Tectonic model for the Proterozoic growth of North America

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ABSTRACT

This paper presents a plate-scale model for the Precambrian growth and evolution of the North American continent. The core of the North American continent (Canadian shield) came together in the Paleoproterozoic (2.0–1.8 Ga) by plate collisions of Archean continents (Slave with Rae-Hearne, then Rae-Hearne with Superior) as well as smaller Archean continental fragments (Wyoming, Medicine Hat, Sask, Marshfield, Nain cratons). The resulting Trans-Hudson orogen was a collisional belt similar in scale to the modern Himalayas. It contains mainly reworked Archean crust, but remnants of juvenile volcanic belts are preserved between Archean masses. The thick, buoyant, and compositionally depleted mantle lithosphere that now underlies North America, although dominantly of Archean age, took its present shape by processes of collisional orogenesis and likely has a scale of mantle heterogeneity similar to that exhibited in the overlying crust.

In marked contrast, lithosphere of southern North America (much of the continental United States) was built by progressive addition of a series of dominantly juvenile volcanic arcs and oceanic terranes accreted along a long-lived southern (present coordinates) plate margin. Early juvenile additions (Pembine-Wausau, Elves Chasm arcs) formed at the same time (1.84–1.82 Ga) the core was assembling. Following final assembly of the Archean and Paleoproterozoic core of North America by 1.8 Ga, major accretionary provinces (defined mainly by isotopic model ages) were added by arc-continent accretion, analogous to present-day convergence between Australia and Indonesia. Also similar to Indonesia, some accreted terranes contain older continental crustal material [Archean(?) Mojavia], but the extent and geometry of older crust are not well known. Accretionary provinces are composed of numerous 10 to 100 km scale terranes or blocks, separated by shear zones, some of which had compound histories as terrane sutures and later crustal-assembly structures. Major northeast-trending provinces are the Yavapai province (1.80–1.70 Ga), welded to North America during the 1.71–1.68 Ga Yavapai orogeny; the Mazatzal province (1.70–1.65 Ga), added during the 1.65–1.60 Ga Mazatzal orogeny; the Granite-Rhyolite province (1.50–1.30 Ga), added during the 1.45–1.30 Ga tectonic event associated with A-type intracratonic magmatism; and the Llano-Grenville province (1.30–1.00 Ga), added during the 1.30–0.95 Ga broader Grenville orogeny. During each episode of addition of juvenile lithosphere, the transformation of juvenile crust into stable continental lithosphere was facilitated by voluminous granitoid plutonism that stitched new and existing orogenic boundaries. Slab roll back created transient extensional basins (1.70 and 1.65 Ga) in which Paleoproterozoic quartzite-rhyolite successions were deposited, then thrust imbricated as basins were inverted. The lithospheric collage that formed from dominantly juvenile terrane accretion and stabilization (1.8–1.0 Ga) makes up about half of the present-day North American continent. Throughout (and as a result of) this long-lived convergent cycle, mantle lithosphere below the accretionary provinces was more hydrous, fertile, and relatively weak compared to mantle lithosphere under the Archean core.

Keywords: Proterozoic, Rodinia, Laurentia, continent assembly, North America

INTRODUCTION AND SCOPE

The formation and long-term behavior of continental lithosphere requires a plate-scale, time-integrated understanding of crust and mantle formation and modification events within single continents, in association with continued attempts to reconstruct the cycle of supercontinent formation and fragmentation in the Precambrian. North America offers among the most complete geologic, geophysical, and isotopic data sets of any continent and so is an important case study for continental evolution. It was also centrally located in the Precambrian supercontinents of Nuna (1.8–1.6 Ga) and Rodinia (1.1–0.9 Ga); therefore, improved understanding of the evolution of North America needs to be directly linked with studies of past supercontinent reconstructions (e.g., Li et al., 2007).

The purpose of this paper is to present a visual model for the development of the Precambrian core of North America (i.e., Laurentia) via a series of time-slice maps (Figs. 1–20) and animations (Animations 1 and 2[1]). These visualizations may be useful both for nongeology audiences and for introductory geology teaching as a graphic display of plate tectonic models for continental growth and the time-integrated record preserved in continents. For more advanced audiences, we hope this model will stimulate critical debate about the tectonic evolution of North America, resulting plate-scale heterogeneity, processes that shape and modify continents, and piercing points that can be used to match the margins of ancestral North America (Laurentia) to neighboring continental margins of past supercontinents. We recognize that any model at the scale of a whole continent and over 1 b.y. of history is necessarily simplified, based on variable quality data, and limited by the biases and incomplete knowledge of the authors.

METHODS

This compilation of pre-Neoproterozoic tectonic belts in southern Laurentia is based primarily on geologic and geochronologic data from exposed Proterozoic outcrops (~10% of

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Geosphere; August 2007; v. 3; no. 4; p. 220–259; doi: 10.1130/GES00055.1; 20 figures; 2 animations.

Animation 1. Powerpoint sequence of the sequential assembly and growth of Laurentia from ca. 2.0 Ga through ca. 0.535 Ga. Powerpoint slides follow the order of Figures 2–20 in the text.

Animation 2. Quicktime movie of the assembly and growth of Laurentia from ca. 2.0 Ga through ca. 0.535 Ga. Original positions of Archean terranes are unconstrained; assembly of the Laurentian core is shown relative to the present-day positions of the Rae and Hearne provinces.

the continental U.S.; Reed and Harrison, 1993; Reed, 1993), and also includes data from drill holes in the mid-continent (Van Schmus et al., 1996, 2007). Extrapolation of contacts and structural trends into areas covered by Phanerozoic sedimentary rocks (~90% of the continental U.S.) is aided by interpretation of aeromagnetic data (North American Magnetic Anomaly Group, 2002). Aeromagnetic maps have been used in the interpretation of basement lithotypes and structure, including the western United States (e.g., Lidiak, 1974; Ramberg and Smithson, 1975; Cordell and Grauch, 1985; Finn et al., 2001; Sims and Stein, 2003; Grauch et al., 2003; Finn and Sims, 2005), Canada (e.g., Pilkington et al., 2000; Ross et al., 1991), Australia (e.g., Gunn et al., 1997), and elsewhere (e.g., Henkel, 1991).

Aeromagnetic anomalies primarily reflect variations in magnetization properties of Proterozoic crystalline basement (Grauch et al., 2003). Magnetic intensities reflect distribution of magnetite and other iron-bearing minerals in the crust. Sedimentary strata generally have little magnetic character and are transparent in regional aeromagnetic mapping (Finn and Sims, 2005). Large Cenozoic volcanic fields can obscure magnetic patterns in the basement, but for much of the western United States, Cenozoic volcanics tend to have short-wavelength anomalies related to known vent areas and flows such that the generally more persistent basement magnetic fabric is still decipherable. In areas where there are outcrops (e.g., Rocky Mountains), granite and granodiorite plutons often are characterized by relatively high amplitude magnetic highs (20–200 nT), and this character has been used to extrapolate the contacts of plutons into covered areas (Karlstrom et al., 2004; Finn and Sims, 2005). In the southern Rocky Mountains, long-wavelength (>50 km), high-amplitude (>500 nT above base values) magnetic anomalies commonly correspond to 1.4 Ga plutons (Finn and Sims, 2005). In this paper, this has been invoked for interpretations of the distribution and large areal extent of granitoid plutons in the mid-continent (Fig. 1).

Caveats to this approach of correlating prominent aeromagnetic highs with granitoid plutons are numerous. For example, while magnetite-bearing granitoids are common and are strongly magnetic, two-mica peraluminous granitoids are moderately magnetic and may not be uniformly magnetized; ilmenite-bearing granitoids tend to be nonmagnetic; and some granitoids form strong aeromagnetic lows (e.g., the 1.1 Ga Pikes Peak pluton of Colorado), presumably due to the orientation of the remnant magnetization. Plutons of different ages can have similar signatures (Karlstrom et al., 2004), and so the age assignments herein are largely conjectural in areas far from outcrops. Nevertheless, our interpretations of voluminous stitching plutons are consistent with the shapes, surface areas, and scale of heterogeneity of granitoids as mapped in areas of good outcrops, and compatible with a regional magnetic potential map that shows magnetic highs corresponding to areas of voluminous 1.4 Ga granite and/or rhyolite magmatism (Finn and Sims, 2005). Thus, we believe that the maps show realistic portrayals of the tectonic grain and nature of heterogeneity, and perhaps even a crude estimate of the percentage of granitoid plutons versus country rock of the subsedimentary basement across the continental United States. However, nearly all contacts on these map reconstructions have some degree of uncertainty, and we have not tried to portray the variable levels of uncertainty.

Crust of rifted margins tends to be thinned and modified during extension, after which thinned crustal blocks can become incorporated and overprinted as they are reaccreted to convergent plate margins. Therefore we rely heavily on Nd and Pb isotopic data to indicate an average crustal model age for a given province. When model ages are within 10–100 m.y. of U-Pb crystallization ages, we refer to the terranes as juvenile. However, when model ages are significantly older than crystallization ages (e.g., for portions of the Trans-Hudson orogen, Mojave province, and the pre-Appalachian rifted margin), the model ages commonly represent a mixture of sources with different mantle separation ages that can be interpreted in different ways. (1) They may represent the presence of older crustal blocks in the subsurface that were sampled by magmas during later partial melting in the crust and/or tectonically imbricated with younger crust. (2) They may represent detritus from older terranes that become mixed in with a younger terrane in a proportion to produce a mixed model age. While we attempt to distinguish areas of juvenile crust of a given age from areas of reworked older crust (reworked in the sense of 1 and/or 2 above), the mixed model ages do not yield unique tectonic interpretations. In modern analogs such as the Indonesian region, all types of mixing of crustal ages are common: older crustal fragments form the foundation of portions of younger island arcs (e.g., under Sumatra but not Java), tectonic imbrication is common, and older detritus can be transported long distances (thousands of kilometers) in arc-trench systems (Hamilton, 1979).

The following sections discuss and illustrate, in chronological order, our model for the key tectonic events and provinces that were involved in the sequential growth of Laurentia during the Proterozoic.

ASSEMBLY OF THE LAURENTIAN SHIELD: ARCHEAN PROVINCES AND EARLIEST PALEOPROTEROZOIC JUVENILE CRUST, 2.4–2.0 Ga

The assembly history of the core of the Laurentian craton was summarized in numerous papers (e.g., Hoffman, 1988; Corrigan et al.,

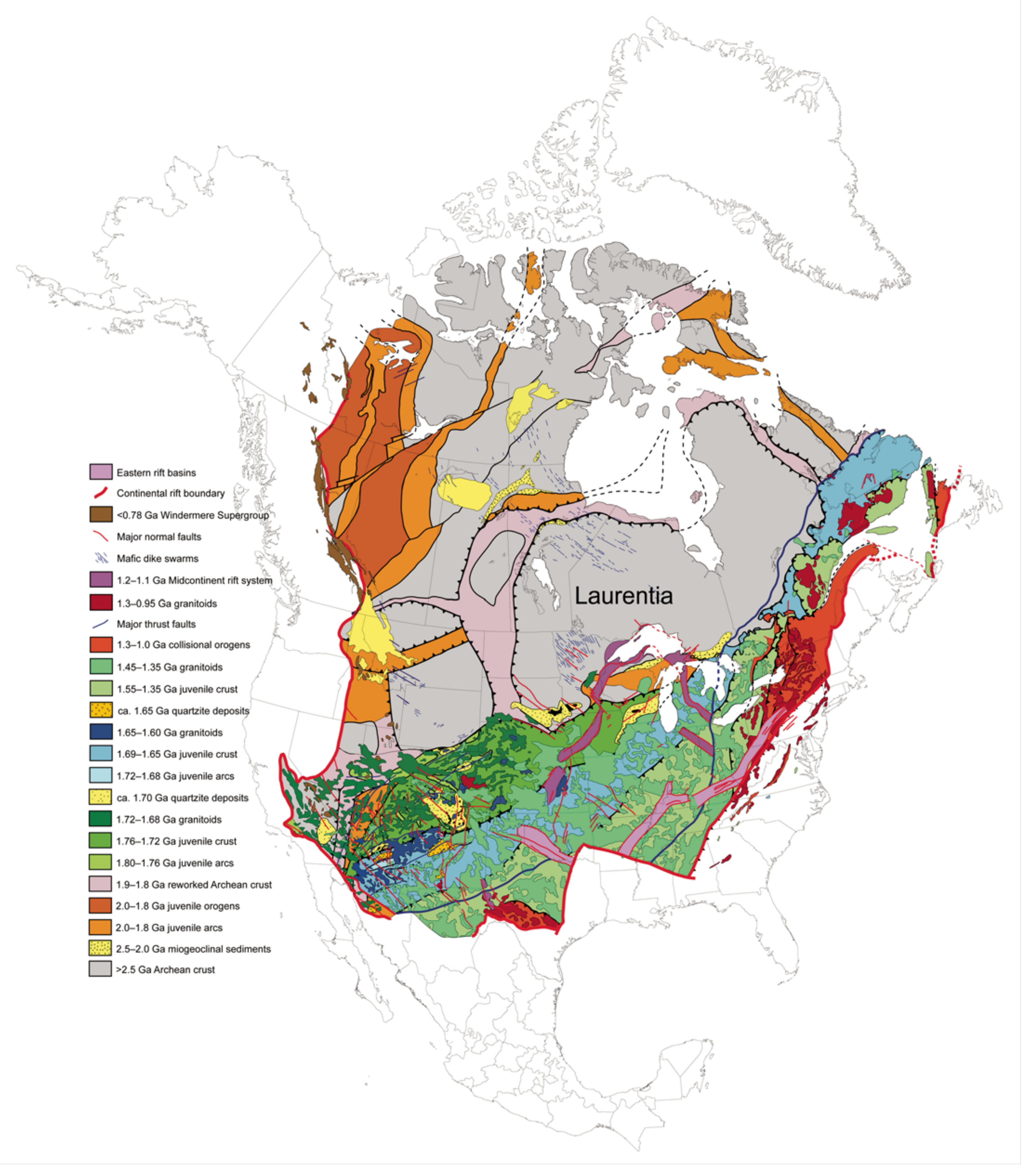

Figure 1. Depiction of Archean through Neoproterozoic basement features of Precambrian North America (Laurentia). Significant terranes, orogenic belts, basins, rifts, and structural features are highlighted by individual colors. Background outline of North American states and provinces based on Geological Society of America Decade of North American Geology (DNAG) spherical Transverse Mercator projection (centered on 100°W meridian; Snyder, 1987); initial work on the U.S. portion of this map based on Reed (1993).

2005; Percival et al., 2004). Figures 2–8 show a sequence of Paleoproterozoic collisional events between Archean continental fragments that resulted in assembly of a continental core of a scale compatible with modern continents. The fragments record a long history of Archean tectonism (e.g., Bleeker, 2003; Percival, 2004), after which many underwent 2.45–2.1 Ga rifting, as shown by thick miogeoclinal successions on the edges of Archean cratons (Corrigan et al., 2005; Fig. 2). The precollisional (pre–1.9 Ga) paleogeographies of these Archean cratons, for example their orientations and the separation between, are poorly known, and so the relative positions of the cratons in Figures 2–8 are not well defined. However, in the terms of Hoffman (1988), Paleoproterozoic collisional amalgamation of these cratonic blocks resulted in the birth of the North American craton. These continent-continent collisional events occurred diachronously from ca. 1.96 to 1.83 Ga for Slave-Rae-Hearne assembly (Figs. 2–4) and from 1.83 to 1.80 Ga for the collision of the Superior craton (Figs. 6 and 7), comparable in spatial and temporal scale with the >50 m.y. of collision between the Indian and Asian continents that formed the modern Himalayas. At least some of the Paleoproterozoic collisions probably consumed large ocean basins (Corrigan et al., 2005; Bleeker, 2003; Bleeker and Ernst, 2006), and the remnants of 1.96–1.80 Ga juvenile volcanic arc belts left between the colliding Archean blocks likely reflect only a small fraction of the oceanic material caught between the continents. This is because juvenile arcs and oceanic fragments may have tended to be thrust at higher crustal levels over the stronger colliding Archean continents, and only small volumes are left at the present middle crustal levels of exposure (Corrigan et al., 2005). Large areas of 1.9–1.8 Ga reworked Archean crust in the Trans-Hudson orogens (Figs. 6 and 7) are interpreted to reflect both buried Archean crustal elements (like the Sask craton; Bickford et al., 2005; Ansdell, 2005) and eroded Archean materials that were tectonically reworked and isotopically mixed with materials from the juvenile belts.

Archean terranes in southern Laurentia represent extensions of more northern Canadian provinces within the Laurentia core. Examples include the Minnesota River Valley and Wyoming provinces. The Minnesota River Valley province is a southward continuation of a composite Superior province (Bickford and Van Schmus, 1985; Schmitz et al., 2006; Bickford et al., 2006). The Wyoming province has been interpreted to be (1) a complex southward continuation of the Hearn province (Ross, 2002); (2) a separate Archean microcontinent (Foster et al., 2006); or (3) a fragment rifted off the southern Superior province, rotated 180° and translated westward, based on the similarities between 2.4 and 2.1 Ga mafic dikes as well as miogeosynclinal packages of the Huronian and Snowy Pass Supergroups (Roscoe and Card, 1993; Harlan et al., 2003a; Dahl et al., 2006). We do not address this early history in detail, but show Wyoming (Fig. 7) to have collided with the growing core of collisional Archean fragments by 1.80 Ga (Mueller and Frost, 2006; Foster et al., 2006).

The Archean Wyoming province consists of an older core of 3.6–3.0 Ga gneisses (some containing 4.0 Ga zircons) rimmed by, and variably reworked by, several younger roughly concentric magmatic and/or tectonic belts (Houston, 1993). Chamberlain et al. (2003) divided the Archean into a series of subprovinces, including a Late Archean to Middle Archean core rimmed by a southward-younging series of arc-related magmatic additions, deformational belts, and Late Archean supracrustal rocks, that were added to the Wyoming craton ca. 2.90–2.55 Ga. In the southernmost Wyoming province, basement older than 2.7–2.5 Ga is overlain by the 2.4–2.1 Ga miogeoclinal Snowy Pass Supergroup, which represents rifting of the Archean craton (Fig. 2; Karlstrom et al., 1983; Karlstrom and Houston, 1984). Additional evidence for early Paleoproterozoic rifting is the Kennedy dike swarm, a set of northeast-striking 2.01 Ga mafic to ultramafic dikes that parallels the Cheyenne belt (Cox et al., 2000). The Snowy Pass Supergroup is similar in age, thickness, and stratigraphic sequence to the Huronian Supergroup (Roscoe and Card, 1993); this has led to two viable alternate tectonic models. Houston (1993) proposed that both miogeoclinal successions formed on south-facing rift margins at 2.4–2.0 Ga along an already amalgamated Archean nucleus. This would negate models for closure of large oceans across the Trans-Hudson belts between Superior and Wyoming cratons. Alternatively, Roscoe and Card (1993) proposed that the Wyoming and Superior passive margins were conjugate, facing margins that rifted away from each other, then rotated 180° so that they are now both south facing. This interpretation allows for large continental translations between 2.4 and 2.0 Ga rifting and 1.9–1.8 Ga collisions across the Trans-Hudson region.

The Grouse Creek block (Foster et al., 2006; Figs. 2–7) contains Archean rocks of the Grouse Creek, Albion, and East Humboldt Ranges. It could be a separate Archean fragment from the Wyoming province. This is compatible with Pb and Nd data from Mesozoic and Cenozoic rocks of the Snake River Plain that have Archean model ages and with the observation that Paleoproterozoic rocks are in the Wasatch Range in the area between the Grouse Creek and Wyoming blocks (Foster et al., 2006).

Juvenile 2.4–2.0 Ga crust is present in western North America in several locations. The Wopmay orogen contains 1.9 Ga rocks that have Nd ratios suggesting the involvement of 2.4–2.0 Ga crust (Figs. 3–5; Bowring and Podosek, 1989). U-Pb dating of drill cores in the Alberta basement of Canada and work in the western Rae province have also revealed appreciable crust of 2.4–2.0 Ga in western Canada (Ross, 2002). In this case, the main terranes, Buffalo Head and Chinchaga, are interpreted as arcs built on Archean crust, in contrast to similar-aged juvenile arc crust in the Wopmay orogen to the north (the accreted Hottah terrane). Hedge et al. (1983) and Bryant (1988) reported 2.0 Ga dates from Antelope Island in the Great Salt Lake in northern Utah based on two-point chords of variably discordant data from large multigrain zircon suites; these dates may represent mixtures of 2.4–2.0 Archean and 1.8–1.6 Ga Proterozoic rocks. Cavosie and Selverstone (2003) reported 1.9 Ga oceanic crust preserved in Colorado, but this was based on a single high-temperature step in a complex $^{40}Ar/^{39}Ar$ release spectrum on hornblendes that gave an integrated age of 1.4–1.5 Ga, and should not be viewed as a reliable primary age for these rocks.

TRANS-HUDSON OROGEN: REWORKED ARCHEAN CRUST WITH BELTS OF 1.9–1.8 Ga JUVENILE CRUST

The Trans-Hudson orogen represents the 1.85–1.78 Ga amalgamation of the Hearne, Wyoming, and Superior cratons into the cratonic core of Laurentia (Figs. 4–7; Hoffman, 1988; Ross and Villeneuve, 2003). Figures 5 and 6 show the final indentor-type collision analogous to the India-Asia Himalayan collision. Geochronologic (Bickford et al., 2005) and seismic (Lucas et al., 1993, 1994) work has identified a small Archean craton, the Sask block (Ansdell et al., 1995), within the Trans-Hudson juvenile belts in Saskatchewan and Manitoba. Preservation of Trans-Hudson juvenile arcs was likely facilitated by entrapment of Sask crust between the Hearne and Superior cratons during the 1.9–1.8 Ga suturing event (Hajnal et al., 2005; Corrigan et al., 2005). Associated juvenile belts within the Laurentian shield include the Foxe belt in the northeastern region of the Rae province, the La Ronge and Flin Flon arcs south of the Hearne province, the Natsajuaq arc on the northern tip of the Superior province, and the Cape Smith and New Quebec orogens along the eastern margins of the Rae, Superior, and perhaps Nain provinces (Hoffman, 1988; Ross and Villeneuve, 2003; Figs. 4 and 5).

