U-Pb ages (3.8-2.7 Ga) and Nd isotope data from the newly- identified Eoarchean Nuvvuagittuq supracrustal sequence, Superior Craton, Canada

Jean David

Service Géologique de Québec, Ministère des Ressources naturelles de la Faune et des Parcs, 5700 4ième Avenue Ouest, Charlesbourg, QC, Canada G1H 6R1, and GEOTOP-UQAM-McGill – Univsersité du Québec à Montréal, C.P. 8888, Succursale Centre-Ville, Montréal, QC, Canada H3C 3P8. (email: k32460@er.uqam.ca)

Laurent Godin

Department of Geological Sciences and Geological Engineering, Queen's University, Kingston, ON, Canada, K7L 3N6.

Ross Stevenson*

GEOTOP-UQAM-McGill and Département des Sciences de la Terre et de l'Atmosphère - Université du Québec à Montréal, C.P. 8888, Succursale Centre-Ville, Montréal, QC, Canada H3C 3P8,

Jonathan O'Neil and Don Francis

Earth & Planetary Sciences, McGill University, 3450 University St. Montreal, QC, Canada, H3A 2A7

* Corresponding author's email: stevenson.ross@uqam.ca

ABSTRACT

We report new U-Pb ages and Sm-Nd isotope data for an Eoarchean volcano-sedimentary sequence in the northeast Superior Province of northern Quebec, Canada. The Nuvvuagittuq supracrustal sequence is characterized by mafic amphibolite rocks with rare felsic schists, accompanied by ultramafic sills, oxide-rich and quartz-rich iron formations and possible conglomeratic units with peak metamorphism reaching upper amphibolite facies. The sequence outcrops as an open, shallow south-plunging synform with tonalites in the center and around the exterior margins of the fold structure. A 3817 ± 16 Ma age from a felsic schist is interpreted to represent a maximum age for the sequence, while a 3661 ± 4 Ma age from the enclosing tonalites suggests a lower age boundary. Detrital zircons from a conglomeratic unit yielded similar old ages of ca. 3.78 Ga. Younger ages of ca. 3.66 and 3.3 Ga from the conglomeratic unit, the felsic schist and 2.7 Ga ages from an amphibolite gabbro and a pegmatite dyke underline the poly-metamorphic history of the sequence. Initial Nd isotope compositions ($\varepsilon Nd^{3.8}$ = -0.5 to + 2.3) for mafic amphibolites and felsic schists are indicative of a moderately-depleted mantle. Sm-Nd analyses of a 2.7 Ga pegmatite dykes and the ~ 3.66 Ga tonalites both yield Nd depleted mantle model ages (TDM) of ~ 3.9 Ga, indicating that they were derived from a crust of juvenile Eoarchean origin. We propose that fragments of Paleo/Eoarchean terranes that rim the Superior Province represent remnants of a larger terrane(s) that was rifted and dismembered prior to the formation of the 3.0 – 2.7 Ga Superior Craton.

INTRODUCTION

Eoarchean rocks offer a unique window into early Earth evolution processes such as the initial differentiation and subsequent evolution of the Earth's mantle and crustal reservoirs (e.g.; Wilde et al., 2001; Blichert-Toft et al., 1999), as well as possibly yielding the first evidence of the emergence of life (Rosing, 1999; Mojzsis et al. 1996; Schidlowski, 2001; Furnes et al., 2004; Banerjee et al., 2006). The Earth's oldest known crustal rock is the 4.04 Ga Acasta gneiss of the Slave Craton in the Canadian Shield (Bowring and Williams, 1999), although studies of detrital zircons from the Australian Jack Hills locality hint at the formation of a hydrous crust as early as 4.4 Ga (Wilde et al., 2001). The uncertain nature of the protolith of the Acasta gneiss, however, limits the interpretation of geological processes. 3.8-3.7 Ga supracrustal sequences such as those of West Greenland (e.g. Akilia Island and the Isua greenstone belt) and Labrador (Nulliak assemblage) are important because they contain mafic volcanic units of undeniable mantle origin and sedimentary units with preserved primary features (e.g. Collerson and Bridgwater, 1979; Nutman et al., 1984; Appel et al., 1998; Komiya et al., 1999). These sequences provide an important geological framework for the interpretation of geochronological and geochemical data of the Eoarchean Earth.

We report here new ages for the ~ 3.8 Ga supracrustal sequence, the Nuvvuagittuq sequence (David et al., 2002; Cates and Mojzsis, 2007), in the northeastern Superior Province of Canada (Fig. 1). The combined geological, U-Pb geochronology, and Sm-Nd isotope data[1] from the Nuvvuagittuq sequence indicate that this supracrustal sequence represents a juvenile Eoarchean crustal package that was thermally disturbed at ~ 3.66, ~ 3.3 and ~ 2.70 Ga.

