THE SLAVE CRATON: GEOLOGICAL AND METALLOGENIC EVOLUTION

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Introduction

The Archean Slave craton (Figs. 1, 2; Padgham and Fyson, 1992; Bleeker and Davis, 1999a) is a major building block of the Canadian Shield. It is 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 growth of Laurentia (Hoffman, 1988, 1989) from 2.0 to 1.8 Ga, 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 of crust relative to other well-known cratons in Laurentia such as the Superior, Nain, and Rae (Bleeker, 2003, 2004).

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-perhaps a speculative late Archean supercontinent Kenorland (Williams et al., 1991) or, perhaps more likely, a smaller landmass referred to as 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 (Sclavia), even though the shape and size of this supercraton are currently unknown. The salient point is that cratons like the Slave only preserve parts of the much larger tectonic systems in which they were generated.

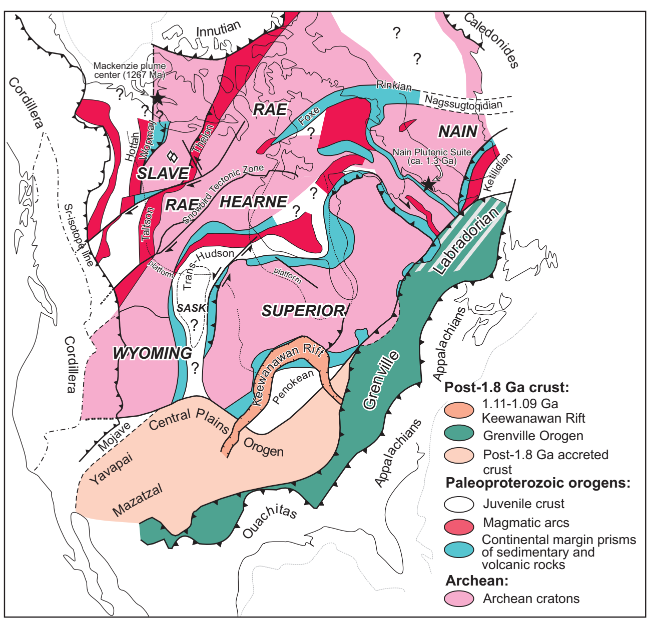

FIG. 1: Tectonic map of the Precambrian basement of North America, showing the location of the Archean Slave craton relative to other first-order crustal elements. Greenland is shown in a pre-drift position. (Modified after Hoffman, 1988; and Ross et al., 1991)

In agreement with this conceptual view, latest Archean events are remarkably homogeneous across the Slave craton and may be used, together with pre-breakup Proterozoic mafic dyke swarms, to help identify neighbouring fragments of Sclavia from among the 35 extant cratons. One such Slave craton-wide event is a voluminous "granite bloom" between ca. 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, thus 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 (Bleeker and Davis, 1999a, b and references therein; Sircombe et al., 2001; Ketchum et al., 2004). The present paper briefly summarizes this crustal growth history and the overall geological evolution of the Slave craton, while highlighting significant metallogenic events preserved within the craton. Significant ore deposits and occurrences are listed in Appendix 1 and will be discussed in terms of their overall setting within the geology of the craton.

Ancient 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 (Figs. 2, 3, and 4; see 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 to gabbroic composition (e.g., Fig. 5a) that yield protolith ages up to ca. 4.03 Ga (Bowring et al., 1989; Stern and Bleeker, 1998; Bowring and Williams, 1999). Although essentially a chance discovery (M. St.-Onge, pers. comm., 2000; Bowring et al., 1989), no other rocks of this age have yet 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 (Fig. 4b; e.g., Sircombe et al., 2001; see Bleeker and Davis, 1999b for a compilation of basement ages).

Interestingly, complementary data from the mantle suggest that at least part of the lithospheric mantle below the central part of the craton may be of similar antiquity

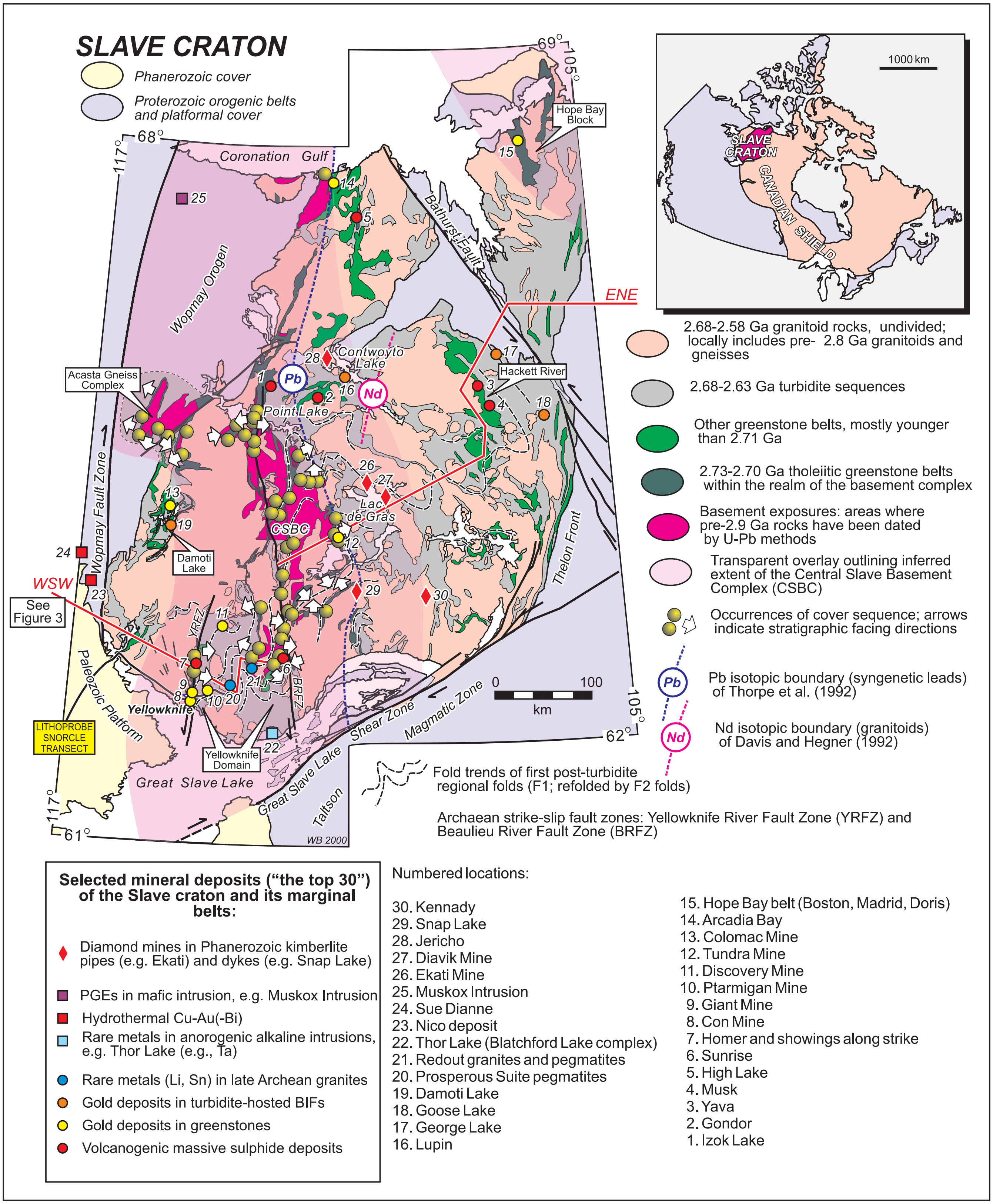

FIG. 2: Simplified geological map of the Slave craton. Localities mentioned in the text are highlighted, as are selected mineral deposits or significant occurrences. Cross-section line (ENE-WSW) refers to the craton-wide structural section shown in Figure 3.

(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-2.9 Ga supracrustal rocks have been found at the base of some greenstone belts (e.g., Ketchum et al., 2004) but form only a small component.

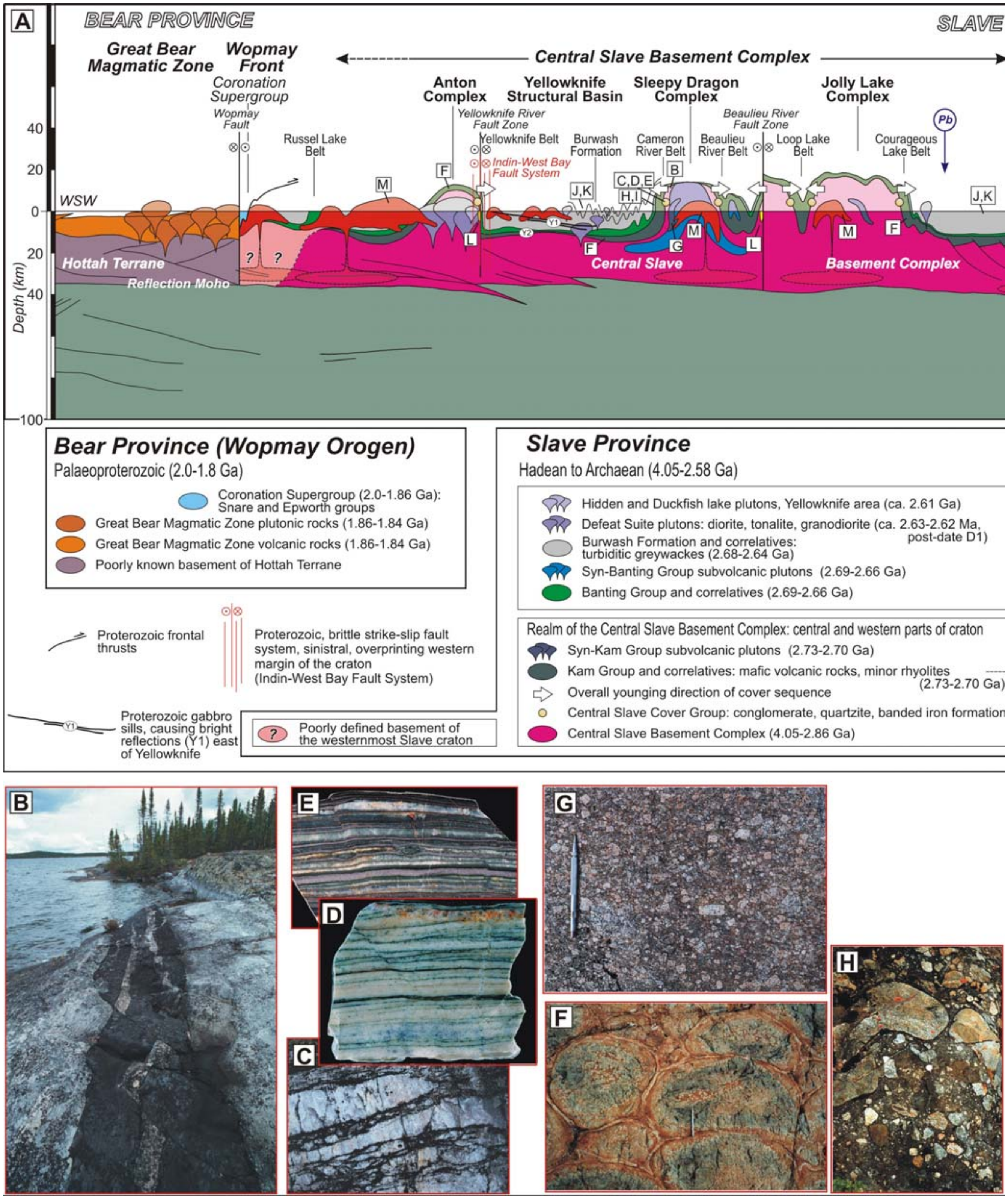

FIG. 3: A. WSW-ENE structural-stratigraphic section across the Slave craton (see Fig. 2 for location of profile). Lettres B, C…m refer to illustrations below section. Present erosion level at 0 km depth; some units are shown above this level in lighter tones for ease of interpretation; no vertical exaggeration. Deep structure in the western part of the profile interpreted from LITHOPROBE's SNORCLE seismic reflection profile (e.g., Cook et al., 1999). Volcanic units <2690 Ma and overlying turbidites overlap the boundary between different basement domains. Field photos: B. Typical 2.95 Ga foliated tonalites of the Central Slave Basement Complex with transposed 2734 Ma mafic dykes. C., D., E. Basal quartz pebble conglomerate, fuchsitic quartzite, and banded iron formation of the Central Slave Cover Group that overlies the basement complex. F. Variolitic pillow basalts of the Kam Group, Yellowknife. G. Syn-Kam Group K-feldspar porphyritic granodiorite pluton in basement below greenstone belts. H. Polymict conglomerate, including 10-30 cm granitoid cobbles, which occurs locally at the base of the younger, 2.69-2.66 Ga, volcanic cycle. I. Carbonate-cemented rhyolite breccia typical for the younger volcanic cycle.