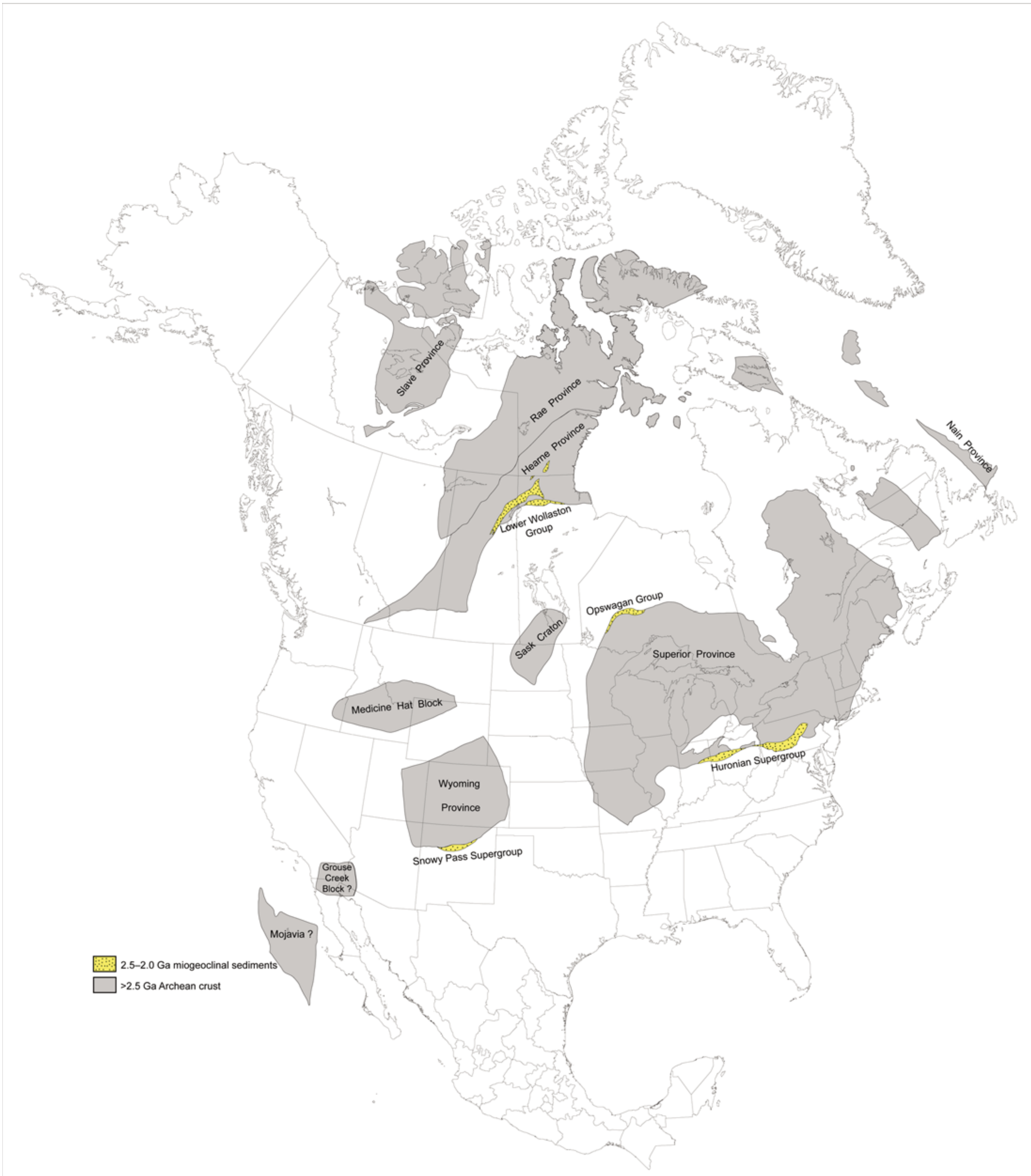

Figure 2. Archean cratons (gray) of the Canadian shield and northwestern United States, including Slave, Rae, Hearne, Superior, Nain, and Wyoming provinces, Sask craton, Medicine Hat and Grouse Creek(?) blocks, and possibly Mojavia. Depicted positions of cratons (older than ca. 2.0 Ga) are unconstrained; the Rae and Hearne provinces are shown in present-day positions. Miogeoclinal sediments (ca. 2.5–2.0 Ga) of the Lower Wollaston and Opswagan Groups and the Snowy Pass and Huronian Supergroups are indicated in stippled yellow.

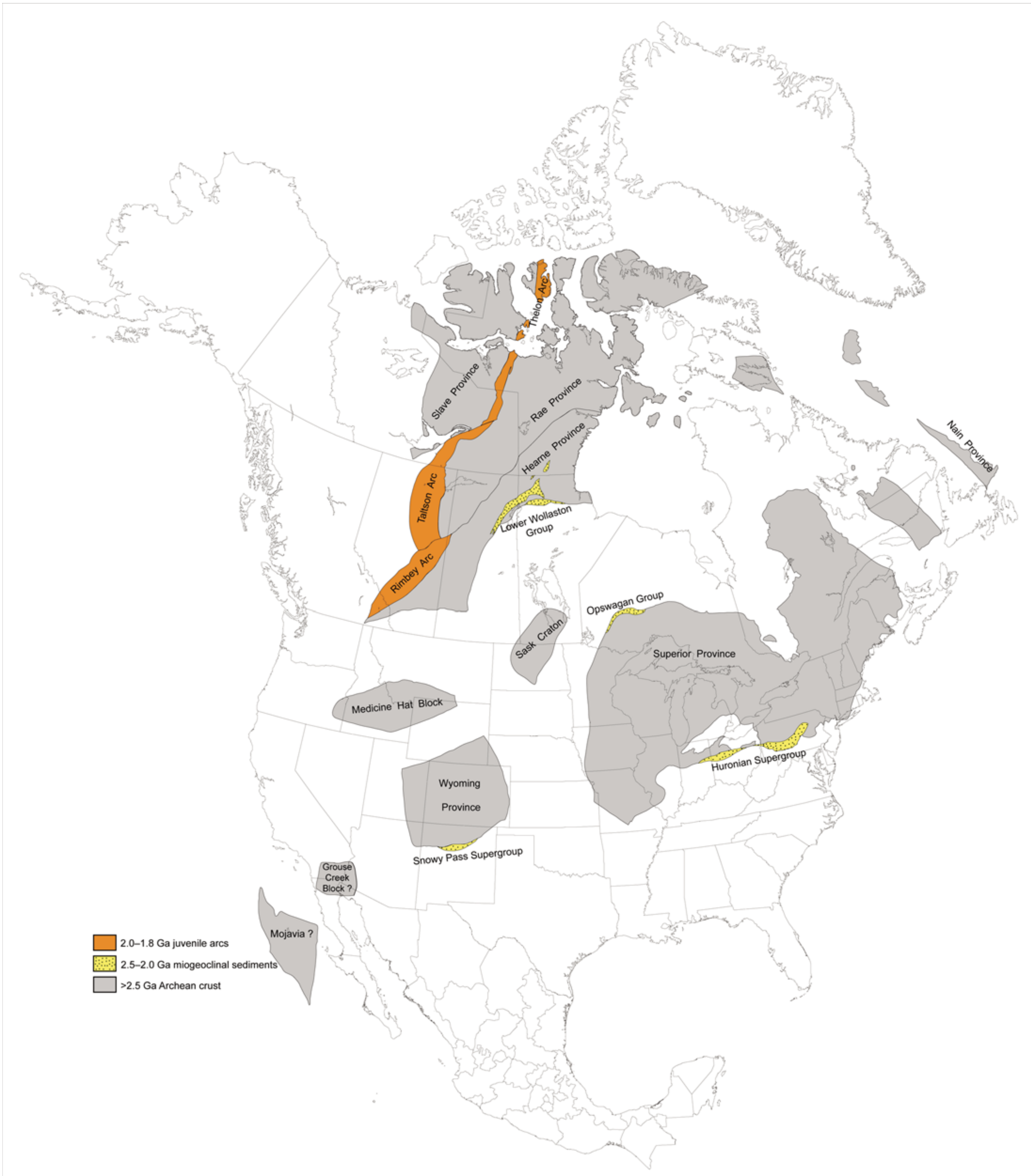

Figure 3. Earliest (ca. 1.96–1.92) juvenile arcs (orange) developed along Slave-Rae collisional zone (Thelon arc) and western margin of Rae and Hearne provinces.

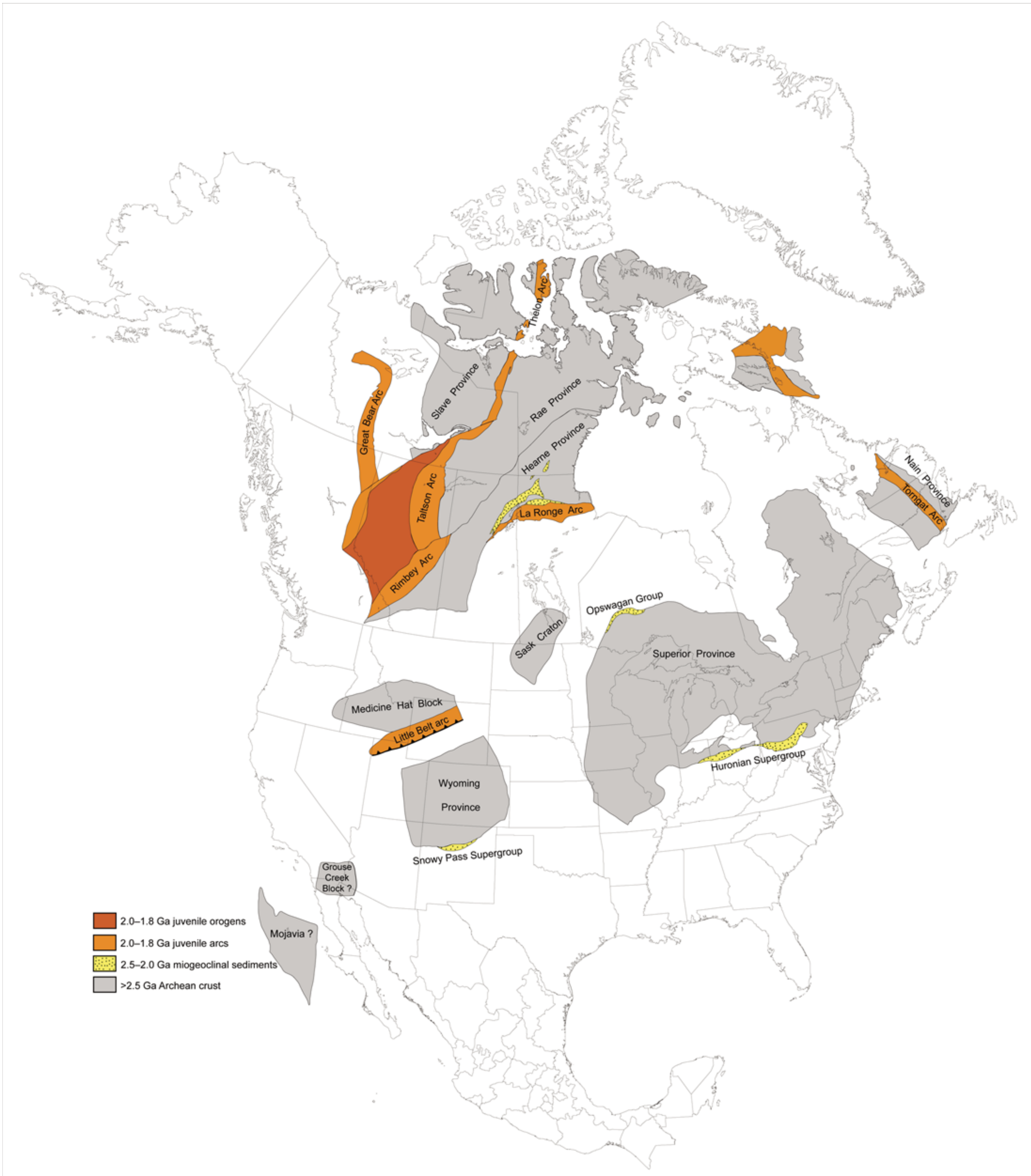

Figure 4. Continued closure of oceans, ca. 1.92–1.86 Ga, accreted La Ronge, Torngat, and Little Belt arcs (orange). Great Bear arc formed outboard of western margin of Slave province.

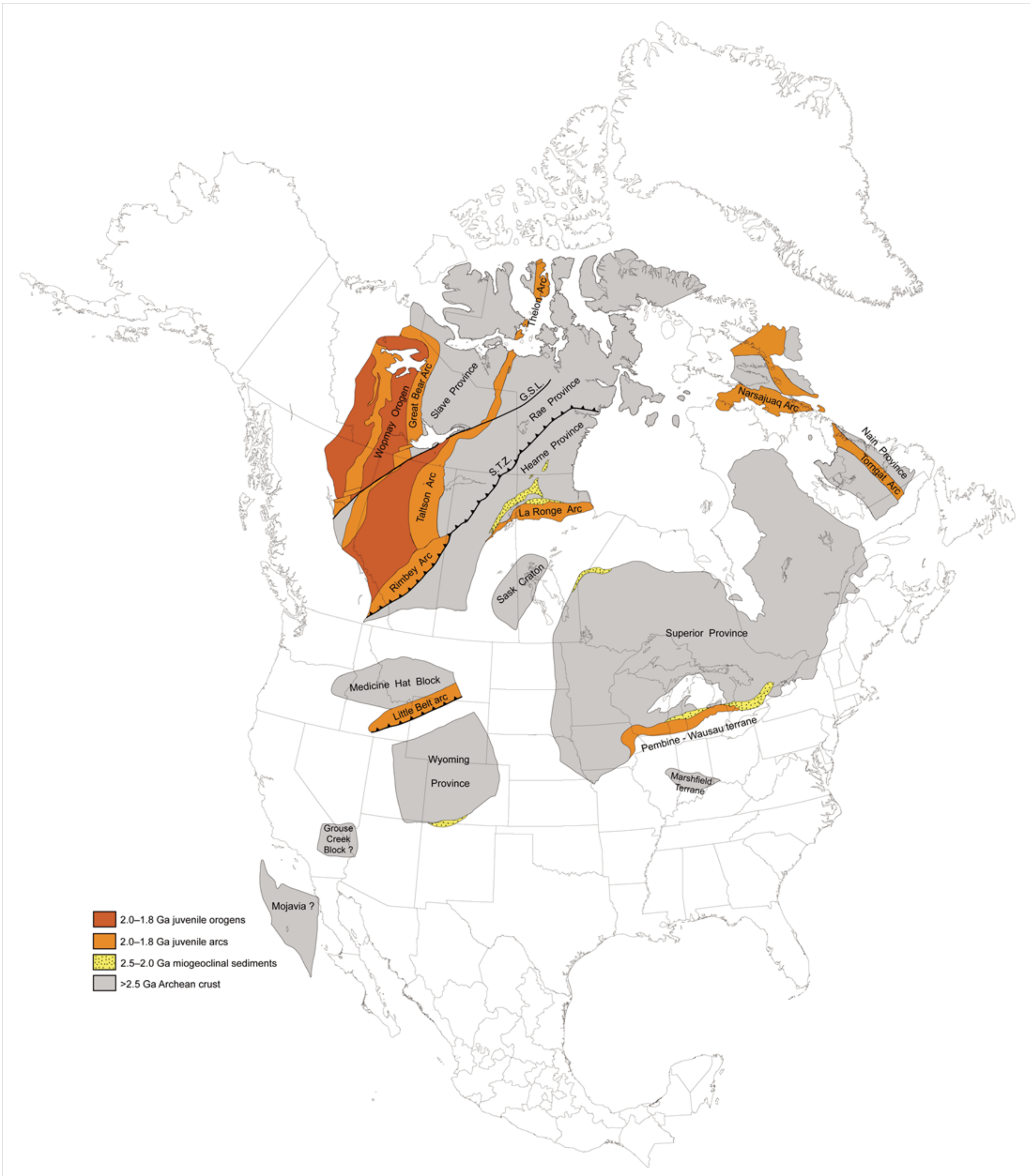

Figure 5. Juvenile arc accretion to Archean microcontinents continued with the Narsajuaq arc and the final suture of the Great Bear arc along the Wopmay orogen (ca. 1.86–1.84 Ga). Early stages of the Penokean orogeny included the accretion of the Pembine-Wausau arc terrane to the southern margin of the Superior craton. Southward thrusting along the Snowbird tectonic zone (S.T.Z.) reattached and shortened the Archean Rae and Hearne provinces. G.S.L.—Great Slave Lake shear zone.

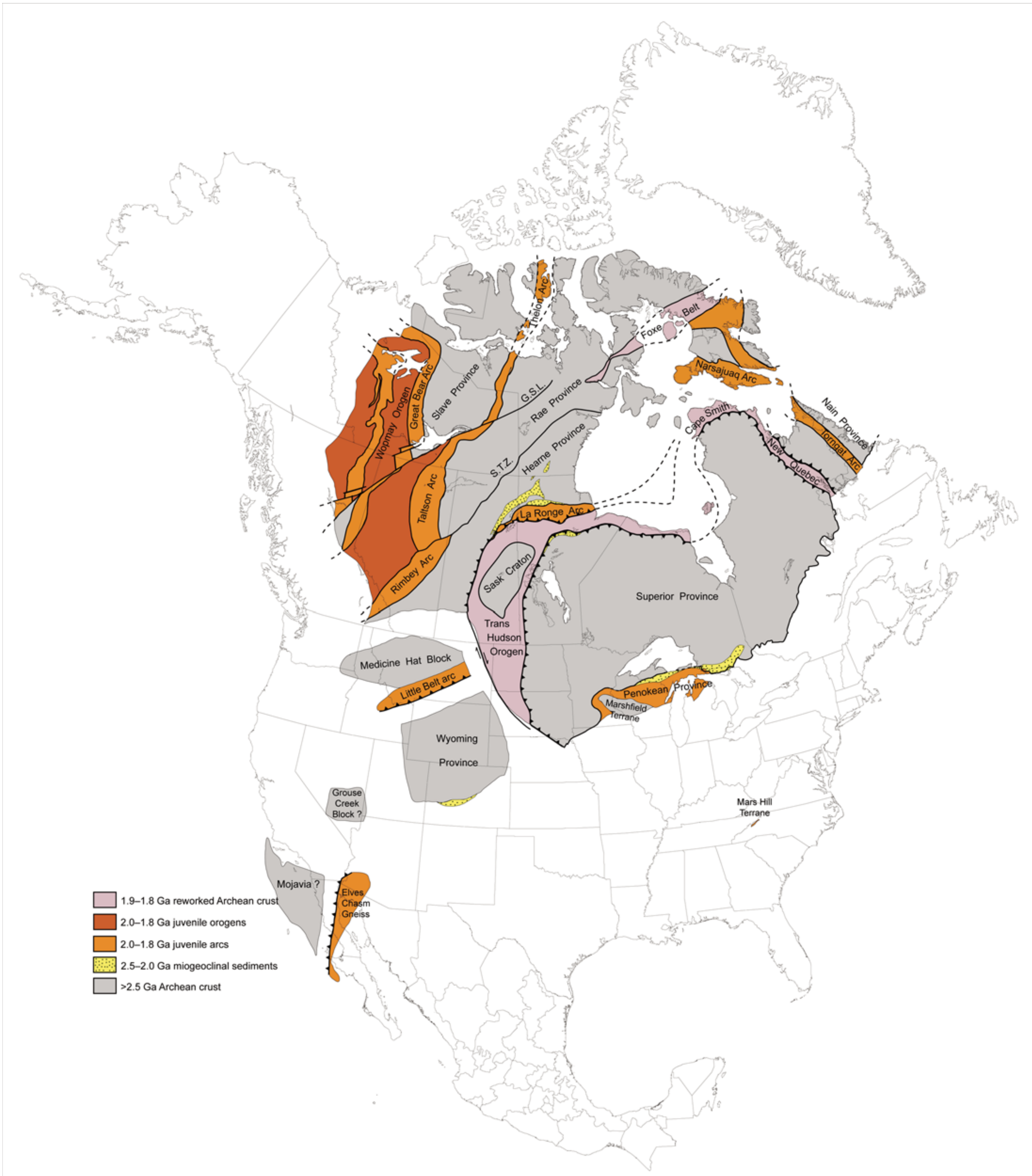

Figure 6. Collision of the Archean Superior craton and Sask craton with the Archean Rae-Hearne provinces along the complex Trans-Hudson orogenic belt (ca. 1.84–1.82 Ga). The Trans-Hudson belt includes Archean fragments, reworked Archean crust, and juvenile arcs along an extensive zone that extends from present-day Montana to Hudson Bay, and is correlated with the Foxe, Cape Smith, and New Quebec orogenic belts in northeastern Canada. Accretion of the Archean Marshfield terrane (gray) to the Pembine-Wausau arc along the southern margin of the Superior province is the final stage of Penokean province orogenesis. Similar-age rocks can also be found in the Mars Hill terrane in North Carolina. S.T.Z.—Snowbird tectonic zone. G.S.L.—Great Slave Lake shear zone.