[1] GSA Data Repository item NUMBER. A comparison of U-Pb dating techniques used in this paper as well as cathadoluminescence imaging of dated zircons are available online at www.geosociety.org, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

GEOLOGIC SETTING

The Archean Superior Province in northern Quebec is dominated by the largely plutonic Minto block. The Minto block is sub-divided into a series of north-south trending domains containing scattered supracrustal sequences, the majority of which are metamorphosed to the upper amphibolite facies (Fig. 1B; Percival et al., 1992; 1994; Percival and Skulski, 2000). The Nuvvuagittuq volcano-sedimentary sequence is located in the westernmost Inukjuak domain of the Minto block, on the shores of Hudson Bay (Fig. 1).

The Inukjuak domain is composed of ca. 2.8-2.7 Ga tonalite-trondhjemite-granite and monzodiorite-granite suites with less abundant mafic-ultramafic plutonic suites (Simard et al., 2003). The Nuvvuagittuq supracrustal sequence, the largest coherent volcano-sedimentary package in the Inukjuak domain, covers an area of 20 km^2 of nearly-continuous outcrop forming a semi-oval shape against the coastline of Hudson Bay (Fig. 2). The supracrustal sequence consists of tight to isoclinally-folded and highly-deformed banded iron formation and amphibolite units, refolded in an open, shallow south-plunging synform (Fig. 2). The sequence is sheathed by a grey-pink, foliated to mylonitic tonalite that is itself intruded by the surrounding younger heterogeneous tonalite suite of the Inukjuak domain, dated at 2750 ± 6 Ma (Simard et al., 2003). A similar grey-pink mylonitic tonalite and a highly strained amphibolitic gabbro occupy the center of the folded supracrustal assemblage (Fig. 2).

The Nuvvuagittuq sequence consists principally of amphibolites, ultramafic and gabbro sills, and iron formation in addition to minor felsic and possible conglomeratic units. Although the sequence underwent significant metamorphism and tectonic transposition, compositional layers are mappable, and sufficient information can be extracted from units that are less deformed to permit a first-order interpretation of possible protoliths. From west to east, the proportion and character of these units progressively change along with an increase in metamorphic grade. In the southwest, amphibolite mineral assemblages are dominated by actinolite and chlorite, progressing to cummingtonite ± anthophyllite ± hornblende ± garnet in the northeast.

Amphibolites

The Nuvvuagittuq sequence is dominated by the presence of amphibolitic rocks that are highly variable in color (dark green, grey, beige) and aspect (massive to banded with 1-5cm-thick bands) due to mineralogy. Massive dark green amphibolites are found along the southwestern margin of the sequence in association with thin banded iron formation units consisting of dark colored magnetite-rich and quartz-rich jasper-colored bands. These massive amphibolites consist predominantly of recrystallized actinolite, chlorite and epidote with minor plagioclase and quartz and are metamorphosed to upper greenschist facies (Fig. 3A,B). Towards the east, the amphibolitic rocks become noticeably lighter in color and consist of cummingtonite + quartz + biotite + plagioclase ± hornblende ± anthophyllite (Fig 3C). Lighter colored amphibolitic rocks are associated with higher concentrations of cummingtonite; and variable proportions of cummingtonite, biotite and quartz lead to a light-dark banded appearance. The abundance of amphibole suggests the metamorphic grade has increased to amphibolite facies. Both the actinolite-bearing cummingtonite-bearing amphibolites are mafic in composition with 40 to 56% SiO_2 and 4 to 16% MgO with the cummingtonite amphibolites having higher Al2O3 contents (O'Neil et al. 2007).

Further east, the amphibolites have an increasingly greater content of cummingtonite and anthophyllite culminating in a distinctive cummingtonite + tremolite + anthophyllite + Mg-chlorite ± talc schist. Whether the appearance of garnet in the cummingtonite-plagioclase amphibolites following a garnet-in isograd (O'Neil et al., 2007) reflects the change in composition of the amphibolites or an increase in metamorphic grade is still to be determined. East of the garnet-in isograd, the amphibolites take on an increasingly orange-beige color and assemblages of biotite + plagioclase + quartz + garnet become more common (Fig. 3D) with occasional assemblages of biotite + cordierite + garnet + sillimanite. These Al-rich mineral assemblages likely reflect alteration of the amphibolites rather than a primary pelitic composition because these rocks are still mafic in character with compositions (40-56 wt% SiO2, 4-16 wt% MgO) similar to those of the massive amphibolites at lower grade (O'Neil et al., 2007).

