Lower and Upper Chisel volcanic rocks

Undivided supracrustal rocks (units J25–J33) occur west of the Kobar Lake Fault, west and north of Richards Lake, in the vicinity of Kormans Lake and along the north shore of Wekusko Lake. Direct correlation of these units with their counterparts in the central part of the area is hindered by a combination of complex faulting and stratigraphic successions that are substantially different from those in the type areas, likely due to along-strike facies changes in units. This, combined with a paucity of geochemical analyses of volcanic rocks in units J25–J33, precludes a meaningful correlation with those units (J1–J24) in the better understood, central part of the map area.

Aphyric mafic flows (unit J25): These mafic flows occur mainly in a 500 m thick succession southwest of Kobar Lake and a 100–200 m thick by 5 km long, northeast-trending sequence that follows the north shore of Wekusko Lake adjacent to the Berry Creek Fault. Other small domains of this unit outcrop on Morgan Lake and south of Cook Lake. Unit J25 and subunit J25a are typically pillowed to massive flows. The massive mafic rock exposed along the north shore of Wekusko Lake is fine grained, aphyric and locally gabbroic textured. Local quartz amygdules and amoeboid breccia in unit J25 on the north shore of Wekusko Lake indicate an extrusive origin there.

Plagioclase-phyric mafic flows (unit J26): Plagioclase-phyric pillowed mafic flows (unit 26) form a 400 m wide, east-facing succession northeast of Richard Lake. They are intercalated with mafic volcanic tuff and lapilli tuff (subunit J29c), and mixed mafic and felsic heterolithic volcanic breccia (unit 30a), overlain by dacite and rhyodacite tuff (unit 31). Together, these units are most likely equivalent to the Lower Chisel strata.

Pyroxene- and pyroxene-plagioclase–phyric mafic flows (unit J27): A north-facing succession of massive and pillowed pyroxene- and pyroxene-plagioclase–phyric mafic flows (unit J27), up to 1 km wide, is located north of Morgan Lake. These flows are of uncertain stratigraphic affinity and could belong to either the Anderson or Chisel sequences. They are most similar in appearance to Chisel mafic flows, but a single analyzed sample from this unit displays the distinctive trace-element and rare-earth-element geochemistry of volcanic rocks belonging to the Anderson sequence. Narrow units of pyroxene- and pyroxene-plagioclase–phyric flows are also located on Morgan Lake, where they are intercalated with Chisel sequence mafic volcaniclastic rocks.

Fine-grained mafic gneiss (unit J28): Thin (20–200 m) units of fine-grained amphibolite (unit J28) occur within the strongly recrystallized felsic volcanic rocks (subunit J12d) west of Cook Lake. The local presence in them of quartz amygdules suggests that they are mainly derived from mafic flows, but some could also be derived from fine-grained mafic sills and dikes.

Mafic monolithic to weakly heterolithic volcanic breccia, and mafic tuff and lapilli tuff (unit J29): Mafic volcaniclastic rocks (unit J29 and subunits J29a–c) occur widely throughout the map area. Where these units bear uncertain stratigraphic relationship to the better known parts of the Anderson and Chisel sequences, they are placed in unit J29.

Mixed mafic and felsic heterolithic volcanic breccia (unit J30): Breccia of uncertain stratigraphic affinity and composed mainly of mafic (subunit J30a) and, less commonly, felsic (subunit J30b) volcanic clasts occurs widely throughout the map area. Mafic breccia (subunit J30a) is typically matrix supported, crudely bedded and similar in many respects to those already described (unit J4, subunit J8a, unit J21). Mafic breccia units (subunit J30a) exposed north of Morgan Lake, along Parisian Creek and on Woosey and Morgan lakes are associated with base-metal sulphide deposits (Penn and Morgan VMS deposits) that are most similar in metal content and rock association (mainly volcaniclastic rocks) to deposits in the Chisel sequence. Mafic breccia (unit 30a) northwest of Richard Lake may be equivalent to the Edwards mafic breccia (subunit J8a). It occupies a similar stratigraphic position relative to the underlying Anderson sequence but on opposite limbs of an $F_{1\text{-}2}$ fold structure cored by the Richard intrusive complex (unit J18). Mafic breccia (unit 30a) and tuff and lapilli tuff (unit 30c) north of Chisel Lake are tentatively correlated with Lower Chisel strata, whereas those northeast of Stall Lake correlate with Upper Chisel strata. Altered breccia along Parisian Creek is tentatively interpreted as a felsic heterolithic volcaniclastic rock (subunit J30b) belonging to the Chisel sequence.

Dacite and rhyodacite tuff, lapilli tuff (unit J31): A plagioclase-phyric felsic tuff and lapilli tuff (unit J31) forms a 1 km long, 200 m wide, west-facing unit west of Richards Lake. It is truncated to the west and northwest, respectively, by the Kobar Lake Fault and a splay from this fault. Texturally and geochemically, this unit is similar to the Powderhouse dacite (unit J11). It also occurs at approximately the same stratigraphic position as the Powderhouse dacite should, but in a much more complex and poorly understood portion of the Chisel sequence.

Fine-grained quartzofeldspathic metamorphic rocks (unit J32): Quartzofeldspathic biotite and local garnet-porphyroblastic leucocratic schist and gneiss (unit J32) occur in the Kormans and Cook lakes area. They are clearly derived from felsic volcanic rocks, as they are along strike from felsic volcanic rocks and locally contain remnant quartz phenocrysts and felsic fragments.