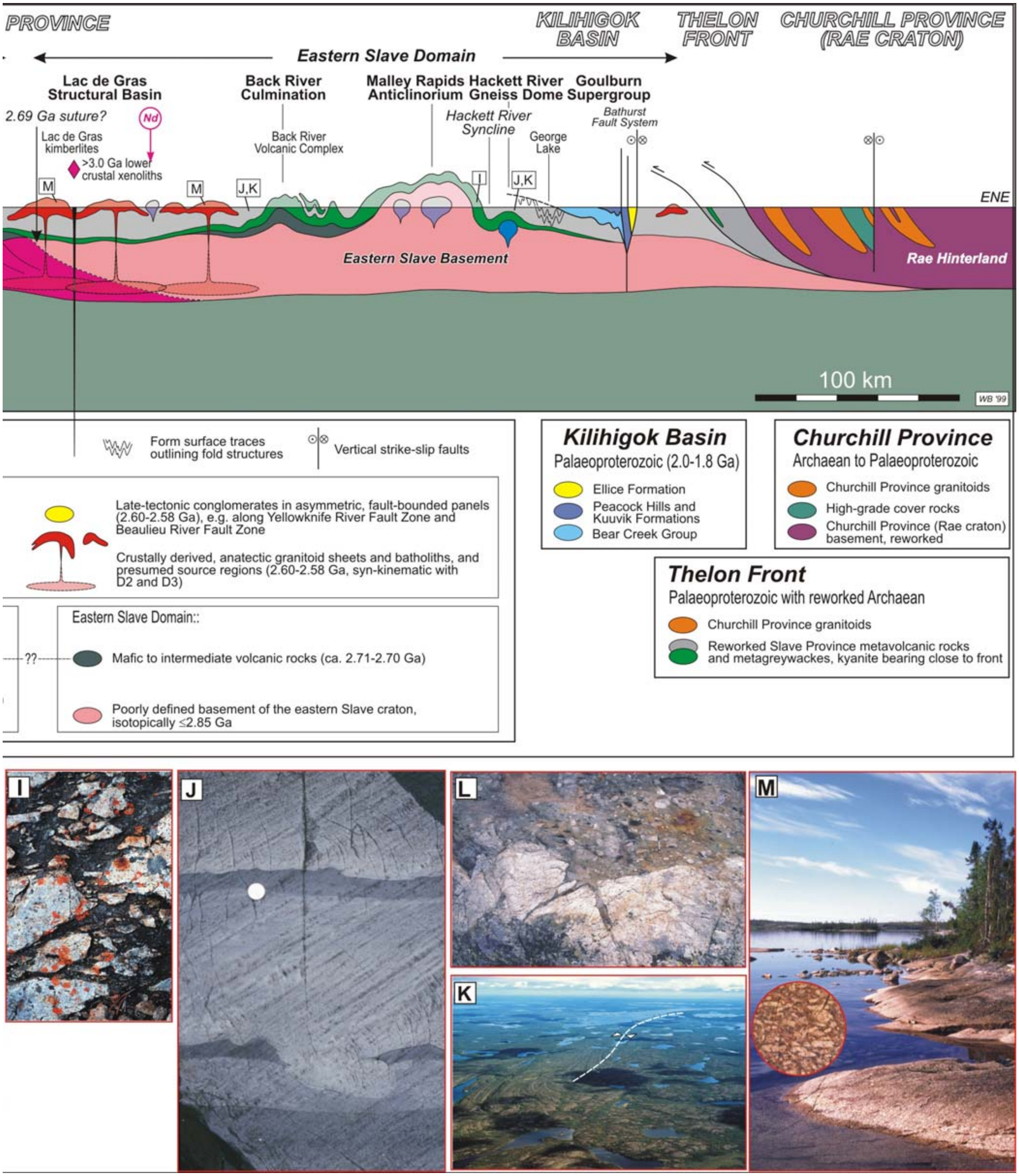

Fig.3 (continued): J. Well-preserved sub-biotite grade turbidites in the core of the Yellowknife structural basin, showing graded bedding and load casts. K. Areal photo of large scale, upright, fold structures in turbidites of the Yellowknife structural basin. L. Late-tectonic conglomerates, <2600 Ma, unconformably overlying unroofed granitoid rocks, Point Lake. M. Late-tectonic, 2585 Ma, K-feldspar megacrystic granite of the Morose Suite in the core of the domal Sleepy Dragon Complex; inset shows 1-2 cm-large K-feldspar megacrysts.

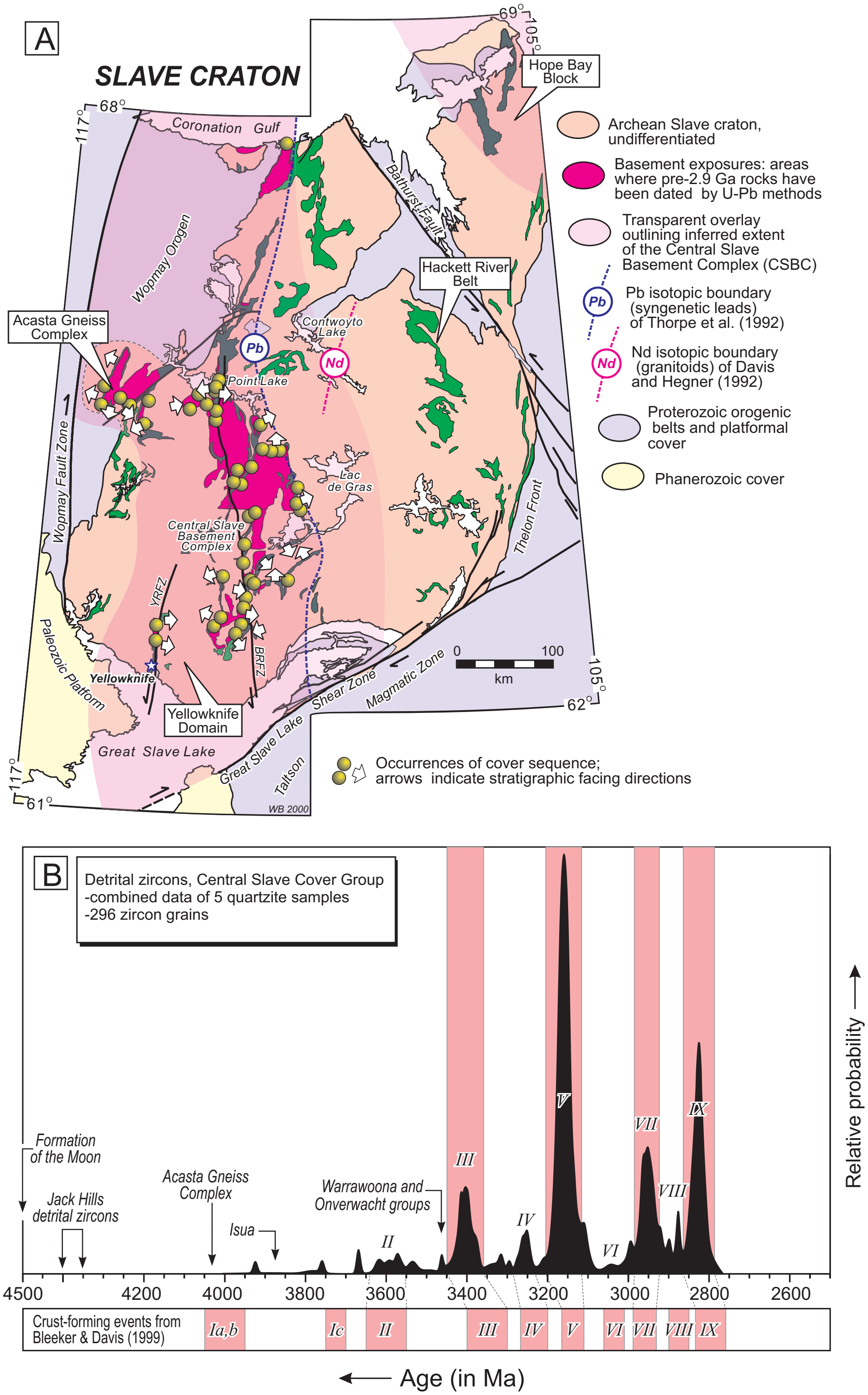

FIG. 4: A. Simplified map showing mininum extent of Mesoarchean to Hadean basement of the Central Slave Basement Complex (CSBC). Spheres (yellow or orange) highlight locations where the diagnostic basement to cover stratigraphy has been observed. Note the "Hope Bay block" in the northeastern Slave craton, which has a number of characteristics that are more similar to the southwestern Slave and its greenstone belts overlying the CSBC (e.g., Yellowknife) than to typical localities of the eastern Slave (e.g., Hackett River). Currently, there are insufficient data to answer the question whether the CSBC reappears in the Hope Bay block. B. Age distribution of the Central Slave Basement Complex, as sample by 296 concordant detrital zircon grains from five quartzite samples across the basement complex (orange spheres in Fig. A: Yellowknife, Cameron River belt, northern Beaulieu belt, Point Lake, and quartzite overlying Acasta basement; see Sircombe et al., 2001). Although biased by detrital sampling and preservation of basement zircons, this age spectrum allows a rapid assessment of major age components of the basement complex. Note significant age peaks starting at ca. 3400 Ma. The depositional ages of these quartzites is ca. 2850-2800 Ma. Roman numerals refer to major crust-forming events in the basement recognized from U-Pb protolith ages (see Bleeker and Davis, 1999b).

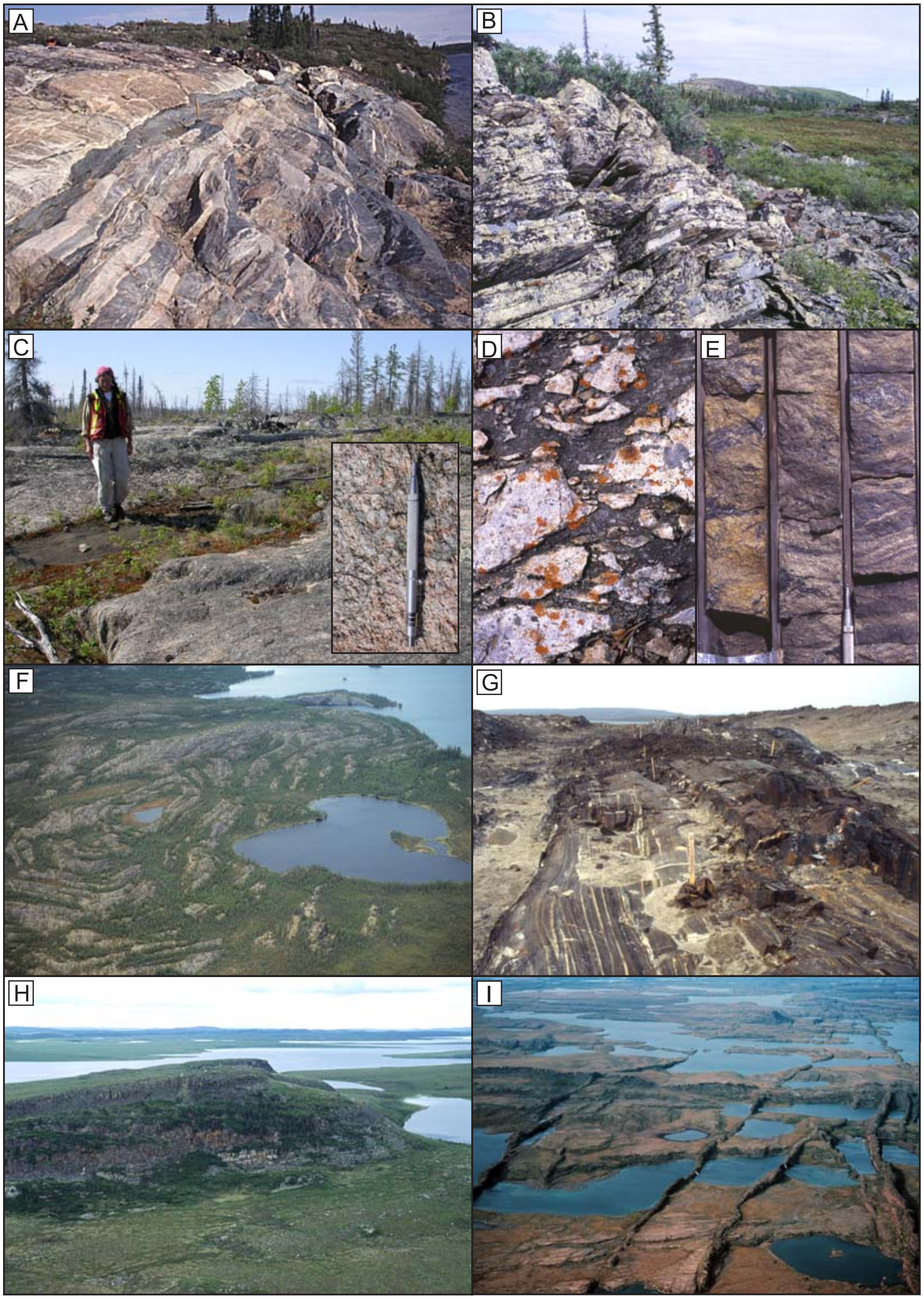

FIG. 5: Key elements of the geology of the Slave craton illustrated by field photographs. A. Cleaned exposures of the Acasta gneisses at their discovery site. Ancient tonalites (4.03 Ga) occur on left side of the picture, and are intruded by highly deformed younger granite sheets and mafic dykes. B. Basal quartzites of the Central Slave Cover Group overlying basement of the Central Slave Basement Complex. Low foreground to the right are low-weathering basement gneisses; dark ridge in background are ca. 2.7 Ga basalts overlying the quartzites. C. Syn-Kam Group quartz-porphyritic tonalite intrusion (ca. 2713 Ma), silling into the northern part of the Yellowknife greenstone belt (geologist Val Jackson for scale). The large sill-like body is cut by somewhat younger mafic dykes that likely fed the upper part of the greenstone belt. Inset shows close-up of altered quartz-porphyritic tonalite. D. Quartz porphyritic rhyolite breccia with carbonate matrix, typical for the uppermost part of 2690-2660 Ma felsic volcanic edifices. E. Massive sulphide mineralization of the Sunrise deposit, associated with ca. 2670 Ma felsic volcanic rocks just below the interface with the Burwash Formation turbidites. F. Thickly bedded sandy turbidites typical of the Burwash Formation in its type area east of Yellowknife. The oblique areal photo shows an F1 syncline refolded by north-northwest trending F2 folds. G. Silicate facies iron formation interlayered with turbiditic greywackes, George Lake, northeastern Slave. This banded iron formation hosts significant epigenetic gold mineralization. H. Passive margin strata of the Coronation Supergroup (Epworth Group) overlying the western margin of the rifted Slave craton, structurally at the base of Wopmay orogen. I. Dense Proterozoic mafic dyke swarms cutting extended Slave crust and its cover.