Ultramafic Sills

On the western limb of the synform, a series of dark brown weathering sills with an ultramafic composition (31% MgO and 2.6% Al_2O_3; O'Neil et al., 2007) are found interlayered among the amphibolites. The sills are boudinaged and can be followed along strike for hundreds of meters as they thin and swell from 1 to 30 m-thick (Fig. 3E). The sills consist of dunite and pyroxenite layers, some with gabbroic margins, containing relic forms of olivine and orthopyroxene but now largely replaced by serpentine + talc + tremolite ± hornblende ± chlorite. The margins of these horizons are commonly composed of fine grained amphibolite that might reflect a chilled margin, thus suggesting that the ultramafic horizons were initially dunite/harzburgite sills. O'Neil et al. (2007) showed that the compositions of the sills fall into two groups. Type-1 sills are Fe-rich and Al-poor and are grouped to the west of Type-2 sills, which are Fe-poor and Al-rich by comparison.

Gabbro/Amphibolitic Gabbro Sills

A number of gabbro bodies are found in the central and eastern portion of the Nuvvuagittuq sequence. These gabbros consist of coarse to medium grained hornblende + plagioclase + quartz ± orthopyroxene ± cummingtonite and have a black and white banded appearance brought on by alternating plagioclase-rich and hornblende-rich bands. They are distinctly coarser and darker than the cummingtonite-amphibolites and are thus mapped as a separate lithology. More homogeneous gabbros are found in association with the mylonitized tonalities found in the synform core of the supracrustal assemblage.

Iron Formation

The character of the iron formations in the Nuvvuagittuq sequence changes significantly from west to east. A thin (< 1 m-thick) banded iron formation with jasper-like silica bands and magnetite rich oxide bands is found on the extreme southwest margin of the exposed sequence. Further east, a 5 to 30 m-thick unit of banded iron formation associated with the ultramafic sills can also be traced along the western limb of the assemblage. The unit is rusty-brown on the weathered surface (Fig 3F) and varies from finely laminated (0.1-1 cm) to coarsely laminated (1-5 cm). Like the ultramafic sills, the iron formation thins and swells in boudins and can be followed for hundreds of meters along the strike of the western limb. The rock consists of alternating light bands of quartz + magnetite with darker bands of magnetite (50%), amphibole (grunerite, hornblende, actinolite; 40%) ± orthopyroxene ± garnet. Garnet is occasionally found in quartz-rich portions of the iron formation.

On the eastern synform limb of the Nuvvuagittuq sequence, banded iron formations are progressively coarser grained and dominated by the silicate-rich horizons to the point that the rock is essentially pure quartz. The quartz-rich content gives the unit a distinct quartzitic/chert aspect; however, no primary sedimentary features remain. Because these units are either spacially associated with banded iron formation or are accompanied by thin iron-oxide rich lamina, they are considered to be a silica-rich facies of the iron formation (Fig. 3G). These silica-rich rocks are beige to white and weather to a light brown or white in color and the units vary from 0.5 to 200 m in width. These units consist of 80-95% quartz with thin schistose layers of magnetite, biotite, chlorite, epidote and pyrite. Occasional thicker iron-oxide rich bands are found (1-2 m-thick) and comprise coarse-grained magnetite (50%), amphibole (grunerite and hornblende; 40%), plagioclase and quartz.

Conglomeratic Unit

In the southwestern part of the Nuvvuagittuq sequence, rare, thin conglomeratic and felsic schist units (50-100 cm-thick) are found interlayered with the amphibolites and iron formation. The conglomeratic units have a fine grained matrix of quartz, biotite, chlorite +/- garnet that support sub-rounded and strained polylithic clasts (5-20 cm) composed for the most part of quartz, amphibolite and felsic schist (Fig 3H, I). Cates and Mojzsis (2007) also report that clinopyroxene, garnet and sulfides may be present in the matrix. However, given the high strain associated with the unit and the erroneous identification of conglomeratic units in similar high strain zones in the ca. 3.8 Ga Isua greenstone belt (e.g. Appel et al. 1998; Fedo and Moorbath 2005), a definitive identification of conglomeratic units at Nuvvuagittuq requires further study. For the time being the reader should keep in mind that the unit is regarded only as possibly of conglomeratic origin.

Felsic Schist

Rare felsic schists composed of plagioclase, biotite and quartz (63-70% SiO_2) are found in the southwestern limb of the synform (Fig. 3J, K). These units are light grey in color, up to 0.5 m-thick, and interlayered and isoclinally folded with amphibolite, conglomeratic units and iron formation. In contrast with the interpretations of Cates and Mojszis (2007), we have not found any evidence such as clear cross-cutting relationship that would support an intrusive nature. On the contrary, this unit displays remarkable concordance with the surrounding units.

Based on the concordant nature of the felsic schist, the progressive change of rock types from the external parts of the synform towards the internal part of the western limb is interpreted to reflect progressive change from a mafic volcanic-dominated environment to a volcano-sedimentary sequence comprised of mafic amphibolite rocks with associated plagioclase-quartz-biotite felsic schists that are overlain by, or interlayered with sedimentary rocks that include a possible polylithic quartz-rich conglomerate and oxide-rich and silicate-rich banded iron formation.

