The Slave Craton From On Top: The Crustal View

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The Archean Slave craton (Fig. 1) is a major building block of the Canadian Shield and one of ca. 35 Archean cratons preserved around the world (Bleeker, 2003). Its amalgamation with the Rae craton, starting at ca. 2 Ga, initiated the climactic 2-1.8 Ga growth of Laurentia (Hoffman, 1989), probably within the broader context of the formation of Earth's first modern supercontinent, Nuna. Much of the Slave craton is old and, within the context of the Laurentian collage, it can be regarded, for all practical purposes, as an exotic fragment relative to other well-known cratons such as the Superior, Nain, and Rae.

As a mere fragment of ancient crust, surrounded by Paleoproterozoic rifted margins, it originated from the break-up of a much larger late Archean landmass, the supercraton Sclavia (Bleeker, 2003). The late Archean and earliest Proterozoic development of Slave crust should thus be viewed in the context of this larger supercraton, even though its shape and size are currently unknown. The critical point is that cratons like the Slave only preserve parts of much larger tectonic systems.

In agreement with this conceptual view, latest Archean events are remarkably homogeneous across the Slave craton and may be used to help identify neighbouring fragments of Sclavia from among the 35 extant cratons. One such event is a voluminous "granite bloom" between 2590-2580 Ma (Davis and Bleeker, 1999). This singular event in the craton's evolution transferred, irreversibly, a significant fraction of heat-producing elements and lower crustal fluids to the upper crust, allowing cooling and stiffening of the lower crust and setting the stage for cratonization and long-term preservation (Bleeker, 2002).

Predating these latest events, the Slave crust preserves a complex and spatially heterogeneous record of crustal growth spanning nearly 1.5 billion years:

Basement complex

Much of the central and western parts of the craton are underlain by ancient and largely crystalline basement—the Central Slave Basement Complex (Bleeker et al., 1999a,b; Ketchum and Bleeker, 2001; Ketchum et al., 2004). Along the Acasta River, this basement complex consists of polymetamorphic gneisses of tonalitic and gabbroic composition that yield protolith ages of ca. 4.03 Ga (Stern and Bleeker, 1998; Bowring and Williams, 1999). Although essentially a chance discovery (Bowring et al., 1989), no other rocks of this age have been found. Apart from a central core with sporadic ages >3.5 Ga (Acasta to Point Lake), the Central Slave Basement Complex is mostly younger with important age modes, from detrital and protolith U-Pb zircon ages, around 3400 Ma, 3150 Ma, 2950 Ma and 2826 Ma (Sircombe et al., 2001).