Paraconglomerate, minor greywacke and mudstone (unit J33): This framework-supported, unsorted volcanic conglomerate forms a unit, up to 600 m wide, that is composed mainly of subangular to rounded felsic volcanic detritus with minor granophyric tonalite. North of Woosey Lake, it is overprinted by a metamorphically recrystallized synvolcanic alteration zone. Outside the map area, on central Woosey Lake, unit J33 is intercalated with felsic volcanic breccia and, within the map area on Woosey and Morgan lakes, it is cut by fine-grained, high-level, probable subvolcanic intrusions (units J17 and J37). Bailes (1980a) suggested that this unit was likely deposited by debris flows but interpreted the deposits to be part of the Burntwood Group. This is now known to be unlikely, as the Burntwood Group rocks postdate volcanism and associated synvolcanic hydrothermal alteration by approximately 50 million years (David et al., 1996).

Snow Creek sequence

The Chisel sequence is overlain by >0.5 km of aphyric, massive to pillowed basalt flows (unit J34) with the MORB-like geochemical characteristics of ocean-floor–back-arc basalt (Bailes and Galley, 1996, 1999). This basalt has an unfaulted basal contact and is truncated upsection by the D_1 Snow Lake (thrust?) Fault. The unfaulted basal contact is consistent with their being a product of intra-arc rifting rather than being structurally emplaced. A similar basalt (unit J35), which is deposited upon volcaniclastic rocks, is exposed on Tramping Lake.

Snow Creek basalt (unit J34): This basalt is a distinctive, uniform, grey-green weathering colour and characterized by pillows with narrow selvages, thin domains of interpillow hyaloclastite and virtual absence of vesicles. Massive divisions of flows tend to be gabbroic textured and, in thicker flows, are virtually indistinguishable from associated bodies of contained gabbro (unit J36). With the exception of a small area at the southwest corner of Snow Lake, the basalt displays no indication of alteration, including an absence of epidotization. The basalt is metamorphically recrystallized to mineral assemblages of the lower almandine-amphibolite facies.

Tramping Lake basalt (unit J35): This 200 m wide unit is geochemically indistinguishable from the Snow Creek basalt (unit J34). It consists of pillowed to massive, aphyric and rarely pyroxene- and plagioclase-phyric basalt (subunit J35a) that is locally intercalated with pyritic and graphitic mudstone (subunit J35b). The basalt typically weathers a uniform buff-green colour and is composed of lower greenschist-facies mineral assemblages. Locally, it is strongly silicified and overprinted by disseminated pyrite.

Undivided intrusive rocks

Gabbro, diorite and quartz diorite (unit J36): A wide assortment of dikes, sills, plugs and stocks of gabbro, diorite and quartz diorite (unit J36) cuts across volcanic rocks of the Snow Lake arc assemblage. The age of these intrusions is generally unknown, although they are not present in, and must therefore predate, Burntwood Group metasedimentary rocks (unit B1). Some of the intrusions are likely feeders to the Anderson and Chisel sequences, but some are younger and may be feeders for 'successor-arc' volcanic rocks. Intrusions of unit J36 north of Chisel Lake cut across Upper Chisel mafic feeder dikes (units J23 and J24), but are themselves cut by the Chisel Lake intrusion (unit P1).

Leucotonalite and quartz diorite (unit J37): Fine-grained, equigranular (subunit J37a) and quartz-phyric (subunit J38b) leucotonalite and quartz diorite outcrop most abundantly on Morgan and Woosey lakes, where they cut volcanic rocks (units J30 and J33) of probable Chisel sequence affinity. They are, in turn, truncated by the Woosey (unit P4) and Ham (unit P6) plutons. They cut across synvolcanic hydrothermal alteration zones and, for this reason, are considered to most likely postdate the Chisel sequence and to be products of early 'successor-arc' magmatism.

Burntwood Group

Sedimentological (Bailes, 1980a, 1980b), stratigraphic (Zwanzig, 1990) and geochronological (David et al., 1996; Machado et al., 1999) constraints on least recrystallized and deformed Burntwood Group rocks suggest that deposition was at about 1.855–1.84 Ga by prograding submarine fans fed by braided river systems draining from adjacent elevated land masses that exposed oceanic volcanic rocks as well as 'successor-arc' plutons and, locally, active 'successor-arc' volcanoes. Horizontal shortening accompanying 1.84–1.81 Ga continental collision tectonics has juxtaposed Burntwood Group sedimentary rocks against earlier-formed assemblages. As a result, contacts of Burntwood Group sedimentary rocks with the older Snow Lake arc assemblage are typically thrust faults (Connors et al., 1999).

Greywacke, siltstone, mudstone and derived paragneiss (unit B1): The Burntwood Group rocks outcropping on Wekusko Lake comprise greywacke, siltstone and mudstone (subunit B1a), with bedforms and sedimentary structures indicating deposition by turbidity currents in a submarine fan environment. They contain well-preserved primary features, such as graded bedding, flame structures, sandstone dikes, synsedimentary folds, scour structures and internal cyclic Bouma bed zonation. Those exposed at Kormans and Snow lakes retain some primary sedimentary structures but are recrystallized to garnet–biotite±staurolite paragneiss (subunit B1b) during a 1.815 Ga (David et al., 1996) episode of regional metamorphism that is centred in the Kisseynew Domain to the north.

Late successor-arc intrusive rocks

This magmatic episode includes a layered ultramafic intrusion (unit P1) of uncertain age and a series of granitoid plutons (units P2–P6) known to be 1.84–1.83 Ga in age. The granitic plutons, which are localized along the south margin of the Kisseynew Domain, intrude volcanic rocks of the eastern Flin Flon Domain and stitch together structurally imbricated rocks of the Snow Lake arc assemblage (units J1–J36) and Burntwood Group (unit B1).

