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Detrital chromites reveal Slave craton's missing komatiite

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ABSTRACT

Komatiitic magmatism is a characteristic feature of Archean cratons, diagnostic of the addition of juvenile crust, and a clue to the thermal evolution of early Earth lithosphere. The Slave craton in northwest Canada contains >20 greenstone belts but no identified komatiite. The reason for this dearth of komatiite, when compared to other Archean cratons, remains enigmatic. The Central Slave Cover Group (ca. 2.85 Ga) includes fuchsitic quartzite with relict detrital chromite grains in heavy-mineral laminations. Major and platinum group element systematics indicate that the chromites were derived from Al-undepleted komatiitic dunites. The chromites have low $^{187}Os/^{188}Os$ ratios relative to chondrite with a narrow range of rhenium depletion ages at 3.19 ± 0.12 Ga. While these ages overlap a documented crust formation event, they identify an unrecognized addition of juvenile crust that is not preserved in the bedrock exposures or the zircon isotopic data. The documentation of komatiitic magmatism via detrital chromites indicates a region of thin lithospheric mantle at ca. 3.2 Ga, either within or at the edge of the protocratonic nucleus. This study demonstrates the applicability of detrital chromites in provenance studies, augmenting the record supplied by detrital zircons.

INTRODUCTION

Quartzite—metamorphosed quartz arenite—is a common basal sedimentary unit in many Archean basement-cover sequences (e.g., Eriksson and Fedo, 1994). The preserved silicate minerals (e.g., detrital quartz and zircons) mostly reflect provenance from felsic to intermediate crust. Detrital zircon U-Pb dating is the preferred tool for provenance determinations and is especially powerful when combined with Lu-Hf and $\delta^{18}O$ data (e.g., Pietranik et al., 2008). However, this method does not constrain the source and flux of any mafic-ultramafic magma to the crust. Some quartzites display a distinct green-weathering color along heavy-mineral laminations, imparted by the mineral fuchsite [$K(Al,Cr)_2(AlSi_3O_{10})(OH)_2$]. Fuchsite forms by substitution of Cr^{3+} for Al^{3+} in muscovite (Sleep et al., 2011) after metamorphic breakdown of detrital chromite [$(Mg,Fe^{2+})(Cr,Al,Fe^{3+})_2O_4$]. Detrital chromites are locally preserved as relict grains within fuchsite, or in heavy-mineral laminations. Chromite crystallizes in mafic-ultramafic magmas and is notably resistant to weathering and transport compared to silicate minerals such as olivine and pyroxene. While detrital zircons provide the opportunity to resolve felsic-intermediate provenance, Re-Os isotopes are a better tracer of mafic-ultramafic crustal growth (e.g., Pearson et al., 2007), and detrital chromites provide the opportunity to document this provenance.

The Central Slave Cover Group (ca. 2.85 Ga) of the Slave craton, northwest Canada, contains fuchsitic quartzite with relict detrital chromite lamination. In this study, chromite mineral separates are studied using major elements, platinum group elements (PGEs), and Os isotopes. The data are used to infer that the provenance of the detrital chromites was from komatiites that are no longer preserved in the bedrock record of the Slave craton, and the findings are compared with crustal evolution models. Our study demonstrates the value of detrital chromite grains to appraise the nature and role of mafic-ultramafic magmatism in the evolution of the lithosphere.

CENTRAL SLAVE COVER GROUP

The Slave craton (Fig. 1A), has a preserved rock record that extends from the Hadean to the Neoarchean. The Central Slave Basement Complex (CSBC) is composed of Hadean to Mesoarchean (4.02–2.85 Ga) tonalite-granite gneiss (Fig. 1B), with locally preserved mafic bodies (Ketchum et al., 2004). The Central Slave Cover Group (CSCG; Fig. 1B) is a supracrustal succession that was deposited unconformably on the CSBC between ca. 2.85 and 2.83 Ga (Isachsen and Bowring, 1997; Bleeker et al., 1999; Sircombe et al., 2001). The CSCG is correlated over hundreds of kilometers and from base to top (Fig. 1B) contains pebble conglomerate, quartzite (locally fuchsite-bearing), felsic volcanic rocks, and banded iron formation (BIF); thin ultramafic sills occur locally (Bleeker et al., 1999). Volcanic and sedimentary rocks of the Yellowknife Supergroup (ca. 2.73–2.60 Ga) were deposited, or structurally emplaced, on top of the CSCG. All supracrustal rocks in the Slave craton were polydeformed and metamorphosed, from greenschist to granulite grade, around 2.60 Ga; the CSCG rocks in this study (Bell and Dwyer Lakes; Fig. 1A; Fig. S1 in the Supplemental Material[1]) were metamorphosed to lower amphibolite facies.