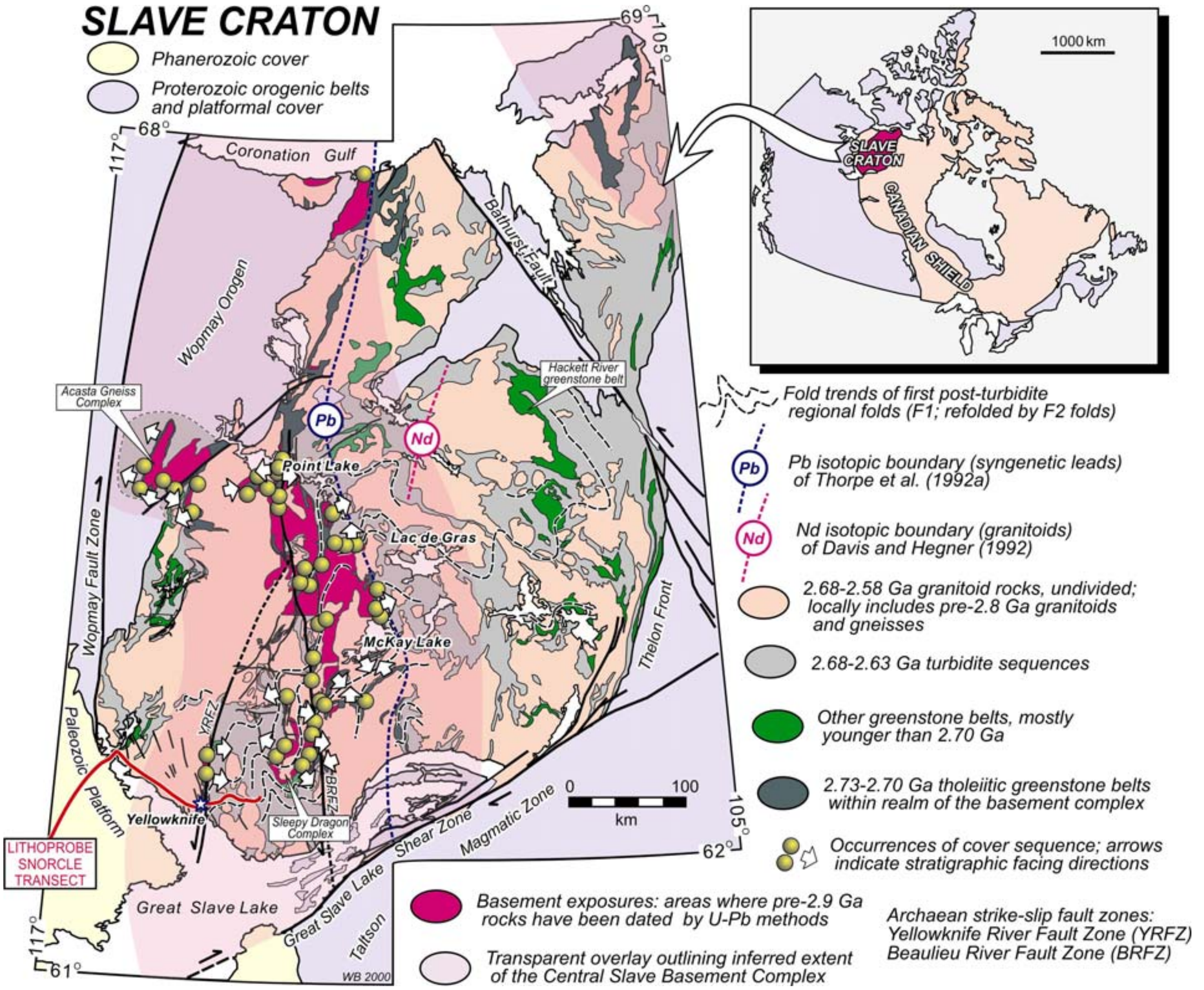

Figure 1. Geological map of the Slave craton. Inset shows location of the Slave craton in the Canadian Shield. Note SNORCLE transect line through Yellowknife.

Interestingly, complementary data from the mantle reveal that at least part of the lithospheric mantle below the central part of the craton is of similar antiquity (Aulbach et al., 2004). Although a crude age zonation can be recognized in the basement complex (Ketchum and Bleeker, 2001), no easily interpretable tectonic pattern has yet emerged. Pre-3.0 Ga supracrustal rocks have been found but form only a small component.

The cover sequence

The contiguous nature of the basement complex, by at least 2.9 Ga, is indicated by a thin but widespread, ca. 2.9-2.8 Ga (Ketchum and Bleeker, 2000) cover sequence of quartzite and banded iron formation (Fig. 2a,b), which marks the onset of the Neoarchean cycle of supracrustal development (Bleeker et al., 1999a). The supermature and commonly fuchsitic quartzites mark the emergence and erosional unroofing of the basement complex in what was probably an aggressive CO_2-rich atmosphere. Abundant detrital chromite may suggest contemporaneous komatiitic volcanism. Similar fuchsitic quartzite sequences occur in many other cratons worldwide, particularly between ca. 3.1 Ga and 2.8 Ga. After 2.4 Ga, mature quartzites are rarely fuchsitic.