Chisel layered ultramafic intrusion (unit P1): This 1.8 km by 9.8 km, north-trending, canoe-shaped, internally zoned ultramafic-mafic body is composed of six units ranging from peridotite to gabbro in composition. They include a marginal gabbro (not shown on map), up to 12 m wide; a peridotite (subunit P1a), up to 1200 m thick; a pyroxenite (subunit P1b), up to 50 m thick; a well-layered lower gabbro (subunit P1c), up to 540 m thick; a massive, homogeneous middle gabbro (subunit P1d), up to 240 m thick; a texturally variable upper gabbro (subunit P1e), up to 125 m thick; and a zoned upper contact (subunit P1f), up to 70 m thick, composed of gabbro, pyroxenite and a 2–3 m thick upper chilled margin (Young, pers. comm., 1994; Bailes et al., 1996). The Chisel Lake pluton truncates the isoclinally F_{1-2} folded, main ore lens of the Chisel Lake massive sulphide deposit. Young (pers. comm., 1994) interpreted the pluton to be late tectonic and emplaced after the culmination of F_3 deformation and during the waning stages of the amphibolite-facies regional metamorphism. The F_1 folds formed ca. 1.84 Ga (Connors, 1996; David et al., 1996) and regional metamorphism at Snow Lake occurred ca. 1.81 Ga (David et al., 1996), indicating that the intrusion postdates the ca. 1.89 Ga volcanism by at least 50 Ma and possibly as much as 80 Ma.

Bujarski pluton (unit P2): This oval pluton, 3.5–4 km in diameter, is one of several large, post–1.84, Ga calcalkaline quartz diorite to granite plutons intruding the volcanic and sedimentary rocks of the Chisel–Anderson lakes area. The Bujarski pluton is composed of medium- to coarse-grained hornblende tonalite (unit P2) and has a U-Pb zircon crystallization age of 1836 +4/–4 Ma (Bailes et al., 1991). An agmatitic outer margin to the pluton contains numerous variably digested xenoliths of the host mafic volcanic rocks. It grades rapidly inwards to a homogeneous equigranular phase that contains up to 10% small xenoliths of the host volcanic rocks. The northeastern margin of the pluton is truncated by a west-northwest–trending fault.

Epp-Morgan intrusive complex (unit P3): This unit forms three small plutons in the southeastern corner of the map area. The intrusions are zoned, with granodiorite cores (subunit P3a) and tonalite, quartz diorite and gabbro margins (subunits P3b–e). The largest intrusion, east of Woosey Lake, is approximately 1 km in diameter. It is characterized by a complex outer margin composed of hornblende-phyric tonalite dike swarms and associated alteration (subunit P3f), intrusion breccia (parts of subunit P3e) and agmatitic and xenolith-rich domains (subunit P3c). The southwestern margin of the pluton is a truncated by the same fault that terminates the northern margin of the Bujarski pluton.

Woosey pluton (unit P4): This northeast-trending, 2 km by 7 km pluton is located north of Woosey Lake and occurs in two segments, separated by a northwest-trending fault. The pluton, which is dominated by foliated, medium-grained, biotite-hornblende granodiorite and tonalite (subunit P4a), has an agmatitic leucotonalite margin (subunit P4b) up to 200 m wide. The pluton contains local zones of chlorite+garnet+biotite–rich rocks that elsewhere occur only in synvolcanic plutons (e.g., units J5 and J18).

Wekusko and Tramping plutons (unit P5): Composed of massive, unfoliated, medium- to coarse-grained, pink-weathering granite and granodiorite (unit P5), these one-mica intrusions with minor primary hornblende, largely altered to biotite, have U-Pb zircon crystallization ages of 1834 +8/–6 Ma (Gordon et al., 1990) and 1837 +8/–6 Ma (David et al., 1996). They are emplaced into isoclinally F_{1-2} folded, Burntwood Group greywacke, siltstone and mudstone (unit B1), with the Wekusko pluton occupying the core of a later, open, northeast-trending F_3 antiform. Similar to other unit P plutons in the Chisel–Anderson lakes area, the Wekusko pluton structurally stitches together volcanic and sedimentary assemblages that were previously juxtaposed during D_{1-2} deformation. The northern margin of the Tramping pluton is truncated by the Berry Creek Fault.

Ham pluton (unit P6): This 16 km by 7 km, north-north-east-trending, massive granodiorite to tonalite intrusion has a

U-Pb zircon crystallization age of 1830 +27/–19 Ma (Gordon et al., 1990). It predates regional metamorphism (Bailes, 1980a) and postdates F_{1-2} folding and movement on the Loonhead Lake Fault (Connors and Ansdell, 1994; Connors, 1996), a major thrust fault that is interpreted to have tectonically transported rocks from the Kisseynew Domain over those of the Snow Lake segment of the Flin Flon Domain. The southeastern margin of the pluton contains abundant xenoliths of foliated country rocks, including some from the adjacent Woosey pluton. The northern portion of the pluton is more foliated than southern portions and contains almandine-amphibolite–facies mineral assemblages formed during ca. 1.81 Ga regional metamorphism.

Late tectonic rocks

Tectonites (units W1 and W2) associated with the Berry Creek Fault zone, and mafic to intermediate intrusions (unit L1) postdate the late successor-arc plutons. The Berry Creek Fault zone truncates the northwestern margin of the Tramping pluton, and unit L1 intrusions typically form straight-walled, narrow dikes that cut Burntwood Group turbidites (unit B1) and calc-alkaline quartz diorite to granite plutons (unit P2 to P6).