The bedrock record of the Slave craton lacks komatiite, a distinguishing feature of volcanic belts in many other Archean cratons (Padgham, 1992; Campbell and Davies, 2017). Sedimentological reconstructions of the CSCG show that it was deposited during continental attenuation of the CSBC, with quartzites deposited in a dynamic tide-influenced estuary while BIFs were deposited in a deeper-water, shelf environment (Mueller et al., 2005). Detrital zircon U-Pb data from CSCG quartzites show ages from Hadean to Mesoarchean (Sircombe et al., 2001) and Hf-depleted mantle age modes spanning the same range (Pietranik et al., 2008).

[1]Supplemental Material. Detailed field observations, chromite separation techniques, analytical techniques (including Microprobe, LA-ICP-MS, Re-Os isotopes, and PGE chemistry) and data tables. Please visit https://doi.org/10.1130/GEOL.S.14524245 to access the supplemental material, and contact editing@geosociety.org with any questions.

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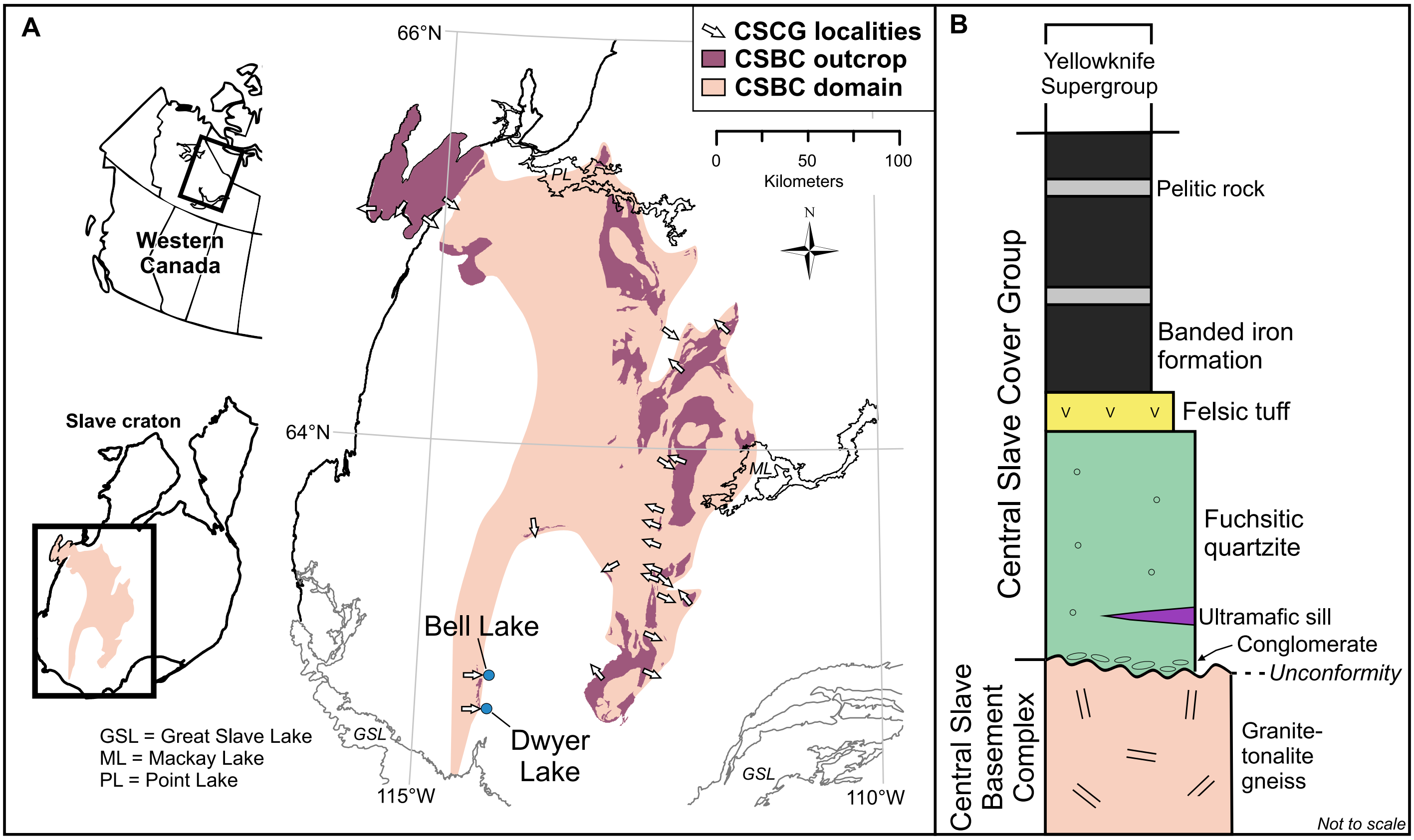

