The Calderian orogeny in Wopmay orogen (1.9 Ga), northwestern Canadian Shield

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ABSTRACT

The Wopmay orogen is a Paleoproterozoic orogenic belt formed in part by the accretion of Hottah terrane, an east-facing continental magmatic arc, to the western margin of the Archean Slave craton at ca. 1.88 Ga. The arc-continent collision was responsible for the Calderian orogeny. Just prior to the collision, arc volcanism of the Hottah terrane had migrated trenchward and changed in composition from an aphyric calc-alkaline to a bimodal tholeiitic suite. The change in magmatism, along with subsidence and consequent high-temperature–low-pressure metamorphism of rocks on the upper plate, is attributed to extension and asthenospheric upwelling during rollback of the lower plate. The upwelling led to regional heating of the crust, melting, and generation of the metaluminous to peraluminous Hepburn intrusive suite within the arc just prior to accretion of Hottah terrane onto the Slave craton. During the collision, a 200-km-long, orogen-parallel swarm of mafic sills was intruded into the lower plate and trench-fill sediments at the passive-margin shelf-slope break. The mafic magmatism may have been induced by slab breakoff. The Calderian orogeny provides an excellent example of the magmatic history of an arc-continent collision.

The Calderian orogeny was short-lived. Within a few million years, an east-dipping subduction zone formed outboard of the accreted terrane, creating volcanic rocks that were erupted atop the eroded Hottah-Slave collision zone. The short duration of the Calderian orogeny seems typical of arc-continent collisions, perhaps because the attenuated arc crust, coupled with the extended crust of the lower plate, never generates severely overthickened crust. Rise of asthenospheric mantle due to slab breakoff generates rapid uplift, erosion, and collapse of the mountains built during collision.

INTRODUCTION

Located in the northwest part of the Canadian Shield (Fig. 1), the Wopmay orogen provides a well-exposed example of a Paleoproterozoic collisional belt and postcollisional magmatic arc. First mapped by Fraser et al. (1960; see also Fraser, 1964), it contains one of the first recognized Precambrian continental margins (Hoffman, 1973) and remains a defining example of Precambrian plate tectonics (Hoffman, 1980; Windley, 1995; Condie, 1997; Stanley, 1999).

Existing models were mostly published prior to the recognition of a large klippe of exotic rocks (Turmoil klippe; Fig. 1) within the internal zone of the orogen. The volcanic and sedimentary rocks of the klippe were previously interpreted as initial rift-facies rocks (Easton, 1981a, 1982; Hoffman and Bowring, 1984; Hoffman et al., 1988; Hoffman, 1989), and the crystalline basement beneath them was interpreted as their metamorphosed equivalents (St-Onge, 1981, 1984a, 1984b; St-Onge and King, 1987a, 1987b; King, 1986). A key finding was presented by Bowring and Grotzinger (1992), who reported U-Pb zircon dates for ash beds in cratonic cover correlative with the passive-margin sequence to demonstrate that the Coronation margin was substantially older than the supposed initial rift rocks.

In this contribution, we meld geochronological data with field observations in the internal zone to reveal that the evolution of the main collisional phase of the orogen, termed the Calderian orogeny, represents a typical short-lived arc-continent collision. We also present an evolutionary model that incorporates progressive rollback of the lower plate and extension of the arc terrane on the upper plate, emplacement of the leading edge of the extended continental arc upon the western margin of Slave craton, failure of the subducting plate during the collision, and the subsequent development of a new subduction zone, or possible propagation of an older one, of opposite polarity outboard of the amalgamated collision zone—all within about 10 m.y. The overall evolution and short-lived nature of the collision are typical of modern arc-continent collisions (Suppe, 1987; Cloos et al., 2005), and it appears that arc-continent collisions are the chief way in which passive margins are converted to active margins.

TECTONIC DIVISIONS OF WOPMAY OROGEN

Wopmay orogen is divided into five major zones, from east to west: Coronation margin, Turmoil klippe, the Medial zone, Great Bear magmatic zone, and Hottah terrane (Fig. 1). Coronation margin contains basement rocks of the Archean Slave craton overlain by a tripartite sedimentary succession representing three distinct tectonic regimes: rift, passive margin, and foredeep (Hoffman, 1973, 1980, 1984, 1989). The supracrustal rocks of the margin were detached from their basement, folded, and transported eastward during the Calderian orogeny. The passive margin to foredeep transition, marking the onset of collision, is dated at 1882 ± 4 Ma by U-Pb chronology of zircons from a volcanic ash bed near the base of the foredeep sequence (Bowring and Grotzinger, 1992). The westernmost zone, Hottah terrane, developed remotely from, but in part contemporaneously with, Coronation margin. It consists of calc-alkaline volcanic and plutonic rocks erupted on and intruded into early Paleoproterozoic continental crust (Hildebrand et al., 1983, 1984; Hildebrand and Roots, 1985). Turmoil klippe

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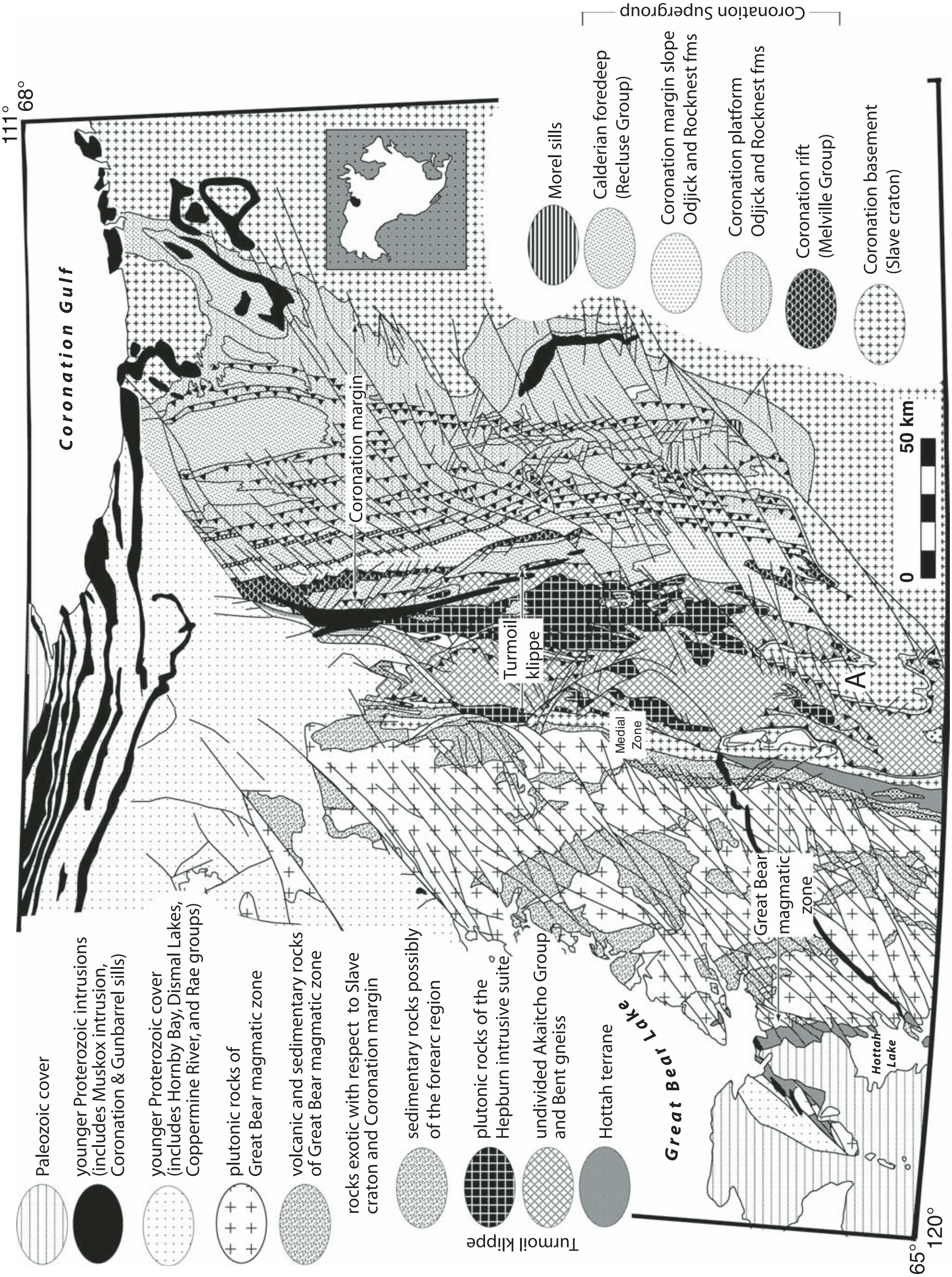

Figure 1. Geological sketch map showing the subdivisions of Wopmay orogen discussed in this paper. A—Acasta gneiss in Exmouth anticline.

is a large erosional remnant of Hottah terrane that structurally overlies the westernmost parts of Coronation margin (Hildebrand et al., 1990, 1991). The klippe contains crystalline basement unconformably overlain by metamorphosed sedimentary and volcanic rocks, all of which are intruded by the composite Hepburn Batholith (Fig. 1). Great Bear magmatic zone is dominated by calc-alkaline volcanic and plutonic rocks. It is a typical magmatic arc built on continental crust (Hildebrand et al., 1987b). It is younger than the Calderian orogeny and forms an overlap assemblage unconformably overlying the Hottah terrane, the extreme western edge of Coronation margin, and Turmoil klippe (Hoffman and McGlynn, 1977; Hildebrand et al., 1990).

The Medial zone of the orogen occurs along the eastern margin of Great Bear magmatic zone and includes rocks of all the other zones—as well as their structural complexities—now tightly folded about northerly trending axes (Hildebrand et al., 1990). Normal faults, unconformably overlain by rocks of the Great Bear magmatic zone, occur within the zone where they placed little metamorphosed pillow basalts of Turmoil klippe over high-grade rocks of the collisional core, and they attest to the gravitational collapse of the orogen (Hildebrand et al., 1990). Based on isotopic and field data, the western edge of Slave craton lies within the Medial zone (Housh et al., 1989; Bowring and Podosek, 1989; Hildebrand et al., 1990). The folds postdate magmatism in the Great Bear magmatic zone, where they mostly trend northwesterly. Where the folds affect rocks of Slave basement and its cover, such as in the Medial zone and eastward, they trend northerly. The orogen is permeated by a through-going system of conjugate transcurrent faults, forming northeast-striking right-lateral and northwest-striking left-lateral domains (Fig. 1).

The folds and the transcurrent faults are related to younger orogenies, which were focused westward of the Great Bear magmatic zone but had effects throughout the orogen (Hildebrand et al., 1987b). Models based on the deep seismic line SNORCLE mostly relate to these younger, more westerly rocks and events (Snyder et al., 2002; Cook et al., 2005). In a general sense, the orogen is exposed in oblique cross section due to a gentle northward plunge caused by much younger events and open, east-northeast–trending, Paleoproterozoic cross-folds that postdate the Calderian orogeny (Hoffman et al., 1988) (Fig. 1).

CORONATION MARGIN

Slave craton constitutes the basement for rocks of the Coronation Supergroup. It is a 500 × 700 km region of Archean lithosphere surrounded by Paleoproterozoic orogenic belts (Hoffman, 1989; Hoffman and Hall, 1993). Regional geological mapping and geochronology led to the recognition of a domain of igneous and metamorphic basement older than 3.0 Ga—locally overlain by 2.8 Ga quartzites and iron formation—in the west-central portion of the craton (Bleeker et al., 1999). The basement rocks include the well-known 4.03–3.6 Ga Acasta gneisses (Bowring et al., 1989; Bowring and Williams, 1999), which are exposed in the core of Exmouth anticline within Wopmay orogen (Fig. 1). Overlying the older basement, there are 2.73–2.62 Ga supracrustal rocks intruded by diverse suites of granitoids, including 2.7 Ga tonalites, granodiorites, and 2.58 Ga potassic granites.

Rocks of Coronation margin lie unconformably on Archean rocks of Slave craton and are collectively termed the Coronation Supergroup (Hoffman, 1973). The supergroup consists of, in ascending order, Melville, Epworth, and Recluse groups (Hoffman, 1981).

The Melville Group is a complex of bimodal volcanic and mainly coarse-grained clastic rocks that are interpreted as a rift-facies assemblage (Hoffman and Pelletier, 1982). It is only exposed in a single northerly trending anticlinorium that is 130 km long by <13 km wide (Fig. 1). Hildebrand and Bowring (1999) suggested that most of the original rift-facies rocks were subducted during collision.

Overlying the Melville Group, there is a sequence of shallow-marine sedimentary rocks known as the Epworth Group, which is interpreted to represent a west-facing passive continental margin (Hoffman, 1973). The Epworth Group is widely exposed and consists of a basal 200–1500 m, westward-thickening succession of storm-dominated siliciclastic rocks known as the Odjick Formation. It is overlain by the Rocknest Formation, a 450–1100 m, cyclic, dolomitic, shelf sequence with a rim of supratidal tepee facies flanked by outer-shoreface grainstones and stromatolites, and a sediment-starved submarine debris apron (Grotzinger, 1986a, 1986b, 1986c). U-Pb zircon geochronology of volcanic ash beds in strata correlative with the Epworth Group provide an age of 1969 ± 1 Ma for the lower part of the Odjick Formation (Bowring and Grotzinger, 1992). Terminal drowning of the Rocknest shelf (Fig. 2) is marked by stromatolite patch reefs of the form-genus *Tungussia*, overlain by deep-water siliciclastics of the basal Recluse Group (Grotzinger, 1986b).