Mineralization

There are few if any known mineral occurrences of note within the Mesoarchean (or older) crystalline basement complex, a statistic that is generally mirrored by other ancient gneiss complexes in cratons around the world. To a first degree, this poor endowment probably correlates with the virtual lack of supracrustal rocks within such complexes.

Indirectly, however, the presence of the ancient basement complex may have exerted controls on several classes of younger mineral deposits:

- Seafloor hydrothermal deposits or occurrences within Neoarchean bimodal rift volcanic rocks that overlie faulted basement, for instance the basal greenstones of the Yellowknife and Courageous Lake belts.
- Late Archean evolved granites and their associated pegmatite swarms, some of which are enriched in rare elements (Sn, Ta, Li). Such enrichment typically correlates with multiple cycles of crustal fractionation.
- Diamondiferous kimberlites in the Lac de Gras region (Fig. 6), which appear to have ascended through the edge of the basement complex and its ancient lithosphere (Fig. 2b).

The association of diamondiferous kimberlites with "low-geotherm" Archean cratons and their mantle keels is well known ("Clifford's rule"). Whether Mesoarchean or older crustal rocks are particularly favourable within the context of Archean cratons is not clearly established, but appears a question worth testing against a global database. The distribution of economic diamond deposits in the Slave craton (Fig. 2) helps to bring this question into focus.

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 cover sequence of quartzite and banded iron formation (Fig. 7; see also Fig. 5b), the Central Slave Cover Group (Bleeker et al., 1999a). This sequence, which is locally intruded by ultramafic sills (Fig. 7), marks the onset of the Neoarchean cycle of supracrustal development (Bleeker et al., 1999a).

The supermature and commonly fuchsitic quartzites that are characteristic of this sequence mark the emergence and erosional unroofing of the basement complex in what was probably an aggressive, CO_2-rich, Archean 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, indicating a lesser role for detrital chromite (and komatiites) in the post-Archean world.

Mineralization

The Central Slave Cover Group hosts some of the more prominent banded iron formations (BIF) of the Slave craton, although most are thin (1-10 m) and variable in composition along strike, changing from oxide iron formation into silicate-rich varieties or merely ferruginous chert. Locally, however, folding has thickened the highly magnetic BIF into substantial thicknesses (e.g., at Amacher Lake, on the eastern flank of the Sleepy Dragon Complex), resulting in some of the highest amplitude total field magnetic anomalies in the Slave craton. Overall, the BIFs appear of low economic value, although some may possibly host epigenetic gold mineralization and may be under explored for this commodity. However, most iron formation-hosted gold mineralization appears to be associated with BIFs hosted in low to medium-grade turbidite packages.

FIG. 6: North America's first diamond producer, the Ekati Mine of the central Slave craton. Main picture shows the flat barren lands of the central Slave craton, with several pipe-like kimberlite bodies being excavated in circular open pits. Inset (upper left) shows a close-up of one of the partially excavated pipes. Several mm-size gem quality diamond octahedral are shown on upper right (photos provided by David Snyder and Grant Lockhart).

Fuchsitic quartzites below the iron formations (Fig. 7, photo C) are enriched in detrital minerals, including highly stable heavy minerals such as chromite, zircon, and rutile. Individual, dark, detrital chromite grains are a characteristic feature of these otherwise white to grey quartzites (Bleeker et al., 1999a). Commonly, these chromite grains have undergone variable reaction towards bright green fuchsitic mica during metamorphism and deformation. In a few localities, chromites are concentrated in seams of "black sand", but clearly such concentrations are too small to be of economic interest. If road access were available, some of the green-white quartzite would make attractive building or decorative stone. In Greenland, India, and Australia, similar quartzites are often quarried for this purpose. Elsewhere in the world, quartzites similar to these contain paleoplacer deposits of gold and/or uranium. An initial survey of such potential in the Slave craton was carried out by Roscoe (1990).

Ultramafic sills (or flows?) intruded the cover sequence in several places, and locally contain seems of magmatic chromite (Covello et al., 1988). Economic concentrations have not been found. In one remote locality, on the south shore of Desteffany Lake, the author found sulphide concentrations adjacent to ultramafic rocks within the cover sequence. Overall, the volume of komatiitic rocks is limited, not only at this stratigraphic interval but throughout the Slave craton.

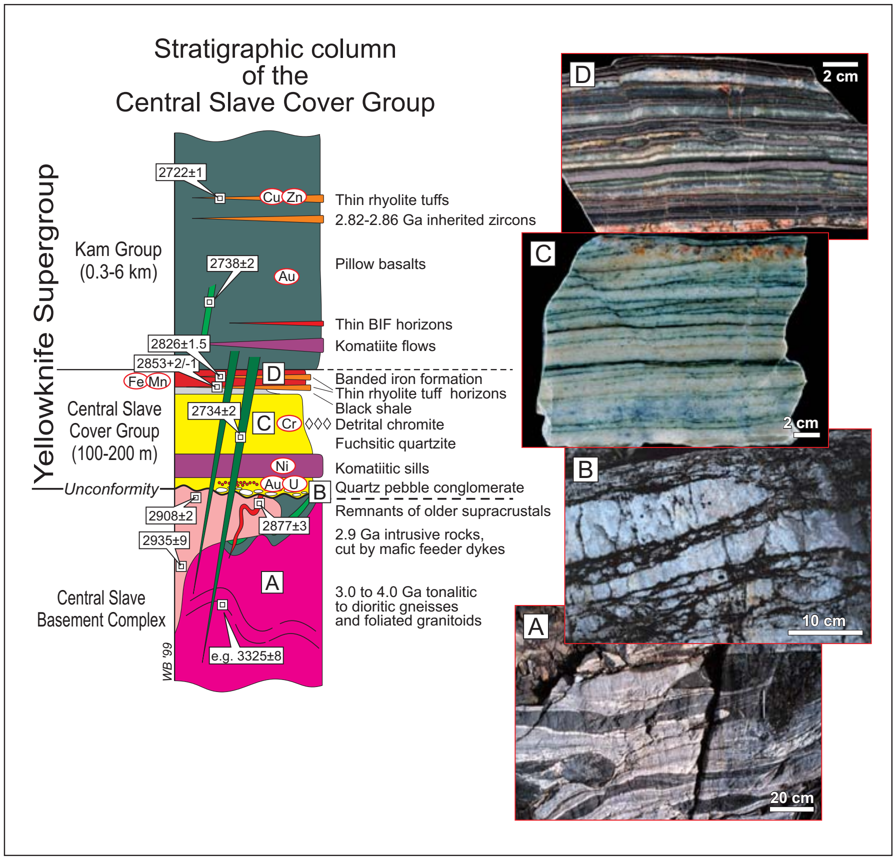

FIG. 7: Generalized stratigraphic column of the autochthonous cover of the Central Slave Basement Complex-the Central Slave Cover Group (Bleeker et al., 1999a). Photos A-D illustrate characteristic lithologies. Some of the age data were reported in Isachsen and Bowring (1997), Bleeker et al. (1999a, b), and in Ketchum and Bleeker (2000).

Ca. 2.73-2.70 Ga Tholeiitic Volcanism

Wherever the thin cover sequence is recognized, it is overlain by a thick and extensive sequence of tholeiitic basalts, with minor komatiite and rhyolite tuff intercalations (Figs. 7, 8). In the Yellowknife greenstone belt, this basalt-dominated volcanic sequence (Fig. 8) is known as the Kam Group (Helmstaedt and Padgham, 1986; Bleeker et al., 1999a). Possible correlative basalt successions (Fig. 9) are known across the basement domain, as far east as the Courageous Lake belt, and at least as far north as around the Exmouth antiform in the Acasta area. This basalt sequence typically consists of several hundred meters to several kilometres of pillowed and massive flows, with thin felsic horizons, and intruded by numerous dykes and sills of several generations (Fig. 8).

Well-dated components of this basalt-dominated sequence yield ages from >2738 Ma to 2697 Ma (Isachsen and Bowring, 1997; Davis et al., 2004; and unpublished data). In Yellowknife, the top of the sequence is represented by voluminous basaltic flows and intercalated felsic volcanic rocks of the Yellowknife Bay Formation, dated at ca. 2700 Ma (Fig. 8). In support of the overall regional correlation, similar ca. 2700 Ma ages have been obtained from Courageous Lake and Acasta areas. Stratigraphy, dense dyke swarms, and isotopic data link the basalt sequence to the basement (Henderson, 1985; Bleeker et al., 1999a, b; Bleeker, 2002, and references therein; Northrup et al., 1999; Cousens, 2000).

If the broad regional correlation of these basalts is valid, the magnitude of volcanism approaches LIP (large igneous province) proportions (areal distribution >100,000 km^2, typical thickness 1-6 km). The widespread basaltic volcanism probably accompanied protracted rifting of the basement complex, possibly assisted by mantle plume activity. The stratigraphy in Yellowknife is compatible with such a rifting interpretation. At the top of the Kam Group, bimodal volcanic rocks of the Yellowknife Bay Formation become progressively more intercalated with volcaniclastic sediments, before final intrusion by thick tholeiitic sills. One of these sills, the Kam Point gabbro sill, has a preliminary baddeleyite age of ca. 2697 Ma (Fig. 8).

Mineralization

A volcanically active rift environment, characterized by bimodal volcanism and minor aprons of volcaniclastic sedimentary rocks, is a highly favourable environment for seafloor hydrothermal activity and the formation of volcanogenic massive sulphide deposits. Indeed numerous showings of sulphidic horizons occur throughout the basalt-dominated greenstone belts of the west-central Slave.

Of particular interest are intercalated felsic volcanic flows and/or sills, which are direct indicators of proximity to a differentiated magmatic center, and thus a long-lived sub-volcanic heat source. The Bell Lake quartz-porphyritic tonalite sill (Fig. 5c) and the rhyolitic Townsite Formation, dated at 2713Π2 Ma and ca. 2709 Ma, respectively (Davis et al., 2004), are examples of such proximal felsic volcanic rocks in the Yellowknife greenstone belt. Hydrothermal alteration and minor sulphide mineralization is known from the Yellowknife Belt (e.g., the Homer Lake showing, and horizons northeast of Bell Lake), but to date no significant deposits have been found. Similarly, despite at least a first wave of exploration across the other basaltic greenstone belts of the west-central Slave craton, the author is not aware of any major discoveries.

From a mantle perspective, it seems inconceivable that events associated with the voluminous basaltic volcanism recorded across the ancient basement terrain did not involve thinning or at least modification of the lithospheric mantle below the Central Slave Basement Complex. Large-scale melting was probably triggered by adiabatic rise of asthenospheric mantle. Perhaps, then, the ca. 2.7 Ga basaltic volcanism may have contributed to the highly depleted mantle compositions underlying the core of the craton.

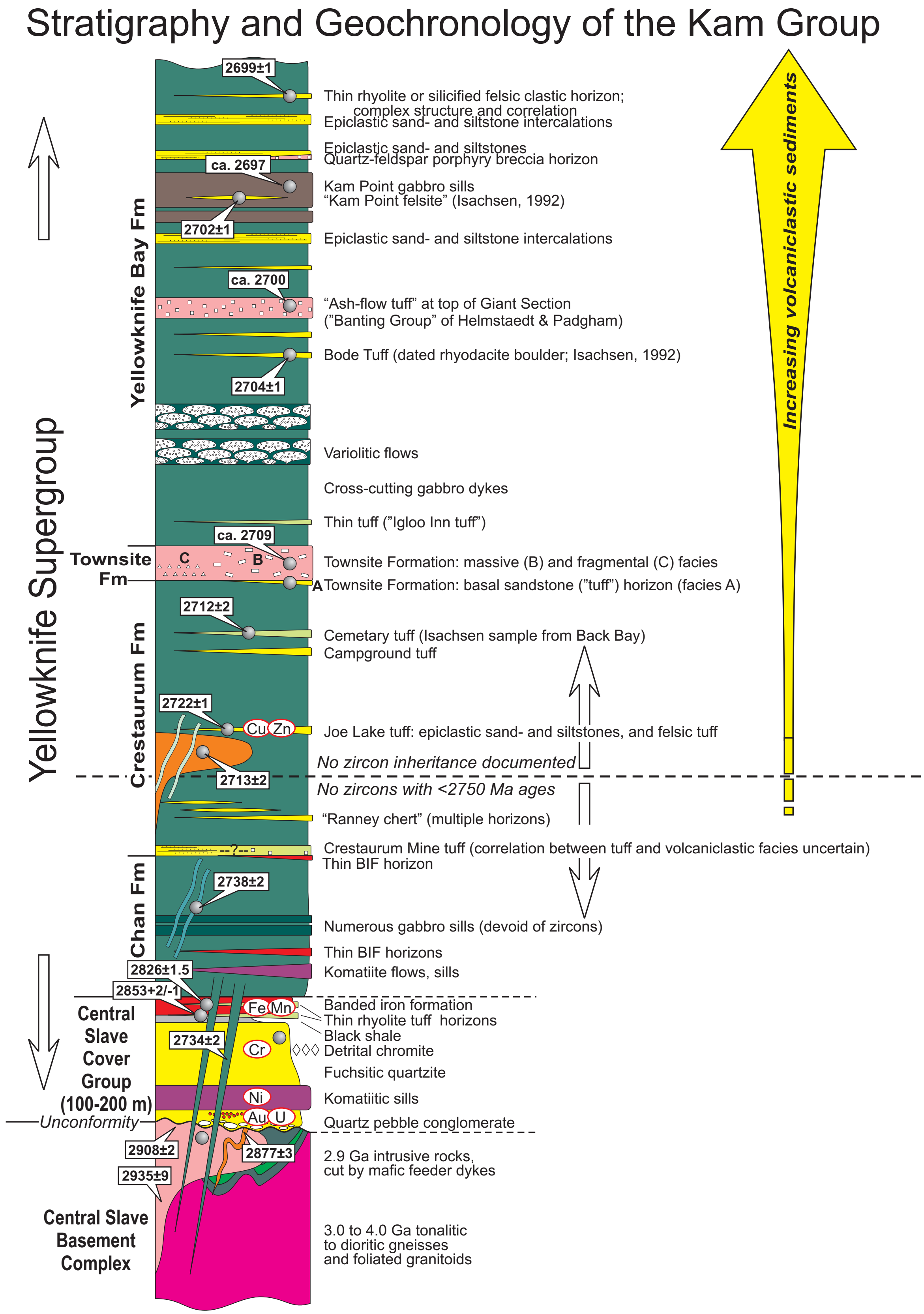

FIG. 8: Stratigraphy and geochronology of the Kam Group, Yellowknife greenstone belt. All major units have now been dated, showing a uniform monotonic younging upwards through the pile. Sources of age data: Isachsen and Bowring, 1997; Bleeker et al., 1999a; Ketchum and Bleeker, unpublished data; Davis et al., 2004. A cross-cutting gabbro dyke in the Chan Formation, dated recently at 2738 Ma (J. Ketchum, pers. comm., 2004), shows that much of the Chan Formation is >2738 Ma. The youngest event recognized to date are the large intrusive gabbro sills at Kam Point, with a preliminary baddeleyite age at 2697 Ma. A large quartz-porphyritic tonalite sill (see Fig. 5c) has been dated at 2713 Ma and acted as a heat source for seafloor hydrothermal alteration.

