UNITED PLATES OF AMERICA, THE BIRTH OF A CRATON: Early Proterozoic Assembly and Growth of Laurentia

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> *The fact that two provinces of the Canadian Shield have been together during post-Cambrian time does not necessarily mean that they were formed close together ...*
>
> J. Tuzo Wilson (1962)

INTRODUCTION

Cratons are large areas of continental lithosphere that have remained coherent and relatively rigid since the Precambrian. Laurentia, the North American craton, is one of the oldest and largest. It includes the Precambrian shields of Canada and Greenland, the covered platform and basins of the North American interior, and, in this paper, the reactivated Cordilleran foreland of the southwestern United States.

Laurentia owes its existence to a network of Early Proterozoic[1] orogenic belts (Figure 1). Many of the belts are collison zones preserving only the deformed margins of formerly independent microcontinents composed of Archean crust. Other belts contain accreted Early Proterozoic island arcs and associated intraoceanic deposits. The assembly of Laurentia in the Early Proterozoic may be compared with the assembly of Eurasia in the Phanerozoic.

This review of geological and geochronological data concerns the nature

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[1] Archean = before 2.5 Ga; Early Proterozoic = 2.5–1.6 Ga; Middle Proterozoic = 1.6–0.9 Ga; Late Proterozoic = 0.9–0.6 Ga; Ga = 10^9 yr before present; Myr = 10^6 yr.
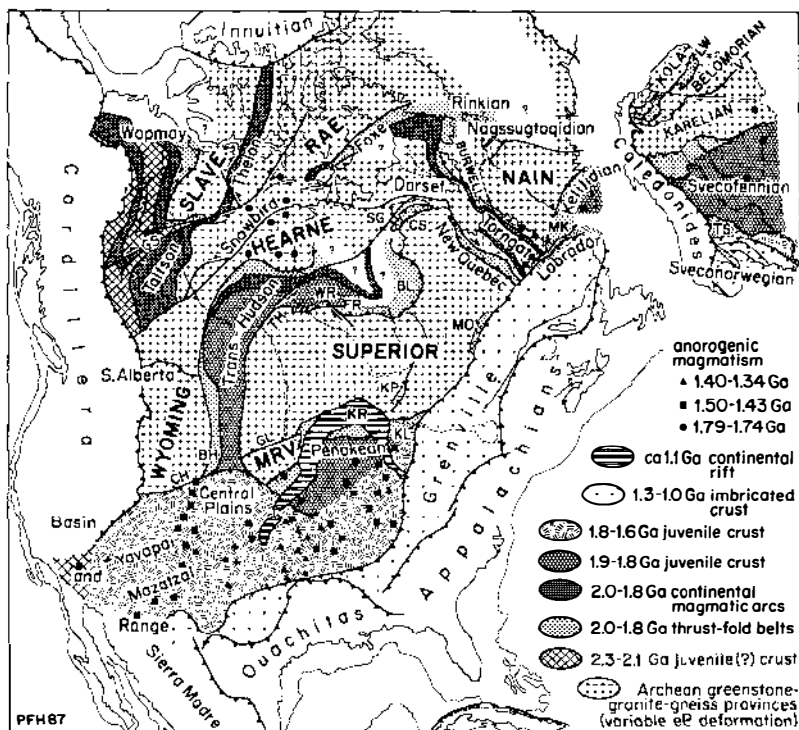
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Figure 1   Precambrian tectonic elements of Laurentia. The Baltic shield is shown in a pre-Iapetus reconstruction, and Greenland is restored prior to rifting from North America. Uppercase names are Archean provinces; lowercase names are Proterozoic and Phanerozoic orogens. Abbreviations: BH, Black Hills inlier; BL, Belcher belt; CH, Cheyenne belt; CS, Cape Smith belt; FR, Fox River belt; GL, Great Lakes tectonic zone; GS, Great Slave Lake shear zone; KL, Killarney magmatic zone; KP, Kapuskasing uplift; KR, Keweenawan rift zone; LW, Lapland–White Sea tectonic zone; MK, Makkovik orogen; MO, Mistassini-Otish basins; MRV, Minnesota River Valley terrane; SG, Sugluk terrane; TH, Thompson belt; TS, Transscandinavian (Småland-Värmland) magmatic zone; VT, Vetrenny tectonic zone; WR, Winisk River fault.

and timing of Early Proterozoic orogenic belts in Laurentia. It complements an earlier review of geophysical data on proposed Proterozoic sutures in Canada (Gibb et al 1983). The underlying theme of both reviews is J. Tuzo Wilson's (1962) admonition, quoted above, regarding the implications of mobilism for Precambrian tectonics. (It is revealing that his first paper espousing mobilism concerns the Precambrian. Field work in the Canadian shield was his initiation to earth science, and he, among the founders of plate tectonics, was the one most interested in its long-term implications.)

New impetus for studying the early history of Laurentia comes, above all, from advances in isotopic geochronology. Precise dating of igneous and metamorphic events provides the most effective means of testing and refining dynamic models for Precambrian orogenic belts based on geological or geophysical data. Greatly improved U-Pb dating methods (e.g. Krogh 1973, 1982, Roddick et al 1987) permit igneous and metamorphic ages to be determined with an analytical precision of better than 0.5% (i.e. ± 5 Myr at 2 Ga). Although the actual uncertainties are likely somewhat greater than the formal analytical errors, zircon U-Pb ages are far more accurate and generally much more resistant to isotopic resetting than Rb-Sr or K-Ar ages for Precambrian rocks. *Unless otherwise stated, all ages quoted in this paper are based on U-Pb analyses.* Most are based on multiple age determinations, and the resulting age limits are rounded to the nearest 10 Myr.

Progress has also been spurred by other developments in geochronology, geophysics, and geology. Sm-Nd isotopic systematics, complementing earlier Pb-Pb and Rb-Sr methods, enables crust newly extracted from the Proterozoic mantle ("juvenile crust") to be distinguished from the products of remelting or mechanical reworking of Archean crust (McCulloch & Wasserburg 1978, Patchett & Arndt 1986). Systematic gravity surveys and interpretation of gravity anomalies permitted the early identification of probable collision zones (Gibb et al 1983, and references therein). Digitized high-resolution aeromagnetic surveys now cover much of the continental interior (Committee for the Magnetic Anomaly Map of North America 1987), providing grounds for extrapolating structural trends and magmatic zones (especially the magnetite-bearing rocks of magmatic arcs) across poorly exposed or poorly mapped parts of the shield and beneath the platformal cover (e.g. Dods et al 1985, Hoffman 1987a).

Some evidence of Early Proterozoic relative motions within Laurentia has been obtained from paleomagnetic studies (e.g. McGlynn & Irving 1981, Irving et al 1984). However, the failure to recognize secondary remagnetizations and other problems (Burke et al 1976) led many paleomagnetists (e.g. McElhinny & McWilliams 1977, Embleton & Schmidt 1979, Piper 1983) to question the early proponents of mobilism (e.g. Gibb & Walcott 1971, Fraser et al 1972, Dewey & Burke 1973). Recently recognized criteria for determining the kinematics of shear zones (Berthé et al 1979, White et al 1980, Lister & Snoke 1984, Passchier & Simpson 1986, Hanmer 1986) assist in determining relative motions between adjacent crustal blocks during collisional orogeny, but the lack of information on relative motions prior to collision remains the most serious obstacle for Early Proterozoic paleocontinental reconstructions. Nevertheless, the recent discoveries of obducted slices of oceanic crust and mantle, including

"sheeted" dike complexes, within Early Proterozoic orogenic belts (e.g. Kontinen 1987, St-Onge et al 1988) corroborates the now prevailing view of geologists, geophysicists, and geochemists that plate tectonics was in operation during the birth of Laurentia.

ARCHEAN PROVINCES AND PROTEROZOIC OROGENS

The main tectonic elements of Laurentia, stripped of its platformal cover, are shown in Figure 1. The problematic "Churchill province" of previous authors is subdivided into the Archean Hearne and Rae provinces to the northwest and the Early Proterozoic Trans-Hudson orogen to the southeast (Hoffman 1988). The Archean provinces (Slave, Rae, Hearne, Wyoming, Superior, Nain) are clustered in the northern two thirds of the craton and underlie most of the Canadian shield. Each province has an Archean basement complex comprising a granite-greenstone terrain or its high-grade equivalent, overlain by erosional remnants of Early Proterozoic sedimentary cover of platformal facies. Variable deformation and metamorphism of the sedimentary cover indicate degrees of Early Proterozoic reactivation of the Archean provinces. In general, reactivation is related in trend and intensity to the orogenic belts that frame the Archean provinces.

Many of the Early Proterozoic orogenic belts appear to represent collision zones between Archean provinces. In the western shield, the Trans-Hudson orogen welds the Hearne and Wyoming provinces to the Superior province, the Southern Alberta orogen welds the Hearne and Wyoming provinces, the Snowbird orogen welds the Rae and Hearne provinces, and the Thelon orogen welds the Slave and Rae provinces. In the eastern shield, the southeastern branch of the Rae province is welded to the Superior and Nain provinces by the New Quebec ("Labrador Trough") and Torngat orogens, respectively; the northeastern branch of the Rae province is welded to the Nain province by the Rinkian-Nagssugtoqidian orogen; and the Foxe fold belt forms a syntaxis at the junction of the north- and southeastern branches of the Rae province. The collisional orogens are typically asymmetric, with one side characterized by a sedimentary prism overthrust toward the Archean foreland, and the other side characterized by a magmatic arc and by great reactivation of the Archean hinterland. The asymmetry presumably reflects the dominant polarity of subduction and consequent thermal regimes during ocean closure. By implication, the Archean provinces were formerly independent microcontinents, although they may have originated cogenetically as rifted fragments of an earlier

continental assembly in the same way that the rifted fragments of Gond-wanaland have been reassembled in Eurasia.

The Early Proterozoic orogenic belts peripheral to the cluster of Archean provinces appear to represent zones of lateral accretion of mainly juvenile Proterozoic crust. Accretion between 2.0 and 1.8 Ga occurred in the Wopmay, Penokean, and Makkovik-Ketilidian orogens, and between 1.8 and 1.6 Ga in the Labradorian, Central Plains, and Yavapai-Mazatzal orogens. In the southwestern United States, the zone of accretion is at least 1200 km wide.

Geochronological data (Table 1) indicate that amalgamation of the Archean provinces occurred between about 2.0 and 1.8 Ga, and that accretion of Laurentia was complete, except for the southeastern part of the Grenville orogen, by about 1.6 Ga. That Laurentia is essentially a product of events occurring between 2.0 and 1.6 Ga is the main conclusion of this review. It gives a mobilistic interpretation to the tectonic episode recognized by Stockwell (1961) as the "Hudsonian orogeny."

The subsequent Proterozoic evolution of Laurentia, including wide-spread Middle Proterozoic "anorogenic" igneous activity (Barager 1977, Emslie 1978, Anderson 1983), the 1.2–1.0 Ga midcontinent (Keweenawan) rift system (Van Schmus & Hinze 1985), and the 1.3–0.9 Ga Grenville orogen (Moore et al 1986, Rivers et al 1988), lies beyond the scope of this review.

Table 1 Comparison of Early Proterozoic orogens in Laurentia

Orogen	Foreland	Hinterland	Obliquity	Magmatism (U-Pb ages)[a]
Thelon	Slave	SW-Rae	right	2.02–1.91 Ga
Wopmay	Slave	arcs	right	1.95–1.84 Ga
Snowbird	Hearne(?)	SW-Rae(?)	(?)	post-1.92/pre-1.85 Ga
S-Alberta	Hearne	Wyoming	(?)	pre-1.85 Ga
Trans-Hudson	Superior	Hearne(+)	left(?)	1.91–1.81 Ga
Cape Smith	Superior	Hearne(?)	nil	1.96–1.84 Ga
New Quebec	Superior	SE-Rae	right	2.14–1.81 Ga
Torngat	Nain	SE-Rae	left	post-2.3/pre-1.65 Ga
Foxe	NE-Rae	SE-Rae	left(?)	1.90–1.82 Ga
Nagssugtoqidian	Nain	NE-Rae	(?)	(?)
Makkovik	Nain	arcs(?)	right	1.86–1.80 Ga
Labrador	Nain(+)	arcs(?)	(?)	1.71–1.63 Ga
Penokean	Superior(+)	arcs	(?)	1.89–1.82 Ga
Yavapai	Wyoming	arcs	(?)	1.79–1.69 Ga
Mazatzal	Wyoming(+)	arcs(+)	(?)	1.71–1.62 Ga

[a] See text for references.

COLLISIONS IN THE NORTH

Thelon Orogen: Rae/Slave Collision

The Slave province is a late Archean (2.7–2.5 Ga) granite-greenstone terrain that served as a foreland for the Thelon orogen (2.02–1.91 Ga) to the east and the Wopmay orogen (1.95–1.84 Ga) to the west (Figure 2). The Thelon orogen resulted from a dextral-oblique collision between the Slave and Rae provinces (Figure 3), followed by indentation of the Rae hinterland by the Slave foreland province (Gibb 1978a, Hoffman 1987a, R. Tirrul & J. P. Grotzinger, in preparation). Indentation was accommodated by dextral-oblique crustal shortening across the Queen Maud uplift, a reactivated Archean(?) granulite-grade domain in the hinterland, and by 600–700 km of right-slip on the Great Slave Lake shear zone, an intracontinental transform structure exposed as a zone of continuous mylonite 25 km wide (Hanmer 1987a, Hoffman 1987a). Crustal wedges were extruded laterally from the indentation zone: The one bounded by the Great Slave Lake and Rutledge-Allen shear zones escaped southwestward; another bounded by the Bathurst and Ellice fault zones escaped north-northwestward. The deformed alluvial-lacustrine Nonacho basin, although undated, is probably related to sinistral wrench faulting associated with tectonic escape from the indentation zone (Aspler & Donaldson 1985).

A cryptic suture zone (Gibb & Thomas 1977), transposed by right-slip and east-dipping dip-slip shear zones (Thompson et al 1986, Henderson et al 1987), is inferred between Archean rocks of the Slave province and a zone to the east characterized by 2.02–1.91 Ga granitic to dioritic plutons (van Breemen et al 1987a,b, Bostock et al 1987). This zone of magnetite-series and subordinate ilmenite-series plutons is expressed as a distinctive belt of magnetic anomalies more than 80 km wide that can be traced for 2550 km from central Alberta to Prince of Wales Island in the central Arctic archipelago (Figure 1). Exposed for 1000 km, the Taltson-Thelon plutonic zone is interpreted as a composite precollisional magmatic arc and postcollisional anatectic batholith.