The Recluse Group is a collision-related foredeep assemblage (Hoffman, 1973). A sediment-starved, transgressive tract (Tree River Formation) consists of glauconitic siltstone and granular ironstone. It is overlain by a thick, shoaling-upward sequence of flat-laminated graphitic-sulfidic shale (Fontano Formation), concretionary nongraphitic shale (Kikerk Formation), argillaceous limestone rhythmite (Cowles

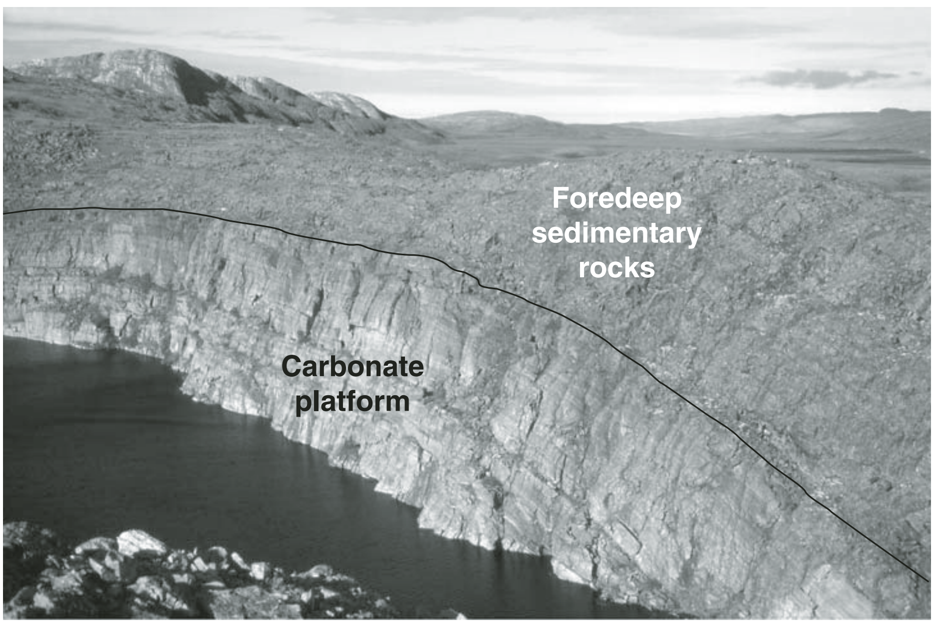

Figure 2. Abrupt contact between carbonate platform (Rocknest Formation) and clastic rocks of the Calderian foredeep (Tree River Formation). The contact represents terminal drowning of the Coronation margin as it was pulled down into the trench beneath the Hottah terrane.

Formation), evaporite solution-collapse megabreccia, and red, cross-bedded, lithic-arenite (Takiyuak Formation). Northerly derived turbidites (Fig. 3) of coarse-grained feldspathic wacke (Asiak Formation) are intercalated with the graphitic-sulfidic and concretionary shales, and their first appearance climbs stratigraphically upsection from west to east across the former shelf area, consistent with eastward migration of the foredeep axis over time (Hoffman, 1973). The volcanic ash bed dated at 1882 ± 4 Ma (Bowring and Grotzinger, 1992) occurs near the base of the graphitic-sulfidic shale, below the lowest turbidites. Combined with the age of 1969 ± 1 Ma for the basal Odjick Formation, it suggests that the Coronation passive margin lasted for ~90 m.y. (Bowring and Grotzinger, 1992).

HOTTAH TERRANE

Hottah terrane, which is exposed east and west of the Great Bear magmatic zone (Fig. 1), consists of crystalline basement, volcanic and sedimentary cover, and a variety of intrusive rocks. Isotopic data and sparse outcrops suggest that rocks of the Hottah terrane underlie most of the Great Bear magmatic zone; there is no evidence that the Slave craton extends farther west than the Medial zone (Bowring and Podosek, 1989; Housh et al., 1989; Hildebrand et al., 1990). Rocks grouped in the Hottah terrane are in part the same age as those of Coronation margin (Fig. 4), but they are vastly different in lithology and tectonic setting. Turmoil klippe (Fig. 1) is a part of the leading edge of the Hottah terrane that was thrust onto the Coronation margin.

West of the Great Bear magmatic zone, basement within the Hottah terrane is termed the Holly Lake metamorphic suite. Within the Turmoil klippe, Hottah basement is named Bent gneiss. Cover rocks to the west are called the Bell Island Bay Group; within Turmoil klippe, they are known as the Akaitcho Group. There is no group name for plutonic rocks of the Hottah terrane west of the Great Bear magmatic zone, but to the east, they are known as the Hepburn intrusive suite, or collectively as the Hepburn Batholith.

Basement within the Hottah terrane has not been the subject of detailed petrographic studies, but its distribution and lithology are well known (Hildebrand et al., 1983, 1984, 1991; Hildebrand and Roots, 1985). The Holly Lake metamorphic suite is dominantly orthogneiss and quartz-plagioclase-biotite ± muscovite ± sillimanite schist with minor garnet amphibolite, all of which are intimately intruded by variably deformed granitoid bodies. Original compositional layering in the schists is completely transposed, and primary sedimentary structures obliterated. The transposed fabric has been isoclinally folded about variable axial planes. The orthogneisses are of intermediate composition, mainly diorite and quartz diorite, with lesser amounts of quartz monzonite and granite. They are L/S tectonites with shallowly plunging axes and variable axial planes.

Additional lithologies common in the basement assemblage, especially near Hottah Lake (Fig. 1), are variably deformed pillow basalt and porphyritic andesite. The metavolcanic rocks were metamorphosed to amphibolite facies, but were locally regressed to greenschist facies. They are intercalated with cordierite paragneiss, psammites, pelitic schists, probable volcaniclastic rocks, minor hematite beds, and conglomerate.

U-Pb analyses of multigrain detrital zircon populations from a sample of biotite schist in the basement yielded dates of ca. 2.1 Ga, and the sandy matrix of a stretched-pebble conglomerate containing quartz, mafic volcanic, and siliceous porphyritic pebbles yielded mainly zircons of uniform size, shape, and color with an upper intercept date of 2278 ± 10 Ma from multigrain analyses (Fig. 4). These data suggest an age for the source of 2.1–2.3 Ga. Furthermore, a sample of metasedimentary rocks that were collected 130 km away to the north-northeast from the main outcroppings of the Hottah terrane, and that are known to lie unconformably beneath rocks of the Great Bear magmatic zone, also yielded detrital zircons dated at ca. 2.1 Ga (Bowring, 1984).

Several types of deformed and metamorphosed plutonic rocks, most commonly hornblende or biotite-hornblende diorite, quartz diorite, granodiorite, and monzogranite, with lesser amounts of biotite syenogranite, occur within the basement of the Hottah terrane. They are variably foliated and, in places, protomylonitic to ultramylonitic. Two samples of deformed plutonic rocks, both of which lie unconformably beneath the cover sequence, yielded U-Pb crystallization ages of 1914 ± 5 Ma and 1902 ± 7 Ma (Hildebrand et al., 1983).

Bell Island Bay Group

This group is known only from the westernmost exposures of Wopmay orogen, near the southeast corner of Great Bear Lake. The best-exposed and most complete sections occur around Hottah Lake (Fig. 1). They sit unconformably atop the Holly Lake metamorphic suite and are unconformably overlain by rocks

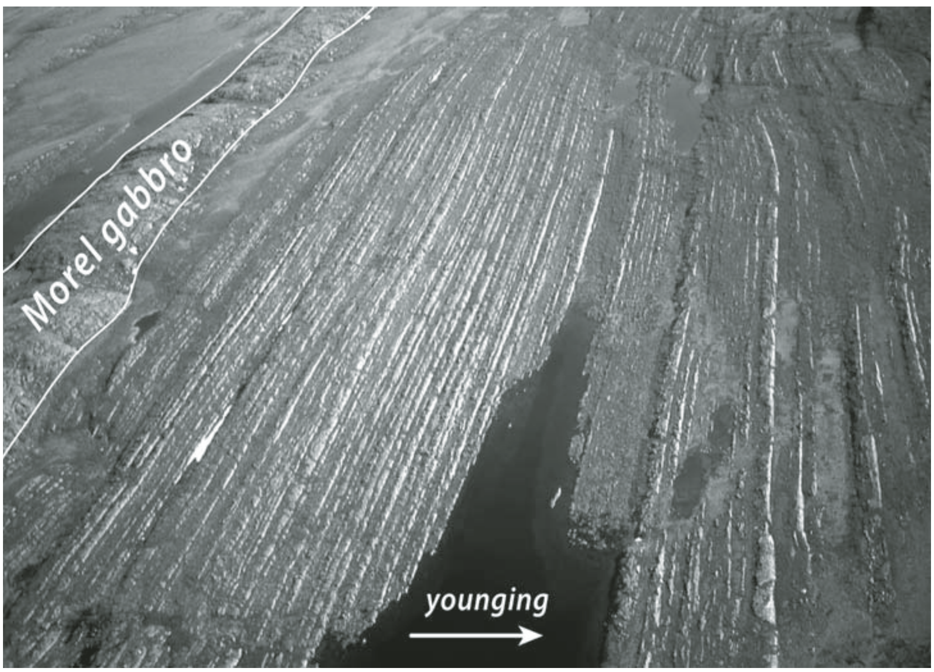

Figure 3. Aerial view of feldspathic-wacke turbidites (Asiak Formation) of the Calderian foredeep. The turbidity currents flowed southward within the foredeep, and their first appearance steps stratigraphically higher from west to east across the former shelf area, consistent with eastward migration of the foredeep axis over time. The turbidites are intruded by a syncollisional Morel gabbro sill. Field of view in the middle ground is ~1.0 km.

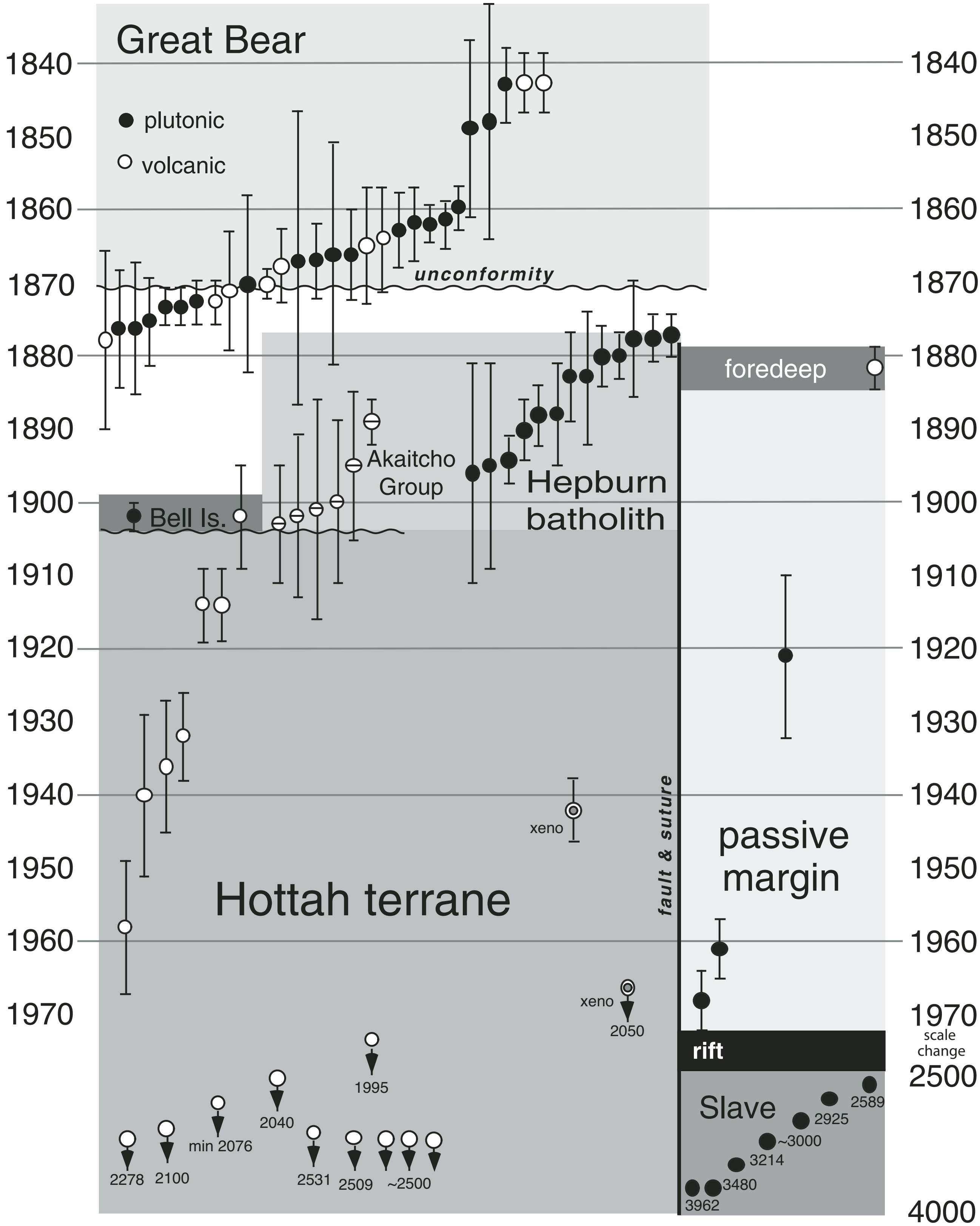

Figure 4. Compilation of U-Pb zircon geochronology for Wopmay orogen. Verticle axis is Ma, note break and scale change for Slave craton. Note that the statistical minimum age for Hepburn batholith is 1874–1880 Ma and that magmatism of the Great Bear magmatic zone must have started by 1871 Ma. Data are from Bennett and Rivers (2006); Bowring (1985); Gandhi et al. (2001); Housh (1989); Reichenbach (1991); and Villeneuve (1988).

of the Great Bear magmatic zone (Hildebrand et al., 1983, 1987b).