Post-2.70 Ga Volcanism

Following ca. 2.7 Ga basaltic volcanism and rifting, most areas in the Slave craton show a transition to calc-alkaline volcanism characterized by abundant felsic and intermediate volcanic rocks, calc-alkaline basaltic rocks, and intercalated volcaniclastic sedimentary rocks (Fig. 9). In nearly all areas, these arc-like rocks are stratigraphically overlain by turbiditic sedimentary rocks (Fig. 9). Ages for the arc-like volcanic rocks are typically in the range of 2690-2660 Ma.

The arc-like volcanic rocks are geochemically juvenile (e.g, Davis and Hegner, 1992). They dominate the eastern part of the craton, where they lack any apparent association with older basement, its cover, and/or the basalt-dominated rift sequence. These observations have led to models in which the eastern Slave represents an exotic juvenile arc (the "Hackett River arc") that collided with the basement domain in the west (e.g., Kusky, 1989, 1990). However, similar arc-like rocks, with identical ages, stratigraphically overlie the basement domain and its cover in the west-central parts of the craton (Fig. 9a), where they can be tied to the basement and bimodal rift volcanic rocks by means of unconformities, cross-cutting feeder dykes, and subvolcanic intrusions (Fig. 10).

It thus appears that, if these rocks were generated in an arc-like setting, this arc was constructed marginal to and on top of the highly extended continental crust of the Central Slave Basement Complex. This suggests a marginal to continental arc setting. The arc was actively extending and evolved into a back-arc basin that was ultimately filled with turbiditic sediments (Figs. 9, 10). The geochemistry of the volcanic rocks and the associated subvolcanic plutons, although juvenile, typically shows strong arc-like signatures (light rare earth and large-ion lithophile element enrichment, Nb depletion) compatible with enriched sources in a supra-subduction zone setting.

Mineralization

The ca. 2687-2660 Ma time interval and the arc-like volcanic sequences are highly favourable for volcanogenic massive sulphide (VMS) mineralization. Nearly all known massive sulphide deposits, including Izok Lake, the Hackett River deposits, and the Sunrise deposit (Fig. 5e; see Fig. 2 for locations) of the southern Slave craton belong to this group. An exception is the High Lake deposit, which is associated with older, ca. 2705 Ma, bimodal volcanic rocks.

Nearly all the VMS deposits of this group occur associated with proximal felsic volcanic rocks at or near the transition to overlying turbiditic metasedimentary rocks. This transition is characterized by rhyodacite to rhyolite complexes, volcaniclastic sediment aprons, thin sulphidic chert horizons and, in some localities, banded iron formations. Carbonate rocks (calc-arenites) are associated with some of the felsic complexes (Fig. 5d). This typical stratigraphic evolution, from shallow water or emergent felsic volcanic complexes to deep-water turbidite sedimentation, suggests active extension and tectonic subsidence of the arc environment, most likely in an overall back-arc setting (Fig. 10). Such an environment of active faulting, active volcanism, thinning lithosphere, and high heat flow, has long been recognized as a classic environment for VMS deposits. Izok Lake and the Hackett River deposits represent some of Canada's largest undeveloped volcanogenic massive sulphide deposits that will be economic as soon as road and coast access is available.

Ca. 2.68-2.66 Ga Sedimentation

Starting 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., 2005). A persistence of volcanic intercalations up-section and late mafic sill complexes suggest a volcanically active extensional setting, perhaps best compared with modern back-arcs. The minimum size of this basin was ca. 400x800 km (Fig. 11a), comparable to that of the Japan Sea, and making it the largest and possibly best-preserved Archean turbidite basin in the world. Like the Japan Sea, the Burwash Basin was largely ensialic, in agreement with inferences by early workers (e.g., Henderson, 1985).

The Burwash Basin fill consists largely of immature greywackes and mudstones, deposited below wave base, and may locally be up to 10 km thick. Intercalated tuff layers have been dated at ca. 2661 Ma (e.g., Bleeker and Villeneuve, 1995). Across the Slave craton, the greywacke turbidites have been given different formational names: the classical Burwash Formation in the Yellowknife Domain; the Contwoyto Formation in central and northern Slave, identical in essentially all aspects to the Burwash Formation further south, except for the presence of intercalated iron formations; the Itchen Formation, a more mud-rich facies in the north-central Slave; and the Beechey Lake Group in the northeastern Slave.

Many of the turbidite beds, particularly those of the Burwash and Contwoyto Formation, are sand dominated with only thin silt to mud sections at the top of the graded beds. In the Yellowknife Domain, thick amalgamated sand beds (2-10 m) are not unusual (Fig. 5f). Petrography, detrital zircons, and geochemical analysis indicate that the greywacke detritus consists of a mixture of mafic and felsic volcanic rocks and uplifted plutonic infrastructure, with only minor input from ancient basement rocks. The main axis of the basin and subsequent structural trends appear to have been northeast-southwest (Fig. 11), distinctly across the north-south isotopic boundaries that track the nature of deep basement (see also Padgham, 1992). This interpretation is based on the following observations:

- Identical Burwash Formation turbidites extend from near Yellowknife (the type area) to the northeastern Slave (Figs. 9, 10, 11a).
- Banded iron formations in the turbidites are restricted to the northwest half of the craton, suggesting a northeast-southwest facies boundary or tectonic trend across the basin (Fig. 11a).
- Earliest folds in the turbidites, which formed between 2650-2630 Ma, have northeasterly trends after qualitative "unfolding" of younger fold generations. Early folds appear to form a systematic northeast-southwest trending fold belt (Fig. 11b).

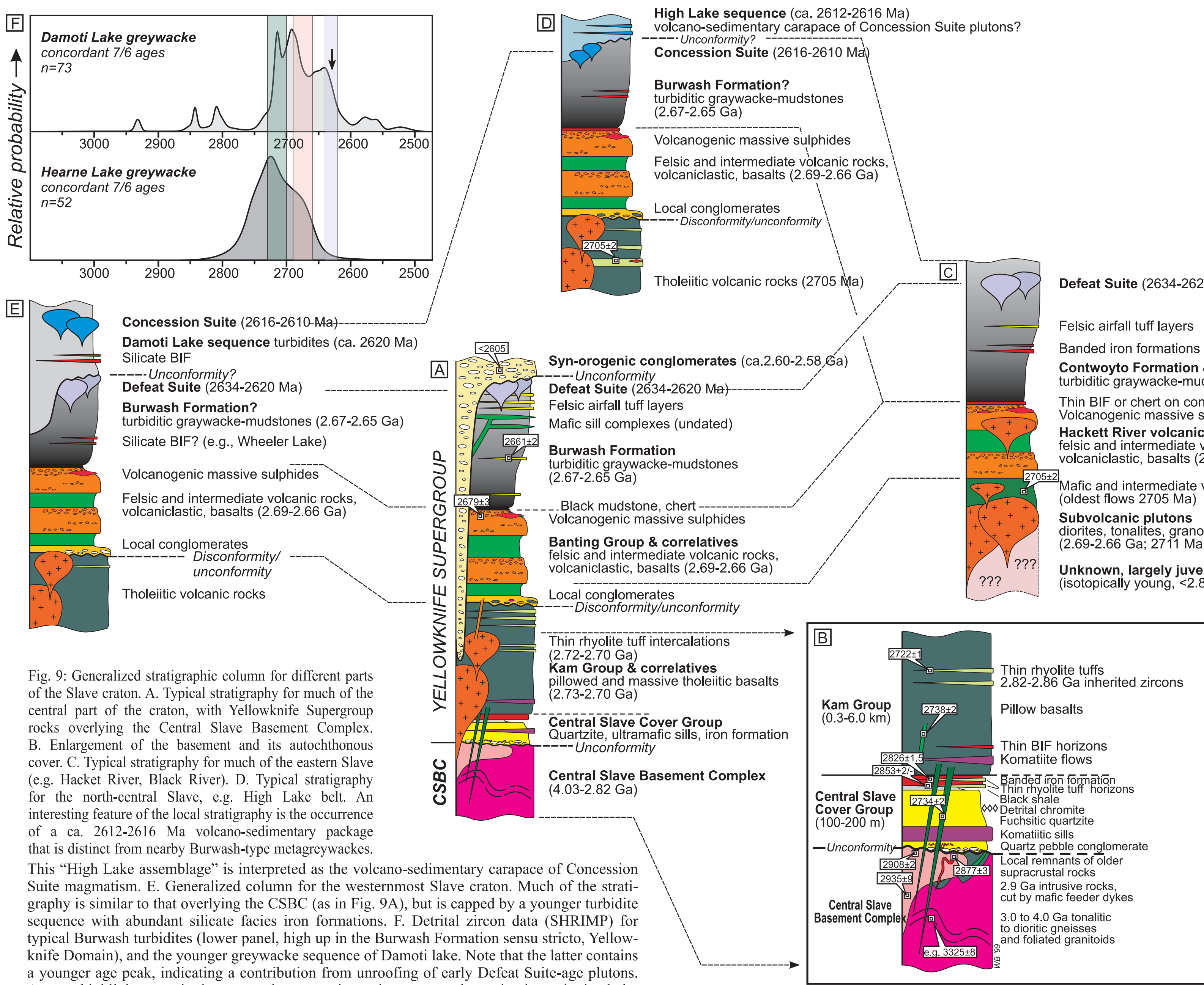

Fig. 9: Generalized stratigraphic column for different parts of the Slave craton. A. Typical stratigraphy for much of the central part of the craton, with Yellowknife Supergroup rocks overlying the Central Slave Basement Complex. B. Enlargement of the basement and its autochthonous cover. C. Typical stratigraphy for much of the eastern Slave (e.g. Hacket River, Black River). D. Typical stratigraphy for the north-central Slave, e.g. High Lake belt. An interesting feature of the local stratigraphy is the occurrence of a ca. 2612-2616 Ma volcano-sedimentary package that is distinct from nearby Burwash-type metagreywackes. This "High Lake assemblage" is interpreted as the volcano-sedimentary carapace of Concession Suite magmatism. E. Generalized column for the westernmost Slave craton. Much of the stratigraphy is similar to that overlying the CSBC (as in Fig. 9A), but is capped by a younger turbidite sequence with abundant silicate facies iron formations. F. Detrital zircon data (SHRIMP) for typical Burwash turbidites (lower panel, high up in the Burwash Formation sensu stricto, Yellowknife Domain), and the younger greywacke sequence of Damoti lake. Note that the latter contains a younger age peak, indicating a contribution from unroofing of early Defeat Suite-age plutons. Arrow highlights a single concordant, precise, zircon age determination obtained by Pehrsson and Villeneuve (1999).