Foreland thrust-fold belts and autochthonous foreland basins related to the Thelon orogen are preserved in two structural depressions (Figure 2). The Goulburn Supergroup in the northeastern Slave province comprises a basal, eastward-thickening wedge (0–500 m) of shallow-shelf quartzite and carbonate (Kimerot platform) that is overlain by the Bear Creek foredeep (Grotzinger & McCormick 1987). The foredeep was initiated by sudden drowning of the Kimerot platform, followed by deposition of a flysch-molasse wedge that thins from 5.5 km in the east to 1.5 km over a syndepositional flexural arch in the foreland. Turbidity currents flowed
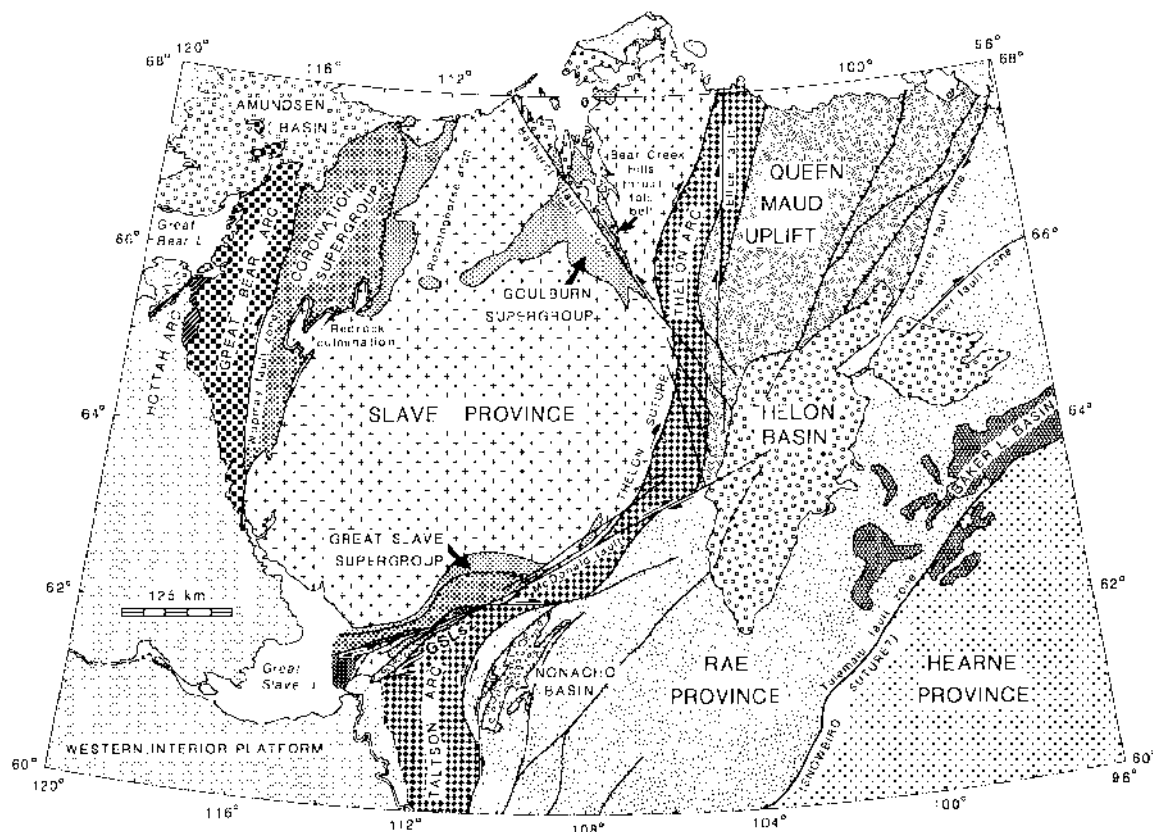
Figure 2 The northwestern Canadian shield, showing the tectonic elements described in the text. The Slave province is bounded by the Thelon orogen on the east and the Wopmay orogen on the west.
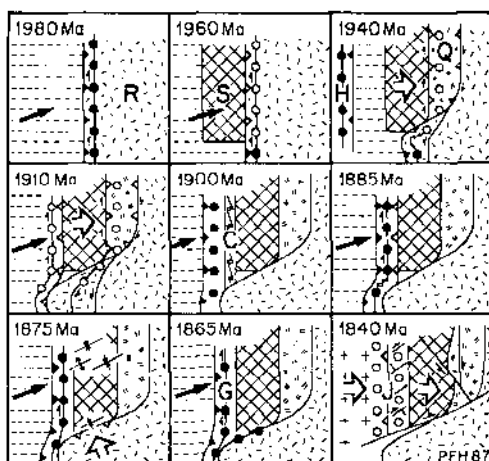
Figure 3   Schematic summary of the plate-tectonic evolution of the Thelon and Wopmay orogens. Solid teeth indicate subduction zones; open teeth indicate intracontinental thrust zones. Solid circles are magmatic arcs above subduction zones; open circles are post-subduction magmatic zones. Horizontal dashed areas are oceanic lithosphere. Abbreviations: C, Coronation Supergroup back-arc basin; G, Great Bear magmatic zone; H, Hottah magmatic zone; J, Johnny Hoe suture(?); Q, Queen Maud uplift; R, Rae province; S, Slave province.

axially in the foredeep, but the overlying fluvial sediments were transported to the west or northwest (Campbell & Cecile 1981, Grotzinger & McCormick 1987). The remnant Bear Creek Hills thrust-fold belt, which is completely exposed in cross section as a result of later folding associated with the Bathurst left-slip fault system, is developed above a WNW-directed ductile sole thrust located near the base of the Kimerot platform (Tirrul 1985, R. Tirrul & J. P. Grotzinger, in preparation). Thin-skinned deformation produced about 60% shortening in the cover transverse to the Thelon orogen. U-Pb zircon dating of a volcanic ash layer near the base of the Bear Creek foredeep establishes a temporal link between foredeep subsidence and the Taltson-Thelon magmatic zone, and provides a preliminary age of about 1.96 Ga for the Rae/Slave collision (S. A. Bowring, personal communication, 1986).

A complex basin in the east arm of Great Slave Lake is located on the southeast margin of the Slave province and is bounded by the Great Slave Lake shear zone (Figure 2). The basin contains three Early Proterozoic sequences, the older two of which are deformed by northwest-directed thrust-nappes and all three by northeast-trending right-slip faults and related folds (Hoffman 1987c). The oldest sequence is the allochthonous Wilson Island Group, a rift-like succession at least 8 km thick in which

bimodal subaerial volcanics (1.93 Ga) are overlain by fluvial to marine siliciclastic metasediments intruded by a 1.89-Ga granite (Bowring et al 1984, Johnson 1986). The Great Slave Lake shear zone overlaps the Wilson Island Group in age and contains tentatively correlative metasedimentary protoliths. The 3–7-km-thick Great Slave Supergroup structurally underlies but is apparently younger than the Wilson Island Group. It unconformably overlies the Great Slave Lake shear zone, and its basal conglomerate contains clasts lithologically similar to the Wilson Island Group. Paleocurrents show that the lower clastics (Sosan Group) of the Great Slave Supergroup were shed from the northeast, probably from the Thelon magmatic zone and Queen Maud uplift, and that the upper clastics were derived from the Wopmay orogen to the southwest (Hoffman 1969). The Wilson Island Group and Great Slave Supergroup were telescoped by northwest-directed thrust-nappes and then intruded by calc-alkaline laccoliths (Badham 1981) that are coeval (1.87–1.86 Ga) with the Great Bear magmatic zone of the Wopmay orogen (Bowring et al 1984b). The laccoliths are overlain erosionally by alluvial-fan deposits related to right-slip on the McDonald fault system. The basin was earlier interpreted as an aulacogen related to the Wopmay orogen (Hoffman 1973), but it is now believed to have originated as an Andaman-type basin (Figure 3) resulting from indentation-extrusion tectonics (Tapponnier et al 1982) related to the Rae/Slave collision (Hoffman 1987a).

Wopmay Orogen: Active Western Margin of the Slave Province

The Wopmay orogen evolved on the active western margin of the Slave province (Figure 2). Dividing the orogen is the meridional Wopmay fault zone, a 10-km-wide belt characterized by mylonite having subhorizontal stretching lineations and by younger brittle faults having many kilometers of west-side-down throw (Hildebrand et al 1987a, J. E. King, in preparation). To the west are two continental calc-alkaline volcanic-plutonic belts (Hildebrand 1981, 1984, Hildebrand et al 1987b), in part spatially superimposed but temporally separated by a short-lived episode of rifting evidenced by subsidence and tholeiitic volcanism (Reichenbach 1986, 1987). Based on U-Pb zircon ages (Bowring 1984), the older Hottah magmatic arc was active from at least 1.95 to 1.91 Ga and was deformed and metamorphosed prior to the rifting event at about 1.90 Ga. The younger, more easterly Great Bear magmatic arc was active from about 1.88 to 1.86 Ga, after which dextral shear parallel to the arc produced broad northwest-trending en echelon folds. A 1.86–1.84 Ga suite of anatectic syenogranites postdates the en echelon folds but predates a system of conjugate trans-

current faults that accommodates east-west shortening and north-south extension affecting the entire orogen (Hoffman 1984, Tirrul 1987a).

East of the Wopmay fault zone, rifting marginal to the Slave province at about 1.90 Ga initiated deposition of the Coronation Supergroup, a west-facing passive-margin shelf-rise prism and succeeding foredeep flysch and molasse (Easton 1981, Hoffman & Bowring 1984, Grotzinger 1986, Hoffman 1987b). Soon thereafter, between about 1.89 and 1.88 Ga, the westerly off-shelf facies of the prism was intruded by a suite of gabbroic-dioritic-tonalitic-granitic plutons (Hoffman 1984, Lalonde 1986) and then translated eastward above a ductile sole thrust, producing a thin-skinned foreland thrust-fold belt in shelf and foredeep sediments to the east (Tirrul 1983, King 1986, Hoffman 1987b, Hoffman et al 1987, Tirrul 1987b). Preservation of inverted metamorphic isograds in upward-facing autochthonous strata indicates that thrusting occurred while the plutons and their metamorphic envelope were at least 250°C hotter than the underlying autochthon (St-Onge & King 1987a,b). Microprobe analyses of zoned poikiloblastic garnets document clockwise pressure-temperature paths for various structural levels in the allochthon and autochthon, with the latter having been uplifted from depths exceeding 30 km (St-Onge 1987). Uplift was accompanied by large-scale basement-cover folding coaxial with meridional synmetamorphic stretching ascribed to a dextral component of regional transpression (Hoffman et al 1987, King et al 1987). Deformation of the Coronation Supergroup has been ascribed to oblique collision of an Atlantic-type margin with an exotic island arc (Hoffman 1980, Hildebrand et al 1983, Hoffman & Bowring 1984). Alternatively, the correlation of 1.90-Ga rifting events east and west of the Wopmay fault zone is compatible with short-lived (1.90–1.88 Ga) evolution of the Coronation Supergroup in a back-arc setting (Hildebrand & Roots 1985, Reichenbach 1986, 1987). Rifting of the back-arc basin may have been induced by an arc-continent collision, accounting for the pre-1.90-Ga deformation of the Hottah arc, by a mechanism similar to that postulated for opening of the active Okinawa Trough (Letouzy & Kimura 1986).

Interpretation of magnetic and gravity anomalies west of the exposed part of the Wopmay orogen, constrained by sparse well data, provides a possible explanation for the post-1.84-Ga conjugate transcurrent faulting in the orogen and the correlative McDonald-Bathurst fault system of the eastern Slave province and Thelon orogen (Hoffman 1980). A linear gravity high bounded by an east-facing "escarpment" in the Bouguer anomaly field 180 km west of the Wopmay fault zone occurs between the Hottah/Great Bear magmatic zones to the east and a parallel 1200-km-long magnetic high to the west, from which a drill core of granodiorite has been dated at about 1.86 Ga (Hildebrand et al 1987b, Hoffman 1987a). The

positive magnetic anomaly is tentatively interpreted as a buried magmatic arc developed above a west-dipping subduction zone at the leading edge of a terrane that collided with the Wopmay orogen along a suture de-lineated by the Bouguer "escarpment." Accordingly, the conjugate transcurrent faulting could be a manifestation of east-west shortening in the foreland of the collision zone.

No exposures of Archean crust are known west of the Wopmay fault zone, and preliminary investigation of Pb and Nd isotopes in the Hottah and Great Bear arcs, as well as U-Pb dating of detrital and xenocrystic zircons, suggests that none was involved in their generation (Bowring & Podosek 1987, S. A. Bowring, personal communication, 1987).

Snowbird Tectonic Zone: Hearne/Rae Collision?

Between the Thelon orogen to the northwest and the Trans-Hudson orogen to the southeast lies a broad region of Archean crust (Figure 4) containing
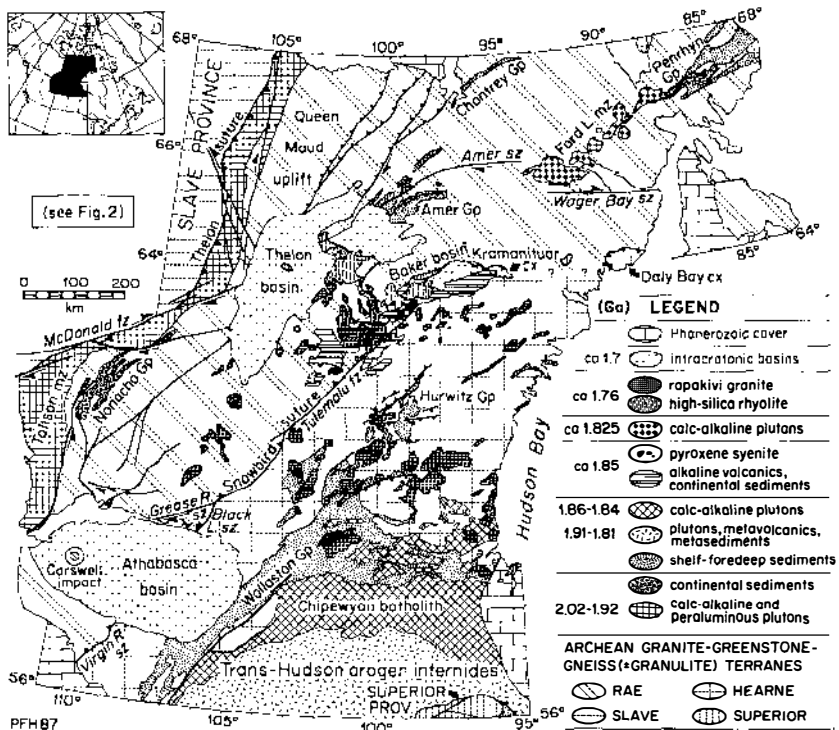

Figure 4 Geology of parts of the Rae and Hearne provinces framed by the Thelon orogen on the west, the Trans-Hudson orogen on the south, the Hudson Bay basin on the east, and the Arctic platform on the north.

scattered outliers of folded sedimentary cover, posttectonic igneous suites, and overlying cratonic basins all of Early Proterozoic age (Davidson 1972, Lewry et al 1985). The region is transected by an anastomosing, northeast-trending crustal break most evident on the horizontal gravity gradient map (Sharpton et al 1987). Segments of the break are recognized geologically as the Virgin River (Lewry & Sibbald 1977), Black Lake (Gilboy 1980, Hanmer 1987b), and Tulemalu (Tella & Eade 1986) fault and/or shear zones. Extending for almost 3000 km from the Rocky Mountains to Hudson Strait (Figure 1), the entire break is here referred to as the Snow-bird tectonic zone. The name is derived from Snowbird Lake in the southeast corner of the District of Mackenzie, where Taylor (1963) de-scribed a zone of mylonite many kilometers wide containing bodies of banded gabbro-anorthosite-pyroxenite, an association now known to occur at many places along the length of zone. The name "Athabasca axis," proposed for the same zone by Darnely (1981), is not retained because its trend is genetically unrelated and almost perpendicular to the depositional and structural axis of the Athabasca basin (Ramaekers 1981).

The initial suggestion that the Snowbird zone might be a suture (Walcott & Boyd 1971, Gibb & Halliday 1974) was based on the associated gravity anomalies. Other evidence comes from magnetic anomalies, which indicate that the Snowbird zone truncates the Taltson magmatic zone of the Thelon orogen at a high angle in the subsurface of central Alberta (Committee for the Magnetic Anomaly Map of North America 1987). Accordingly, the age of suture is less than about 1.92 Ga. It must be older than the 1.85-Ga intrusions and related alkaline volcanics and sediments of the Baker Lake basin (LeCheminant et al 1987a,b), which constitute an overlap assemblage on the central segment of the zone. The Virgin River–Black Lake segment of the Snowbird zone has been interpreted as an intra-continental reactivation structure related to the Trans-Hudson orogen (Lewry & Sibbald 1980), but elsewhere the zone diverges from the trend of the orogen, suggesting an independent origin.