The Bell Island Bay group includes a basal fining-upward sequence of conglomerate and sandstone, and an overlying, but locally interfingering, sequence of generally aphyric dome collapse breccias, ash-flow tuffs, domes, and lavas ranging continuously in composition from basalt to rhyolite (Hildebrand et al., 1984; Reichenbach, 1991). Overall, the rocks are of subgreenschist grade, and primary textures are well preserved (Fig. 5). U-Pb zircon data obtained from a sparsely porphyritic rhyolite flow are consistent with a crystallization age of 1902 ± 2 Ma (Fig. 4), whereas a sample of strained granodiorite collected beneath the unconformity yielded a crystallization age of 1914 ± 2 Ma (Hildebrand et al., 1983; Reichenbach, 1991). Detrital zircons from the basal sandstone gave dates close to 1.96 Ga (Bowring, 1984).

Figure 5. Eutaxitic texture in welded ash-flow tuff, Bell Island Bay Group. Primary volcanic textures are extremely well-preserved in the Bell Island Bay Group, which was part of the arc complex that existed within the Hottah terrane prior to collision with the Slave craton.

Although all of the rocks of the Bell Island Bay Group are altered, probably by low-temperature saline fluids (Reichenbach, 1991), their geochemistry shows that they constitute a calc-alkaline suite (Fig. 6). The basalts have low titanium contents, are enriched in large ion lithophile elements (LILEs), and are depleted in high field strength element (HFSE) content (Reichenbach, 1991). The crystalline basement and the style of volcanism within the Bell Island Bay group—its generally aphyric nature, the continuous range in composition from basalt to rhyolite, and its calc-alkaline affinities—suggest that these rocks were erupted in a volcanic arc constructed on extending continental lithosphere (Reichenbach, 1991).

TURMOIL KLIPPE

Rocks of Turmoil klippe are widely exposed east of the Medial zone, where they sit in thrust contact above rocks of Coronation margin. A zone of mylonitic straight gneiss (Fig. 7), several meters thick, everywhere separates rocks of the klippe from underlying rocks. Mineral lineations are best developed in subjacent quartzite of Coronation margin, which, after unfolding, had an azimuth of NE-SW, indicating oblique

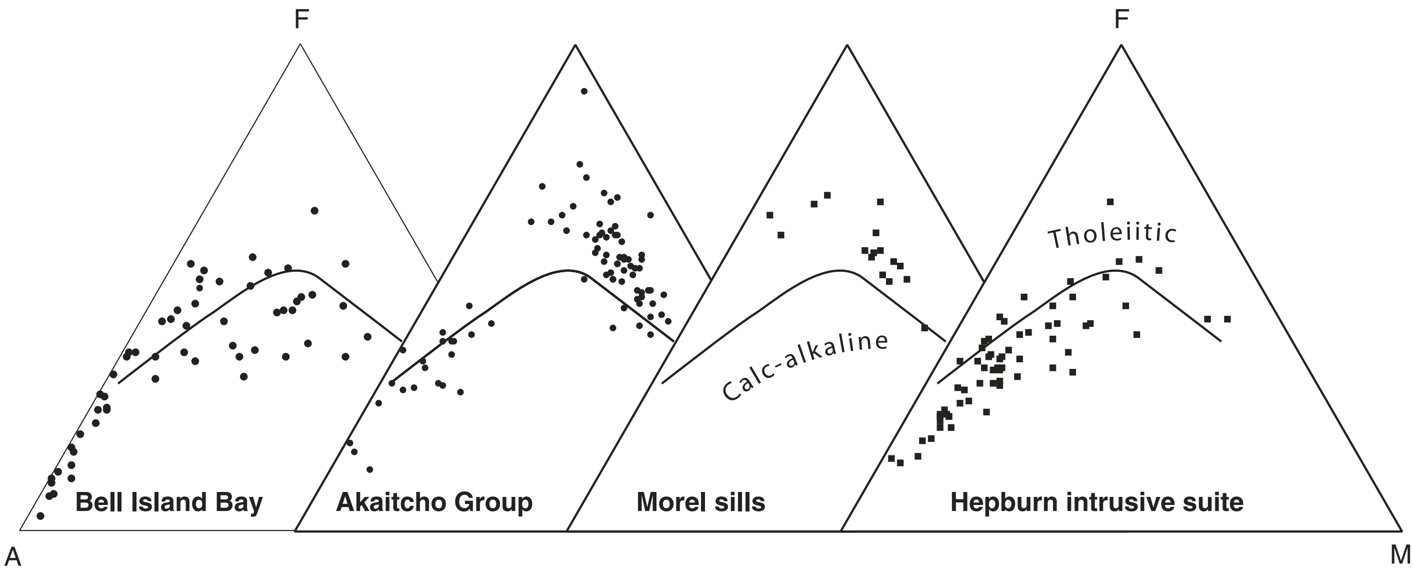

Figure 6. AFM diagrams for various igneous suites discussed in this paper, where A = Na_2O + K_2O, M = MgO, and F = total iron as FeO. Dividing line between calc-alkaline and tholeiitic fields is from Irvine and Baragar (1971). All of the suites shown could have been derived from the interaction of mafic mantle-derived magmas with continental crust. Data are from Hildebrand (1982); Easton (1982); Lalonde (1986); Reichenbach (1986); and unpublished analyses by second author (Hoffman).

emplacement of the klippe relative to the north-trending passive-margin shelf edge (Hildebrand et al., 1991).

The overall gentle regional plunge to the north and regional cross-folds expose an oblique view through various crustal levels of Turmoil klippe. Rocks of the klippe—mainly exposed in a northerly trending syncline-anticline pair—make up large tracts of Bent gneiss and Hepburn Batholith at deep structural levels; a zone of complexly interleaved Bent gneiss and Akaitcho Group with scattered plutons at intermediate structural levels; and dominantly Akaitcho Group intruded by hypabyssal porphyries at shallow structural levels (Hildebrand et al., 1991).

The lower and easternmost level of Turmoil klippe is composed of dominantly Bent gneiss and plutonic rocks of the Hepburn Batholith, with only minor amounts of supracrustal rocks (Fig. 8). The supracrustal rocks only occur on the western side of the batholith, suggesting that the majority of the batholith was intruded along the basement-cover interface. Middle structural levels are characterized by stacked and folded thrust sheets containing rocks of both Bent gneiss and Akaitcho Group (Hildebrand et al., 1991)

In the south, where it is extensively exposed, Bent gneiss is composed dominantly of tonalitic orthogneiss, with subordinate amounts of granitic to dioritic orthogneiss, paragneiss, and amphibolite (Hildebrand et al., 1991). Metasedimentary rocks occur mainly as meter- to decimeter-wide enclaves in the tonalitic gneiss. In places, there are also large coarse-grained and variably strained metagabbroic sills. An assortment of porphyritic and even-grained, but nevertheless gneissic, biotite and biotite-hornblende granitoid rocks ranging from diorite to granite intrudes the gneisses. U-Pb zircon dates from gneissic rocks range from ca. 2.6 Ga to 2.0 Ga, and the oldest ages come from high-grade tonalitic gneisses. This contrasts with the Slave craton, which was buried by turbidites at ca. 2.6 Ga (Isachsen and Bowring, 1994) and contains no basement rocks younger than 2.5 Ga.

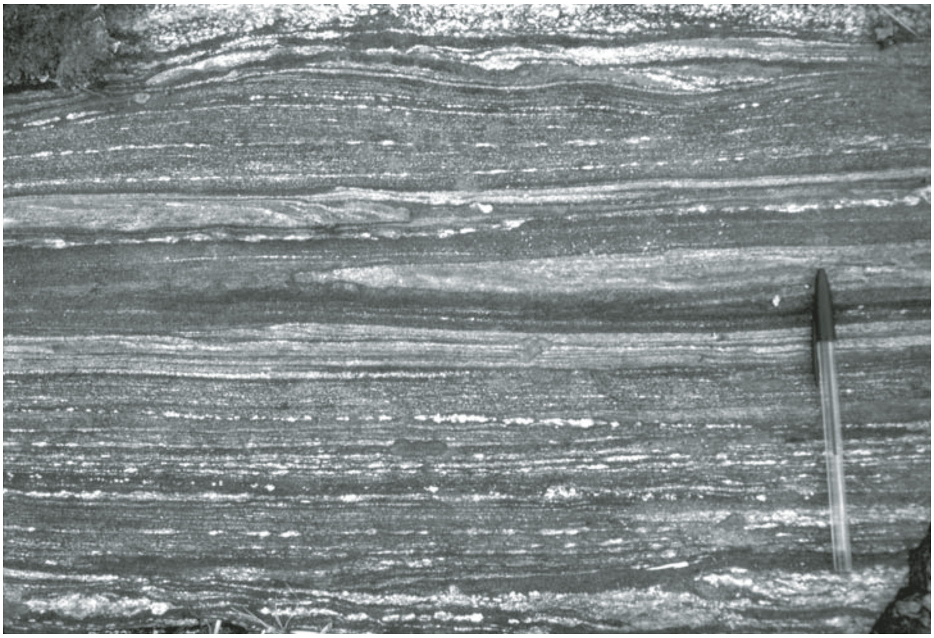

Figure 7. Highly strained gneisses near the base of Turmoil klippe. These mylonitic gneisses occur at the basal contact of Turmoil klippe everywhere it is exposed.

Figure 8. Contorted orthogneiss typical of Bent gneiss.

Akaitcho Group

Although Easton (1982) divided rocks of the northern Akaitcho Group into three subgroups and seven formations, they are more simply viewed as intercalated sedimentary rocks and bimodal basaltic-rhyolitic volcanic rocks, all cut by hypabyssal sills and juxtaposed on thrust and normal faults. The complex structure and facies, not yet definitively unraveled, preclude complete stratigraphic analysis, except near the basal unconformity or in individual fault slices. U-Pb zircon dates (Fig. 4) from rhyolitic lavas and sills range from 1903 Ma to 1889 Ma, whereas multigrain fractions of zircon from an arkosic sandstone yielded detrital zircons in the range 2.6–1.89 Ga (Bowring, 1984).

The upper structural levels of Turmoil klippe are dominated by a single thrust slice, consisting mostly of sedimentary rocks of the Akaitcho Group intruded by sills. Basement within this slice only occurs on its western side, where granitic gneisses are overlain by several kilometers of subarkosic to arkosic turbidites, all intruded by feldspar porphyritic sills, 300–600 m thick (Easton, 1982).

Slices at the intermediate structural levels consist of Bent gneiss unconformably overlain by sedimentary and volcanic rocks of Akaitcho Group. In the north, sedimentary rocks include immature sandstones, typically turbiditic, dolomite, pelite, and cobbly conglomerate. Relict chiastolite occurs locally within the pelites. Volcanic rocks are distinctly bimodal (Fig. 6) and have pillowed and massive basalt (Fig. 9) overlain by subaqueous rhyolitic flows and domes. Rhyolitic sills, mostly plagioclase and potassium feldspar porphyry, and gabbroic sills intrude the volcano-sedimentary pile (Easton, 1980, 1981a, 1982). The basalts were classified as continental tholeiites and oceanic basalts (Easton, 1981a, 1982).

In the southern part of the area, rocks of the Akaitcho Group unconformably overlie Bent gneiss on an erosional surface with <1 m of relief. In most places, the surface is overlain by variable thicknesses of pyritic psammopelite, but locally there are 10–15 cm of grus or pebbly conglomerate sitting beneath the psammopelite. Overlying the psammopelite, there are several meters of carbonate, which preserve tight reclined folds more difficult to see in other lithologies. The carbonate unit is generally overlain by pillow basalts but, locally, by medium-grained psammites. Numerous gabbroic sills intrude the sequence.

Rocks of the Akaitcho Group record sedimentation and volcanism in a subsiding basin. The bimodal nature of the volcanic rocks and the presence of both continental and oceanic tholeiites suggest that the basin formed by extension where basalts were able to invade and penetrate the crust (Easton, 1981a, 1982). Since rocks of the Akaitcho Group are mostly younger than rocks of Coronation margin and were deposited on Hottah basement, they cannot be initial-rift rocks of Coronation margin as originally hypothesized (Easton, 1981a, 1982; Hoffman and Bowring, 1984; King, 1986; St-Onge, 1984a, 1984b).

Hepburn Intrusive Suite

Crystalline basement and cover rocks of the Akaitcho Group within Turmoil klippe have been perforated and intruded by plutons of the Hepburn intrusive suite (Hoffman et al., 1980). These plutons were once interpreted to represent syncollisional magmatism, in large part because of their location within the internal zone, their variably deformed nature, and the peraluminous character of some plutons (Hoffman et al., 1980, 1988; St-Onge, 1987; St-Onge and King, 1987a, 1987b; St-Onge et al., 1982, 1983, 1984; Lalonde, 1986, 1989). However, plutons of the suite are not known to intrude the underlying Slave craton and are confined to Turmoil klippe (Figs. 1 and 10A), where they are concentrated near the basement-cover interface, and to areas within the Medial zone.

Overall, the suite ranges continuously in composition from granite to gabbro. The earliest plutons are gneissic biotite-hornblende monzogranites, garnet monzogranite sheets, and voluminous biotite-muscovite granites (Lalonde, 1986; Hoffman et al., 1980; Hildebrand et al., 1987b). Younger plutons, as inferred from crosscutting relationships, are more massive and include two mica granites and granodiorites, as well as hornblende-biotite-pyroxene diorite, quartz diorite, and gabbro (Fig. 10B). Overall, the Hepburn intrusive suite becomes more mafic in composition and less voluminous with time (Hoffman et al., 1980), the opposite of typical plutonic suites within nonextending continental arcs.

Many of the older plutons contain potassium feldspar megacrysts, sparse opaque oxides with ilmenite dominant, and accessory minerals such as garnet, sillimanite, zircon, allanite, apatite, and tourmaline (Lalonde, 1986). Disrupted metasedimentary enclaves are common (Fig. 11), as are xenocrysts of garnet, muscovite, and sillimanite, especially where there are abundant enclaves. The granites also have low ferric to ferrous iron ratios and heavy oxygen isotope ratios (Lalonde, 1986, 1989). In general, the granitic members of the suite have the typical attributes of peraluminous granites.