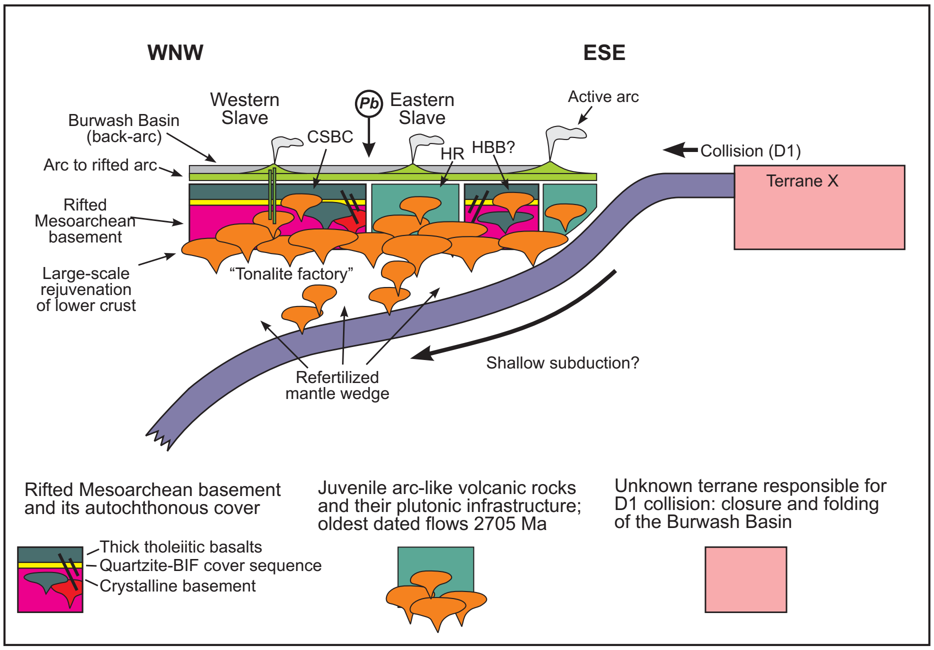

FIG. 10: Tectonic model for the general setting of the Slave craton between ca. 2690 Ma and 2660 Ma. All of the Slave appears to have been situated in a supra-subduction zone setting, with abundant and widespread calc-alkaline volcanism and plutonism. Arc-like assemblage (e.g., HR: Hackett River) were built across rifted basement and evolved into a large back-arc basin filled with turbidites, the Burwash Basin. Stratigraphically lowest, pre-2687 Ma components of the eastern arc-like domain could be exotic, but alternatively could represent juvenile volcanics in narrow back-arc rifts. The Hope Bay block (HBB) could represent a rifted fragment of the Central Slave Basement Complex (CSBC). Voluminous tonalite intrusions rejuvenated much of the lower and mid crust. Collision of an unknown terrane (X) led to closure and F1 folding of the Burwash turbidites.

- The earliest plutonic suite that intrudes folded Burwash strata, the ca. 2630 Ma Defeat Suite, appears to form a northeast-southwest-trending magmatic belt across the southeastern half of the craton (Fig. 11c).

With more and better U-Pb zircon ages, a tentative "volcanic line" of 2661 Ma felsic volcanic complexes, coeval with turbidite sedimentation, has begun to emerge (Bleeker and Davis, in preparation). This volcanic line also trends northeast-southwest and may represent the first recognition of a linear arc system.

Mineralization

The immature greywackes and mudstones of the Burwash Basin contain few primary mineral deposits other than banded iron formations (Fig. 5g). The latter occur intercalated in greywackes scattered across a broad swath in the northwestern part of the craton (Fig. 11c), from the Goose Lake and George Lake areas to the Point Lake area, and from there to the southwestern Slave. Many are highly magnetic. Although of scientific interest for the understanding of facies boundaries, basin evolution, and geochemistry, they are uneconomic in terms of their ferrous metal content.

The principal type of economic mineralization within Burwash Formation metaturbidites is epigenetic gold mineralization hosted by the intercalated banded iron formations. The most important example of this deposit type is the Lupin deposit on the southern shores of Contwoyto Lake, which has been a significant gold producer from 1982 to 2003, yielding 3-4 million ounces of Au (Normin, 2005). Other examples, e.g. George Lake and Goose Lake, occur throughout the northern Slave craton and may become economic with elevated gold prices and better access.

The general model for these deposits is that the host iron formations formed chemical traps for gold-bearing H_2O- CO_2 fluids during metamorphism and deformation. Destabilization of the Au-carrying sulphur complexes, due to interaction with reduced Fe-rich host rocks, led to alteration and gold deposition, either in veins or in fluid-altered and sulphidized zones of the iron formations. The structural timing of these epigenetic deposits is generally syn- to late-kinematic and syn- to late-metamorphic, i.e. consistent with maximum fluid production deeper in the telescoped structural-metamorphic pile. The most likely source for the fluids, and the gold, is metamorphic devolatilization of a voluminous, immature sediment pile and its volcanic substrate. Sporadic iron formations provide accidental traps to the migrating fluids, with discrete structures locally playing a role in increased focusing of fluid flow.

Similar processes also led to gold-bearing quartz veins within metaturbidites (e.g., laminated veins along sheared bedding planes, saddle reefs), but without a specific focusing mechanism these occurrences and deposits tend to be of small size, although locally of high grade. Examples are the Ptarmigan and Discovery mines in proximity to Yellowknife (Fig. 2).

Ca. 2.65-2.63 Ga Closure of the Burwash Basin

Turbidite sedimentation in the Burwash Basin came to an end sometime before 2650 Ma, the age of the oldest recorded granitoid pluton intruding Burwash strata (Point Lake area; W. Mueller, pers. comm.). Subsequent tectonic events record the closure and folding of the Burwash Basin (D1) prior to 2634 Ma (see F1 fold belt in Fig. 11b). The latter age constraint is provided by early plutons of the Defeat Suite, a distinct and possibly subduction-related magmatic suite across the southern (and southeastern) Slave craton (Fig. 11c).

Closure of the highly extended, but largely ensialic back-arc basin allowed considerable shortening and mobility but with a structural style dominated, at least at high structural levels, by fairly systematic, mostly upright, northeast-southwest trending fold trains. At deeper levels, e.g. along the basement-cover interface, the fold trains must have been detached allowing differential shortening of the basement

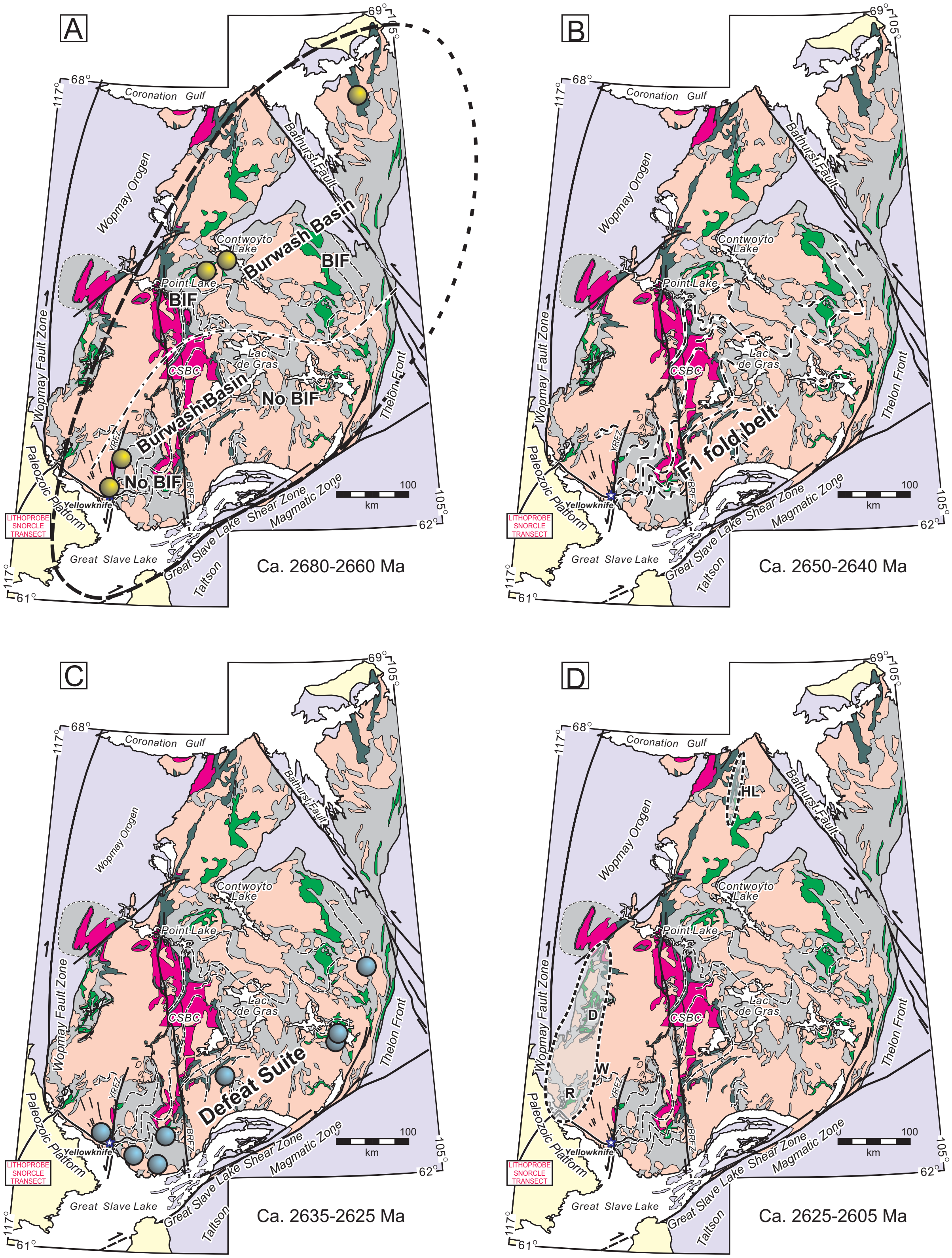

FIG. 11: Thematic maps illustrating key stratigraphic and structural aspects of the Slave craton through time. A. Minimum extent of the ca. 2680-2660 Ma Burwash Basin, based on continuity and geochronological similarity of turbiditic greywackes across large parts of the craton. Dash-dot line separates areas with intercalated iron formations (NW) from those lacking iron formations (SE). Yellow spheres highlight localities with precisely dated 2661 Ma volcanism closely associated with turbidite sedimentation. The linear trend may reflect a 2661 Ma magmatic line in a general arc-like setting. B. General trends of the F1 fold belt in the Burwash Formation and correlatives, trending NE-SW across the craton. C. Defeat Suite plutons dated between 2635 Ma and 2625 Ma. This apparent trend of arc-like plutons parallels the F1 fold belt. D. Areas in the Slave craton with younger volcano-sedimentary packages, that post-date deposition and folding of the Burwash Formation: the ca. 2620 Ma Damoti Lake (D) assemblage and the ca. 2612-2616 Ma High Lake (HL) assemblage. The Damoti Lake assemblage may extend to Russel Lake (R) and possibly Wheeler Lake (W). Its full extent is not known.

and cover.

The folded Burwash strata do not represent an outboard accretionary prism (which would require a trench setting rather than the more likely back-arc setting), and there is no evidence for a discrete "Contwoyto terrane" (cf. Kusky, 1989). The northeast-southwest structural grain of the F1 fold belt is also recognized in the lithospheric mantle (Grütter et al., 1999). Shallow subduction (either from the SE or NW?) may have emplaced distinct mantle slabs (Davis et al., 2003). These processes terminated with docking of an outboard terrane (e.g., Fig. 10), either in the southeast or the northwest; but this terrane is not preserved, however, within the exposed Slave craton. Crustal thickening led to uplift and erosional exhumation of folded Burwash strata and the unroofing of Defeat Suite plutons. Detrital zircons of Defeat Suite age are recorded in younger sedimentary packages (e.g., Fig. 9f).

Mineralization

Folding, incipient crustal thickening, and the onset of regional metamorphism, together with Defeat Suite plutonism, must have initiated devolatilization reactions and metamorphic fluid flow. These events thus likely kick-started the development of epigenetic gold mineralization, but were followed by much more intense metamorphic events ca. 20-30 million years later, during D2.

Arc-generation and subduction processes almost certainly modified the mantle lithosphere below the Slave craton, possibly creating the starting conditions for what is now a thick diamondiferous mantle root (e.g., Davis et al., 2003). Interestingly, trends of similar mantle domains, based on indicator mineral chemistry, appear to parallel the northeast-southwest trends of the Burwash Basin and D1 folding (Grütter et al., 1999).

Post-2.63 Ga Turbidites

Along the northwestern margin of the craton, younger turbidites containing ca. 2630 Ma detrital zircons (Figs. 9e, f and 11d; Sircombe and Bleeker, unpublished SHRIMP data; 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. 2616-2608 Ma tonalite-granodiorite plutons of the Concession Suite.

In the multiply folded, metamorphosed, and intermittently exposed terrain of the western Slave craton, it has proven difficult to distinguish these younger turbiditic greywackes from Burwash Basin turbidites. There is no sharply defined demarcation line that separates the two turbidite packages and recognition of the younger sequence relies largely on the absence of Defeat Suite-age plutons and the presence of <2640 Ma detrital zircons. Preliminary work suggests that the younger turbidite sequence contains abundant intercalated iron formations, mostly of silicate facies, those of the Damoti Lake area representing one of the more significant examples. Many of the iron formations are "lean", comprising background turbiditic greywacke variably enriched in metamorphic garnet, other Fe-rich silicates, and/or disseminated sulphides.

A distinct, post-Burwash Basin, greywacke and/or volcaniclastic sediment package, associated with felsic volcanic rocks and subvolcanic intrusions, and dated at approximately 2716-2712 Ma, occurs along the tightly folded synclinal core of the High Lake greenstone belt of the northern Slave craton (Henderson et al., 2000; see Figs. 9d and 11d). This package is of significance in that it is one of the few examples of a preserved volcano-sedimentary carapace to one of the major plutonic suites, i.e. the coeval Concession Suite.

Mineralization

Types of mineralization within the younger (turbiditic) greywacke packages are similar to those in folded Burwash Basin strata. A principal example of epigenetic gold mineralization is that hosted by silicate facies iron formation in the Damoti Lake area. Similar iron formations occur all along the southwestern edge of the Slave craton, from the Emile River area in the north to the Russel Lake area in the south, and have been moderately explored for gold and base metals. Scattered gold mineralization also occurs in numerous "lean" iron formations throughout the western Slave, e.g. the Wheeler and Germaine Lake areas ("W" in Fig. 11d). The stratigraphic status of the lean iron formations and their host turbidites in the Wheeler-Germaine Lake areas is currently unresolved.

2.60-2.58 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. These processes culminated 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 ?). Overall timing relationships are summarized in Figures 12a and b.