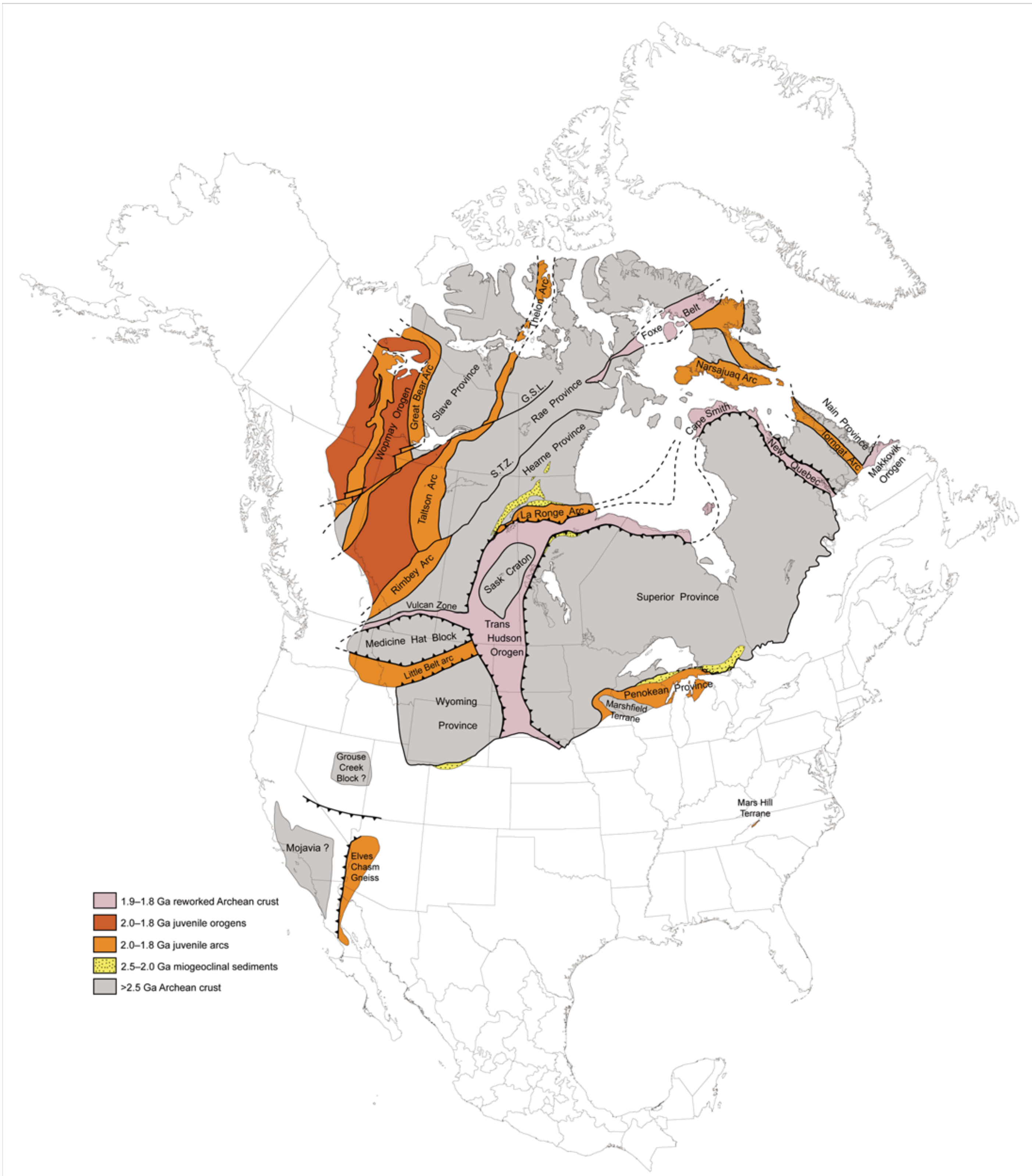

Figure 7. Accretion of the Archean Medicine Hat block (gray), juvenile Little Belt arc (orange) and Archean Wyoming province (gray) along Trans-Hudson-age (1.82–1.80 Ga) orogenic belts (e.g., Vulcan zone, light purple). Coeval Makkovik orogen formed along southern margin of the Nain province in the northeast. S.T.Z.—Snowbird tectonic zone. G.S.L.—Great Slave Lake shear zone.

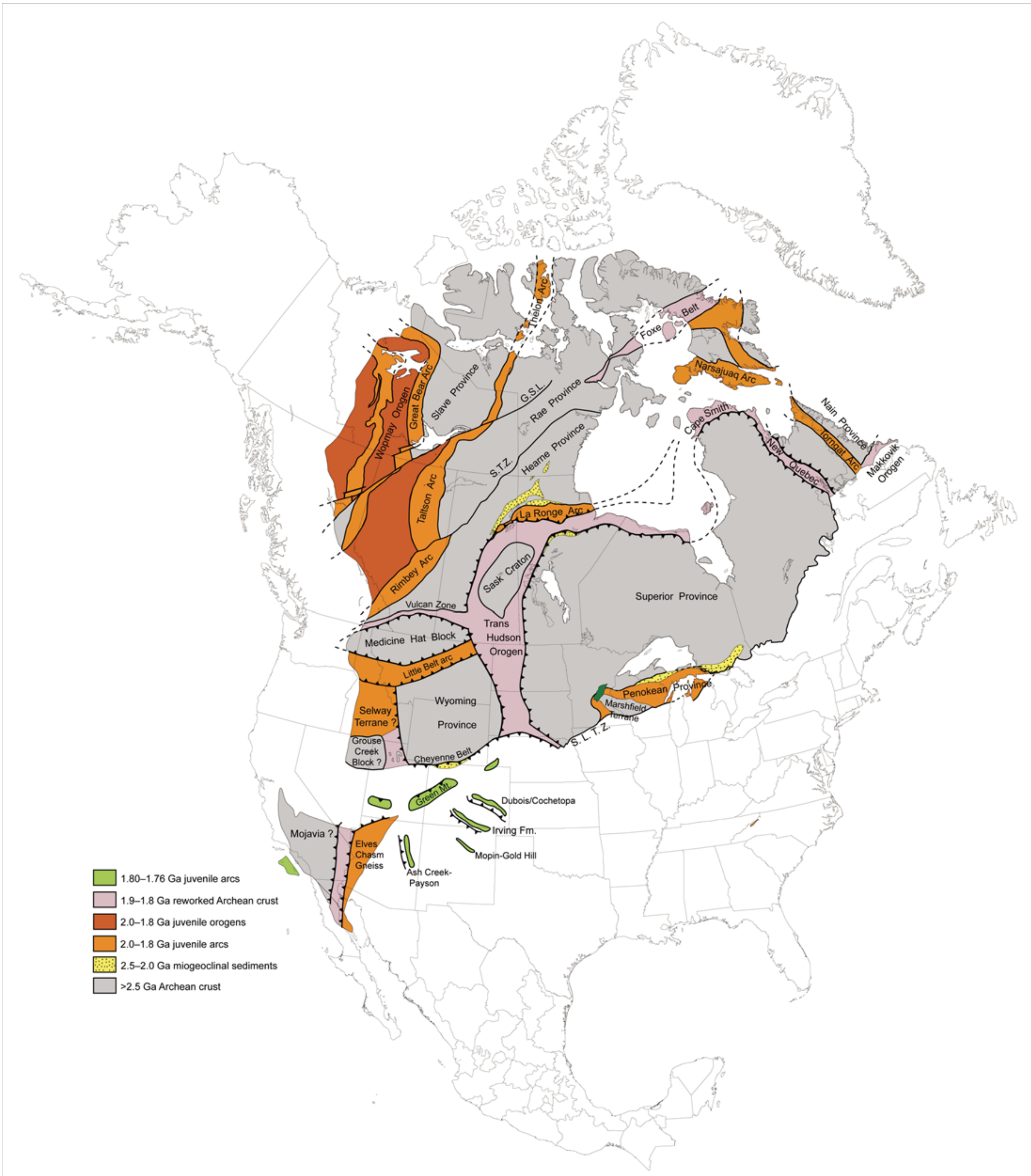

Figure 8. Archean(?) Grouse Creek block and Selway terrane are accreted to the western margin of the Wyoming province, ca. 1.80–1.76 Ga. Early Yavapai-age arcs (Green Mountain, Dubois-Cochetopa, Irving, Mopin–Gold Hill, Ash Creek–Payson) are located outboard of southern margin of assembled Laurentian core terranes, alongside Mojavia–Elves Chasm terrane. S.T.Z.—Snowbird tectonic zone. G.S.L.—Great Slave Lake shear zone.

The Trans-Hudson orogen extends into the northern United States, but only is recognized from outcrops in the Black Hills and scattered drill cores (Dahl et al., 1999). Its southern margin with younger accreted Proterozoic terranes is poorly known (Chamberlain et al., 2003). Other related components of the greater Trans-Hudson orogen include the Vulcan structure, which represents the collision of the southwestern margin of the Hearne province with the Medicine Hat block ca. 1.8 Ga. The Medicine Hat block consists of northwest-trending belts of gneiss (to 3.3 Ga; Villeneuve et al., 1993) and plutonic rocks (2.7–2.6 Ga; Mueller et al., 2002). The Great Falls tectonic zone facilitated the collision of the Medicine Hat block and Wyoming craton. This suturing event must have occurred prior to 1.8 Ga, because it appears to be truncated by the main Trans-Hudson belt (Mueller et al., 1996, 2002). The Selway terrane (Fig. 8; Foster et al., 2006) is defined as a domain of 2.4–1.6 Ga Proterozoic basement that crops out in culminations in the Sevier thrust sheets north of the Grouse Creek block. The juvenile belts and sutures of the Trans-Hudson orogenic system enabled the assembly and stabilization of disparate Archean crustal components into the cratonic core of Laurentia by 1.8 Ga (Figs. 2–8).

PENOKEAN PROVINCE: 1.9–1.8 Ga JUVENILE CRUST

The Penokean province comprises an east-northeast–trending belt of Archean and Paleoproterozoic igneous and metasedimentary rocks that extends from central Minnesota across northern Michigan and the northern coast of Manitoulin Island in Lake Huron before pinching out at the Grenville deformation front in northern Ontario (Figs. 5 and 6; Holm, 1999; Davidson, 1995). The Penokean orogeny (ca. 1.875–1.835 Ga; Van Schmus, 1976), roughly coeval with Trans-Hudson deformation, deformed and metamorphosed Archean basement and Paleoproterozoic supracrustal rocks of the Superior craton along the southern margin of Laurentia. Deformation resulted from accretion of the Pembine-Wausau oceanic arc between 1.88 and 1.86 Ga along the Niagara fault zone (Larue, 1983; Holm, 1999; Holm et al., 2007; Van Schmus et al., 2007). This was followed by accretion of the Archean Marshfield terrane along the Eau Pleine shear zone between 1.86 and 1.84 Ga (Holm et al., 2005, 2007). Oblique convergence related to southward-directed subduction between 1.89 and 1.86 Ga produced north-northwest–directed thrust nappes in Minnesota, flanked by a broad fold belt to the south (Schneider et al., 2002). Transcurrent motion resulted in primarily dextral motion along the Niagara fault zone and produced pull-apart basins to the north. Syntectonic deposition was most pronounced in the Animikie basin foredeep ca. 1.875 Ga (Schneider et al., 2002). Post-tectonic granites (1.836–1.834 Ga; Sims et al., 1989; Schneider et al., 2002) intrude the Penokean orogenic belt and crosscut the Niagara fault zone, which pins the end of the collisional event at 1.835 Ga. Sm-Nd model ages across the Penokean belt are fairly consistent, ca. 2.1 Ga (Barovich et al., 1989; Holm et al., 2005), and can be interpreted as hybrid ages due to mixing of Archean crust in the north with newly accreted juvenile crust to the south (Hoffman, 1988).

Pre–1.75 Ga juvenile crust is also recorded within a narrow northeast-trending band in Ontario and Labrador (Dickin, 1998; 2000), termed the Makkovik orogen (Fig. 7; Scharer, 1991; Corrigan et al., 2005). Nd signatures and model ages suggest arc accretion ca. 1.9 Ga, with late-stage plutonism through 1.75 Ga (Krogh et al., 1992; Dickin, 2000). Another isolated outlier of 1.84–1.80 Ga crust occurs in parts of the Mars Hill terrane in North Carolina (Fig. 6; Ownby et al., 2004). East-northeast orientations along with accretion of juvenile crust and post-tectonic granitoid intrusion suggest that the Penokean and Makkovik orogens can be considered the first of several Paleoproterozoic to Mesoproterozoic accretionary events that enlarged the Laurentian continent by adding and building crust along a southeast-facing convergent margin (Karlstrom et al., 2001).

MOJAVE PROVINCE: ARCHEAN CORE (?), 1.84 ELVES CHASM ARC, AND 1.8–1.7 Ga ARCS BUILT ON AND/OR IMBRICATED WITH OLDER BASEMENT

The Mojave province of the southwestern United States (Mojavia; Figs. 2–9) is characterized by upper amphibolite to granulite grade 1.78–1.68 Ga Paleoproterozoic gneiss exposed in isolated uplifts of the Basin and Range province, containing isotopic evidence for older crustal materials (Bennett and DePaolo, 1987; Karlstrom and Bowring, 1988, 1993; Wooden et al., 1988). Rock types include migmatitic quartzofeldspathic gneisses, pelitic gneisses, amphibolites, and rare quartzites, intruded by a heterogeneous suite of granitoids. The deformational history includes early subrecumbent folding (D1), followed by northwest-southeast penetrative shortening and D2 development of subvertical foliation. Metamorphic grade reaches granulite facies (Young et al., 1989), and may be related to abundant syntectonic plutonism. The oldest dated crustal rocks are 1.78 Ga (Barth et al., 2000) to 1.76 Ga (Wooden and Miller, 1990; Duebendorfer et al., 2001) granitoids; however zircons in metasedimentary gneisses are as old as 2.6–2.83 Ga (Wooden et al., 1994; Duebendorfer et al., 2006).

Elevated Nd and Pb isotope compositions of a variety of Mojave rock types suggest the presence of pre–1.8 Ga (and possibly Archean) crustal material reworked during 1.8–1.7 Ga orogenesis (Bennett and DePaolo, 1987; Wooden and Miller, 1990; Wooden and DeWitt, 1991; Ramo and Calzia, 1998). However, the nature of the older crust remains poorly understood and likely includes a combination of the following possible origins: (1) basement blocks in the subsurface (middle to lower crust) older than 1.8 Ga, as documented by the 1.84 Ga Elves Chasm gneiss in the Grand Canyon (Hawkins et al., 1996); (2) in situ 2.0–2.4 Ga crustal blocks (one of the options in Wooden and Miller, 1990); (3) possible Archean crustal blocks in the subsurface (see following); and/or (4) Archean detrital grains deposited in paragneisses and metasedimentary rocks and/or incorporated in plutons, as supported by detrital and inherited zircons of Archean age within gneisses of the New York Mountains (Wooden et al., 1994). Some have interpreted the Mojave province to be a fundamentally distinct terrane or microplate (Wooden et al., 1988; Condie, 1992; Duebendorfer et al., 2001); others envision it as part of the same 1.8–1.7 Ga arc system as in the Yavapai province (on the basis of similar crystallization ages; see following), but built on subsurface fragments of older crust similar to present-day Sumatra or the eastern Aleutian arc (Karlstrom and Bowring, 1993; Karlstrom et al., 2003), and incorporating far-traveled detritus (Wooden and Miller, 1994) such as the Indus fan sediments adjacent to present-day Sumatra.

The Mojave province has higher average Σ_{Nd} (+1 to −3), older Nd model ages (2.5–1.8 Ga), higher average Th/U (>4), and higher average $^{207}Pb/^{204}Pb$ (15.38) relative to the adjacent Yavapai province (+3 to +5, 1.8–1.6 Ga; <4, and 15.27). The boundary between the provinces has been drawn as a 75-km-wide mixed zone, bounded on the east and west, respectively, by the west-up thrust sense Crystal (Ilg et al., 1996) and Gneiss Canyon (Karlstrom et al., 2003) shear zones. The internal mixed zone contains small domains that may exhibit Pb isotope characteristics of either province, or transitional values (Wooden and DeWitt, 1991; Duebendorfer et al., 2006). This zone has been interpreted tectonically as (1) juvenile volcanic- and sediment-filled rift basins within stretched Mojave crust (Wooden and DeWitt, 1991; Duebendorfer et al., 2006), and (2) thrust imbrication of Mojave and Yavapai crust (Karlstrom et al., 2003; Karlstrom and Williams, 2006), and mixed-in Archean detritus (from the Wyoming province or an unknown Archean source terrane), with

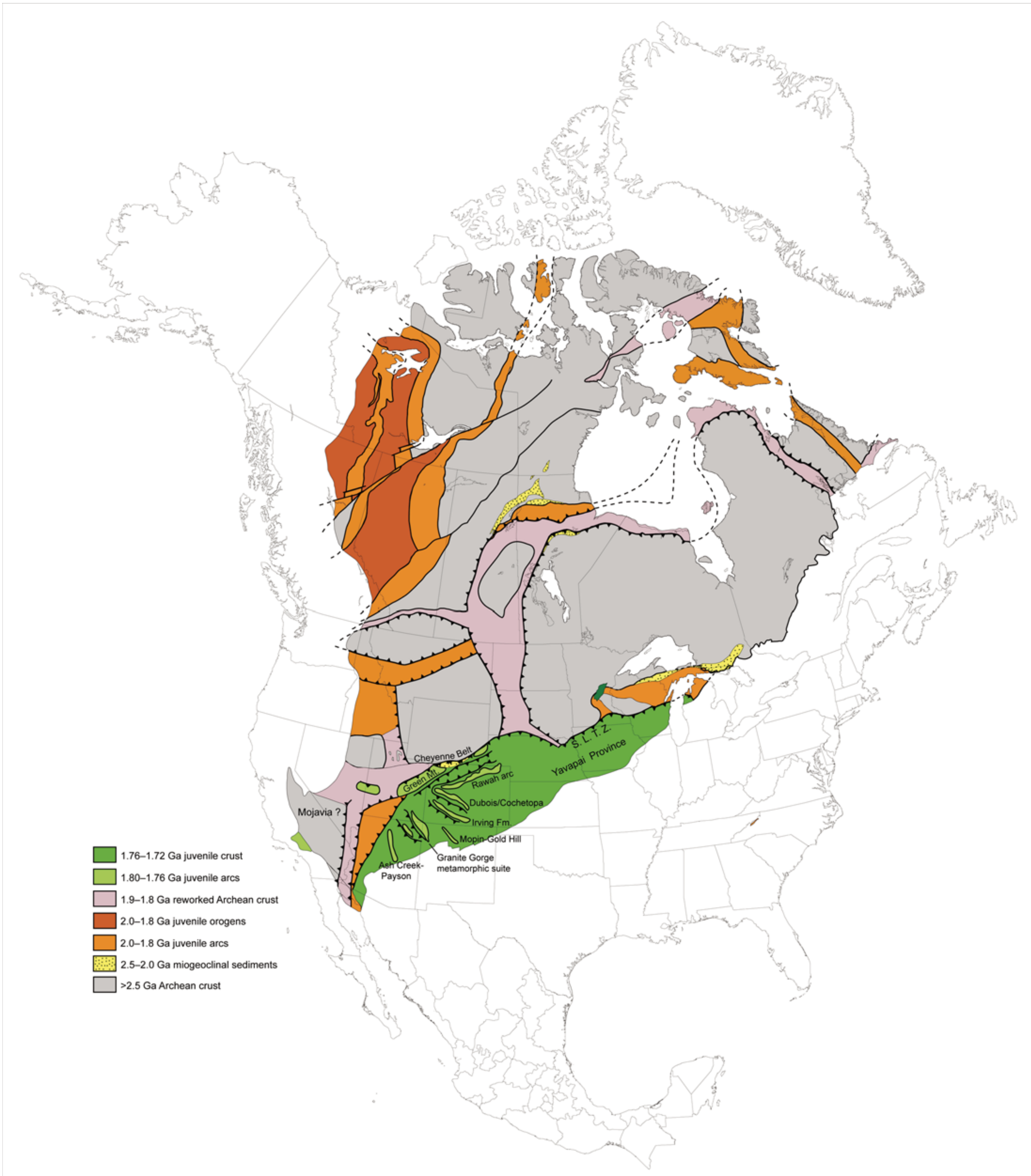

Figure 9. Accretion of the Archean(?) Mojavia terrane (gray) and associated reworked Paleoproterozoic crustal components (e.g., Elves Chasm gneiss, orange) culminated during Yavapai orogenesis. The Yavapai province is largely an assembly of oceanic arc terranes (ca. 1.76–1.72 Ga) in a complex geometry, analogous to the present-day Banda Sea region of the Indonesian archipelago. Suturing of juvenile Yavapai crust occurred along the Cheyenne belt (southern margin of Wyoming province) and the Spirit Lake tectonic zone (S.L.T.Z., southern margin of western Superior and Penokean provinces).

resulting variable influence of inherited older crust during differentiation of 1.78–1.70 Ga arc rocks (Wooden and Miller, 1990).

In these respects, the tectonic relationship between the Mojave and Yavapai provinces remains enigmatic and requires additional isotopic and structural studies. However, our preferred alternative (Figs. 2–9) depicts the Mojave province as a possible Archean fragment (Mojavia) based on the Nd model ages of 2.4–2.7 Ga from Death Valley (Ramo and Calzia, 1998), and the questionable presence of an Archean rock in the Turtle Mountains (Wooden and Miller, 1990). The extent of such Archean crust in the Mojave province (if any) is unknown. Alternatively, the amount of Archean detritus subducted and incorporated into plutons of a younger arc needed to satisfy the Nd data is estimated to be 14%–32% (Ramo and Calzia, 1998) to 17%–42% (Bennett and DePaolo, 1987). Nd model ages of 1.8–1.9 Ga in the eastern Mojave province suggest that Elves Chasm–type 1.84 Ga crust could underlie significant areas and may extend in the subsurface west to at least the Gneiss Canyon shear zone (Karlstrom et al., 2003). West of this shear zone, both Nd and Pb isotopic data suggest a westward increase in older crustal components, and therefore we postulate a wider zone of isotopic mixing between the Mojave and Yavapai provinces than depicted by Wooden and DeWitt (1991) and Duebendorfer et al. (2006). One possible explanation for the general westward increase in Nd model ages in the Mojave province (Bennett and DePaolo, 1987; Ramo and Calzia, 1998) is the juxtaposition of an Archean block with a 1.84 Ga block and tectonic mixing of these blocks (Figs. 6–8). Major shear zones are located in and at the edges of this zone of transition (Fig. 8; Ilg et al., 1996; Quigley, 2002; Karlstrom and Williams, 2006), and may indicate early zones of weakness, possibly at the edges of discrete crustal blocks or terranes that were reactivated as high-strain zones and transposed to more northeastern trends during the Yavapai orogeny (Figs. 8 and 9). Nd isotopic data showing mantle separation model ages of 2.0–1.8 Ga in parts of the Yavapai province (Coleman et al., 1996) and Mazatzal province (Eisele and Isachsen, 2001) also suggest a component of older crust in the subsurface and/or mixing of older detritus.

YAVAPAI PROVINCE: 1.80–1.70 Ga JUVENILE CRUST ASSEMBLED DURING THE 1.71–1.68 Ga YAVAPAI OROGENY

The northeast-trending Yavapai province has been defined (e.g., Bowring and Karlstrom, 1990; Karlstrom and Bowring, 1993; Karlstrom and Humphreys, 1998) as the wide zone of dominantly juvenile crust that extends from Arizona to the area of Colorado south of the Cheyenne belt, then northeastward in the subsurface to the mid-continent region (Van Schmus et al., 2007; Fig. 9). The Yavapai province is the product of the accretion of dominantly juvenile arc crust from 1.80 to 1.70 Ga, including probable outboard development and collisions of arcs from 1.78 to 1.72 Ga (Figs. 8 and 9), and an orogenic peak ca. 1.71–1.68 Ga (Yavapai orogeny) that resulted in a progressive amalgamation of Yavapai crust to Laurentia (Fig. 10). Some areas have been identified that show isotopic evidence or inherited zircon evidence for the presence of older components (e.g., in Colorado; Hill and Bickford, 2001), and Archean zircons have been documented in some supracrustal successions (Selverstone et al., 2000; Hill and Bickford, 2001; Jones, 2005). However, the amount of Archean and pre–1.80 Ga Paleoproterozoic material incorporated into the Yavapai province is interpreted to be small based on Nd model ages of 2.0–1.8 Ga, just older than crystallization ages of 1.8–1.7 Ga (DePaolo, 1981). Similarly, Pb isotope studies suggest that this terrane was derived primarily from juvenile mantle material (Aleinikoff et al., 1993).