The interpretation of the Snowbird zone as a suture remains hypo-thetical. No magmatic arc related to the Snowbird zone has been clearly identified, although 2.4–2.0 Ga granites occur 150 km northwest of the zone at Lake Athabasca (Van Schmus et al 1986) and ~1.88-Ga granite occurs about 100 km northwest of the zone near Baker Lake (LeCheminant et al 1987b). There are as yet few kinematic data for the Snowbird zone. Hanmer (1987b) presents preliminary evidence that the Tantato wedge of mafic to felsic granulite-facies mylonite north of the Athabasca basin was driven southwestward, compatible with tectonic "escape" from a zone of convergence between the Hearne and Rae provinces to the northeast. There, granulites of the Rae province are juxtaposed against relatively

low-grade Archean rocks of the Hearne province. To the northeast, the Tulemalu fault zone contains fragments of high-pressure ($\sim$1.1 GPa) garnet-clinopyroxene granulite (Tella & Eade 1986). Farther to the northeast, between the Baker Lake and Hudson Bay basins, layered mafic-felsic granulite-anorthosite complexes appear to be thrust north- or northwestward onto lower-grade rocks of the Rae province (Schau et al 1982, Gordon 1987).

The Hearne and Rae provinces have much in common. Both contain some very old rocks; granites between 3.33 and 2.95 Ga occur widely in the Rae province on the Melville Peninsula (Wanless 1979), near Baker Lake (A. N. LeCheminant, personal communication, 1987), and at Lake Athabasca (Van Schmus et al 1986), and a 3.48-Ga gneiss occurs in the Hearne province 200 km northeast of the Athabasca basin (Wanless 1979, W. D. Loveridge & K. E. Eade, in preparation). Both provinces contain late Archean greenstone belts and associated granites. Those in the Hearne province (Davidson 1970a,b) contain submarine mafic-intermediate-felsic volcanics and associated graywackes and are about 2.7 Ga (Mortensen & Thorpe 1987); those in the Rae province (Schau 1982, Taylor 1985) include ultramafic-mafic-felsic volcanics about 2.8 Ga, associated with cross-bedded quartzarenite, and an epizonal granite-rhyolite ash-flow tuff association about 2.6 Ga (LeCheminant et al 1984, 1987b).

Both provinces preserve erosional remnants of Early Proterozoic platformal sedimentary cover that predate 1.85 Ga. Both the Hurwitz Group of the Hearne province (Bell 1970, Eade & Chandler 1975) and the Amer Group of the Rae province (Tippett & Heywood 1978, Frisch & Patterson 1983, Frisch et al 1985, Patterson 1986) have a prominent orthoquartzite in their lower part; a middle part containing mixed carbonate, pelite, and localized mafic volcanics; and an upper part dominated by fine-grained felspathic sandstone. Deformation of the Hurwitz Group is of two trends, producing a basin-and-dome interference pattern at their intersection. East-striking folds and south-dipping thrusts parallel the Seal River fold belt of the Trans-Hudson orogen in the south; northeast-striking folds and north- to northwest-dipping thrusts parallel the Snowbird zone in the north and west (Eade 1974, 1987). The folds and thrusts involve the Archean basement, and deformation and metamorphic grade increase toward both the Trans-Hudson orogen (Pearson & Lewry 1974, Schledewitz 1978) and the Snowbird zone (Davidson 1970a, Eade 1986, Tella et al 1986). In the Rae province, folds and thrusts affecting the Amer Group and its basement trend northeasterly, but there is no obvious increase in deformation or metamorphism toward the Snowbird zone. Patterson (1986) interprets the thrusts in the Amer Group to have north to northwest vergence, but this is questionable because several of the

thrusts ramp down into the basement in that direction, and stratigraphic footwall cutoffs (Tippett & Heywood 1978, Patterson 1986) are more compatible with northeast-directed thrusting.

Extensive posttectonic Early Proterozoic igneous suites straddle the Snowbird zone (Figure 4) and distinguish the Hearne and Rae provinces from the other Archean provinces. A suite of 1.85–1.84 Ga lamprophyre dikes and pyroxene-syenite stocks, associated with alkaline volcanics and continental sediments of the Baker Lake basin (Blake 1980, LeCheminant et al 1981, 1987a,b), overlaps the central part of the Snowbird zone. A younger suite of 1.78–1.75 Ga high-silica rapakivi granites and rhyolite ash-flow tuffs, associated with continental sediments, is widely distributed west of Hudson Bay. The tectonic significance of the magmatism is con-jectural, but Hoffman (1980) suggests that the alkaline suite was localized by east-directed "wedging" of the Slave province on the McDonald-Bathurst fault system (Figure 3).

Southern Alberta "Rift": Wyoming/Hearne Collision?

The buried Precambrian basement of southwestern Saskatchewan, southern Alberta, and Montana, lying between the exposed parts of the Hearne and Wyoming provinces, is composed mainly of Archean or reworked Archean crust (Peterman 1981, Frost & Burwash 1986, Peterman & Futa 1987). The Hearne and Wyoming provinces may therefore be mutually continuous. Alternatively, an easterly trending crustal dis-continuity that crosses southern Alberta between latitudes 50° and 50°30′N (Thomas et al 1987, Sharpton et al 1987) is a possible suture between the two provinces (J. W. Peirce, personal communication, 1986).

Kanasewich et al (1969) interpreted the southern Alberta structure (Fig-ure 1) as a buried Precambrian rift related to the Middle Proterozoic Belt-Purcell basin on the basis of Bouguer and magnetic anomalies and seismic reflection profiling. Their seismic profile shows that the lower crust dips southward beneath the "rift" but is uplifted sharply at its southern margin. A compatible asymmetry is indicated by the gravity field, which shows a negative anomaly coincident with the "rift," a sharp positive anomaly cresting close to its southern margin, and a broad positive anomaly cresting 30–70 km north of the rift. The seismic evidence for a thickened crust and the nature of the gravity field are not typical of old, thermally mature rifts. The observations are more compatible with a foredeep (foreland basin) and associated flexural bulge to the north, result-ing from north-directed thrusting at the south margin of the "rift" (cf Karner & Watts 1983). Contrasting trends of regional magnetic anomalies, having a strong northwest "grain" south of the "rift" but an irregular north

to northeast "grain" north of the "rift" (Committee for the Magnetic Anomaly Map of North America 1987), suggest the juxtaposition of different crustal blocks (Green et al 1985b). Accordingly, the Wyoming province may have collided with the Hearne province along a south-dipping suture zone, producing a foredeep consistent with the polarity implied by the seismic and gravity data.

The age of the proposed collision is conjectural, but its presumed eastward extension appears to be truncated by and therefore older than the subsurface part of the Trans-Hudson orogen (Green et al 1985b, Thomas et al 1987). This implies a pre-1.9 Ga collision age but does not preclude reactivation of the suture as a rift arm during deposition of the Belt-Purcell basin (Kanasewich et al 1969) or as a transform structure during Late Proterozoic rifting (Lis & Price 1976) and early Paleozoic subsidence (Bond & Kominz 1984) of the Cordilleran passive margin.

Trans-Hudson Orogen: Hearne-Wyoming/Superior Collision

As exposed in the provinces of Saskatchewan and Manitoba, the 500-km-wide Trans-Hudson orogen forms a dogleg, convex to the northwest, bounded by the Hearne and Superior provinces (Figures 1, 5). Southward, the orogen has been outlined in the subsurface between the Wyoming and Superior provinces as far as South Dakota (Green et al 1985a, Klasner & King 1986, Thomas et al 1987), where it appears to be truncated by the Central Plains orogen (Sims & Peterman 1986). To the northeast, the main part of the orogen passes beneath the Paleozoic Hudson Bay basin (Gibb & Walcott 1971, Coles & Haines 1982, Gibb 1983, Sharpton et al 1987), but its southeastern margin is discontinuously exposed south of Hudson Bay and along the coast and offshore islands of eastern Hudson Bay (Baragar & Scoates 1981, Mukhopadhyay & Gibb 1981). The northwest-dipping Sugluk suture zone in northernmost Quebec (see Figure 8) may represent the pinched extension of the orogen, its internal zone having been obducted southward and preserved as an infolded klippe (the Cape Smith thrust-fold belt) within the Superior province (Hoffman 1985, Doig 1987, St-Onge & Lucas 1988).

Essentially, the orogen comprises an internal zone of intraoceanic rocks flanked by ensialic external belts (Stauffer 1984). The Wathaman-Chipewyan batholith (Figure 5) is interpreted to be an ensialic magmatic arc (Ray & Wanless 1980, Lewry et al 1981, Fumerton et al 1984, Meyer 1987), and its position bordering the Hearne province implies a gross tectonic polarity for the orogen, in which the Superior province is the foreland and the Hearne province is the hinterland.
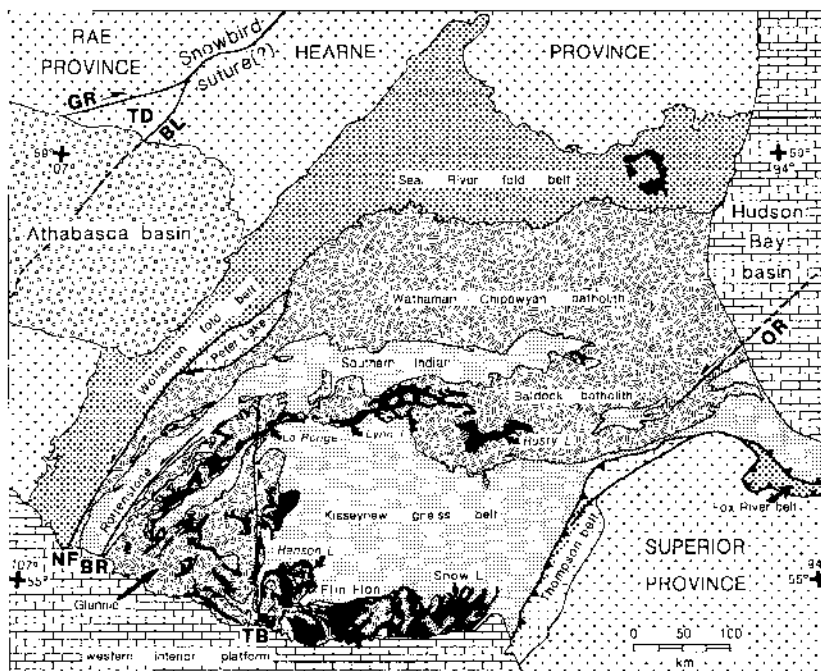
Figure 5 Tectonic elements of the Trans-Hudson orogen as exposed in northern Saskatchewan and Manitoba. Volcanic belts are shown in black. Major shear zones: BL, Black Lake; BR, Birch Rapids; GR, Grease River; NF, Needle Falls; OR, Owl River; TB, Tabbernor; TD, Tantato granulite-facies mylonite domain.

EXTERNAL BELTS   The ensialic belts of the Superior margin are exposed in four segments named, from southwest to northeast, the Thompson, Fox River, Belcher, and Cape Smith belts. The most complete stratigraphic sequence occurs in the Belcher belt, an orogenic salient in eastern Hudson Bay. On the coast, an arcuate west-dipping autochthonous sequence exposes, in oblique cross section, a basal east-trending rift-valley prism, 70 km wide, overlain by postrift shelf strata that onlap the basement north and south of the rift (Chandler 1984, 1987). The islands to the west belong to an arcuate meridional belt of doubly plunging folds that have been transported eastward, relative to the autochthon, on a sole thrust that carries 7–9 km of strata (Jackson 1960). The fold belt exposes at least 3.4 km of shelf carbonates and clastics containing a unit of plateau basalt up to 0.9 km thick (Ricketts & Donaldson 1981). The shelf strata are overlain paraconformably by foredeep deposits (Hoffman 1987b), comprising a transgressive sequence of quartzarenite, ironstone, tholeiitic-komatiitic volcanics, and euxinic shale, followed by a regressive sequence of gray-

wacke turbidites and fluvial arkose (Ricketts & Donaldson 1981, Ricketts
et al 1982, Baragar 1984). There are no U-Pb ages for the Belcher belt,
but Pb isochrons for flows in the shelf and foredeep sequences are
1.96 ± 0.08 and 1.81 ± 0.03 Ga, respectively (Todt et al 1984), and Sm-Nd
isotopic data indicate that both sequences contain flows contaminated by
Archean crust (Chauvel et al 1987).

The Cape Smith belt (see Figure 7) is an erosional remnant of a thin-
skinned, south-vergent thrust-fold belt preserved as a doubly plunging
synclinorium resulting from two episodes of basement-involved refolding
(St-Onge et al 1986, St-Onge & Lucas 1988). The volcanic and sedimentary
rocks record the evolution of the rifted north margin of the Superior
province from continental, through transitional oceanic, to true oceanic crust
(St-Onge et al 1988). The continental rift sequence (Povungnituk Group)
is represented by basal imbricates of semipelite, ironstone, and proximal to
distal arkosic submarine-fan deposits, which pass stratigraphically upward
into light-rare-earth-element-enriched basalt, minor rhyolite (1.96 Ga),
volcaniclastic deposits, and fault-scarp breccias (Hynes & Francis 1982,
St-Onge & Lucas 1988). The transitional crust (Chukotat Group) occurs
in structurally higher thrust sheets and is dominated by pillowed basalt
distinguished by a komatiitic to midoceanic-ridge-basalt-(MORB)-like
tholeiitic chemistry (Francis et al 1983). Both the continental and tran-
sitional oceanic crust sequences are intruded by layered peridotite-
gabro sills (St-Onge & Lucas 1988). Structurally overlying the Chukotat
Group is a metamorphosed ophiolite suite exposed in inverse stratigraphic
order (St-Onge et al 1988). From structural bottom to top, the ophiolite
consists of thrust sheets composed of (*a*) laminated graphitic pelite and
semipelite; (*b*) pillowed basalt cut by mafic dikes, "sheeted" dikes, and
gabbro cut by dikes; and (*c*) layered mafic-ultramafic cumulates. Early
thrusting occurred in a "piggy-back" sequence and was followed by "out-
of-sequence" thrusting, with both forms of thrusting rooted on a sole
thrust at the basement-cover contact in the preserved belt (St-Onge &
Lucas 1988). Thermal relaxation following the early thrusting resulted in
syndeformational metamorphism and is probably responsible for minor
tonalite to granite plutons (1.84 Ga) that intrude the ophiolitic thrust
sheets (St-Onge & Lucas 1988).

The east-trending Fox River belt (Figure 5) and Sutton inlier south of
Hudson Bay are poorly exposed but include steeply north-dipping thrust
stacks of relatively thin basal shelf sediments and overlying foredeep flysch,
the latter associated with differentiated tholeiitic-komatiitic sills and flows
(Bostock 1971, Scoates 1981, Baragar & Scoates 1981). The 2-km-thick
Fox River sill and the extensive NNE-trending Molson dike swarm of the
northwestern Superior province are coeval at 1883 ± 2 Ma (Scoates &

Macek 1978, Heaman et al 1986), but the tectonic significance of the magmatism is unclear. The north margin of the Fox River belt is a probable north-dipping thrust separating subgreenschist-facies rocks to the south from amphibolite-facies metasedimentary paragneisses similar to those of the Kisseynew belt (Figure 5).