Figure 1. (A) Distribution of the Central Slave Basement Complex (CSBC) and Central Slave Cover Group (CSCG) in Slave craton, northwest Canada. Sample locations for Bell and Dwyer Lakes are identified. (B) Generalized stratigraphy of the CSCG.

SAMPLES AND METHODS

We collected chromite-bearing fuchsitic quartzite samples from CSCG outcrops at Bell and Dwyer Lakes (Fig. 1A; Fig. S1). Analytical work was conducted at the University of Alberta, Canada. The chromite-rich laminations in seven samples were examined petrographically, and major oxides of chromite grains were analyzed in thin sections and mineral separates using a JEOL 8900 electron probe microanalyzer (EPMA). Time-series spectra of PGEs were obtained by laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS). Four samples were selected for heavy mineral separation and chemistry: three from Bell Lake and one from Dwyer Lake. The chromite separates were pooled by sample in ~0.1 g multigrain aliquots and leached in HCl before digestion for PGE and isotope analyses. Osmium (Os) isotopes and abundances were measured by negative thermal ionization mass spectrometry (N-TIMS) using Faraday cups equipped with 10^{12} Ω amplifiers (Liu and Pearson, 2014) on a Thermo Fisher Triton Plus at the Arctic Resources Geochemistry Laboratory (ARGL), University of Alberta. PGE and Re abundances of the same bulk-rock dissolution aliquots were analyzed using isotope dilution techniques on either an Element 2 ICP-MS or Nu Attom ICP-MS at the ARGL. Field observations, chromite separation, analytical techniques, and data tables are documented in the Supplemental Material.

PETROGRAPHY, COMPOSITION, AND Re-Os MODEL AGE OF DETRITAL CHROMITES

The detrital chromites define thin black laminations (1–3 mm) with variable cluster density and distance between individual laminae (Figs. 2A and 2B). Individual chromites range from 0.16 to 0.40 mm (average 0.25 mm) and are coarser than accessory rutile and detrital zircon (Fig. 2C). Chromites are surrounded by fuchsite (Figs. 2D and 2E) and exhibit subangular (Figs. 2F–2J) to subrounded morphologies (Figs. 2K–2N), with some grains possessing relict octahedral spinel shapes (Figs. 2F and 2G). Evidence of transport and reworking, probably by river and wave action, is displayed by abrasive rounding with multiple pits and microfracturing along grain boundaries (Figs. 2L–2N). A few chromites have exsolution lamellae (~1–2 µm; Fig. 2L) and multiple shaped inclusions (<5 µm) of silicates and sulfides (Figs. 2M and 2N).

The chromites have Cr# between 0.58 and 0.70 (average 0.64; $n = 90$) and extremely low (<0.05) Mg# (Fig. 3A; Table S1). The Mg# and substitution of Zn^{2+} and Mn^{2+} for Fe^{2+} (Fig. S2) indicate modification of the less robust divalent ions, as is typical for metamorphosed chromite (Barnes, 1998, 2000). The magmatic Al^{3+}-Cr^{3+} trend (Fig. 3B) indicates that during prograde metamorphism, the alteration of chromite to Al-poor chromian-magnetite (ferritchromite) was insignificant (e.g., Wang et al., 2005). This is supported by unzoned core-to-rim patterns of the trivalent ions (Cr^{3+}, Al^{3+}, Fe^{3+}; Fig. S2; Table S1). In particular, the low and constant Fe_2O_3 (Fig. 3C) precludes major re-equilibration between primary chromite and chromian-magnetite, which would lead to an increase in Fe^{3+} and simultaneous decrease of both Cr^{3+} and Al^{3+} in the spinel structure (Barnes, 2000; Wang et al., 2005). The multigrain chromites have chondrite-normalized bulk Os-Ir-Ru abundances that are elevated relative to Pt-Pd, and a positive Ru anomaly (Fig. 3D). Laser ablation of individual chromite grains revealed a generally uniform PGE distribution, specifically for Os-Ir-Ru, in time-series spectra (Fig. S3).