While peak temperatures were attained in the lower crust, large basement-cored domes were amplified by buoyancy driven deformation (Fig. 3); 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 may have formed as late as ca. 2580 Ma (Sircombe and Bleeker, unpublished SHRIMP data; see detrital zircon age spectra in Fig. 12b). 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.

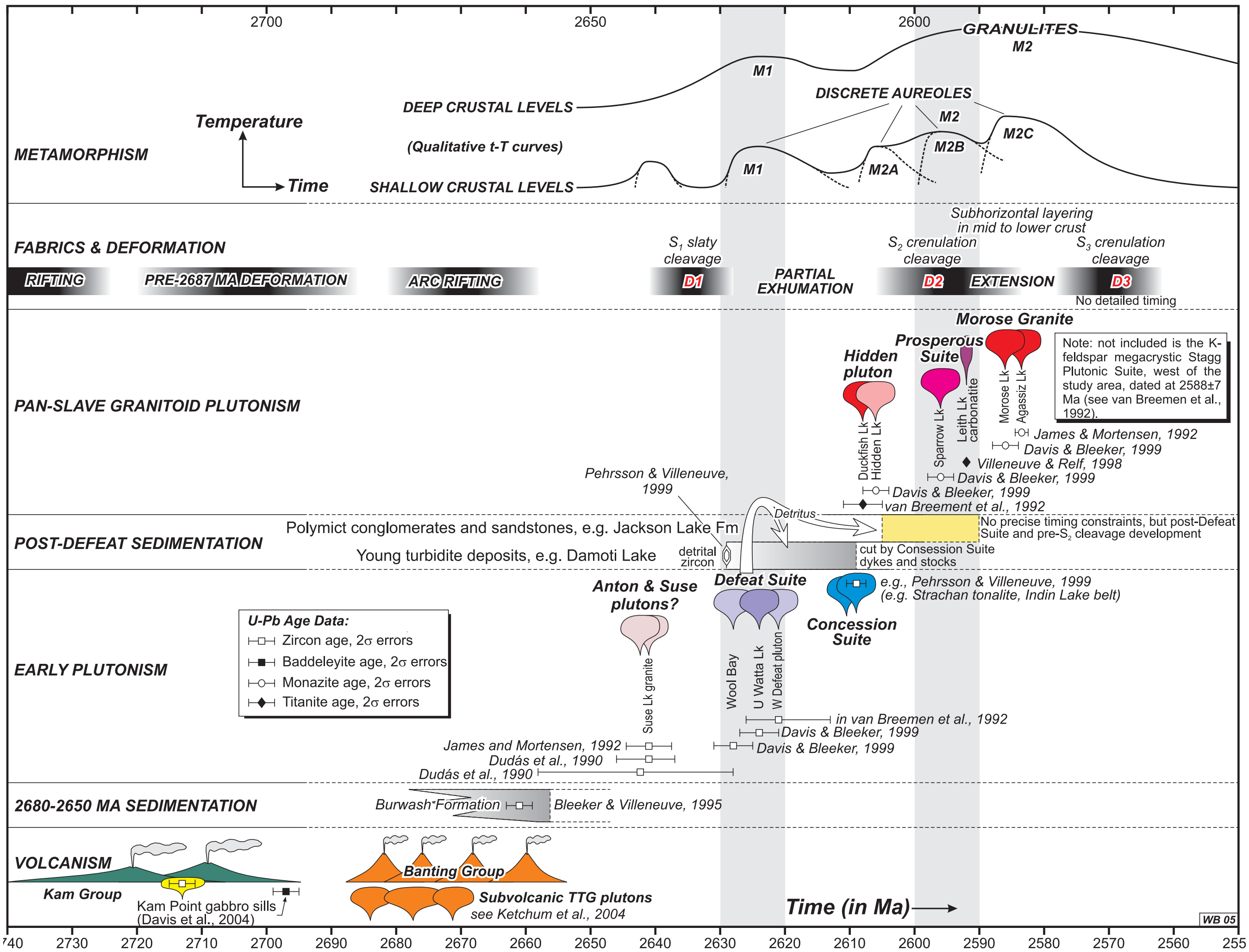

FIG. 12: Time charts of stratigraphic and structural events in the Slave craton. A) Detailed chart for key events in the Yellowknife Domain (updated from Davis and Bleeker, 1999).

Mineralization

Strong penetrative regional deformation, culminating between 2600 Ma and 2590 Ma, as determined from syn-kinematic granite sheets (Davis and Bleeker, 1999), represents the most obvious deformation throughout most of the Slave craton. It must have driven moderate to significant crustal thickening and led to the main thermal peak of regional metamorphism in most areas. This D2 deformation and the associated metamorphism were the main driver for epigenetic gold mineralization throughout the Slave craton. In the Yellowknife greenstone belt, it led to formation of the ca. 15 million ounce Con-Giant Au deposit, along a complex system of mostly reverse shear zones. As is typical for this class of deposits, the Con-Giant system occurs mostly within moderate to strongly deformed basaltic rocks, in proximity to a regional stratigraphic break, the Yellowknife River Fault Zone. An asymmetric synclinal panel of syn-orogenic conglomerates (the Jackson Lake Formation) occurs along this fault zone. Identical relationships are observed in several other major Archean gold camps, most notably Timmins, Kirkland Lake, and Kalgoorlie. The critical control common to all these camps is localization of Au mineralization within significant bends of the regional fault zones; these bends were most likely dilational during emplacement of the gold-bearing quartz veins.

Although numerous other volcanic-hosted gold vein systems are known from the Slave craton, some of which were briefly in production in the past, only one other major camp has emerged in recent years. This camp occurs in the Hope Bay belt, on the Coronation Gulf coast of the Slave craton, and consists of a string of deposits (Boston, Doris, Madrid) that are being readied for production. Elsewhere, despite significant past exploration, overall potential for this class of deposits remains excellent. Several greenstone belts throughout the Slave craton have very similar structural-stratigraphic characteristics to that of the Yellowknife belt, including a thick, folded turbidite pile adjacent to a basaltic greenstone belt, a regional deformation zone, and a young conglomerate package. Best examples are the Point Lake and Arcadia Bay areas.

Another class of mineral deposits related to final orogenesis is that of rare-element enriched granitoids, particularly highly evolved anatectic granites and their pegmatites. Tin (cassiterite) and Li (spodumene) were briefly mined from such pegmatites in the Yellowknife Domain, but other peg-

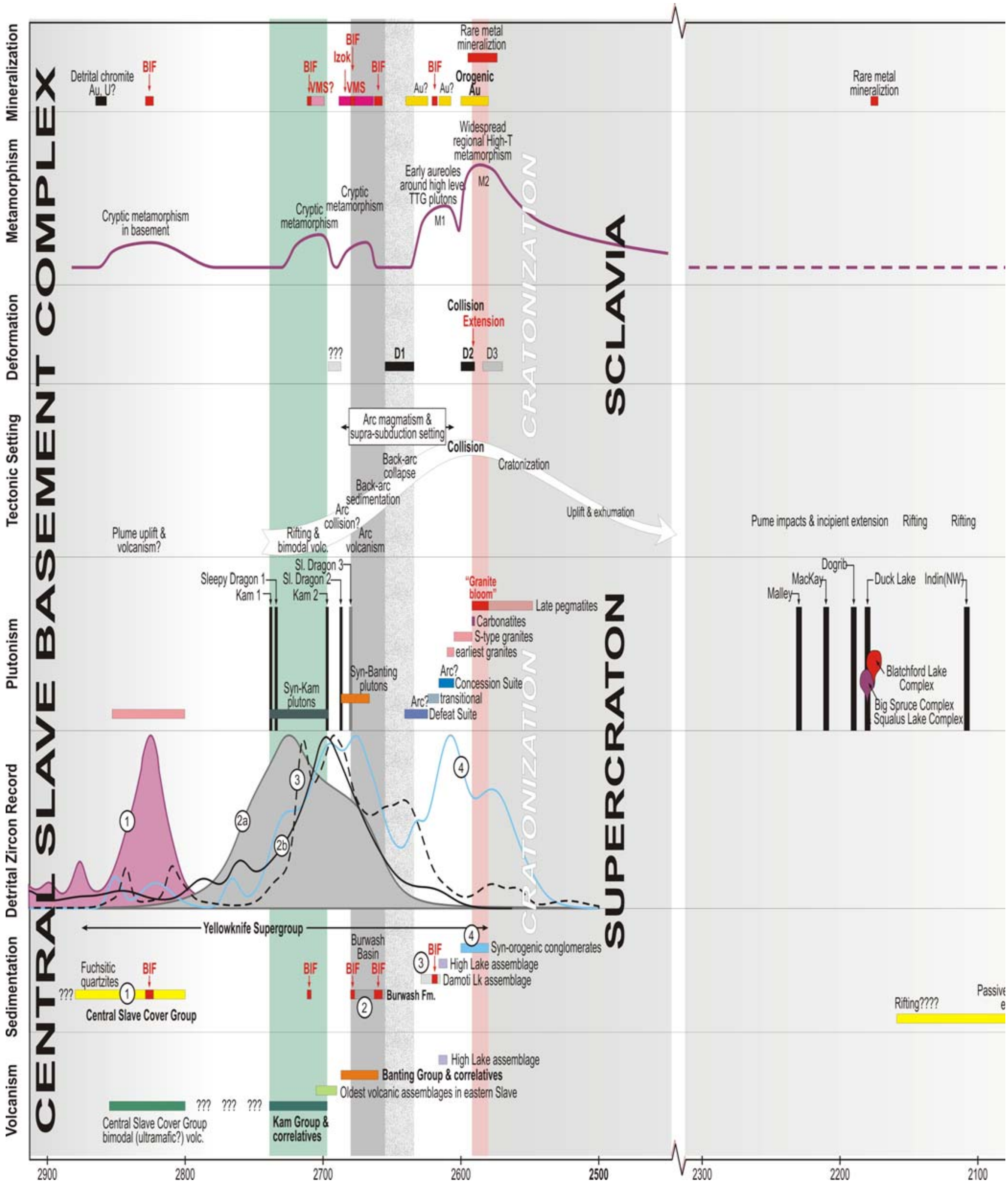

FIG. 12: B) Extended time chart for entire Slave craton (excluding pre-2900 Ma history of the Central Slave Basement Complex). Note breaks in horizontal scale (time) at two places. Typical detrital age spectra (1, 2a, b, 3, and 4) are shown for some of the main stratigraphic units: 1, quartzites of the Central Slave Cover Group; 2, turbiditic greywackes of Burwash Formation (s.s., 2a) and correlatives in the eastern Slave (Back River, 2b); the Damoti Lake turbidites (3); and syn-orogenic conglomerates (4). Known mafic dyke swarms are shown by narrow black bars (or grey, if poorly dated).) (Figure continued on next page)

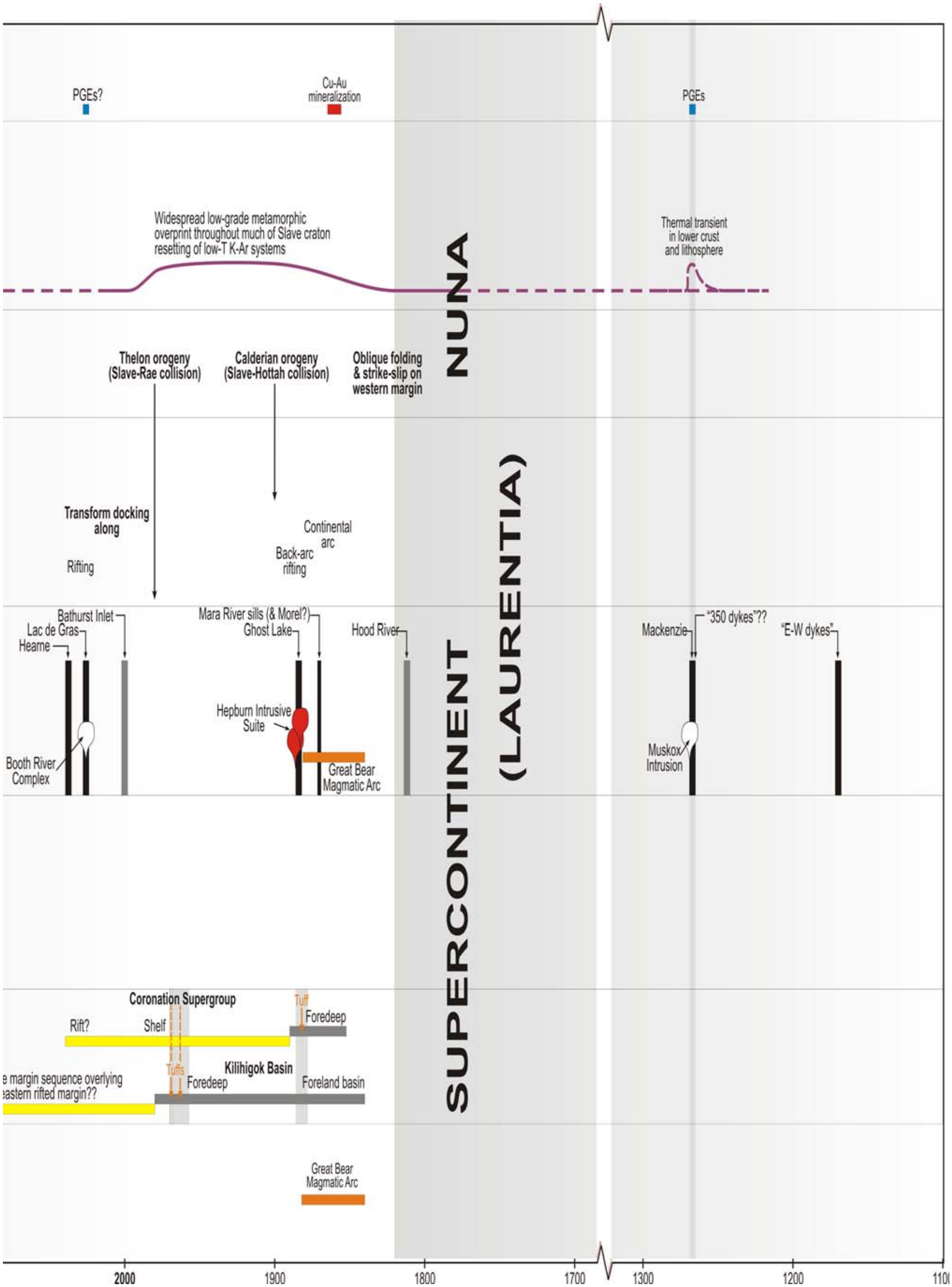

FIG. 12B) (continued): (For caption see previous page.)

matite fields are known throughout the Slave craton. From a global metallogeny point of view, these occurrences are of interest as they typically correlate with ancient, multiply recycled, felsic crust.