Ilg et al. (1996) and Hawkins et al. (1996) reported and documented the presence of the 1.84 Ga Elves Chasm block in Grand Canyon. Such terranes are expected within dominantly accretionary orogens (Karlstrom et al., 1993; Jessup et al., 2005). However, we see no evidence that “Trans-Hudson- Penokean crust, or fragments thereof, underlie much of southern Laurentia” (Bickford and Hill, 2007, p. 169). We do not believe that the presence of Archean detritus in some successions and not others in the Yavapai province provides adequate documentation to infer distinct tectonic blocks and large translations between blocks (cf. Selverstone et al., 2000) without more detailed understanding of deformational history and the extent and provenance of various metasedimentary successions.

Accretion and associated deformation took place during several pulses within a long orogenic progression. The oldest rocks include 1.80–1.75 Ga granite-greenstone associations that consist of metabasalt, metaandesite, and metarhyolite, and associated volcanogenic metasedimentary rock intruded by calc-alkaline tonalite to granodiorite plutons. Fragments of these old arcs are widely dispersed within the Mojave and Yavapai provinces. The oldest rocks dated (1.80–1.78 Ga) are found in widely separated areas. (1) Just south of the Cheyenne belt, in the Green Mountain block, are 1.79–1.78 interlayered calc-alkalic (Condie and Shadel, 1984), bimodal, subaerial, submarine volcanic and volcaniclastic rocks and volcanogenic massive sulfide deposits. These rocks are cut by layered mafic complexes (1.774 ± 2 Ga quartz diorite, 1.768 ± 8 Ga gneissic inclusion; Pallister and Aleinikoff, 1987; 1.781 Ga (Pb/Pb) gabbro; Snyder, 1980). Nd data indicate that these rocks were not derived from older crust (Chamberlain, 1998), and represent mantle-derived oceanic arc materials. (2) In the Needle Mountains of southern Colorado, mafic gneisses (metavolcanics) of the Irving Formation (1.80–1.79 Ga) are intruded by tonalitic to granodioritic intrusives of the Twilight Gneiss (1.78–1.77 Ga; Gonzales et al., 1994). Similar rocks are the 1.77–1.76 Ga Dubois succession of Colorado (Bickford et al., 1989), the 1.77–1.75 Ga Mopin–Gold Hill successions of northern New Mexico, and 1.76 Ga granitoids that are intruded by the Payson ophiolite in central Arizona (Fig. 9; Karlstrom et al., 1997). These successions are dominated by mafic metavolcanic rock, but also include a significant proportion of felsic and intermediate metavolcanic rock. Locally they preserve evidence for marine deposition, including pillow basalt and turbidite sedimentary structures. The metavolcanic rocks are tholeiitic to calc-alkaline, the plutons are calc-alkaline, and the combined association is interpreted to have formed in arcs in a marine setting (Condie, 1986). (3) The Transverse Ranges of the Mojave province contain 1.78–1.76 Ga tonalities and granodiorites (Barth et al., 2000).

The alleged bimodality of some of the volcanics (Bickford and Hill, 2007), used to argue against an arc origin for these rocks, has not been well documented, and additional studies need to be conducted in the context of modern petrologic studies of major and trace element tectonic discrimination diagrams. Generally, regardless of silica content, these rocks are interpreted to show major and trace element indications of arc associations (Bickford et al., 1989; Boardman and Condie, 1986; Knoper and Condie, 1988). In terms of modern analogs, basalt-rhyolite associations are relatively common in modern oceanic arc systems (Hamilton, 1979); bimodality of the volcanic rocks, if ultimately documented as a characteristic of the Yavapai province, is also compatible with the arc model for crustal growth.

The original distribution and orientation of juvenile 1.8–1.75 Ga arcs (Figs. 8 and 9), and hence subduction-zone orientation and polarity of the earliest juvenile Paleoproterozoic additions, remain poorly understood given the intensity of later northwest-southeast shortening near the end of the Yavapai orogeny. However, in southwestern Colorado, a general northwest strike of foliation is preserved (Jessup et al.,

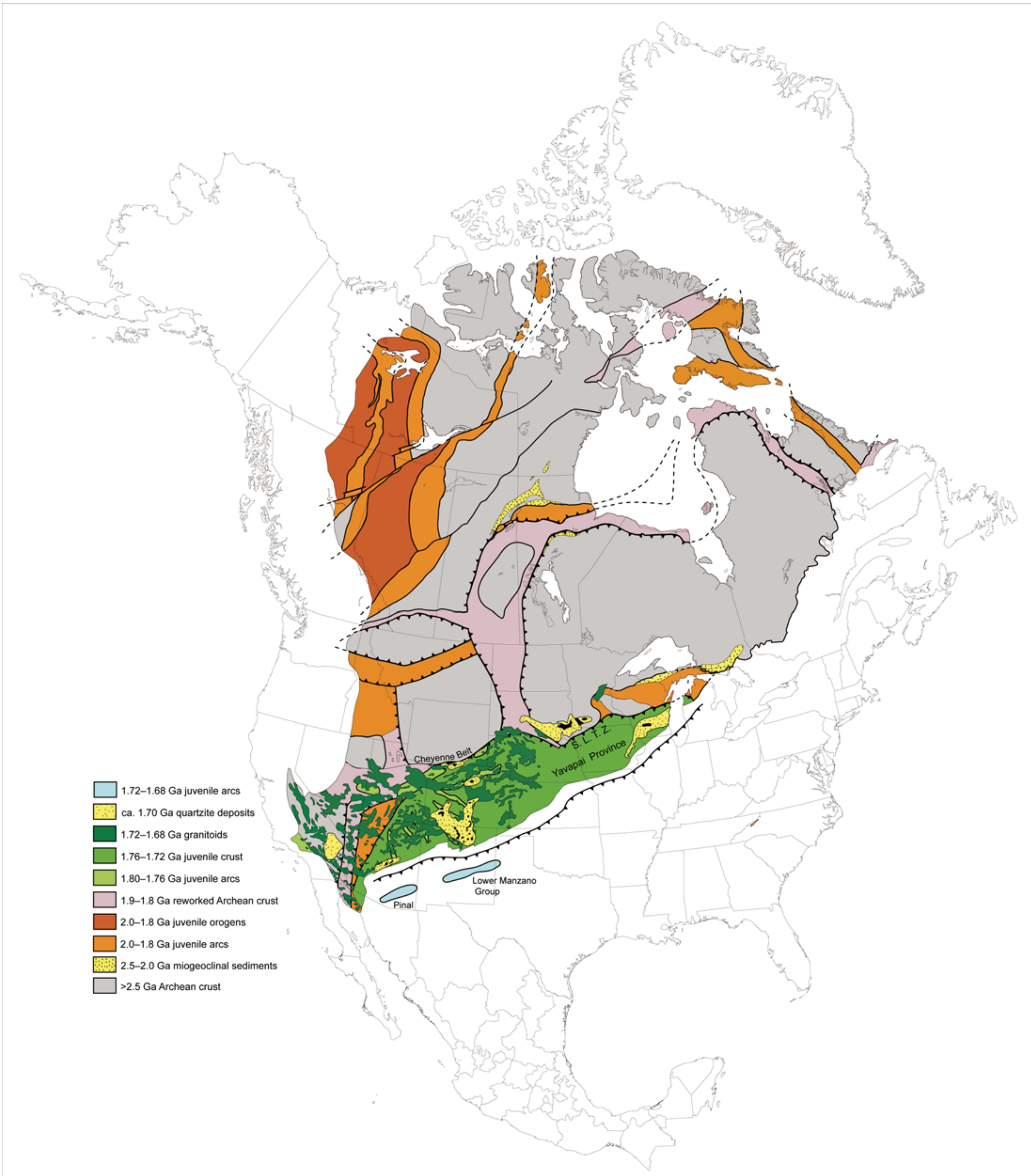

Figure 10. Accretion of Yavapai arcs was closely followed by voluminous intrusion of granitoids (dark green; ca. 1.72–1.68 Ga) that stitched existing province boundaries and helped to stabilize juvenile crust. Distinctive quartzite-rhyolite successions (stippled yellow) occurred during the Yavapai orogeny (ca. 1.70 Ga). Pinal and Lower Manzano Group arcs (light blue) are outboard of Yavapai southern convergent margin. S.L.T.Z.—Spirit Lake tectonic zone.

2005) that is mimicked by northwest-trending volcanic and/or plutonic belts that are progressively younger toward the northeast (Irving-Dubois-Cochetopa; Fig. 9). This is interpreted as a complex oceanic arc system (somewhat similar to the present-day Banda arc) related to several potential subduction systems, followed by collision and imbrication of arcs of somewhat distinct ages.

Younger series of greenstone complexes throughout the Yavapai province (Fig. 9) range in age from 1.75 to 1.71 Ga. For example, the Rawah block south of the Green Mountain block contains volcanogenic supracrustal rocks (younger than 1.76 Ga) and granites (1.746 and 1.735 Ga; Premo and Van Schmus, 1989). In southern Colorado, the 1.74–1.72 Ga Cochetopa succession (Bickford et al., 1989) may have originally been in depositional contact with the older Dubois arc succession (Shonk, 1984), but the contact is now sheared (Knoper and Condie, 1988). A similar age range for arc rocks is in New Mexico, where the 1.72 Ga Pecos greenstone belt is in close proximity to the 1.77–1.75 Ga Mopin and Gold Hill successions. The Cochetopa and Pecos rocks are basalt and/or rhyolite successions (including pillow basalt) with interlayered metasedimentary rocks and local ultramafics, iron formation, and massive sulfide deposits, all intruded by calc-alkaline mafic plutons. Major and trace element studies indicate that these rocks were formed in an arc or backarc setting, but one that was perhaps more evolved than the Dubois succession (Bickford et al., 1989), and possibly developed in a continental-margin setting (Boardman and Condie, 1986; Knoper and Condie, 1988) as the offshore arcs approached Laurentia (Jessup et al., 2005). In central Arizona, arc rocks of the Yavapai province include the 1.76–1.73 Ga Ash Creek block, which contains intermediate volcanics, and 1.76 Ga basement intruded by the Payson ophiolite (Dann, 1991). In the Grand Canyon arc rocks range in age from 1.76 to 1.71 Ga, similar to the age range of arc plutons in the Mojave province and in central Arizona. In Figure 9 we show several inferred arcs, many that developed on oceanic crust, but some that straddled older Mojave crust. We envision that some of these arcs may have evolved over 20–40 m.y., compatible with the life cycle of modern arcs, but the number, geometry, and evolution of the crust forming elements in the Yavapai province remain poorly known.

The best agreed upon suture zone in the Precambrian of the western United States is the Cheyenne belt, a northeast-striking subvertical set of shear zones that records the 1.78–1.75 Ga collision of Proterozoic arc terranes with the Archean Wyoming province and its rift-related miogeoclinal cover sequence (Hills and Houston, 1979, Karlstrom and Houston, 1984; Chamberlain, 1998). Integrated seismic and geologic investigations of the deep structure of the Cheyenne belt in the Sierra Madre and Park Range reveal a complex deep structure involving an interwedging in the crust and upper mantle between Archean and Proterozoic rocks, and a north-dipping high-velocity anomaly in the upper mantle that is interpreted as a segment of Paleoproterozoic mantle that was underthrust northward during suturing (Karlstrom et al., 2002, 2005; Tyson et al., 2002; Morozova et al., 2005). This north-dipping mantle structure is interpreted to project upward toward the Farwell Mountain structure, just south of the Cheyenne belt (Tyson et al., 2002; Morozova et al., 2005). The subduction system across which the suturing took place was overall south dipping, as suggested by the absence of arc granitoids in the southern Wyoming province. Reactivations of the zone took place until ca. 1.6 Ga (Duebendorfer et al., 2006), attesting to the long-term weakness of this zone. The difference in lithospheric character across the zone, locked in during suturing, is interpreted to explain the different responses of Colorado versus Wyoming lithospheres during numerous later tectonic events (Karlstrom and Humphreys, 1998).

Other potential sutures that now juxtapose once-separate lithospheric plates or microplates have also been proposed. In some cases potential sutures separate subtly different arc terranes in the Yavapai province. Tyson et al. (2002) and Morozova et al. (2005) proposed a suture zone between the Green Mountain arc and the Rawah arc based on seismic studies, lithologic and age contrasts, and metamorphic grade change across the Farwell Mountain shear zone. Exotic rock types such as interlayered marble, amphibolite, calc-silicate, and metachert that may be altered marine exhalites (Foster et al., 1999) crop out along the Farwell Mountain–Lester Mountain suture zone. Siliceous pod rock (Snyder, 1988), with lenses of quartz + sillimanite ± muscovite, may record hydrothermal alteration. There are boudins of small mafic–ultramafic bodies, including chromite- or spinel-bearing amphibole peridotites, orthoamphibole-rich rocks, dunite, wehrlite, and harzburgite (Snyder, 1980), and deformed pillow basalts. These are interpreted as tectonic slivers of ophiolite within a sedimentary accretionary complex. The Farwell Mountain–Lester Mountain suture zone contains several tectonite fabrics, including north-verging, subrecumbent, isoclinal folds that are probably the surface expression of south-dipping seismic reflectors (Tyson et al., 2002). A younger, overprinting foliation may represent shortening and steepening of fabrics due to continued convergent tectonism.

In Arizona, the Crystal shear zone in the Grand Canyon is interpreted to be the eastern shear zone of a wide middle crustal accretionary duplex zone that marks the suture between the Mojave and Yavapai provinces (Ilg et al., 1996; Karlstrom et al., 2003; Karlstrom and Williams, 2006). Ultramafic rocks are mapped that have been tectonically emplaced within mixed turbidite and volcanic successions (1.75–1.72 Ga) and intruded by calc-alkaline batholiths. The ultramafics have chemistries and mineralogies consistent with deep crustal cumulates (Low et al., 2002), but may include ophiolite fragments. Their tectonic emplacement within supracrustal successions indicates a high degree of allochthoneity, likely more than tens of kilometers (Karlstrom and Williams, 2006). Lithologies that are consistent with some combination of dismembered ophiolite components and marine arc settings are also present in numerous other places across the Yavapai province (Shaw and Karlstrom, 1999; Quigley, 2002; Cavosie and Selverstone, 2003; Tyson et al., 2002; Strickland et al., 2003). Our view is that these are best explained as middle crustal remnants of regionally extensive accretionary complexes (Karlstrom and Williams, 2006).

Suture zones are often identified based on the presence of ophiolites, but the Payson ophiolite of the southernmost Yavapai province was interpreted by Dann (1997, p. 364) to have "developed in situ as a distinct extensional phase during the complex evolution of an Early Proterozoic arc." In this model, it may have developed near the southern edge of 1.76 Ga Yavapai province crust as an intraarc basin formed as a pull-apart structure related to arc-parallel strike-slip faults, rather than as a back-arc basin formed by roll back. Dann (1997) used Indonesian analogs (Marinduque intraarc basin in the Philippines), reinforcing the general model for a complex association of extensional and transcurrent structures within an overall convergent arc accretion zone. Nd and isotopic data suggest that there may be older crust beneath the Payson ophiolite similar in composition to the eastern Mojave province (Coleman et al., 1996). Following infilling of the basin with turbidites (1.72 Ga), this block was deformed and accreted to the rest of the Yavapai block by ca. 1.7 Ga (Dann, 1991).

Much research over the past two decades has focused on combined structural and geochronologic studies to determine timing of deformation. When viewed regionally, these studies show a near continuum of deformation between 1.78 and 1.68 Ga within the Yavapai province. The observed regional range in volcanic and granitoid batholith ages from 1.8 to 1.75 Ga are interpreted to record subduction-related tectonism and outboard collisions to form the 1.8–1.75 Ga arcs. These events overlap with the timing of early D1 deformational structures of 1.75–1.73 Ga and with blooms of syntectonic granitoids (Fig. 10). Welding of arc terranes to North America is interpreted to have been a progressive deformation from ca. 1.72 to 1.68 Ga, based on timing of deformations documented from different areas and polyphase reactivations of older terranes during younger events. In southern Colorado, monazite dating and aureoles around syntectonic plutons suggest that pervasive shortening across the orogens (D2) took place at 1.71–1.70 and 1.69–1.67 Ga (Shaw et al., 2001; Jessup et al., 2005). This is similar to the Grand Canyon, where metamorphism and plutonism were a progressive D1-D2 shortening that was active from 1.71 to 1.69 Ga, based on the growth of metamorphic monazite (Hawkins et al., 1996). These dates represent a Yavapai orogenic peak that lasted for tens of millions of years. An interesting component of this deformational history is that decompression of some blocks took place by collisional exhumation involving thrusting and reverse faulting plus erosion during arc collision at the same time that other blocks were being buried by these thrusts (Karlstrom et al., 2003). Deformation at elevated temperatures involved middle crustal flow on both steep D2 and shallow D1 fabrics, with continued interaction of shortening and fabric reactivation during decompression (Dumond et al., 2007).

Structural studies of folds and foliations indicate that contractional deformation dominates the rock fabrics, and so we interpret the Yavapai orogeny in terms of a long-lived convergent plate margin orogen along a southward-growing Laurentia. Transcurrent deformations have been well documented structurally in Arizona as partitioned shear related to overall convergence (Bergh and Karlstrom, 1992). While transpressional structures are expected in any convergent setting and many such structures might yet be revealed by detailed structural studies, there have been no regional-scale (more than tens of kilometers) strike-slip displacements along transcurrent shear zone systems mapped or adequately documented in the Yavapai province. Large-scale transcurrent motions were speculated across the Buckhorn Creek shear zone based on small areas with shallow lineations in northern Colorado (Selverstone et al., 1997; Cavosie and Selverstone, 2003), but these need additional structural documentation to decipher their timing and regional significance. The preponderance of dip-slip stretching lineations for both D1 and D2 suggests a dominance of contractional deformation during later stages of crustal assembly.

The Yavapai orogeny also affected areas of the mid-continent and is documented as a series of magmatic episodes at 1.80, 1.775, and 1.75 Ga that overlapped with the Penokean orogeny and the assembly of the Laurentian core (Figs. 8–10). These are interpreted in terms of subduction flip from south dipping in the Penokean orogeny to north dipping, interpreted to explain the southeast migration of granitoid magmatism from 1.80 to 1.75 Ga (Holm et al., 2005).

Metamorphic studies indicate a polyphase middle crustal metamorphic history for most of the rocks of the Mojave, Yavapai, and Mazatzal provinces. Williams and Karlstrom (1996) proposed that most rocks underwent a clockwise-looping pressure-temperature path by which they were taken from surface to 10–20 km depths via a combination of thrusting, recumbent folding, then crustal thickening. Large regions record isobaric metamorphism at some crustal level (Karlstrom and Williams, 2006). Large areas of central Arizona and central Colorado were buried to peak metamorphic pressures near 3 kbar (10 km depths; Karlstrom and Williams, 1998, 2006; Karlstrom et al., 2002; Jessup et al., 2005). The Grand Canyon and some areas of the Mojave province represent large regions that reached 6 kbar (20 km levels; Ilg et al., 1996; Duebendorfer et al., 2001; Dumond et al., 2007). Large regions in New Mexico record isobaric metamorphic conditions of 3.5–4.5 kbar (12–15 km; Grambling, 1981). Large lateral metamorphic temperature gradients of ~300 °C at constant pressure suggest a Paleoproterozoic metamorphic style involving pluton-enhanced middle crustal metamorphism (Williams and Karlstrom, 1996; Karlstrom and Williams, 2006; Dumond et al., 2007). This style of metamorphism is characterized by abrupt lateral temperature gradients at near-constant metamorphic pressures and should not be confused with juxtapositions of crustal levels. For example, the apparent granulite facies conditions in the Grand Canyon (Dumond et al., 2007) and the Mojave (e.g., Duebendorfer et al., 2001) may in part be a reflection of large volumes of plutonic rocks in these areas rather than increased crustal depth.

Other workers have proposed different names for orogenies in local areas or in adjacent regions that all overlap in time with, and we view as components of, the Yavapai orogeny. The term Ivanpah orogeny was used in the eastern Mojave province for 1710–1680 deformation in the New York Mountains area (Wooden and Miller, 1990). The term Central Plains orogeny was used to describe the eastward extensions into the mid-continent of the Colorado Proterozoic rocks (Sims and Peterman, 1986). Medicine Bow orogeny was used to describe the 1.78–1.75 Ga continent-arc collision that produced the Cheyenne belt (Chamberlain, 1998). Colorado province (Reed et al., 1987) and Colorado orogeny (Sims and Stein, 2003) were used to describe 1.78–1.70 tectonism in Colorado. Our view, in keeping with the 100–1000 km scale of orogens and the observed diachronous nature of tectonism in convergent zones, is that these different areas are part of the same orogenic system. Our use of the term Yavapai orogeny follows Karlstrom and Bowring (1988) and Karlstrom et al. (2001) and encompasses all of these areas. We envision a series of separate oceanic arcs that developed diachronously outboard of Laurentia and became welded together and to Laurentia across the Cheyenne belt and other sutures from 1.78 to 1.68 Ga. This is compatible with the 100 m.y. time span for many orogens. It is important to continue to evaluate temporal and spatial partitioning of deformation to better understand the geometry of the accreted blocks, the age and extent of any older crustal substrates, and the timing of discrete events within this long continuum of arc accretion and assembly.

MAZATZAL PROVINCE: 1.70–1.60 Ga JUVENILE CRUST ADDED TO LAURENTIA DURING THE 1.65–1.60 Ga MAZATZAL OROGENY

The Mazatzal province contains ca. 1.68–1.60 Ga crust that is interpreted to have formed in continental margin volcanic arcs and back-arc-related supracrustal successions that extend from the southwestern United States through correlative rocks of the northern mid-continent region to the Labradorian orogeny in the Canadian Maritime provinces (Fig. 11). The term Mazatzal orogeny has been used to include all deformation younger than 1.65 Ga, including deformation that extends into a foreland zone on the edge of the Yavapai orogeny inboard from the boundary with Mazatzal crust (Karlstrom and Bowring, 1993; Shaw and Karlstrom, 1999). The oldest rocks in the Mazatzal province are 1.68–1.65 Ga volcanogenic greenstone successions (Karlstrom et al., 2004) that typically include basalt and basaltic andesite (including pillow volcanics), dacitic tuff, and rhyolite. Geochemical data support an oceanic basalt origin for some of these rocks (Condie, 1980), and Nd and Pb isotopic data highlight crust that has a slightly younger mantle derivation model age (1.8–1.7 Ga) than the adjacent Yavapai province (Bennett and DePaolo, 1987; Wooden and DeWitt, 1991; Aleinikoff et al., 1993).