Tonalitic Units

The margin of the Nuvvuagittuq sequence consists of a strongly foliated to mylonitized tonalite (Fig. 3L) to granodiorite (57-73 wt% SiO_2) that is itself intruded by a Neoarchean heterogeneous tonalite Boizard suite (Simard et al., 2003). The core of the synform is also composed of highly-strained tonalite and amphibolitic gabbro containing fine grained amphibolitic enclaves. This tonalite and gabbro core is characterized by a well-defined, almost horizontal, south-trending mineral elongation lineation and a very weakly-developed, east-striking, sub-horizontal foliation, suggesting a fault contact with the structurally underlying Nuvvuagittuq sequence. The supracrustal sequence and its core are crosscut by a series of late meter-thick pegmatite dykes (quartz + K-feldspar + muscovite).

General Structural Setting

The supracrustal sequence, including the felsic unit, is affected by isoclinal folding and transposition (D_1) that is locally overprinted by shearing, probably associated with the emplacement of the highly strained internal and border tonalites (D_2). These two episodes are themselves overprinted by the open synform (D_3), which folds the Nuvvuagittuq sequence and the ca. 2.8 Ga Inukjuak Domain tonalites (Nadeau, 2003).

The age of the highly strained internal and border tonalites should therefore yield a minimum age for the sequence because they are not affected by the older isoclinal folding. Similarly, the dating of the felsic volcanic unit should constrain the age of the sequence. Although most units have undergone variable degrees of transposition, the lack of a high-strain zone or a strain gradient associated with the felsic volcanic unit and its contacts argues against tectonic interleaving.

METHODOLOGY

Five rock types yielded sufficient zircon or monazite for U-Pb geochronology to constrain ages of crystallization and metamorphism of the sequence. Samples for U-Pb geochronology were crushed and heavy minerals were separated on a Wilfly table. Further separation of the heavy minerals was done using heavy liquids. At this point a representative fraction of the zircons present were picked and mounted in epoxy for laser ablation analysis. For U-Pb isotope dilution analyses, the remaining zircons were separated into magnetic and non-magnetic fractions using a Franz magnetic separator. Selected individual grains were abraded to minimize the effects of Pb loss (Krogh, 1982) or to remove metamorphic overgrowths. The selected grains were weighed, mixed with a ^{205}Pb-^{233}U-^{235}U spike and subsequently dissolved in an HF-HNO3 solution. Uranium and lead were separated using an anion exchange resin and the resulting uranium and lead fraction were loaded on single Re filaments and analyzed using a solid source VG Sector mass spectrometer with a Daly detector. The isotopic ratios are corrected for fractionation, spike, blank and initial common lead, and regressions were calculated using IsoplotEX (Ludwig, 2003).

Zircons for laser ablation U-Pb dating were analyzed with an Eximer laser coupled to a GV ISOPROBE ICP-MS multicollector at the GEOTOP-UQAM-McGill laboratories. The laser is a 193 nm Lambda Physik Compex 102-ArF that delivers a maximum energy of 200 mJ per pulse for a period of 25 ns. The analyses were performed following a protocol modified after Valariano et al. (2004). The laser was used in "raster" mode over a distance of 80 μm with a beam diameter of 35 μm and a frequency of 8 Hz. It was found that the sensitivity and stability of the Pb signal could be increased by mixing the gas from the reaction cell with a blank solution from the CETAC Aridus desolvating nebulizer system via a Y junction before entering the torch box. The Pb isotope compositions were corrected for a daily instrument drift in mass discrimination and elemental fractionation using the ‘standard bracketing’ method with an in-house standard dated at 2761±1 Ma (n=25) by TIMS-ID. Thus, under constant analytical conditions the instrument drift could be corrected according to the power law and as a function of time. Error propagation was calculated according to Horstwood et al. (2003) using the external reproducibility of the standard calculated for each analytical session yielding variations of 0.08% to 0.17% (1σ) for the $^{207}Pb/^{206}Pb$ ratios and 0.6% to 1.5% (1σ) for the $^{206}Pb/^{238}U$ ratios. An illustration of the reproducibility of the U-Pb LA-MC-ICP-MS and TIMS analyses is shown in Figure A1.

Rock samples for Sm-Nd analyses were crushed to powder form and Sm and Nd were separated at GEOTOP using the protocol of Henry et al. (1998) and measured using a multicollector VG54 mass spectrometer with the addition of a ^{150}Nd-^{149}Sm spike. Repeated measurements of the LaJolla Nd standard yielded a value of $^{143}Nd/^{144}Nd$= 0.511849 ± 12 (n=21). The Sm and Nd concentrations and the $^{147}Sm/^{144}Nd$ ratios have an accuracy of 0.5% that corresponds to an average error in the initial Nd isotopic composition of 0.5 εNd units. Total Blank concentrations were <150 pg for Sm and Nd combined. Model ages (TDM) are calculated assuming a linear depleted mantle with a present day value of εNd = +10 and $^{147}Sm/^{144}Nd$ = 0.2136 (Jacobsen, 1988).