Tectonites (unit W1 and W2): Felsic to mafic tectonite are part of the Berry Creek Fault, a regional structure that can be traced for more than 100 km to the southwest and 25 km to the northeast of the Chisel–Anderson lakes map area. Movement on the Berry Creek Fault is both pre- and postmetamorphic, and includes both ductile and brittle deformation (Krause and Williams, 1993). The late brittle deformation on the fault is clearly recognized by associated concentrations of hematized, carbonate-filled fractures and by offset of the metamorphic isograds formed during the ca. 1.81 Ga regional metamorphism. Offset during the brittle faulting is sinistral and dominantly strike slip. Premetamorphic and, typically, ductile offset is less easily defined due to overprinting by fault-associated alteration and by fabrics and textures produced during subsequent regional metamorphism. Tectonites of units W1 and W2 are characterized by tectonic lamination, abundant boudin structures, dismembered quartz veins and overgrowth by actinolite porphyroblasts generated during regional metamorphism. A stretching lineation in a mylonitic foliation formed in the Tramping Lake pluton adjacent to the Berry Creek Fault plunges moderately to steeply towards the northeast to north-northeast (Krause and Williams, 1993). According to Krause and Williams (1993), the stretching lineations and microfabrics, such as S/C fabrics and shear bands, indicate sinistral south-side-down movement on the Berry Creek Fault.

Mafic to intermediate dikes and intrusions (unit L1): Together, Burntwood Group turbidite (unit B1) and calcalkaline quartz diorite to granite plutons (units P2–P6) postdate most magmatic activity, but on Wekusko Lake they are locally cut by small stocks and irregularly shaped mafic to intermediate intrusions (subunits L1a, b) and a swarm of east-southeast–trending diabase dikes (subunit L1b). Gabbro and pyroxenite (subunit L1c) exposed on Rice Island contain associated nickel sulphide mineralization. Dikes of subunit L1b are straight walled, have chilled margins and are up to 10 m wide. They are characterized by tabular plagioclase phenocrysts (1–3 cm long) with or without hornblende pseudomorphs after pyroxene. The dikes cut across and are not deformed by F_{1-2} and F_3 folds.

Quartz-feldspar porphyry (unit L2): A small felsic porphyritic intrusion appears to intrude tectonite (unit W1) of the Berry Creek Fault west of Tramping Lake. The unit is exposed in an isolated outcrop within the projected trace of the Berry Creek Fault but its relationship to the tectonite is uncertain, as no contacts with the tectonite are exposed.

Structural history

Numerous detailed structural studies carried out in the Flin Flon Domain have identified deformation events spanning early accretion (1.88–1.87 Ga) to late tectonic continental collisions (1.83–1.77 Ga; Syme, 1995; Connors, 1996; Lucas et al., 1996; Ansdell et al., 1999; Connors et al., 1999). Deformation events that would have accompanied the pre–1.85 Ga tectonic accretion and amalgamation have yet to be recognized in the Chisel–Anderson lakes map area, probably due overprinting by subsequent deformation and metamorphism. The oldest recognized structures (D_1) in the area, which are best developed in such supracrustal units as the Burntwood Group sedimentary rocks (unit W1), are ≤1.84 Ga in age. Four deformation events have been recognized in the Chisel–Anderson lakes area (D_1–D_4; Krause and Williams, 1999). The first two (D_1 and D_2) are interpreted to result from the southward tectonic transport of Burntwood Group greywacke turbidite from the Kisseynew Domain over the Flin Flon Domain (Krause and Williams, 1999). Both the D_1 and D_2 events are characterized by tight isoclinal folding (e.g., west shore of Wekusko Lake and Anderson Lake anticline), and associated low-angle 'thrust' faults (e.g., McLeod Road Fault). The D_3 event is interpreted to result from northwest-southeast transpressional shortening that accompanied syn– to post–peak regional metamorphism (Connors et al., 1999). It is largely manifested by upright, open to closed F_3 folds (e.g., Threehouse Lake synform) and associated, steeply dipping faults (e.g., Berry Creek Fault). The D_4 structures, which are rare in the Chisel–Anderson lakes map area, are east-trending open upright folds that overprint F_3 folds (Connors, 1996; Krause and Williams, 1999).

A structural study (Zwanzig, 1999), involving the Burntwood and Missi groups at the margin of the Kisseynew Domain but extending into the northwestern part of the Snow Lake area, has yielded a more complex history than the studies closer to Snow Lake. The work in the northwest has recognized an early premetamorphic stage of D_1 high-level (brittle) thrusting of the Amisk collage and possibly the Snow Lake assemblage toward the northeast or east over the sedimentary rocks in the Kisseynew Domain. This is based on presently inverted thrust sheets that originally placed early volcanic rocks over the Burntwood Group. In that scenario, the D_1 and D_2 southward tectonic transport represents a period of extensive ductile backfolding, overturning and tightening of the earliest structures, and southwest back-thrusting along the major faults (e.g., McLeod Road Fault).

The main D_1 and D_2 minor structures and fabrics (folds, lineations, foliations related to southerly transport) are plotted together in the Chisel–Anderson lakes area, as they are difficult to distinguish from one another at the outcrop scale. The early brittle faulting, if present in the Snow Lake area, has left little or no fabric imprint. Both D_1 and D_2 deformation formed tight isoclinal folds, which are ubiquitous in the 1.84 Ga

Burntwood Group turbidite (unit B_1), and associated thrust faults. A prominent stretching lineation or shape fabric, commonly manifested as clast elongation or mineral lineation, is interpreted to be a product of combined D_{1-2} deformation. Although it is commonly at an angle to F_{1-2} fold axes, it is interpreted to be a product of ductile transport during D_{1-2} deformation, which formed tongue-shaped (sheath?) noncylindrical folds. According to Zwanzig (1995, 1999), this occurred after tectonic burial in the earlier fold-thrust system. The stretching lineations are commonly subparallel to F_3 fold axes, possibly reflecting influence on the orientation of D_3 fold structures by anisotropy generated by the previous D_{1-2} deformation events in the Chisel–Anderson lakes area.

