# The nature of volcano-plutonic relations and the shapes of epizonal plutons of continental arcs as revealed in the Great Bear magmatic zone, northwestern Canada

**Robert S. Hildebrand[1], Paul F. Hoffman[2,*], Todd Housh[3,†], and Samuel A. Bowring[4]**

[1]*Department of Geology, University of California, One Shields Avenue, Davis, California 95616-8605, USA*
[2]*Department of Earth and Planetary Sciences, Harvard University, 20 Oxford Street, Cambridge, Massachusetts 02138, USA*
[3]*Department of Geological Sciences, The University of Texas at Austin, 1 University Station C1100, Austin, Texas 78712-0254, USA*
[4]*Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Building 54-1126, Cambridge, Massachusetts 02139, USA*

## ABSTRACT

**The shapes of plutons and their emplacement mechanisms, the connection between the volcanic and plutonic realms, and the development of batholiths have been of interest to geologists since they realized that plutons were once low viscosity magma. These issues have proven difficult to resolve because there are few places that have enough relief to expose the critical relations. The Great Bear magmatic zone, a Paleoproterozoic continental arc located in northern Canada's Wopmay orogen, provides an informative field setting to resolve some of these issues because the rocks are generally non-metamorphosed and were broadly folded such that calderas, stratovolcanoes, and a wide variety of plutons are exposed in oblique cross-section on fold limbs in an area of subdued topographic relief.**

**Early mafic plutons intruded co-magmatic pillow basalt piles as thin sheets with aspect ratios of 10–15. Plutons of intermediate composition, temporally associated with andesitic stratocones, have flat or slightly domical roofs and flat floors, and aspect ratios in the range of 5–10. Granodioritic to monzodioritic plutons that cut thick sequences of ash-flow tuff and related volcaniclastic rocks are generally sheet-like bodies with aspect ratios of 10–20, except where they intrude calderas and form resurgent plutons. Granitic plutons intrude at slightly deeper crustal levels, are generally younger, and typically have miarolytic cavities, pegmatites, and associated dike swarms. The granites have flat roofs and floors but generally have lower aspect ratios than the intermediate composition plutons.**

**Cycles—where magma bodies were first partially evacuated by eruption, then were re-energized and rose into their own ejecta to form plutons—span the compositional range from basalt to rhyodacite. The cycle of eruption, with the partial evacuation of chambers and subsequent rise of remaining magma to even higher levels in the crust explains why it is generally so difficult to link volcanic eruptions to specific plutons.**

**The overall development of the Great Bear magmatic zone—from small-scale local eruptions of basalt to voluminous eruptions of intermediate composition ash-flow tuff followed by wide-scale emplacement of granitic plutons—is interpreted to represent input of subduction-related magmas, which led to progressive heating, melting, and wholesale upward differentiation of the crust beneath the arc.**

## INTRODUCTION

Continental magmatic arcs are typically curvilinear zones, hundreds of km long, where voluminous volcanic and plutonic rocks result from the interaction of subduction-generated basalt and continental crust (Hildreth and Moorbath, 1988; Hildreth, 2007; Hughes and Mahood, 2008). Because matching individual plutons with specific volcanic rocks is difficult, one of the great unknowns in our understanding of continental arcs is the relationship of the exposed plutonic rocks to the volcanic rocks (Bachmann et al., 2007). The problem arises in part because there are few places in the world where both are well exposed. For the most part active arcs display the volcanic suprastructure, whereas older eroded equivalents appear to be dominated by gregarious plutonic rocks (Hamilton and Myers, 1967) with only a few mostly unrelated volcanic rocks preserved mainly as pendants and in the cores of calderas (Busby-Spera, 1984; Fiske and Tobisch, 1994; Saleeby et al., 1990; Pitcher, 1993).

On a smaller scale, some of the longest standing problems in igneous petrology involve the three-dimensional (3D) shape of plutons and how they create space for themselves (Daly, 1905, 1933; Lawson, 1914; Grout, 1927; Chamberlin and Link, 1927; Benn et al., 1998; Petford et al., 2000; Glazner and Bartley, 2006; Yoshinobu and Barnes, 2008). The shapes of plutons that constitute Cordilleran batholiths and continental arc terranes were generally considered to be in the form of "bell-jars" with steep sidewalls based mainly on plutonic exposures with several kilometers of structural relief in both the Sierra Nevada batholith and Coastal batholith of Peru (Bateman, 1992; Cobbing et al., 1981; Myers, 1975; Bussel and Pitcher, 1985; Cruden, 1998). Attempting to explain the geometries of compositional and petrological variations in plutons, other workers have argued convincingly that they see oblique cross-sections through tilted plutons (Flood and Shaw, 1979; Hildebrand, 1984b, 1986;

*Present address: School of Earth and Ocean Sciences, University of Victoria, Box 1700, Victoria, British Columbia V8W 2Y2, Canada
†Deceased.

*Geosphere*; December 2010; v. 6; no. 6; p. 812–839; doi: 10.1130/GES00533.1; 13 figures; 1 table.

Barnes et al., 1986; John, 1988; Haeussler and Paterson, 1993; Verplanck et al., 1999; Miller and Miller, 2002; Walker et al., 2007), some of which are tadpole-shaped (Hutchison, 1982; Hopson and Dehlinger, 1989). Still others, such as Cruden and McCaffrey (2001), used observations combined with geophysical and theoretical methods to conclude that many plutons form two end members: wedge-shaped and tabular. Similarly, Haederle and Atherton (2002) used gravity models to suggest that the entire Coastal batholith of Peru is a thin sheet similar to that suggested by Hamilton and Myers (1967) for Cordilleran batholiths in general.

The main reason why it is so difficult to evaluate the shapes of plutons is that, even in the highest mountain ranges on Earth, there isn't enough structural relief to see both roofs and floors of the same bodies. What is needed to resolve the shapes of plutons and their connections to volcanic rocks are areas where batholiths are folded, or possibly exposed in tilted fault blocks (Seager, 1981; Miller and Miller, 2002), yet are neither metamorphosed nor highly-strained. Over eighty years ago, Grout (1927) suggested that the Canadian shield was the best place to resolve problems relating to the "form, origin, history, and nature of batholiths" and we believe that there was some truth to his remarks.

The Great Bear magmatic zone—located in the northwestern Canadian shield and now mapped in considerable detail over 2° of latitude (Fig. 1)—is a Paleoproterozoic example of a magmatic arc developed upon continental crust (Hildebrand et. al., 1987a). It is one of the best places in the world to study the shapes of plutons because rocks of the zone are well-exposed in glacially-polished outcrops (Fig. 2A, B, C), are of low metamorphic grade, have little penetrative fabric, and are broadly-folded such that sections up to 10 km through the arc crust are visible despite subdued topographic relief (Fig. 2D). The plutons of the zone intrude related volcanic rocks and are epizonal (~1–5 km depth) as documented by early synvolcanic unroofing, razor-sharp contacts, narrow metamorphic aureoles, and preservation of thick sections of subgreenschist grade volcanic and sedimentary rocks. The folding allows us to see and map oblique cross-sections that display both floors and roofs of many plutons, from which we conclude that most of the plutons, except massive biotite granites, are not hippo-shaped, but rather thin and sheet-like in overall form. In this paper we emphasize the map relations that document the shapes of a variety of plutons within the Great Bear magmatic zone, discuss the overall evolution of the composite batholith, visit the relationship of plutonism to volcanism, and touch on their implications for other continental magmatic arcs.

## GEOLOGICAL SETTING

The Great Bear magmatic zone is the youngest magmatic belt within Wopmay orogen, a Paleoproterozoic orogen located in the northwestern Canadian Shield (Fig. 1). The orogen formed when Hottah terrane, a Paleoproterozoic arc-bearing microcontinent, collided with the western margin of the Archean Slave craton at ~1.884 Ga (Bowring and Grotzinger, 1992). During the collision, known as the Calderian orogeny, rift, shelf-rise and overlying foredeep rocks of Proterozoic age, which formed a west-facing package atop the Slave craton, and crystalline basement/forearc cover and plutons of Hottah terrane, were tectonically shortened and transported eastward along a basal décollement (Hildebrand et al., 2010). Eastward subduction following collision led to continental arc magmatism of the Great Bear magmatic zone during the period 1872–1843 Ma (Fig. 1; Hildebrand et al., 1987a). This model is supported by the Lithoprobe seismic experiment, which imaged mantle reflectors interpreted as relicts of eastwardly directed subduction dipping beneath the Great Bear magmatic zone (Cook et al., 1999; Cook and Erdmer, 2005). Shut-down of arc magmatism at ~1.84 Ga was followed by oblique folding of the arc rocks and finally, east-west shortening, accommodated by north-south extension, resulting in an orogen-wide swarm of conjugate transcurrent faults (Fig. 1). Younger Proterozoic dikes and sills, entirely unrelated to the events discussed here, occur throughout the region (Buchan et al., 2009).

## GREAT BEAR MAGMATIC ZONE

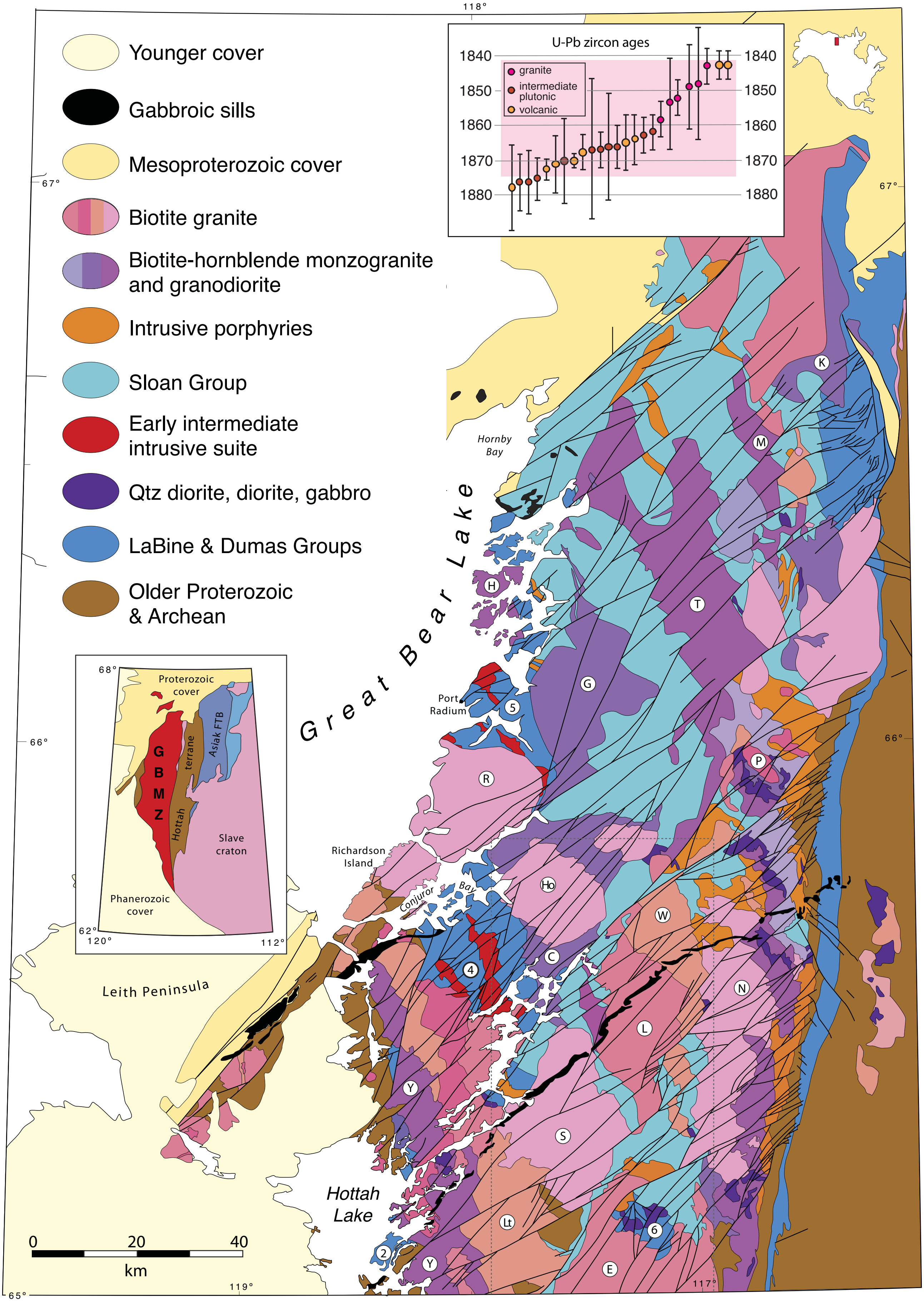
The Great Bear magmatic zone (Fig. 1) is an ~100 km-wide belt of mostly subgreenschist-facies volcanic and plutonic rocks that crop out over a strike length of 450 km. They unconformably overlie and intrude rocks deformed and metamorphosed during the Calderian orogeny (Hildebrand et al., 2010). The zone can be traced along strike for an additional 500 km southward beneath a thin veneer of Paleozoic cover and to the north bends westward then continues for another 300 km (Coles et al., 1976; Hildebrand and Bowring, 1984; Hoffman, 1987; Hildebrand et al., 2010). Thus, its two-dimensional extent is comparable to modern continental arc terranes. The overall structure of the zone is crudely synclinal (Hoffman and McGlynn, 1977; Hildebrand and Bowring, 1984) such that the oldest supracrustal rocks occur along its eastern and western margins, although sparse exposures of the lower part of the pile and its basement occur elsewhere within the zone.

The volcanic and sedimentary rocks of the Great Bear magmatic zone are collectively included within the MacTavish Supergroup, which is divided into three groups (Hoffman, 1978, 1982; Hoffman and McGlynn, 1977; Hildebrand, 2010). The LaBine group crops out along the western side of the zone; the temporally correlative Dumas group along the eastern side; and overlying both in the central part of the zone are rocks of the Sloan group. All of the plutonic rocks are informally termed the Great Bear batholith. We use the term pluton to refer to intrusive magma emplaced within a single outer contact that was mappable in the field. Individual plutons may have resulted from single or multiple pulses of injection. In our terminology, a batholith is simply an amalgamation of plutons that are more or less spatially and temporally related.

Except for early tholeiitic basalts (Reichenbach, 1991; Figures 2E, F), the volcanic rocks of the MacTavish Supergroup are mostly calc-alkaline and span the entire compositional range from basalt to rhyolite (Hildebrand et al., 1987a). Intermediate composition rocks dominate, and even where more siliceous compositions are abundant, andesitic lavas occur intercalated with the more siliceous units (Hildebrand, 1981, 1983, 1985). Modes and geochemistry of plutonic rocks are virtually identical to batholiths of continental arcs and Cordilleran-type batholiths (Hildebrand et al., 1987a).

In the west, LaBine Group consists of voluminous eruptions of intermediate to siliceous ash-flow tuff that when erupted, led to collapse of calderas, many of which were subsequently filled with porphyritic andesitic lavas, breccias

**Figure 1. (inset) Location of Great Bear magmatic zone (GBMZ) within Wopmay orogen. Geological sketch map showing the general geology of the northern Great Bear magmatic zone. Circled numbers refer to locations of subsequent figures. Circled letters are individual plutons: C–Calder quartz monzonite; E–Ellington Lake pluton; G–Gilleran pluton; Ho–Hooker megacrystic granite; K–Kamut pluton; L–Lever Lake pluton; Lt–Longtom Lake pluton; N–Nadin pluton; P–Handley Page intrusive suite; R–Richardson pluton; S–Self Lake pluton; T–Torrie pluton; W–Wylie pluton; Y–Yen pluton. Dashed box–location of Figure 12. (inset) U-Pb zircon dates (2 σ errors) from volcanic and plutonic rocks of the northern Great Bear magmatic zone (Bowring, 1985).**


Younger cover
Gabbroic sills
Mesoproterozoic cover
Biotite granite
Biotite-hornblende monzogranite and granodiorite
Intrusive porphyries
Sloan Group
Early intermediate intrusive suite
Qtz diorite, diorite, gabbro
LaBine & Dumas Groups
Older Proterozoic & Archean
U-Pb zircon ages
granite
intermediate plutonic
volcanic
1840
1850
1860
1870
1880
118°
67°
66°
65°
119°
117°
68°
62°
120°
112°
Proterozoic cover
Asiak FTB
terrane
Hottah
G B M Z
Slave craton
Phanerozoic cover
Hornby Bay
Great Bear Lake
Port Radium
Richardson Island
Conjuror Bay
Leith Peninsula
Hottah Lake
0
20
40
km
K
M
H
T
G
5
R
P
Ho
W
4
C
N
L
Y
S
Lt
6
2
E

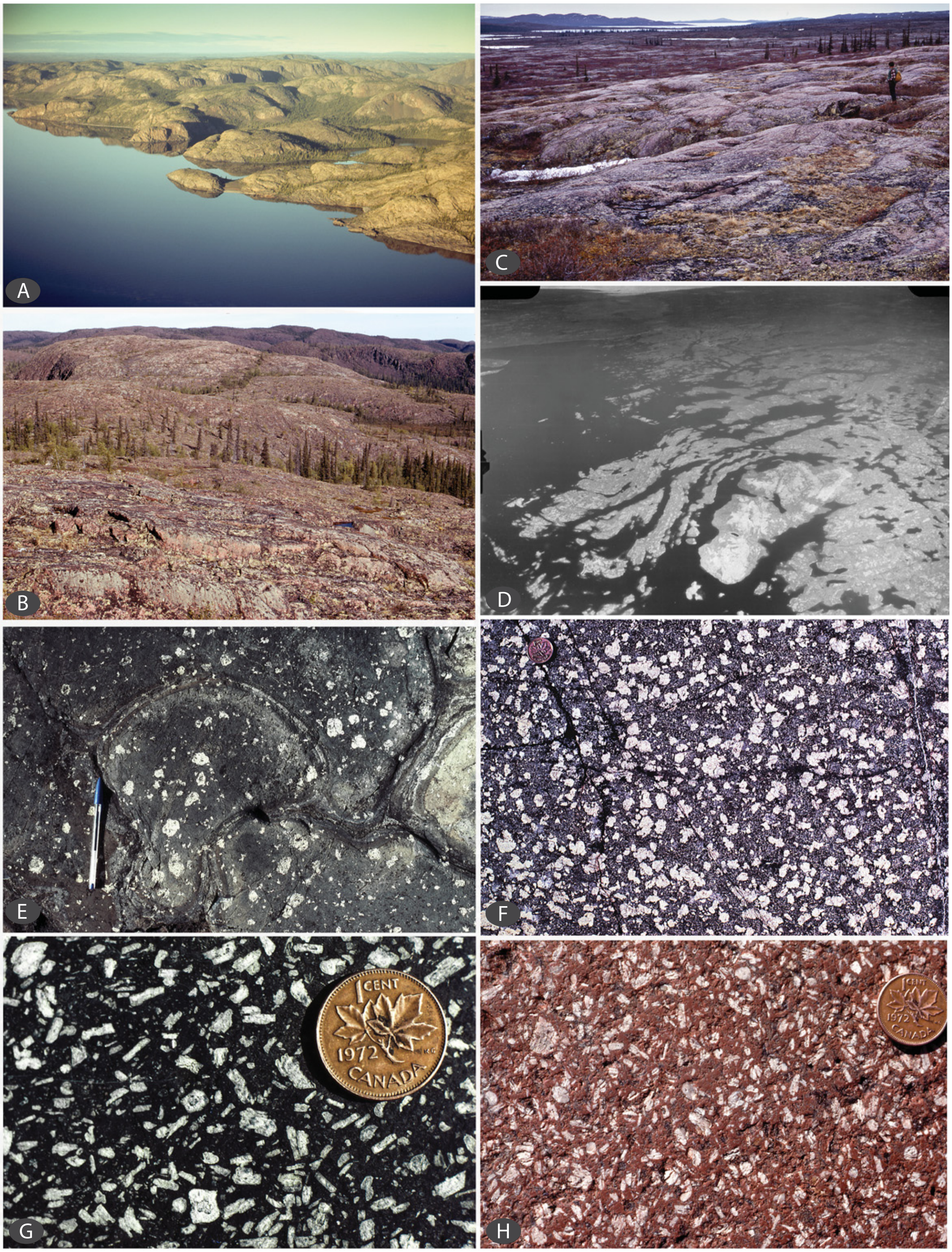



**Figure 2 (*continued on following pages*). Field photographs showing representative magmatic rocks and intrusive contacts within the Great Bear magmatic zone. The photographs provide the reader with the ability to evaluate the quality of outcrops and the validity of our comparison of rocks and textures there with younger equivalents despite the remoteness of the area. (A) oblique aerial view of the eastern shore of Great Bear Lake just south of Port Radium showing part of a 3 km thick section of andesite; (B) in many areas where rocks of the Sloan group are exposed, outcrop is nearly 100%; (C) typical view of low-standing terrain dominated by intrusive rocks with higher-standing volcanic belts in the distance; (D) oblique aerial view looking northward over Hottah Lake to Great Bear Lake, courtesy of the RCAF, showing folds along the eastern shore of Hottah Lake (compare with Fig. 3); in the extreme center-right one can see a prominent white ridge, which represents a giant quartz stockwork, typical of the western Great Bear magmatic zone (Furnival, 1938); these resistant stockworks assist in mapping transcurrent faults through otherwise massive granitic plutons; (E) glomeroporphyritic pillow basalts of Bloom Basalt at Hottah Lake; (F) glomeroporphyritic sill of Fishtrap Gabbro; (G) typical plagioclase porphyritic andesitic lava; (H) oxidized porphyritic andesite;**