U-Pb GEOCHRONOLOGY AND Nd ISOTOPES

Data for the U-Pb TIMS and laser ablation multi-collector inductively coupled mass spectrometry (LA-MC-ICP-MS) analyses are given in Tables 1 and 2, respectively.

Plagioclase-Quartz-Biotite Schist (POR1091f)

A plagioclase-quartz-biotite schist (POR1091f) was collected from Assemblage 2 on the southwestern limb of the synform (Fig. 3J, K). The concordant unit is light grey in color, up to 0.5 m-thick, and is interlayered and isoclinally folded with amphibolitic schist and the conglomeratic unit.

Other examples of this schist occur in the immediate area and are all interlayered with the mafic amphibolite and/or the conglomeratic unit. Zircons extracted from this unit form a homogeneous population of small prismatic grains (50-100 μm) and pyramidal terminations. Cathodoluminescence imaging (Fig. A2A) reveals that the internal structures of these zircons are dominated by primary oscillatory zoning patterns with smaller secondary structures (<20 μm) that are visible as highly luminescent zones with isometric textures or banded rims (Vavra et al., 1999). Fifteen single zircons were analyzed by the thermal ionization mass spectrometry and isotope dilution method (TIMS-ID). All zircons from this sample show evidence of recrystallization, overgrowths, inclusions and microfractures. Zircons 1-6 are the least recrystallized, least fractured and contain the least number of inclusions, whereas zircons 7-15 are of lesser quality. The zircons yield discordant (2.0-8.9%) $^{207}Pb/^{206}Pb$ ages ranging from 3572 Ma to 3752 Ma and Th/U ratios (0.5-0.2) that decrease with decreasing $^{207}Pb/^{206}Pb$ ages (Fig. 4A). Regression of zircons 1-6 (David et al., 2002) yields an upper intercept (crystallization) age of 3817 ± 16 Ma (MSWD = 1.7) and a lower intercept of 2450 ± 80 Ma, respectively (Fig. 4A). Regression of all 15 fractions yields a lower age of 3779 ± 21 Ma with a higher degree of scatter (MSWD 19). Eight individual zircons were also analyzed for U-Pb geochronology by LA-MC-ICP-MS and yielded a range of discordancy that overlaps the TIMS analyses (Fig. 4A).

The predominance of primary oscillatory zoning patterns in these zircons indicate magmatic crystallization with the secondary structures likely reflecting either solid-state recrystallization or overgrowths produced during high-grade metamorphism (Vavra et al., 1999; Hoskin and Black 2000). The extensive discordancy among the zircons is likely a product of both ancient and recent Pb-loss caused by at least two metamorphic events at ca. 3.66 Ga and 2.70 Ga, respectively (see below). Given the magmatic nature of the zircons, the 3817 ± 16 Ma age is considered to be the age of crystallization of the protolith of the schist (a felsic volcanic unit) and the 3779 ± 21 Ma age reflects dispersion of the data by younger events.

Plagioclase-Quartz-Biotite Schist (POR134)

A second sample of plagioclase-quartz-biotite schist was collected from within an amphibolitic band on the western flank of the supracrustal sequence. The zircons from this sample are brownish sub-euhedral stubby prisms with abundant inclusions. Cathodoluminescence images of these zircons (Fig. A2B) shows that they are poorly luminescent with unzoned cores or large diffuse bands and lack the complex magmatic oscillatory zoning (Vavra et al., 1999) found in the preceding sample. Some grains are characterized by concentric secondary overgrowths. Seven individual zircons were analysed by the isotope dilution technique (TIMS). The zircons were found to be U-poor (< 100 ppm) resulting in high Th/U ratios of 0.88-1.1. Four analyses yielded $^{207}Pb/^{206}Pb$ ages between 3365.9 and 3360.7 Ma, which upon regression yielded an upper intercept age of 3366.3±2.5 Ma (Fig. 4B). Three analyses produced slightly younger $^{207}Pb/^{206}Pb$ ages that lie along a discordia between 3365 and 2700 Ma suggesting metamorphism at 2700 Ma that may have produced the secondary overgrowths on some of the zircons from this sample. In spite of the high Th/U ratios (1) of these zircons, the low U concentrations and quasi-absence of magmatic crystallization structures suggest that the zircons of this lithology crystallized from fluids associated with a Mesoarchean metamorphic event.