Rocks of the suite form a continuous compositional series that is calc-alkaline (Fig. 6). In a general way, rocks with more than ~60% SiO_2 are corundum-normative; those with less are diopside-normative (Fig. 12). The rocks have elevated concentrations of rare earth elements (REE) and the strongly fractionated patterns typical of many arc rocks (Lalonde, 1986).

Previous interpretations of the geology indicated that the Hepburn intrusive suite had a thermal aureole in its wall rocks as defined by mineral isograds (St-Onge, 1981, 1987; St-Onge and King, 1987a, 2987b). However, more recent field work has revealed that most, if not all, of the high-grade rocks in Turmoil klippe are basement gneisses rather than cover, and that metamorphism shows no obvious relationship to the intrusive suite (Hildebrand et al., 1991). Although the pressure-time (*P-t*) paths of St-Onge (1987) related to pre-Akaitcho metamorphism in the basement and not to Calderian metamorphism, rocks of Coronation margin directly to the east of, and structurally beneath, Turmoil klippe were metamorphosed to andalusite and biotite grade, and the grade increases structurally upward toward the klippe.

Interestingly, the plutons of the batholith are almost entirely localized near the basement-cover interface within the structurally lowest and easternmost slice(s) of Turmoil klippe, and rocks of the Akaitcho Group occur only on the western side of the batholith, where they are exposed on the east limb of Robb River syncline.

Figure 9. Slightly flattened pillow basalt of Akaitcho Group.

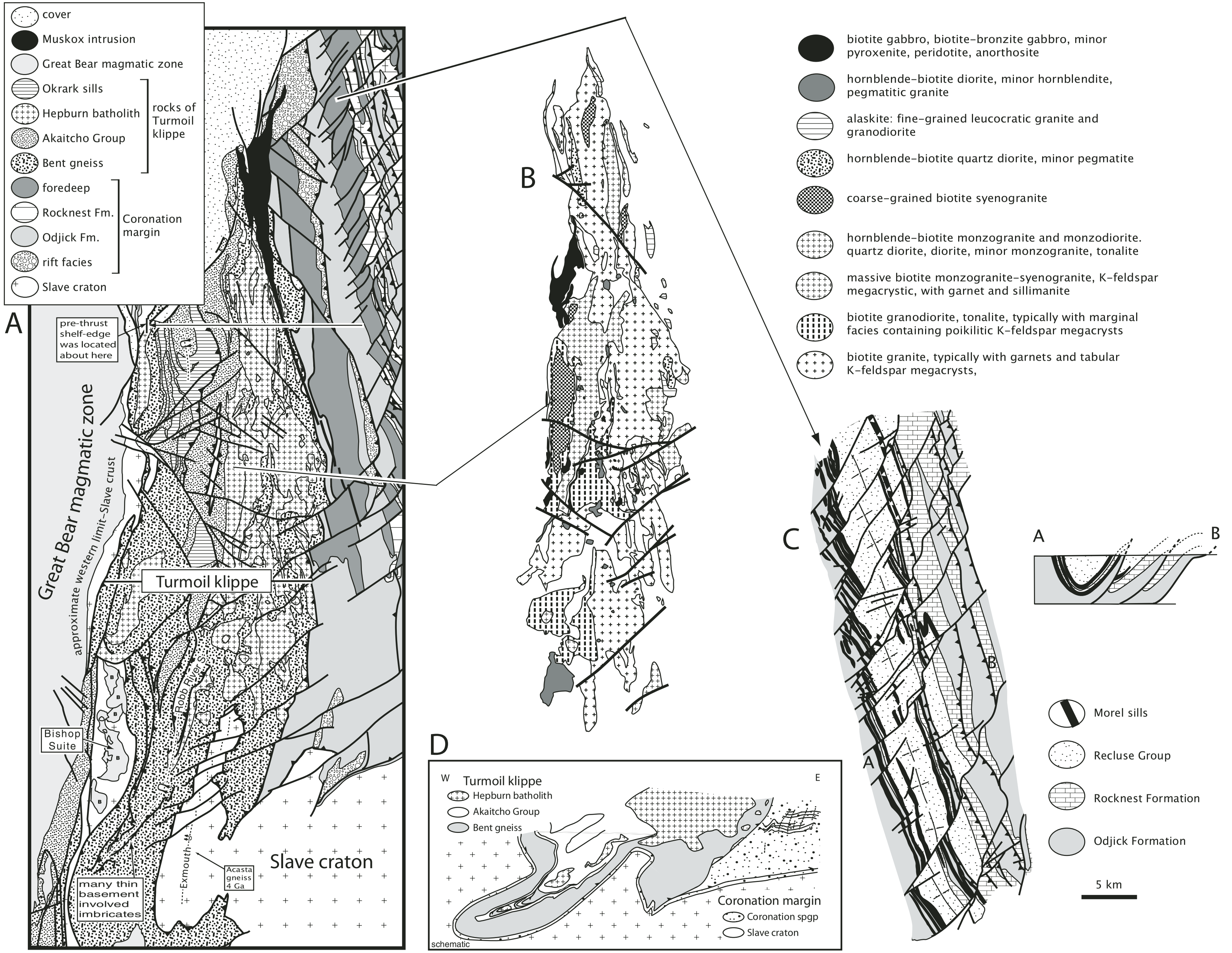

Figure 10. Geological sketch maps illustrating geological relationships in Wopmay orogen. (A) Sketch map showing the geological setting of the Turmoil klippe, Bent gneiss, Akaitcho Group, and Hepburn intrusive suite. Robb River S—Robb River syncline. (B) Various plutonic units within the main mass of the Hepburn intrusive suite. (C) Detailed map of the relationships at the shelf edge of the Rocknest formation showing thrust faults and Morel sills. (D) Schematic composite cross section through the internal zone showing the different structural levels within Turmoil klippe (spgp—Supergroup).

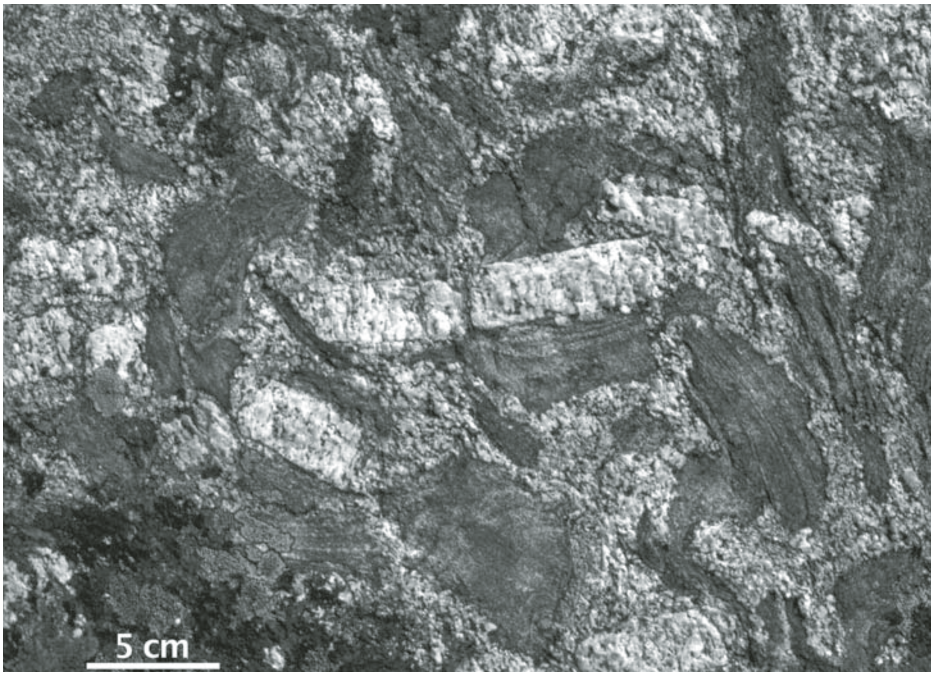

Figure 11. Porphyritic granite of Hepburn Batholith choked with metasedimentary enclaves. Although the dominant lithology is peraluminous granite, plutons within the batholith range in composition from gabbro to granite. Based on crosscutting relationships, magmas of the Hepburn Batholith became more mafic with time, more metaluminous, and tended to be smaller in volume. Older magmas were strongly peraluminous, full of metasedimentary inclusions, and locally garnet and sillimanite.

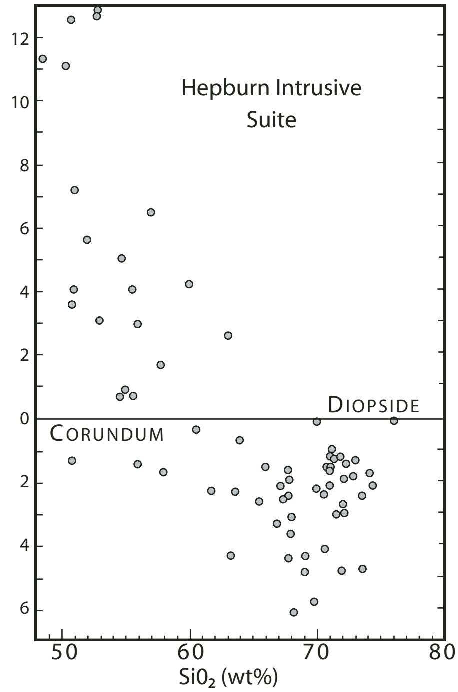

Figure 12. Normative values plotted versus SiO_2 for plutons of the Hepburn Batholith. Note that there is a general trend for peraluminosity to increase with increasing silica content. Because magmas of the batholith became more mafic with time, in a crude way, the diagram shows the ages of the plutons, with younger magmas to the left and older to the right. Figure is modified from LaLonde (1986).

However, large siliceous sills, known as Okrark sills (Fig. 10A), that intrude the sedimentary rocks of the structurally highest levels of the klippe may be shallow equivalents of the batholith.

Large areas of the western Slave craton are 3.0–>4.0 Ga, which is so much older than the dominantly Paleoproterozoic age of basement with Turmoil klippe that its isotopic signature should show up in any intrusions derived from it. Common Pb data obtained from leached feldspars of various magmatic suites are shown in Figure 13. The two plutons of the Bishop suite are clearly distinct from the igneous products that involved Hottah crust (i.e., Hepburn intrusive suite, Akaitcho Group, and Hottah terrane rocks). The Bishop suite plutons are related to the Great Bear magmatic zone, and they intruded the Turmoil klippe *after* its structural emplacement above the Coronation margin. Therefore, they bear an Archean isotopic signature that is lacking in the Hepburn intrusive suite and other constituents of Hottah terrane. Neodymium isotopes (Fig. 13) also indicate a Hottah, not a Slave, source for the Hepburn intrusive suite. These data support the field observations that rocks of the Hepburn intrusive suite were not derived from, nor did they pass through and assimilate with, Slave basement (Bowring, 1984; Bowring and Podosek, 1989; Housh et al., 1989). Instead, they were derived from Paleoproterozoic basement and were emplaced within Turmoil klippe prior to its emplacement on the Coronation margin.

Surprisingly, U-Pb zircon geochronology in the Hepburn intrusive suite is not generally complicated by inheritance (Fig. 4), and magmatic ages of individual plutons range from 1.9 Ga to 1.88 Ga (Bowring, 1984). Thus, plutons of the suite partially overlap in age, but they are mostly younger by a few million years than volcanism of the Akaitcho Group. Plutonism ceased at about the time the foredeep developed around 1.88 Ga. One garnetiferous pluton, with an obvious inherited component, yielded an age of 1942 ± 2 Ma for the inherited zircons and 1887 ± 4 Ma for the magmatic zircons (Bowring, 1984). Overall, the rocks of Hepburn Batholith yielded no evidence of an Archean inherited component. This lends further credence to the interpretation that rocks of the Hepburn intrusive suite are exotic with respect to Slave craton and the Coronation margin.

MOREL SILLS

A 200-km-long, 10-km-wide swarm of dominantly low-K_2O gabbroic intrusions, known as Morel sills, is localized at the shelf-slope break of the Coronation margin (Figs. 1, 10C, and 14). The term "foredeep magmatism" is used because they intrude trench-fill turbidites of the foredeep sequence as well as underlying passive-margin strata, yet they are deformed by the folds and thrusts of the foredeep inner slope (Hoffman, 1987). That the sills are intrusive and not structurally emplaced is proved by fine-grained, chilled, marginal zones and by narrow metamorphic aureoles in their wall rocks. They must have been intruded within the axial zone of the foredeep (Hoffman, 1987).