Both D_1 and D_2 faults typically define contacts between lithotectonic assemblages and are curvilinear due to subsequent folding, partly during D_3 deformation. The Snow Lake assemblage cores a large, map-scale allochthon that is expressed in the Chisel–Anderson lakes map area as a doubly plunging F_1 antiform (Anderson Bay anticline). The F_1 folds and D_1 thrust faults are bracketed in age by the crosscutting 1.84–1.83 Ga (David et al., 1996) granitic plutons (units P2–P6) and the youngest 1.84 Ga detrital zircon (Machado et al., 1999) in the Burntwood Group greywacke (unit B1).

Fold structures belonging to D_2 were originally identified by Krause and Williams (1999) at Snow Lake, but can also be demonstrated to occur locally in Burntwood Group sedimentary rocks on Wekusko Lake. Both Connors et al. (1999) and Krause and Williams (1999) indicated that D_2 deformation, similar to the earlier D_1 deformation, involves low-angle 'thrust' faults.

Deformation structures belonging to D_3 are ubiquitous in the eastern Flin Flon Domain. In the Chisel–Anderson lakes map area, they are mainly manifested as east-northeast–trending, open F_3 folds (e.g., Threehouse synform), which are syn– to post–peak regional metamorphism (ca. 1.81 Ga; David et al., 1996). Other manifestations of D_3 deformation are S_3 foliations, which are axial planar to the folds. The crescentic configuration of the Mcleod Road Fault and F_{1-2} folds in the Snow Lake assemblage juvenile arc is a product of broad folding during D_3 deformation.

Evidence of pre–peak metamorphic D_3 folding at File Lake (File Lake synform cut by undeformed metamorphic isograds; Bailes, 1980a) to late-metamorphic shearing (e.g., Berry Creek Fault on Wekusko Lake) indicate that D_3 represents protracted northwest compression followed by late sinistral transpression, block uplift and metamorphic retrogression.

Alteration

The volcanic sequences of the Snow Lake arc assemblage contain prominent zones of alteration that visibly affect approximately 25% of the volcanic strata and associated synvolcanic intrusions (Figure 8). These alteration zones, which were formed by synvolcanic hydrothermal activity, were subsequently recrystallized during 1.81 Ga regional metamorphism and can be readily recognized by their unique metamorphic mineral assemblages (Figures 9 and 10). Centimetre-scale, euhedral crystals of chlorite, phlogopitic biotite, amphibole, muscovite, garnet and staurolite are common within metamorphically recrystallized semiconformable alteration zones. Recrystallized discordant alteration zones also contain coarse-grained kyanite and andalusite. The alteration zones are depicted on the accompanying map by black-line patterns over the (in some instances interpreted) original lithological units. Three periods of robust hydrothermal activity are identified within the Snow Lake arc assemblage (Bailes and Galley, 1999).

The first, and oldest, hydrothermal event (Figures 9a, b, e–g) is focused within the Anderson–Stall and Daly felsic extrusive complexes (unit J3), and is interpreted to be genetically related to formation of the Anderson, Stall and Rod Cu-Zn VMS deposits, in the former complex, and the Raindrop and Pot Lake base metal occurrences, in the latter (Figure 8). This hydrothermal activity also affected the underlying Sneath intrusive complex (unit J5), where it is manifested, most typically, as 'fracture-controlled' zones of alteration. These are interpreted to result from 'collapse' of the hydrothermal system into the subvolcanic intrusion as it cooled. The alteration caused by this hydrothermal activity includes both stacked, semiconformable zones, 3–5 km long and as much as 1000 m in total thickness, and discordant alteration 'pipes' that commonly terminate upsection at known VMS deposits and occurrences. These alteration zones are truncated by a late phase of the Sneath intrusive complex, which is considered evidence of the coeval nature of the volcanism, alteration and plutonism.

The second hydrothermal event took place at the end of Anderson sequence volcanism. This hydrothermal activity resulted in silica and epidote addition to Welch basalt (unit J1) at the top of the Anderson sequence (Figure 9c–d). A 300–500 m thick zone of alteration was produced directly underlying the 'Foot-Mud tuff-exhalite' at the top of the Anderson sequence. Both the alteration and the Foot-Mud exhalite have the same strike length as the underlying Sneath subvolcanic intrusion (unit J5).

The third hydrothermal event is located within Lower Chisel volcanic rocks (Figure 10) and is spatially associated with Lower Chisel intrusive rocks, in particular dikes of the 'Powderhouse' dacite (unit J16) and early phases of the Richard intrusive complex (unit J18). Resultant alteration by the hydrothermal fluids, which is largely confined to the Edwards mafic volcaniclastic rocks, is interpreted to be synvolcanic and related to generation of VMS deposits, as it only affects strata underlying the Chisel–Chisel North–Ghost–Lost VMS horizon (units J6–J13). The alteration zone is truncated to the west by late phases of the Richard intrusive complex (unit J18) and by late faulting. An alteration system underlying the Cook Lake and Bomber VMS occurrences trends to the southwest into a large alteration system along Parisian Creek, Woosey Lake and Morgan Lake; it may represent the western extension of the 'Chisel footwall' alteration system.