Conglomeratic Unit (POR29)

Thirty individual zircons from the conglomeratic unit in Assemblage 2 were analyzed by LA-MC-ICP-MS to investigate possible ages of sedimentary provenance (Sample POR29). Most of the zircons are highly fractured and fall into two groups: 1) abundant small, brown, sub-rounded prism fragments with primary internal oscillatory and sector zoning that likely reflect magmatic crystallization, and 2) less abundant colorless to brownish equant xenomorphic grains that are either structureless or have transgressive recrystallization patterns (Fig. A2C). The first group yielded $^{207}Pb/^{206}Pb$ ages from 3770 Ma to 3258 Ma whilst being increasingly discordant and an age of 3787 ± 25 Ma upon regression (MSWD=2.9, Fig. 4C, open symbols). The second group yielded ages from 3661 Ma to 2739 Ma (dark symbols, Fig. 4C) with the least discordant ages clustering at 3649 ± 10 Ma (MSWD=3.6; light-grey symbols) and the most discordant ages lying on or close to discordia lines stretching between 2.7 Ga and 3.65 or 3.79 Ga.

Although both zircon fractions appear to be affected by subsequent thermal (metamorphic or magmatic) events and by recent Pb-loss, the > ca. 3.76 Ga detrital zircon fragments from the conglomerate with primary crystallization structures are interpreted to be derived from the erosion of igneous units within the supracrustal sequence. The younger zircon populations (2.70-3.65 Ga) show strong evidence of recrystallization or metamorphic growth that likely formed during different metamorphic events including a ca 3.2-3.4 Ga event also found in the second felsic schist sample (POR134).

Mylonitized Tonalite (POR23)

A sample of the mylonitized tonalite (POR23) was collected from the western flank of the Nuvvuagittuq sequence (Fig. 2). Zircons from this unit are light brown, stubby, well-faceted prisms and characterized by primary oscillatory and sector zoning with occasional resorption textures that are indicative of a magmatic origin (Fig A2D). U-Pb isotope dilution analyses of 4 single zircons yielded a crystallization age of 3659 ± 3 Ma (MSWD=1.6) and LA-MC-ICP-MS analyses of thirteen zircons yielded an identical age (within error) of 3661 ± 4 Ma (Fig. 4D). Rare cores were observed in the zircons but did not yield any difference in ages. The magmatic structure of the zircons and high Th/U ratios (0.85) suggest that the above ages are crystallization ages. The age of the tonalite is therefore contemporaneous with the ca. 3.6 Ga metamorphic event identified in the conglomeratic sample.

Amphibolitic Gabbro (POR30) / Pegmatite (POR1091G)

Small equant zircons were recovered from an amphibolitic gabbro (POR30) associated with the tonalite in the center of the synform structure. The zircons exhibit a mottled internal structure and radial-sector zoning (Fig. A2E) that is characteristic of zircons crystallized under upper amphibolite to granulite metamorphic facies conditions (Vavra et al., 1999). Thus the 2693 ± 3 Ma (Fig. 4E) age determined from two single grain U-Pb TIMS analyses is interpreted to reflect a metamorphic age.

The age of a cross-cutting pegmatite dyke (POR1091G) was determined by U-Pb analysis of four single monazite grains that yielded an age of 2686 ± 4 Ma upon regression (MSWD=1.9**,** Fig. 4E) and is interpreted to represent its emplacement age .

Sm-Nd Isotopes

Sm-Nd isotope data and methodology are reported in Table 3. Analyses of the plagioclase-quartz-biotite schist, the mylonitic tonalites and the pegmatites all yield depleted mantle model ages (T_{DM}) of ca 3.9 Ga, indicating that they were derived from a similar (juvenile) Eoarchean source (Fig. 5). This similarity in model ages indicates that the pegmatite dykes (εNd^i = -9) and tonalites (εNd^i = -0.6 to -1.8) formed by the anatexis of an ancient crust, without incorporation of younger Archean material. Initial εNd values of the felsic schist and associated amphibolites yield values ranging from -0.5 to + 2.3 with T_{DM} ages of 3.8-3.9 Ga. These isotopic compositions are consistent with a depleted mantle origin and exceed those predicted by progressive depletion models for the mantle (DePaolo, 1981; Jacobsen, 1988). However, our results indicate a less depleted source than the range of values obtained from southern West Greenland (e.g., Bennett et al., 1993; Blichert-Toft et al., 1999), or Labrador (Collerson et al., 1991).

DISCUSSION

The geochronology and interpretation of Eoarchean rocks is fraught with controversy because of the complexity of the zircon compositions and field relations imposed by multiple and often severe deformational and metamorphic events that obscure both the age of emplacement and the nature of the protolith. For example, ca. 3.8 Ga cores in zircons in tonalitic orthogneisses in southwest Greenland have been interpreted to reflect either the emplacement age or older zircon inherited by a younger (ca 3.6 Ga) event (e.g. Whitehouse et al., 1999; Nutman et al,. 1996; Mojzsis et al., 1996).