In the northeast-trending Thompson belt, equivalents of the Fox River belt strata are tightly infolded with Archean basement (Weber & Scoates 1978, Peredery et al 1982). Foliations in the basement and cover dip very steeply to the southeast, and stretching lineations in mylonite zones are generally subvertical throughout the Thompson belt (Fueten et al 1986). Kinematic indicators in the western part of the belt show that the Superior province has moved up relative to the Trans-Hudson orogen (W. Bleeker, personal communication, 1987). The bulk of the deformation postdates the 1.88-Ga Molson dikes and predates 1.79–1.77 Ga pegmatites; post-metamorphic cooling is recorded by a concordant titanite age of 1.72 Ga (Machado et al 1987).

The northwest marginal zone of the orogen includes the Wollaston and Seal River fold belts (Figure 5) of Saskatchewan and Manitoba. The cover sequence on the margin of the Hearne province is thought to include an early synrift assemblage, subsequent shelf quartzite, and late arkosic synorogenic deposits (Lewry & Sibbald 1980, Stauffer 1984). However, the stratigraphic order and its interpretation are uncertain because of poor outcrop, complex structure involving upright basement-cover folds superimposed on polyphase recumbent folds (Lewry & Sibbald 1980), and upper-amphibolite- to granulite-facies metamorphism (Lewry et al 1978, Schledewitz 1978). There are numerous granitic bodies, particularly in the Seal River belt (Schledewitz 1986), but it is difficult in the absence of U-Pb ages to ascertain the relative proportions of granite related to the Wathaman-Chipewyan batholith (1.86–1.85 Ga; Meyer 1987) to the south, the 1.75-Ga rapakivi suite of the Hearne province, and Archean basement inliers.

The Black Hills uplift in South Dakota (Redden & Norton 1975) exposes a fold belt marginal to the Wyoming province near the junction of the Trans-Hudson and Central Plains orogens (Sims & Peterman 1986). According to Redden et al (1987), Archean basement is overlain by synrift clastic sediments that were intruded by a 2.17-Ga mafic sill, then deformed and eroded, prior to deposition of younger rift- and shelf-facies clastic sediments and mafic volcanics. The shelf quartzites, which contain a 1.97-Ga tuff bed, are laterally equivalent to and overlain by thick turbidites capped by 1.88-Ga alkalic tuffs. The entire sequence was recumbently folded about easterly axes, refolded about NNW-trending axes, and intruded by the 1.72-Ga Harney Peak granite.

The Black Hills uplift lies just west of the North American Central Plains electrical conductivity anomaly (Jones & Savage 1986, and references therein), which trends northward from the Cheyenne belt of southeastern Wyoming (Figure 1) to the Canadian shield, where it follows the curvature of the Rottenstone–Southern Indian gneiss belt (Figure 5) and continues eastward to Hudson Bay. The conductivity anomaly is generally regarded as being related to a suture zone within the Trans-Hudson orogen, and it may coincide with the deep subsurface limit of Archean crust contiguous with the Wyoming and Hearne provinces. Thomas et al (1987) question this interpretation because of differences in trend between the conductivity anomaly and horizontal gravity gradients. However, the latter emphasize near-surface density variations and may reflect the structure of Early Proterozoic allochthons that need not parallel the deep-seated limit of autochthonous Archean crust.

INTERNAL BELTS The internal zone of the Trans-Hudson orogen is a complex of plutonic, metavolcanic, and metasedimentary rocks, varying proportions of which have been used to distinguish different belts (Figure 5). The Baldock and Wathaman-Chipewyan belts are compound calc-alkaline batholiths of granodiorite and granite, with subordinate tonalite, diorite, and gabbro. The La Ronge, Lynn Lake, Rusty Lake, Glennie Lake, Hanson Lake, and Flin Flon belts consist mainly of plutonic, metavolcanic, and subordinate metasedimentary rocks of greenschist to lower amphibolite grade. The Kisseynew and Rottenstone–Southern Indian belts are composed of upper-amphibolite-grade metasedimentary paragneiss and felsic-to-mafic orthogneiss.

The internal zone was long thought to be a typical "granite-greenstone-gneiss" complex of Archean age (Harrison 1951); its true age of 1.9–1.8 Ga (discounting 1.8–1.7 Ga Rb-Sr isochrons reset during metamorphism) was first indicated by Pb isotopic ratios of syngenetic massive-sulfide mineralization (Sangster 1972, 1978). Recent U-Pb zircon dating has confirmed the Early Proterozoic age (Van Schmus & Schledewitz 1986, Baldwin et al 1987, Gordon et al 1987a,b, Syme et al 1987, Van Schmus et al 1987a), showing that the earliest known volcanism occurred in the Lynn Lake belt at 1.91 Ga, followed by volcanism at 1.89–1.88 Ga in the La Ronge, Rusty Lake, Glennie Lake, Hanson Lake, and Flin Flon belts. Plutonism occurred between 1.88 and 1.83 Ga in all the volcanic-plutonic belts and in the Rottenstone–Southern Indian gneiss belt, contemporaneous with intrusion of the Wathaman-Chipewyan batholith between 1.86 and 1.84 Ga. Peak metamorphism and anatectic plutonism in the Kisseynew gneiss belt occurred at 1.82–1.81 Ga. The only dated Archean rocks occur in the Peter Lake orthogneiss domain (Figure 5) and

in small structural culminations within the Hanson Lake (Van Schmus et al 1987a) and Glennie Lake (J. Chiarenzelli, personal communication, 1987) belts. Sm-Nd isotopic data (Chauvel et al 1987, Hegner & Hulbert 1987, Thom et al 1987) indicate that the magmas of the Flin Flon volcanic and plutonic rocks were derived from sources having no perceptible crustal residence time, but that contributions from older crust increase toward the northwestern Archean hinterland. These data indicate that magma generation involved little or no Archean crust in the southeastern part of the internal zone, but they do not rule out subsequent underthrusting by Archean crust (cf Green et al 1985a). The isotopic data strongly support the view that most of the rocks exposed in the internal zone represent juvenile 1.9–1.8 Ga crust of intraoceanic origin (Bailes 1971, Stauffer 1974, 1984, Green et al 1985a) bounded by an Andean-type continental margin to the northwest (Ray & Wanless 1980, Lewry et al 1981, Fumerton et al 1984, Meyer 1987).

The volcanic-plutonic belts contain structurally dismembered piles of mainly submarine, mafic-through-felsic, metavolcanic rocks having the major- and trace-element characteristics of Cenozoic island arcs (Stauffer et al 1975, Gilbert et al 1980, Syme 1985, 1987, Bailes & Syme 1987, Thom et al 1987, Watters et al 1987). They include tholeiitic basalts, high-alumina/low-titania basalts and basaltic andesites, and calc-alkaline andesite-dacite-rhyolite centered complexes. Mafic-to-felsic heterolithic breccias and redeposited volcaniclastic debris are common. The volcanic piles are intruded by synvolcanic gabbro, diorite, and tonalite sheets and stocks, above which extensive mineralized alteration haloes are developed (Baldwin 1980, Barham & Froese 1986, Bailes et al 1987). In the Bear Lake block east of Flin Flon (Bailes & Syme 1987), a thick pile of arc-like mafic flows is overlain by a ferrobasalt-ferrogabbro-rhyolite association succeeded by andesitic turbidites, a sequence possibly related to intraarc rifting. Elsewhere, volcanic piles are intercalated with and grade laterally into volcaniclastic turbidites (Bailes 1980a). The intercalated volcanic and sedimentary rocks (Amisk and Wasekwan groups of the Flin Flon and Lynn Lake belts, respectively) and related plutons are overlain unconformably by alluvial fanglomerate and sandstone (Missi and Sickle groups of the same belts) (Milligan 1960, Mukherjee 1974). Alluvial sedimentation accompanied D1 (discussed in what follows) folding (Stauffer & Mukherjee 1971, Bailes et al 1987), and the 1.83-Ga age of a rhyolite tuff in the Missi Group (Gordon et al 1987a) indicates that this occurred at least 50 Myr after Amisk volcanism.

The transitions into the Kisseynew gneiss belt from the bounding Flin Flon and Lynn Lake volcanic-plutonic belts represent changes in both primary facies, from volcanics to turbidites (Zwanzig 1976, Bailes 1980a,b,

Ashton & Wheatley 1986), and marked increases in metamorphic grade from chlorite-biotite-muscovite schist to K feldspar-cordierite-garnet paragneiss (Bailes & McRitchie 1978, Bailes 1980b, Jackson & Gordon 1985, 1986). It has been suggested that turbidite sedimentation occurred in a zone of high heat flow because of the spatial correlation between turbidite facies and high metamorphic grade, and because of the high-temperature (750°C)/moderate pressure (0.5 GPa) character of the metamorphism. This would favor an immature back-arc basin (Lewry 1981) over a fore-arc setting (Green et al 1985b). However, the current view that turbidite sedimentation was coeval with the 1.88-Ga Amisk volcanism (e.g. Bailes 1980a), rather than with the 1.83-Ga Missi sedimentation (Harrison 1951), implies a 60–70 Myr lapse between sedimentation and peak metamorphism. The enormous volume of metasediment in the Kisseynew belt suggests proximity to a contemporaneous collision zone, consistent with isotopic evidence for a greater contribution from older crust in the Kisseynew belt than in the bounding volcanic-plutonic belts (Chauvel et al 1987).

Although complex in detail, much of the structural evolution of the internal zone can be described in terms of two main phases of deformation, here referred to as D1 and D2 (e.g. Froese & Moore 1980, Stauffer 1984, Bailes et al 1987, Lewry et al 1987). The dominant map-scale structures are broad, upright D2 folds plunging gently to the northeast. They refold systems of stratigraphically distinct thrust-nappes containing originally recumbent, near-isoclinal D1 folds (e.g. Bailes & Syme 1987, Lewry et al 1987). The nappes are bounded by faults and shear zones having northwest-trending stretching lineations. Preliminary analysis of stratigraphic cutoffs and kinematic indicators suggests that both southeast- and northwest-vergent thrusting is present, with the former predominant. For example, the Railway fault system at Flin Flon appears to be a folded east- or southeast-vergent thrust fan (cf Bailes & Syme 1987), whereas the McLeod Road fault at Snow Lake, 120 km to the east, appears to be a northwest-vergent thrust (cf Froese & Moore 1980). Preliminary observations of rotated porphyroclasts in mylonite of the northwest-dipping Stanley shear zone indicate southeast-vergent thrusting betweeen the La Ronge and Glennie Lake belts; however, shear bands with top-side-down sense of movement occur in the northwest-dipping McLennan Lake tectonic zone within the La Ronge belt (cf Lewry & Slimmon 1985).

Although the gross structural evolution (i.e. D1 thrust-nappes deformed by broad upright D2 folds) appears similar throughout the internal zone, the timing of the thermal peak with respect to the structural evolution varies across the zone. In the northwest, strong synmetamorphic fabrics were developed during D1 thrusting; in the southeast, the dominant syn-

metamorphic fabrics are related to the upright D2 folds. Accordingly, stretching lineations in the northwest were folded during D2 folding and have moderately large rakes in the D1 foliation planes; stretching lineations in the southeast at Snow Lake plunge gently to the northeast, coaxial with the D2 folds (Bailes et al 1987). The latter observation implies that D2 folding was accompanied by hinge-line extension, compatible with deformation involving a significant component of noncoaxial strain.

One possible model for the tectonic evolution of the orogen (Figure 6A) involves meridional convergence and collision between the Superior and Hearne provinces. Convergence was accommodated by prevailing north-dipping subduction of oceanic lithosphere resulting in the development of an accretionary prism(s) in which dominantly south-directed D1 thrusting occurred. During collision, the accretionary complex was subjected to sinistral transpression between the ENE-trending continental margins,
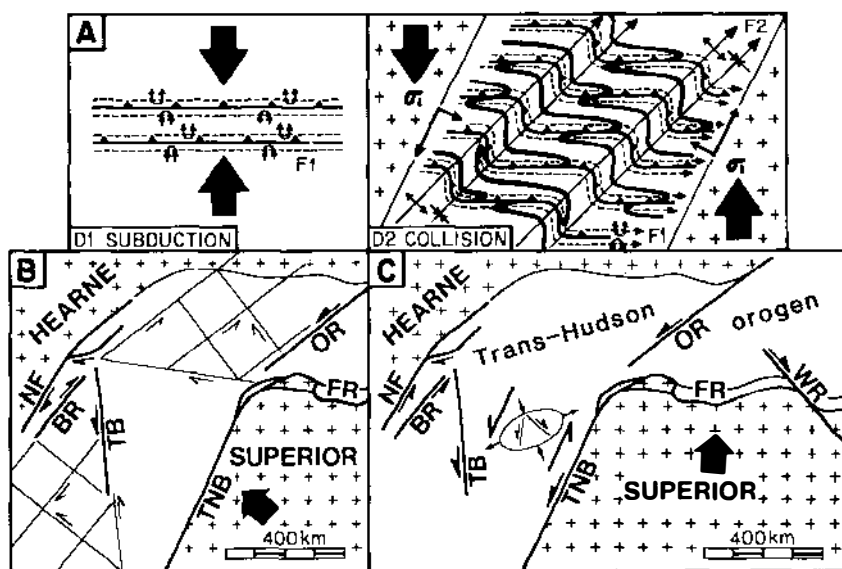

Figure 6   (A) Two-stage structural model for the internal zone of the Trans-Hudson orogen discussed in the text. (B) Slip-line field model proposed by Gibb (1983), assuming north-westward convergence of the Superior province relative to the Hearne province. (C) Model compatible with structural model (A), assuming northward convergence of the Superior province relative to the Hearne province. Note the difference in sense of movement on the Needle Falls and Birch Rapids shear zones in the two models (B) and (C). Abbreviations: BR, Birch Rapids shear zone; FR, Fox River belt; NF, Needle Falls shear zone; OR, Owl River shear zone; TB, Tabbernor fault zone; TNB, Thompson belt; WR, Winisk River shear zone.

causing the D2 folding and counterclockwise rotation of the D1 lineations. The model is consistent with the maximum development of the Wathaman-Chipewyan batholith on the latitudinal margin of the orogen and with the interpretation of the Tabbernor fault (Figure 5) as a meridional transform fault linked to north-dipping subduction beneath the Lynn Lake arc (Lewry 1981). The model predicts a component of left-slip between the Kisseynew and Thompson belts (Green et al 1985a), for which preliminary evidence has been observed (W. Bleeker et al, in preparation). An alternative model (Gibb 1983), based on slip-line field analysis of the transcurrent shear zones, implies late-stage northwest-southeast-oriented convergence between the Superior and Hearne provinces (Figure 6B). This model predicts right-slip motion on the Needle Falls and Birch Rapids shear zones, where preliminary indications of left-slip have been observed (J. F. Lewry, personal communication, 1987). More structurally oriented field work is urgently needed to clarify the tectonic evolution of this key orogenic belt.

Kapuskasing Uplift: Intracratonic Thrust Related to the Trans-Hudson Orogen?