The chromites have low $^{187}Os/^{188}Os$ values of 0.1051–0.1057 (Table S2). Rhenium is consistently very low (less than the limit of detection of 7 pg; Table S2), making it possible to establish rhenium depletion model ages (T_{RD}) between 3.21 Ga and 3.17 Ga (3.19 ± 0.06 Ga, 2σ weighted mean) using an ordinary chondrite reference mantle model (Walker et al., 2002). Propagating an ~0.10 b.y. uncertainty that arises

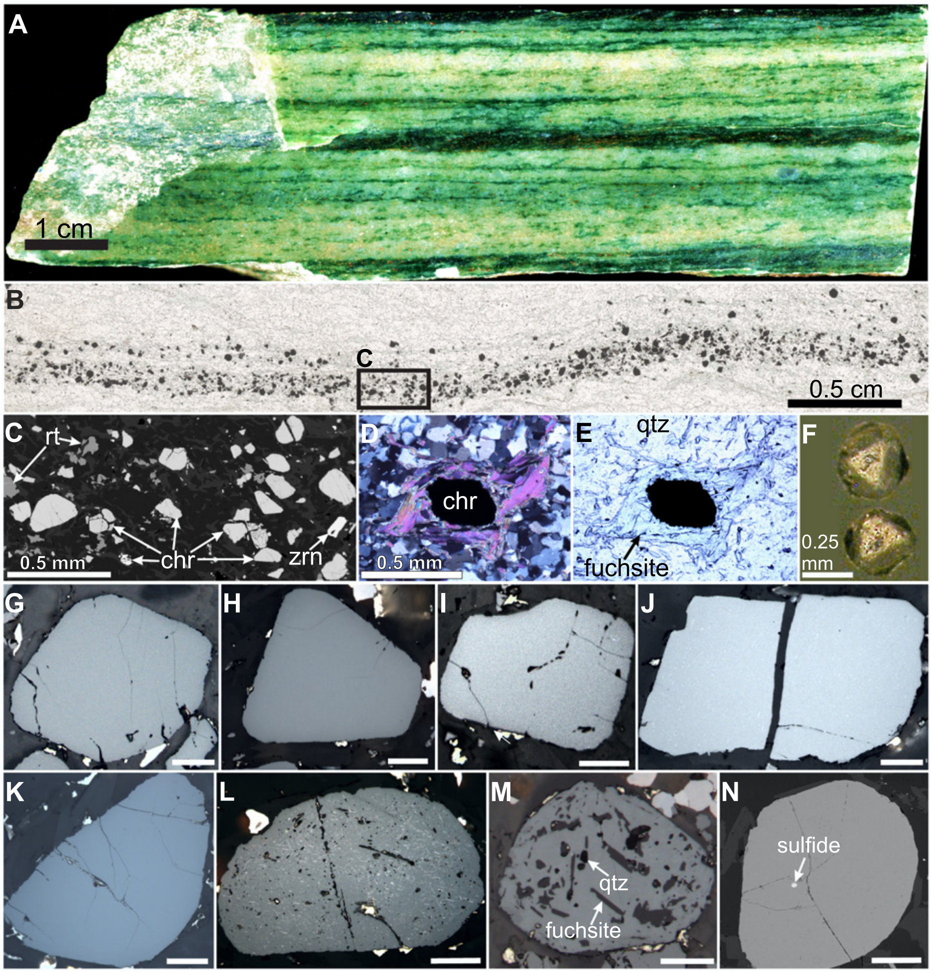

Figure 2. (A) Quartzite sample from Bell Lake (northwest Canada) with laminae accentuated by fuchsite and detrital heavy minerals. (B) Photomicrograph (plane-polarized light, PPL) of chromite-rich laminae (opaque) in quartzite. (C) Backscattered-electron image (inset in B) showing variably shaped detrital chromite (chr) and zircon (zrn) grains, and rutile (rt). (D) Crossed-polarized light and (E) PPL photomicrographs of single chromite within fuchsite; qtz—quartz. (F) Photomicrograph (incident light) of two chromites displaying spinel symmetry. (G–N) Photomicrographs (reflective light) of chromite displaying subangular morphologies (G–J) and subrounded morphologies (K–N). Scale bar = 50 μm. A minority of grains contain exsolution microlamellae (L) and inclusions of silicate (M) and minor sulfide (N).

from uncertainty in the mantle reference reservoir yields a mean age of 3.19 ± 0.12 Ga.