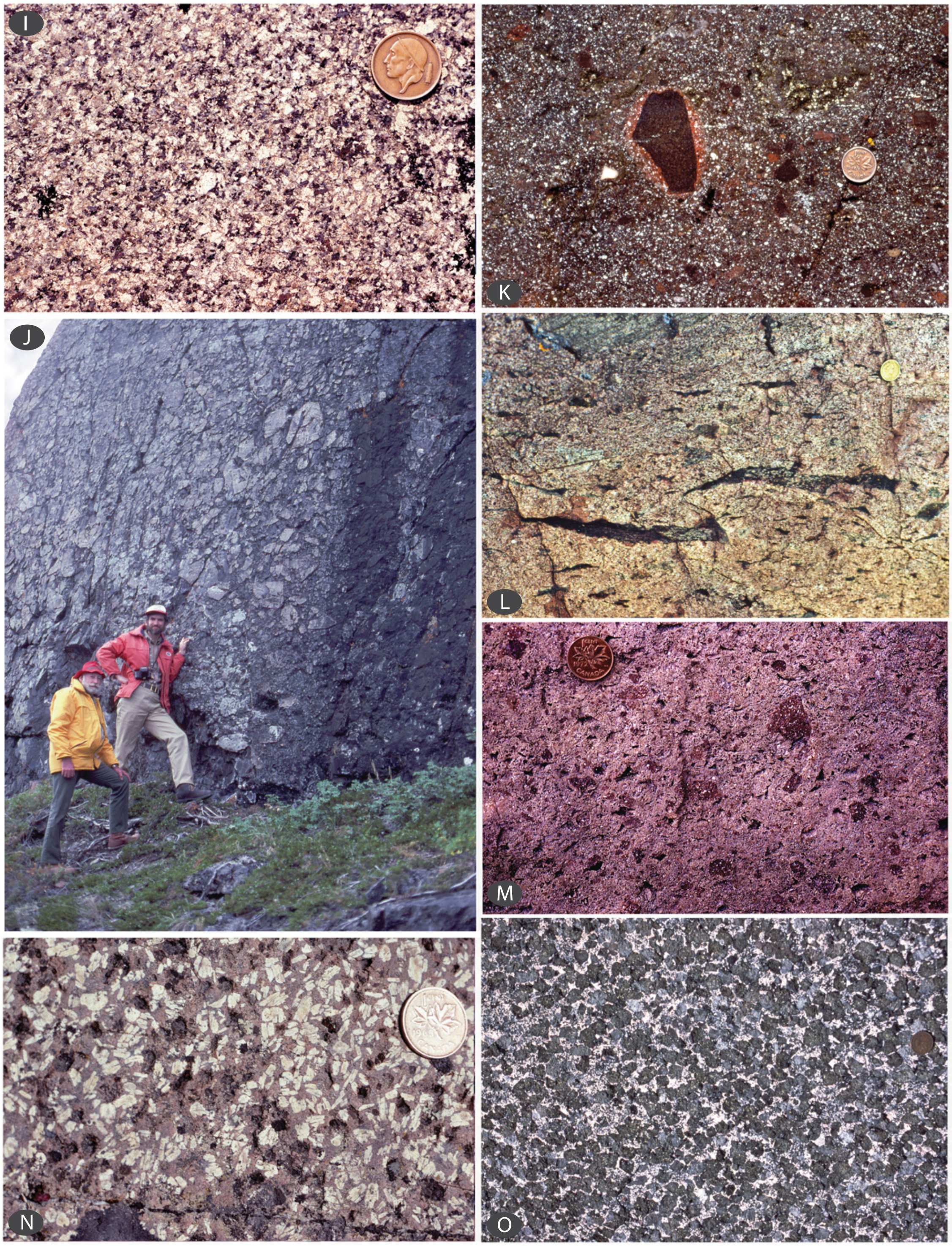



**Figure 2 (*continued*). (I) seriate textured Balachey pluton; (J) geologists at the buttress unconformity formed along the Clut cauldron margin; the mesobreccia here is dominated by clasts of Balachey pluton and is less than 1 m thick on the cliff-face; (K) crystal- and lithic-rich intracauldron facies White Eagle tuff of Clut cauldron; (L) outflow facies tuff showing typical preservation of original welding textures; in thin section relict shards are observed; (M) densely-welded tuff of Sloan group, note abundant cognate lithic fragments; (N) porphyritic border phase of Rainy Lake intrusive complex (compare with andesite of Figure 2G and 2H; (O) poikolitic gabbro;**

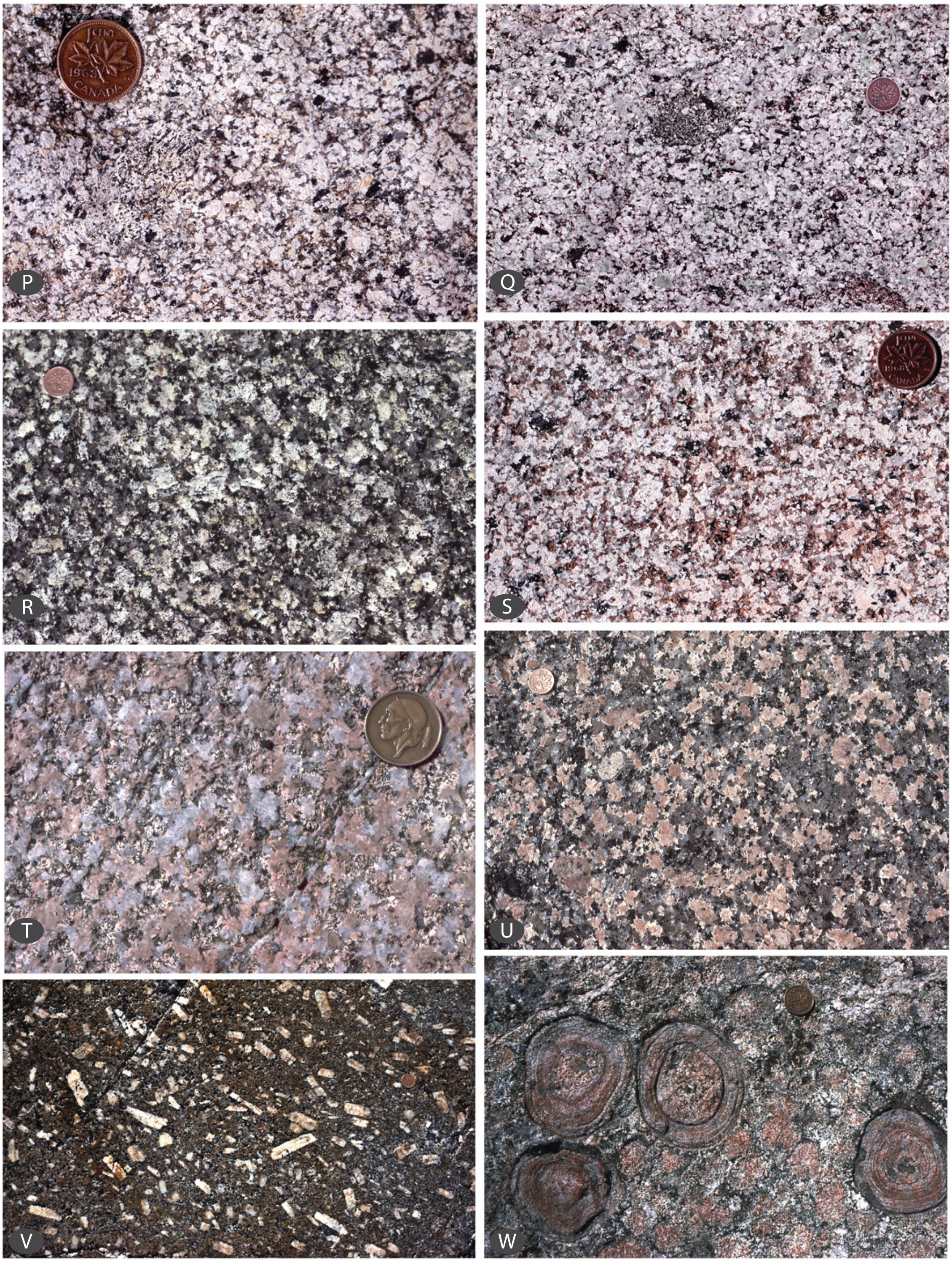



**Figure 2 (*continued*). (P) hornblende quartz diorite (note cognate (?) blob of finer-grained material beneath penny) and weak alignment of hornblende prisms; (Q) granodiorite of Yen pluton showing typical hornblende prisms, seriate plagioclase, large quartz, and enclave; (R) seriate textured hornblende-biotite bearing Calder quartz monzonite; (S) biotite-hornblende granodiorite-monzogranite of Torrie pluton showing typical seriate plagioclase: (T) medium-coarse grained Richardson granite; (U) poorly-formed phenocrysts of potassium feldspar in biotite granite; (V) strongly porphyritic granite showing dominantly euhedral phenocrysts of potassium feldspar; (W) orbicular diorite, in some cases (left) layers formed around cognate enclaves of diorite suggesting that they were suspended in melt;**

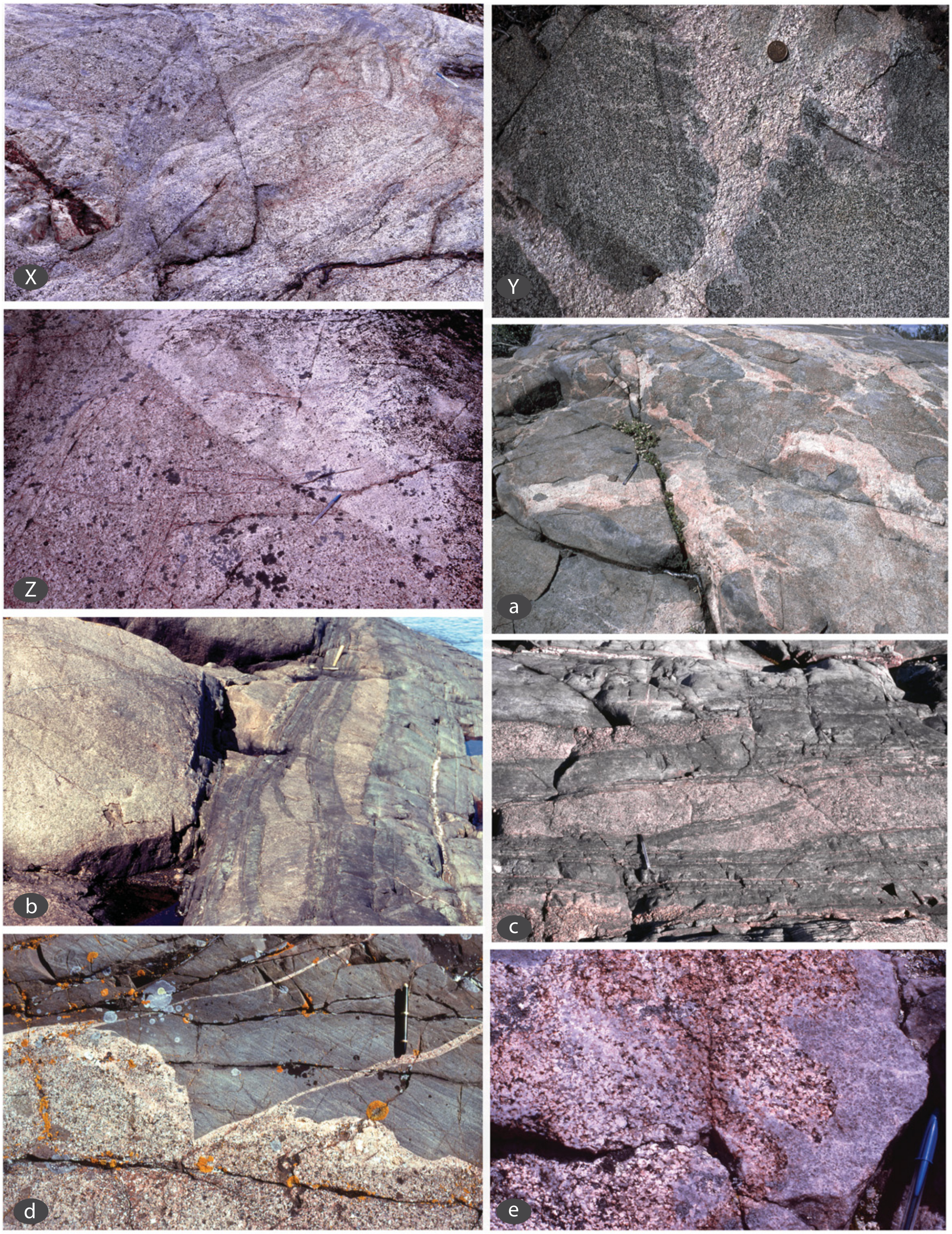



**Figure 2 (*continued*). (X) compositional layering at contact of Yen pluton; pen in upper right for scale; (Y) magma mingling between dioritic blobs and monzogranite (note fine-grained zone at contact); (Z) typical sharp granite-granite contact; (a) magma mingling texture, such features are rare in the Great Bear magmatic zone; (b) steeply-dipping contact between Richardson granite (left) and sedimentary rocks (right) showing narrow sill complex at the formerly horizontal contact; (c) detail of Richardson contact showing invasion of granite along bedding planes even at very fine scales, which suggests that the body was dominantly melt at the time of intrusion; the prominent prong of metasedimentary rock extending from the pen to the upper right is the same prong as that located in the middle foreground of photo b; (d) razor-sharp contact of granite near Port Radium showing pegmatitic border phase and thin dikes intruding microdiorite; (e) contact of Yen pluton with quartz arenite; note that the sandstone is disaggregated and incorporated within the pluton (see Fig. 2Q);**

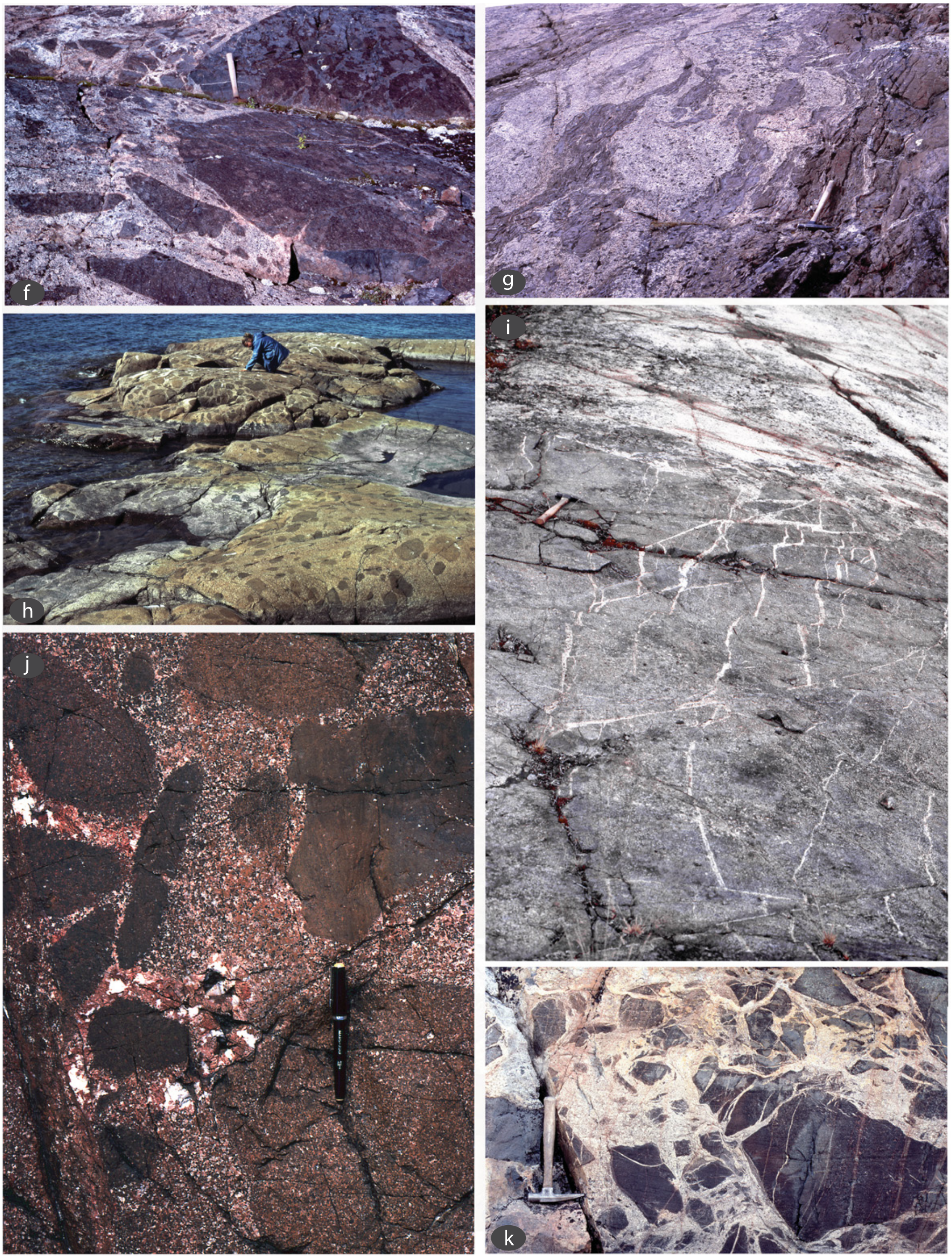



**Figure 2 (*continued*). (f) blocky xenoliths formed where Yen pluton intruded andesite; (g) blob-like irregular-shaped xenoliths formed where Yen pluton intruded metasedimentary rocks; (h) abundant enclaves of uncertain origin in granite at Port Radium; (i) small-scale dike swarm at contact of granodiorite-monzogranite with diorite illustrating how pluton invades rigid isotropic wallrock along pre-existing fractures (note completely engulfed dark blocks in the distance); (j) detail of the outcrop in (h) showing crowded enclaves and pegmatite filled miarolytic cavities; (k) dominantly angular inclusions with different generations of dikelets breaking enclaves into smaller pieces.**

and tuffs erupted from andesitic stratovolcanoes (Fig. 2G, H). In places, preserved deposits of the stratocones are up to 3km thick. The ignimbrites of the LaBine group are highly variable, but generally contain 5%–35% phenocrysts, and range from small volume andesitic tuffs, erupted from andesitic stratocones, to voluminous, monotonously-intermediate and compositionally-zoned dacite and rhyodacitic sheets of outflow facies tuff and caldera fill (Fig. 2J, K, L). Contemporaneous eastern facies are predominantly fluvio-lacustrine with thin outflow-facies tuffs and minor lavas, collectively known as the Dumas Group. In general, stratigraphically-highest rocks of the Sloan Group occur within the central part of the zone and constitute sections many kilometers thick dominated by monotonous crystal-rich, densely-welded ignimbrite, typically dacitic with abundant cognate lithic fragments (Fig. 2M), and interbedded with only minor lava flows and sedimentary rocks. Detailed facies studies are incomplete, but preserved thicknesses of ash-flow tuff within the Sloan Group indicate that it comprises complexly nested calderas formed by cataclysmic eruptions of monotonously-intermediate ignimbrite.