No feeder dikes for the intrusions were found in the field despite the resistance of mafic rocks to erosion relative to the foredeep strata (Fig. 14). Since the sills form a northerly trending swarm, their feeders likely have the same strike as the margin and the Calderian folds and

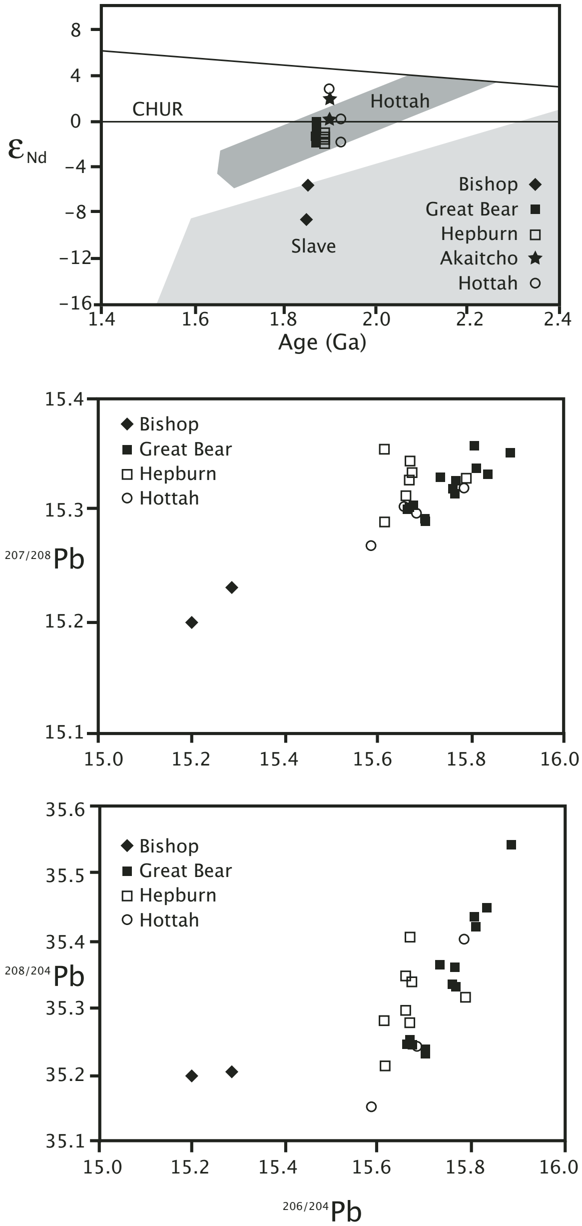

Figure 13. (A) Relationship of age and initial Nd at that age for rocks of various magmatic suites of Wopmay orogen. Prospective end-member sources are shown. Figure is after Bowring and Podosek (1989). Some magmatism of the Great Bear magmatic zone spilled over to the east, where it melted and intruded Slave craton. Those plutons are known as the Bishop suite, and they have markedly different Nd signatures from other magmatic suites. CHUR—Chondritic Uniform Reservoir. (B) Common Pb of feldspars, extracted from different magmatic suites of Wopmay orogen and leached, illustrate the similarity of rocks belonging to Hottah, Great Bear, and Hepburn suites. Note that plutons of the Bishop suite have markedly different Pb signatures from other magmatic suites. Data are from Housh et al. (1989). Both the Nd and Pb data provide strong support for the concept based on field relations that rocks of the Hepburn Batholith are exotic with respect to Slave craton and were derived from Hottah basement.

thrusts. Hildebrand and Bowring (1999) used this as evidence that the sills reflected east-west extension during their emplacement.

STRUCTURE OF THE COLLISIONAL ZONE

The structure of Coronation margin is complex in detail, but it involved an early thin-skinned fold-thrust event—during which Paleoproterozoic cover was foreshortened and translated eastward relative to cratonic basement—followed by thick-skinned events that deformed the orogen into large-scale folds with amplitudes of 5–15 km, first coaxial and later oblique to the thin-skinned structures (Hoffman et al., 1988).

The thin-skinned deformation produced closely spaced, steeply ramped thrusts and related folds developed in a prograde ("piggy-back") sequence above a continuous sole thrust (Fig. 15). The sole thrust is located 100–300 m above the basement surface, and numerous reversals of vergence, coupled with the steep ramp-angles and lack of change in average structural level, indicate that the active belt was a wedge of low taper (Tirrul, 1983; Hoffman et al., 1988). Estimates of shortening between the shelf edge and the frontal thrust are ~45%, but these are minimum estimates because conservative values of slip were applied to thrusts of unknown displacement (Tirrul, 1983). The estimated shortening restores the shelf-edge and Morel sill swarm almost to the Medial zone (Fig. 10A)—coincident with the western edge of Slave basement (Hoffman, 1984; Bowring and Podosek, 1989; Housh et al., 1989; Hildebrand et al., 1990; Cook et al., 1999).

The structure of Turmoil klippe is not as well resolved, but some broad generalities have emerged. The structure is tripartite and three main structural levels preserved, from bottom to top: (1) dominantly polydeformed Bent gneiss, (2) thrust-fault slices carrying Bent gneiss and unconformably overlying sedimentary rocks of the Akaitcho Group, and (3) imbricate sedimentary and volcanic rocks of the Akaitcho Group.

Other than in the thrust slices at intermediate structural levels, the basement-cover contact occurs only along the western margin of the Hepburn Batholith, which implies that the batholith is exposed in cross section from east to west across the axial region of Exmouth anticline (Figs. 1 and 10). If this interpretation is correct, then the batholith was rotated close to 90° during later folding, as originally suggested by King (1986).

At intermediate levels within the klippe where thin basement-cover thrust slices dominate, the structure appears more complex, but this may be because the basement-cover contact makes an excellent marker to resolve structural complexities. Hepburn intrusions occur near the basement-cover interface, but they are sparse. The general structure consists of folded imbricate thrust sheets typically carrying both basement and cover. The thrusts sheets are very thin, in places only a few meters thick, and yet they are remarkably continuous along strike, since they are traceable for many kilometers.

The thrust faults truncate penetrative fabrics in the basement, have no apparent associated mineral lineation, and carry members of the Hepburn intrusive suite. Reclined, tight to isoclinal folds of bedding and a crudely horizontal axial planar foliation are particularly well-developed in the supracrustal rocks. Many of the folds are outcrop-scale and are distinctly asymmetrical, and the asymmetry suggests topside-over-bottom movement to the west. Therefore, Hildebrand et al. (1991) inferred a westward vergence for many of the thrust faults within the klippe.

Due to the regional northward plunge, the highest structural levels within Turmoil klippe are found in the northern part of the area. There, the Robb River syncline (Fig. 10) is cored by a thrust slice containing thick sequences of turbidites cut by rhyolitic sills. The thrust slice is bounded on the east by a major thrust fault (Easton, 1981a) and on the west by the basal

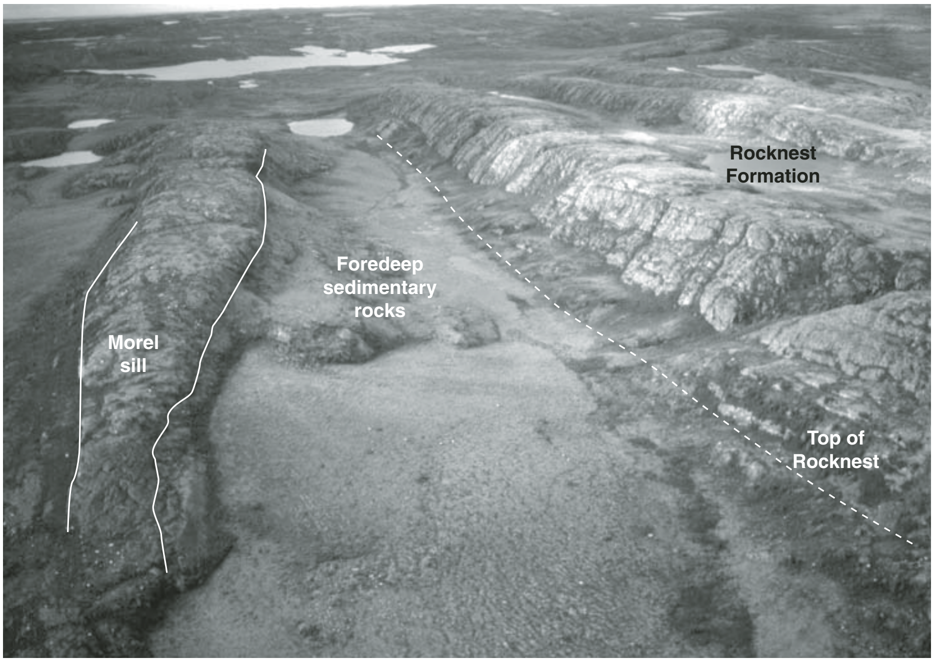

Figure 14. Aerial view of Morel sill (left) near the Rocknest shelf edge.

thrust of the Turmoil klippe. It is likely that additional thrusts occur within the turbidite-sill package, but lithologies are not distinct enough to resolve individual faults at mapped scale. Bent gneiss only outcrops in the western part of the slice, which, if it is depositional basement for the supracrustal package, suggests that the thrust at the base of this slice splays from the sole thrust and ramps upsection to the east.

Although Turmoil klippe as a whole was thrust eastwardly over Coronation margin, rocks within the klippe were, at least in part, thrust westwardly as discussed previously (Hildebrand et al., 1991). Relationships between the thrusting in Coronation margin and that in Turmoil klippe are unresolved, except that thrusting in the klippe must be younger than 1.88 Ga, the age of the youngest rocks involved in thrusting ,and older than final emplacement of the klippe. Most likely the westward-directed thrusts in the klippe are back thrusts formed as the Turmoil klippe ramped over the western edge of Coronation margin—more or less contemporaneously with the thin-skinned easterly directed thrusts.

The coaxial (north-trending) thick-skinned folding event is younger than rocks of the 1.875–1.84 Ga Great Bear arc (discussed later herein) because it folds those rocks as well as rocks of the Calderian collision zone (Hildebrand et al., 1990). Some workers (Hoffman et al., 1988; St-Onge, 1987; St-Onge and King, 1987a, 1987b; King, 1986) suggested—based on metamorphic data thought to be syncollisional, but now known to predate the collision—that the thick-skinned folds developed as a second phase of deformation during the Calderian orogeny. This deformational event involved minimal thrusting, at least as far as we were able to determine in the field, and instead led to basement-involved folds with amplitudes of 5–15 km. Although not part of the Calderian orogeny, the thick-skinned deformation is important because it, and an even younger set of broad, open cross-folds, largely controls the large-scale map patterns within the orogen (Fig. 1). For example, supracrustal rocks within the Turmoil klippe are preserved in the core of a thick-skinned syncline—the Robb River syncline—and the Acasta gneiss is exposed to the east in the adjacent thick-skinned Exmouth anticline (Fig. 10A).

GREAT BEAR MAGMATIC ZONE

Although not the subject of this paper because its rocks clearly postdate the Calderian orogeny, volcanic and sedimentary rocks of the Great Bear magmatic zone (Fig. 1) constrain several key aspects of the Calderian orogeny, so a short description is warranted. The Great Bear magmatic zone occupies most of the western exposed part of Wopmay orogen, and it is an ~100-km-wide belt of dominantly subgreenschist-facies volcanic and plutonic rocks that outcrop over a strike length of 450 km and unconformably overlie rocks deformed and metamorphosed during the Calderian orogeny. The zone can be traced southward for an additional 500 km along strike beneath a thin veneer of Paleozoic cover (Coles et al., 1976; Hildebrand and Bowring, 1984; Hoffman, 1987). To the north, the magnetic anomaly curves sharply to the west and continues for an additional 300 km (Fig. 16). Thus, it is comparable in size to the great continental magmatic arcs of the circum-Pacific region.

Except for early tholeiitic basalts, the volcanic rocks of the Great Bear magmatic zone are mostly calc-alkaline and span the entire compositional range from basalt to rhyolite (Hildebrand et al., 1987b). Intermediate composition rocks dominate, and, even where more siliceous compositions are abundant, andesitic lavas occur intercalated with the more siliceous units. Modes and geochemistry of plutonic rocks are virtually identical to classical "Cordilleran-type" batholiths (Hildebrand et al., 1987b). The overall structure of the zone is crudely synclinal (Hoffman and McGlynn, 1977; Hildebrand and Bowring, 1984) such that the oldest supracrustal rocks occur in the east and west, although sparse exposures of the lower part of the pile and its basement occur elsewhere within the zone. Overall, there is little doubt but that the Great Bear magmatic zone represents a typical continent arc (Hildebrand et al., 1987b). In the model presented here, no oceanic lithosphere remained to be subducted east of the Great Bear magmatic zone, and thus the arc was generated above a subduction zone descending from the west. The east-dipping zone of seismic reflectors at mantle depths beneath the Great Bear magmatic zone is presumed to be a fossil relic of this subduction zone (Cook et al., 1999).

The oldest ages of volcanic rocks in the Great Bear magmatic zone (Fig. 4) are statistically indistinguishable from the ash bed in the Calderian foredeep dated at 1882 ± 4 Ma (Bowring and Grotzinger, 1992). As the ash bed was deposited before Calderian deformation in the area of the foredeep where the ash bed occurs, the length of time between the end of Calderian deformation and initiation of Great Bear magmatism must have been short, <9 m.y. given the analytical uncertainties in the ages (Fig. 4).