PROVENANCE OF DETRITAL CHROMITES

The CSCG detrital chromite grains show compositional similarities in Cr# and Al-Cr-Fe^{3+} systematics to metamorphosed chromite grains from komatiites, rather than boninite, orogenic peridotite, mid-oceanic ridge basalt (MORB), or layered igneous complexes (Fig. 3C). The Cr# is similar to metamorphosed chromites from Al-undepleted komatiite, possibly of komatiitic dunite affinity (Cr# ∼60–85), rather than Al-depleted komatiite (Cr# >85; Fig. 3C; Barnes and Roeder, 2001). Komatiite provenance versus komatiitic basalt is supported by (1) PGE patterns that resemble komatiite chromites (except for a smaller positive Ru anomaly; Fig. 3D), and (2) uniform PGE distribution in chromite as revealed by time-series ablation data that demonstrate Os-Ir-Ru are hosted as solid solution in the chromite structure rather than as PGE-bearing nuggets (Fiorentini et al., 2004). The small Ru anomaly indicates slow cooling of the chromites during crystallization (Locmelis et al., 2011; Pagé and Barnes, 2016) and supports a komatiitic dunite source, possibly at the base of a thick flow or a sill. The relatively coarse chromites observed in the CSCG quartzite would be expected in a komatiitic dunite source, and, naturally, these larger grains would have higher preservation potential during weathering, transport, and deposition.

The Bell and Dwyer chromites have indistinguishable T_{RD} model ages that cluster at ca. 3.2 Ga (Fig. 4A). This is interpreted as the time of chromite formation in the komatiitic melt. Because chromite grains from the different locations are indistinguishable in chemistry and T_{RD} ages, we interpret a komatiitic dunite source of