All of the supracrustal rocks of the belt are cut by epizonal plutons (Fig. 2C), several of which are interpreted to be synvolcanic based on field relations and by U-Pb geochronology (Bowring, 1985). Although we cannot match any single pluton with a specific eruptive product, in a general sense—and based on U-Pb geochronology and unconformities that deroof plutons—there is little doubt that the batholith intruded its own ejecta. The plutonic rocks are grouped into four suites, based on composition and temporal relations (Hildebrand et al., 1987a). The oldest are gabbroic and comagmatic with early pillow basalts. Slightly younger is a compositionally diverse suite of relatively small intermediate to siliceous bodies, collectively named the “early intermediate intrusive suite.” They are spatially and temporally related to andesitic volcanism. Younger yet are huge plutons, corporated as the “granodiorite-monzogranite suite,” that are temporally and compositionally related to voluminous eruptions of ash-flow tuff of the Sloan Group. The youngest plutons are large bodies of granite with only minor eruptive equivalents. Overall, plutonism within the Great Bear magmatic zone became more siliceous with time.

All of the volcanic rocks and all of the plutons are folded (Fig. 2D). Throughout most of the zone, the folds trend northwest but as the axial traces approach the eastern margin of the zone, they bend progressively N-S, parallel to the edge of the Slave craton (Hildebrand et al., 1987b, 1990; Hildebrand and Bowring, 1988). The folds plunge gently everywhere. The regional map (Fig. 1) shows that many plutonic contacts parallel the northwest-trending or northerly striking axial traces of the younger folds and are exposed in cross section. Stratigraphic sections several km thick are exposed on single fold limbs and throughout most of the zone there is no megascopically recognizable planar fabric related to folding with the exception of cleavage within mudstones and poorly-welded tuffs, located mostly in the east where the folds are tighter than elsewhere due to the buttressing effect along the edge of the Slave craton (Hoffman, 1984; Hildebrand and Bowring, 1988).

The east-west shortening that created the northeasterly-trending, dextral transcurrent faults distorted the original trends of fold axes and contacts because the individual fault blocks and their bounding faults rotated in a counterclockwise fashion during shortening of the entire orogen (Freund, 1970; Tirrul, 1987). The overall effect is to cause originally northerly-trending contacts to trend more northwesterly today.

The majority of volcanic and plutonic rocks have U-Pb ages between ~1872 Ma and 1860 Ma, but some granites and a few dated volcanic rocks have ages as young as 1843 Ma (Fig. 1; Bowring, 1985). Oblique folding and transcurrent faulting are younger than 1843 Ma and predate a WNW-trending mafic dike swarm dated at 1740 +5 / −4 Ma (Irving et al., 2004).

### Mafic Intrusions in the Early Basaltic Lava Pile

The oldest magmatic rocks of the Great Bear magmatic zone are 2–3 km of basaltic lavas of the LaBine group exposed along the western margin of the zone (Fig. 1) at Hottah Lake and Conjuror Bay (McGlynn, 1975; Hoffman and McGlynn, 1977; Hildebrand, 1982; Hildebrand et al., 1983; Reichenbach, 1991). The lavas are cut by gabbroic sills (Fig. 3; Table 1), which range in thickness from a few tens of meters to one kilometer and are many kilometers wide. In general they are at least 10–15 times wider than they are thick. Texturally, they range from fine-grained diabase to coarse-grained equigranular, and aphyric to plagioclase porphyritic. Texturally, mineralogically, and chemically the gabbros are similar to the basalts they intrude, except that they are coarser-grained. For example, both sills and basalts contain identical glomeroporphyritic clots of plagioclase (Fig. 2 E, F), and as the clots occur in chilled selvages of pillows and border phases of sills, the lavas and sills are interpreted to be comagmatic (Hildebrand et al., 1983; Reichenbach, 1986). The sills wrap around fold noses defined by bedding in sedimentary and basaltic wallrocks (Fig. 3). Nearly everywhere their razor-sharp contacts dip parallel to bedding.

### The Early Intermediate Intrusive Suite: Plutons Associated with Andesitic Stratovolcanoes

Along the western margin of the Great Bear magmatic zone are accumulations up to 3 km thick of strongly plagioclase-porphyritic andesitic lavas (Fig. 2 G, H), breccias, and related epiclastic rocks that are collectively interpreted to constitute the deposits of stratocones (Hoffman et al., 1976; Hoffman and McGlynn, 1977; Hildebrand, 1981, 1984a). In several cases the cones developed within pre-existing calderas (Hildebrand, 1984a, b). The andesites and their substructure are intruded and considerably altered by a suite of intermediate composition plutons (Figs. 1, 4, 5; Table 1), collectively termed the “early intermediate intrusive suite” (Hildebrand et al., 1987a).

Several plutons of the suite were studied in considerable detail (Furnival, 1939; Tirrul, 1976; Hildebrand, 1984a, b, 1986; Cherer, 1988; Reardon, 1992) as they and their alteration haloes host rich polymetallic veins of native Ag, Ni-Co arsenides, and pitchblende. The plutons are 5–25 km in diameter and 1–2 km thick, of monzonite, monzodiorite, diorite and granite. For the most part they intruded 2–3 km beneath the surface and their contacts with wall rocks are sharp. Xenoliths are sparse. Nearly all of the members of the suite have well-developed alteration haloes up to 1 km wide, comprising an inner albite zone, a central magnetite-apatite-actinolite zone, and an outer pyritic zone (Hildebrand, 1986; Reardon, 1992). At least three of the intrusions were de-roofed soon after emplacement as younger rocks unconformably overlie the intrusions and/or clasts of the intrusions occur in stratigraphically younger units that are themselves cut by plutons and folded.

Overall, the plutons are compositionally heterogeneous, but uniformly medium-grained with a well-developed seriate texture defined by plagioclase (Fig. 2I). Subhedral-euhedral hornblende and biotite are the dominant ferromagnesian minerals but where preserved, border phases (Fig. 2N) locally contain relict clinopyroxene and serpentinized orthopyroxene. Magnetite is the dominant opaque mineral throughout the intrusions.

The plutons crop out in two main areas: the Conjuror Bay-Camsell river area and in the area around Port Radium (Fig. 1). More detailed maps (Figs. 4, 5) show the geological setting and relations, whereas the characteristics of individual plutons are listed in Table 1. In general, the plutons form sheet-like bodies with flat,

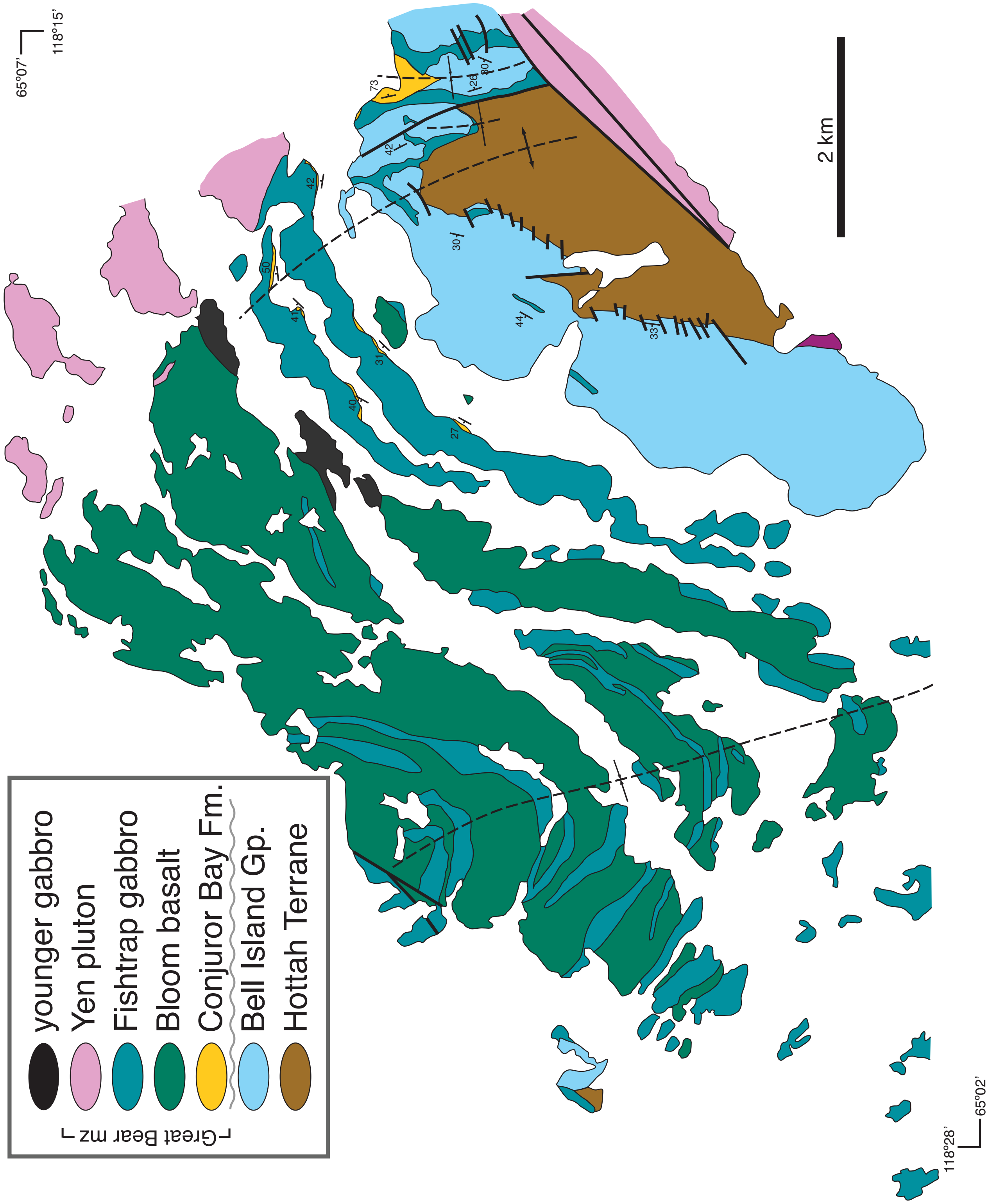



**Figure 3. Geological map showing folded Fishtrap Gabbro and Bloom Basalt at the base of the Great Bear magmatic zone.**

TABLE 1. CHARACTERISTICS OF INDIVIDUAL PLUTONS

| Pluton Name | Suite | Contact Relations | Composition | Compositionally and/or Texturally Zoned | Exposed Length | Exposed Thickness | Eruptive Equivalents | Setting | Map | References | notes |
|---|---|---|---|---|---|---|---|---|---|---|---|
| **Fishtrap gabbro** | Early gabbro | sharp intrusive ll bedding | pyroxene gabbro | Upper part fine-grained with amygdules Plagioclase glomerocrystic central part Lower parts pyroxene | >15 km | 1 km | Pillow basalts | Intrudes early pillow basalts | 2 | McGlynn, 1975: Hildebrand et al., 1983; Reichenbach, 1986, 1991 | Plutons exposed in oblique cross-section and wrap around several fold noses (Figure 3). Glomerocrystic plagioclase in both lavas & sills |
| **Rainy Lake** | Early Intermediate intrusive suite | sharp intrusive ll bedding | cpx-hb monzonite monzodiorite | Border phase with 30-35% plag phenocrysts (Figure 2N)); Upper pseudosyenite, Central monzonite, Plagioclase cumulate monzodiorite | >11 km | 1.5 km | Possibly Camsell River andesite | Intrudes at the base of intracauldron andesite pile | 4 | Tirrul, 1976; Hildebrand, 1984b; 1985; 1986 | Pluton exposed in oblique cross-section; magnetite-apatite-actinolite alteration halo above the roof and superimposed over upper part of pluton (see Figure 4) |
| **Balachey pluton** | Early Intermediate intrusive suite | sharp intrusive ll bedding; unconformably overlain by caldera collapse breccia | hb monzonite-monzodiorite | Medium- to coarse-grained, seriate-textured hornblende quartz monzonite near roof and grades to a qtz monzodiorite at depth (Figure 2I) | >20 km | unknown | Possibly Camsell River andesite | Intrudes intracauldron andesite pile | 4 | Hildebrand, 1984a: 1984b; 1985; 1986; Bowring, 1985 | Pluton exposed in core of anticline; magnetite-apatite-actinolite alteration halo above a flat roof; pluton deroofed during subsequent caldera collapse (Figure 2J). 1875±5 Ma (U-Pb zircon & stratigraphy); Figure 4 |
| **Glacier Lake** | Early Intermediate intrusive suite | sharp intrusive ll bedding | bio-hbd granodiorite hbd-bio granite | Plagioclase porphyritic border phase. Little variation in silica | 4-5 km | 500-750 m | Echo Bay andesite | Intrudes andesite pile | 5 | Hoffman et al., 1976: Hildebrand, 1983; Cherer, 1988: Reardon, 1989: 1990: 1992 | Boulders of pluton in conglomerate interbedded with uppermost andesitic lavas. Pluton exposed in oblique cross section on limb of syncline. |
| **Contact Lake** | Early Intermediate intrusive suite | flat-floored laccolith | hb-biotite monzonite-monzodiorite; granite-granodiorite | Compositionally inversely zoned with qtz monzonite-monzodiorite near roof and granite-granodiorite near floor | 8 km | 0-2km | none documented | Intrudes andesite pile | 5 | Hildebrand, 1983; Cherer, 1988: Reardon, 1989: 1990: 1992 | Exposed in oblique cross-section on limb of syncline. Mag-ap-act alteration. Evidence for deroofing; Figure 5 |
| **Bertrand Lake** | Early Intermediate intrusive suite | sharp intrusive ll bedding | biotite-hb monzodiorite | Not conspicuously zoned | 7km | 1-1.5km | none documented | intrudes base of andesite pile | 5 | Hildebrand, 1983; Reardon, 1989 | Pluton exposed in oblique cross section with flat floor and roof. Strong alteration above roof. Figure 5 |
| **Ellington diorite** | Early Intermediate intrusive suite | sharp intrusive ll bedding | hbd diorite | Not conspicuously zoned | 10 km | 500m | none documented | Intrudes diverse stratigraphy. | 6 | Pelletier, 1988 | Amydules close to roof; Figure 6 |
| **McLaren pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | bio-cpx-hbd qtz diorite to bio syenogran | Highly variable composition that varies at the scale of the body and outcrop | 25 km | 2 km | none documented | Intrudes ash-flow tuffs, lavas, porphyries, and Hottah terrane | 9 | Housh, 1989 | On regional basis, contacts are concordant with bedding in wall rocks, but locally the body cuts bedding at variable angles. Figure 7. |
| **Hogarth pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | hbd-biotite granodiorite; monzogranite, qtz diorite | Not conspicuously zoned | >12km | >3-4 km | Cornwall tuff | Cuts up section through andesite and sedimentary rocks. | 10 | Hildebrand, 1983; 1984a | Domical crest with block faulted roof rocks. Pluton likely represents a resurgent pluton in Cornwall caldera. Lower contact not exposed. 1876±8 Ma; Figure 8. |
| **Calder quartz monzonite** | Granodiorite-monzogranite suite | sharp intrusive | hbd-biotite qtz monzonite; monzogranite, | Not conspicuously zoned | 20-30 km but intruded by plutons | unknown | White Eagle tuff | Intrudes intra-caldera White Eagle tuff and collapse breccias. | 12 | Hildebrand, 1984a: 1984b; 1985; Bowring, 1985 | Intruded core of Clut caldera, evidence for uplift of area following caldera collapse 1868±2Ma (U-Pb zircon); Figure 12 |
| **Torrie pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | bio-cpx-hbd qtz diorite to bio syenogran | Highly variable on all scales with gradational contacts between rock types | 100 km | ~4km (?) | none documented | Intrudes thick ash-flow tuffs and minor sedimentary rocks; intruded by younger plutons. | 11 | Hildebrand et al.,1987; Hoffman et al., 1976; Hoffman, 1978; 1982; | Intrudes thick ash-flow tuffs of the Sloan group. Nearly flat-concordant roof; irregular in detail but flat on a pluton scale. Floor also horizontal; Figure 11. |

(*continued*)

TABLE 1. CHARACTERISTICS OF INDIVIDUAL PLUTONS (*continued*)

| Pluton Name | Suite | Contact Relations | Composition | Compositionally and/or Texturally Zoned | Exposed Length | Exposed Thickness | Eruptive Equivalents | Setting | Map | References | notes |
|---|---|---|---|---|---|---|---|---|---|---|---|
| **Gilleran pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | hbd-biotite granodiorite; monzogranite, | Variable with northern and southern exposures biotite-hbd monzogranite; central regio biotite-hb syenogranite | 45km | 0-4 km | none documented | Intrudes thick ash-flow tuffs of Sloan Gp and lavas/seds of LaBine Gp. | 11 | Hoffman et al., 1976; Hoffman, 1978; 1982; Hildebrand, 1983 | Concordant roof exposed along NE and W sides, concordant floor locally exposed in north; Figure 9. |
| **Kamut pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | bio-cpx-hbd qtz diorite to bio syenogran | Highly variable on all scales with gradational contacts between rock types | 40km | unknown | none documented | Intrudes thick ash-flow tuffs of Sloan Gp and lavas/seds of Dumas Gp. | 1 | Hoffman, 1978; 1982 | Floor concordant and parallel to stratigraphy; cuts up section; floor exposed in western part and is concordant with bedding. Figure 1 |
| **Yen pluton** | Granodiorite-monzogranite suite | sharp intrusive ll bedding | biotite-hbd granodiorite; monzogranite, | Highly variable on an outcrop scale; 10-25% ferromagnesian minerals; in places k-spar porphyritic; local marginal layering. Euhedral hornblende prisms (Figure 2Q) | >100km | unknown | none documented | Intrudes Hottah terrane, pillow basalt, Bell Island Bay Gp | 1 | Hildebrand and Roots, 1985 | Lower contact mainly horizontal but locally climbs up or down section; exposed in core of syncline. Where intrudes qtz arenites, becomes very qtz rich; Figure 1 |
| **Handley Page Intrusive Series** | Granodiorite-monzogranite suite | sharp between phases and with wallrock | granite to gabbro | 3 distinct phases, from oldest to youngest and from margin to core: (1) cpx-biohbd diorite, qtz diorite minor clinopyroxenite, (2) hbd-bio monzogranite-granodiorite, potassium feldspar porphyritic biotite syenogranite. Contacts between phases are sharp. | 45km | unknown | none documented | Intrudes Sloan Gp | 1 | Gibbins (1988); Hildebrand et al., 1987 | Major and most minor elements vary systematically with SiO2. Figure 1 |
| **Ellington pluton** | Granite suite | sharp intrusive ll bedding | biotite granite | Not conspicuously zoned | 25-30 km | unknown | none documented | Intrudes a diverse assembage of sedimentary & volcanic | 12 | Pelletier, 1988 | Intrusive floor extensively exposed; roof not found in map area; Figures 6, 12. |
| **Richardson pluton** | Granite suite | sharp intrusive ll bedding | porphyritic; non-porphyritic biotite granite | Not conspicuously zoned | 20 km | unknown | none documented | intrudes at or close to basement-cover contact | 15 | HIldebrand, 1985; Bowring, 1985 | Upper contact is horzontal and planar on a small scale but locally climbs up/down section. Has sill complex where it intrudes bedded rocks (Figures 2b,c). Floor is crudely horizontal and intrudes Hottah gneiss. 1858±2 Ma (U-Pb) |
| **Self Lake pluton** | Granite suite | sharp intrusive ll bedding | porphyritic; non-porphyritic biotite granite | Not conspicuously zoned | 30 km | unknown | none documented | intrudes at or close to basement-cover contact | 12 | | Planar, concordant roof along NE contact; planar, horizontal floor along SW corner; Figure 12. |
| **Lever Lake pluton** | Granite suite | sharp intrusive ll bedding | porphyritic; non-porphyritic biotite granite | Not conspicuously zoned | 30 km | unknown | none documented | intrudes ash-flows sedimentary rocks, older plutons | 12 | | Planar, concordant roof in SW, planar, horizontal floor, with minor protuberances in NE; Figure 12. |
| **Longtom Lake pluton** | Granite suite | sharp intrusive ll bedding | porphyritic; non-porphyritic biotite granite | Not conspicuously zoned | 30-35 km | unknown | none documented | Intrudes Hottah terrane, Bell Island Bay Gp | 12 | | Planar, concordant roof in SW; Figure 12. |
| **Hooker megacrystic granite** | Granite suite | sharp intrusive | porphyritic biotite granite | Not conspicuously zoned | 20 km | unknown | none documented | Intrudes Calder quartz monxonite | 12 | HIldebrand, 1984b; 1985 | Planar roof exposed; oval shape overall suggests steep sidewalls; Figure 12. |