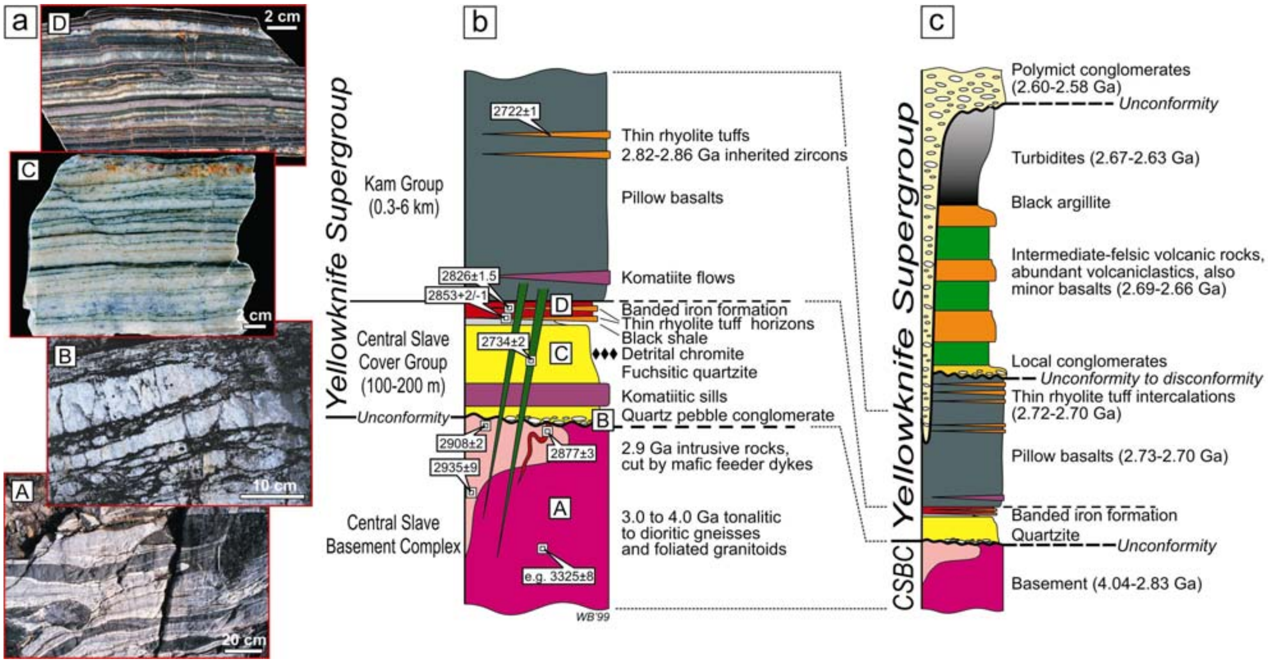

Figure 2. Generalized stratigraphy of the Slave craton. a) Photos (A...D, keyed to Fig. b) illustrating critical features of basement and cover. b) The basement complex and its cover sequence. c) Overall stratigraphy of the craton.

Ca. 2.73-2.70 Ga tholeiitic volcanism

The cover sequence is everywhere overlain by a thick and extensive sequence of tholeiitic basalts, with minor komatiite and rhyolite tuff intercalations (Fig. 2b). Although these lavas extruded in a subaqueous environment, regional correlations suggest a basalt sequence approaching LIP proportions (areal distribution >100,000 km^2, typical thickness 1-6 km). Stratigraphy, dense dyke swarms, and isotopic data link this basalt sequence to the basement (Bleeker, 2002 and references therein; Cousens, 2000). Well-dated components of this basalt-dominated sequence yield ages from 2734 Ma to 2697 Ma (Isachsen and Bowring, 1997; our unpublished data). The widespread basaltic volcanism probably accompanied rifting of the basement complex, possibly assisted by mantle plume activity.