The Kapuskasing uplift (Figure 1) is a composite northeast-trending, 500-km-long zone of granulite and upper-amphibolite-facies rocks that transects the east-west Archean granite-greenstone and metasedimentary-gneiss belts of the central Superior province. The belts on either side of the uplift correlate lithologically and geochronologically, and therefore the uplift is likely intracratonic in origin (Percival & Card 1983, 1985). The deep structure of the uplift has been inferred from paleomagnetism (Ernst & Halls 1984), seismic reflection profiling (Cook 1985), metamorphic geobarometry (Percival & McGrath 1986), and modeling of gravity and magnetic anomalies (Percival & McGrath 1986). The east side of the uplift is bounded by a system of west-dipping, crustal-scale thrusts. In the north, the west side is defined by an east-dipping back-thrust, and the uplift has the geometry of a pop-up; in the central part, the west side is defined by a system of west-dipping normal faults, and the uplift has the geometry of a perched thrust-tip; and in the south the uplift is a simple west-dipping slab (Percival & McGrath 1986). Geometrically, the uplift resembles the Laramide structures of Wyoming and adjacent states. Integrated isotopic studies (U-Pb zircon and sphene; K-Ar and $^{40}Ar/^{39}Ar$ hornblende, biotite, and whole-rock; Rb-Sr biotite) suggest that the high-grade rocks cooled nearly isobarically in the lower crust from about 2.6 Ga until they were brought to the surface by thrusting at about 2.0–1.9 Ga (J. A. Percival, personal communication, 1986). Thrusting must be older than carbonatite and alkaline igneous complexes that intruded

the uplift and its boundary faults at 1.87–1.86 Ga (T. E. Krogh, personal communication, 1984; Bell et al 1987). Gibb (1978b, 1983) suggests a genetic link between the Kapuskasing uplift and the parallel Thompson belt of the Trans-Hudson orogen, 1250 km to the northwest. This suggestion is compatible with the possible synchroneity of the Kapuskasing uplift with foredeep(?) magmatism in the Fox River belt (1.88 Ga; Heaman et al 1986) and with deformation in the Thompson belt (post-1.88/pre-1.79 Ga; Machado et al 1987). Geometrically, however, the Gibb (1983) model is difficult to reconcile with right-slip on the Winisk River fault evident from magnetic anomalies (Committee for the Magnetic Anomaly Map of North America 1987).

New Quebec ("Labrador Trough") Orogen: Rae/Superior Collision

The northeast margin of the Superior province and the opposing margin of the Nain province (Figure 1) are bounded by the New Quebec and Torngat orogens, respectively. The orogens share a common 180–280-km-wide Archean hinterland (Ashwal et al 1986, Machado et al 1988), occupied by the George River drainage basin and Ungava Bay (Figure 7). The hinterland appears to represent, together with the Dorset fold belt of southern Baffin Island (Jackson & Taylor 1972), a southeastern extension of the Rae province. The sedimentary-volcanic belt related to the Rae/Slave collision (Kearey 1976) is widely known as the "Labrador Trough," a name that this author has been urged to abandon in order to avoid confusion with the recently named, geologically younger "Labrador orogen" (Thomas et al 1985) and "Trans-Labrador batholith" (Gower & Owen 1984) to the southeast (Figure 7). The proposed name suits because most of the belt lies in New Quebec, not Labrador, and because the belt is a structural depression but not a primary depositional "trough" (e.g. Dimroth 1981). The belt has long been considered as being continuous with the Trans-Hudson orogen (e.g. Baragar & Scoates 1981), but this remains to be demonstrated geochronologically and assumes that the Rae and Hearne provinces were already joined when they collided with the Superior province.

The New Quebec orogen contains an 800-km-long, southwest-vergent, foreland thrust-fold belt (Wardle 1982, Boone & Hynes 1987) that has been eroded at its north end across a transverse flexural culmination parallel to the Cape Smith belt (Hoffman 1985) and redeformed at its south end by northwest-vergent structures of the Grenville orogen (Rivers & Wardle 1985, Rivers & Chown 1986). The belt is bounded to the southwest by the flexurally arched Archean foreland of the Superior
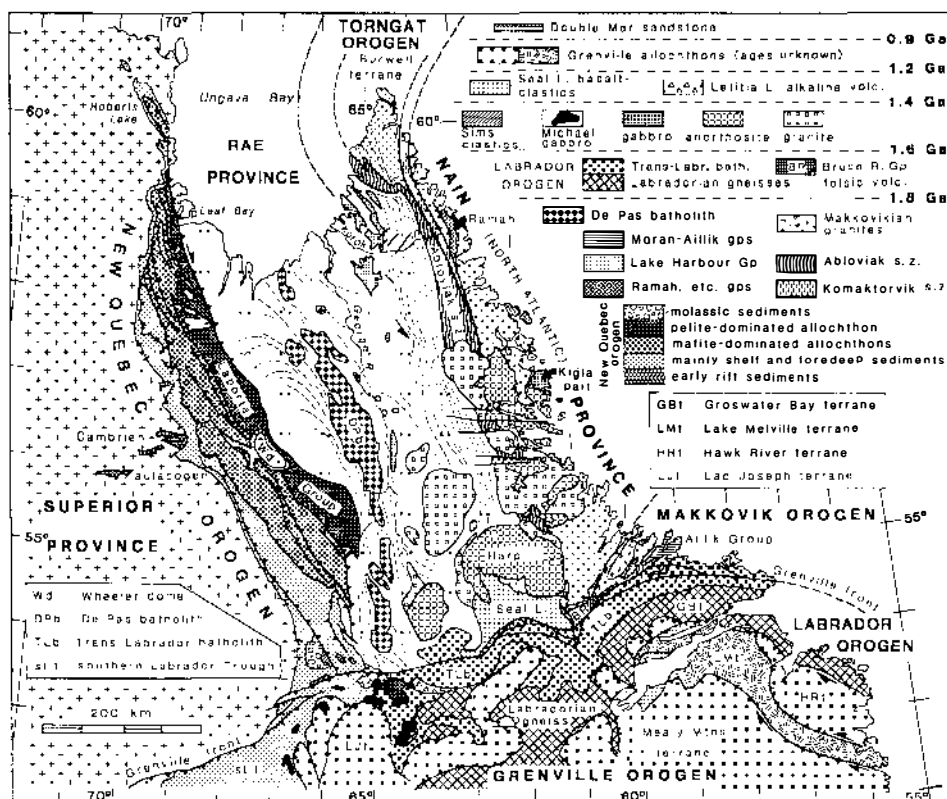
Figure 7 Geology of northeastern Quebec ("New Quebec") and Labrador. The New Quebec ("Labrador Trough") orogen is the Early Proterozoic collision zone between the Archean Superior and Rae provinces. The Torngat orogen is the Early Proterozoic collision zone between the Archean Rae and Nain provinces. The Makkovik and Labrador orogens involved accretion of Early Proterozoic crust onto the Nain province and the Rae-Nain assembly, respectively. Middle Proterozoic anorogenic granite-anorthosite-gabbro suites intruded the Rae and Nain provinces and the Torngat orogen. During the late Middle Proterozoic Grenville orogeny, all of the older crustal elements were telescoped by means of northwest-directed thrusting.

province and to northeast by the allochthonous Archean hinterland of the Rae province. From southwest to northeast, the belt comprises an autochthonous sedimentary zone, a medial zone of igneous and sedimentary allochthons that can be more or less confidently correlated with the autochthonous cover, and a metasedimentary zone (Laporte Group) of uncertain ancestry. The ductile sole thrust of the belt, characterized by northeast-

trending stretching lineations and sheath-folds that have been refolded by northwest-trending basement-involved folds, is exposed in the structural culmination on the west side of Ungava Bay (Figure 7).

The autochthonous strata consist of three unconformity-bounded sequences that overstep sequentially onto the Archean foreland (Le Gallais & Lavoie 1982). The oldest sequence is exposed mainly in the medial allochthons but also is observed in the autochthonous Cambrian Lake aulacogen (Clark 1984). From the base, it comprises fluvial redbeds (Seward Subgroup), marine-shelf quartzite and carbonate (Pistolet Subgroup), foredeep black shale and turbidites (Swampy Bay Subgroup), and a regressive carbonate reef-complex (Denault/Abner Formation). The middle sequence begins with a transgressive high-energy quartzarenite (Wishart Formation), overlain by an iron-formation (Sokoman Formation) and a second foredeep flysch (Menihek Formation). The third sequence is a fluvial molasse (Chioak/Tamarack River Formation) containing westerly derived clasts of basement and middle-sequence lithologies. The overall sequence resembles that of the central Appalachians in having two successive foredeeps developed above initial-rift and passive-margin prisms (Hoffman 1987b).

The thrust sheets to the northeast carry sequences in part correlative with those already described but also containing thick piles of mafic flows and sills, locally with small felsic centers (Baragar 1960, Dimroth 1971). Although correlations are uncertain, major tholeiitic suites appear to occur in the Swampy Bay and Menihek foredeeps, as do alkalic(?) suites in the basal rift sequence and in the Sokoman iron-formation. The allochthon structurally beneath the Laporte Group is composed of mafic and ultramafic rocks (Doublet Group) of unknown stratigraphic affinity. The basal rift-related volcanics have not been dated, but rhyolites tentatively included in the Swampy Bay foredeep are 2.14 Ga (T. E. Krogh & S. A. Bowring, personal communications discussed in Hoffman & Grotzinger 1987). Syenite associated with the Sokoman iron-formation and a glomerophyric gabbro sill in the Menihek foredeep are 1.88 Ga (R. Parrish, personal communication, 1987), and a similar sill in the correlative(?) Hellancourt Formation is 1.87 Ga (Machado et al 1988). These are maximum ages for thrusting and metamorphism. Pb isotope ages for galena mineralization in various stratigraphic units (Clark & Thorpe 1987) are in broad agreement with the U-Pb zircon ages. These preliminary geochronological data indicate that while initial rifting in the New Quebec orogen (pre-2.14 Ga) greatly predated rifting in the Cape Smith belt (1.96 Ga), the Wishart-Sokoman-Menihek sequence (1.88 Ga) was deposited during the tectonic evolution of the Cape Smith belt (1.96–1.84 Ga).

The Laporte Group consists of arkosic semipelitic schist that may rep-

resent a passive continental-rise prism belonging to the Superior province (Wardle & Bailey 1981) or a fore-arc basin deposited on the leading edge of the Rae province (van der Leeden et al 1988). The northeast boundary of the Laporte zone west of Ungava Bay is a northeast-dipping, post-metamorphic fault having a major component of dextral transcurrent slip (Goulet et al 1987). There, basement just east of the Laporte zone is overlain by coarse grits and metatuffs, intruded by metatonalite, possibly representing a proximal fore-arc environment (Poirier et al 1987). The tectonic significance of the domal Archean basement inliers within the Laporte zone (e.g. Wheeler dome; see Figure 7) is uncertain; they may be either tectonic windows exposing the autochthonous Superior province beneath the allochthonous Laporte zone or structural culminations of allochthonous basement on which the Laporte metasediments were deposited. Metamorphic grade in the Laporte zone increases eastward from greenschist to granulite facies, and the thermal peak was preceded by the pressure acme of 0.8 GPa, suggesting that metamorphism was a consequence of thermal relaxation following thrusting (Perreault et al 1987). Peak metamorphism postdates emplacement of the domal basement inliers, and its age is given by U-Pb monazite dates of 1.79 Ga for the basement inliers and 1.78 Ga for basement on the east margin of the belt, which have U-Pb zircon ages of 2.88 and 2.71 Ga, respectively (Machado et al 1988). As thrusting and metamorphism appear to be continuous throughout the New Quebec orogen, it is inferred that thrusting in response to the Rae/Superior collision occurred after 1.87 Ga (the age of gabbro sills emplaced prior to thrusting) and before 1.79 Ga (Machado et al 1988).

Intruding the Rae province 25–100 km east of the trough is the NNW-trending De Pas batholith (Figure 7), a calc-alkaline plutonic belt coincident with a pre- to synplutonic dextral granulite-grade shear zone involving Archean basement in the north and the Laporte Group in the south (Taylor 1979, van der Leeden et al 1988). Preliminary zircon ages for the batholith are 1.84 and 1.81 Ga (S. A. Bowring & T. E. Krogh, personal communications, 1986), and a monazite age of 1.81 Ga dates the metamorphism of an Archean migmatitic host rock of the batholith (Machado et al 1988). The batholith is interpreted as a continental magmatic arc related to dextral oblique convergence between the Superior and Rae provinces (van der Leeden et al 1988); however, its age relative to the inferred collision suggests that plutonism may be partly or entirely due to postcollisional anatectic melting. The southern part of the batholith and adjacent Archean rocks are thrust westward on structures that parallel the Torngat orogen (Figure 6). This deformation clearly postdates the thin-skinned thrusting in the New Quebec orogen (cf Wardle 1982), suggesting that the Torngat orogen is the younger of the two collision zones.

Torngat Orogen: Rae/Nain Collision

Torngat orogen is the name proposed for the zone of intense Early Pro-
terozoic deformation and metamorphism between the Rae and Nain prov-
inces, best exposed in the Torngat Mountains of northern Labrador (Fig-
ure 7). The orogen is thought to be a compound suture zone, bisected in
the north by the Burwell terrane (Taylor 1979, Korstgård et al 1987), a
possible Archean(?) microcontinent.

The Torngat orogen is a mirror image of the New Quebec orogen,
thrusting in the Nain foreland being east-vergent and transcurrent shear
in the Rae hinterland being sinistral (Korstgård et al 1987). In the foreland,
sedimentary cover of the Ramah Group is preserved in a 110 by 10 km
fold belt, which comprises a lower siliclastic shelf sequence overlain by
foredeep flysch intruded by mafic sills (Knight & Morgan 1981). Meta-
morphism of the cover increases from greenschist to upper amphibolite
facies toward the western margin of the fold belt, which is truncated by
west-dipping basement-rooted thrusts (Morgan 1979, 1981, Mengel 1985).

To the west of the Ramah Group is a crustal-scale shear zone that
bifurcates northward into the Komaktorvik and Abloviak shear zones,
which bound the eastern and western sides of the Burwell terrane, respec-
tively (Wardle 1983, Korstgård et al 1987). The former is an asymmetric
zone, up to 20 km wide, of east-vergent thrusting and sinistral transcurrent
shear, in which the latitudinal pre–Ramah Group dikes of the Nain prov-
ince foreland are progressively transposed. Metamorphic grade increases
westward across the zone, where granulite-facies rocks characterized by
large bodies of sheared Archean anorthosite occur (Wardle 1983). Farther
to the west, the Abloviak shear zone is marked by a belt, 5–45 km wide,
of garnetiferous mylonite derived from pelitic diatexite. The mylonite
contains a subhorizontal stretching lineation and coincides with a marked
negative magnetic anomaly that contrasts with the linear magnetic highs
corresponding to the adjacent granulite-facies orthogneisses. The magnetic
low is continuous for 900 km along strike, from the Grenville front to
Resolution Island off the southeast tip of Baffin Island (Figure 8). A
sinistral kink in the Abloviak shear zone occurs near the southern ter-
mination of the Burwell terrane (Taylor 1979). The absence of a kink in
the adjacent Komaktorvik shear zone suggests that the Abloviak shear
zone may be the older of the two shear zones.