Cratonization and Beyond

The youngest significant granite plutons of the Slave craton are ca. 2585-2580 Ma. Only a few pegmatites are known to be significantly younger. Between ca. 2590 Ma and 2580 Ma, an enormous volume of granite was generated throughout much of the Slave craton. This "granite bloom", driven by moderate tectonic overthickening (D1-D2) and high intrinsic heat production, irreversibly transferred a significant fraction of heat-producing elements and lower crustal fluids (and Au) to the upper crust. In the lower crust, it must have involved large-scale migration of anatectic granitoid magmas, significant horizontal channel flow of partially molten rocks, development of a horizontal layering, and flattening of the Moho discontinuity into a stable density configuration. Collectively, these processes allowed the lower crust to cool and stiffen, over several tens of millions of years. U-Pb geochronology of lower crustal xenoliths shows that high-grade metamorphic reactions and zircon growth continued to about 2510 Ma at depth (Davis et al, 2003b). Finally, sufficient cooling allowed the crust to mechanically couple with the mantle (Bleeker, 2002). The end product was cratonic crust of high relative strength.

Following cratonization, there is a ca. 300 million time interval for which there are few recorded events within the Slave craton (Fig. 12b). Ca. 2.45 Ga magmatism, known from many other cratons around the world (e.g., Heaman, 1997), so far appears to be absent from the Slave craton.

At 2230 Ma, the northeast-trending Malley mafic dyke swarm, transecting the central Slave craton, provides the first evidence for mantle-driven magmatism and attempted rifting events. Involving at least ten different dyke swarms and associated extension events, the Slave craton finally broke out of its ancestral Sclavia supercraton between 2200 Ma and 2000 Ma. Details remain sketchy. Almost certainly, the eastern margin of the Slave craton, now involved in and overridden by the Thelon orogen, became established as a passive margin well before the western margin. The latter is referred to as the so-called Coronation margin and later became involved in Wopmay orogen (e.g., Hoffman, 1980; Bowring and Grotzinger, 1992; Hildebrand and Bowring, 1999). Once liberated out of the confines of its ancestral supercraton, the Slave continental microplate must have experienced a drift phase as an independent craton, before being progressively incorporated into the growing Laurentian collage and the supercontinent of Nuna.

Paleoproterozoic amalgamation processes varied along the margins of the Slave craton. In the east, the Slave acted as a lower plate, being overridden by the west-vergent Thelon orogen. In the south, along the shores and islands of Great Slave Lake, deformation was mainly transpressional along a long-lived transform boundary. In the west, at least two arc terranes (Hottah and the Great Bear Magmatic Zone) were involved, followed by oblique folding and late-stage, dextral, strike-slip deformation along the Wopmay Fault Zone. Development of the Great Bear arc, between about 1880 Ma and 1840 Ma, likely involved subduction of Paleoproterozoic lithosphere below the western Slave craton (e.g., Bostock, 1998).

Post-dating the assembly of Laurentia and Nuna, the Slave craton, particularly along its margins, became partially buried beneath intra-continental Proterozoic basins. At ca. 1269-1267 Ma, the craton was partly uplifted and intruded by the giant Mackenzie dyke swarm, radiating from a plume center west of Victoria Island (Barager et al., 1996; LeCheminant and Heaman, 1989). This is the last major event affecting the core of the craton, although some younger mafic magmatic events affect its edges (e.g. the ca. 780 Ma Hottah sheets). Since that time, Slave crust has been "bobbing" gently up and down, with interior seas expanding and receding across the craton. Ordovician and Cretaceous sedimentary rocks and fossils are known as wall rock fragments in some of the central Slave kimberlites.

Despite the relative stability at the surface, melting events were triggered in the subcontinental mantle lithosphere, leaving their traces as clusters of kimberlites across the craton. From the several hundreds of kimberlites now known across the craton, the following ages have been recorded: Cambrian, Siluro-Ordovician, Permian, Jurassic, Cretacous, and finally Eocene (e.g., Heaman et al., 2003). It are Eocene (ca. 55-50 Ma) kimberlite pipes of the Lac de Gras area in the central Slave craton that now support two highly profitable diamond mines, Ekati and Diavik. Several other diamond mines are in various stages of development. In just over a decade, diamonds have become the most profitable commodity within this ancient craton.

Summary

The Slave craton is a relatively small Archean craton with a geological knowledge base that is relatively mature. However, the following major questions remain:

1. What is the nature of the Hope Bay block in the northeast part of the craton? Does cryptic ancient basement reappear in this part of the craton? Is it perhaps a rifted fragment of the Central Slave Basement Complex?
2. What is the tectonic significance of pre-2687 Ma volcanic rock in the eastern Slave? Do they form remnants of a ca. 2.7 Ga, exotic, intra-oceanic juvenile arc that collided with the extended Central Slave Basement Complex between 2697-2687 Ma? If so, where is the suture? Or do these volcanics represent the oldest fill of narrow backarc-like troughs, formed by progressive rifting of the Central Slave Basement Complex in an overall arc setting?
3. What is the detailed outline of particular magmatic (e.g., Defeat and Concession suites) and sedimentary (e.g., the post-Defeat turbidite basin along the western margin of the Slave) belts?
4. How did events inferred from the crustal evolution contribute to or interfere with formation of the subcontinental mantle lithosphere below the Slave craton?
5. How far does Slave mantle lithosphere extend below the Rae craton to the east?

6. What is the detailed break-up history for each of the margins of the Slave craton? In other words, how was the Slave fragment liberated out of the supercraton Sclavia?
7. What is the detailed depositional record associated with rifting, thermal subsidence, and finally collision, along each of the margins of the Slave craton.?
8. And does Slave lithosphere extend all the way to the Innutian front (Fig. 1) in the high arctic?

To many of these first-order questions, we currently have only rudimentary answers. More sophisticated answers will require more complete and more refined data sets. In particular, a greatly expanded geochronological database, both in quantity and precision, in conjunction with targeted field work across the craton and its marginal belts, would quickly advance the state of knowledge.

In terms of mineral potential, much of the craton and all significant greenstone belts have seen at least a first wave of exploration for major commodities. These investigations quickly discovered a number of large VMS deposits (e.g., Izok Lake), which await road access for economic production. Gold potential remains high, particularly in more remote greenstone belts that may not have seen the required level of drill testing. In this respect, the Point Lake greenstone belt and its extension further north appears attractive as it has all the major attributes of a world-class gold camp.

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Appendix 1

Appendix 1 (in progress)lists approximately 50 of the principal ore deposits and showings of the Slave craton and its marginal belts. It represents an attempt to collect much of the relevant information on a single sheet of paper, although many of the entries are still incomplete. The NORMIN database (Narmin, 2005) and Padgham (1992) were used as starting points. Deposits are grouped into genetic types and by approximate stratigraphic order, with ilders mineralization types at or near the bottom of the table. Genetic groupings are discussed in the main text.

Appendix 1: Listing of main mineral deposits and showings by mineralization type and approximate stratigraphic or time sequence.

Deposit Type & Name	# *	Principal Commodity	Status	Host rocks	Age
Diamonds in kimberlites:					
Ekati	26	Gem diamonds	Producing mine, several kimberlite pipes	Kimberlite pipes, Eocene	52-56 Ma
Diavik	27	Gem diamonds	Producing mine	Kimberlite pipes, Eocene; mostly volcaniclastic facies kimberlite	52-56 Ma
Jericho	28	Gem diamonds	Permitting stage	Kimberlite pipes, Jurassic	173 Ma
Snap Lake	29	Gem diamonds	Producing mine	Kimberlite dykes, Siluro-Ordovician	523-535 Ma
Kennady Lake (Gahcho Kue)	30	Gem diamonds	Advanced exploration	Kimberlite pipes, Cambrian	542 Ma
Proterozoic hydrothermal Cu-Au (IOCG):					
Lou Lake, NICO	23	Cu, Au, Co, Bi	Feasibility	Disseminated and vein-type mineralization, calc-alkaline intrusion related	ca. 1860 Ma
Sue-Diane	24	Cu, Au, Ag, U, Fe	Advanced exploration	Volcanic-hosted, calc-alkaline suite	ca. 1860 Ma
Proterozoic alkaline intrusion-related rare-element deposits:					
Thor Lake	22	Ta, Nb, Be	Advanced exloration	Blatchford Lake intrusive complex	ca. 2180-2175 Ma
Proterozoic mafic intrusion-related deposits:					
Muskox Intrusion	25	PGEs	Advanced exploration	Layered mafic intrusion, lopolithic dyke	1269 Ma
Booth River Complex		V, PGEs?	Exploration?	Large layered intrusion	2026 Ma
Late Archean rare-element enriched granites and pegmatitites:					
Hidden Lake, Prosperous Lake	20	Li, Be	Exploration trenches	Late stage pegmatite dykes, probably Prosperous Suite related	ca. 2595 Ma
Upper Ross Lake (Peg Tantalum)	21	Be, Ta, Nb, Sn	Producer during WWII	Late stage granites and pegmatites, Redout Suite	ca. 2592 Ma
Gold in turbidite-hosted BIF:					
Lupin	16	Au-Ag	Past producer (1982-2003); c&m	Iron formations hosted by Contwoyto Formation turbidites	ca. 2665-2660 Ma
George Lake	17	Au	Advanced exploration	Silicate (oxide) BIFs in low-grade metaturbidites (Beechey Lake Group)	2680-2660 Ma
Goose Lake	18	Au	Advanced exploration	Silicate (oxide) BIFs in low-grade metaturbidites (Beechey Lake Group)	2680-2660 Ma
Damoti Lake	19	Au, Ag	Advanced exploration	Silicate facies BIFs in Damoti Lake assemblage turbidites	ca. 2620 Ma
Wheeler (Germaine) Lake area		Au, Ag	Exploration	Lean silicate BIFs in metaturbidites (Burwash Formation??)	ca. 2660 Ma??
Vein-hosted gold in folded turbidites:					
Discovery Mine	11	Au	Past producer (1949-1969)	Quartz veins in folded turbidites	ca. 2650-2600 Ma
Ptarmigan Mine	10	Au, Ag	Past producer; c&m	Quartz veins in folded turbidites	ca. 2650-2600 Ma
Shear and vein-hosted gold in volcanic rocks:					
Giant Yellowknife	9	Au, Ag	Past producer; c&m	Sheared and altered mafic volcanics	ca. 2600-2580 Ma
Con Mine	8	Au, Ag	Past producer; c&m	Sheared and altered mafic volcanics	ca. 2600-2580 Ma
Ormsby Zone		Au	Advanced exploration	Quartz veins and alteration in an amphibolite panel; rare pillow structures	
Nicholas Lake		Au, Ag	Advanced exploration	Veined and altered granodiorite	
Tundra Gold Mine	12	Au, Ag	Past producer; abandoned	Quartz veins along contact between felsic volcanics and Burwash Formation turbidites	
Tundra-Fat		Au	Advanced exploration	Sheared and altered felsic volcaniclastic rocks	
Salmita		Au, Ag	Past producer; abandoned	Veins in mafic volcanic package	
Arcadia	14	Au, Ag	Exploration, some zones drilled	Volcanic rocks and tonalite	
Hope Bay: Boston, Madrid, Doris	15	Au, Ag	Starting production	Sheared and altered mafic volcanics	
Colomac	13	Au, Ag	Past producer; c&m	Sheared and altered porphyries?	
Kim		Au	Advanced exploration	Sheared and altered mafic volcanics	
Volcanogenic massive sulphide deposits in arc-like volcanics:					
Izok Lake	1	Cu, Zn, Pb, Ag	Waiting for road access	Felsic volcanic complex	ca. 2684 Ma
Gondor	2	Zn, Ag, Cu, Pb, Au	Advanced exploration	Felsic and intermediate volcaniclastic rocks	
Yava	3	Zn, Cu, Ag, Pb, Au	Drilled	Felsic and intermediate volcaniclastic rocks	ca. 2680 Ma
Musk	4	Zn, Pb, Cu, Ag, Au	Advanced exploration	Felsic volcanic rocks	ca. 2680 Ma
Hackett River, A Zone		Zn, Pb, Cu, Ag, Au	Advanced exploration	Felsic volcanic rocks	ca. 2680 Ma
Hackett River, Boot		Cu, Zn, Pb, Ag			ca. 2680 Ma
Hackett River, Cleaver		Cu, Zn, Pb, Ag			ca. 2680 Ma
Hackett River, Stringer Zone		Cu, Zn, Pb, Ag			ca. 2680 Ma
High Lake, Ab	5	Cu, Zn, Ag	Advanced exploration	Felsic volcanics	ca. 2705 Ma
High Lake, D-Zone	5	Zn, Cu, Ag, Pb, Au	Advanced exploration		ca. 2705 Ma
High Lake, W-Zone	5	Cu, Zn, Pb, Ag	Advanced exploration		ca. 2705 Ma
Sunrise	6	Zn, Pb, Ag, Cu, Au	Advanced exploration	Felsic volcanic complex near volcanic-sedimentary interface	ca. 2670 Ma
Bear		Zn, Pb, Ag, Cu, Au	Advanced exploration		
Boot Lake		Cu, Zn, Pb, Ag			
Creek Zone Mat		Cu, Zn, Pb, Ag			
Deb		Cu, Zn, Pb, Ag	Advanced exploration		
East Cleaver Lake		Cu, Zn, Pb, Ag			
Hood #10		Cu, Zn, Pb, Ag			
Hood #10		Cu, Zn, Pb, Ag			
Hood #41		Cu, Zn, Pb, Ag			
Kennedy Lake (BB+Lk+Cuzone)		Zn, Pb, Ag, Cu	Advanced exploration		
Kennedy, 1. zone		Cu, Zn, Pb, Ag			
Kennedy, BB zone		Cu, Zn, Pb, Ag			
Kennedy, Cu zone		Cu, Zn, Pb, Ag			
Lark		Cu, Zn, Pb, Ag			
Lark		Cu, Zn, Pb, Ag			
Len		Cu, Zn, Pb, Ag			
Len		Cu, Zn, Pb, Ag			
Susu Lake		Cu, Zn, Pb, Ag			
Turnback, OK		Cu, Zn, Pb, Ag			
Turnback, XL		Cu, Zn, Pb, Ag			
Volcanogenic sulphides in bimodal rift volcanics of Kam Group type:					
Homer Lake 1	7	Ag, Zn, Pb, Au, Cu	Showings, trenches; drilled	Weakly mineralized felsic rocks in bimodal rift successio	ca. 2720-2700 Ma
Homer Lake 2		Ag, Zn, Pb, Au, Cu	Showings, trenches	Weakly mineralized felsic rocks in bimodal rift succession	ca. 2720-2700 Ma
Bell Lake		Cu, Zn	Showings	Weakly mineralized felsic rocks in bimodal rift succession	ca. 2720-2700 Ma
Courageous Lake		Cu, Zn	Showings	Weakly mineralized felsic rocks in bimodal rift succession	ca. 2720-2700 Ma
Point Lake		Cu, Zn	Showings	Weakly mineralized felsic rocks in bimodal rift succession	ca. 2720-2700 Ma
Banded iron formations:					
Dwyer Lake		Fe; epigenetic Au?	Prospective for Au?	Banded iron formations at the top of the Central Slave Cover Group	ca. 2826 Ma
Patterson Lake		Fe; epigenetic Au?	Prospective for Au?	Banded iron formations at the top of the Central Slave Cover Group	2850-2800 Ma
Amacher		Fe; epigenetic Au?	Prospective for Au?	Banded iron formations at the top of the Central Slave Cover Group	2850-2800 Ma
e.g., Point Lake; others		Fe; epigenetic Au?	Prospective for Au?	Minor banded iron formations within basalt-dominated volcanic packages	ca. 2720-2700 Ma
Contwoyto Formation		Fe; epigenetic Au	Prospective for Au?	Banded iron formations in turbidites	2680-2660 Ma
Back River		Fe; epigenetic Au	Advanced exploration for Au	Banded iron formations in turbidites	2680-2660 Ma
George Lake		Fe; epigenetic Au	Advanced exploration for Au	Banded iron formations in turbidites	2680-2660 Ma
Goose Lake		Fe; epigenetic Au	Advanced exploration for Au	Banded iron formations in turbidites	2680-2660 Ma
Damoti Lake		Fe; epigenetic Au	Advanced exploration for Au	Banded iron formations in Damoti Lake turbidites	2625-2615 Ma
Mesoarchean paleo-placer Au, U, Cr:					
Dwyer Lake		Cr as detrital chromite	Curiosity	Supermature quartzites of the Central Slave Cover Group	>2853 Ma
Other quartzites		Au, U paleo-placers?	Prospects?	Supermature quartzites of the Central Slave Cover Group	2900-2800 Ma