Along the eastern margin of the Great Bear zone, sandstones, siltstones, mudstones, basalt, and locally stromatolitic dolomite, unconformably overlie the rocks deformed and metamorphosed during the Calderian orogeny. Stubby tongues of granite-dominated talus occur within the finer-grained sedimentary rocks and were derived from local westerly facing scarps created by west-side-down normal faults (Hoffman and McGlynn, 1977; Hildebrand et al., 1990; Hildebrand and Bowring, 1988). Thin ash-flow tuffs fill paleovalleys, and distinctive porphyritic intrusions cut the entire sequence. Elsewhere within the zone, mature quartz arenite directly above the unconformity is overlain by

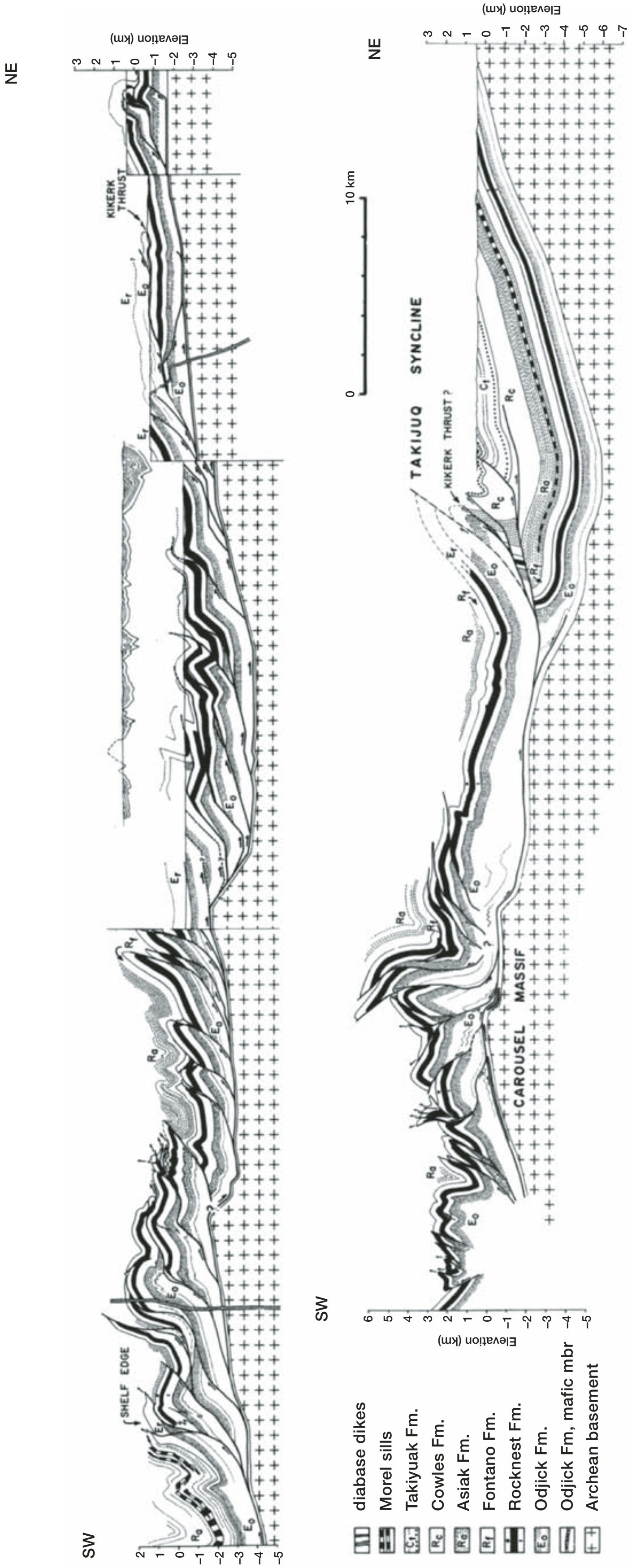

Figure 15. Balanced composite cross sections through the fold-and-thrust belt by Rein Tirrul (Hoffman et al., 1988). Line A–H is in the northern part of the belt, whereas line K–L is in the south (Fig. 1). The Odjick and Rocknest Formations are the dominant formations of the passive margin, whereas the Fontano, Asiak, Cowles, and Takiyuak Formations are formations of the foredeep. No vertical exaggeration.

intercalated ash beds and fine-grained epiclastic rocks (Hildebrand, 1994). Despite detailed mapping, we did not find sedimentological evidence for high-standing terrain to the east, which suggests the possibilities that: (1) mountain drainage funneled sediment in other directions; (2) the mountainous terrain had collapsed; (3) mountains never existed there; or (4) some combination of the first two possibilities.

DISCUSSION

During the 1970s, Wopmay orogen was generally interpreted as a Cordilleran-type margin with easterly dipping subduction and backarc thrusting without collision (Hoffman, 1973; Hoffman and McGlynn, 1977; Badham, 1978), but subsequent mapping led to models in which the orogen was interpreted as an arc-continent collision zone (Hoffman, 1980; Hildebrand, 1981). It was the failure to recognize crystalline basement in the Turmoil klippe and the dating of the Akaitcho Group (Fig. 4), given the assumption that it represented initial rift magmatism (Hoffman, 1980), that led to the hypothesis that the Coronation margin was short-lived (Hoffman and Bowring, 1984) and developed in a backarc setting (Reichenbach, 1991; St-Onge and King, 1987a, 1987b; Lalonde, 1989). Here, we suggest that the Hottah terrane and Turmoil klippe, with its cover of Akaitcho Group, are better interpreted to be exotic with respect to Slave craton on the basis of U-Pb geochronology, isotopic constraints, lithological differences, and overall stratigraphy. The basement rocks in Slave craton and Hottah terrane are different in age. More importantly, calc-alkaline plutons and volcanic rocks of Hottah terrane and Turmoil klippe are largely the same age as passive-margin sedimentation on the Coronation margin (Bowring, 1984; Bowring and Grotzinger, 1992).

Magmatic History of Hottah Terrane before the Hepburn Intrusive Suite

Arc magmatism within the Hottah terrane–Turmoil klippe changed style, composition, and location with respect to time. Older magmatism occurred in the west, where it was subaerial. Outpourings of nearly aphyric magma within the Bell Island Bay Group, ranging in composition from basalt to rhyolite, led to thick cooling units of ash-flow tuff, as well as lavas, and domes. The nearly aphyric nature of the intermediate-siliceous volcanic rocks points to low residence time in the crust and suggests that the area was extending somewhat as the calc-alkaline suite was erupted (Reichenbach, 1991). To the east within Turmoil klippe, vast amounts of tholeiitic basalt—mostly younger

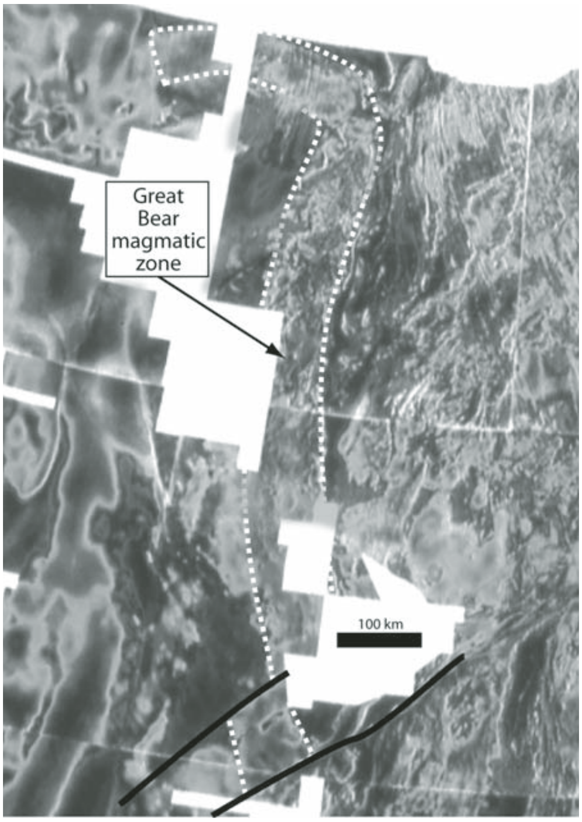

Figure 16. Aeromagnetic map of the Great Bear magmatic zone showing extent to south and lengthy curve to west at its northern end.

than magmatism of the Bell Island Bay Group (Fig. 2)—were erupted and intruded into a subsiding subaqueous sedimentary basin (Easton, 1982). The presence of chiastolite in intercalated sedimentary rocks indicates high-temperature–low-pressure metamorphism, compatible with extension and magmatic heat advection (Vernon et al., 1993). Overall, the facies of precollision rocks atop Hottah terrane indicate a temporal and geographic progression from subaerial, calc-alkaline volcanism in the west to subaqueous eruptions of tholeiitic basalt in the east. Volcanic rocks are slightly younger to the east, suggesting that magmatism within the arc was migrating in that direction, that is, toward the ocean, and erupting through crust under extension.

The best modern analog for the oceanward migration of arc magmatism, known from nearly every modern subduction zone in the circum-Pacific region, is subduction rollback, which generates its magmatism coincident with rollback of the subduction hinge in the fashion originally proposed by Elsasser (1971) and amplified by others more recently (Dewey, 1980; Kincaid and Olson, 1987; Garfunkel et al., 1986; Uyeda and Kanamori, 1979; Carlson and Melia, 1984; Royden, 1993a; Royden and Burchfiel, 1989; Spence, 1987). Lateral migration, or retrograde slab motion as it is sometimes called, is common in modern subduction zones (Garfunkel et al., 1986) and occurs when the rate of subduction exceeds the rate of plate convergence (Dewey, 1980; Royden and Burchfiel, 1989; Royden, 1993a, 1993b). Some workers interpret the rollback of the subducting plate to subduction of progressively older, colder, and denser oceanic lithosphere and a progressive increase in slab dip, but because there is no simple relationship between slab dip and age of the lithosphere (Cruciani et al., 2005; Manea and Gurnis, 2008), the subduction angle during rollback does not necessarily change, at least until terminal collision. Nevertheless, old lithosphere does favor the formation of backarc basins because its greater density leads to greater rollback (Molnar and Atwater, 1978).

The rollback and lateral migration causes extension in the overriding plate and oceanward migration of the arc front, such that arcs above retreating subduction zones are erupted in extensional regimes that actively migrate across ocean basins (Karig, 1970; Hamilton, 1988; Apperson, 1991; Hamilton, 1995; Royden, 1993a, 1993b). During, and just after, extreme arc extension, the typical calc-alkaline suite is replaced by a bimodal suite of tholeiitic volcanic rocks (Clift et al., 1995). A useful modern analog of an arc on continental crust undergoing extension is the dominantly rhyolitic Taupo volcanic zone of New Zealand, where active arc extension is propagating southward onto continental crust (Cole, 1986, 1990).

In the Taupo zone, huge volumes of rhyolitic magma were erupted as the arc crust was stretched ~80% and broken due to the retreat of the subducting slab. The active zone of extension occurs in the easternmost part of the basin (Davey et al., 1995), which suggests that extension is migrating eastward toward the trench with time, just as expected if it is due to trench rollback. The magmas erupted through the extending crust are dominantly rhyolite, but they include high-alumina basalt, andesite, and minor dacites, which resulted from the mixing of andesitic and rhyolitic magmas (Graham et al., 1995). Although the rocks are somewhat bimodal now, presumably with greater extension, the quantity of basalt would increase and, with sufficient extension, would be truly bimodal as they are farther north, where the extending arc crust is transitional to oceanic.

Thus, we conclude that rocks of the Bell Island Bay and Akaitcho Groups, as well as slightly older calc-alkaline plutons of the Hottah terrane, were all part of an arc built on early Proterozoic Hottah basement by west-dipping subduction of oceanic lithosphere connected to Slave craton (Fig. 17A). Because the locus of magmatism closely tracks the location of the subducting slab in modern subduction systems, it seems likely to have been the same for subduction during the Proterozoic. We suggest that the rapid eastward migration of magmatism and the regional extension were caused by rollback of the subduction hinge (Fig. 17B). In this scenario, initial magmatism in the basin would have been distinctly bimodal because the magmas would have been able to rise rapidly to the surface without extensive crustal assimilation or mixing.

Hepburn Intrusive Suite

Continued mafic magmatism—and a change in stress regime from extension to compression as the forearc region began to impinge on Slave craton—caused the basalts to interact more intensely with Hottah crust and sedimentary cover to generate hybrid magmas of the Hepburn intrusive suite. The magma batches rose diapirically toward the surface until they encountered pelitic sediments of the arc basin–forearc region. The magmas were able to assimilate some pelite but were frozen in place as the increased alumina content elevated the solidus and/or as they assimilated H_2O and became saturated (Burnham, 1979). Thus, they are arc-type plutons that gained their peraluminous nature from assimilation of wall rocks in the zone of emplacement, not their source region (Bowring, 1984; Lalonde, 1989). The overall trend from siliceous

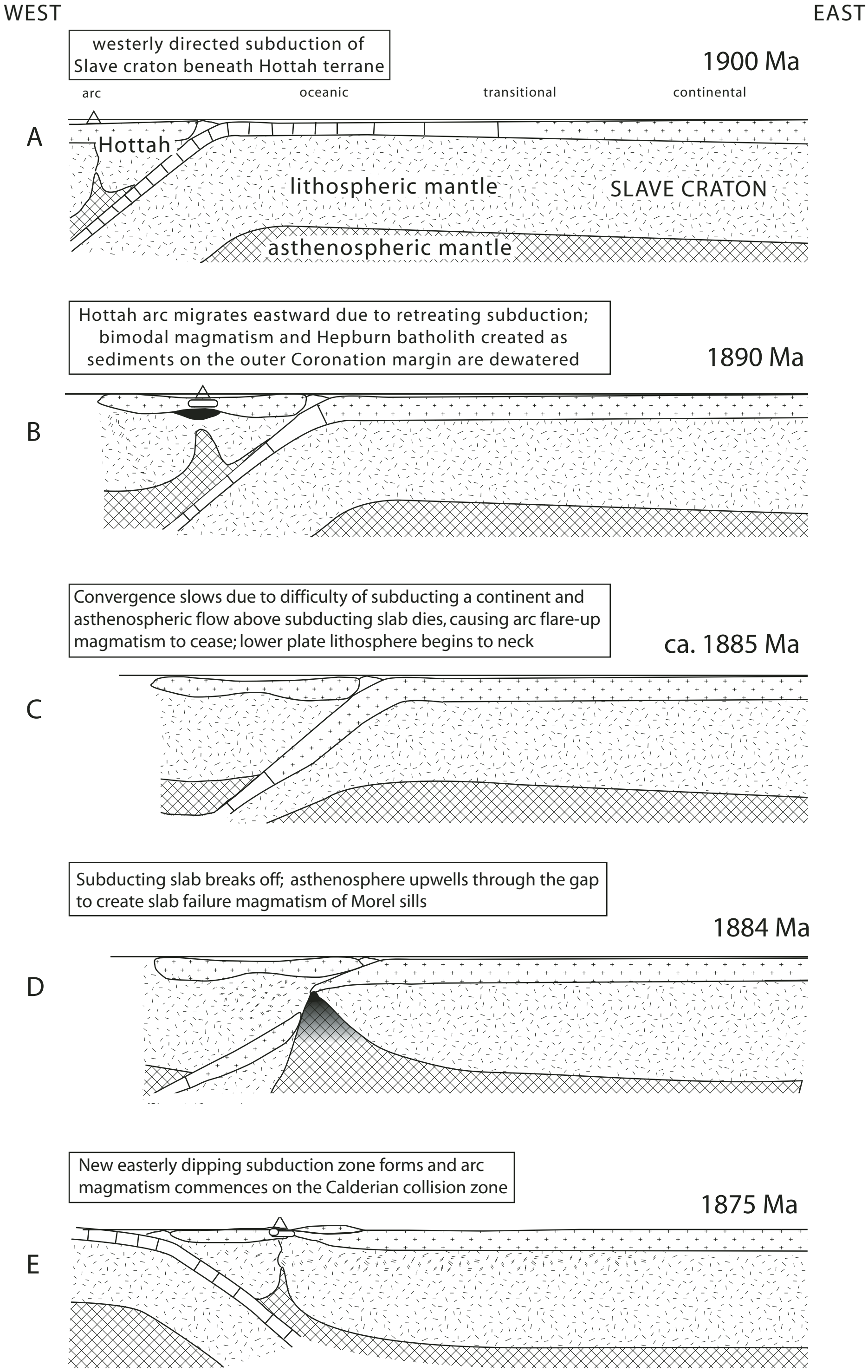

Figure 17. Geological model for the evolution of Wopmay orogen. (A) Normal subduction of oceanic crust and lithosphere of Slave plate beneath the arc-bearing Hottah plate. (B) Retrograde motion of the subduction system causes extension in the Hottah plate; arc magmatism migrates trenchward and changes from calc-alkaline to bimodal tholeiitic; and Hepburn intrusive suite is emplaced. (C) As the leading edge of Slave plate is subducted, Turmoil klippe is emplaced on it. (D) Failure of Slave plate causes asthenospheric mantle to rise upward, melt, and enter the torn edge of Slave craton, where they form the Morel sills. (E) Within 9 m.y. of slab breakoff, a new east-dipping subduction zone causes magmatism of the Great Bear magmatic zone.

to intermediate/mafic and from foliated to nonfoliated suggests that the latest magmas formed as extensional stresses increased somewhat, such that there was less melting and assimilation of continental crust with increased time.