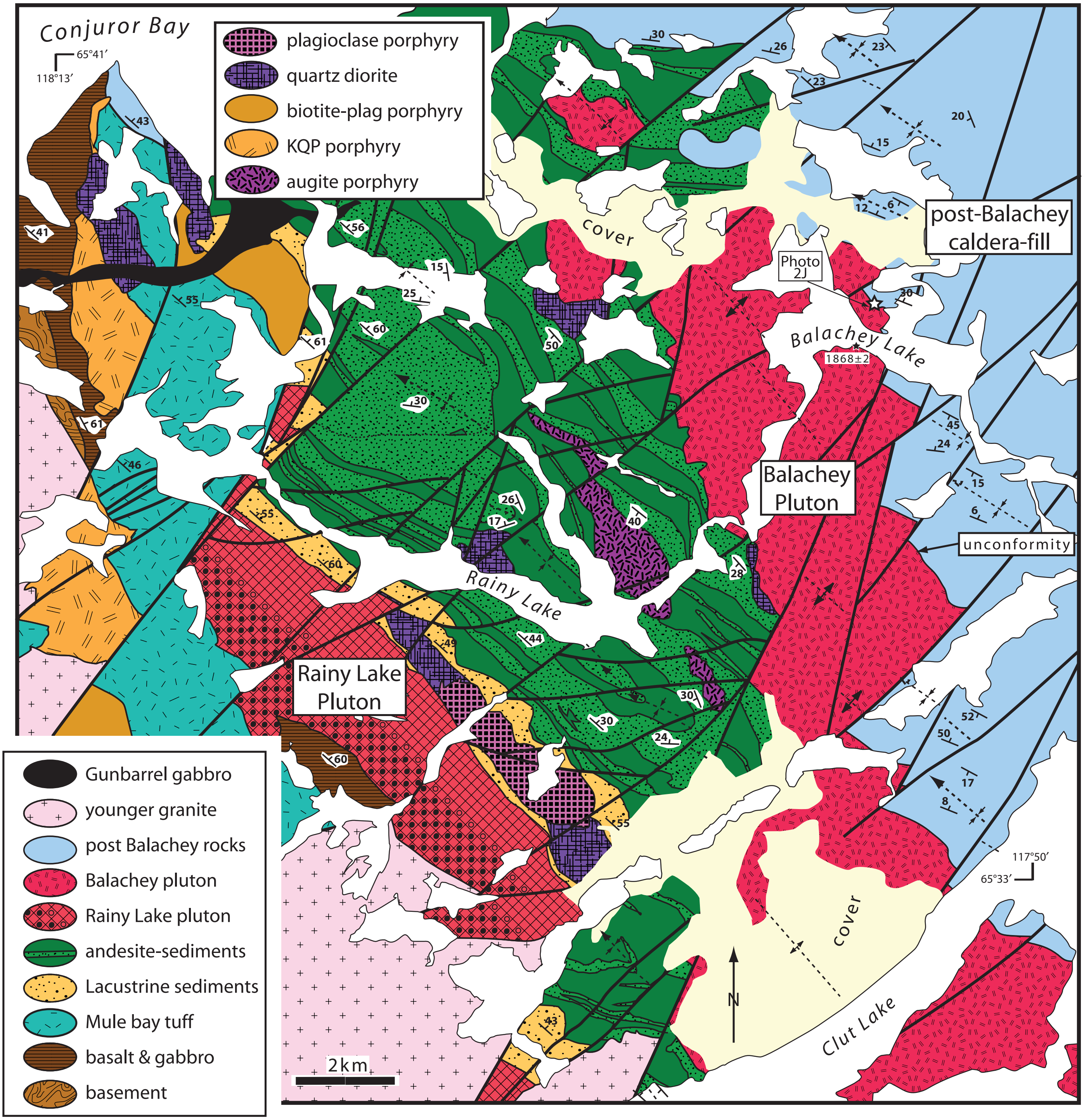



**Figure 4. Geological sketch map of the Balachey-Rainy lakes area showing the geological setting of the Balachey and Rainy Lake intrusions (early intermediate intrusive suite) discussed in the text. The thick northeastwardly-dipping Mule Bay tuff located in the western part of the figure is interpreted to represent intracauldron facies tuff. The overlying lacustrine sediments and thick accumulation of andesites were deposited within the Black Bear caldera following ash-flow eruptions (Hildebrand, 1984b; 1985). The Rainy Lake and Balachey plutons intruded rocks of the caldera complex and the Balachey pluton was deroofed during formation of Clut caldera, the fill (White Eagle tuff and breccias) of which is shown in light blue. The U-Pb zircon age of the Balachey is 1868 ± 8 Ma, the resurgent Calder quartz monzonite (Fig. 12), which intrudes the fill of Clut caldera has a U-Pb zircon age of 1868 ± 2 Ma, and intracaldera facies White Eagle Tuff yielded a U-Pb age of 1873 ± 6 Ma. (Bowring, 1985). Note that the Rainy Lake body and those just above its roof are exposed in oblique cross section; whereas the Balachey pluton occupies the core of an anticline.**

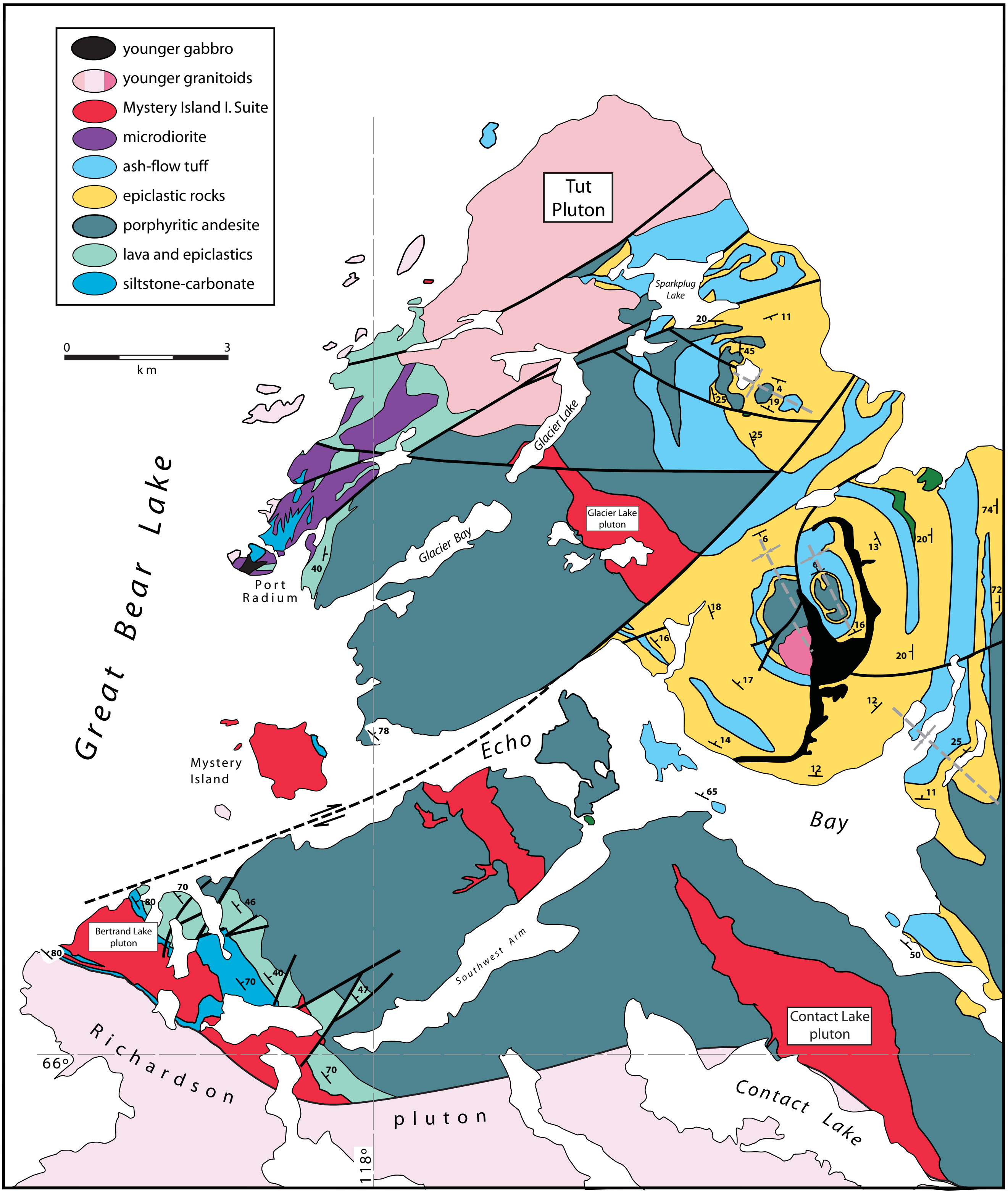



**Figure 5. Geological sketch map showing the setting for members of the early intermediate intrusive suite in the Echo Bay area (Hildebrand, 1983). The plutons in the belt crop out in oblique cross-section within the northeasterly-dipping limb of a major syncline. Similar alteration in the Port Radium area to that observed above the roofs of other plutons of the suite suggest that a member of the suite formerly occurred there but is now cut-out by younger granite. The steeply-dipping roof of the Richardson pluton, formerly horizontal over extensive distances, is shown in the southern part of the figure. The sill complex at the contact, shown in Figure 2 (b, c), occurs where the contact meets Great Bear Lake.**

concordant roofs and floors. Another early pluton, a diorite, intrudes farther east at Ellington Lake (Pelletier, 1988), and also has both roof and floor that are dominantly concordant with bedded wall rocks (Fig. 6).

### Granodiorite-Monzogranite Suite

The granodiorite-monzogranite suite is one of the major plutonic suites of the zone. Most contacts of the plutons are concordant with bedding in their wall rocks, and trend northwest (Fig. 1; Table 1), parallel to the regional fold axes. There is little doubt that they are folded along with their wall rocks as contacts and narrow thermal aureoles ranging in grade up to hornblende hornfels facies swing around fold noses with wall rock stratigraphy (Figs. 7, 8, 9). U-Pb ages of the plutons are generally indistinguishable from those obtained from the ash-flow tuffs they intrude (Bowring, 1985), which, along with similar compositions and textures, suggests in the broad sense that the plutons and tuffs may be genetically related. Most of the members of this suite are large plutons, ranging up to 100 km across, with flat, concordant floors and roofs, but at least two members, both of which appear to be resurgent plutons in the cores of calderas, have domical roofs (Table 1).

Most commonly, members of the suite are medium-grained bodies of granodiorite and monzogranite as the name of the suite suggests; however, more siliceous and more intermedi-

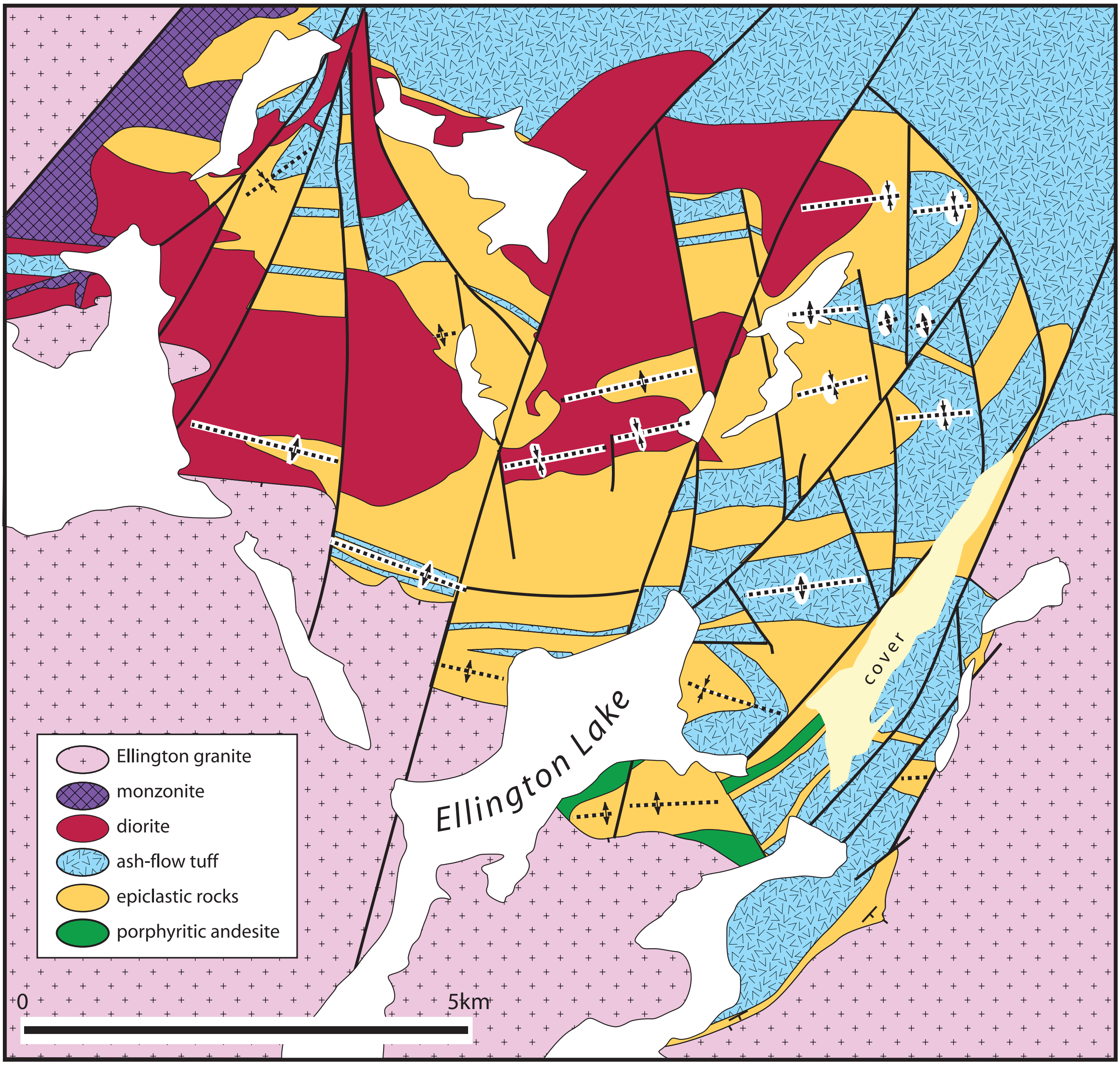



**Figure 6. Geological sketch map showing diorite and part of the Ellington pluton near Ellington Lake after Pelletier (1988). Note that supracrustal rocks dip beneath the flat floor of the Ellington pluton.**