In the well-exposed Manzano Group of New Mexico (Karlstrom et al., 2004), the volcanic rocks grade upward into thick metasedimentary successions that contain chert at the base, then

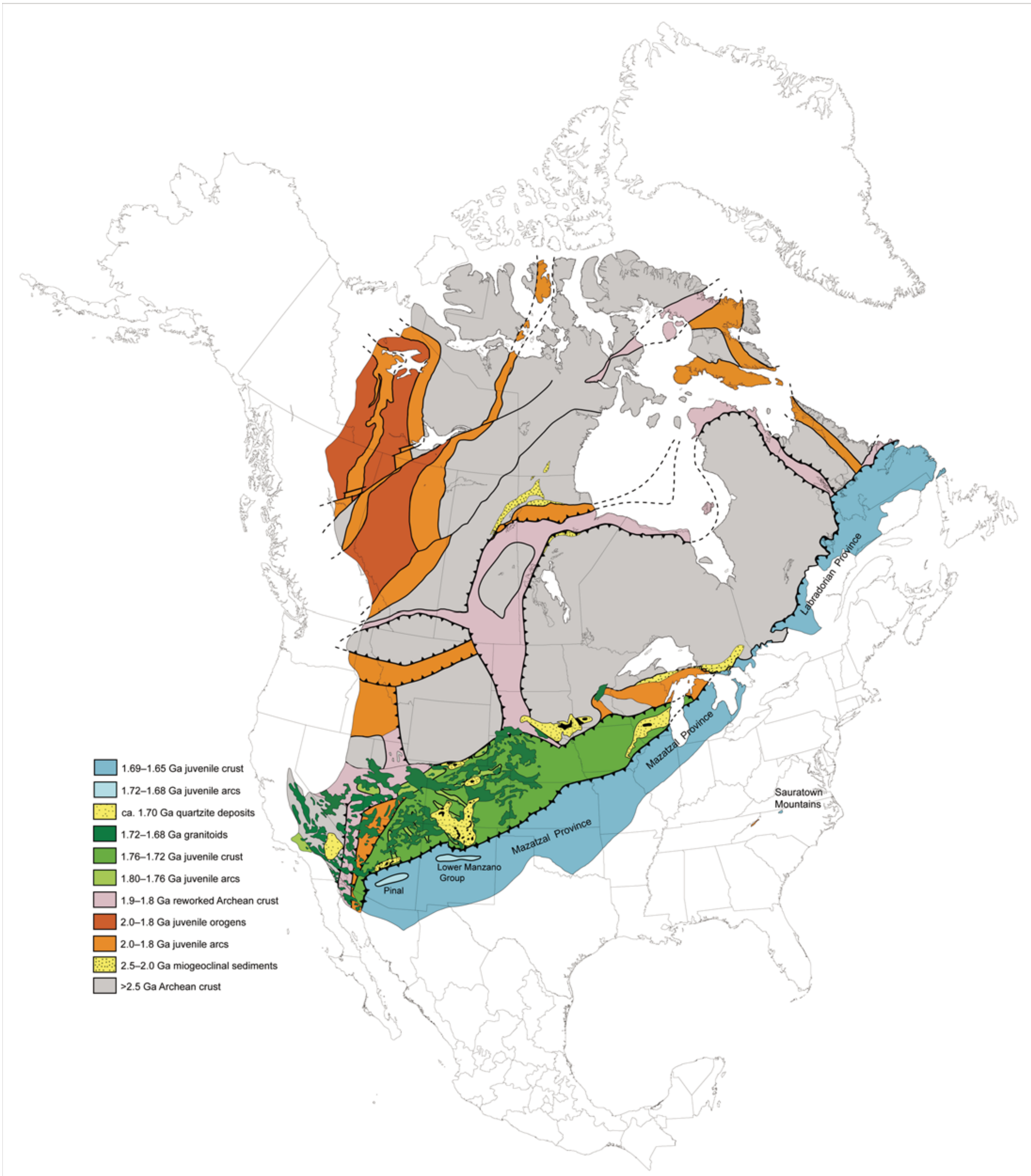

Figure 11. The Mazatzal province (ca. 1.69–1.65 Ga) is well documented in the southwestern United States, where it includes the Pinal arc and Lower Manzano Group (light blue). Coeval basement in northeastern Canada has been called the Labradorian province, and is tentatively correlated with the Mazatzal province across the poorly exposed mid-continent. Mazatzal-age outcrops are also documented in the Sauratown Mountains window in North Carolina.

are progressively more mature upward through lithic arenite and quartzarenite. Metarhyolites yield dates perhaps as old as 1.70–1.68 Ga, but most are ca. 1.67–1.66 Ga, which is interpreted as the age of the volcanic and/or sedimentary succession. Calc-alkaline plutons intrude the volcanic rocks and yield dates of 1.66–1.65 Ga over a wide region (Fig. 12). These plutons, with their similar-age volcanic country rocks, make up most of the exposed crust of the Mazatzal province, with some blocks dominated by outcrops of granitoids (Sunflower block, Arizona; Karlstrom and Bowring, 1993). Isolated 1.68–1.65 Ga plutons may extend as far north as the Cheyenne belt (Premo and Van Schmus, 1989) and as far northwest as the Mojave province (Wooden and Miller, 1990; Barth et al., 2000). This distribution suggests that the 1.66–1.65 Ga magmatic event affected a large region, with subduction-related magmatism in the Mazatzal province and A-type magmatism in intracratonic regions (Anderson and Cullers, 1999; Anderson and Bender, 1989). Deformation began before emplacement of syntectonic 1.65 Ga granitoids and outlasted deposition of 1.60 Ga rhyolites (Luther et al., 2005), suggesting a progressive deformation in thrust systems linked in time and space to both basin formation and granitoid emplacement.

A suture zone between the Yavapai and Mazatzal province lithospheric blocks has been proposed along the Jemez lineament in northern New Mexico (Karlstrom et al., 2002, 2005; Magnani et al., 2004, 2005). Deep seismic reflection images show oppositely dipping zones of reflections in the deep crust that converge in the region of the Jemez lineament. These are interpreted in terms of a bivergent orogen and interwedging of crustal blocks, similar to those in the modern Alpine system (Schmid et al., 1996). Teleseismic images and receiver function interfaces suggest that the crustal suture extends downward into a zone of deep (to >150 km) lithospheric velocity contrast that may have formed initially as a hydrated subduction zone scar, then was reactivated during Cenozoic asthenospheric upwelling (Karlstrom et al., 2005). This zone also corresponds in position with several other indications of a crustal province boundary, i.e., change in Pb isotope signatures indicating different crustal compositions (Wooden and DeWitt, 1991), the southern limit of pre–1.7 Ga rocks at the surface (Karlstrom and Humphreys, 1998), and a zone of long-lived weakness, reactivation, and magmatism along the Jemez volcanic lineament (Aldrich 1986).

Mazatzal province rocks in Arizona include mainly supracrustal successions of the Sunflower and Pinal blocks (Figs. 10 and 11). The Mazatzal block contains a cover sequence that is of Mazatzal age (1.70–1.68 Ga) and a basement of Yavapai age (the 1.73 Ga Payson ophiolite and the 1.76 Ga granitoids it intrudes). The nature of the basement rocks to the Pinal and Sunflower blocks remains largely unknown. In the Mazatzal block, supracrustal rocks progress upsection from ophiolitic basement of the Payson ophiolite to 1.72 Ga East Verde River sequence turbidites, to mixed metavolcanics and metasediments of the Alder Group, to 1.70 Ga rhyolites of the Red Rock rhyolite to the Mazatzal Group quartzite (Cox et al., 2002). The southern boundary of the Mazatzal block, the subvertical Slate Creek shear zone, is a several kilometer wide zone of tectonic mélange (Roller, 1987) that, based on interpretation of magnetic anomalies, is the southwestward continuation of the Jemez lineament suture in New Mexico (Karlstrom et al., 2004).

In southern Arizona, Eisle and Isachsen (2001) identified an accretionary boundary between the 1.647–1.63 Ga rocks of the Cochise block east-southeast and 1.68–1.65 Ga Pinal block west and northwest of the proposed suture (Fig. 12). The boundary shear zone contains thick sections of pillow volcanics and composite dikes that may be fragments of an ophiolite suite within an accretionary prism that developed at a subduction-type continental margin (Swift and Force, 2001). Deformation took place between 1.678 and 1.655 Ga (Eisle and Isachsen, 2001). Nd model ages for the Pinal block are 1.8–2.0 Ga, more similar to the eastern Mojave province and apparently not explained by the observed age range of detrital zircons of 1.731–1.678 Ga, suggesting older crust in the subsurface. Nd model ages for the Cochise block are ca. 1.70 Ga, indicating that separation of crust from a depleted mantle took place <40 m.y. before crystallization of the felsic igneous rocks, which suggests that the Cochise block is juvenile (Eisle and Isachsen, 2001).

The Mazatzal orogeny involved ca. 1.65–1.60 Ga deformation that affected the 1.67–1.60 Ga rocks of the Mazatzal province and propagated northward into the previously assembled Yavapai basement (Bauer et al., 1993; Karlstrom and Bowring, 1993; Shaw and Karlstrom, 1999; Duebendorfer et al., 2001). Metamorphic and structural studies suggest that early metamorphism accompanied thrusting and shortening as supracrustal rocks were tectonically buried to depths of ~10 km. Based on overprinting relationships documented in several studies (Karlstrom, 1999; Ilg et al., 1996), we infer that structures evolved from premetamorphic low-angle thrusts (now preserved only in the lowest grade blocks; e.g., Mazatzal block; Doe and Karlstrom, 1991), to ductile thrust belts (such as the Needle Mountains in Colorado and the Manzano Mountains in New Mexico). At somewhat deeper levels (10 km, 3 kbar) in areas such as the Needle Mountains (Harris et al., 1987), early thrusts and recumbent folds are overprinted by penetrative northeast-trending subvertical foliation and shear zones that record northwest-southeast horizontal shortening and general shear and represent the dominant orogenic fabric in much of the southwestern United States

The metamorphic character and field gradients of Mazatzal-age metamorphism (ca. 1.65 Ga) remain poorly understood. The Alder Group, Red Rock Rhyolite, and Mazatzal Group of central Arizona were deposited ca. 1.70 Ga, and were deformed before 1.68 Ga (Labrenze and Karlstrom, 1991), intermediate between the Yavapai orogeny of the Grand Canyon (1.71–1.68 Ga) and 1.65–1.60 Ga Mazatzal deformation of New Mexico (Luther, 2006). The metamorphic grade of the Mazatzal orogeny in Arizona varies from very low (<350 °C; Gillentine et al., 1991) to ~580 °C (Williams, 1991). The Hondo Group in northern New Mexico and Colorado and other rocks that exhibit near Al-silicate triple point metamorphism were also deposited near the end of the Yavapai event on top of an angular unconformity over unroofed Yavapai basement (Jessup et al., 2005), and do not record Yavapai-aged tectonism. The main amphibolite facies metamorphism of these units, 1.45–1.35 Ga, overprinted and obscured the degree of thrust-related burial and metamorphism that took place in the Mazatzal orogeny. Thus, all available data suggest that Mazatzal-aged regional metamorphism may have reached upper greenschist facies, with slightly higher grades near syntectonic plutons. Mazatzal-aged tectonism in the mid-continent region involved the deformation of the 1.75–1.65 Ga Baraboo interval quartzites. This was interpreted by Holm et al. (2005) in terms of accretion of the Mazatzal arc, which caused south-verging folding of quartzites and mild reheating of Penokean crust to the north, as documented by Ar-Ar thermochronology.

Intense metamorphism and plutonism took place in southern Labrador during the approximate time of the Mazatzal orogeny (ca. 1.71–1.62 Ga; Fig. 11). This has been termed the Labradorian orogeny (Gower et al., 1992; Dickin, 2000). Labradorian plutonic rocks and associated anorthosites with slightly older Nd model ages can be traced from the eastern margin of Labrador along a southwest trend to the eastern shores of Georgian Bay (Ashwal et al., 1986; Scharer, 1991; Dickin, 2000). The extent of Mazatzal-Labradorian crust in the mid-continent is largely undefined, although (limited) Nd model ages (Van Schmus, 1976; Van Schmus et al., 1996, 2007) suggest continuity through

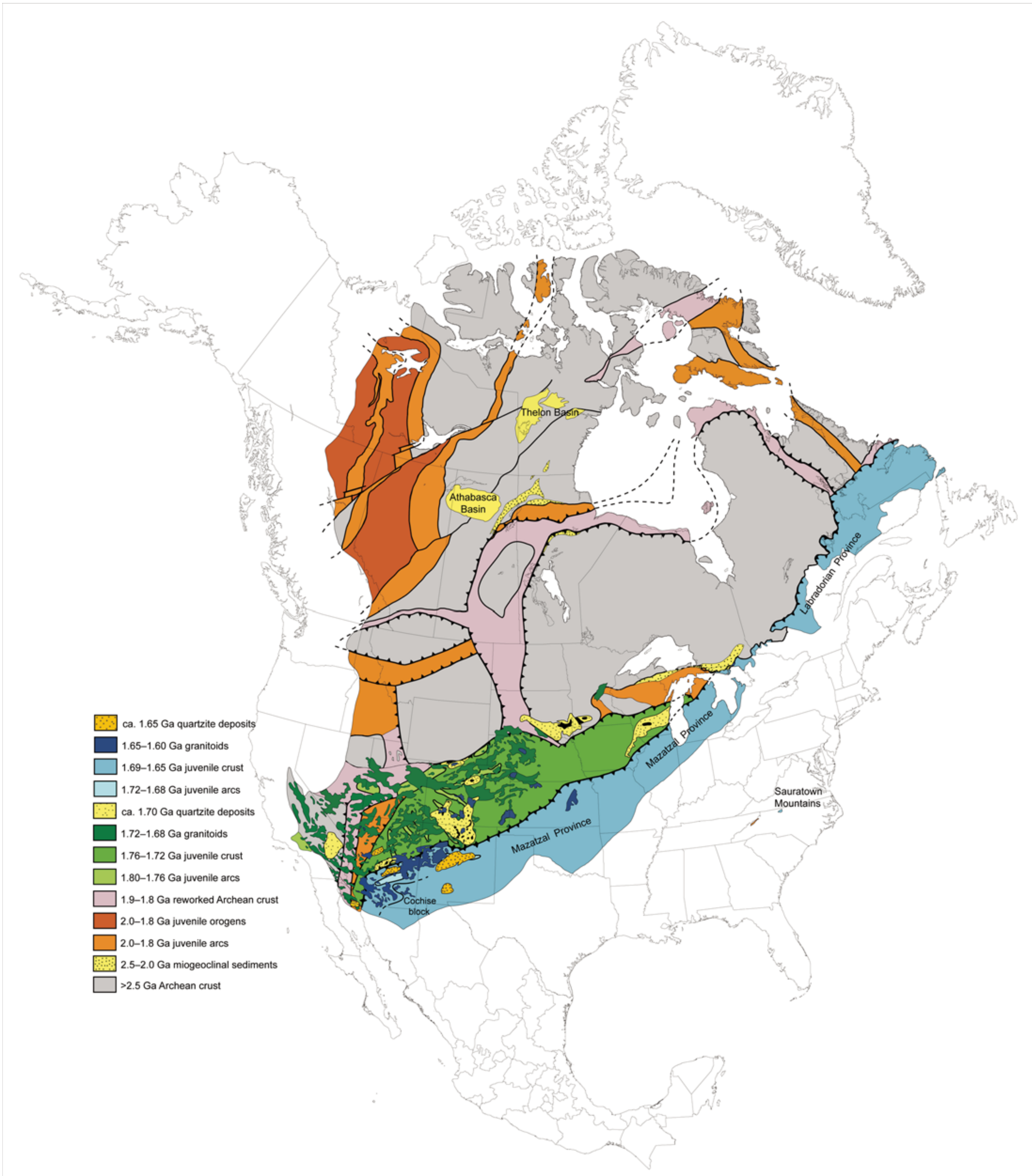

Figure 12. Mazatzal-age plutons (dark blue; ca. 1.65–1.60 Ga) stitched juvenile Mazatzal and older Yavapai crust. Quartzite-rhyolite successions (stippled yellow) stratigraphically occur at the upper end of the Mazatzal orogeny. Correlative basin deposits (Athabasca and Thelon basins; yellow) are found in the northern interior of the Canadian shield.

the Great Lakes–mid-continent region (Fig. 11). Mazatzal-age rocks also crop out within the Sauratown Mountains window in northwestern North Carolina (Fig. 11; Bream et al., 2004). Similar to other Proterozoic rocks within the Appalachian orogen (e.g., the Mars Hill terrane and other younger components mentioned in the following), provenance and Pb isotope analyses suggest a Gondwanan affinity (Hatcher, 1984; Hatcher et al., 2004). These basement blocks were likely not accreted to the Laurentian craton until the Paleozoic Appalachian orogen.

RHYOLITE-QUARTZITE ASSOCIATIONS: SYNTECTONIC COVER ASSEMBLAGES (1.70–1.65) BETWEEN OROGENIC PULSES

Across western Laurentia, from the Mojave province to the mid-continent, volcanogenic greenstone complexes of the Yavapai and Mazatzal provinces are unconformably overlain by supracrustal successions dominated by rhyolite, quartzite, and pelitic schist (Figs. 10–12). Examples include the 1.70–1.69 Ga Uncompahgre Group of Colorado (Harris, 1990), 1.69 Ga Hondo Group of New Mexico (Soegaard and Eriksson, 1985), Mazatzal Group of Arizona (Cox et al., 2002), and Baraboo and related quartzites of the mid-continent (Holm et al., 1998). These successions are unusually thick and chemically ultramature, first-cycle quartzarenite successions. Primary structures are locally well preserved and indicate fluvial to shallow-marine deposition on a southward-deepening siliciclastic shelf (Soegaard and Eriksson, 1985). Quartzites contain distinctive aluminosilicate horizons and Mn-rich layers (e.g., near the Vadito-Ortega contact) that contain Mn-andalusite and other Mn- and rare earth element–rich minerals (piemontite, gahnite), and appear to represent early hydrothermal alteration and/or exhalative deposits (Grambling, 1981).

The seemingly unique formation of chemically ultramature, first-cycle quartzites may be related to Earth's ocean-atmosphere system as it transitioned from a CO_2-H_2S–dominated system to one of oxygen, causing distinctive Proterozoic weathering conditions (e.g., Medaris et al., 2003) possibly involving high CO_2 and low pH. In addition, the absence of stabilizing plants, presence of microbial mats (Dott, 1983), depositional conditions involving extreme wind and water abrasion (Dott, 1983), diagenetic conditions involving deep laterite formation and removal of labile materials (Cox et al., 2002), and basin dynamics involving newly stabilized and weak lithosphere (Karlstrom et al., 2005) probably all contributed to development of the quartzites. The association with high-K, high-alkali, caldera-related rhyolites suggests that quartzite basins developed in extensional environments, perhaps due to slab roll back and extension of the newly stabilized lithosphere.

The quartzites range in age from 1.70 to 1.60 Ga and, both in New Mexico and Arizona, there is evidence that the tectonic regimes responsible for the quartzite-rhyolite deposition took place at least twice. Both the 1.70 Ga (Fig. 10) and the 1.65 Ga (Fig. 12) quartzite-rhyolite successions were clearly syntectonic in a regional sense; volcanism and deposition were occurring in these basins at the same time that plutonism and deformation were occurring in nearby regions or at deeper crustal levels (Karlstrom and Bowring, 1993). The lower unconformable contact of the 1.70 Ga quartzites at least locally represents late to post-Yavapai unroofing of middle crustal rocks (Gibson, 1990; Conway and Silver, 1989; Jessup et al., 2005). Basins that formed above this angular unconformity represent fundamental pauses and probable extensional events ca. 1.70 within the otherwise nearly progressive 1.8–1.6 Ga Yavapai-Mazatzal accretionary orogens. This transient extensional setting quickly returned to shortening during the Mazatzal orogeny, when the quartzites were tectonically buried to 5–20 km depths and deformed into some of the largest and most distinctive structures (folds, thrusts, shear zones) in the orogenic belt. A likely model for these basins involves roll back of a north-directed subducting slab, causing localized and short-lived extension of the newly developed upper plate lithosphere, then subsequent inversion of the basins during continued convergence (Giles et al., 2002; Holm et al., 2005; Betts and Giles, 2006).

MESOPROTEROZOIC TECTONISM: 1.55–1.35 Ga MAGMATISM, METAMORPHISM, AND DEFORMATION

Following a tectonic lull from 1.60 to 1.55 Ga, interpreted as a period of stabilization of North American lithosphere (Bowring and Karlstrom, 1990), juvenile terrane and arc accretion resumed along the lengthy southeastern margin of Laurentia. Van Schmus et al. (1996) delineated a significant mid-continental crustal boundary that extends from northwestern Mexico along a northeast trend to Ontario. The boundary divides crust with a Nd model age older than 1.55 Ga on the northwestern side from crust with a Nd model age younger than 1.55 Ga to the southeast, and may represent a major suture zone recording the collision of ca. 1.55–1.4 Ga juvenile crustal block against the south- and east-facing margin of Laurentia (the southeastern edge of the Mazatzal province). In the mid-continent of the United States, 1.55–1.3 Ga crust southeast of the Nd boundary has been termed the Granite-Rhyolite province (Lidiak et al., 1966; Bickford and Van Schmus, 1985). Similar-aged basement probably extends east of the Grenville deformation front, perhaps as far as the New York–Alabama lineament (Fig. 13). In eastern Laurentia, 1.55–1.3 Ga crust is bounded to the north by Labradorian (Mazatzal) basement, and to the south 1.55–1.3 Ga crust is interfingered with thrust slices of Elzivir and younger crust (Dickin and Higgins, 1992; Guo and Dickin, 1996; Dickin, 1998, 2000). Outliers of 1.55–1.35 Ga crust within the Appalachian orogen include the Goochland terrane (Owens and Samson, 2004) and the Toxaway and Tallulah Falls domes (Bream et al., 2004; Hatcher et al., 2004).