Recently, Collins and Richards (2008) argued that S-type granitoids originate in backarc or intraplate settings where increased slab dip causes melting of backarc basinal sedimentary rocks following a crustal thickening event. In Wopmay orogen, the S-type batholith was viewed to have possibly developed in a backarc setting (Hoffman et al., 1988; St-Onge and King, 1987a, 1987b), but we now recognize that the Hepburn intrusive suite developed on the upper plate just prior to collision. We suggest a similar tectonic setting for the Australian (Tasman) examples.

Any process that pumps basaltic magma into the forearc region, or an extensional basin within the arc, might produce similar composition rocks. Therefore, we considered other possible mechanisms than that favored here. For example, we evaluated the effects of ridge subduction, which creates a slab window beneath the overriding plate (Thorkelson and Taylor, 1989; Hole et al., 1994; Madsen et al., 2006). A well-studied example of ridge-subduction magmatism is located in Alaska where Paleocene-Eocene ridge subduction produced a suite of calc-alkaline, but peraluminous, plutons ranging in composition from gabbro to granite (Bradley et al., 2003; Kusky et al., 2003). Another example is located in the area east of New Guinea where the Woodlark spreading ridge has entered the Solomon Trench (Taylor and Exon, 1987).

Although not generated by the same process, ridge subduction does allow mantle-derived melts to interact with forearc basement and sedimentary cover, just as does slab rollback, to create a suite of plutons that are compositionally similar to those of the Hepburn Batholith. In the case of Wopmay orogen, the extension and the magmatism of the Akaitcho Group–Hepburn intrusive suite occurred just before collision with Slave craton, so they cannot have been related to ridge subduction.

One peculiar twist, demonstrated in experiments but not yet documented in the field, is the syncollisional subduction of the entire forearc region of an extending overriding plate (Chemenda et al., 2001; Boutelier et al., 2003). Experiments by these authors only modeled oceanic arcs, but they suggest the possibility that the entire leading edge of the overriding plate back to the zone of lithospheric thinning could founder and be subducted. This mechanism, if valid for continental lithosphere, could provide an explanation for the occurrence of the Hepburn intrusive suite occurring so close to the apparent leading edge of Turmoil klippe.

Our model for plutons of the Hepburn intrusive suite is somewhat different than that recently proposed for the Donegal intrusions of Ireland by Atherton and Ghani (2002) despite the similarities between the suites. They argued that the Main Donegal granites were the result of slab failure and consequent heating of the upper plate by hot asthenosphere impinging on the base of the crust after the closure of Iapetus, whereas we argue that the Hepburn Batholith was generated prior to slab failure by rollback of the lower plate.

Morel Sills

In Wopmay orogen, it is the younger Morel sills—emplaced into the lower plate during the collision—that are the apparent signature of slab breakoff. The Morel sills, as discussed previously, intrude passive-margin and foredeep cover on Slave craton and are a result of slab breakoff in the usual sense (Price and Audley-Charles, 1987; Sacks and Secor, 1990; Davies and von Blanckenburg, 1995; Davies, 2002; Levin et al., 2002; Haschke et al., 2002). In this model, when the strength of the continent-oceanic slab is exceeded by the competing forces of bouyancy of the attached continent and downward pull of the oceanic lithosphere, the oceanic part of the plate tears off at its weakest point and sinks into the mantle due to its greater density (Fig. 17D). Different workers place the weakest zone at different places (Davies and von Blanckenburg, 1995; Atherton and Ghani, 2002; Cloos et al., 2005), but in Wopmay orogen, the abrupt western edge to the Slave craton within the Medial zone likely marks the zone of detachment (Hildebrand and Bowring, 1999).

This is consistent with palinspastic reconstruction of the thrust belt, which places the restored shelf edge of the Rocknest Formation in this region (Tirrul, 1983). It probably marks the eastern limit of normal faulting during initial rifting of Coronation margin, because to the east, there are no normal faults mapped that cut the unconformity at the base of the passive-margin prism despite nearly continuous outcrop across strike. Thus, Hildebrand and Bowring (1999) concluded, at least in Wopmay orogen, that the weakest part of the system was not at the oceanic-continent interface but at the eastern limit of upper-crustal extension. Cloos et al. (2005) argued that the break-off zone occurs in the lower crust beneath the continent itself, due to weak coupling between the upper and lower crust, rather than the rather strongly coupled oceanic and transitional crust, but the result is approximately the same.

Another example of intense lower-plate magmatism, which appears to reflect slab failure, occurred during the Grampian orogeny, when 470 ± 2 Ma mafic magmas of the Insch gabbro suite intruded Dalradian, lower-plate sediments during arc-continent collision (Dewey, 2005). In this collisional belt, rift deposits appear to have been torn away with the descending slab, much the same as in Wopmay orogen. Thus, in ancient collisional zones, we believe that the presence of mafic magmatism that postdates and intrudes passive-margin sedimentary rocks and, in places, foredeep deposits, yet predates or is even synchronous with thrusting and metamorphism is an indication of slab failure. We know of no other mechanism that can cause margin-parallel magmatism in a cold passive-margin setting. Furthermore, many collisional orogens have sparsely preserved rift assemblages, and their general absence points to slab failure as a likely mechanism for their demise.

Because arc magmatism so closely tracks the subduction zone, the progression from extension, with oceanward magmatic migration in the upper plate, through the development of voluminous Cordilleran-type batholiths, to failure of the lower plate documents the progressive subduction, rollback, and failure of subduction of an old cratonic margin (Dewey, 1980; Hamilton, 1995, Royden, 1993a, 1993b). The rollback can produce a wide variety of features in the arc, such as extensional basins and even new oceanic crust, but the key developmental scheme to look for in older orogens is the progressive migration of arc magmatism toward the suture and the possible sudden development of Cordilleran-type batholiths as the water-rich sedimentary apron of the continental margin enters the subduction zone (Hildebrand, 2009). In long-lived systems, the upper plate in the collisional belt might include a series of collapsed marginal basins and remnant, rifted arcs, all sitting behind the youngest arc. Compositional changes may also help resolve the tectonic setting and/or processes, as magmatism in Hottah terrane changed from a calc-alkaline basalt-andesite-dacite-rhyolite suite to a dominantly tholeiitic bimodal suite and then back again to a calc-alkaline batholithic suite emplaced just prior to terminal collision.

Rollback and migration of magmatism can continue until subduction starts to involve continental crust. When the edge of a craton and its overlying continental margin is subducted, the thick, cold and dense oceanic slab is competing against the buoyancy of the continent. If the continental mass is small, or without thick lithospheric mantle, it might be subducted and recycled, but where it is large with a well-developed subcontinental root, subduction is impossible. Either the entire system grinds to a halt due to the inability of the oceanic slab to overcome the buoyancy of the continent, or—especially in the normal case of oblique subduction—the subducting slab breaks off and sinks into the mantle. As the slab tears, asthenosphere rises upward to melt and generate mafic magmatism, which then can rise into the collision zone, where it intrudes sedimentary rocks of the passive margin during thrusting. During slab breakoff, magmas may or may not penetrate the upper plate, depending on the rate at which breakoff, and hence lithospheric necking, occurs (Cloos et al., 2005). In Wopmay orogen, there does not appear to have been any obvious upper-plate magmatism of the appropriate age. It is possible that the very youngest mafic rocks of the Hepburn intrusive suite were generated by slab failure, but the absence of any isotopic signature of the Slave craton within the plutons constrains their intrusion to predate emplacement of the Turmoil klippe. This could only happen if the slab breakoff was strongly diachronous such that plate convergence continued to pull the Slave craton beneath the Hottah terrane after the local slab tear developed, and even then, it implies that Turmoil klippe and the thrust belt in Coronation margin formed after slab tearing. A north-northeasterly trending mafic dike swarm, known as the Ghost swarm, occurs within the western Slave craton south of the exposed belt of Morel sills. Recent dating of baddeleyite from three different dikes (Buchan et al., 2009) yielded ages of 1884 ± 6, 1884 ± 2, and 1886 ± 5 Ma, which suggest that they were part of the slab breakoff magmatism and provide additional evidence that mafic magmas were entering the subducting plate at that time.

Hildebrand and Bowring (1999) expanded on the implications of slab breakoff for the preservation of the subducting plate. One likely possibility is that most, if not all, of the rift deposits on the margin may be recycled into the mantle along with the subducted slab. The absence of rift sequences and the mafic magmatism intruding the lower-plate prior to and/or during thrusting appear to be key diagnostics indicating slab breakoff.

Short-Lived Collision

When the subducting slab fails, the descending plate is freed from its oceanic anchor, thin-skinned thrusting ceases, and the lower plate immediately starts to rise isobarically. This leads to rapid uplift in the collision zone, which could generate gravitational failure and collapse of the thickened orogen. In Wopmay orogen, there are large-displacement normal faults within the Medial zone (Hildebrand et al., 1990) that attest to the gravitational collapse of the region above the hypothesized break-off zone.

Another aspect of the Calderian orogeny, not overtly discussed in this paper, but hinted at and worth noting in more detail, is its short duration: less than 9 m.y. (Fig. 4) from collision until eruption of postcollisional magmas of the Great Bear magmatic zone. Within this period, mafic magmas flooded the lower plate at the relict shelf margin during slab breakoff, Turmoil klippe was emplaced upon the Slave margin, and rocks of the collision zone were rapidly eroded and/or gravitationally collapsed—all prior to ~ca. 1872–1870 Ma, when volcanic and sedimentary rocks of the postflip Great Bear arc were deposited on the eroded collision zone. This implies that either high mountains never formed or that any mountains that developed must have collapsed and/or eroded rather rapidly.

Dewey (2005) favored crustal thinning and extensional collapse driven by subduction rollback and buoyancy forces to explain the lack of high mountains during the Grampian orogeny. When rocks on Coronation margin were pulled beneath the attenuated leading edge of Hottah terrane, they were thus pulled beneath very thin crust. This, when combined with lower-plate attenuation just prior to and during slab failure, could work to limit high mountains from ever forming or perhaps reducing their longevity because crustal thickness is never extreme. This is essentially the retreating subduction model of Royden and Burchfiel (1989) and Royden (1993a, 1993b), which suggests that strongly retreating subduction boundaries do not create high-standing mountains during collision. Within Wopmay orogen, the generally low metamorphic grade of the Calderian hinterland, the thin-skinned, low-taper thrust belt, and the domination of the foredeep by orogenic flysch are all expected characteristics in the retreating subduction model as elucidated by Royden (1993a). In these types of collisions, subduction ceases rapidly when thick buoyant continental crust enters the subduction zone and the dense oceanic slab tears off. This decapitates the collisional zone from its gravitational driving force, although Laramide-style, thick-skinned deformation can occur in the foreland if the breakoff is diachronous.

While a 9 m.y. time frame for collision and collapse of the orogen might appear short, the island of Taiwan provides an excellent modern analog and demonstrates that it can be very short. There, oblique collision between Eurasia and the actively retreating Luzon arc provides a timeline of 4–5 m.y. for arc-continent collision in the south, slab breakoff, collapse of the mountain belt in the northern part of the island, and initiation of oppositely directed subduction beneath the Ryukyu arc (Viallon et al., 1986; Suppe, 1987; Lallemand et al., 2001; Huang et al., 2006). Relief of 3–4 km on Taiwan disappears rapidly to the north as the mountains collapse by gravitational failure, mass wasting, and stream erosion (Fig. 18). Overall, the 3–4-km-high collisional mountain chain of central Taiwan was tectonically reduced to the 2000-m-deep basin of the Okinawa Trough within 3 m.y. as the subducting slab failed and the Okinawa Trough propagated into the orogen (Teng, 1996).