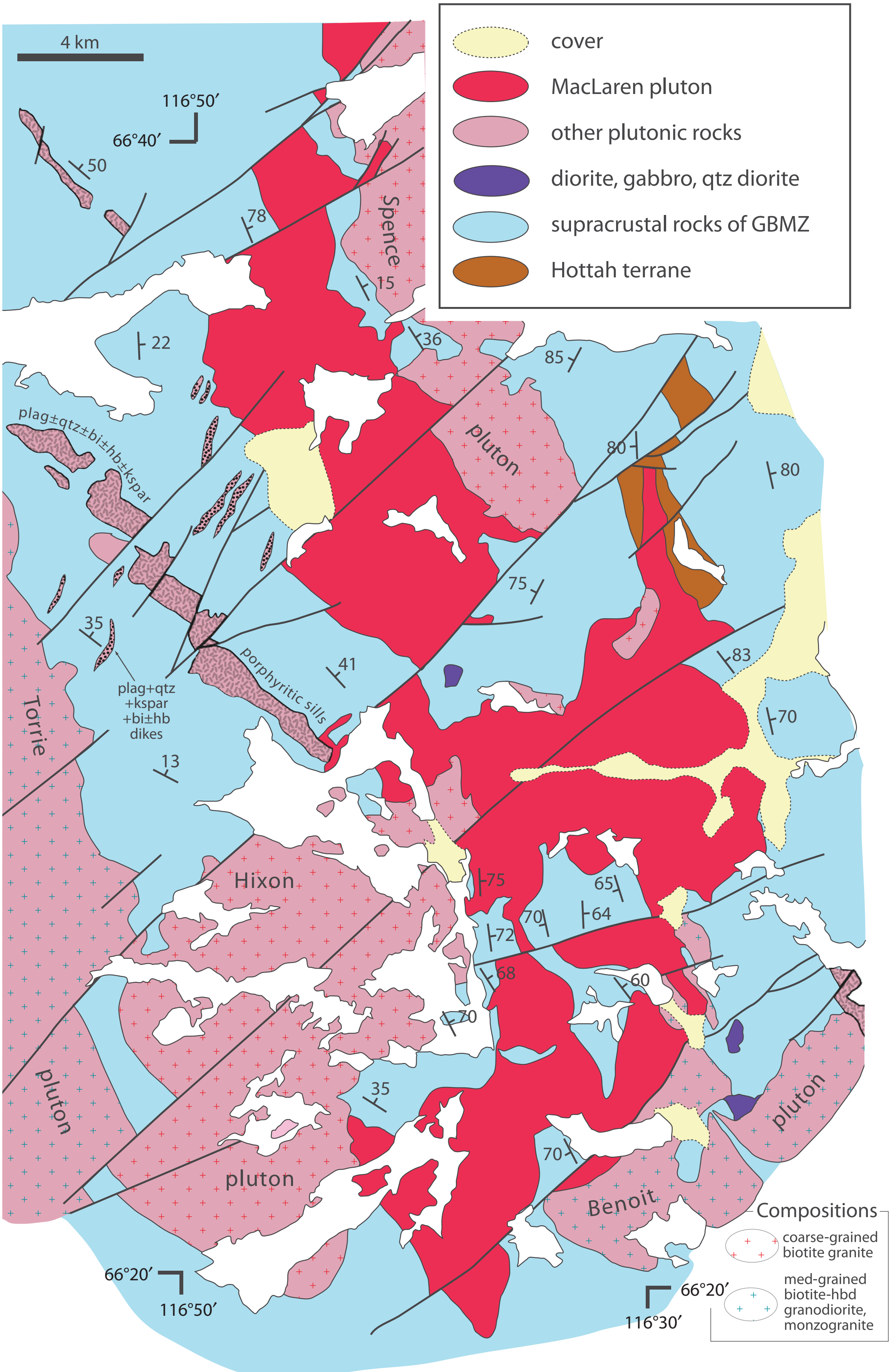



**Figure 7. Geological sketch map of the MacLaren pluton (Housh, 1989).**

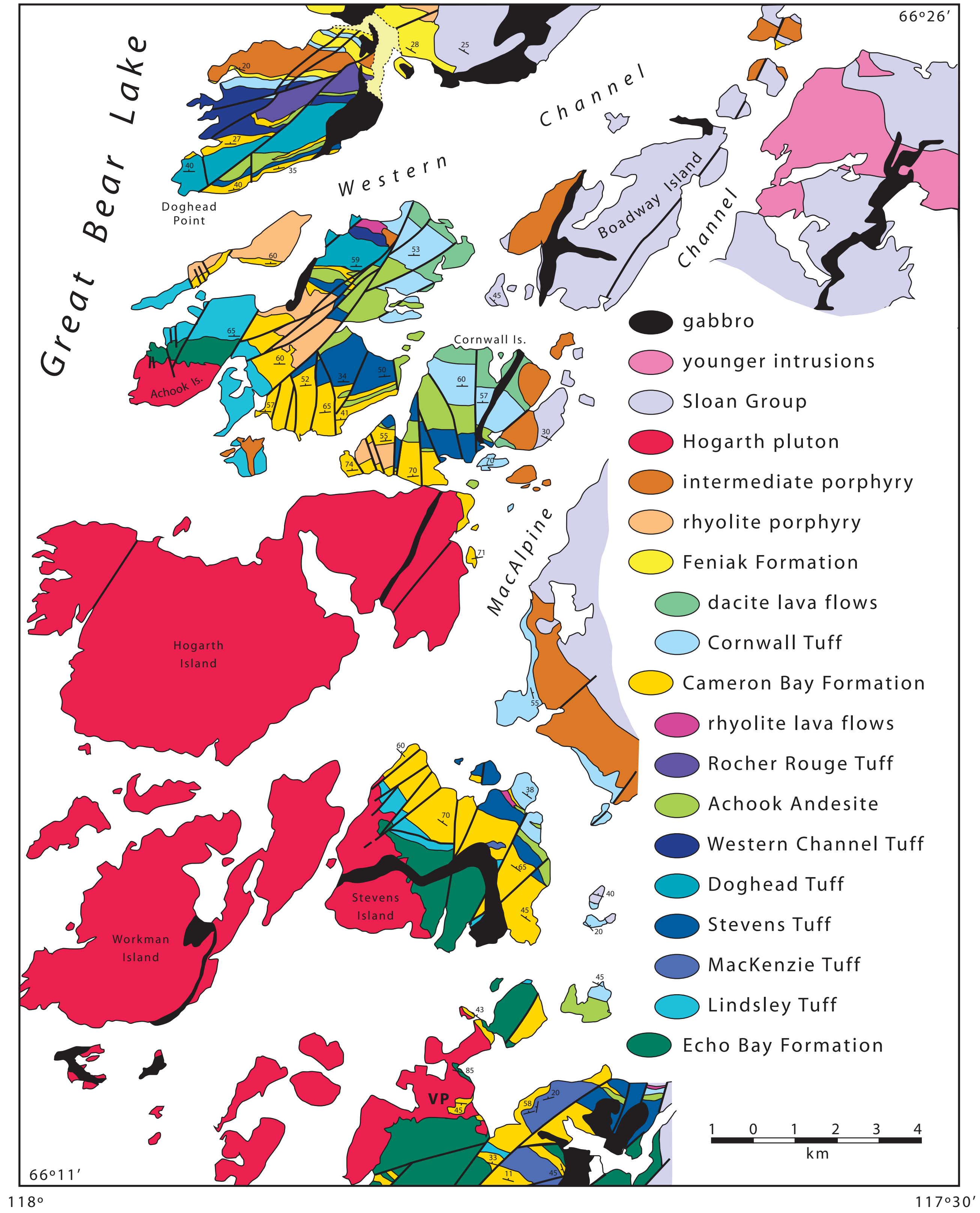



**Figure 8. Geological sketch map of the Hogarth resurgent pluton. Except for VP–Vance Peninsula, where there are folds involving the Hogarth pluton, the area represents a faulted north-northeastwardly-dipping homocline. On Stevens and western Achook islands the pluton intrudes lower stratigraphic levels than on Cornwall and eastern Achook islands, where there is a series of northerly-trending block faults interpreted to have formed when the pluton intruded and domed intracauldron fill of Cornwall caldera (Hildebrand, 1981; 1984a).**

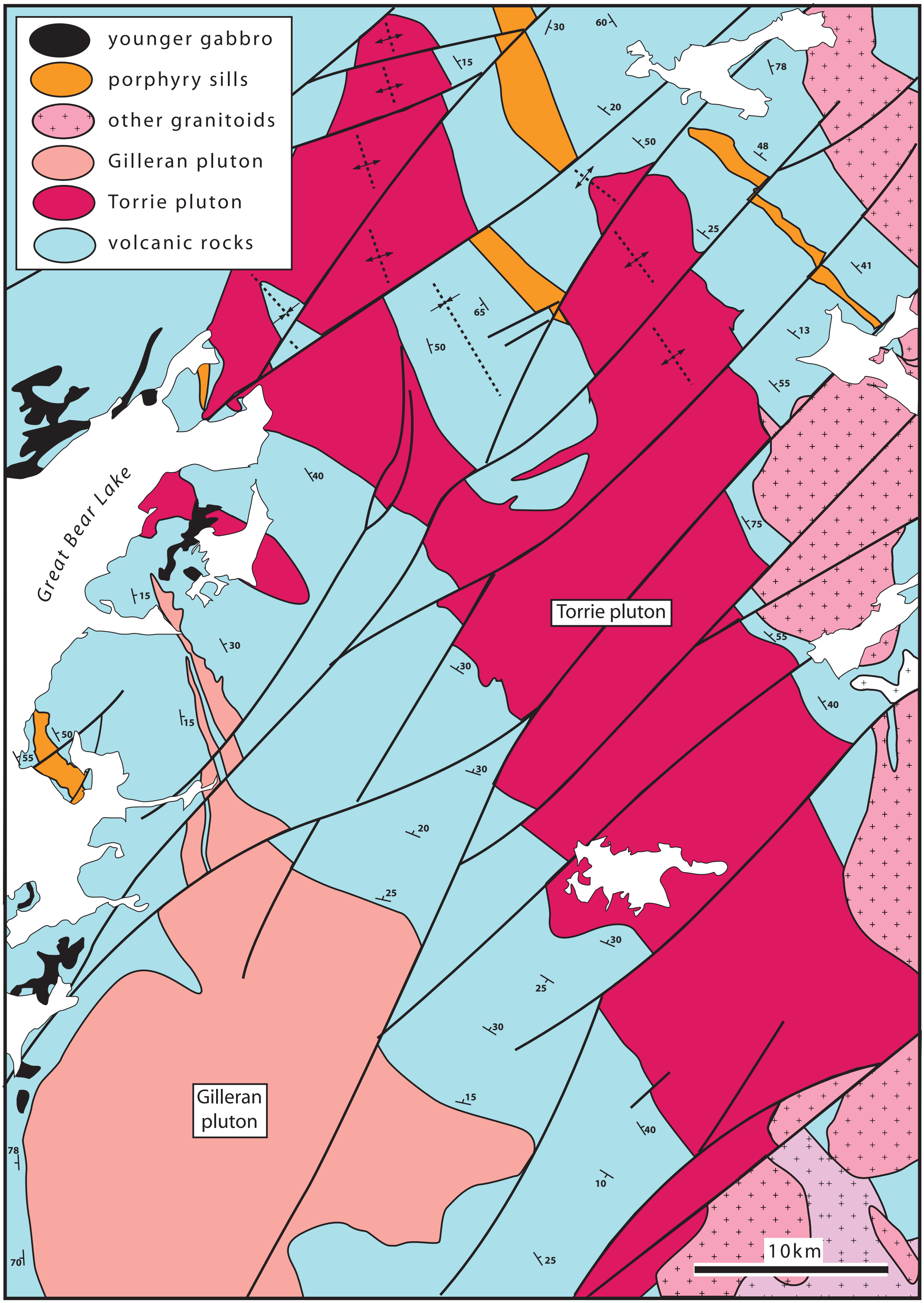



**Figure 9. Geological sketch map of the Torrie and Gilleran plutons (Hoffman, 1978, 1982).**

ate compositions occur in virtually every body (Fig. 2 P, Q, R, S). In fact, members of this suite have internal modal (Fig. 10) and chemical (Fig. 11) variations nearly as great as the entire batholith. Most members of this suite are characterized by seriate plagioclase and 5%–25% subhedral to euhedral biotite and hornblende with hornblende predominating (Hildebrand et al., 1987a). Anhedral quartz and potassium feldspar fill interstices between plagioclase and ferromagnesian minerals. A few percent clinopyroxene, typically rimmed by hornblende, are locally common. Slender needles of apatite, euhedral zircon and sub-euhedral titanite are ubiquitous accessory minerals. Titanomagnetite ± ilmenite are the dominant opaque phases. The compositional variations between phases within the plutons are gradational and occur over tens of centimeters to tens of meters. Obvious signs of basalt-granite mingling were only found in one small area. There, mafic pillows with well-developed selvages (Fig. 2Y, a) occur in more siliceous rocks.

There is ample evidence in the form of xenoliths and net vein breccias, at least at the final site of pluton emplacement, for wall-rock stoping. In general, the stoped blocks are mostly seen adjacent to the roofs (Fig. 2f, g, I, k), and as there are no accumulations of blocks on the floors, must have been digested. Locally, where plutons cut quartz arenite, they are much more quartz rich apparently owing to disaggregation of the sandstone and incorporation of individual grains (Fig. 2e). Only along the margin of one pluton did we find discontinuous layers, 1–10 cm thick, of variable mineralogical proportions. The layering is more or less parallel to the margins and truncations of successive layers are common (Fig. 2X).

In a detailed study of several of the plutons, Housh (1989) documented that there is no simple relationship between rock types and suggested that convection was not strong enough, or perhaps did not last long enough, for the body to homogenize the batch(es) of magma that filled the chamber or redistribute them consistent with their densities. As such, the variations are due to either complex melting and/or mixing processes or heterogeneities in the source region.

### Granites

The younger granites are mainly biotite granites and, in contrast to other plutons of the zone, are rather homogeneous bodies (Fig. 2T, U, V). Many of the plutons are strongly porphyritic, containing subhedral to euhedral potassium feldspar phenocrysts to 5 cm in a fine- to medium- grained matrix of anhedral quartz, feldspar, and tiny flakes of biotite. Prismatic hornblende is locally present but everywhere less abundant than biotite. Titanite is rare, but allanite, zircon, apatite, magnetite, ilmenite, monazite, and flourite are locally common accessory minerals. Most members of the suite are slightly corundum normative.

The plutons of this suite differ from all others within the zone in that they commonly have zones with abundant pegmatite, miarolytic cavities, granophyre and aplite (Fig. 2j). Additionally, regionally extensive dike swarms appear in many cases to emanate from individual plutons of this suite (Hildebrand et al., 1987b).

Contacts with wall rocks are everywhere sharp (Fig. 2Z, d). Locally there are spectacular sill complexes (Fig. 2b, c) or swarms of partially digested wall rocks (Fig. 2 h, j). The intimate nature of the sills and their compositional similarity to the main body suggest that the magma was highly fluid and little crystallized at the time of emplacement. This is different from the older three magmatic suites, which, judging from the porphyritic nature of their quenched border phases, were partly crystalline.

The younger granites have contacts that tend to be more discordant to supracrustal rocks than the older, more intermediate-composition plutons. Numerous plutons of this suite have lengthy intrusive contacts that parallel bedding in their wall rocks but in places the contacts climb up or down stratigraphy on the scale of kilometers (Figs. 1, 5, 12).

In the southern part of the mapped area—and based upon the more extensive outcrops of Hottah basement and lowermost Great Bear stratigraphy relative to the north—slightly deeper levels of the zone are exposed and granites are much more common (Fig. 1). The contacts of the granites tend to parallel bedding in supracrustal rocks and parallel axial traces of folds even where the traces swing around to a northerly orientation along the eastern margin of the zone (Fig. 1). Such a geometry could only occur if the plutons were folded and this is reflected in the common northwesterly trends of their contacts. However, the granites also have long stretches where their contacts trend orthogonal to the axial traces of the folds; that is, they strike northeast where axial traces of folds trend northwest. The only way to satisfy such a geometry is for the granitic plutons to be much thicker than more intermediate-composition plutons and have steep side-walls. Therefore, they may be more analogous to the bell jar plutons of other workers (Cobbing et al., 1981; Bussell and Pitcher, 1985; Pitcher, 1993).

## SHAPES OF PLUTONS AND DEPTH OF EMPLACEMENT

Magmas of nearly every composition, except the biotite granites, created sill-like plutons, 500m-4km or so thick, with aspect ratios of five to ten or even greater. Intrusions related to the early basaltic eruptions have aspect ratios in the range 10–15. Those magmas, which form dikes and inclined sheets where they intrude basement and older supracrustal rocks, only form sills at the highest structural levels. Plutons of the early intermediate intrusive suite form sheet-like bodies both within and at the base of possibly comagmatic andesitic stratovolcanoes. Those plutons have aspect ratios of 5–10 at a minimum. The monotonous granodiorite-monzogranite bodies, which intrude the thick dacitic ash-flow tuff sheets, have aspect ratios of 10–30. Only the granites appear to be thicker, although due to the lack of supracrustal rocks

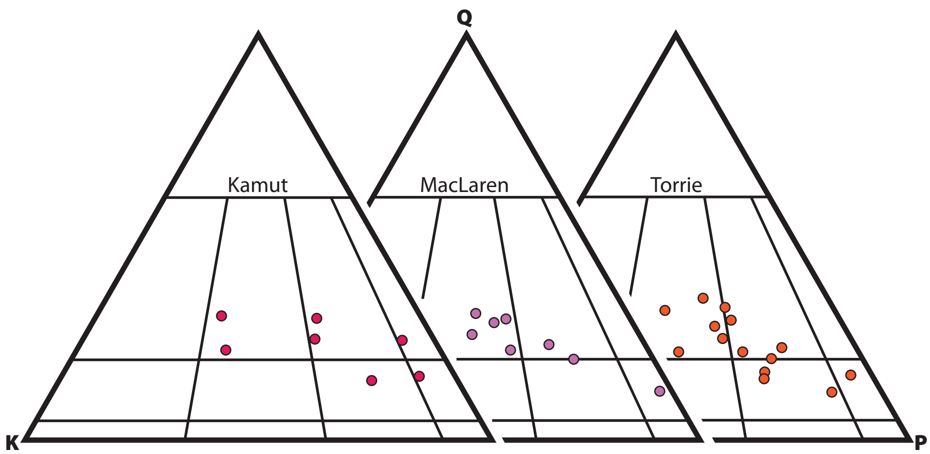



**Figure 10. Modal analyses of three plutons illustrating the compositional variations typical within bodies of the monzogranite-granodiorite suite. Modified from Hildebrand et al., 1987a.**

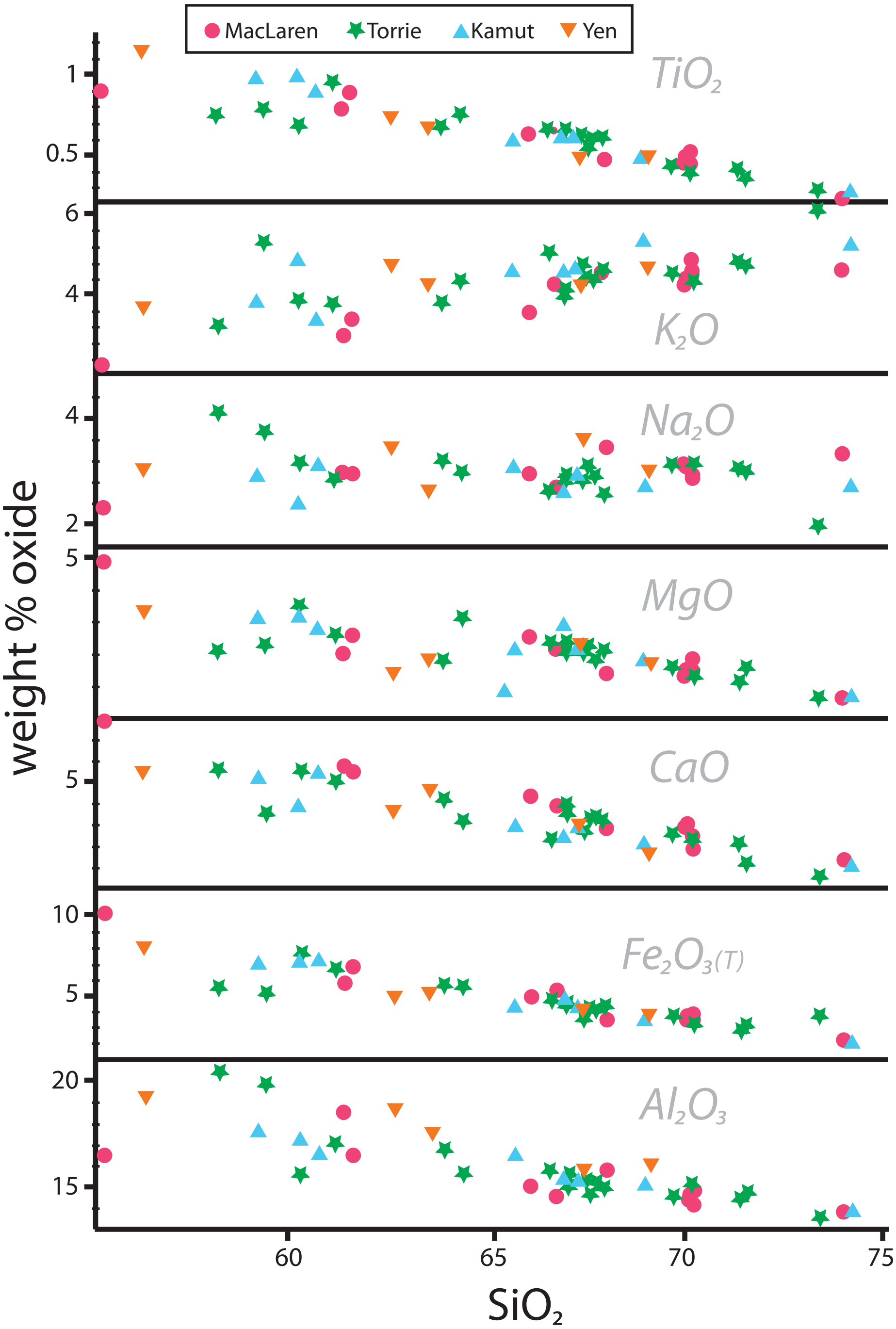



**Figure 11. Harker diagrams for four plutons of the granodiorite-monzogranite suite illustrating the breadth of compositional variations within individual plutons. The dominant lithology in each of these bodies is granodiorite-monzogranite but locally compositions are highly variable and overall $SiO_2$ in each varies more than 20%. Modified from Hildebrand et al., 1987a.**

in their zone of intrusion, accurate thickness estimates were difficult to obtain. An estimate made from cross sections suggests that they have aspect ratios of ~2–4. None of the bodies were obviously composite; that is, sharp internal contacts caused by multiple intrusions or sheets, were not recognized.