Economic geology

The eight producing and past-producing VMS mines in the Snow Lake arc assemblage of the Chisel–Anderson lakes map area occur in an approximately 6 km thick section that records, in its stratigraphy and geochemistry, a temporal evolution in geodynamic setting from a 'primitive arc' (Anderson sequence), to a 'mature arc' (Chisel sequence) and then to an 'arc-rift' (Snow Creek sequence). The 'primitive arc' contained three Cu-rich VMS mines (Anderson, Stall, Rod) and the 'mature arc' contained four Zn-rich VMS (Chisel, Lost, Ghost,

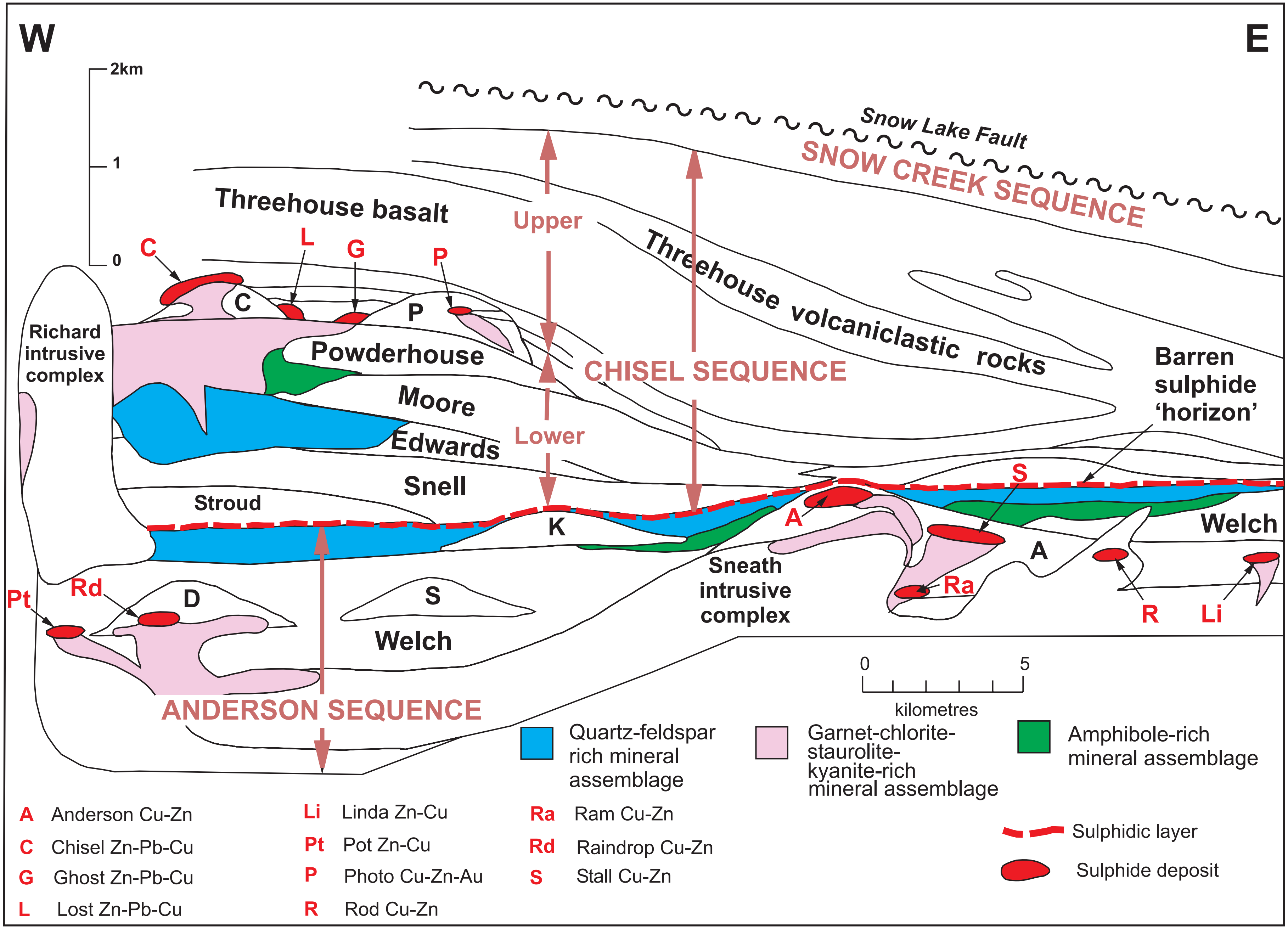

Figure 8: *Schematic cross-section of regional-scale alteration zones affecting rocks of the eastern two-thirds of the Snow Lake arc assemblage. Units and notations are identical to those used on Figures 4 and 5.*