Near the Korok River west of the Abloviak shear zone, a fold belt
involving granitoid basement and sedimentary cover of probable Archean
and Early Proterozoic age, respectively, has been sinistrally sheared under
metamorphic conditions decreasing westward from granulite to amphibo-
lite facies (Wardle 1984). The cover contains quartzite, marble, and euxinic
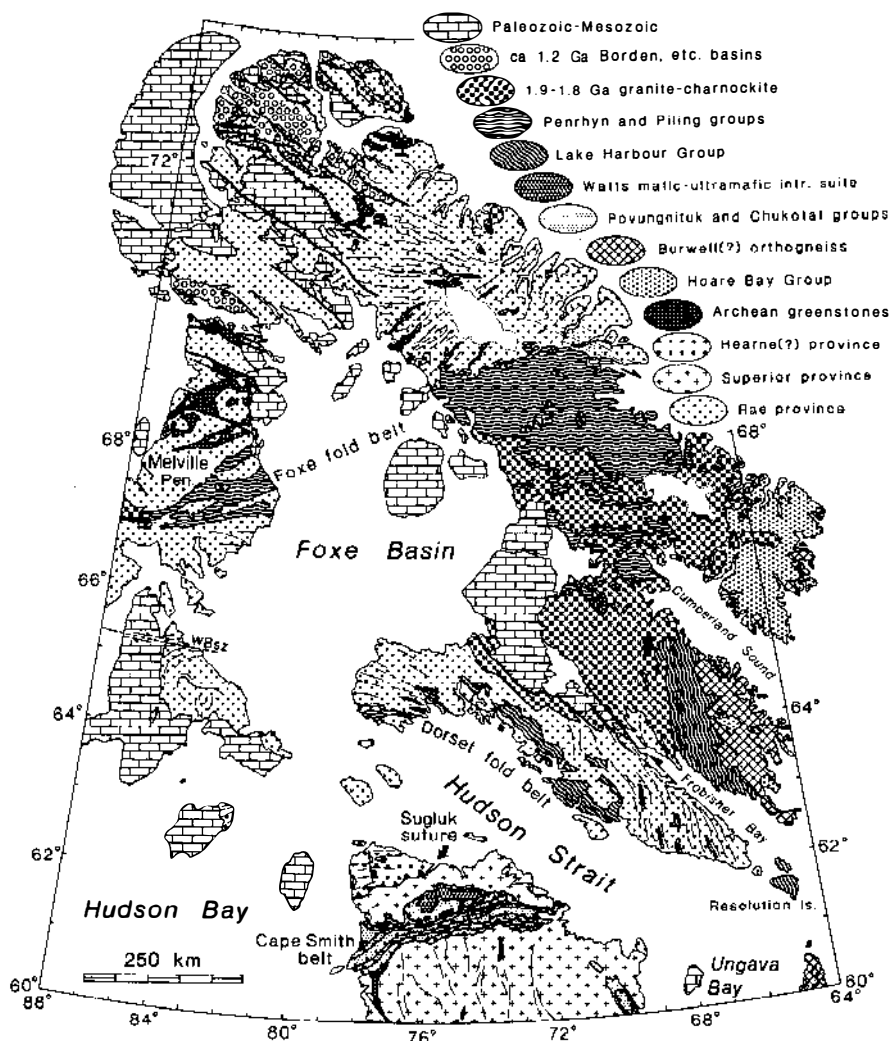
Figure 8 Geology of the Baffin Island, Foxe. Basin, and Hudson Strait area, showing possible extensions of the Archean Rae and Hearne provinces and of the Burwell terrane. Interpretations for southern Baffin Island are extremely tenuous because of the absence of U-Pb geochronology. Abbreviation: WBsz, Wager Bay shear zone.

pelite, resembling the Lake Harbour Group of the Dorset fold belt on southern Baffin Island (Jackson & Taylor 1972, Taylor 1979). The Korok River and Lake Harbour sediments are interpreted as epicontinental deposits of the Rae province, lithologically similar to the Amer Group of

the northwestern Rae province and cover strata of the Foxe fold belt (Figure 8) in the northeastern Rae province. The protoliths of the Abloviak shear zone possibly include off-shelf correlatives of the Korok River (epi-Rae) and/or Ramah (epi-Nain) sedimentary cover.

The age of the Torngat orogen is very poorly constrained. It must be younger than the Ramah Group, which unconformably overlies dikes that have a minimum (Rb-Sr) age of about 2.3 Ga, and is older than the 1.65-Ga Trans-Labrador batholith (Figure 7). A single zircon age of about 1.91 Ga has been obtained from a mylonite in the Komaktorvik shear zone southwest of the Kiglapait intrusion (U. Schärer, personal communication, 1987), but the significance of this age is uncertain. If meridional basement-involved thrusts in the southern New Quebec orogen (Figure 7) are cogenetic with the Torngat orogen, as suggested earlier, then the Torngat orogen must be younger than 1.88 Ga, possibly younger than 1.81 Ga.

South Baffin Batholith and Its Bounding Fold Belts

Northern Baffin Island is underlain by a 2.9–2.7 Ga granite-greenstone-gneiss terrane that represents an extension of the Rae province of the northern District of Keewatin (Jackson & Taylor 1972). It includes scattered nonfoliated granites and charnockites of possible Early Proterozoic age (Jackson & Morgan 1978) and the Borden and related basins of Middle Proterozoic age (Jackson & Iannelli 1981). South-central Baffin Island contains a 250,000 km^2 charnockite-granite batholith bounded by three fold belts (Figure 8): the Foxe fold belt to the north, the Dorset fold belt to the southwest, and the Hoare Bay fold belt to the east (Jackson & Taylor 1972). The batholith is 1.90–1.87 Ga (Pidgeon & Howie 1975, Henderson 1985a) and is intrusive into granulite-facies equivalents of the bounding fold belts (Jackson & Morgan 1978).

The Early Proterozoic Foxe fold belt of Baffin Island extends westward onto Melville Peninsula (Figure 8) and eastward, prior to the opening of Baffin Bay, into the Rinkian belt (Escher & Pulvertaft 1976) of central west Greenland (Figure 1). The Early Proterozoic cover strata of the belt comprise a thin lower sequence of quartzite, schist, and marble of platformal facies, and an upper sequence of ferruginous pelite, locally with mafic-ultramafic flows and sills, and a great thickness of graywacke turbidites, resembling foredeep flysch (Morgan et al 1976, Henderson & Tippett 1980, Henderson 1983). The cover sequence is known as the Piling Group on Baffin Island, the Penrhyn Group on Melville Peninsula, and the Karrat Group in west Greenland, all possibly correlative.

On central Baffin Island, the upper sequence has been shortened to produce tight, closely spaced, steeply inclined, ENE-trending folds that predate the metamorphic peak (Morgan 1983). The upright folds pass

downward into polyphase recumbent folds possibly associated with a ductile detachment zone separating the cover from the basement. Lobate basement-cored nappes trending SSE occur near the southern margin of the belt (Henderson et al 1979, Henderson 1985a). These early folds are refolded by two sets of upright basement-involved folds (Henderson & Tippett 1980, Henderson 1985a,b). ENE-trending lobate basement-cored antiforms and cuspate synforms (Morgan 1983) were formed during waning metamorphism, which increases in grade outward and structurally downward from greenschist facies in the central part of the belt to upper-amphibolite facies along the northern margin and eastern end of the belt and granulite facies along its southern margin (Jackson & Morgan 1978). A younger set of ESE-trending folds is developed toward the east end of the fold belt and affects the Archean rocks far to the north, producing recumbent folds that are spectacularly exposed in the fiords along the northeast coast of Baffin Island. A 1.81-Ga tonalitic pegmatite is syntectonic with the ESE-trending folds, dating the deformation that has been tentatively attributed to dextral transcurrent shear parallel to the coast (Henderson & Loveridge 1981).

On Melville Peninsula, the Foxe fold belt is characterized by belt-parallel stretching lineations and sheath folds, superimposed on a large-scale basement-cover nappe system, that converge to the southwest, where the cover pinches out (Henderson 1983, 1984). The charnockite-granite batholith of Baffin Island does not appear on Melville Peninsula, but a zone of composite calc-alkaline plutons extends for 350 km southwestward from the west end of the fold belt to the Wager Bay shear zone (Figure 4). The plutons were emplaced at 1.83–1.82 Ga, during waning metamorphism of the fold belt, and slightly predate pegmatites (1.82–1.81 Ga) that are synkinematic with respect to the dextral transcurrent Wager Bay shear zone (Henderson et al 1986, LeCheminant et al 1987c).

The Dorset fold belt appears to be coextensive with the western hinterland of the Torngat orogen (Jackson & Taylor 1972). The fold belt comprises amphibolite- to granulite-grade flyschoid metasediments; subordinate quartzite, marble, and rusty pelite; rare mafic bands; and granitoid orthogneiss. Although no internal stratigraphy, basement/cover relations, or U-Pb dating is available for this superbly exposed 700-km-long belt, the presence of platformal facies suggests that much of the orthogneiss may be basement to the metasediments. The belt is wrenched along a northwest-trending sinistral shear zone occupied by Frobisher Bay (Figure 8). East of the batholith is the Hoare Bay Group, an undated assemblage of mainly semipelitic schist and gneiss, lesser mafic metavolcanic and ultramafic rocks, and granitoid orthogneiss. The more westerly parts of this assemblage have been compared lithologically with rocks

of the Foxe and Dorset fold belts, but other parts more closely resemble Archean greenstone belts north of the Foxe fold belt (Jackson & Taylor 1972). The age of the Hoare Bay Group is unknown, but the fold belt east of the batholith may represent an extension of the Burwell terrane of northern Labrador. Prior to the opening of Baffin Bay, the Hoare Bay belt was situated adjacent to the Nagssugtoqidian belt (Figure 1) of west Greenland (Myers 1984).

The tectonic significance of the 1.90-Ga charnockite-granite batholith of southern Baffin Island and its bounding Early Proterozoic fold belts is enigmatic, as are their relations to the Nagssugtoqidian belt of Greenland. A pivotal problem is whether the presumed Archean basement of the Dorset fold belt is an extension of the Rae province (Figure 1). If so, the Trans-Hudson and Snowbird orogens do not extend northeastward beyond Hudson Bay (e.g., Lewry et al 1985) but must curve around the northeast margin of Superior province and link up with the New Quebec orogen. These important problems may remain unresolved until high-resolution magnetic anomaly data are available for Foxe Basin and Hudson Strait.

ACCRETION IN THE SOUTH

The north-central part of Laurentia was assembled through a rapid succession of microcontinental collisions, from 1.96 to about 1.80 Ga in age, involving the Archean Slave, Rae, Hearne, Wyoming, Superior, and Nain provinces (Figure 1). The remaining one third of the craton consists mostly of juvenile Early Proterozoic crust accreted prior to 1.6 Ga (Patchett & Arndt 1986). It is best to describe the Early Proterozoic accreted terranes south of the Nain, Superior, and Wyoming provinces independently, because they are exposed in widely separated areas and it is not yet certain how much accretion occurred before or after the collisions to the north.

The Early Proterozoic accreted terranes of Laurentia are correlative with belts in the Baltic shield (Figure 1), with which they may have been coextensive prior to opening of the Iapetus paleoocean (Gower & Owen 1984, Gower 1985). There are three sets of coeval accreted belts, decreasing in age from north to south. The Penokean orogen of the Great Lakes area, the Makkovik orogen of Labrador (Gower & Ryan 1986), the Ketilidian orogen of south Greenland (Allaart 1976), and the Svecofennian orogen of the Baltic shield (Gaal & Gorbatschev 1987) are all 1.9–1.8 Ga. The Yavapai orogen extending through northwestern Arizona and Colorado (Karlstrom & Bowring 1987), the Central Plains orogen (Sims & Peterman 1986), and the Killarney belt of Lake Huron (van Breemen & Davidson 1987) are 1.8–1.7 Ga. The Mazatzal orogen of southeastern Arizona and

New Mexico (Karlstrom & Bowring 1987), the Labrador orogen (Thomas et al 1986), and the Trans-Baltic (or Småland-Värmland) belt (Gaal & Gorbatschev 1987) are all 1.7–1.6 Ga. Collectively, an area of new crust up to 1200 km wide and 5000 km long was accreted in less than 300 Myr.

Isotopic data indicate that Early Proterozoic crust underlies the external parts of the Grenville orogen (Ashwal et al 1986, Schärer et al 1986, Schärer & Gower 1987, van Breemen & Davidson 1987, Rivers et al 1988) and the correlative Sveconorwegian orogen (Demaiffe & Michot 1985), but that internal parts of the orogen on both sides of the Atlantic consist mainly of rocks younger than 1.3 Ga.

Accretion to Nain Province: Makkovik and Labrador Orogens

The Nain province is bounded to the southeast by the Ketilidian orogen of south Greenland (Allaart 1976) and the coextensive Makkovik orogen (Gower & Ryan 1986), which occupies a wedge-shaped area of coastal Labrador truncated to the south by the Trans-Labrador batholith (Figure 7). The Ketilidian orogen and the correlative Svecofennian orogen of the Baltic shield have external zones developed on or adjacent to Archean crust; their internal zones are composed of juvenile 1.9–1.8 Ga crust (Patchett & Bridgwater 1984, Kalsbeek & Taylor 1985, Huhma 1986, Patchett & Kouvo 1986). The Makkovik orogen corresponds to the external zones of the Ketilidian orogen, but possible equivalents of the internal zones occur as enclaves within the younger (1.7–1.6 Ga) Labrador orogen to the south.

In the Makkovik orogen (Figure 7), Archean basement continuous with the Nain province is overlain unconformably by the Early Proterozoic Moran Lake Group (Ryan 1984) and structurally by the correlative lower Aillik Group (Marten 1977). Both groups consist of a basal quartzarenite overlain by graywacke-semipelite and mafic metavolcanic rocks, which are in fault contact to the east with the upper Aillik Group, comprising calc-alkaline, dacitic to rhyolitic volcanic and associated clastic sedimentary rocks (Gower et al 1982, Gandhi 1984, Ryan 1984). These rocks were deformed during an early phase of north(?)-directed low-angle thrusting (including basement-cover imbrication), a main phase of upright base-ment-involved folding of NNE trend, and a late phase of northeast-trend-ing dextral shearing (Marten 1977, Clark 1979, Korstgård & Ermanovics 1984, Gower & Ryan 1986). Metamorphic grade increases southeastward into the orogen from lower-greenschist facies at the Nain province margin to middle-amphibolite facies. Thrusting postdated the 1.86–1.81 Ga upper Aillik volcanics; the main upright folding event occurred during or after the emplacement of 1.81–1.80 Ga granites, and waning metamorphism is

dated at 1.79 (U-Pb monazite) and 1.76 (U-Pb titanite) Ga (Schärer et al 1987, Ermanovics et al 1987). In the Ketilidian orogen, synmetamorphic(?) granites are 1.81–1.80 Ga and postmetamorphic rapakivi granites are 1.76–1.74 Ga (van Breemen et al 1974, Gulson & Krogh 1975, Patchett & Bridgwater 1984).

The recently defined Labrador orogen (Thomas et al 1985, 1986, Wardle et al 1986, Schärer et al 1986) is a broad, ENE-trending zone of 1.71–1.63 Ga gneissic, plutonic, and volcanic-sedimentary rocks that make up most of the northeastern Grenville orogen and, locally, extend north of the Grenville front. These rocks were deformed and metamorphosed up to granulite facies during the Labradorian orogeny, which culminated at about 1.65 Ga, before being structurally telescoped and metamorphosed again during the Grenvillian orogeny about 1.03–0.97 Ga. Rocks of the Labrador orogen overlie, intrude, or truncate the Rae and Nain provinces and the New Quebec, Torngat, and Makkovik orogens (Figure 7).

The zonation of the Labrador orogen comprises a northern volcanic-sedimentary belt (Bruce River Group) preserved in the foreland of the Grenville orogen, a Grenvillian parautochthonous zone containing the 1.65-Ga Trans-Labrador batholith flanked to the south by amphibolite-grade Labradorian para- and orthogneisses, and a zone of Grenvillian allochthons (folded about southeast-plunging axes) composed of granulite-grade Labradorian gneisses. The 1.65-Ga Bruce River Group contains a lower assemblage of alluvial fanglomerate, sandstone, and minor bimodal lava flows; a middle sequence of volcaniclastic arenites; and an upper association of potassic calc-alkaline mafic-intermediate-felsic lava flows, pyroclastic flows, related intrusions, and derived volcaniclastic sediments (Baragar 1981, Ryan 1984, Schärer et al 1987). The 500-km-long Trans-Labrador batholith consists mainly of calc-alkaline granite and grano-diorite, along with subordinate early gabbro-diorite bodies, and is syn- to posttectonic with respect to the Labradorian orogeny (Gower & Owen 1984, Thomas et al 1986). The amphibolite- and granulite-grade Labradorian metasediments are generally monotonous pelitic to psammitic gneisses, containing strongly to weakly deformed granitic to gabbroic and anorthositic intrusions older than about 1.63 Ga. Isotopically, the Labradorian gneisses and associated intrusions appear not to contain any crustal components older than about 1.71 Ga (Wardle et al 1986, Schärer et al 1986). Thus, the Labradorian orogeny involved major accretion of new continental crust that was later incorporated in the eastern Grenville orogen.