The 3817 ± 16 Ma age of the quartz-plagioclase-biotite schist represents the oldest age ever determined for a rock from the Superior Province. Cates and Mojzsis (2007) collected a rock at the same locality as our sample and obtained a minimum age of ca. 3750 Ma using ion microprobe U-Pb analyses of zircons. Whether these ages reflect the age of the supracrustal sequence depends on its origin and relationship with the associated amphibolites. Repeated field investigations of the plagioclase-quartz-biotite schists have failed to find evidence of cross-cutting relationships with neighboring units or specific strain zones that would indicate that the schists were tectonically interleaved into the sequence. The felsic schist with its confined stratigraphic position, lack of cross-cutting relationships with surrounding units, magmatic character of the ca. 3.8 Ga zircons, and its association with amphibolite units could indicate a volcanic origin which would also imply that the Nuvvuagittuq supracrustal sequence represents a ca 3.8 Ga volcano-sedimentary assemblage. However, given the high amount of strain associated with these rocks, we cannot completely discard the scenario that the schist may have been tectonically transposed within a possibly younger (< 3.8 Ga) supracrustal assemblage.

The second felsic schist sample (POR134) did indeed yield a younger age (3366.4 ± 2.3 Ma), however the zircons of this sample lacked the magmatic zoning characteristics found in the first sample

(Fig A2A, B) and have a recrystallized appearance. Thus we interpret this younger age as a metamorphic age. This younger zircon population was not identified in POR1091f and there appears to be no ca. 3800 Ma zircons in the second sample, suggesting that the second sample (POR134) was metamorphosed at ca 3366 Ma whereas sample POR1091f largely escaped the effects of this metamorphic event. This may be simply because the composition of sample POR1091f is less conducive to secondary zircon growth and/or because the flanks of the supracrustal assemblage were more severely metamorphosed compared to the southern fold hinge where sample POR1091f was collected.

The majority of the zircons from the conglomeratic sample yield two age groups. Zircons of the first group are the most abundant and have a sub-rounded detrital character with complex zoning patterns and yield a ca. 3.8 Ga age. Zircons of the second group are extensively recrystallized (featureless structures; Fig A2C) and yield younger ages that are consistent with younger metamorphic events. The most prominent group of younger zircons yields an age of ca 3660 Ma which overlaps with the age of the tonalite suite. A few zircons yielded discordant Pb-Pb ages of ca 3.2-3.3 Ga and likely reflect metamorphic ages similar to those found in the second felsic schist (POR134). Although more study of the conglomeratic unit is warranted to confirm its clastic origin, these ages are consistent with a supracrustal assemblage >3660 Ma and perhaps as old as 3800 Ma that was metamorphosed at ca 3.66, 3.36 and 2.70 Ga. In a similar unit, Cates and Mojzsis (2007) found one zircon >3700 Ma of possible detrital character and a number of other younger zircons.

The Nuvvuagittuq tonalites were emplaced at ca 3.66 Ga with negative initial εNd values (-0.6 to -1.8) that are consistent with an origin from recycled ca 3.8 Ga crust (Fig. 5). This suggests that the tonalites were formed during metamorphism and this ca 3.66 Ga metamorphic event may also have promoted zircon growth within the conglomeratic sample of the supracrustal assemblage. An even younger metamorphic event is recorded by the age of the amphibolitic gabbro and pegmatite at 2.7 Ga and here again, the initial εNd values for the pegmatite are consistent with an origin via anatexis of 3.8 Ga crust. Thus the Nuvvuagittuq supracrustal assemblage is characterized by at least four discrete events; crystallization of the plagioclase-biotite-quartz schist at ca 3.8 Ga followed by metamorphism and anatexis at 3.66 Ga to form the Nuvvuagittuq tonalite suite, metamorphism and zircon growth in the felsic schists at ca 3.37 Ga and a final metamorphic event and anatexis at 2.7 Ga to form the late pegmatites. These youngest thermal event ages (ca. 2.7 Ga) are consistent with the age of regional Neoarchean metamorphism (2.70-2.68 Ga) that characterizes a large part of the Minto Block (Percival and Skulski, 2000; Simard et al., 2003). This Neoarchean metamorphic event likely resulted in zircon growth in the amphibolitic gabbros and formation of pegmatitic dykes.

Regional Implications

Supracrustal units similar to the rock types of the Nuvvuagittuq sequence are found as enclaves incorporated in the Neoarchean granitoid suites of the Inukjuak Domain (Simard et al., 2003). Regional Sm-Nd isotopic studies of these Neoarchean granitoids yield model ages of 3.9 to 3.2 Ga, indicating that the Eoarchean rocks were more extensive, but were dismembered or totally recycled by younger tectonic and magmatic events (Stevenson et al., 2006). However, at ca 3.8 Ga, the plagioclase-biotite-quartz schist is 300 Ma older than the next oldest unit in the Superior Province (e.g. 3.5 Ga Minnesota River Valley gneisses; Bickford et al., 2004) and, as such, provokes questions about its affinity with the province. Because of its 3817 ± 16 Ma age, the quartz-plagioclase-biotite schist evokes comparison with the geological history of the Itsaq Gneiss Complex of West Greenland and the Nain Province of Labrador (#6-7, Fig. 6). Geological similarities between these regions include (compiled from Nutman et al., 1984; 2001; 2004; Whitehouse and Kamber, 2005): 1) Magmatic events at 3.9-3.8 Ga; 2) lithologies such as mafic volcanic rocks, banded iron formation and possible clastic sediments; 3)