If the model developed here is correct, then many arc-continent collisions should involve upper-plate extension and oceanward migration of arc magmatism, due to rollback of the subducting slab and shutdown of that magmatism, all followed by magmatism in either or both the lower and upper plates as the subducting slab fails. Peraluminous batholiths arise when large volumes of arc-generated magma rise into the sedimentary basins of the retreating and extending arc where they interact with pelitic sediments. Some such batholithic flare-ups may owe their ultimate origins to rapid and voluminous dehydration of the leading edge of the subducting continent and its slope-rise sediments as they enter the zone of dehydration–melt generation beneath the arc. This would occur right before foundering of the subduction system and slab break-off.

During and after slab failure, the orogen collapses, in part due to isostatic uplift of the partially subducted continent and partly due to the rise of hot asthenosphere through the torn slab. Within a few million years of slab failure, new arc magmatism, related to oppositely directed subduction, begins on top of the eroded collision zone. Short-lived orogeny may be characteristic of arc-continent collision and might be diagnostic.

The events in Wopmay orogen are fairly representative of other arc-continent collisions ranging from Paleoproterozoic to the present. In the Paleoproterozoic Penokean orogen of the Lake Superior region, where the Pembine-Wausau arc terrane collided with the southern margin of the Archean Superior craton, the upper-plate precollisional arc was strongly extensional and characterized by both calc-alkaline and tholeiitic volcanic rocks, later intruded by both flare-up and postcollisional, subduction-reversal batholiths (Schultz and Cannon, 2007). During the Cretaceous-Tertiary Cordilleran orogeny, western North America was partially subducted beneath an arc-bearing superterrane, and there are distinct periods of magmatism, including precollisional extensional arc, Cordilleran-type flare-up magmatism, slab breakoff magmatism, and postcollisional, subduction-reversal arc magmatism (Hildebrand, 2009). On the island of New Guinea, the Tertiary collision between the Australian and the Pacific plates involved the partial subduction and breakoff of the leading edge of Australia beneath the Melanesian arc, associated upwelling syncollisional magmatism, and postcollisional arc magmatism (Cloos et al., 2005).

In all four cases, the old, dense, and thick craton was the lower plate, which, when combined with similar relations in virtually every other orogenic belt, such as the Paleozoic Taconic (Williams, 1979) and Grampian (Dewey, 2005) orogens, as well as ongoing collisions in Taiwan (Suppe, 1987) and present-day northern Australia (Hamilton, 1979), suggests that new arcs—and hence, subduction zones—do not form by collapse of old and strong oceanic crust adjacent to continental margins (Cloetingh and Wortel, 1986) but instead form from young oceanic crust (Cloetingh et al., 1989). It appears that pristine ocean basins like the Atlantic must be "infected" with a parasite—a segment of subduction zone from elsewhere—which can then propagate to ultimately destroy its ocean-basin host (Mueller and Phillips, 1991). This implies that passive margins are converted to active margins mainly through subduction reversal following arc-continent collision.

CONCLUSIONS

(1) The evolution of the Calderian orogeny is best viewed as an arc-continent collision between the Slave craton and Hottah terrane. However, events before the collision played important roles in the assembly of the orogen. Precollisional rollback of the westerly descending Slave plate led directly to extension within the Hottah arc (Fig. 17). The extension caused magmatism in the Hottah terrane to switch from a calc-alkaline to a bimodal basalt-rhyolite assemblage.

(2) The leading edge of the Hottah terrane was thrust over rocks of the Coronation margin (Fig. 17) and is preserved in the Turmoil klippe. The leading edge of the klippe lies some 55 km east of the western edge of the Slave craton, as inferred from outcrop and isotopic studies. The original shelf edge of the Coronation margin restores nearly to the Medial zone, and the isotopic evidence that no Archean basement exists beneath the Great Bear magmatic zone suggests that any Slave crust formerly located there was subducted along with the oceanic slab.

(3) Upper-plate plutonism of the Hepburn Batholith progressed from larger-volume peraluminous granites and tonalites to smaller-volume metaluminous diorites, quartz diorites, gabbros, and pyroxenites. The batholith was emplaced along the interface between basement gneiss and its volcano-sedimentary cover. The magmas froze when they assimilated alumina and water from the basinal sediments. According to our model, Hepburn Batholith is

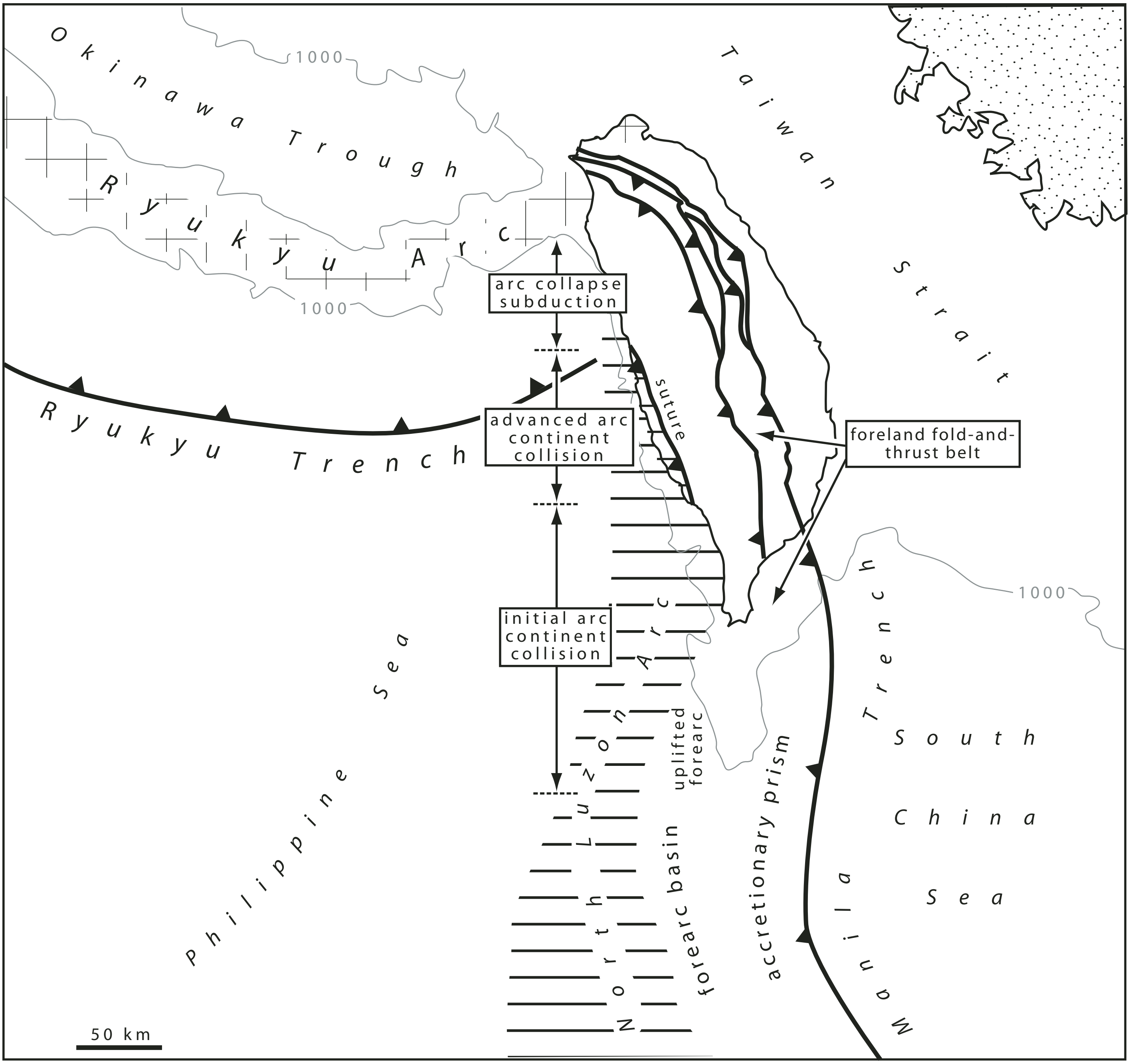

Figure 18. The arc-continent collision on Taiwan (after Suppe, 1987; Huang et al., 2006). The map is reversed (east is toward the left) for easier comparison with Wopmay orogen. A south-to-north transect of Taiwan shows all the phases of arc-continent collision, from incipient collision to collapse of the orogenic belt, which takes place in 4–5 m.y. In this reversed figure, the Philippine Sea plate is moving obliquely toward the upper right relative to Eurasia.

contained entirely with the Turmoil klippe and was emplaced into the arc during the 20 m.y. prior to the collision.

(4) As the arc collided, the oceanic portion of the Slave plate failed in extension and sank into the mantle (Fig. 17). Slab failure caused mantle-derived melts to rise into the formerly cold Slave plate, where they invaded the passive-margin shelf-edge to create a linear swarm of gabbroic sills. The mafic magma intruded undeformed passive-margin sediments and trench-axis turbidites yet experienced all the folding and thrusting associated with their incorporation into the forearc accretionary prism. They represent syn-collisional trench-axis magmatism associated with slab failure.

(5) Within 9 m.y. of the collision, subduction had stepped outboard of the newly accreted arc and was descending in the opposite direction—eastward in present-day coordinates (Fig. 17E). The new subduction led to the eruption of volcanic rocks of the Great Bear magmatic zone (Fig. 19), which lie unconformably on the eroded Calderian collision zone. The short

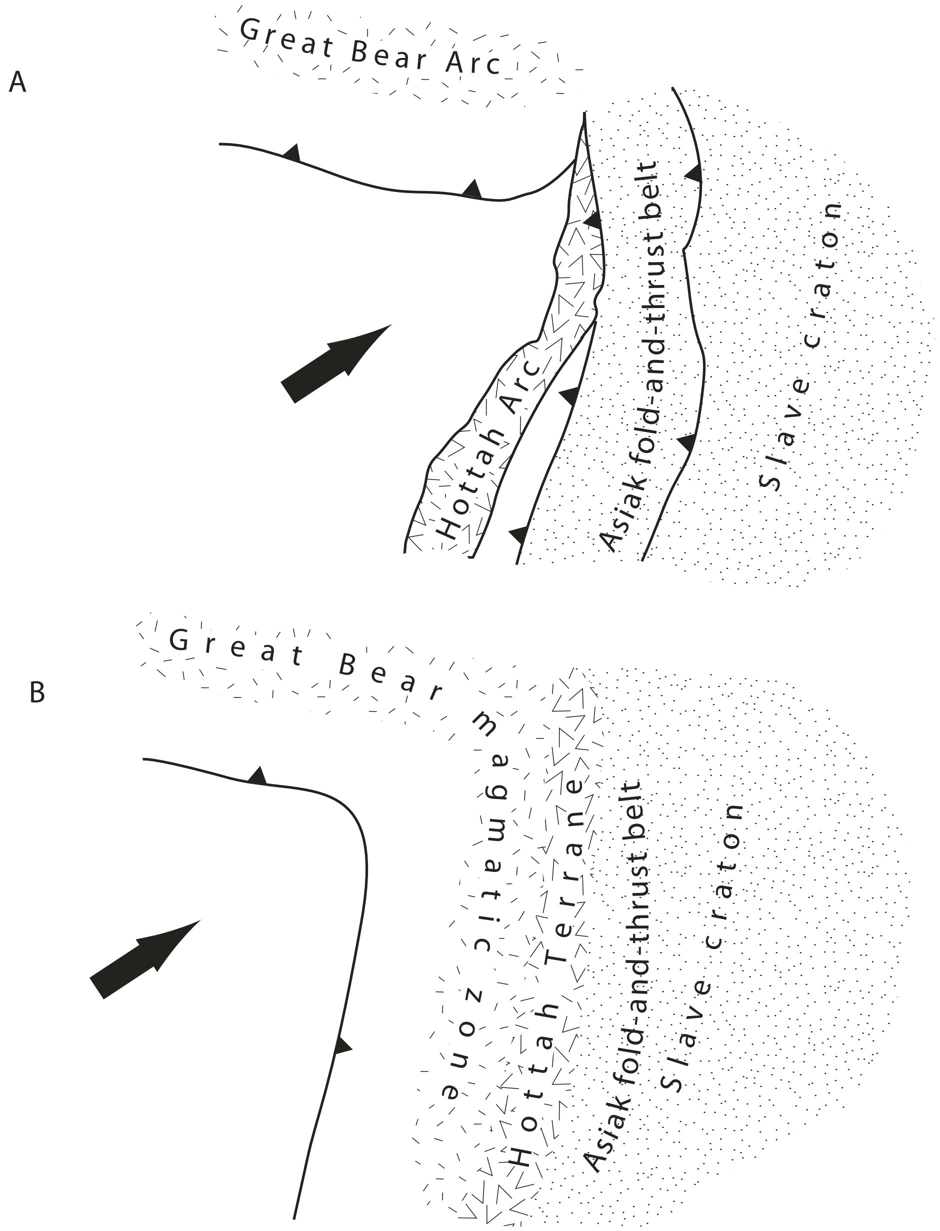

Figure 19. Schematic maps illustrating (A) possible large-scale plate geometry during and after arc-continent collision in Wopmay orogen. The obliquity of the collision is indicated by the NE-SW lineations at the base of the Turmoil klippe. This is similar to the current Taiwan geometry as shown in Figure 18. Note that magmatism of the Great Bear magmatic zone is analogous to the modern-day Ryukyu arc. (B) Following accretion of the Hottah arc to Slave craton, a new subduction started up in a few million years west of the amalgamated terrane. This model explains the huge bend seen in the positive magnetic anomaly of the Great Bear magmatic zone (Fig. 15).

length of time (<9 m.y.) between the onset of collision and Great Bear magmatism, coupled with the lack of voluminous detritus in the axial depression of the Great Bear arc, implies that the collision zone never had high relief, or that it collapsed rapidly. The timing, at least, is compatible with that for the active arc-continent collision and subduction polarity reversal in Taiwan, which takes place over an interval of 4–5 m.y. (Suppe, 1987).

(6) Old cratons are typically the lower plate in arc-continent collisions, which are the principal mechanism for turning passive into active margins.

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