Accretion to the Superior Province: Penokean Orogen

The evolution of the Early Proterozoic southern margin of the Superior province is recorded in the Penokean orogen exposed around Lake

Superior (Figure 9). This 1.90–1.83 Ga orogen forms an embayment on the Superior province margin, also occupied by the younger Keweenawan rift system. Uplift of the Penokean shield south of Lake Superior is of Keweenawan (1.14 Ga) age (Peterman et al 1985) and may be a flexural consequence of loading by rift volcanics, focused inside the U-shaped bend in the rift system (Peterman & Sims 1986). The westward extension of the orogen appears to be truncated in the subsurface by the 1.80–1.63 Ga Central Plains orogen (Sims & Peterman 1986). Its eastward extension is truncated by or incorporated within the Grenville orogen. The Otish and
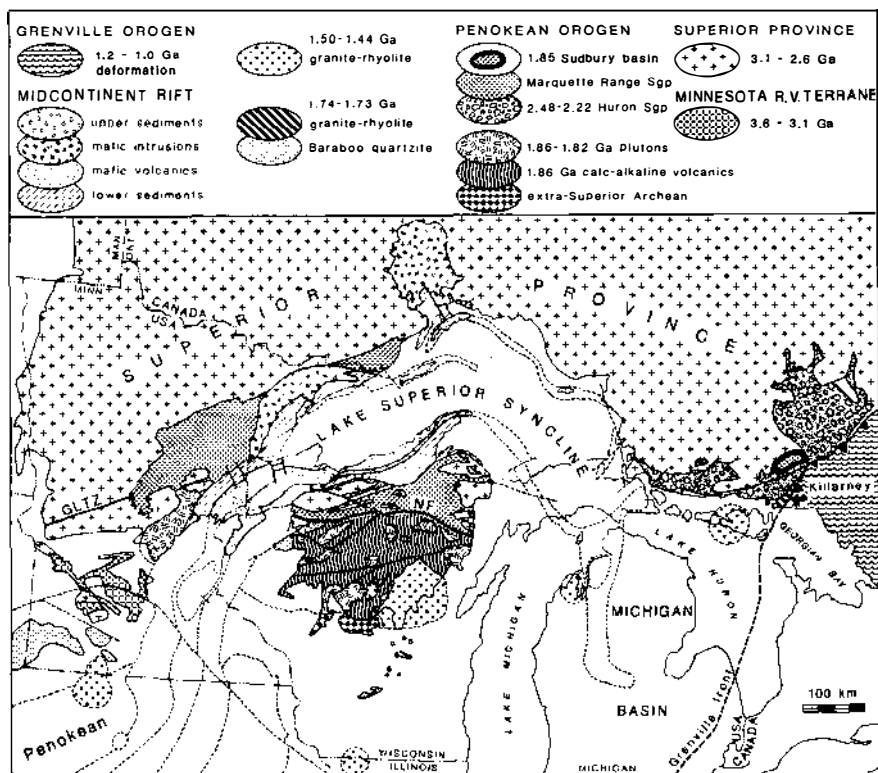

Figure 9 Precambrian tectonic elements of the Great Lakes region, modified from P. K. Sims (personal communication, 1985). The Great Lakes tectonic zone (GLTZ) is a possible suture, occluded by late Archean granitic intrusions, between the Superior province and the Minnesota River Valley terrane. The Niagara Fault zone (NF) separates the ensialic Marquette Range Supergroup from ensimatic magmatic-arc terranes accreted during the ca 1.85 Ga Penokean orogeny. Note the small area of Archean rocks south of the Penokean island-arc terrane, and the post-Penokean (ca 1.75 Ga) granite-rhyolite inliers west of Lake Michigan and in the Killarney area of Georgian Bay.

Mistassini outliers (Chown & Caty 1973, Rivers & Chown 1986) in the foreland of the Grenville orogen in Quebec may also be related to the Penokean orogen.

The Niagara fault zone divides the Penokean orogen into a northern domain of sedimentary and minor tholeiitic volcanic rocks deposited on Archean basement of the Superior province and the Minnesota River Valley terrane, and a southern domain composed mainly of Early Proterozoic volcanic and plutonic rocks resembling island-arc suites (Morey et al 1982, Greenberg & Brown 1983, Schulz 1983, Klasner et al 1985). The absence of Early Proterozoic arc-type igneous rocks north of the Niagara fault zone favors models involving south-dipping subduction terminated by collision of island arc(s) against a south-facing passive continental margin (Cambray 1978, Larue & Sloss 1980, Schulz et al 1984).

The deformed continental margin includes two different depositional prisms: the Huron ("Huronian") and Marquette Range supergroups. East of Lake Superior, the Huron Supergroup is a southward-thickening siliciclastic prism (Roscoe 1968) deposited between 2.48 and 2.22 Ga (Krogh et al 1984, Corfu & Andrews 1986). The prism has an aggregate maximum thickness of 12 km and progressively onlaps the Superior province to the north, from which most of the sediment was derived (Card 1984). Polyphase deformation of its southern part into easterly trending folds and northward-vergent thrusts (Zolnai et al 1984) has been ascribed to the Penokean orogeny (Brocoum & Dalziel 1974). Much of the deformation postdated the 1.850-Ga (Krogh et al 1984) Sudbury impact event (Rousell 1984), and most of it predated the 1.74-Ga Killarney granite (Davidson 1986). However, folding began before intrusion of the 2.22-Ga (Corfu & Andrews 1986) Nipissing diabase, and prograde metamorphism of Huronian strata occurred prior to the Sudbury event (Card 1978).

Conversely, the Marquette Range Supergroup south of Lake Superior and its equivalents to the northwest may be closely related to the Penokean orogeny in age and origin, although this remains to be demonstrated radiometrically. They contain lower assemblages (Mille Lacs and Chocolay groups) in which immature clastics and mafic volcanics are overlain by shelf-type quartzarenite and dolomite (Larue 1983). Overstepping these rocks is a more extensive upper sequence (Animikie Group), in which a basal trangressive littoral quartzarenite (Ojakangas 1983) is overlain successively by an iron-formation, mafic tholeiites, and a thick succession of turbidites (Morey 1983, Cambray 1987). Schulz (1987a) relates the Animikie Group to initial continental rifting, whereas Hoffman (1987b) suggests that the lower assemblages represent rift and passive-margin deposits, and that the Animikie Group filled a foredeep related to the Penokean orogeny. The foredeep model predicts a synorogenic age for

Animikie volcanism and possible derivation of the Animikie turbidites from the Early Proterozoic arc terrane to the south.

During the Penokean orogeny, the Marquette Range Supergroup underwent early thin-skinned, northward-directed thrusting, followed by upright basement-involved folding of easterly trend (Holst 1982, 1984, Sims et al 1987). The upright folds are typically doubly plunging (Morey et al 1982), and the gross meridional alignment of successive culminations, expressed by elliptical basement domes, is suggestive of transverse cross folding. Peak metamorphism postdates thrusting (Klasner 1978) and ranges in grade from anchizonal to the sillimanite + muscovite zone in pelites of the Marquette Range Supergroup and its correlatives (Morey 1978).

The southern limit of the Marquette Range Supergroup is the 0–12-km-wide Niagara fault zone, a broadly arcuate, convex northward system of fault slices composed of strongly flattened, steeply dipping rocks having dominantly downdip stretching lineations (Larue & Ueng 1985, Sedlock & Larue 1985). Although structurally dismembered, ophiolitic crustal components are present, and they appear to floor the calc-alkaline volcanic rocks occurring to the south (Schulz 1987b). Mafic-intermediate-felsic volcanic rocks of calc-alkaline character and related tonalitic-granitic intrusions occupy a 140-km-wide zone south of the suture (Greenberg & Brown 1983, Schulz 1983). The felsic volcanics are about 1.86 Ga, and the intrusions range in age from 1.86 to 1.82 Ga (Van Schmus 1980, Peterman et al 1985). Deformation attributed to mainly subhorizontal compression under metamorphic conditions ranging from lower-greenschist to upper-amphibolite facies occurred within the same time interval (Maass 1983), perhaps culminating prior to 1.84 Ga (Maass et al 1980). The Penokean magmatic arc terrane, 120–140 km wide, separates Archean rocks continuous with the Superior province from a zone of high-grade Archean (2.8 Ga) rocks in central Wisconsin invaded by 1.89–1.82 Ga tonalites (Van Schmus & Anderson 1977, Maass et al 1980, Anderson & Cullers 1987). The central Wisconsin Archean rocks and the arc terrane are believed to have been accreted to the craton during the Penokean orogeny (Schulz et al 1984).

Following a magmatic hiatus of 60 Myr, a suite of epizonal syenogranites and related rhyolites was emplaced within and south of the exposed Penokean orogen at about 1.76 Ga (Van Schmus 1980, Van Schmus & Bickford 1981). Metaluminous and peraluminous granites and rhyolites of this age exposed in south-central Wisconsin comprise an anorogenic igneous suite associated with dikes of tholeiitic basalt and andesite (Anderson et al 1980, Smith 1983). In the Killarney area northwest of Lake Huron, a wedge-shaped complex of 1.74–1.73 Ga epizonal granites

and related rhyolites truncates Huronian rocks of the Penokean fold belt to the northwest and is truncated in turn by the Grenville orogenic front to the southeast (Davidson 1986, van Breemen & Davidson 1987). Rocks of the Killarney wedge are massive adjacent to the Penokean belt, but deformation increases toward the southeast, producing a steeply dipping, northeast-striking foliation and associated gently plunging stretching lineation. The deformation predates 1.40-Ga pegmatite dikes and may be causally related to the ~1.63-Ga resetting of Rb-Sr isochrons in the Killarney complex, also observed in the 1.76-Ga granites of southern Wisconsin. If the deformation occurred at ~1.63 Ga, it would be coeval with the important Labradorian and Mazatzal orogenies observed in the extreme east and southwest of Laurentia, respectively.

A 1.50–1.42 Ga epizonal granite-rhyolite suite is exposed in the St. Francois Mountains inlier of southeast Missouri and occurs widely in the subsurface of Illinois, Indiana, Missouri, and Kentucky (Anderson 1983, Bickford et al 1986, Van Schmus et al 1987b). This suite is coeval with isolated intrusions to the north, such as the Wolf River batholith exposed in central Wisconsin and the Manitoulin granite in the subsurface of northern Lake Huron (Van Schmus et al 1975a,b). Nd isotopic ratios (Nelson & De Paolo 1985) indicate that the subsurface granite-rhyolite suite was mostly derived from crust having a model age of crust-mantle separation ~1.9 Ga, indicating little or no involvement of Archean crust. In contrast, Penokean granites and younger granites exposed within the Penokean orogen have Nd model ages of ~2.3–2.1 Ga and were probably derived from crustal sources containing mixed Archean and Early Proterozoic components.

Accretion to the Wyoming Province: Yavapai and Mazatzal Orogens

Most of the southwestern United States is underlain by Early Proterozoic crust accreted to the south margin of the Wyoming province between about 1.8 and 1.6 Ga (Figure 10). The main exposures of Early Proterozoic crust are in the Cordilleran front ranges extending from southern Wyoming through Colorado to northern New Mexico, in the ranges of central and southern New Mexico bordering the Rio Grande rift, in the transition zone between the Colorado Plateau and the Basin and Range province of Arizona and adjacent states, and along the San Andreas fault system of southern California (Condie 1981). The correlative Central Plains orogen (Sims & Peterman 1986) of the midcontinent truncates the respective southerly and southwesterly extensions of the Trans-Hudson and Penokean orogens in the subsurface of Nebraska (Van Schmus & Bickford 1981, Arvidson et al 1984, Bickford et al 1986).
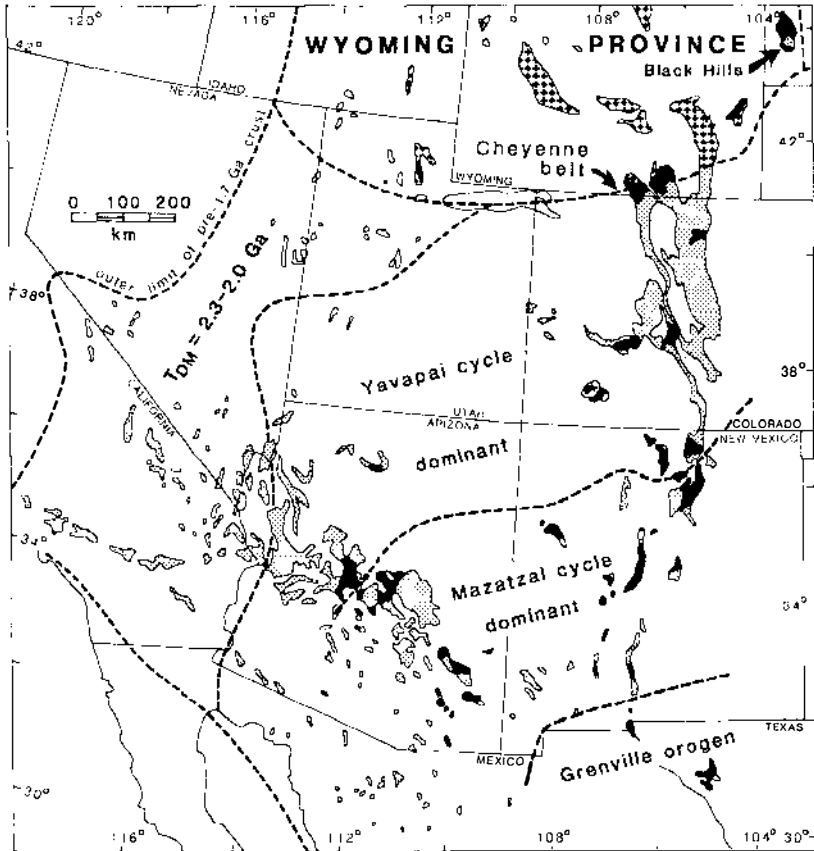
Figure 10 Distribution of Early Proterozoic volcanic and sedimentary rocks (solid) accreted to the Archean Wyoming province, modified from Condie (1986). Plutonic rocks (stippled) include both Early Proterozoic intrusions related to the accreted terranes and Middle Protcrozoic postaccretionary intrusions. Rocks of the Yavapai cycle evolved between 1.79 and 1.69 Ga, and those of the Mazatzal cycle between 1.71 and 1.61 Ga. The distribution of crust having "model" (i.e. crust-mantle separation) ages of 2.3–2.0 Ga is from Bennett & De-Paolo (1987).

Isotopic ratios of Pb, Nd, and Sr indicate that the crust of Arizona, New Mexico, and Colorado has a mantle-separation age of roughly 1.8 Ga, and that little or no Archean crust is present (Zartman 1974, DePaolo 1981, Condie 1982, Stacey & Hedlund 1983, Nelson & DePaolo 1984, 1985, Wooden et al 1987). In contrast, Mesozoic-Tertiary granites in western Utah, southern Nevada, and southern California have Nd model ages of 2.3–2.0 Ga (Farmer & DePaolo 1984, Bennett & DePaolo 1987),

and Pb isotopic data require an Archean component for crustal genesis in this area (Wooden et al 1987). However, the older model-age province, as exposed in southern California, underwent lower-granulite-grade metamorphism at about 1.71 Ga (Wooden et al 1986), essentially coeval with the main orogenic event in Arizona.