Bimodal A-type granites and associated anorthosites were intruded between 1.48 and 1.35 Ga within the Granite-Rhyolite province (Fig. 14) and are also extensively dispersed throughout Paleoproterozoic crust farther west (Van Schmus et al., 1996; Windley, 1993; Karlstrom and Humphreys, 1998). Although commonly termed anorogenic, there is increasing evidence for an orogenic link (McLelland et al., 1996; Corrigan and Hanmer, 1997; Karlstrom et al., 2001) involving both continental arc magmatism and the collision of juvenile rocks against the eastern and southern margins of Laurentia. In eastern and central Laurentia these rocks have for the most part been overprinted by the Grenville orogenic cycle and concealed by Paleozoic mid-continent cover, respectively. However, extensive magnetic highs in the mid-continent (North American Magnetic Anomaly Group, 2002) may represent extensive granitoid intrusions in covered basement rocks (Fig. 14). Good exposures of juvenile volcanic and intrusive rocks crop out in southeastern Canada (Dickin and Higgins, 1992; Gower and Tucker, 1994; Rivers, 1997), the St. Francois Mountains of Missouri (Bowring et al., 1992; Van Schmus et al., 1996), northwestern Texas (Mosher, 1998; Patchett, 1989; Barnes et al., 1999), southern New Mexico (Barnes et al., 1999), and parts of northern Mexico (Patchett and Ruiz, 1989).

Geochemistry of 1.48–1.35 Ga plutons in the southwestern United States ranges from peraluminous to metaluminous, even when plutons are the same age and in close proximity (Thompson and Barnes, 1999). This suggests that middle crustal granitic magmatism was strongly influenced by a heterogeneous lower crust. Petrogenetic models suggest that differentiation of tholeiitic magmas plus variable crustal contamination can produce the distinctive A-type chemistry and variability of these plutons (Frost

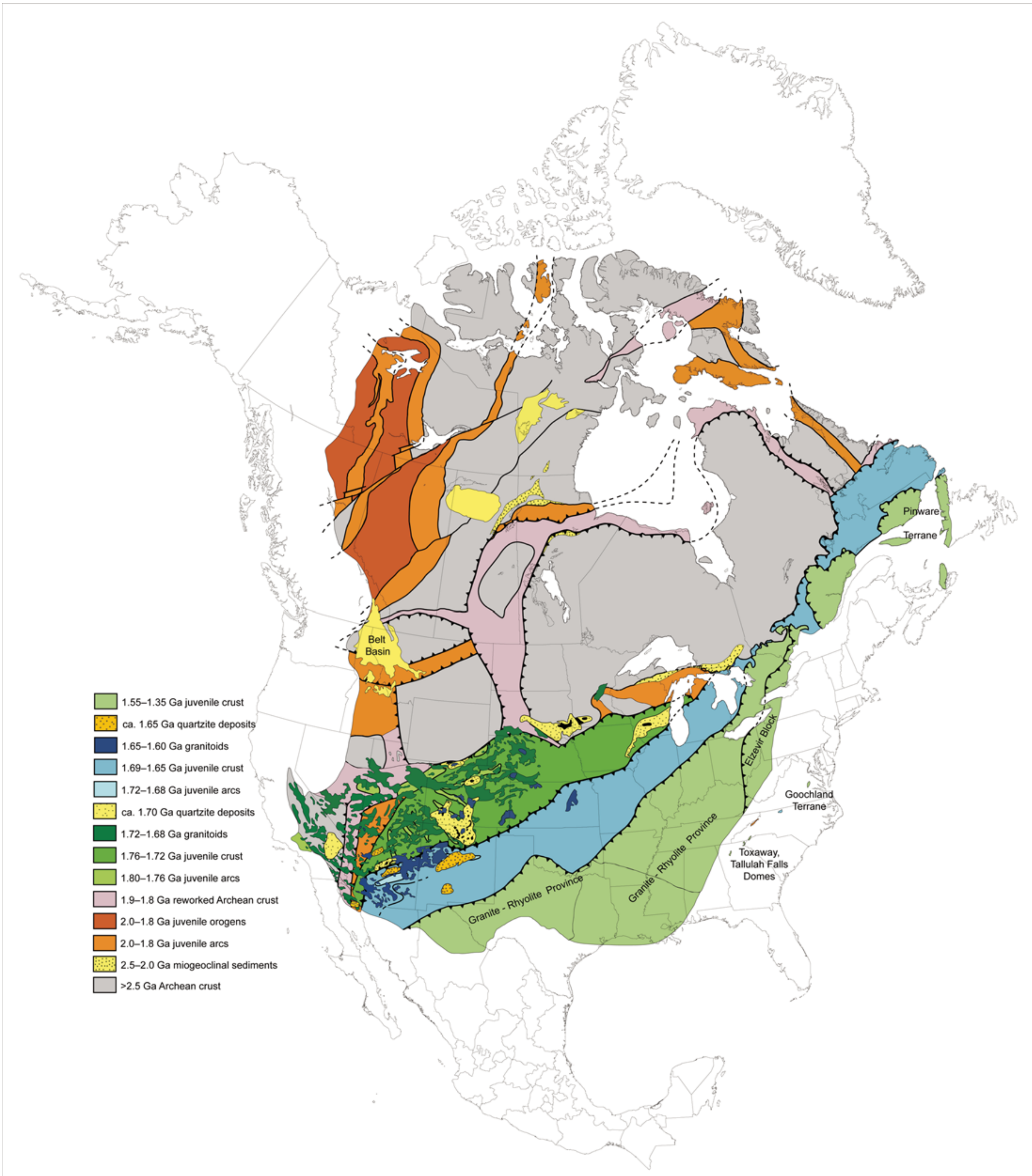

Figure 13. The Granite-Rhyolite province (ca. 1.55–1.35 Ga) is an extensive juvenile terrane extending from the southwestern United States through the mid-continent. Correlative blocks in the northeast include the Elzevir block and the Pinware terrane. Southern Appalachian outliers include the Goochland terrane and the Toxaway and Tallulah Falls domes.

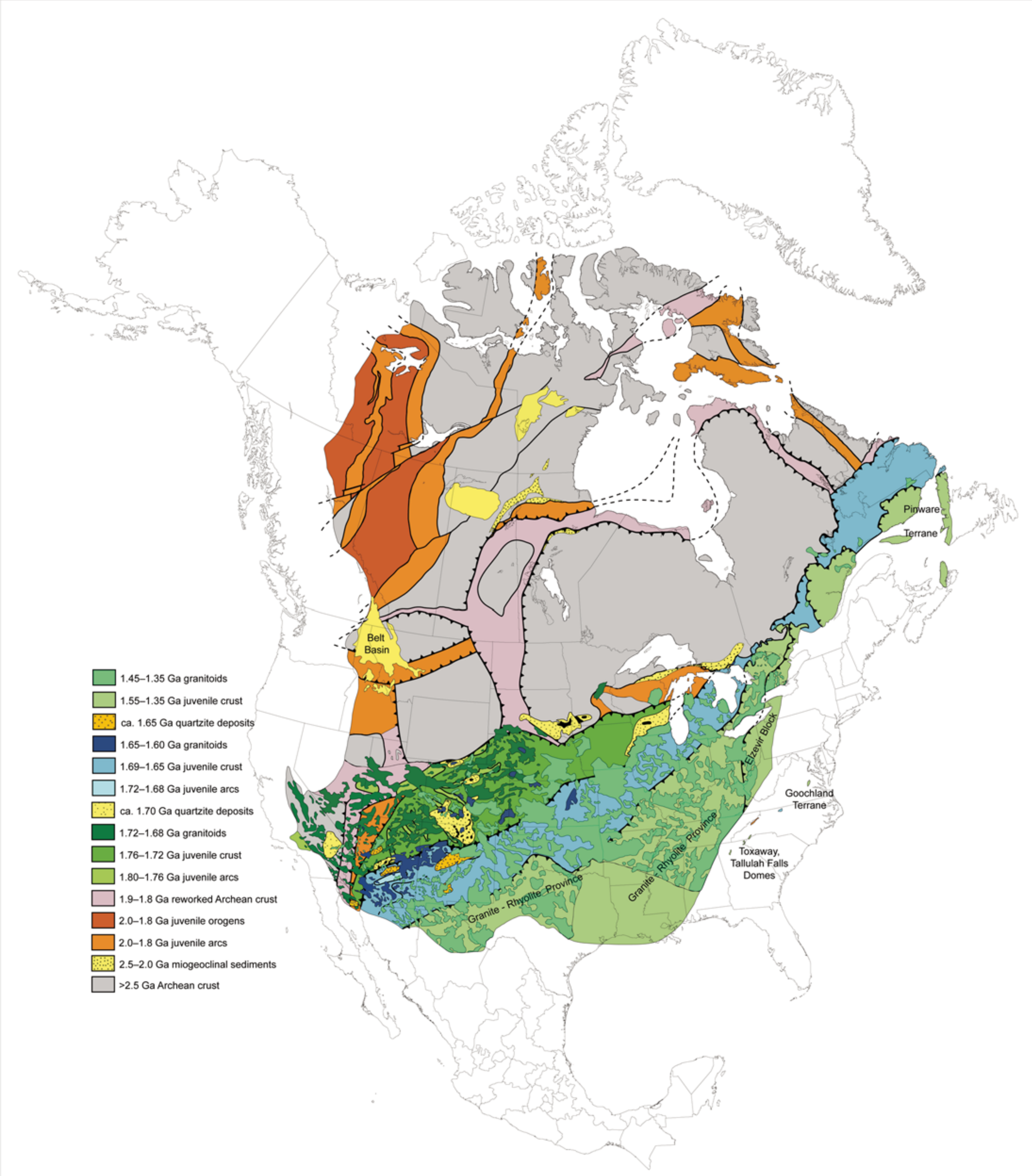

Figure 14. The most significant Proterozoic plutonic event that intruded southern Laurentia produced mostly A-type granitoids that stitch the Granite-Rhyolite, Mazatzal, and Yavapai provinces (ca. 1.48–1.35 Ga). Correlative with this is the formation of the extensive Belt basin in northern Idaho.

and Frost, 1997). In addition, preliminary seismic and xenolith results (Crowley et al., 2006) and Sm-Nd studies (Gonzales et al., 1994) support the hypothesis that a regional 1.4 Ga basaltic underplate, derived from mantle tectonism, may be a major constituent of the lower crust of the southwestern United States and part of the linked system of ca. 1.4 Ga crustal magmatism, metamorphism, and deformation (Karlstrom and Humphreys, 1998; Karlstrom et al., 2001). The extensive record of A-type magmatism, with sporadic activity over a huge area and a long interval (1.6–1.0 Ga), can be facilitated by a variety of tectonic settings. However, the onset of broadly simultaneous and bimodal intracratonic magmatism at some distance from the margin of southeastern Laurentia suggests that continental backarc processes controlled the morphology of the margin and provided the thermal energy required to produce magmas.

In the southwestern United States, 1.48–1.35 Ga deformation and regional metamorphism took place in areas that extended more than 1000 km inboard from the southern margin of Laurentia. Most areas show a component of northwest shortening, interpreted by Nyman et al. (1994) and Karlstrom and Humphreys (1998) to reflect far-field stresses that induced deformation in thermally softened rocks at large distances from a transpressive plate margin to the south. In many areas rocks remained at depths of ~10–15 km, as shown by the average metamorphic pressures of 3–4 kbar in aureoles of 1.4 Ga granites (Thompson et al., 1996; Nyman and Karlstrom, 1997; Read et al., 1999; Williams et al., 1999). One model is that the inferred shortening directions for this time period (west-northwest; Nyman et al., 1994) was due to transpressive deformation along preexisting structures within the now-intracratonic Paleoproterozoic orogenic belts (Karlstrom et al., 2001). Shaw et al. (2005) suggested that an orogenic plateau existed from 1.45 to 1.35 Ga and involved a 100 m.y. progressive intracratonic response to continued transpressive convergence along southern Laurentia. They speculated that topographically driven syncontractional extension may reconcile evidence for northwest contraction, regional reheating of the middle crust, evidence for basaltic underplating, and petrogenetic models requiring upwelling tholeiitic parental melts.

Major sedimentary sequences were deposited in Laurentia at 1.5–1.3 Ga, including the Belt-Purcell Supergroup of western Laurentia (Fig. 13) that accumulated tens of kilometers of sediment between 1.47 and 1.35 Ga (Evans et al., 2000), probably in several pulses of basin subsidence accompanied by episodes of mafic magmatism at 1.47, 1.455, 1.44, and 1.38 Ga (Chamberlain et al., 2003). Most of the Belt detritus came from western sources, with only minor input from the south (Ross et al., 1992; Ross and Villeneuve, 2003). The tectonic setting of the Belt basin remains controversial (Ross and Villeneuve, 2003); models include its origin as a rift basin (Evans et al., 2000; Chamberlain et al., 2003), intracratonic lake (Winston, 1986), delta (Cressman, 1989), impact basin (Sears and Alt, 1989), trapped ocean basin (Hoffman, 1988), and western North American transpressive margin, with the Belt basin analogous to a Caspian Sea–type pull-apart basin (Ross and Villeneuve, 2003).

GRENVILLE-AGE TECTONISM AND FINAL ASSEMBLY OF RODINIA: 1.3–0.9 Ga

Tectonism over a protracted period from ca. 1.3 to 0.9 Ga took place in Laurentia and on many other continents, and culminated in continent-continent collisions that facilitated the final assembly of Rodinia (Moores, 1991; Dalziel, 1991). For Laurentia this concluded an 800 m.y. history of accretionary orogenesis along an east- and southeast-facing, predominantly convergent margin (Fig. 15). The initial stage of the Grenville orogenic cycle, termed the Elzeverian orogeny by Moore and Thompson (1980), sutured the Elzevir and Frontenac blocks to the eastern margin of Laurentia ca. 1.3–1.2 Ga. Subsequent widespread intrusion of the anorthosite-mangerite-charnockite-granite (AMCG) plutonic suite from ca. 1.19 to 1.11 Ga likely resulted from orogenic collapse of overthickened crust (McLelland, 1996). The Ottawan orogeny (Moore and Thompson, 1980; Grenvillian orogeny of Rivers, 1997) spans the interval ca. 1.09–0.98 Ga and induced renewed northwest thrusting and imbrication of terranes in southeastern Canada. Deeper crustal levels are exposed in the Adirondack Mountains and Eastern Grenville terranes, where the effects of Ottawan orogenesis are principally high-grade metamorphism, plutonism, and ductile folding. This was the culminating event of the Grenville orogenic cycle and probably the final stage in the assembly of Rodinia, prior to continental breakup from ca. 0.78 to 0.535 Ga.

The Grenville deformation front marks the western extent of Grenville-age deformation and extends southwest from the coast of Labrador, through eastern Michigan, and then south through northwestern Alabama (Fig. 15). Within the mid-continent region the Grenville front bisects 1.5–1.33 Ga rocks of the eastern Granite-Rhyolite province. North of the Great Lakes, the Grenville front is east of the Sm-Nd line of Van Schmus et al. (1996), and trends through Mazatzal-age and Labradorian crust.

The Elzevirian orogeny, from 1.3 to 1.2 Ga, comprised outboard and backarc magmatism (Moore and Thompson, 1980), during which the Elzevir and Frontenac terranes were amalgamated against the southeastern margin of Laurentia with consequent collision, deformation, and high-grade metamorphism. The Elzevir and Frontenac blocks (along with the Central Granulite terrane) could be underlain by 1.5–1.45 Ga basement that rifted off the eastern margin of Laurentia during the period 1.45–1.35 Ga (McLelland et al., 1996), although no direct evidence of basement older than 1.35 Ga has been documented. The type Elzevirian arc is a cal-alkaline backarc or marginal basin system (Smith and Holm, 1987), the Frontenac block being the eastern (passive?) margin of the rifted Elzevir terrane (Davidson, 1995; McLelland et al., 1996). Low-grade rocks of the Frontenac block do not record the amphibolite facies metamorphism of the Elzivir block, which suggests that the Frontenac block was significantly shallower in the crust throughout the Grenville orogenic cycle. The Bancroft terrane, which consists of low-grade marginal rocks along the western margin of the Elzivir block, may be an allochthonous slice of the Frontenac block that was thrust over the Elzivir block during Elzevirian or Ottawan orogenesis (Davidson, 1995). Nonetheless, the present-day close spatial juxtaposition of these blocks was likely due to Elzevirian imbrication, with continued shortening during the Ottawan orogeny.

The Central Granulite block extends south from central Quebec through the Adirondack Highlands, south of which Grenville basement is overlain and obscured by deformed Paleozoic rocks of the Appalachian orogenic cycle. Tonalitic and high-silica granitoids (ca. 1.35–1.3 Ga) predominate within the Adirondack Highlands and extend northward into Quebec. No evidence of older basement has been documented, although 1.3 Ga granitoids intrude thick quartzite sequences in places, which suggests that older basement may underlie the Adirondack arc (Daly and McLelland, 1991). McLelland et al. (1996) suggested that 1.5–1.33 Ga basement extends beneath the Central Granulite block and as far east as the Green Mountain massif, the northernmost extent of the Eastern Grenville terranes. If true, the Central Granulite block was presumably another piece of rifted ca. 1.5 Ga crust that was reaccreted to the eastern margin of Laurentia during the Elzevirian orogeny.

The emplacement of large anorthosite and AMCG plutons followed Elzivirian orogenesis from ca. 1.19 to 1.11 Ga (Fig. 16), as overthickened crust underwent delamination and extensional collapse (Corrigan and Hanmer, 1997; McLelland et al., 1996). This period has

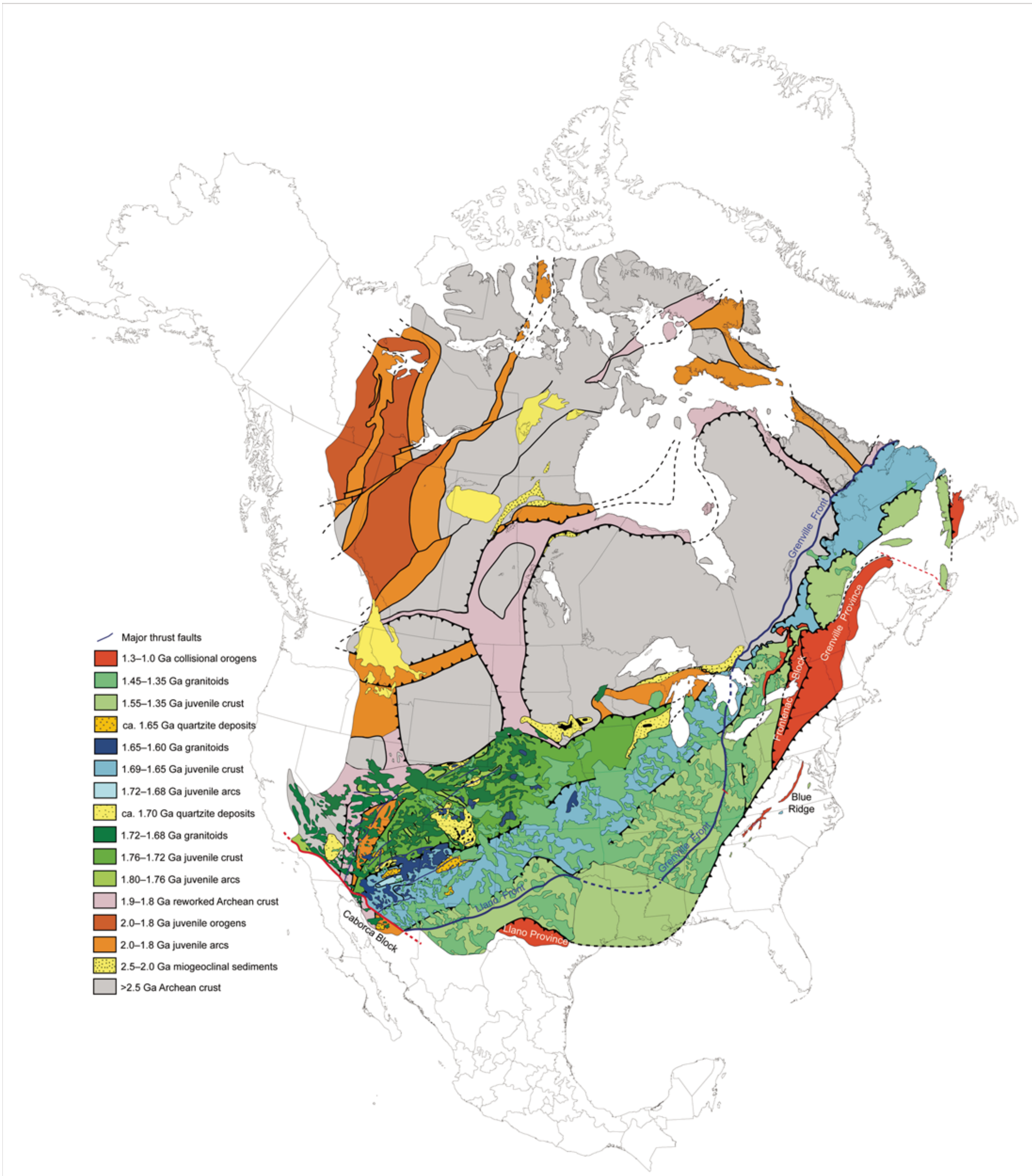

Figure 15. The Grenville and Llano provinces (ca. 1.3–1.0 Ga) represent the culminating continent-continent collisions that assembled Rodinia. Appalachian outliers are found along the length of the Blue Ridge. Deformation extended as far inland as the Grenville and Llano Fronts (dark blue line). Regionally significant transcurrent faulting included the southeastward transfer of the Caborca block (southwestern U.S.) along the Mojave-Sonora megashear.