The older, more intermediate and mafic composition plutons tend to feather out laterally in the paleohorizontal plane, whereas in many cases the younger granites appear to have steep sidewalls. Because the floor typically parallels the roof of these plutons there is no evidence that the plutons are triangular with inclined floors and flat roofs.

From the amount of volcanic cover above plutons, other than the biotite granites, they were emplaced 1–4 km beneath the surface. Rocks within the alteration haloes of the early intermediate intrusive suite yield paleodepths, based on paleofluid compositions preserved in inclusions, of a km or less (A.K. Somarin, 2010, personal commun.). Several of the plutons were deroofed shortly after emplacement, so the exact amount of cover is unknown as it was eroded away; but using stratigraphic reconstructions and unconformities, similar emplacement depths seem reasonable.

In spite of voluminous volcanic eruptions and excellent exposures of the roofs of scores of plutons we do not generally see cupolas above their roofs. The upper contacts of most plutons are remarkably planar. Only in the cases of two resurgent calderas and the Contact Lake pluton, do we see dome-like or laccolithic plutons, but if they are part of a larger body it must have been much deeper as it is not exposed.

The Great Bear magmatic zone and, in fact, all of Wopmay orogen appears to have a slight northward plunge created by much younger events, likely Phanerozoic in age. An interesting observation that came out of our studies is that the biotite granite bodies are generally concentrated at apparent lower structural levels than the more intermediate composition plutons. This can be seen by examining Figure 1, and noting the prevalence of granites in the southern and northeastern portions of the study area. There they intrude the lowest stratigraphic units of the Great Bear magmatic zone and its basement. Based only on density, the granites should have been buoyant enough to rise to any level, but they didn't. It may be that they were too viscous to rise any farther, or that they contained more water than other bodies, and so rapidly exsolved volatiles and froze. Because many of the granitic plutons contain miarolytic cavities lined with pegmatite, and because the magma that formed dozens of rhyolite porphyries, many quite substantial in size, did rise to high levels in

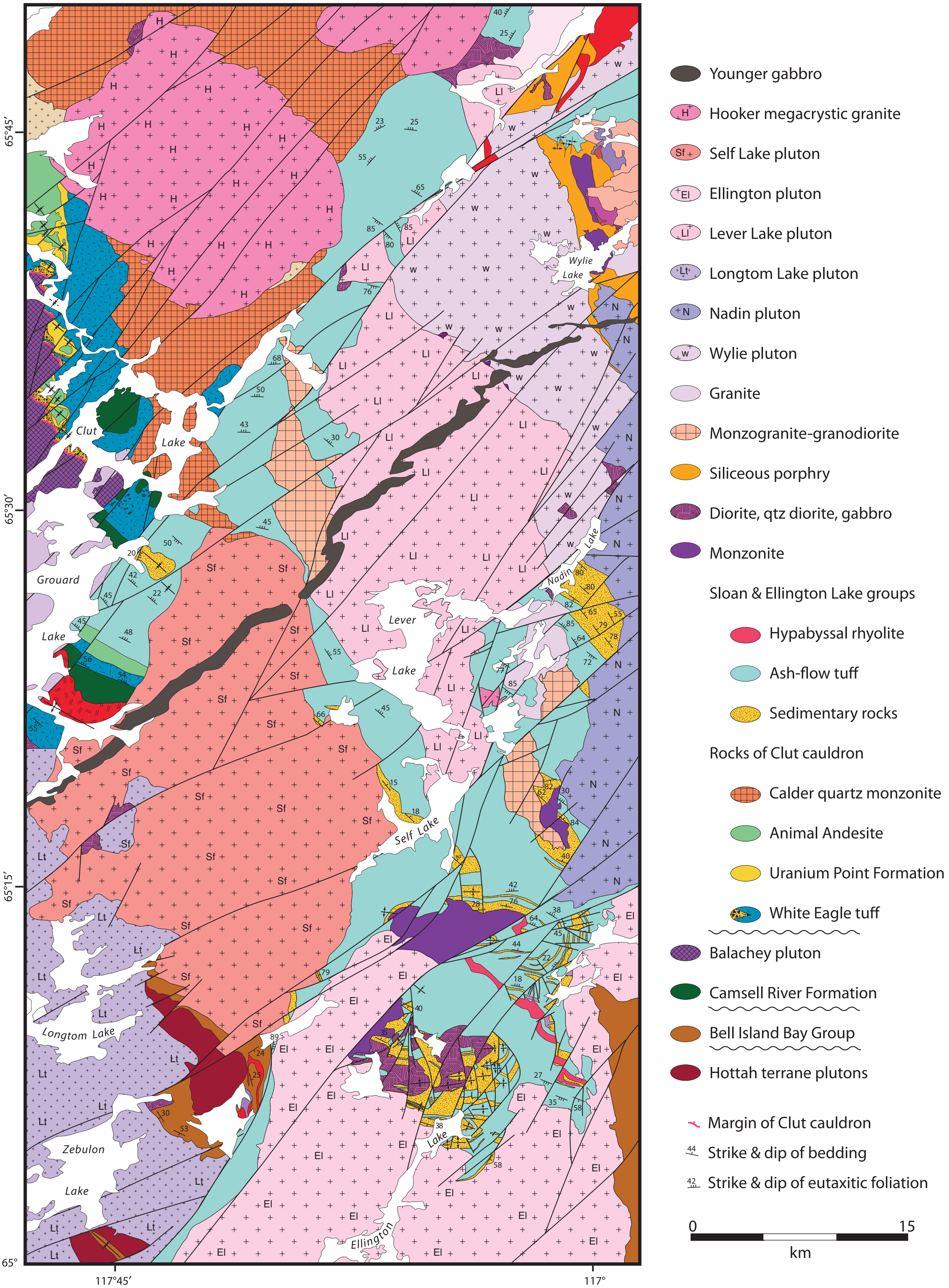



Figure 12. Geological sketch map showing central part of the Great Bear magmatic zone (see Figure 1 for location) and relations of granitoid plutons discussed in the text.

the volcanic suprastructure, the second hypothesis is more probable.

We believe a major control on the shape of plutons in the epizone is lateral spreading and elevation of their roof rocks because (1) it is mechanically much easier than depressing their floors (Paterson and Fowler, 1993; Petford et al., 2000); and (2) because we don't see inclined floors of plutons (Cruden, 1998). Because the intermediate composition bodies are able to spread out laterally in thick sequences of densely-welded tuff or accumulations of andesite flows, well-defined bedding planes do not appear critical to this process. However, when one compares plutons of the Great Bear that intrude bedded rocks versus more massive units, the contacts are invariably much more planar in detail where they intrude the former.

Supporting these ideas are the type of sills and enclaves found adjacent to the roofs of the plutons themselves. Where the wall rocks are well-bedded, there are typically meter-to centimeter thick sills of intrusive rock and flat enclaves formed as the plutons spall off pieces of wall rock; but where the wall rocks are more massive, enclaves are irregular, and if equigranular igneous rock, are blocky and equant. Where deformed metasedimentary rocks constitute the wall rock, the enclaves are quite irregular and blob-like in overall form (Fig. 2 f, g, i, k).

We nowhere saw evidence near these plutons for forceful emplacement at the outcrop scale. There simply aren't folds, faults, or ductile features at that scale that one can relate to the emplacement of the plutons, even though in many cases plutons intrude bedded rocks that would show deformation. In cases of resurgent plutons in calderas, such as the Hogarth and the Calder, it is clear the plutons were able to lift their roofs, but in the more typical case a pluton is many kilometers to tens of kilometers across and despite superb outcrop we nowhere saw evidence that they depressed their floors. Overall, the plutons were emplaced at such shallow levels in the crust, that it is difficult to envision that they did anything but lift their roofs.

## COMPOSITIONAL VARIATIONS

Most of the studies of pluton emplacement and temporal development were done in mesozonal or deeper rocks (Matzel et al., 2006; Coleman et al., 2004; Glazner et al., 2004; Miller et al., 2007) and may or may not be applicable to epizonal plutons, their processes, or even their timescales. Working downwards, some workers attempted to reconstruct plutonic processes from the volcanic realm (Smith, 1979; Bachmann and Bergantz, 2004, 2008; Bachmann et al., 2002, 2007; Hildreth, 2007; Reid, 2008). The oblique crustal sections of the Great Bear magmatic zone span the two and so offer an uncommon opportunity to evaluate and test models of magmatism in the epizone where volcanism and plutonism are presumably connected.

Several of the granodioritic-monzogranitic plutons are more than 100 km across, and while there are typically internal variations down to the scale of a hand specimen, there are no obvious intrusive contacts or sharp compositional breaks. In a few places near the wall-rock contacts there are compositional layers (Fig. 2X), but in general they are rare. Housh (1989) studied several plutons and using a combination of Pb and Nd isotopes, petrography, and whole rock chemistry, corroborated the field observations by demonstrating that they are multicomponent mixes. However, he was unable to discern where these mixes were assembled or how. The processes that fill a 100 km wide pluton are simply unknown. Are they open systems, filled incrementally, in a series of pulses; or are they inflated once and then remain as closed systems, or does some combination of the two operate? Open system behavior would seem to fit the data so long as the pulses were frequent enough not to allow sufficient cooling for sharp internal contacts to develop; yet it seems as though incremental pulses (Miller, 2008; Miller et al., 2009) couldn't reproduce the seemingly random compositional variations throughout the bodies. The general lack of cumulate or anti-cumulate layers in these large plutons suggests that either the bodies did not have enough liquid to allow crystal fractionation (sinking, floating, filter pressing), or possibly that convection was particularly vigorous. The problem with strong convection is that the bodies are not well mixed and there is typically no layering at the margins. We suggest that the plutons were likely emplaced more or less at one time as crystal mushes with ~40% plagioclase and hornblende phenocrysts, then rapidly froze, preserving internal heterogeneities and inclusion trains.

Most of the large ash-flow tuff sheets within the LaBine group are compositionally zoned, yet for the most part, we do not see obvious systematic compositional zonation or compositional layering within plutons such as the Hogarth or Calder bodies (Table 1). Small volume crystal-poor rhyolitic lavas and porphyritic intrusions suggest that some separation of liquid and crystals took place, but for the most part, even rhyolitic ignimbrites contain 10%–25% broken phenocrysts. Ignimbrites of the Sloan group are weakly zoned, if at all, and are best classified as monotonous intermediates (Hildreth, 1981), yet generally they do not have the composition variability that exists in many of the monzogranite-granodiorite plutons that intrude them. Both compositionally zoned and monotonous intermediates are common in younger belts as well, yet compositionally-zoned subvolcanic plutons are also apparently rare (Lipman, 2007).

We mapped ~20,000 km$^2$ of the Great Bear magmatic zone, and obvious field evidence for magma mixing and mingling is sparse, unless one relates the internal compositional variations to nearly complete mixing of different magmas. There are exceptions, however, and in the best-exposed area, only a few hundreds of meters across, obvious mafic pillows with well-developed selvages occur in more siliceous rocks (Fig. 2Y, a). Also common in that area are mafic orbicules (Fig. 2W), coarsely-poikiolitic gabbro (Fig. 2O), and granodiorite-quartz monzonite choked with mafic inclusions. These relations attest to the invasion of a siliceous system by mafic magma and likely streaming of volatiles (Moore and Lockwood, 1973). The mixed zone may be easily recognized because it was generated by direct introduction of mafic magma into a siliceous body relatively close to the solidus. Within the batholith as a whole, however, the limited occurrence of obvious mixing and mingling textures, such as pillows and blobs, suggests that *epizonal mixing* of different bodies of mafic and siliceous magma was not a very important process in constructing magmas of the Great Bear magmatic zone and possibly other arcs as well. If so, then do the common heterogeneities in the magma bodies come from bulk melting of source rocks or do different melts mix during ascent to create the observed inhomogeneities? One possibility, not often considered, is that the magmas were originally more ordered, say in a zoned chamber, but then rose upwards following the disturbance caused by partial evacuation during eruption, to become more chaotic. This is consistent with the observation that nearly every one of the intermediate composition plutons, including both the early intermediate suite and the granodiorite-monzogranite suite, vary widely in composition and contain ~40% seriate plagioclase and several percent hornblende prisms, which are readily interpreted as early crystallizing phases. Thus, it may be that there were too many crystals for the bodies to be able to convect sufficiently to mix when finally emplaced.

In addition to the general lack of evidence for magma mixing, and despite many kilometers of structural relief, we also do not see mafic magmas beneath the floors of the plutons, which indicates that any influx of mafic magmas to re-energize or remobilize the magma chambers must have taken place at deeper levels before the magma bodies rose higher in the crust. Likewise, in the Southern Rocky

Mountain volcanic field of Colorado no epizonal plutons show evidence for mixing of siliceous and mafic magma (Lipman, 2007), so it is hard to understand how magmatic underplating could represent a viable widespread possibility to rejuvenate magma bodies at the highest structural levels, even though it is recognized at some volcanoes (Bacon and Lanphere, 2006; Andrews et al., 2008). This does not mean that we don't believe that underplating is a viable process at depth, for it might readily occur there and provide a heat source to re-energize magma chambers after initial eruptions.

We generally don't see cumulate or obvious mush zones in the lower parts of the plutons. Where there are cumulate facies, such as with plagioclase in the Rainy Lake pluton, they are obvious. The general lack of cumulates suggests that magma chambers were not mostly evacuated, nor tapped of the rhyolitic melt component (Bachmann and Bergantz, 2004), to leave behind crystal accumulations.

In the entire area, we mapped only one concentrically zoned complex, the Handley Page intrusive series (Fig. 1; Table 1), which is similar to the Tuolumne Intrusive Suite (Bateman and Chappell, 1979; Bateman, 1992). The series includes diorites and minor gabbroic rocks that were intruded by younger and more siliceous bodies progressively inward to a granitic core (Hildebrand et al., 1987b; Gibbins, 1988). These types of complexes clearly have sharp contacts, which reveal age relations, and are a different type of plutonic suite than the majority of bodies in the Great Bear zone. The concentrically zoned suites were originally considered to show reasonably clear evidence for the involvement of crystal fractionation to produce some of their compositional variations (Bateman and Chappell, 1979; Kistler et al., 1986), but recent studies (Coleman et al., 2004; Gray et al., 2008) suggested major roles for other processes. Overall, the general lack of crystal-poor rhyolitic ignimbrites and the rarity of cumulate rocks in the crustal column of the Great Bear magmatic zone and elsewhere suggests that crystal fractionation is not the predominant mechanism that creates broad compositional variations within arc terranes, at least at mid- to upper-crustal levels. Varying degrees of partial melting of lowermost crustal rocks, mixing with mantle-derived basalt, and assimilation of wall rocks during ascent and emplacement might best produce the compositional variations found in continental arcs.

## RELATIONSHIP OF PLUTONISM TO VOLCANISM

When one looks at the relationships between plutonism and volcanism in the Great Bear magmatic zone, an interesting pattern emerges. After each eruption type, plutons of similar composition rose into the volcanic suprastructure. At their current level, the plutons, which by every measure appear comagmatic with their wallrocks, cannot have fed them because they intrude them and are thus demonstrably younger. This cyclical relationship repeats itself over most of the history of the zone.

The oldest volcanic rocks of the Great Bear magmatic zone are the tholeiitic basalts (McGlynn, 1975), which were intruded by comagmatic sill-like plutons. The next phase of development was a mixed assemblage of caldera-forming ignimbrite eruptions ranging in composition from dacite to rhyolite and andesitic stratocones. The calderas were intruded by resurgent plutons of similar composition, whereas the andesitic piles were intruded by plutons of the early intermediate intrusive suite. In both cases, the plutons are strikingly similar to their wallrocks, but are younger. Following the mixed caldera-stratocone phase were voluminous eruptions of dacitic-rhyodacitic ash-flow tuff preserved in the central part of the zone as the Sloan group. The voluminous granodiorite-monzogranite plutons intruded the dacites-rhyodacites, and are mineralogically similar to them, but appear to have greater compositional variability.

Following the large-volume dacitic eruptions and emplacement of the granodiorite-monzogranite suite, there were only a few minor eruptions of siliceous lava and tuff; but abundant large tabular plutons of biotite granite were emplaced near the base of the volcanic pile. These plutons have few known eruptive equivalents, despite many having dike swarms. It is, of course, possible that the volcanic rocks associated with the granitic plutons were eroded, but as the entire Great Bear magmatic zone was folded soon after cessation of magmatism (Hildebrand et al., 1987a) it seems unlikely that there would be few such rocks preserved.

Thus, there are two trends to explain: (1) the overall development of the zone from small-scale eruptions of basalt to voluminous granodiorite-dacite magmatism followed by emplacement of abundant biotite granites (Fig. 13); and (2) the smaller-scale rise and emplacement of plutons into what for all appearances appears to represent their own volcanic ejecta (Fig. 13). These smaller-scale events are independent of composition as they occur in basalts, andesites, dacites, and rhyodacites.

The smaller scale trends of magmatism involving the emplacement of plutons into their own ejecta reveals why it is so difficult to directly relate particular plutons to a given volcanic eruption or sequence of eruptions; not only in the Great Bear magmatic zone, but also elsewhere (Walker et al., 2008). Whereas the plutons clearly rose into their respective volcanic suprastructures and therefore must postdate the volcanic eruptions, two possibilities are obvious: (1) the magma chambers from which the plutons were derived erupted the magmas and then continued their rise toward the surface with smaller volumes and fewer volatiles; or (2) the magma bodies erupted from depth and it was entirely different bodies of magma that later rose and were emplaced into the volcanic suprastructure. We prefer the first possibility because we don't see similar plutons or deflated chambers at depth, and because the plutons we do see don't seem to have erupted in situ, so were likely partly drained of volatiles during previous eruptions. If correct, it means that we are, in fact, looking at bodies of the same magma that erupted at the surface, but at a slightly later stage in their evolution. This implies that the magmas were moderately mixed and didn't fractionate between their initial eruption and their final emplacement. It also implies that there were discrete magma batches that rose from depth, lifted their roofs, and spread laterally. In many cases the plutons were rapidly deroofed. It is difficult to envision these chambers, lacking in sharp internal contacts, starting out as thin sills that were progressively assembled by crack-seal growth or other relatively small-volume replenishment mechanism (Coleman et al., 2004; Glazner et al., 2004; Bartley et al., 2008; Annen, 2009; Miller et al., 2009). Furthermore, as pointed out by Lipman (2007), progressive growth by crack seal appears to preclude large volume eruptions as there would never be enough melt to feed such eruptions.

If the magmas feeding the plutons erupted at the surface, how could they later rise into their own volcanic ejecta, given the obvious loss of volatiles during eruption? The important point is that eruptions reflect episodes of degassing within the upper parts of magma chambers. That is, the eruptions don't evacuate the entire magma chamber, but merely draw from its uppermost portions (Smith, 1979). For example, a circular pluton 50 km in diameter with an aspect ratio of 10 would contain nearly 10,000 $km^3$ of crystalline mush and magma, so a major ash-flow eruption of 1000 $km^3$ would only lead to a 10% drawdown of the chamber.