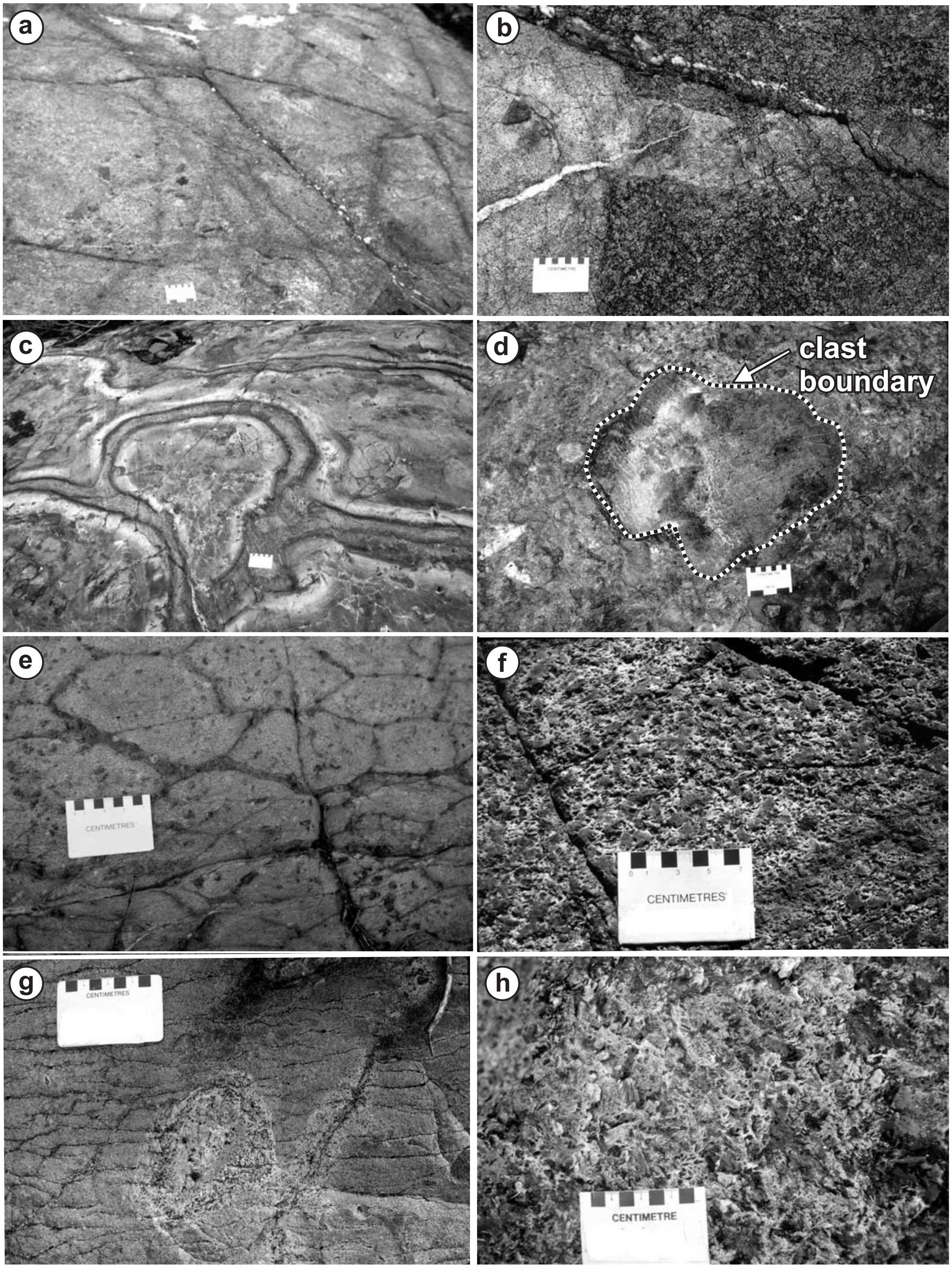

Figure 9: *Altered, metamorphically recrystallized Anderson sequence ('primitive arc') rocks:* ***a)*** *feldspar-destructive alteration (now chlorite-quartz [feldspar-poor] domains) adjacent to fractures, Sneath intrusive complex (subunit J5a);* ***b)*** *staurolite-chlorite–bearing altered tonalite (dark) cut by unaltered aplitic phase, Sneath intrusive complex (subunit J5a);* ***c)*** *albitized and silicified Welch basalt, 15 m below the 'Foot-Mud horizon' and the top of the Anderson sequence (subunit J1a);* ***d)*** *broken fragment of silicified/albitized Welch basalt in interflow heterolithic breccia (subunit J1b), attesting to near-seafloor alteration of flows (*see *Skirrow and Franklin 1994);* ***e)*** *staurolite prophyroblasts (dark) concentrated in polygonal joints that likely provided primary permeability for hydrothermal alteration, Daly rhyolite (subunit J3b);* ***f)*** *staurolite and biotite porphyroblasts (dark) adjacent to 'pipe-like' alteration zone within 100 m of the Raindrop VMS deposit (subunit J3b);* ***g)*** *oval domain of 'bleached', incipiently altered rhyolite with amphibole porphyroblasts (crosscut by recrystallized S_1 cleavage), Anderson-Stall rhyolite (subunit J3b);* ***h)*** *staurolite, biotite and kyanite (KY) porphyroblasts in Ram 'pipe-like' alteration zone (subunit J3b), 1.5 km from the Stall Lake VMS deposit.*