Three main stages of Proterozoic crustal development have been recognized from geological relationships and U-Pb zircon dating throughout much of the Arizona-Sonora-New Mexico-Colorado area (Silver 1969, 1984, 1987, Silver et al 1977a). The Yavapai cycle (1.79–1.69 Ga) involved the generation of volcanic-plutonic suites and associated graywacke-pelite facies, interpreted as relics of islands arcs and related sedimentary basins (Anderson 1978, 1987, Condie 1982, 1986). These rocks were consolidated at about 1.70 Ga during an episode of deformation, metamorphism, and plutonism referred to as the Yavapai orogeny (Karlstrom & Bowring 1987). The Mazatzal cycle (1.71–1.62 Ga) is characterized both by subaerial felsic volcanics and shelf-facies arenites that overlie Yavapai-type rocks in central Arizona and northern New Mexico and by turbidites in central Colorado, southeastern Arizona, and southern New Mexico for which no basement is known. These rocks experienced northwest-directed folding and thrusting, followed by plutonism (1.65–1.62 Ga), referred to as the Mazatzal orogeny (Karlstrom & Bowring 1987). The third stage involved the widespread emplacement of "anorogenic"calc-alkaline to alkaline plutons (rapakivi granite, granodiorite, anorthosite, syenite) and related volcanics in two pulses (1.50–1.42 and 1.40–1.34 Ga; Silver et al 1977b, Anderson 1983, Silver 1984, Thomas et al 1984, Bickford et al 1986, Van Schmus et al 1987b). Northwestern Arizona is dominated by rocks of the first stage (Yavapai cycle), whereas southeastern Arizona and southern New Mexico are dominated by rocks of the second stage (Mazatzal cycle). However, the effects of the two orogenic cycles overlap in much of the terrane (e.g. Silver 1984, Karlstrom et al 1987, Karlstrom & Bowring 1987, Reed et al 1987).

Although the general character of the Arizona-New Mexico-Colorado terrane and its boundary with the Wyoming province are reasonably well understood, the specific details of the tectonic evolution remain controversial because of their intrinsic complexity and the discontinuous exposure of the terrane. One scenario involves the initial collision of an island arc with the Wyoming province as a consequence of southeast-dipping subduction, followed by arc-polarity reversal and progressive southward accretion of an Andean margin above a northwest-dipping subduction zone (Hills & Houston 1979, Karlstrom & Houston 1984, Anderson 1986, Condie et al 1987, Reed & Premo 1987). The role of cyclic back-arc basins during accretion is stressed by Condie (1982, 1986),

particularly for the sediment-dominated zone in central Colorado (Condie et al 1987, Reed et al 1987). In contrast, the sediment-dominated zone of southeastern Arizona and southern New Mexico is interpreted as a prograded trench-forearc complex (Anderson 1986, Swift 1987). Unlike other modelers, who invoke progressive southward accretion, Karlstrom & Bowring (1987) view central Arizona as having been assembled by nonsystematic juxtaposition of crustal blocks having disparate histories along strike-slip and dip-slip shear zones of various ages.

WYOMING   The Cheyenne belt forms the boundary between the Wyoming province and the accreted terranes to the south, and it is exposed in the Medicine Bow Mountains and Sierra Madre of southern Wyoming (Karlstrom et al 1983, Karlstrom & Houston 1984, Duebendorfer & Houston 1987). North of the boundary, an Early Proterozoic passive-margin sedimentary prism (Snowy Pass Supergroup) is exposed in two south- to southeast-dipping thrust sheets and an underlying autochthon floored by Archean basement. The sediments were intruded at about 2.0 Ga by mafic sills and dikes that predate thrusting. The autochthonous cover sequence comprises quartzite, conglomerate, phyllite, and minor marble of the Deep Lake Group. The allochthonous Libby Creek Group consists of a lower thrust sheet of quartzite, diamictite, and phyllite, and an upper thrust sheet in which stromatolitic dolomite, nearly 2 km thick, is overlain by mafic flows(?) and laminated ferruginous slate. The slate may mark the onset of foredeep sedimentation, related to abortive subduction of the margin beneath the arc terranes to the south (Hills & Houston 1979).

The boundary zone is best exposed in the Medicine Bow Mountains, where it consists of a pair of steeply dipping northeast-trending mylonite zones, between which granitic orthogneiss and amphibolite of unknown ages separate the Snowy Pass Supergroup to the north from an Early Proterozoic metamorphic-plutonic assemblage to the south. The latter includes metamorphosed graywacke, pelite, intermediate-mafic volcanics, peridotite, and 1.78–1.76 Ga dioritic to granitic plutons. The mylonite zones record two components of simple shear: a dominant synmetamorphic southeast-side-up dip-slip component, and a subordinate dextral strike-slip component. The dominant downdip stretching lineation resulted from northwest-directed thrusting on planes that were subsequently steepened. This is consistent with the observed increase in metamorphic grade to the southeast across the boundary, including inverted metamorphic zonation in the southeast-facing upper Snowy Pass Supergroup (Duebendorfer 1987).

COLORADO   South of the Cheyenne belt, metavolcanic rocks are exposed in southern Wyoming (Condie & Shadel 1984) and southern Colorado

(Bickford & Boardman 1984, Boardman 1986, Boardman & Condie 1986), but the intervening area is characterized by turbiditic metasediments and derived paragneiss. In southern Wyoming, felsic volcanism occurred at 1.79 Ga, plutonism at 1.78–1.74 Ga, and the peak of metamorphism and deformation at about 1.77 Ga (Reed et al 1987). In south-central Colorado, calc-alkaline volcanism at 1.77–1.76 Ga and syntectonic plutonism at 1.76–1.75 Ga were followed by felsic volcanism and plutonism at 1.74–1.70 Ga and a major deformation at about 1.71 Ga (Reed et al 1987). In south-western Colorado, 1.79–1.73 Ga volcanic and plutonic rocks were deformed and metamorphosed (Yavapai orogeny) prior to intrusion of 1.69-Ga plutons, collectively forming a basement complex to quartzites and pelites (Uncompahgre Group) that were deformed by NNW-directed thin-skinned thrusting and subsequent ENE-trending thick-skinned folding during the Mazatzal orogeny (Gibson et al 1987). In the sediment-dominated zone of northern and central Colorado, syntectonic plutons range from 1.75 to 1.67 Ga and major deformation occurred at 1.68–1.67 Ga (Reed et al 1987). The current interpretation of these relations (Condie et al 1987, Reed & Premo 1987) is that the northern volcanic belt represents an island arc that collided with the Wyoming province at about 1.77 Ga, that the southern volcanic belt represents a continental-margin arc developed above a northwest-dipping subduction zone immediately following the arc-continent collision, and that the dominantly meta-sedimentary zone represents a composite back-arc basin that closed as a result of collision(s) farther south.

ARIZONA   Both the Yavapai and Mazatzal cycles are represented in central Arizona and northern New Mexico, but relations between them are uncertain. A northeast-trending boundary that angles midway between Flagstaff and Phoenix in Arizona separates a northwestern domain dominated by the older cycle from a southeastern domain dominated by the younger cycle at the present erosion level (Silver & Ludwig 1986, Karlstrom & Conway 1986, Karlstrom et al 1987, Karlstrom & Bowring 1987). The boundary fault is a relatively late high-angle structure (the southeast side having moved down), but Karlstrom et al (1987) propose that the fundamental boundary is a subhorizontal surface on which rocks of the southeastern domain were thrust over unrelated rocks to the northwest during the Mazatzal orogeny. Thin-skinned thrusting is well documented, and their evidence for tectonic juxtaposition of unrelated terranes is that intense ductile compressional deformation and contemporaneous emplacement of peraluminous granite at a depth of at least 8 km were occurring in the northwestern domain at 1.70 Ga (Yavapai orogeny) while, at the same time, alkaline rhyolite was being erupted on little-deformed Yavapai-

type rocks in the southeastern domain. Thus, peraluminous magmatism accompanied by strong horizontal compression in the northwest occurred simultaneously with alkaline magmatism unaccompanied by deformation in the southeast. Karlstrom et al (1987) suggest that the one outlier of rhyolite and quartzite representing the Mazatzal cycle overlying deformed Yavapai rocks in the northwestern domain may be a klippe emplaced by thin-skinned thrusting. The interpretation of central Arizona as being composed of "accreted terranes" is extended by Karlstrom & Bowring (1987) to include three composite terranes of contrasting history, super-posed along northwest-directed subhorizontal thrusts, and subsequently chopped up by late high-angle strike-slip and dip-slip faults.

NEW MEXICO   Northern New Mexico is similar to the southeastern terrane of central Arizona in that mafic-felsic arc-type volcanics and immature sediments, intruded by plutons, of the Yavapai cycle were not tightly folded prior to deposition of shelf quartzite (Ortega Group) of the Mazatzal cycle (Bowring & Condie 1982, Grambling & Codding 1982, Bowring et al 1984a, Silver 1984, Soegaard & Eriksson 1986). Unlike the thin-skinned thrust-fold belt of central Arizona, the Mazatzal orogeny in northern New Mexico was characterized by progressive ductile shearing that produced large-scale polyphase recumbent folds (Grambling & Codding 1982, Hol-combe & Callender 1982, Williams & Grambling 1987). Deformation accompanied prograde metamorphism that peaked after folding at region-ally uniform conditions near the aluminosilicate triple point and was followed by slow isobaric cooling (Grambling 1986).

In southern New Mexico and southeasternmost Arizona, the Mazatzal cycle is represented by turbiditic metasediments (Pinal schist), possibly correlative with shelf-facies quartzite and conglomerate to the northwest; by minor mafic and felsic metavolcanic rocks; and by many 1.65-Ga and younger plutons (Silver 1978, 1987, Condie & Budding 1979, Bowring et al 1983, Copeland & Condie 1986, Conway & Silver 1987). The turbidites are interpreted as deep-sea or trench deposits incorporated into a south-east-facing accretionary prism, based on sedimentological and structural evidence (Swift 1987), in accord with the arc-trench progradation model of Anderson (1986) for the Mazatzal terrane. However, this model implies that synsedimentary shearing and thrusting should be southeast-directed, in contrast to the northwest-directed deformation generally observed in the Mazatzal orogen.

The southwestern United States and the central Trans-Hudson orogen of Manitoba and Saskatchewan are important as the only parts of Lau-rentia where juvenile Early Proterozoic crust is extensively exposed through a significant range of crustal depths and metamorphic grades.

They appear to have much in common with the Archean ("granite-green-stone-gneiss") provinces of Laurentia, aside from the obvious differences in age and certain petrochemical details.

DISCUSSION

Archean crustal provinces are clustered in the northern part of Laurentia; its southern part is underlain by juvenile Early Proterozoic crust. The Canadian shield is not representative of the craton as a whole, but rather is biased in favor of Archean crust. The Early Proterozoic crust is preferentially hidden beneath Phanerozoic sedimentary cover. The narrowest part of the ring-shield enclosing the Hudson Bay basin is underlain by Early Proterozoic crust of the Trans-Hudson orogen. As the shield is believed to have been an area of positive relief through Phanerozoic time, the question arises as to why the Archean lithosphere is more buoyant than Early Proterozoic lithosphere. As depleted mantle has a lower density than fertile mantle due to its lower garnet content (Oxburgh & Parmentier 1978), the relative buoyancy of the Archean provinces may reflect a relatively depleted underlying mantle lithosphere, as inferred from isotopic ratios in carbonatites intruding the Superior province (Bell & Blenkinsop 1987). The presumed secular decline in mean temperature of the asthenosphere implies a greater depth and volume of melting accompanying mantle upwelling in the Archean, and consequently a thicker, more depleted mantle lithosphere (Sleep & Windley 1982, Bickle 1986). Various mechanisms have been discussed whereby such a depleted mantle "tectosphere" might develop beneath the Archean cratons (e.g. Oxburgh & Parmentier 1978, McKenzie 1984, Pollack 1986, MacGregor & Manton 1986, Kramers 1987).

Laurentia contains at least six Archean provinces sutured along Early Proterozoic orogenic belts. The pattern of intersecting orogenic belts bounding provinces that behaved as relatively stable platforms during the Early Proterozoic is comparable to composite Phanerozoic continents such as Eurasia, formed by amalgamation of microcontinents (many of common Gondwanaland ancestry) and intervening island arcs. Interpretation of the Early Proterozoic orogenic belts as collision zones resulting from subduction of oceanic lithosphere is consistent with their sedimentary, structural, metamorphic, and magmatic asymmetry. This is well displayed in the Thelon, Wopmay, Trans-Hudson, New Quebec, Torngat, Penokean, and Cheyenne belts, all of which feature foredeep basins, thrust-fold belts, sparse mafic magmatism, and relatively low metamorphic grades in their forelands, contrasted with relatively high metamorphic grades, major calc-alkaline magmatic belts, and complex patterns of ductile defor-

mation in their hinterlands. Their forelands were dominated by thrusting; their hinterlands by transcurrent shearing accompanying arc magmatism, a consequence of oblique subduction (Fitch 1972), and manifestations of "propagating extrusion tectonics" (Tapponnier et al 1982) in response to continent-continent indentation.

Phanerozoic orogens represent a spectrum of cases between "accretionary" and "collisional" end members, and the same is true for the Early Proterozoic orogens of Laurentia. The orogens of southern Laurentia, the Svecofennian orogen, and possibly the Wopmay orogen involved the accretion of broad zones of juvenile Early Proterozoic crust (Patchett & Arndt 1986). In the Trans-Hudson orogen, a 350-km-wide central zone of juvenile Early Proterozoic crust is trapped between opposing Archean provinces. The Thelon, Snowbird, New Quebec, and Torngat orogens are characterized by the "tight" suturing of adjacent Archean provinces, without the preservation of intervening intraoceanic relics. Differences between Early Proterozoic and Phanerozoic orogens probably exist, but generalizations are premature given the great variability in history, style, and erosion level among orogens of either age.

The amalgamation of the Archean provinces of Laurentia took place in only about 150 Myr. The times of collision, based on ages of foredeep sedimentation and/or cessation of arc-type magmatism, are estimated to be about 1.96 Ga for the Thelon orogen, between 1.92 and 1.85 Ga for the Snowbird zone, between about 1.85 and 1.83 Ga for the Trans-Hudson orogen, and between 1.87 and 1.81 Ga for the New Quebec orogen. Accretion in the Wopmay orogen began about 1.91 Ga, in the Penokean orogen about 1.85 Ga, in the Makkovik orogen about 1.81 Ga, and in the Cheyenne belt about 1.75 Ga. Thus, the inter-Archean collision events occurred between about 1.96 and 1.81 Ga, and accretion of intraoceanic terranes is documented between about 1.91 and 1.63 Ga. The limited time span required to assemble most of Laurentia is the single most startling conclusion to emerge from this synthesis.

Many important problems remain. Did the Archean provinces originate independently, or were they products of the breakup of one or more large continents? Why is crust formed between 2.5 and 2.0 Ga apparently of limited distribution (e.g. northwest Africa, northeast South America, North American Cordillera, and possibly western Wopmay orogen)? What is the significance of the apparent progressive southward growth of Laurentia and its Baltic extension? How large did Laurentia ultimately become, and are its extensions, removed by post-1.6-Ga rifting, to be found on other continents? What is the origin and significance of the extensive Middle Proterozoic "anorogenic" magmatism? Was the discordant trend of the 4000-km-long Grenville orogen dictated by Middle Proterozoic

rifting? Where is the Grenville hinterland? Is North America a fragment of Middle and/or Late Proterozoic supercontinent(s) rifted at about 0.6 Ga? Such questions demand a global synthesis of the extant Precambrian lithosphere.

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