Post-2.70 Ga volcanism

Ca. 2.7 Ga rifting and basaltic volcanism initiated a complex sequence of events including craton-wide, 2.69-2.66 Ga, typically calc-alkalic volcanism and sub-volcanic intrusive activity. These largely juvenile arc-like rocks dominate the eastern part of the craton, while in the central and western part they stratigraphically overlie basement and its basaltic cover (Fig. 2c). Hence, the most likely setting appears an arc- or back arc-like environment that was constructed on highly extended continental crust. The craton-scale stratigraphic relationships and lack of a suture do not easily support models that invoke collision of an exotic, juvenile arc terrane.

Ca. 2.68-2.66 Ga sedimentation

At ca. 2680 Ma, a broad turbidite basin—the Burwash Basin—developed across much of the craton and progressively buried the volcanic substrate (e.g., Ferguson et al., 2004). Late mafic sill complexes and other evidence suggest a volcanically active extensional setting, perhaps best compared with modern back-arcs. The minimum size of this basin was ca. 400x800 km, comparable to that of the Japan Sea, making it the largest and possibly best preserved Archean turbidite basin in the world. Like the Japan Sea, it was largely ensialic, in agreement with suggestions by many early workers (e.g., Henderson, 1985). The main axis of the basin and subsequent structural trends (Fig. 1a) are northeast-southwest, distinctly across the north-south isotopic boundaries that track the nature of deep basement. With more and better U-Pb zircon ages, a tentative volcanic line of 2661 Ma felsic volcanic complexes is recognized, parallel to the Burwash Basin.

Ca. 2.65-2.63 Ga closure of the Burwash Basin

Subsequent tectonic events record the closure and folding of the Burwash Basin (D1), prior to 2634 Ma, the age of a distinct and probably subduction-related magmatic suite (Defeat) across the southern Slave craton. Closure of the highly extended, but largely ensialic basin allowed considerable shortening and mobility but with a structural style dominated by fairly systematic, mostly upright, northeast-southwest trending fold trains. The folded Burwash strata do not represent an outboard accretionary prism and there is no “Contwoyto terrane” (cf. Kusky, 1989). The northeast-southwest D1 structural grain is also recognized in the lithospheric mantle. Shallow subduction (either from the SE or NW?) may have emplaced distinct mantle slabs (Davis et al., 2003).

Post-2.63 Ga turbidites

Along the northwestern margin of the craton, younger turbidites containing ca. 2630 Ma detrital zircons (e.g., Pehrsson and Villeneuve, 1999) record a migration of tectonic activity to the northwest. Deposition was coeval with uplift and erosional unroofing of Defeat plutons and tightly folded Burwash Formation strata. Shortly following their deposition, these younger turbidites were shortened and intruded by ca. 2615-2610 Ma tonalite-granodiorite plutons.

2.60-2.59 Ma: final orogenesis

Starting at ca. 2600 Ma, the entire craton was affected by cross-folding and significant further shortening (D2), characterized by broadly north-south structural trends, and probably in response to final collision along a distant active margin of Sclavia. Moderate overthickening of the crust led to HT-LP metamorphism, widespread anatexis, the appearance of S-type granites, and a hot and weak lower crust, culminating in ca. 2590 Ma extension and the regional “granite bloom”. The intrusion of carbonatites (Villeneuve and Relf, 1998) and involvement of other mantle-derived melts indicate a role for mantle processes (delamination?). While peak temperatures were attained in the lower crust, large basement-cored domes were amplified by buoyancy driven deformation; lower crustal devolatilization reactions mobilized gold-bearing fluids; and syn-orogenic clastic basins formed and were immediately infolded into tight synclines (Bleeker, 2002). At least one of these syn-orogenic clastic basins formed as late as ca. 2580 Ma (Sircombe and Bleeker, in prep.). Late strike-slip faulting overprinted and truncated the synclinally infolded clastic basins. The lower crust cooled (Bethune et al., 1999), finally coupled with the mantle, and the Slave (within Sclavia) became a craton.

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