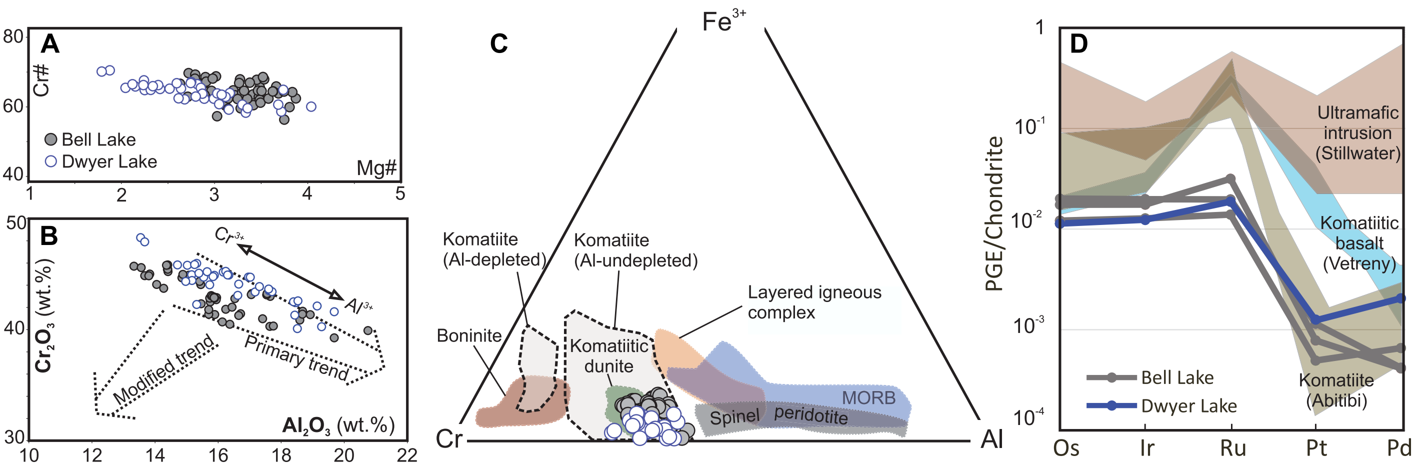

Figure 3. (A) Cr# versus Mg# and (B) Al_2O_3 versus Cr_2O_3 from Bell and Dwyer Lakes (northwest Canada) detrital chromite grains. (C) Ternary Al-Cr-Fe^{3+} classification diagram. Fields of komatiite, komatiitic dunite, boninite, and mid-oceanic ridge basalt (MORB) are from Barnes and Roeder (2001); peridotite is from Dick and Bullen (1984); layered igneous complex data are from Rollinson et al. (2010). (D) Chondrite-normalized platinum group elements (PGEs) compared with chromite from Archean komatiite and ultramafic intrusions. Abitibi (Ontario, Canada) data are from Puchtel et al. (2004); Stillwater (Montana, USA) data are from Pagé et al. (2012); Vetreny (Russia) data are from Puchtel et al. (2016).

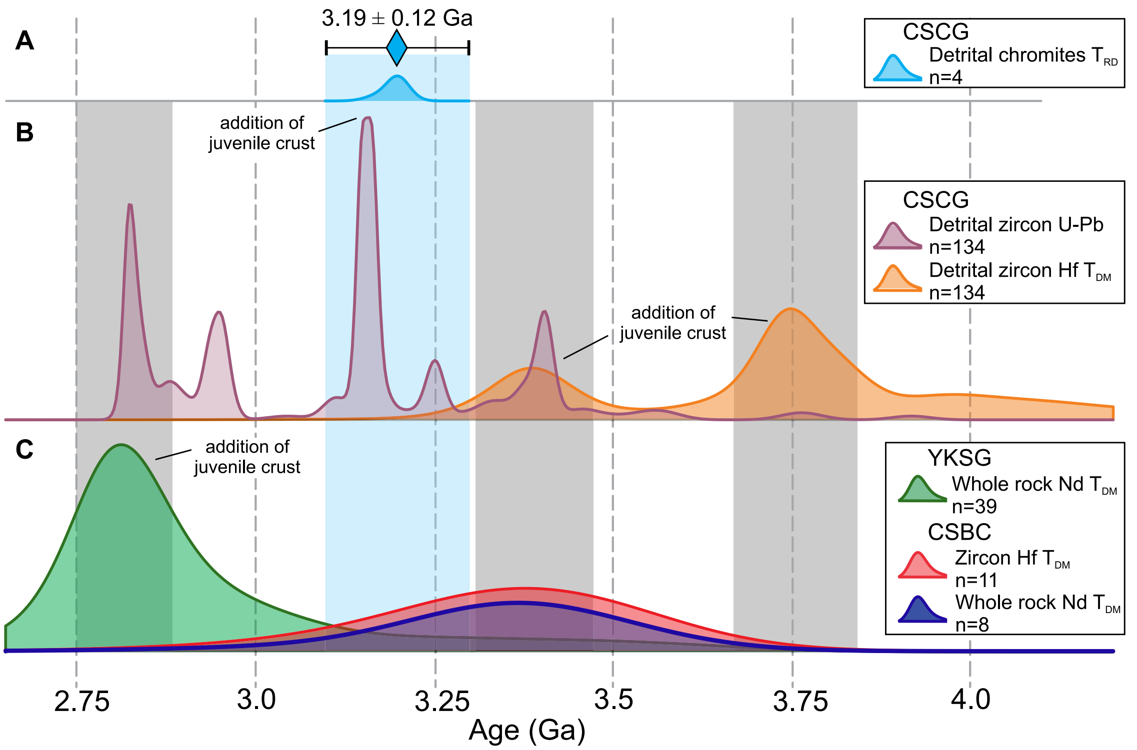

Figure 4. (A) Rhenium depletion model ages (T_{RD}) of four chromite samples from the Central Slave Cover Group (CSCG), northwest Canada. (B) Relative probability curve for detrital zircon U-Pb (Sircombe et al., 2001) and Hf-depleted mantle ages (T_{DM}; Pietranik et al., 2008) from CSCG quartzites. (C) Whole-rock Nd T_{DM} for Yellowknife Supergroup (YKSG) and Central Slave Basement Complex (CSBC) rocks (Northrup et al., 1999; Cousens, 2000; Cousens et al., 2002) and zircon Hf T_{DM} from the CSBC (Reimink et al., 2019). T_{DM} ages represent the addition of juvenile crust (gray vertical bars). This study identifies addition of juvenile crust at ca. 3.2 Ga (blue vertical bar).

relatively uniform composition. The timing of this komatiite magmatism broadly coincides with the most significant mode in detrital zircon ages from the CSBC, a time of felsic-intermediate crustal growth (Fig. 4B). Notably, the addition of ultramafic crust at ca. 3.2 Ga is not represented in detrital zircon Hf-depleted mantle ages (T_{DM}) in the igneous zircon from the CSBC or in the whole-rock Nd T_{DM} ages from the CSBC and Yellowknife Supergroup (Fig. 4C). Therefore, the detrital chromite results provide a record of an otherwise invisible addition of ultramafic magma to the crust.