With re-accumulation of volatiles within the chamber, the magma may erupt again, or if volatiles are not concentrated, say near the roof, the magma may rise toward the surface, and thus have the possibility to intrude its own ejecta. Due to the increased crystallization, solidification, and loss of volatiles, subsequent eruptions would probably be smaller and less *plutons*

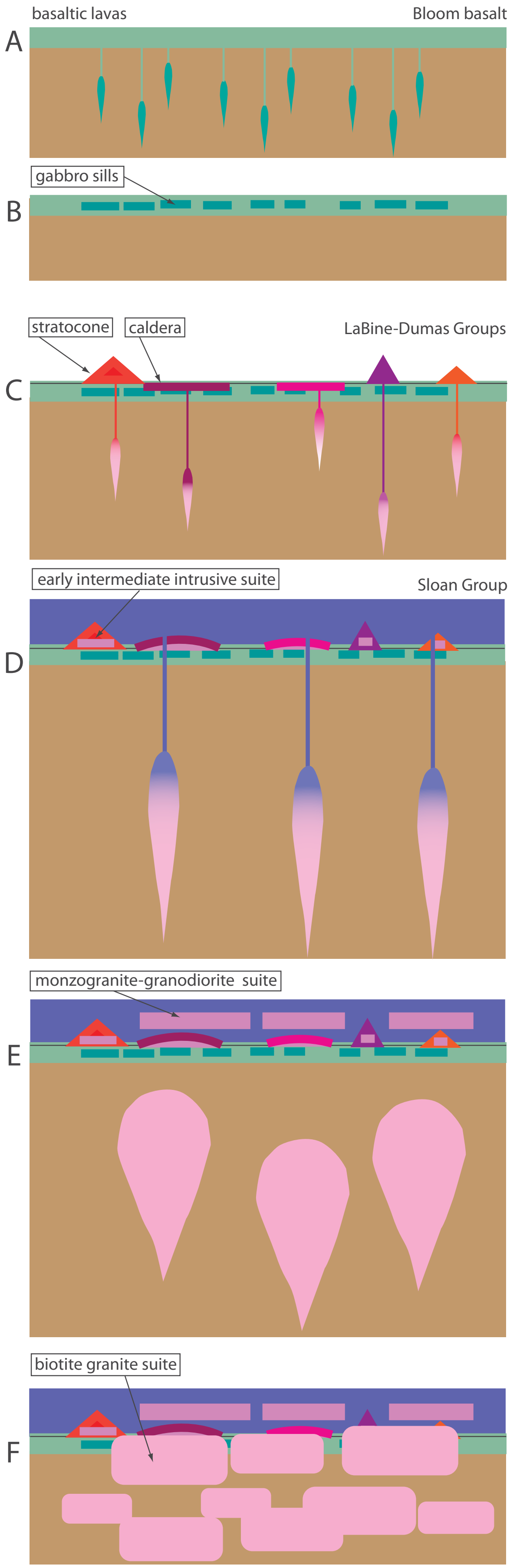



**Figure 13. Interpretive evolution of the Great Bear magmatic zone: (A) lavas of Bloom basalt are erupted from magma chambers that are either not seen or preserved; (B) basaltic magma from the previously-erupted (partially evacuated?) chambers migrates upwards to intrude their own ejecta, where they form sills; (C) new intermediate and siliceous magma batches erupt from compositionally-zoned magma chambers to form ash-flow tuffs and lavas; (D) re-energized magmas from these chambers rise and intrude their erupted equivalents—calderas and stratocones—where they form domical resurgent plutons in the calderas and sill to laccolithic-shaped masses in and just beneath the stratocones; new, and larger, compositionally zoned magma bodies erupt voluminous ash-flow tuffs of the Sloan Group; (E) re-energized magma from the chambers that erupted the tuffs of the Sloan Group migrates upwards to intrude their own ejecta and form large sheet-like plutons of compositionally heterogeneous, but dominantly granodiorite and monzogranite composition; (F) large bodies of granite transgress the crust; the young, and large, granitic bodies apparently didn't typically vent at the surface, and rose to about the level of the Hottah-Great Bear contact, before freezing. The observations that magma batches over the compositional range basalt to rhyodacite, intrude their own ejecta, leads us to hypothesize that the continued rise of formerly compositionally zoned and partially crystalline magma partially mixes the remaining magma such that, when emplaced into its much cooler ejecta, new composition gradients do not have time to form. This explains why it is so difficult to match plutons with their eruptive products.**

explosive, if they occurred at all. Where these bodies of magma formed calderas, their subsequent rise into the caldera produces central plutons and resurgent doming of the intracauldron-facies tuff.

A possible method to re-energize a magma chamber without physically introducing basalt into the chamber itself comes from detailed study of the Soufriere Hills volcano in Monserrat where Couch et al. (2001) described a process called convective self-mixing. In that process a magma body is heated from below, but the two magmas themselves never mix. Instead, plumes of the lowermost part of the upper magma body are heated from below and become gravitationally unstable and rise into the overlying magma. This leads to intermingling of thermally different parts of the same magmas body.

The main problem in applying any model in which basalt provides a direct heat source to re-energize previously and partially evacuated magma chambers is that we don't see extensive bodies of gabbro, except in the oldest part of the zone. One way around this conundrum is for the energizing culprits to have been rising magmas of intermediate composition, similar to the original melts in bulk composition. In that way, different bodies of intermediate composition magmas can mix within the partially evacuated chamber, and leave sparse evidence, perhaps only disequilibrium mineral zoning and resorption and/or isotopic variations.

An additional contributing factor—which could operate in concert with partial drawdown—is the concept of gas sparging, a process in which gases streaming upwards from deeper magmas rejuvenate magma bodies (Bachmann and Bergantz, 2003, 2006). Recharging bodies of wet, crystal-rich magma with volatiles and heat could add enough energy to assist their rise into the volcanic suprastructure (Huber et al., 2010).

Except for the late granites, we didn't find evidence for eruptive conduits such as vesiculated dikes above plutons that would suggest that the bodies erupted after or during final emplacement. The lack of miarolitic cavities, except in the younger granites, suggests that hornblende and biotite were able to incorporate the bulk of magmatic water contained within the bodies.

Whatever the cause of their remobilization, it appears that many magma bodies in the Great Bear magmatic zone erupted then migrated toward the surface to intrude their own ejecta. This process might explain why elsewhere it is typically so difficult to match a pluton with its eruptive products. Any regular chemical and mineralogical zoning that formed in the body just prior to its original eruption was redistributed, stirred, or mixed during its subsequent rise toward the surface, and when the magma arrived in its final plutonic configuration it was too crystalline to convect sufficiently to mix well. Furthermore, because the plutons are essentially large sills, they would have cooled rapidly, which would also have tended to quell convection or fractionation. The dominance of strongly variable compositions demonstrates that, even though many bodies are of sufficient volume, they could not have fed the monotonously intermediate ignimbrites of the Sloan Group from their present locations.

In addition to our field data presented here, Christiansen (2001) developed a model for the Yellowstone volcanic field in which, following major ash-flow eruptions, magma chambers quickly reestablished positive magma pressure and continued to rise buoyantly. More recently, Lipman (2007) pointed out two lines of evidence that suggested to him that magma bodies must erupt at depth then migrate upwards: (1) geological and geophysical data are consistent with the subvolcanic batholith beneath the central San Juan mountains migrating upwards *after* ignimbrite eruptions and collapse because it is not exposed in the floors of nested calderas that cumulatively had 10–15 km of subsidence; and (2) Quaternary calderas in the western U.S., the Andes, and elsewhere don't have shallow geophysical signatures that would indicate voluminous shallow chambers of largely liquid magma beneath them.

## OVERALL EVOLUTION OF THE ARC

The evolution of magmatism within the Great Bear magmatic zone has features in common with other continental arcs (Hildebrand and Bowring, 1984; Hildebrand et al., 1987a) and although it is obvious that no single scheme can explain all the variation and styles in magmatism observed at convergent margins, some broad generalities have emerged. Magmatism started with outpourings of basalt, but volumetrically minor dacites were erupted toward its end. The basaltic phase was followed by a stage where andesitic eruptions were punctuated and intermingled with ash flow eruptions and caldera collapse at eruptive centers. As magmatism continued, it developed into more substantial eruptions of intermediate composition ash-flow tuffs with only minor andesite and rhyolite. Within the Great Bear magmatic zone the final stage of magmatism was the rise and emplacement of tabular granitic plutons.

Obviously, given the complexities of folding and faulting with the Great Bear magmatic zone, we cannot reasonably estimate magmatic flux in either the crust or mantle over time, but the overall development of the zone, from basalt to granite, suggests progressive heating and melting of the crust beneath the arc such that larger volumes of crustal material were processed with time. In an early review of this process Hildebrand and Bowring (1984) concluded that most continental arcs have their base near sea level, which suggested to them that the crust was not appreciably thickened despite the overall influx of basalt. They proposed a model in which the material lost from the region by airfall roughly equaled the mass of basalt arriving at the base of the crust. To this effect must be added any regional extensional effects such as, for example, those caused by rollback of the subducting slab. A survey of currently active continental arcs, such as occur in the Cascades, Japan, New Zealand, Central America, the Kamchatkan, Alaskan and Antarctic peninsulas, the Philippines, Italy, and Sumatra, shows that most remain at or near sea level, albeit with individually high-standing volcanoes, for tens of millions of years, despite continual subduction of oceanic lithosphere and its abyssal veneer of sediment (Hildebrand and Bowring, 1984; Hildebrand, 2009b). This is different than Cordilleran-type batholiths (Hamilton, 1969a, b), such as the Sierra Nevada and Coast Plutonic batholiths, which are flare-ups (Ducea, 2001; de Silva, 2008) characterized by nearly complete crustal re-organization and extreme crustal thickening. They likely owe their origins to a somewhat different process than that caused purely by subduction of oceanic crust and its abyssal cover (Hildebrand, 2009; DeCelles et al., 2009).

## CONCLUSIONS AND IMPLICATIONS FOR OTHER CONTINENTAL ARCS

The folded 1875–1844 Ma Great Bear magmatic zone exposes scores of plutons, ranging in composition from basalt to granite, with varying degrees of exposure of their floors and roofs. Over the full range of compositions the plutons have dominantly flat roofs and floors. The major exceptions are a few laccolithic bodies and resurgent plutons in the cores of calderas, which have domical roofs. In the Great Bear most plutonic bodies have aspect ratios of 5–15 over the compositional range of basalt to rhyodacite. Only the biotite granites tend to be thicker, but they still have aspect ratios greater than unity.

Epizonal plutons in the Great Bear magmatic zone do not appear to have erupted. However, their likely precursors—more or less equivalent bodies located at greater depths—produced monotonously-intermediate and compositionally zoned eruptions in the form of ash-flows and unzoned eruptions as lavas. These eruptions apparently did not severely deplete the magma chambers, so following eruption, the magma bodies rose into their own ejecta. This could happen because the initial eruptions did not draw down the magma chamber sufficiently to cause them to stagnate. Instead they had the ability to regain positive magma pressure, perhaps because the previous eruptions only drew down a small percent of their volume, or because they were re-energized either by new magmas or by gas sparging. While epizonal plutons are close compositionally and petrographically to their wall rocks, they did not feed those eruptions from their level of emplacement, but instead they represent the final phase in the dynamic lives of complex magmatic systems.

**ACKNOWLEDGEMENTS**

We'd like to thank our numerous field assistants for their excellent mapping, and our competent bush pilots, who kept us safe in the air and on the water. This paper was near completion when Todd Housh died unexpectedly. We will miss his rigorous intellect and intense interest in geological processes. Reviews by Olivier Bachmann, three anonymous sources, and Mike Williams, resulted in a more focused and improved paper.

**REFERENCES CITED**

Aguirre, L., Pitcher, W.S., and Cobbing, E.J., 1985, Geological Map of the Coast batholith (1:500000 scale), *in* Pitcher, W.S., Atherton, M.P., Cobbing, E.J., and Beckinsale, R.D., eds., Magmatism at a plate edge: the Peruvian Andes: Halsted Press, New York.

Andrews, B.J., Gardner, J.E., and Housh, T.B., 2008, Repeated recharge, assimilation, and hybridization in magmas erupted from El Chichón as recorded by plagioclase and amphibole phenocrysts: Journal of Volcanology and Geothermal Research, v. 175, p. 415–426, doi: 10.1016/j.jvolgeores.2008.02.017.

Annen, C., 2009, Implications of incremental emplacement of magma bodies for magmatic differentiation, thermal aureole dimensions and plutonism-volcanism relationships: Tectonophysics, doi: 10.1016/j.tecto.2009.04.010.

Bachmann, O., and Bergantz, G.W., 2003, Rejuvenation of the Fish Canyon magma body: a window into the evolution of large-volume silicic magma systems: Geology, v. 31, p. 789–792, doi: 10.1130/G19764.1.

Bachmann, O., and Bergantz, G.W., 2004, On the origin of crystal-poor rhyolites: extracted from batholithic crystal mushes: Journal of Petrology, v. 45, p. 1565–1582, doi: 10.1093/petrology/egh019.

Bachmann, O., and Bergantz, G.W., 2006, Gas percolation in upper-crustal silicic crystal mushes as a mechanism for upward heat advection and rejuvenation of near-solidus magma bodies: Journal of Volcanology and Geothermal Research, v. 149, p. 85–102, doi: 10.1016/j.jvolgeores.2005.06.002.

Bachmann, O., and Bergantz, G., 2008, The magma reservoirs that feed supereruptions: Elements, v. 4, p. 17–21, doi: 10.2113/GSELEMENTS.4.1.17.

Bachmann, O., Dungan, M.A., and Lipman, P.W., 2002, The Fish Canyon magma body, San Juan volcanic field, Colorado; rejuvenation and eruption of an upper crustal batholith: Journal of Petrology, v. 43, no. 8, p. 1469–1503, doi: 10.1093/petrology/43.8.1469.

Bachmann, O., Miller, C.F., and de Silva, S.L., 2007, The volcano-plutonic connection as a stage for understanding crustal magmatism: Journal of Volcanology and Geothermal Research, v. 167, p. 1–23, doi: 10.1016/j.jvolgeores.2007.08.002.

Bacon, C.R., and Lanphere, M.A., 2006, Eruptive history and geochronology of Mount Mazama and the Crater Lake region, Oregon: Geological Society of America Bulletin, v. 118, p. 1331–1359, doi: 10.1130/B25906.1.

Barnes, C.G., Rice, J.M., and Gribble, R.F., 1986, Tilted plutons in the Klamath Mountains of California and Oregon: Journal of Geophysical Research, v. 91, p. 6059–6071, doi: 10.1029/JB091iB06p06059.

Bartley, J.M., Coleman, D.S., and Glazner, A.F., 2008, Incremental emplacement of plutons by magmatic crack-seal, *in* Petford, N., and Sparks, S., eds., Plutons and Batholiths: Transactions of the Royal Society of Edinburgh (Wallace Pitcher Memorial Issue), v. 97, no. 4, p. 383–396.

Bateman, P.C., 1992, Plutonism in the Central Part of the Sierra Nevada Batholith, California: United States. United States Geological Survey Professional Paper 1483, 186 pages.

Bateman, P.C., and Chappell, B.W., 1979, Crystallization, fractionation, and solidification of the Tuolomne Intrusive Series, Yosemite National Park, California: Geological Society of America Bulletin, v. 90, p. 465–482, doi: 10.1130/0016-7606(1979)90<465:CFASOT>2.0.CO;2.

Benn, K., Sawyer, E.W., and Cruden, A.R., 1998, Preface to special issue: extraction, transport and emplacement of granitic magmas: Journal of Structural Geology, v. 20, p. 5–9, doi: 10.1016/S0191-8141(98)00065-0.

Bowring, S.A., 1985, U-Pb zircon geochronology of early Proterozoic Wopmay orogen, N. W. T., Canada: An example of rapid crustal evolution; unpublished Ph.D. thesis, Lawrence, University of Kansas, 148 p.

Bowring, S.A., and Grotzinger, J.P., 1992, Implications of new chronostratigraphy for tectonic evolution of Wopmay orogen, northwest Canadian Shield: American Journal of Science, v. 292, p. 1–20, doi: 10.2475/ajs.292.1.1.

Buchan, K.L., Ernst, R.E., Davis, W.J., Villeneuve, M., van Breemen, O., Bleeker, W., and Hamilton, M.A., Söderlund. U., 2009, Proterozoic Magmatic Events of the Slave Province, Wopmay Orogen and Environs: Geological Survey of Canada, Open File 5985.

Busby-Spera, C.J., 1984, Large-volume rhyolite ash-flow eruptions and submarine caldera collapse in the lower Mesozoic Sierra Nevada, California: Journal of Geophysical Research, v. 89, no. B10, p. 8417–8427, doi: 10.1029/JB089iB10p08417.

Bussell, M.A., and Pitcher, W.S., 1985, The structural controls of batholith emplacement, *in* Magmatism at a Plate Edge: The Peruvian Andes; Blackie & Son Limited, Glasgow, Scotland, p. 167–176.

Cherer, R.M., 1988, Petrology and geochemistry of three plutons of the Mystery Island Intrusive Suite, District of Mackenzie, Northwest Territories: unpublished B.Sc. thesis, University of Toronto, 144p.

Chamberlin, R.T., and Link, T.A., 1927, The theory of laterally spreading batholiths: The Journal of Geology, v. 35, p. 319–352, doi: 10.1086/623419.

Christiansen, R.L., 2001, The Quaternary and Pliocene Yellowstone Plateau volcanic field of Wyoming, Idaho, and Montana: United States Geological Survey, Professional Paper 729-G, 145p.

Cobbing, E.J., Pitcher, W.S., Wilson, J.J., Baldock, J.W., Taylor, W.P., McCourt, W., and Snelling, N.J., 1981, The geology of the Western Cordillera of northern Peru; Natural Environmental Research Council, Institute of Geological Sciences, Overseas Memoir 5, 143p.

Coleman, D.S., Gray, W., and Glazner, A.F., 2004, Rethinking the emplacement and evolution of zoned plutons: Geochronological evidence for incremental assembly of the Tuolumne Intrusive Suite: California Geology, v. 32, p. 433–436.

Coles, R.L., Haines, G.V., and Hannaford, W., 1976, Large scale magnetic anomalies over western Canada and the Arctic: a discussion: Canadian Journal of Earth Sciences, v. 13, p. 790–802, doi: 10.1139/e76-082.

Cook, F.A., and Erdmer, P., 2005, An 1800 km cross section of the lithosphere through the northwestern North American plate: lessons from 4.0 billion years of Earth's history: Canadian Journal of Earth Sciences, v. 42, p. 1295–1311, doi: 10.1139/e04-106.

Cook, F.A., van der Veldon, A.J., Hall, K.W., and Roberts, B.J., 1999, Frozen subduction in Canada's Northwest Territories: Lithoprobe deep lithospheric reflection profiling of the western Canadian shield: Tectonics, v. 18, p. 1–24, doi: 10.1029/1998TC900016.

Couch, S., Sparks, R.S.J., and Carroll, M.R., 2001, Mineral disequilibrium in lavas explained by convective self-mixing in open magma chambers: Nature, v. 411, p. 1037–1039, doi: 10.1038/35082540.

Cruden, A.R., 1998, On the emplacement of tabular granites: Journal of the Geological Society, v. 155, p. 853–862, doi: 10.1144/gsjgs.155.5.0853.

Cruden, A.R., and McCaffrey, K.J., 2001, Growth of plutons by floor subsidence: Implications for rates of emplacement, intrusion spacing and melt-extraction mechanisms: Physics and Chemistry of the Earth. Part A: Solid Earth and Geodesy, v. 26, p. 303–315, doi: 10.1016/S1464-1895(01)00060-6.

Daly, R.A., 1905, The classification of igneous intrusive bodies: The Journal of Geology, v. 13, p. 485–508, doi: 10.1086/621251.

Daly, R.A., 1933, Igneous Rocks and the depths of the earth, New York, McGraw-Hill, 598p.