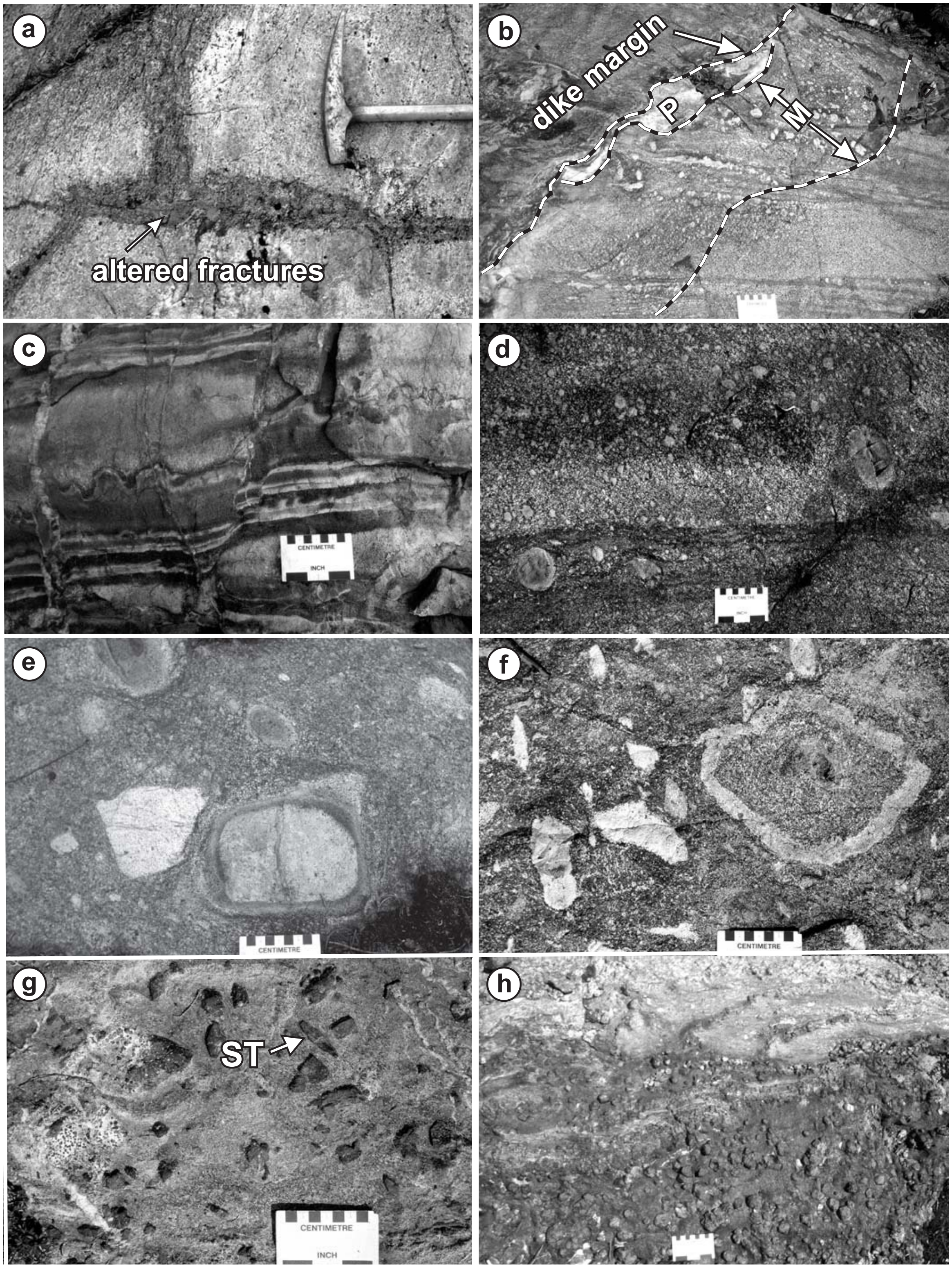

Figure 10: *Altered, metamorphically recrystallized Chisel sequence ('mature arc') rocks:* ***a)*** *garnet-chlorite metamorphic mineral assemblages overprinting synvolcanic Fe-Mg–metasomatized fractures, Richard intrusive complex (subunit J18a);* ***b)*** *synvolcanic mafic dike (unit J15) with adjacent zones of pervasive (P) and mottled (M) albitization and silcification overprinting Edwards bedded mafic tuff (subunit J8a); note amoeboid margin of dyke;* ***c)*** *zone of pervasive albitization and silicification (white) crosscutting bedded Edwards mafic wacke (subunit J8b); note load structures;* ***d)*** *orbicular domains of albitization and silicification (some with epidosite cores) overprinting bedded Edwards mafic tuff (subunit J8a);* ***e)*** *epidosite alteration domain with amphibole and quartz-feldspar rims overprinting fragment in heterolithic Edwards debris-flow breccia bed (subunit J8a);* ***f)*** *altered Moore pillow-fragment breccia characterized by albitized and silicified fragments and Fe-Mg–metazomatized matrix recrystallized to garnet-amphibole–rich mineral assemblages (subunit J10d);* ***g)*** *staurolite (ST)-garnet-chlorite-biotite–rich mineral assemblages in a discordant Fe-Mg alteration zone cutting altered Moore pillow-fragment breccia shown in photo F (subunit J10d);* ***h)*** *garnet-chlorite–rich, metamorphically recrystallized, Fe-Mg alteration zone cutting across albitized/silicified Edwards formation breccia (subunit J8a).*

Chisel North) and one Au-rich Cu-Zn (Photo) mines, as well as the recently discovered Lalor Zn-Cu deposit.

The Snow Lake arc assemblage has stratigraphic and geochemical evidence that both the VMS-hosting Anderson sequence and Chisel sequence volcanic rocks were affected by intra-arc rifting. Evidence for rifting in the Anderson sequence includes the combination of boninite, low-Ti basalt and isotopically juvenile rhyolite flows, an association of rock types that has been attributed in both modern and Phanerozoic arcs to high-temperature hydrous melting of refractory mantle sources in an extensional and/or proto-arc environment. Rifting in the Chisel sequence is indicated by the presence of the stratigraphically overlying Snow Creek 'arc-rift' basalt, with MORB-like geochemistry. Indirect support for rifting in the Chisel sequence is provided by voluminous volcaniclastic detritus; prominent synvolcanic dike sets; and local presence of highly fractionated, differentiated magma series. The VMS deposits of both the Anderson and Chisel sequences are spatially associated with isotopically juvenile rhyolite flows and related subvolcanic felsic intrusive complexes in bimodal basalt-rhyolite successions. The felsic magmas (rhyolite flows and intrusive complexes) are isotopically more juvenile than the intercalated mafic flows, thus precluding their derivation by fractionation in a high-level magma chamber of the same parent magmas that produced the associated mafic flows. One interpretation is that the felsic magmas were produced by high-temperature partial melting of the base of the oceanic lithosphere; this explains the bimodal nature of the extrusive volcanic sequences and the isotopically juvenile character of the felsic magmas.

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