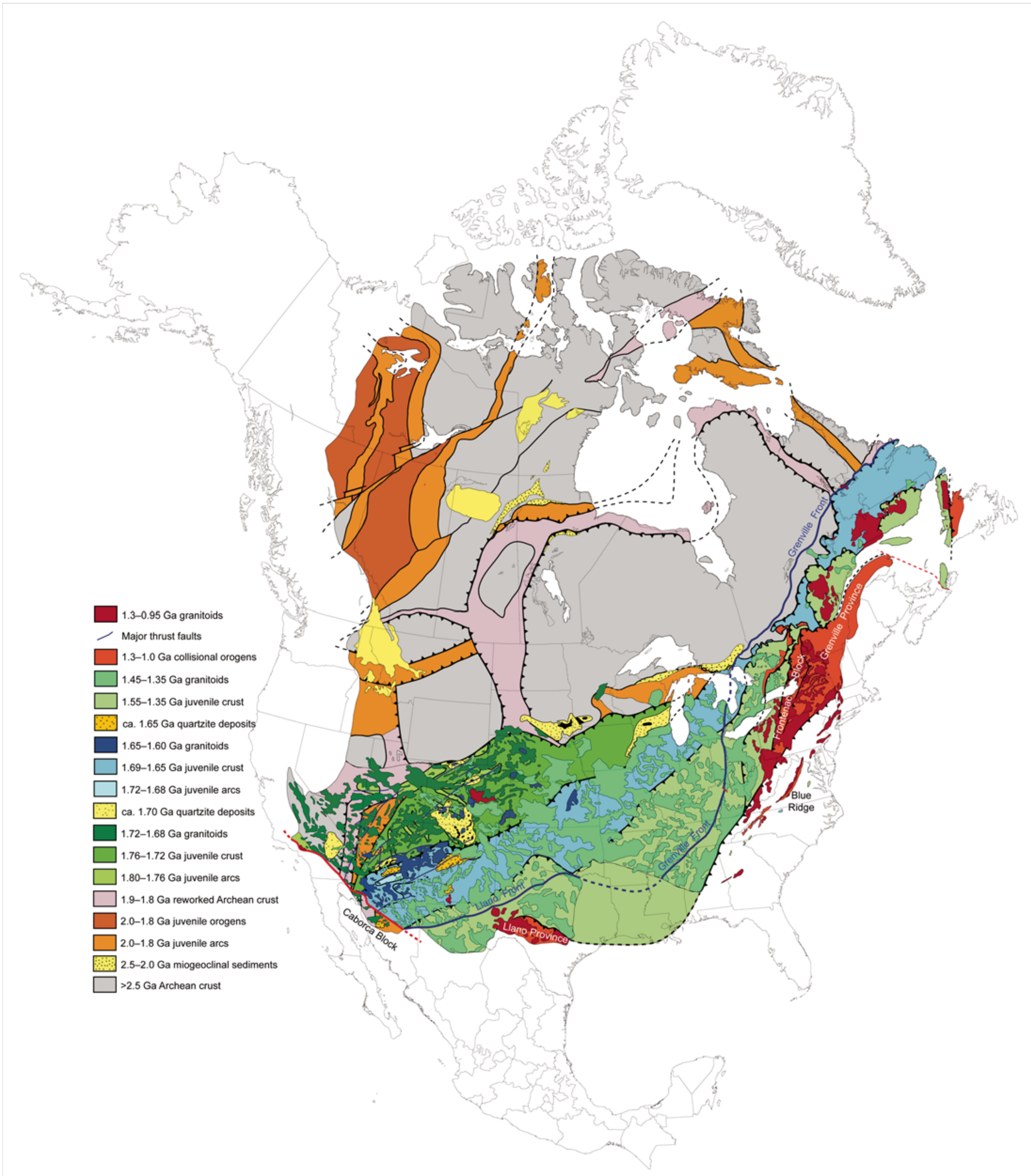

Figure 16. Late Grenville granitoid intrusions (dark red; ca. 1.30–0.95 Ga) stitched juvenile Grenville and older terranes. Grenville-age plutons are found as far west as Colorado.

been called the collisional Shawingian phase of the Grenville orogeny in eastern Canada (Rivers, 1997). However, within the Adirondack Highlands and eastern Grenville terranes of the United States, AMCG plutonism seems to have occurred during a relatively quiescent, extensional period prior to renewed convergence during the Ottawan orogeny. Voluminous anorthosite massifs and AMCG granitoids are a prominent component of almost all exposures of Grenville-related rocks, from southeastern Canada through northern Georgia (Davidson, 1995; McLelland et al., 1996; Aleinikoff et al., 2000). It is reasonable to suspect that these intrusive rocks are equally as prominent within eastern Laurentian basement that currently underlies deformed Paleozoic rocks of the Appalachian orogenic cycle; if so, the Grenville AMCG event is of a scale similar to that of the ca. 1.4 Ga A-type granitoid intrusional event, and suggests that similar tectonic environments may have existed along the then-eastern margin of Laurentia during both of these periods.

The Ottawan orogeny (ca. 1.09–1.03 Ga; Moore and Thompson, 1980) involved widespread deformation along the eastern margin of Laurentia, when major collisions appear to have taken place with one or more large continental masses to the southeast. Possible candidates include Amazonia (Hoffman, 1991; Dalziel, 1997), the Rio de Plata craton (Dalziel, 1997; Meert and Torsvik, 2003), and/or the Kalahari craton (Dalziel et al., 2000; Loewy et al., 2003). During this collisional event large thrust slices driven to the northwest produced imbrication of much of the Canadian Grenville province (Davidson, 1995). In deeper, more ductile regimes, such as the Adirondack Mountains of New York and the eastern Grenville terranes, extensive folding helped accommodate contraction (McLelland et al., 1996). Amphibolite to granulite facies metamorphism took place in almost all areas of the orogen. Metamorphism in most of the province peaked ca. 1.07 Ga (McLelland et al., 2001), but high-grade conditions continued locally until 0.98 Ga (Scharer et al., 1986), the time at which movement ceased along the Grenville deformation front (Haggart et al., 1993). Late- to post-tectonism plutonism is abundant throughout the Central Granulite block and eastern Grenville terranes. Late syntectonic intrusions occurred in the Grandfather Mountain window of North Carolina ca. 1.08 Ga (Carrigan et al., 2003), and in the northern Blue Ridge of Virginia beginning 1.07 Ga (Aleinikoff et al., 2000). Post-tectonic, intracratonic leucogranites intruded from 1.05–1.04 Ga (Berkshire massif of Massachusetts and northern Blue Ridge of Virginia) to 1.02–1.01 Ga (Hudson Highlands of New York; Aleinikoff et al., 2000). Post-tectonic plutonism related to the Ottawan orogeny concluded ca. 0.96 Ga with intrusion of the Mt. Holly complex of the Green Mountains, Vermont (Ratcliffe et al., 1991).

Grenville-age (ca. 1.30–1.0 Ga) and older terranes extend from Vermont through northern Georgia, east of the New York–Alabama lineament, and crop out as isolated basement blocks amid multiply deformed rocks of the Appalachian orogenic cycle. Elzivirian-related (ca. 1.3–1.25 Ga) oceanic arc rocks crop out along the southern extent of the Adirondack Highlands and are also prominent within the Green Mountain and Berkshire massifs of Vermont and western Massachusetts, respectively (McLelland and Chiarenzelli, 1990; Ratcliffe et al., 1991; McLelland et al., 2001). This suggests that an island arc system developed outboard of the main Elzivirian continental arc early in the Grenville orogenic cycle and was accreted to the Laurentian mainland during the Elzivir or Ottawan orogenies. Evidence for Elzivirian orogenesis farther south is restricted to metavolcanic rocks of the Baltimore Gneiss (ca. 1.25 Ga; Aleinikoff et al., 1997) and supracrustal rocks of the Sauratown Mountains window (ca. 1.23 Ga; McConnell, 1990) and eastern Blue Ridge (Carrigan et al., 2003) in North Carolina.

The eastern Grenville terranes have been interpreted as rifted fragments of Laurentia that were reattached to the mainland during Appalachian orogenic events (Hatcher, 1984; Carrigan et al., 2003). This would suggest that their basement could be composed of 1.5–1.33 Ga rocks, equivalent to the proposed basement of the Elzivir and Frontenac blocks (McLelland et al., 1996). However, the Mars Hill terrane of North Carolina contains 1.8–1.6 Ga rocks (Fullagar and Gulley, 1999; Carrigan et al., 2003), and pre–1.5 Ga basement has also been proposed for parts of the Blue Ridge of Virginia (Hatcher, 1989; Hatcher et al., 2004). The closest Laurentian terrane older than 1.55 Ga is the Mazatzal province, thousands of kilometers to the west. Therefore, these blocks of Proterozoic basement are either far-traveled fragments from west of the Granite-Rhyolite province, or more likely they were originally exotic to Laurentia (e.g., West Africa or South American Proterozoic terranes) and accreted during Appalachian orogenesis.

The Llano front, on the west side of the Texas embayment (Thomas, 1991), is considered to be a western extension of the Grenville deformation front. The Llano front is defined on the basis of gravity anomalies (Adams and Keller, 1995) and deformational style, with deformed rocks situated south of the front (Mosher, 1998). A boundary defined by Nd model ages is located north of the deformational front; Paleoproterozoic Nd model ages (1.74–1.64 Ga) are spatially restricted to north of the front in northern Texas. Younger 1.50–1.35 Ga Nd model ages occur south of the isotopic boundary and on both sides of the deformational front (Barnes et al., 1999). This is similar to the Grenville front in the mid-continent, in that the Llano front is a 1.3–1.0 Ga deformational boundary developed within juvenile 1.55–1.30 Ga crust, not a ca. 1.1 Ga crustal age boundary.

The nature of the Grenville-aged orogenic cycle in Texas is subtly different from its character in the eastern United States. Significant exposures of Llano province rocks include the Llano uplift and the Van Horn outcrops in western Texas. In the Llano area, the rocks are polydeformed and have been metamorphosed to upper amphibolite–granulite facies. The northern part of the uplift contains 1.37 Ga rocks of North American affinity, based on age, composition, and Nd isotopes, and may be a southern extension of the Granite-Rhyolite province (Reese et al., 2000). These rocks are tectonically interleaved with 1.33–1.26 Ga backarc and serpentinite assemblages that were shortened and thrust onto the Laurentian margin. The continent-continent collision that imbricated these packages culminated from 1.23 to 1.12 Ga, and late-tectonic plutonism continued until 1.07 Ga (Reese et al., 2000).

Exposures in the Van Horn and Franklin Mountains document Grenville deformation in western Texas between ca. 1.4 and 1.0 Ga (Bickford et al., 2000). Rift or backarc deposition occurred between 1.38 and 1.33 Ga, roughly coeval with continued granitoid magmatism in the southwestern United States and similar to early stages of the Elzivirian orogeny in northeastern Laurentia. At 1.26–1.25 Ga sedimentary and volcanic sequences were deposited in extensional basins along the southern margin of Laurentia and in intracratonic basins within Laurentia. Many of these rift basins are associated with transcurrent tectonism (Soegaard and Callahan, 1994). Deformation possibly related to the Ottawan orogeny is limited to transpressional thrusting at 1.035 Ga that deformed rift fill sequences in the footwall, beneath overriding ca. 1.35 Ga volcanic and metasedimentary rocks.

Magmatism ca. 1.1 Ga was regionally bimodal within the Paleoproterozoic crust, reminiscent of the ca. 1.4 Ga A-type event. The coeval Red Bluff granite in the Franklin Mountains and the Pikes Peak granite of Colorado are both similar in composition and age and are interpreted to represent fractional crystallization from a mildly alkaline basalt parent (Smith et al., 1999). Several tectonic settings have been proposed for 1.1 Ga bimodal magmatism and, as for the ca. 1.4 Ga A-type plutons, basaltic underplating is a likely cause. Barnes et al. (1999) favored an extensional setting, but the overlap in age between basaltic magmatism (1.16 Ga for the Pecos complex and 1.1 Ga for Keweenawan and diabase dikes of the Southwest), granitic magmatism (1.12 Ga in the Franklin Mountains, 1.09 Ga at Pikes Peak, 1.20–1.10 Ga in the Llano uplift), and deformation and metamorphism (1.20–1.10 Ga in the Llano uplift) suggests that any extensional tectonism should be viewed within the overall context of regional Grenville convergence, as suggested by Howard (1991).

GRENVILLE-AGE INTRACRATONIC EXTENSION

During the Grenvillian orogeny, northwest-directed contraction at the southern margin of Laurentia was accompanied by intracratonic extension and voluminous mafic magmatism (e.g., the 1.27 Ga MacKenzie and Sudbury mafic dike swarms, the 1.1 Ga Midcontinent rift, and 1.1 Ga Central Basin platform and diabase sheets in Arizona; Fig. 17). Intracratonic basins formed from 1.25 to 1.10 Ga in New Mexico and Colorado (the De Baca Group and Los Animas Formation, mainly known from the subsurface), similar to deposition and normal faulting of the 1.25–1.10 Ga Unkar and 1.35–1.10 Ga Apache Groups in Arizona (Wrucke, 1989; Timmons et al., 2001). During the same period a set of northwest-striking extensional faults formed over much of Laurentia. These were influential in the creation and/or subsequent reactivation of important post-Rodinia lineaments such as the Mojave-Sonora megashear (Fig. 15), the Texas-Mogollon-Walker lineament, the Uncompahgre lineament, the Oklahoma aulacogen, and the Lewis and Clark–Oklahoma–Alabama lineament (Thomas, 1991; Marshak and Paulsen, 1996; Marshak et al., 2000; Timmons et al., 2001).

Grenville-age extension likely facilitated at least a portion of the exhumation of western United States basement rocks that were at depths of ~10 km at the end of the 1.45–1.35 Ga event. Exhumation may have taken place by erosion of an elevated 1.4 Ga plateau (Karlstrom and Morgan, 1995; Karlstrom and Humphreys, 1998; Shaw et al., 2005), with pulses of differential uplift along extensional faults at 1.25 Ga, 1.1 Ga, and 0.8 Ga (Timmons et al., 2001), such that middle crustal rocks across the region had been exhumed too close to the surface by Cambrian time to create the Great Unconformity (Powell, 1876).

BREAKUP OF RODINIA

Existing models for the breakup of Rodinia (Moores, 1991; Dalziel, 1991; Karlstrom et al., 1999; Li et al., 2007) indicate a diachronous disassembly, with early stages of rifting along the western margin of Laurentia occurring between 0.78 and 0.68 Ga (Fig. 18), followed by the main pulse of rifting along the eastern margin of Laurentia between 0.62 and 0.55 Ga (Fig. 19). Although there is considerable uncertainty about the identity of the western conjugate continents (e.g., Sears and Price, 2000; Meert and Torsvik, 2003; Li et al., 2007), rifting along the western margin of Laurentia is considered to have separated the western continents (Australia, Antarctica, south China, Siberia) from Laurentia, thereby opening a paleo–Pacific Ocean. Intrusion of the Gunbarrel dikes (Gunbarrel mafic magmatic event of Harlan et al., 2003b) from the northern Canadian territories to northwestern Wyoming occurred during this period, between 0.782 and 0.775 Ga. Mafic intrusion along the western margin of Laurentia also included ca. 0.779 Ga dikes in the northern Canadian shield (LeCheminant and Heaman, 1989) and the Franklin igneous event at 0.723 Ga (Heaman et al., 1992), all of which suggest a major extensional event during this period. Farther north, along the paleo-Pacific margin of Canada, synrift to passive margin sequences of the Windermere Supergroup constrain breakup to the Neoproterozoic (after 0.78 Ga; Ross, 1991). In the Grand Canyon region, east-west extension and synrifting deposition of the 0.800–0.742 Ga Chuar Group was possibly an inboard record of rifting at the western plate margin (Timmons et al., 2001; Karlstrom et al., 2003). Thermal subsidence of the Cordilleran miogeocline, hence drift of the last western conjugate continent, may have taken place ca. 0.57 Ga (Bond et al., 1985; Colpron et al., 2002). Evidence for synrift magmatism at 0.685 Ga, rifting in central Idaho (Lund et al., 2003), and 0.57 Ga rifting in southern Canada and the western United States (Colpron et al., 2002) reinforces the interpretation for multistage rifting in western Laurentia.

There is robust evidence that a failed rifting event occurred along the eastern margin of Laurentia during the same period as the successful rifting of western Rodinia. Rift-related igneous activity between 0.76 and 0.70 Ga is documented in the southern Appalachians (e.g., Mount Rogers Formation; Su et al., 1994; Aleinikoff et al., 1995). However, complete breakup along the eastern margin of Laurentia probably did not initiate before ca. 0.62 Ga. Abundant ages for synrift igneous intrusions have been documented along the eastern Laurentia rift margin, from Newfoundland (0.62–0.57 Ga; Williams et al., 1985; Cawood et al., 2001) through the southern Appalachians (0.62–0.55 Ga; Aleinikoff et al., 1995; Owens and Tucker, 2003). Structural evidence of low-angle, east-directed detachments, synrift sedimentation, and evidence for subsequent thermal uplift also suggest significant extension during this period (Thomas, 2006). Earliest Cambrian post-rift sedimentary sequences above rift-related unconformities document the rift to drift transition and opening of the Iapetus ocean.

The final stage of breakup along the eastern margin of Laurentia involved the rifting of the Argentina Precordillera terrane from the Ouachita embayment region of southeastern Laurentia (Fig. 14; Thomas, 1991; Thomas and Astini, 1996). Grenville ages and Pb isotopic ratios of basement rocks in the Precordillera terrane of western Argentina indicate a Laurentian origin (Kay et al., 1996) and equate with structural and stratigraphic controls on the Early Cambrian (ca. 535 Ma) rifted margin of the Ouachita embayment (Thomas, 1993; Thomas and Astini, 1999). The subsequent transfer of the Precordillera terrane across the Iapetus ocean and accretion to the western margin of Gondwana (Ramos, 2004) necessitate relocation of the active Iapetus spreading ridge from (present day) central Alabama to a parallel north-trending zone in central Texas (Fig. 20). The mechanisms for this ridge jump are not fully understood; however, the coincident formation of the Realfoot rift and Oklahoma aulacogen (including intrusion of bimodal igneous rocks; Thomas, 2006) suggests at least two incipient triple junctions in the southeastern United States in the Early Cambrian that combined to initiate a new spreading ridge–transform system.

DISCUSSION AND IMPLICATIONS

Our model for Proterozoic assembly of North America is based on the paradigm that plate tectonics operated in the Proterozoic in essentially the same way as today in terms of mobility of plates and subduction-related processes (Cawood et al., 2006), although heat regimes, rates of plate motion, and convection dynamics may have differed (Condie, 2005). This concept has been debated, especially for times before 2.5 Ga (e.g., Hamilton, 2003; Kerrich and Polat, 2006), but we follow the prevailing view that plate tectonic models have been extremely successful at explaining diverse data sets from Precambrian rocks (Windley, 1995). The scale of ocean basins that closed to form observed orogenic belts, as with many Phanerozoic examples, remains elusive,

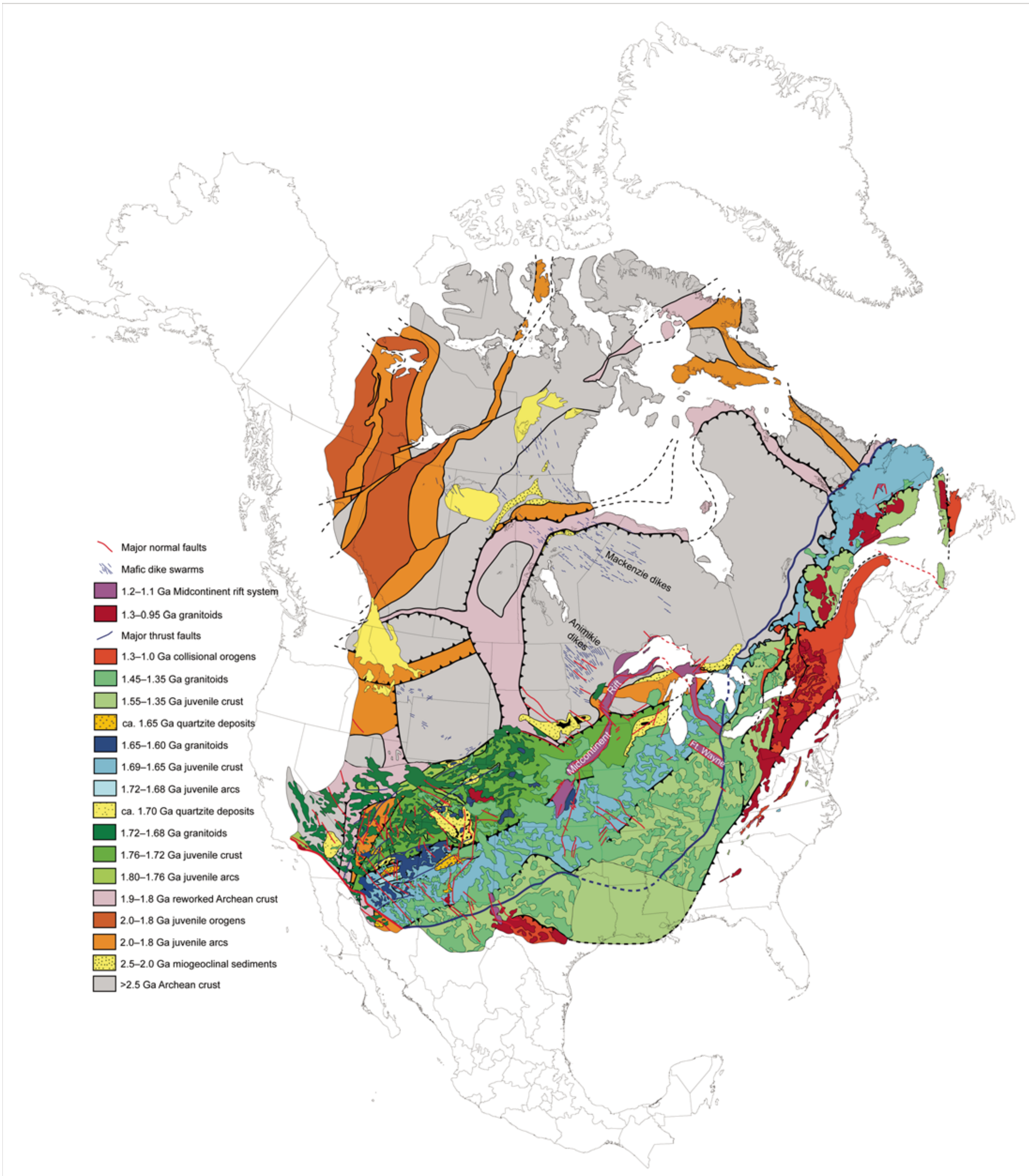

Figure 17. Coincident with Rodinia assembly (ca. 1.2–1.1 Ga), intracontinental extension occurred along the Midcontinent rifts (including the Keweenawan and Fort Wayne rifts). Regionally significant extensional faults (red lines) occur throughout southern Laurentia. Extensive intrusions of mafic dikes (dark blue lines) occurred west and south of Hudson Bay (Mackenzie dikes) and in the Minnesota region (Animikie dikes).