DeCelles, P.G., Ducea, M.N., Kapp, P., and Zandt, G., 2009, Cyclicity in Cordilleran orogenic systems: Nature Geoscience, v. 2, p. 251–257, doi: 10.1038/ngeo469.

De Silva, S., 2008, Arc magmatism, calderas, and supervolcanoes: Geology, v. 36, p. 671–672, doi: 10.1130/focus082008.1.

Ducea, M., 2001, The California arc: Thick granitic batholiths, eclogitic residues, lithospheric-scale thrusting, and magmatic flare-ups: GSA Today, v. 11, no. 11, p. 4–10, doi: 10.1130/1052-5173(2001)011<0004:TCATGB>2.0.CO;2.

Fiske, R.S., and Tobisch, O.T., 1994, Middle Cretaceous ash-flow tuff and caldera-collapse deposit in the Minarets Caldera, east-central Sierra Nevada, California: Geological Society of America Bulletin, v. 106, p. 582–593, doi: 10.1130/0016-7606(1994)106<0582:MCAFTA>2.3.CO;2.

Flood, R.H., and Shaw, S.E., 1979, K-rich cumulate diorite at the base of a tilted granodiorite pluton from the New England batholith, Australia: Journal of Petrology, v. 87, p. 417–425.

Freund, R., 1970, Rotation of strike slip faults in Sistan, southeast Iran: The Journal of Geology, v. 142, p. 188–200, doi: 10.1086/627500.

Furnival, G.M., 1938, The large quartz veins of Great Bear Lake, Canada: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 30, p. 843–850.

Furnival, G.M., 1939, Geology of the area north of Contact Lake, N. W. T: American Journal of Science, v. 237, p. 476–499, doi: 10.2475/ajs.237.7.476.

Gibbins, S., 1988, The petrology and geochemistry of the Handley Page intrusive series: A concentrically-zoned intrusive sequence in the Great Bear magmatic zone, Northwest Territories: Unpublished B.Sc. thesis, Carleton University, Ottawa, 68p.

Glazner, A.F., and Bartley, J.M., 2006, Is stoping a volumetrically significant pluton emplacement process: Geological Society of America Bulletin, v. 118, p. 1185–1195, doi: 10.1130/B25738.1.

Glazner, A.F., Bartley, J.M., Coleman, D.S., Gray, W., and Taylor, R.Z., 2004, Are plutons assembled over millions of years by amalgamation from small magma chambers?: GSA Today, v. 14, no. 4/5, p. 4–11, doi: 10.1130/1052-5173(2004)014<0004:APAOMO>2.0.CO;2.

Gray, W., Glazner, A.F., Coleman, D.S., and Bartley, J.M., 2008, Long-term geochemical variability of the Late Cretaceous Tuolumne Intrusive Series, central Sierra Nevada, California: Geological Society of London: Special Publications, v. 304, p. 183–201, doi: 10.1144/SP304.10.

Grout, F.F., 1927, Fundamental problems of the batholiths and methods for their attack: The Journal of Geology, v. 35, p. 311–318, doi: 10.1086/623418.

Haederle, M., and Atherton, M.P., 2002, Shape and intrusion style of the Coastal batholith, Peru: Tectonophysics, v. 345, p. 17–28, doi: 10.1016/S0040-1951(01)00204-9.

Haeussler, P.J., and Paterson, S.R., 1993, Tilting, burial, and uplift of the Guadalupe igneous complex, Sierra Nevada, California: Geological Society of America Bulletin, v. 105, p. 1310–1320, doi: 10.1130/0016-7606(1993)105<1310:TBAUOT>2.3.CO;2.

Hamilton, W.B., 1969a, The volcanic central Andes—A model for the Cretaceous batholiths and tectonics of North America, *in* McBirney, A.R., ed., Proceedings of the Andesite Conference: Oregon Department of Geology and Mineral Industries, Bulletin 65, p. 175–184.

Hamilton, W.B., 1969b, Mesozoic California and the underflow of Pacific mantle: Geological Society of America Bulletin, v. 80, p. 2409–2430, doi: 10.1130/0016-7606(1969)80[2409:MCATUO]2.0.CO;2.

Hamilton, W., and Myers, W.B., 1967, The nature of batholiths; United States Geological Survey, Professional Paper 554C, 30p.

Hildebrand, R.S., 1981, Early Proterozoic LaBine Group of Wopmay orogen: remnant of a continental volcanic arc developed during oblique convergence, *in* Campbell, F.H.A., ed., Proterozoic basins in Canada: Geological Survey of Canada Paper 81–10, p. 133–156.

Hildebrand, R.S., 1982, A continental arc of early Proterozoic age at Great Bear Lake, Northwest Territories: unpublished Ph.D. dissertation, Memorial University of Newfoundland, St. John's, 237p.

Hildebrand, R.S., 1983, Geology, Echo Bay-MacAlpine Channel area, District of Mackenzie, Northwest Territories: Geological Survey of Canada Map 1546A.

Hildebrand, R.S., 1984a, Folded cauldrons of the early Proterozoic LaBine Group, northwestern Canadian shield: Journal of Geophysical Research, v. 89, p. 8429–8440, doi: 10.1029/JB089iB10p08429.

Hildebrand, R.S., 1984b, Geology of the Camsell, River-Conjuror Bay area, Northwest Territories: early Proterozoic cauldrons, stratovolcanoes and subvolcanic plutons: Geological Survey of Canada Paper 83–20, p. 42 p.

Hildebrand., R.S., 1985, Rainy Lake-White Eagle Falls area, District of Mackenzie: Geological Survey of Canada Map 1546A.

Hildebrand, R.S., 1986, Kiruna-type deposits: their origin and relationship to intermediate subvolcanic plutons in the Great Bear magmatic zone, northwest Canada: Economic Geology, v. 640–659.

Hildebrand, R.S., 2009, Did westward subduction cause Cretaceous-Tertiary orogeny in the North American Cordillera? Geological Society of America Special Paper 457, 71p.

Hildebrand, R.S., 2009b, A collisional origin for the great Cordilleran batholiths of western North America: Geological Society of America Abstracts with Programs, v. 41, no. 7, p. 589.

Hildebrand, R.S., 2010, Geological synthesis of Northern Wopmay Orogen / Coppermine Homocline, Northwest Territories—Nunavut: Geological Survey of Canada Open File Map 6390 (1:500,000 scale).

Hildebrand, R.S., and Bowring, S.A., 1984, Continental intra-arc depressions: A non-extensional model for their origin, with a Proterozoic example from Wopmay orogen: Geology, v. 12, p. 73–77, doi: 10.1130/0091-7613(1984)12<73:CIDANM>2.0.CO;2.

Hildebrand, R.S., and Bowring, S.A., 1988, Geology of parts of the Calder River map area, central Wopmay orogen, District of Mackenzie, *in* Current Research, Part A: Geological Survey of Canada, Paper 88-1C, p. 199–205.

Hildebrand, R.S., and Roots, C.F., 1985, Geology of the Rivière Grandin map area (Hottah terrane and western Great Bear magmatic zone), District of Mackenzie, *in* Current Research, Part A: Geological Survey of Canada, Paper 85-1A, p. 373–383.

Hildebrand, R.S., Bowring, S.A., and Housh, T., 1990, The medial zone of Wopmay orogen, District of Mackenzie, *in* Current Research, Part C: Geological Survey of Canada, Paper 90-1C, p. 167–176.

Hildebrand, R.S., Hoffman, P.F., and Bowring, S.A., 1987a, Tectonomagmatic evolution of the 1.9 Ga Great Bear magmatic zone, Wopmay orogen, northwestern Canada: Journal of Volcanology and Geothermal Research, v. 32, p. 99–118, doi: 10.1016/0377-0273(87)90039-4.

Hildebrand, R.S., Hoffman, P.F., and Bowring, S.A., 2010, The Calderian orogeny in Wopmay orogen (1.9 Ga) northwestern Canadian shield: Geological Society of America Bulletin, v. 122, p. 794–814, doi: 10.1130/B26521.1.

Hildebrand, R.S., Bowring, S.A., Andrew, K.P.E., Gibbins, S.F., and Squires, G.C., 1987b, Geological investigations in Calder River map area, central Wopmay orogen, District of Mackenzie, *in* Current Research, Part A: Geological Survey of Canada, Paper 87-1A, p. 699–711.

Hildebrand, R.S., Bowring, S.A., Steer, M.E., and Van Schmus, W.R., 1983, Geology and U-Pb geochronology of parts of the Leith Peninsula and Riviére Grandin map areas, District of Mackenzie, *in* Current Research, Part A: Geological Survey of Canada, Paper 83-1A, p. 329–342.

Hildreth, W., 1981, Gradients in silicic magma chambers—Implications for lithospheric magmatism: Journal of Geophysical Research, v. 86, p. 10153–10192, doi: 10.1029/JB086iB11p10153.

Hildreth, W., 2007, Quaternary magmatism in the Cascades–Geologic perspectives: United States Geological Survey Professional Paper 1744, 125p.

Hildreth, W., and Moorbath, S., 1988, Crustal contributions to arc magmatism in the Andes of Central Chile: Contributions to Mineralogy and Petrology, v. 98, p. 455–489, doi: 10.1007/BF00372365.

Hoffman, P.F., 1978, Sloan River map-area (86K): Geological Survey of Canada, Open File Map 535.

Hoffman, P.F., 1982, The Northern Internides of Wopmay Orogen, District of Mackenzie, Northwest Territories: Geological Survey of Canada, 1576A (with marginal notes).

Hoffman, P.F., 1984, Wopmay Fault Zone and its role in the evolution of Wopmay Orogen: Geological Association of Canada, Abstracts with Programs, v. 9, p. 74.

Hoffman, P.F., 1987, Continental transform tectonics: Great Slave Lake shear zone (ca. 1.9Ga), northwest Canada: Geology, v. 15, p. 785–788, doi: 10.1130/0091-7613(1987)15<785:CTTGSL>2.0.CO;2.

Hoffman, P.F., and McGlynn, J.C., 1977, Great Bear Batholith: a volcano-plutonic depression, *in* Baragar, W.R.A., Coleman, L.C. and Hall, J.M., eds, Volcanic Regimes in Canada: Geological Association of Canada, Special Paper 16, p. 170–192.

Hoffman, P.F., Bell, I.R., and Tirrul, R., 1976, Sloan River map area (86K), Great Bear Lake, District of Mackenzie, *in* Current Research, Part A: Geological Survey of Canada, Paper 76-1A, p. 353–358.

Hopson, C.A., and Dehlinger, D.A., 1989, Depth dependency of granite type, illustrated by tilted plutons in the North Cascades, Washington: Geological Society of America Abstracts with Programs, v. 21, no. 5, p. 95.

Housh, T.B., 1989, Petrochemistry of granitoids from the northern Great Bear magmatic zone: unpublished Ph.D. dissertation, Washington University, St. Louis, 347 p.

Huber, C., Bachmann, O., and Manga, M., 2010, Two competing effects of volatiles on heat transfer in crystal-rich magmas: Thermal insulation vs defrosting: Journal of Petrology, v. 51, p. 847–867, doi: 10.1093/petrology/egq003.

Hughes, G.R., and Mahood, G.A., 2008, Tectonic controls on the nature of large silicic calderas in volcanic arcs: Geology, v. 36, p. 627–630, doi: 10.1130/G24796A.1.

Hutchison, W.W., 1982, Geology of the Prince Rupert-Skeena map area, British Columbia: Geological Survey of Canada Memoir 394, 116 p.

Irving, E., Baker, J., Hamilton, M., and Wynne, P.J., 2004, Early Proterozoic geomagnetic field in western Laurentia: implications for paleolatitudes, local rotations and stratigraphy: Precambrian Research, v. 129, p. 251–270, doi: 10.1016/j.precamres.2003.10.002.

John, B.E., 1988, Structural reconstruction and zonation of a tilted midcrustal magma chamber; the felsic Chemehuevi Mountains Plutonic Suite: Geology, v. 16, p. 613–617, doi: 10.1130/0091-7613(1988)016<0613:SRAZOA>2.3.CO;2.

Kistler, R.W., Chappell, B.W., Peck, D.L., and Bateman, P.C., 1986, Isotopic variations in the Tuolumne Intrusive Suite, central Sierra Nevada, California: Contributions to Mineralogy and Petrology, v. 94, p. 205–220, doi: 10.1007/BF00592937.

Lawson, A.C., 1914, Is the Boulder "batholith" a laccolith?-a problem in ore genesis: University of California Publications: Bulletin of the Department of Geology, v. 8, p. 1–15.

Lipman, P.W., 2007, Incremental assembly and prolonged consolidation of Cordilleran magma chambers: Evidence from the Southern Rocky Mountain volcanic field: Geosphere, v. 3, no. 1, p. 1–29, doi: 10.1130/GES00061.1.

McGlynn, J.C., 1975, Geology of the Calder River map area (86F), District of Mackenzie, *in* Report of Activities, Part A: Geological Survey of Canada Paper 75-1A, p. 339–341.

Matzel, J.E.P., Bowring, S.A., and Miller, R.B., 2006, Time scales of pluton construction at different crustal levels: Examples from the Mount Stewart and Tenpeak intrusions, North Cascades: Geological Society of America Bulletin, v. 118, p. 1412–1430, doi: 10.1130/B25923.1.

Miller, C.F., and Miller, J.S., 2002, Contrasting stratified plutons exposed in tilt blocks, Eldorado Mountains, Colorado River rift, NV, USA: Lithos, v. 61, p. 209–224, doi: 10.1016/S0024-4937(02)00080-4.

Miller, C.F., Furbish, D.J., Walker, B.A., Claiborne, L.L., Koteas, G.C., Bleick, H.A., and Miller, J.S., 2009, Growth of plutons by incremental emplacement of sheets in crystal-rich host: Evidence from Miocene intrusions of the Colorado River region, Nevada, USA: Tectonophysics, doi: 10.1016/j.tecto.2009.07.011.

Miller, J.S., 2008, Assembling a pluton…one increment at a time: Geology, v. 36, p. 511–512, doi: 10.1130/focus062008.1.

Miller, J.S., Matzel, J.E.P., Miller, C.F., Burgess, S.D., and Miller, R.B., 2007, Zircon growth and recycling during the assembly of large composite arc plutons: Journal of Volcanology and Geothermal Research, v. 167, p. 282–299, doi: 10.1016/j.jvolgeores.2007.04.019.

Moore, J.G., and Lockwood, J.P., 1973, Origin of comb layering and orbicular structure, Sierra Nevada batholith, California: Geological Society of America Bulletin, v. 84, p. 1–20, doi: 10.1130/0016-7606(1973)84<1:OOCLAO>2.0.CO;2.

Myers, J.S., 1975, Cauldron subsidence and fluidisation: mechanisms of intrusion of the Coastal Batholith of Peru into its own volcanic ejecta: Geological Society of America Bulletin, v. 86, p. 1209–1220, doi: 10.1130/0016-7606(1975)86<1209:CSAFMO>2.0.CO;2.

Paterson, S.R., and Fowler, T.K., Jr., 1993, Re-examining pluton emplacement processes: Journal of Structural Geology, v. 15, p. 191–206, doi: 10.1016/0191-8141(93)90095-R.

Pelletier, K.S., 1988, Geology of the Ellington volcanic-sedimentary complex, Great Bear magmatic zone, Northwest Territories: unpublished MSc. thesis, Carleton University, Ottawa, 113 p.

Petford, N., Cruden, A.R., McCaffrey, K.J.W., and Vigneresse, J.-L., 2000, Granite magma formation, transport and emplacement in the Earth's crust: Nature, v. 408, p. 669–673, doi: 10.1038/35047000.

Pitcher, W.S., 1993, The nature and origin of granite, Blackie, London, 321 p.

Read, H.H., 1957, The granite controversy, Thomas Murby & Co., London, 430 p.

Reardon, N.C., 1989, The Mystery Island Intrusive Suite and associated alteration haloes, Great Bear Lake, District of Mackenzie, *in* Current Research, Part

C: Geological Survey of Canada, Paper 89-1C, p. 37–42.

Reardon, N.C., 1990, Altered and mineralized rocks at Echo Bay, N.W.T., and their relationship to the Mystery Island Intrusive Suite, *in* Current Research, Part C: Geological Survey of Canada, Paper 90-1C, p. 143–150.

Reardon, N.C., 1992, Magmatic-hydothermal systems and associated magnetite-apatite-actinolite deposits, Echo Bay, Northwest Territories: unpublished M.Sc. thesis, University of Ottawa, 154 p.

Reichenbach, I.G., 1985, Stratigraphy and structure of the early Proterozoic Bell Island Group, District of Mackenzie, *in* Current Research, Part B: Geological Survey of Canada, Paper 85-1B, p. 151–160.

Reichenbach, I.G., 1986, An ensialic marginal basin in Wopmay orogen, northwestern Canadian Shield; unpublished M.Sc. thesis, Ottawa, Ontario, Carleton University, 120 p.

Reichenbach, I.G., 1991, The Bell Island Bay Group, Remnant of an early Proterozoic ensialic marginal basin in Wopmay orogen, District of Mackenzie: Geological Survey of Canada, Paper 88–28, 43p.

Reid, M.R., 2008, How long does it take to supersize an eruption: Elements, v. 4, p. 23–28, doi: 10.2113/GSELEMENTS.4.1.23.

Saleeby, J.B., Kistler, R.W., Longiaru, S., Moore, J.G., and Nockleberg, W.J., 1990, Middle Cretaceous silicic metavolcanic rocks in the Kings Canyon aream, central Sierra Nevada, California, *in* Anderson, J. L., ed., The nature and origin of Cordilleran magmatism; Boulder, Colorado, Geological Society of America, Memoir 174, p. 251–270.

Seager, W.R., 1981, Geology of the Organ Mountains and southern San Andres Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources Memoir 36, 97 p.

Smith, R.L., 1979, Ash-fow magmatism, *in* Chapin, C. E., and Elston, W. E., eds., Ash-Flow Tuffs: Geological Society of America Special Paper 180, p. 5–27.

Tirrul, R., 1976, The geology of the Rainy Lake igneous complex, District of Mackenzie, Northwest Territories: unpublished B.Sc. thesis, Queen's University, Kingston, 115 p.

Tirrul, R., 1987, Geology, Calliope Lake area, Wopmay orogen, District of Mackenzie; Geological Survey of Canada Map 1654A, scale 1:50,000.

Verplanck, P.L., Farmer, G.L., McCurry, M., and Mertzman, S.A., 1999, The chemical and isotopic differentiation of an epizonal magma body: Organ Needle pluton, New Mexico: Journal of Petrology, v. 40, p. 653–678, doi: 10.1093/petrology/40.4.653.

Walker, B.A., Jr., Lang, N.P., Claiborne, L.L., Miller, C.F., and Hazlett, R.W., 2008, The Spirit Mountain batholith and Secret Pass Canyon volcanic center: Perspectives on silicic magmatism from the mid to upper crust: Geological Society of America Abstracts with Programs, v. 40, no. 1, p. 74.

Walker, B.A., Jr., Miller, C.F., Lowery Claiborne, L., Wooden, J.L., and Miller, J.S., 2007, Geology and geochronology of the Spirit Mountain batholith, southern Nevada: Implications for timescales and physical processes of batholith construction: Journal of Volcanology and Geothermal Research, v. 167, p. 239–262, doi: 10.1016/j.jvolgeores.2006.12.008.

Yoshinobu, A.S., and Barnes, C.G., 2008, Is stoping a volumetrically significant pluton emplacement process?: Geological Society of America Bulletin, v. 120, p. 1080–1081, doi: 10.1130/B26141.1.

Manuscript Received 3 July 2009
Revised Manuscript Received 8 July 2010
Manuscript Accepted 12 July 2010