The coexistence of detrital chromites and variably aged zircons within quartzite demonstrates that a large segment of Slave crust was weathered and eroded prior to 2.85 Ga (Fig. 4; Sircombe et al., 2001). Continental margin breakup, a depositional environment conducive to quartz arenite deposition (Mueller et al., 2005), may have reduced the preservation potential of thermodynamically unstable mineral assemblages (e.g., olivine and pyroxene), hence removing ultramafic rocks from the CSBC. Alternatively, such ultramafic crust could have been rifted away during the Mesoarchean to Neoarchean; regardless, it has disappeared, and detrital chromites are the only remaining evidence of ca. 3.2 Ga komatiitic magmatism in the Slave craton.

NATURE OF THE ARCHEAN LITHOSPHERE

The CSCG detrital chromites are interpreted to have been derived from the weathering of an Al-undepleted komatiitic dunite, age-coincident with a major mode in U-Pb detrital zircon ages (Fig. 4). Aluminum-undepleted komatiites are sourced from shallow mantle depths with high degrees of melt extraction, facilitated by a thin lithospheric "lid" (Arndt et al., 2008). Enhanced heat flow associated with this komatiite event appears to have generated broadly coeval juvenile and evolved magmatism (Fig. 4).

The ca. 3.2 Ga komatiite is younger than suggested peridotitic diamond formation at ca. 3.5 Ga (Westerlund et al., 2006; Helmstaedt, 2009) but older than the Neoarchean amalgamation of the bulk of the craton's lithospheric root, constrained from bedrock evidence (Davis et al., 2003) and peridotite Re-Os systematics (Heaman and Pearson, 2010). Further, the tonalite gneisses of the CSBC crystallized until ca. 2.85 Ga (e.g., Ketchum et al., 2004), and the genesis of such rocks is not consistent with the presence of a ubiquitous, preexisting ≥150 km lithospheric root prior to that time (Davis et al., 2003; Reimink et al., 2019), even if such a root may have been present locally. The ca. 3.2 Ga komatiitic magmatism originated in a region of thinned lithosphere, likely much less than 100 km, either on the protocraton nucleus, or peripheral to the nucleus. This supports lithospheric models proposed for the Yilgarn craton (Mole et al., 2014), for example, and indicates that prior to stabilization of the Slave craton in the Neoarchean, lithospheric architecture on and around the growing protocraton was variable, with much larger gradients in lithospheric thickness than are evident today. Further, the preservation of detrital chromites, along with zircons from the tonalitic crust and coeval portions of the CSBC, clearly documents a key period of juvenile ultramafic and felsic crust addition at ca. 3.2 Ga to what became the building blocks of the Slave craton.

CONCLUSIONS

Detrital chromite grains are preserved within ca. 2.85 Ga fuchsitic quartzite of the CSCG. Grain sizes and major and platinum group element systematics indicate that the chromites were sourced from Al-undepleted komatiitic dunites. The chromites have a narrow range of rhenium depletion model ages with an average age of 3.19 ± 0.12 Ga. The detrital chromites are the only remaining evidence that komatiite existed in, or around, the building blocks of the Slave craton. The komatiites were generated in a region of thinned lithosphere, either at the edge of, or within crustal blocks of the (>3.2 Ga) proto–Slave craton. The chromites identify juvenile addition to the crust that is not preserved in the bedrock exposures or the zircon isotopic data.

In contrast to detrital zircons, detrital chromites provide evidence of the mafic-ultramafic contribution to continental crust building and can serve as an important window into the nature of the lithospheric mantle at their site of genesis. While Re-Os detrital chromite dating is in its infancy, the approach in this study can be applied to other crustal terranes (Hadean/Archean through the Phanerozoic) where detrital chromites are present, potentially identifying mafic-ultramafic crust that is no longer preserved.

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