*#s refer to those in Figure 2.

Appendix 1(continued): Listing of main mineral deposits and showings by mineralization type and approximate stratigraphic or time sequence.

Deposit Type & Name	Geological Environment	Deposit Size	Metal Ratios	Other	Comments	Latitude	Longitude	NTS Sheet	References
Diamonds in kimberlites:									
Ekati	Lac de Gras area, central Slave craton, Lac de Gras structural basin					64.7167	-110.6064	076D10	Normin, 2005; Heaman et al., 2003 and references therein
Diavik	Lac de Gras area, central Slave craton, Lac de Gras structural basin					64.4997	-110.2372	076D08	Normin, 2005; Heaman et al., 2003 and references therein
Jericho	Contwoyto Lake, north-central Slave cratc								Heaman et al., 2002, 2003 and references therei
Snap Lake	South-central Slave craton, Central Slave Basement Complex					63.5925	-110.7281	075M10	Normin, 2005; Heaman et al., 2003 and references therein
Kennady Lake (Gahcho Kue)	Southeastern Slave craton					63.4358	-109.2100	075N06	Normin, 2005; Heaman et al., 2003 and references therein
Proterozoic hydrothermal Cu-Au (IOCG):									
Lou Lake, NICO	Southern Great Bear Magmatic Zone	ca. 42 Mtonnes				63.5483	-116.7586		Goad et al., 2000; Ghandhi et al., 2001
Sue-Diane	Southern Great Bear Magmatic Zone	ca. 17 Mtonnes				63.7586	-116.9128		Goad et al., 2000; Ghandhi et al., 2002
Proterozoic alkaline intrusion-related rare-element deposits:									
Thor Lake	Anorogenic (rift-related?) intrusion along southern margin					62.1161	-112.5969	085I02	Normin, 2005
Proterozoic mafic intrusion-related deposits:									
Muskox Intrusion	Proximal to Mackanzie event plume center								
Booth River Complex	North-central Slave craton, overlain by Kilihigok Basin					66.8394	-109.0731	076K14	Normin, 2005
Late Archean rare-element enriched granites and pegmatitites:									
Hidden Lake, Prosperous Lake	Yellowknife Domain, southwestern Slave craton					62.3081	-112.8036	085I07	Normin, 2005
Upper Ross Lake (Peg Tantalum)	Yellowknife Domain, southwestern Slave craton					62.7442	-113.1083	085I11	Normin, 2005
Gold in turbidite-hosted BIF:									
Lupin	Contwoyto Lake area, north-central Slave craton					65.7647	-111.2250		Normin, 2005
George Lake	Northeastern Slave craton, George Lake synclinorium					65.9258	-107.4764	076G13&14	Normin, 2005
Goose Lake	Northeastern Slave craton, George Lake synclinorium					65.5439	-106.4278	076G09	Normin, 2005
Damoti Lake	Indin Lake belt, western Slave craton					64.1383	-115.1136	086B03	Normin, 2005
Wheeler (Germaine) Lake area	Southwestern Slave craton, west of Yellowknife Domain								
Vein-hosted gold in folded turbidites:									
Discovery Mine	Yellowknife Domain, folded turbidites of the Burwash Formation					63.1883	-113.8972	085P04	Normin, 2005
Ptarmigan Mine	Yellowknife Domain, folded turbidites of the Burwash Formation					62.5192	-114.1972	085J09	Normin, 2005
Shear and vein-hosted gold in volcanic rocks:									
Giant Yellowknife	Yellowknife greenstone belt					62.4989	-114.3628	085J08&09	Normin, 2005
Con Mine	Yellowknife greenstone belt					62.4333	-114.3681	085J08	Normin, 2005
Ormsby Zone	Yellowknife Domain					63.1722	-113.9253	085P04	Normin, 2005
Nicholas Lake	Yellowknife Domain					63.2472	-113.7617	085P04	Normin, 2005
Tundra Gold Mine	Courageous Lake greenstone belt					64.0400	-111.1744	076D03	Normin, 2005
Tundra-Fat	Courageous Lake greenstone belt					64.1178	-111.2706	076D03	Normin, 2005
Salmita	Courageous Lake greenstone belt					64.0750	-111.2411	076D03	Normin, 2005
Arcadia	Anialik River greenstone belt, Coronation Gulf					67.6869	-111.3400	076M11	Normin, 2005
Hope Bay: Boston, Madrid, Doris	Hope Bay greenstone belt					67.6494	-106.3881	076O09	Normin, 2005
Colomac	Indin Lake greenstone belt					64.3975	-115.0856	086B06	Normin, 2005
Kim						64.3167	-115.2706	086B06	Normin, 2005
Volcanogenic massive sulphide deposits in arc-like volcanics:		Tonnage	Cu:Zn:Pb	Ag (ppm)					
Izok Lake	Northern Point Lake greenstone belt	10,800,000	15:77:8	66		65.6311	-112.7989	086H10	Normin, 2005; Padgham, 1992
Gondor	Central volcanic belt	7,500,000	6:87:7	45	avg. for 3 drill holes	65.5628	-111.7969		Normin, 2005; Padgham, 1993
Yava	Hackett River greenstone belt	1-2,000,000	12:75:13	103		65.6044	-107.9383	076G12	Normin, 2005; Padgham, 1994
Musk	Hackett River greenstone belt	300,000	10:79:11	343		65.3231	-107.6208	076G05	Normin, 2005; Padgham, 1995
Hackett River, A Zone	Hackett River greenstone belt	5,000,000	2:84:14	280		65.9172	-108.3631	076F16	Normin, 2005; Padgham, 1996
Hackett River, Boot	Hackett River greenstone belt	5,000,000	4:82:14	176					Normin, 2005; Padgham, 1997
Hackett River, Cleaver	Hackett River greenstone belt	4,000,000	5:83:12	160					Normin, 2005; Padgham, 1998
Hackett River, Stringer Zone	Hackett River greenstone belt	4,000,000	20:45:35	291	low grade				Normin, 2005; Padgham, 1999
High Lake, Ab	High Lake greenstone belt	2,400,000	83:17:0			67.3814	-110.8500	076M07	Normin, 2005; Padgham, 2000
High Lake, D-Zone	High Lake greenstone belt	2,800,000	35:62:3	33		67.3758	-110.8436	076M07	Normin, 2005; Padgham, 2001
High Lake, W-Zone	High Lake greenstone belt								Normin, 2005; Padgham, 2002
Sunrise	Beaulieu River greenstone belt	2,057,000	1:67:32	367	probable	62.9000	-112.3794	085I16	Normin, 2005; Padgham, 2003
Bear	Beaulieu River greenstone belt	809,700	1:72:27	218	29 holes	62.8919	-112.3931	085I16	Normin, 2005; Padgham, 2004
Boot Lake						67.1025	-110.8983	076M02	Normin, 2005; Padgham, 2004
Creek Zone Mat									Normin, 2005; Padgham, 2005
Deb		1,118,000	24:75:1	20	drilling	64.0017	-111.2325	075M14 & 076D03	Normin, 2005; Padgham, 2006
East Cleaver Lake									Normin, 2005; Padgham, 2007
Hood #10			20:80:1	30					Normin, 2005; Padgham, 2008
Hood #10		150,000	59:41:0		low grade				Normin, 2005; Padgham, 2009
Hood #41		300,000	27:73:0	16					Normin, 2005; Padgham, 2010
Kennedy Lake (BB+Lk+Cuzone)						63.0322	-110.9483	075M02	Normin, 2005; Padgham, 2011
Kennedy, 1. zone		70,000	0:86:14	150	trenches				Normin, 2005; Padgham, 2012
Kennedy, BB zone		150,000	100:0:0	?	drilling				Normin, 2005; Padgham, 2013
Kennedy, Cu zone		970,000	0:93:7	102	drilling				Normin, 2005; Padgham, 2014
Lark			4:89:7	?	2 drill holes				Normin, 2005; Padgham, 2015
Lark			5:81:10	?					Normin, 2005; Padgham, 2016
Len			0:32:68	54	2 drill holes				Normin, 2005; Padgham, 2017
Len			0:34:66	42	trenches				Normin, 2005; Padgham, 2018
Susu Lake		142,500	100:0:0	?	9 holes				Normin, 2005; Padgham, 2019
Turnback, OK			0:62:38	395					Normin, 2005; Padgham, 2020
Turnback, XL			12:72:16	157	25 holes				Normin, 2005; Padgham, 2021
Volcanogenic sulphides in bimodal rift volcanics of Kam Group type:									
Homer Lake 1	Yellowknife greenstone be		3:32:65	84	only VMS in Yellowknife bel	62.6544	-114.2994	085J09	Normin, 2005; Padgham, 202(
Homer Lake 2	Yellowknife greenstone belt		2:55:43	45	only VMS in Yellowknife belt				Normin, 2005; Padgham, 2021
Bell Lake	Yellowknife greenstone belt								
Courageous Lake	Courageous Lake greenstone belt								
Point Lake	Point Lake greenstone belt								
Banded iron formations:									
Dwyer Lake	At base of Yellowknife greenstone belt								
Patterson Lake									
Amacher	At base of Beaulieu greenstone belt								
e.g., Point Lake; others	Minor iron formations within basalt-dominated volcanic packages								
Contwoyto Formation	Contwoyto Lake area, north-central Slave craton								
Back River	Overlying the Back River volcanic complex and in turbitidites								
George Lake	Northeastern Slave craton								
Goose Lake	Northeastern Slave craton								
Damoti Lake	Indin Lake belt, western Slave craton								
Mesoarchean paleo-placer Au, U, Cr:									
Dwyer Lake									Bleeker et al., 1999a
Other quartzites									Roscoe, 1990