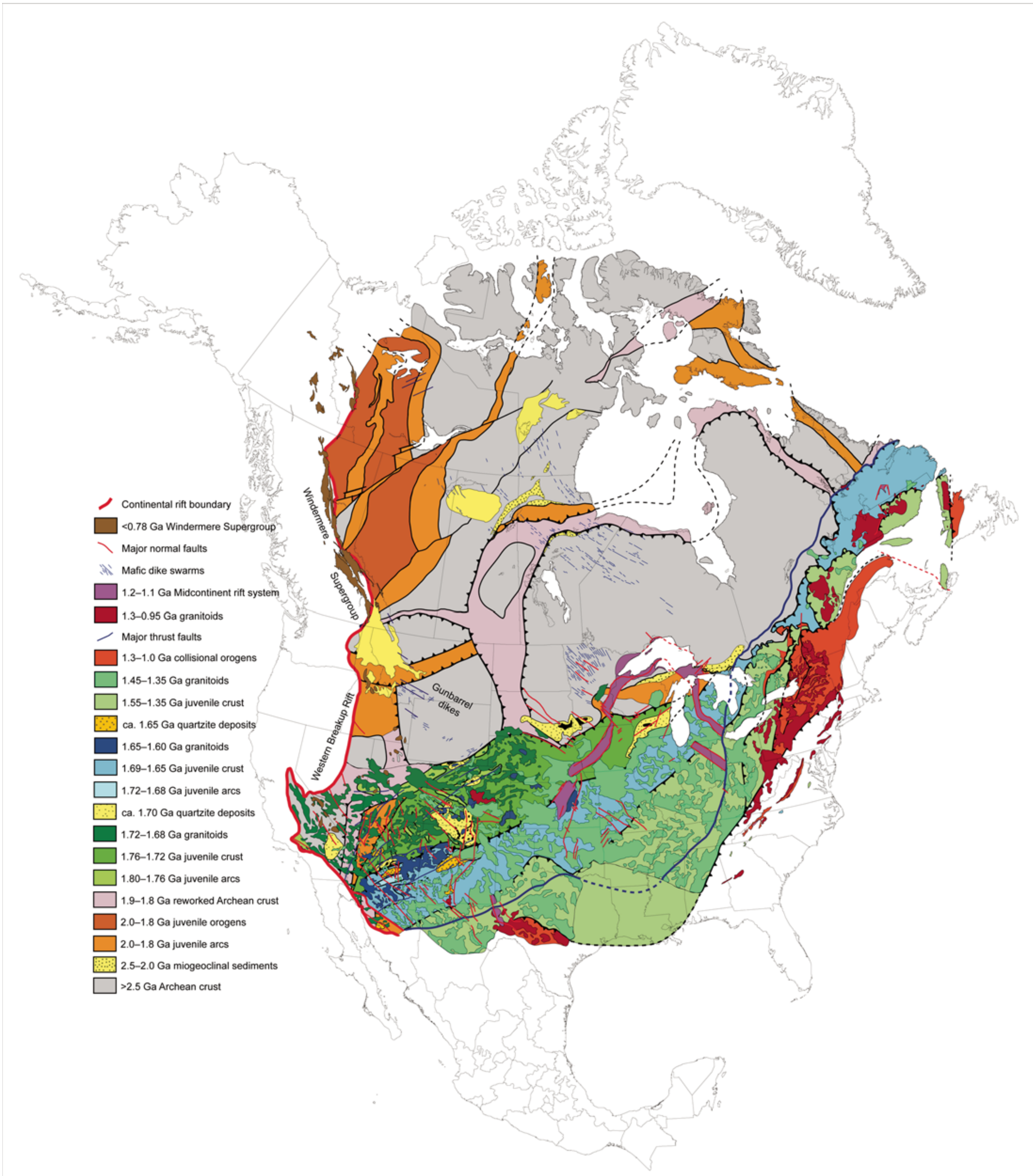

Figure 18. Early stages of the breakup of Rodinia occurred along the west coast of Laurentia (ca. 0.78–0.68 Ga). Evidence for west coast extension included intrusion of the Gunbarrel dikes (bold blue lines) and deposition of the Windermere Supergroup (brown). Bold red lines show rift boundaries along the western margin of Laurentia.

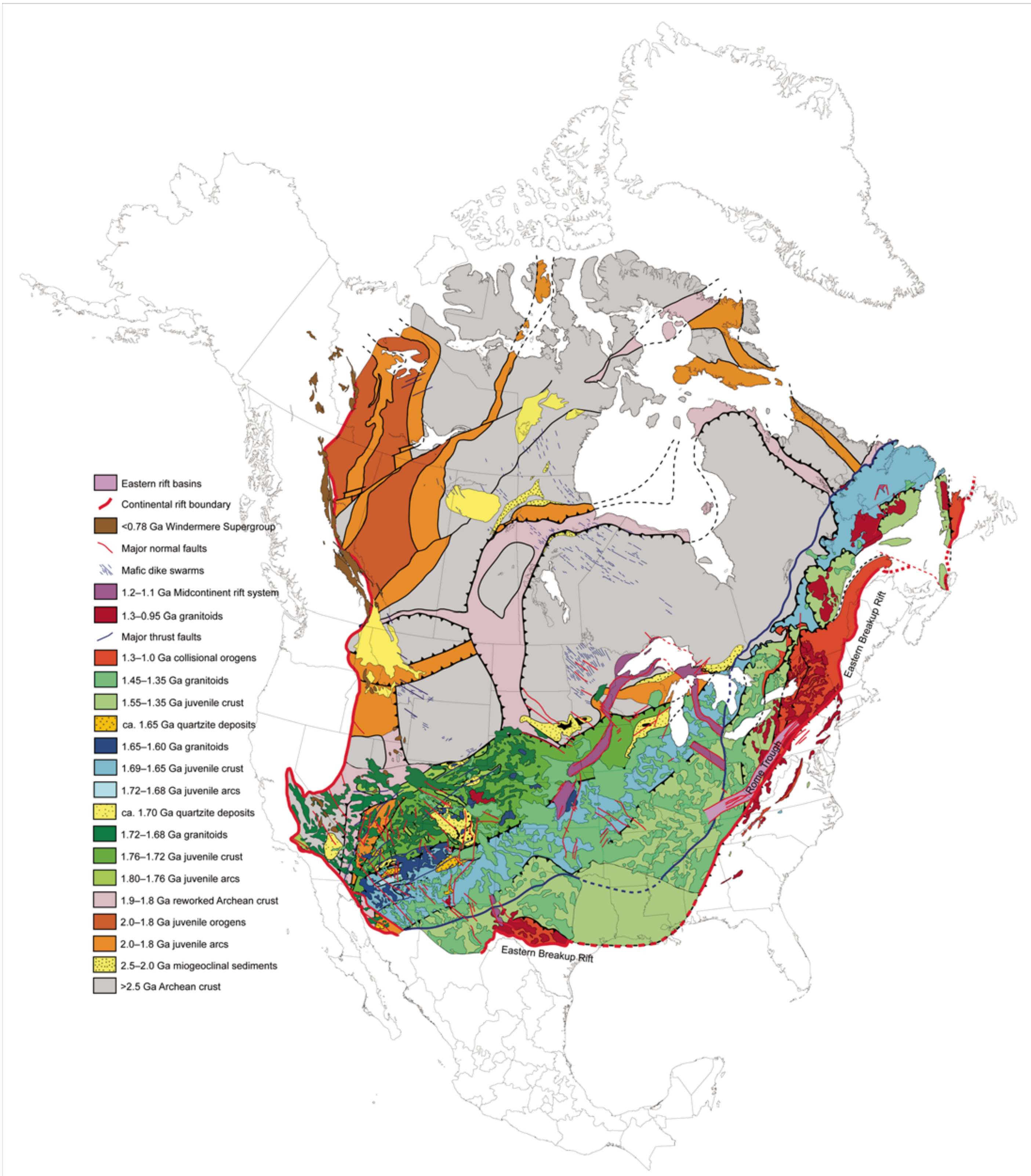

Figure 19. Following failed rifting between 0.720 and 0.68 Ga, the main phase of Rodinia breakup along the eastern margin of Laurentia (ca. 0.62–0.55 Ga) created the Iapetus ocean. Related rift basins include the Rome trough (light purple). Proterozoic basement terranes of the Appalachian Mountains were originally on the eastern (Gondwana) side of the rift, and were reaccreted to Laurentia during the Paleozoic Appalachian orogenic cycle (see text for details). Bold red lines show rift boundaries along eastern margin of Laurentia.

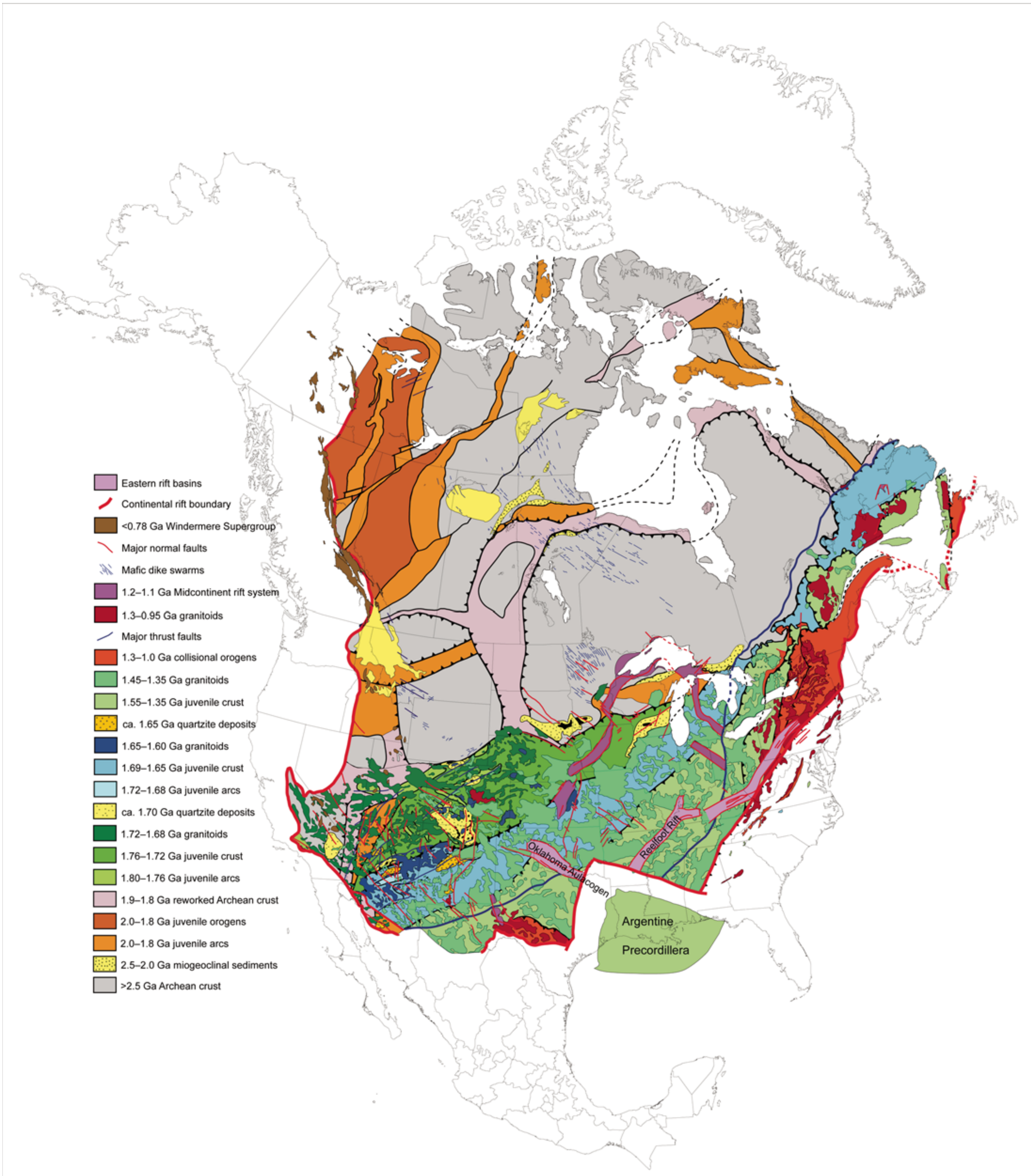

Figure 20. The final phase of eastern rifting detached the Argentine Precordillera microcontinent from the Ouachita (Texas) embayment region of southern United States (ca. 0.535 Ga). The Precordillera terrane is currently located in western Argentina. Associated failed rift arms include the Reelfoot rift and Oklahoma aulacogen (light purple). Bold red lines show rift boundaries.

but ophiolites are well documented from the Trans-Hudson orogen (Scott et al., 1992) and the Yavapai province (Dann, 1997). Seismic images provide good evidence for the existence of subduction scars that are interpreted to represent paleosuture zones in Archean (Calvert et al., 1995) and Proterozoic orogens of northern Laurentia (Cook et al., 1999; Korja and Heikkinen, 2005). In southern Laurentia, seismic images (from the CDROM Experiment; Karlstrom et al., 2001, 2005) indicate complex paleosuture geometries at the Cheyenne belt and across the Jemez lineament and have led to models involving tectonic wedging and changing subduction polarities (Tyson et al., 2002; Magnani et al., 2004). These studies present models where oceanic tectonic elements (island arcs, backarc basins, seamounts, accretionary prisms) became amalgamated via subduction processes into a distinctive type of juvenile lithosphere that was strongly segmented (presumably reflecting assembly of Indonesian-style terranes) and hydrated (reflecting the effects of subduction processes on the stabilizing lithosphere). Based on xenolith work, accreted arcs are interpreted to still be attached to their (compositionally distinct) lithospheric mantle domains, and different types of lithosphere have been persistently different in terms of fertility for melting over billions of years (Karlstrom and Humphreys, 1998). These results point to a unique type of Proterozoic lithosphere in southern Laurentia, perhaps similar to accretionary orogens world wide (Windley, 1992), that is nearly as thick (>150–200 km) and buoyant (due to basalt depletion) as Archean lithosphere, but that has been weaker throughout its history.

In terms of North America crustal growth, our model has roots with Engel (1963), who used North America as an example of continental growth, and DePaolo (1981), who applied Nd studies to estimate mantle derivation ages of juvenile additions in Colorado. Thus, North America may be a primary example of a continent for which much of the continental material formed after the Archean (Reymer and Schubert, 1986; Condie, 1990; Pallister et al., 1990; Taylor and McLennan, 1995). At the global scale, crustal growth curves have been debated since Armstrong (1981), who proposed a model in which most continental material on Earth (low density, differentiated continental crust) formed in the Archean, with subsequent shuffling and shallow recycling of Archean crust and mantle lithosphere since then. Laurentia offers an important field laboratory to test alternative models of continental growth on Earth. The core of the Laurentian craton offers a case study for the extent of juvenile material preserved at deep levels in collisional orogens (Corrigan et al., 2005). Southern Laurentia is a test for the unresolved question of how much older crust and/or lithospheric mantle is hidden beneath the dominantly juvenile accretionary provinces. Bickford and Hill (2007) suggested that much of the southwestern United States may be underlain by Trans-Hudson and/or Penokean crust. In contrast, we recognize the presence of slightly older (1.84 Ga) juvenile arcs that probably had no association with the Trans-Hudson or Penokean orogens (e.g., Elves Chasm arc), as well as the potential for small Archean crustal fragments in the subsurface (Mojavia).

Using Indonesian or Aleutian analogs for the common association of both oceanic and continental arcs in modern orogens, we view the >1000-km-wide Proterozoic orogen in southern Laurentia in terms of several conceptually distinct (but temporally overlapping) processes of crustal growth (Bowring and Karlstrom, 1990).

1. First-stage accretion (differentiation) of mafic, intermediate, and felsic material from the mantle took place via subduction processes during the multistage evolution of outboard arcs. This is the arc-accretion model challenged by Bickford and Hill (2007). We continue to document that timing of magmatism and tectonism (from U-Pb zircon dating) shortly followed (by 10–100 m.y.) mantle separation ages, as inferred from Nd model ages and Pb isotope studies. These data indicate that significant volumes of new continental crust formed as juvenile arcs, arc fragments, and various oceanic terranes outboard of southern Laurentia from 1.8 to 1.0 Ga.

2. Assembly of a mixture of juvenile terranes and older crustal fragments took place progressively in southern Laurentia during a series of orogenic episodes: Yavapai, Mazatzal, Granite-Rhyolite, and Grenville. Analogous to younger convergent orogens, the timing of specific events varies from location to location along the margin, but at the scale of the plate, Laurentia grew southward (present orientation) by arc-continent collisions from 1.9 to 1.0 Ga along a global-scale, long-lived convergent and/or transpressive plate margin (Karlstrom et al., 2001). Arc lithologic associations and style of deformation suggest that tectonism was subduction-dominated during this interval. The multistage assembly process was an important second step in growth of continental crust via crustal thickening, tectonic imbrication of felsic sedimentary successions, and further differentiation and modification of continental crustal materials during suturing.

3. Further stabilization of continental crust took place via crustal melting that was instigated by crustal thickening and resulted in blooms of granitoid magmatism near the end of each orogenic pulse that stitched sutures and helped stabilize the continent. In addition, lithospheric differentiation led to mafic underplating, modification of the Moho, and depletion and stabilization of subcrustal mantle lithosphere (Keller et al., 2005). Following Bowring and Karlstrom (1990), we name this the "arc accretion-assembly-stabilization model"; the result of all three of these processes was the formation of one of the most voluminous accretionary orogens on Earth (Reymer and Schubert, 1986; Windley, 1992).

Apparent challenges to this model (Duebendorfer et al., 2006; Bickford and Hill, 2007) seem mostly semantic. Bickford and Hill (2007, p. 169) argued that "new models involving arc accretion, the involvement of preexisting crust, and crustal extension must be developed"; our view is that we have published these and other elements of uniformitarian models in numerous papers since Karlstrom and Bowring (1988, 1993) and Bowring and Karlstrom (1990). The presence of 1.84 Ga juvenile crust beneath ~1.75 Ga arcs, as documented best in the Elves Chasm area of Grand Canyon (Ilg et al., 1996; Hawkins et al., 1996), does not compromise the arc-accretion model, but rather expands the known time range of formation of arc crust. A key test for the arc accretion-assembly-stabilization model is to determine the extent of Archean crust beneath the southern Laurentian orogens that may have been available to be tectonically recycled (as opposed to detrital inputs) to form new continental lithosphere. Isotopic data suggest that there are not large volumes of Archean material in southern Laurentia, so we continue to view arc differentiation processes as the most efficient way to produce the first-cycle evolved material of the Proterozoic part of the continent.

Extensional deformation has also been proposed as a component of the early evolution of the orogen, based on the concept that basalt-rhyolite (bimodal) associations are restricted to extensional environments and are not present in arcs (Bickford and Hill, 2007). However, discussions of extension (e.g., Wooden and DeWitt, 1991; Duebendorfer et al., 2006) seem largely semantic and need to more adequately define the process or tectonic setting envisioned. Extension is simply part of most three-dimensional strain fields at a variety of scales, and is important, for example, during pluton emplacement (Hutton, 1988). Extension is also one of the possible models to explain the presence of mixed isotopic zones where addition of basalt takes place in small rift basins that developed in older crust (Duebendorfer et al., 2006). However, no early extensional structures have been documented, no major dike swarms are known in this isotopically mixed zone that would support extension, and the observed mixed isotopic values can be adequately explained in terms of tectonic imbrication, sedimentary mixing, and/or plutonic inheritance. In terms of tectonic environments, there are numerous extensional (lithospheric thinning) tectonic environments that are entirely compatible (and expected) within complex Indonesian-style convergent systems and observed complex networks of linked subduction systems–transforms–spreading ridges (Hamilton, 1979). Tectonic regimes that involve localized crustal or lithospheric thinning include intraarc and backarc extension, slab roll back in forearc basins, and transpressive pull-apart basins. In our view, these should not be termed rift models. In a general sense, the magmatic addition of basalt to a thinning lithosphere that takes place in these settings is unlikely to produce appreciable volumes of new continental crust. Dike swarms that are produced by such processes (e.g., Bleeker and Ernst, 2006), while present in parts of the Canadian shield (e.g., Mackenzie, Animikie, Gunbarrel dike swarms of Figs. 17 and 18), have not been documented in southern Laurentia. Appreciable volumes of basalt may have been added as an underplate during the ca. 1.4 Ga tectonic event (Keller et al., 2005; Crowley et al., 2006), but we view this in terms of lithospheric recycling and remelting rather than as appreciable growth by magmatic additions from below. For tectonic interpretations, we focus on the regional contractional stain field (Karlstrom and Humphreys, 1998) and the orogenic connections that may have been going on to the south during the intracratonic A-type magmatism (Karlstrom et al., 2001).

Following Karlstrom and Bowring (1988), the model presented in this paper focuses on orogenic provinces (~1000 km length scale) in terms of isotopic constraints for different age crustal domains. However, we also continue to depict tectonostratigraphic terranes (~100 km scale; e.g., Fig. 8) as the microplates that became welded together to form the larger crustal provinces. Smaller tectonic blocks (~10 km scale) within a segmented lithosphere, as discussed by Karlstrom et al. (2005), are too small to show up in our time slice maps, but we interpret them to be the result of segmentation by shear zones during crustal assembly (e.g., Bergh and Karlstrom, 1992). Continued studies need to better resolve the different scales of crustal additions and interactions between arc accretion and terrane assembly in the development of new continental lithosphere. Better resolution is also needed for the extent of older crust in the subsurface, paleogeography of evolving subduction systems, identification of sutures between different tectonic terranes (e.g., arcs, backarcs, ocean islands, and plateaus), and timing of assembly events. Our model depicts a wide heterogeneously deforming orogenic collage in which outboard collisions caused important reactivation of previously accreted crust.

In terms of the supercontinent cycle, the mosaic of Archean blocks welded into the core of the continent by 2.0–1.8 Ga orogens is similar to that of other cratons, such as western and northern Australia (Myers et al., 1996), Siberia (Sears and Price, 2003), and Baltica (Gorbatschev and Bogdanova, 1993), indicating the likelihood that the birth of North America (Trans-Hudson orogen) took place in the context of assembly of a larger Paleoproterozoic supercontinent. This supercontinent was named Nuna by Hoffman (1997). The 1000-km-long northeast-trending accretionary belts shown in our model are truncated at late Precambrian rift zones, indicating they also extended into adjacent continents within the Mesoproterozoic–Neoproterozoic supercontinent of Rodinia (Li et al., 2007). The integrated geologic history of these belts in southern Laurentian thus can provide a data-rich fingerprint to help test alternate models for the configuration of continental blocks in Proterozoic reconstructions.

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