Crustal recycling at ca. 3600 Ma; and 4) hydrothermal alteration of mafic units. The presence of mafic schists rich in amphiboles such as cummingtonite and anthophyllite is similar to altered schists described in the hydrothermally altered "garbenscheifer" unit at Isua (Nutman et al., 1984; Rosing et al., 1996). These are but broad similarities, but they suggest that either Eoarchean tectonic evolution was governed by a series of global coeval events or that these three regions may have once been part of a larger composite Eoarchean terrain.

Nutman et al. (2001) suggest that the ca. 300 Myr history of the Itsaq Gneiss Complex reflects elements preserved from a once larger Eoarchean terrane that included Eoarchean elements of the Nain Province in Labrador. The similarity in age and geological history between the Itsaq Gneiss Complex, Nain Province, and the Nuvvuagittuq sequence raises the possibility that they were once constituents of that larger Eoarchean terrane, which was dismembered by successive Meso- and Neoarchean events. For example, the predominantly Neoarchean Superior Craton contains fragments of terranes of Eoarchean heritage, preserved mostly along its margins (Fig. 6). In the northeastern Superior Craton, the Nuvvuagittuq sequence lies on the western margin and Wardle et al. (2002) have shown that the Nain Province likely once formed the craton's eastern margin before it was rifted from the Superior Craton at ca 2.0-2.1 Ga. In the western Superior Craton, the northern margin of the Superior Craton consists of Neoarchean rocks of the Northern Superior superterrane with an Eoarchean heritage, as indicated by inherited zircons, detrital zircons, and Sm-Nd T_{DM} model ages that range from 3.7- 3.3 Ga (#2-3, Fig. 6; Böhm et al., 2003; Skulski et al., 2000). Finally, the southern margin of the Superior Craton contains the ca. 3.5 Ga Minnesota River Valley gneisses (#5, Fig. 6; Bickford et al., 2004). We suggest that these Paleoarchean to Eoarchean terranes/margins of the Superior Craton represent remnants of one or several larger Paleoarchean/Eoarchean terrane(s) that were rifted apart and reassembled by later orogenesis. This is consistent with the view that the Superior Craton may represent a collage of Neoarchean terranes that formed during the Superior Orogeny, after a 3.0 Ga rifting event (Percival et al, 1994). Supporting evidence for a Superior Orogen is perhaps best preserved in the Western Superior Province where evidence for rifting is found in the form of early ca 3.0 Ga platformal assemblages with overlying rift related komatiite-basalt associations (e.g. Thurston, 2002 and references therein) and ca 3.0 TTG suites with isotopic evidence for recycled older ca 3.4 Ga crust (Tomlinson et al., 1999; Henry et al., 2000). In the Minto Block of the northeastern Superior Province, Nd isotope studies have demonstrated that the central terrains are juvenile (Goudalie, Lake Minto) while the eastern (Inukjuak, Tikkerutuk) and western (Douglas Harbour) terranes contain recycled pre-3.0 Ga crustal components (Stern et al. 1994; Rabeau, 2001; Stevenson et al. 2006).

The age of the Nuvvuagittuq sequence demonstrates that the Superior Craton contains an important Eoarchean component and Nd isotope studies of these rocks and Eoarchean rocks of Labrador (Collerson et al., 1991), the Itsaq Gneiss complex (e.g. Bennett et al., 1993; Blichert-Toft et al., 1999) all yield isotopic compositions indicative of a depleted mantle origin. The juvenile nature of these rocks and lack of older (pre-3.8 Ga) detrital zircons from these areas (e.g. Nutman et al., 2004; Davis, 1996) further suggest there was a lack of older continental crust to inherit. This implies that the ca. 4.0 Ga Acasta gneiss (#8, Fig 6; Bowring and Williams, 1999) and the ca. 4.4 Ga detrital zircons from Jack Hills (Wilde et al., 2001) and 3.6-3.96 Ga detrital zircons from the Beartooth Mountains of the Wyoming Craton (#1, Fig. 6; Mueller et al., 1992; 1998) formed independently from the Paleoarchean/Eoarchean terrain identified above.

CONCLUSIONS

New ages from the Nuvvuagittuq supracrustal assemblage in northern Quebec indicate a protracted geological history spanning a billion years from ca 3.8-2.7 Ga. The ca 3.8 Ga ages determined from the Nuvvuagittuq supracrustal assemblage (this study, Cates and Mojzsis, 2007; David et al., 2002) are the oldest in the Superior craton and are at least 300 Ma older than the previous oldest units such as the ca 3.5 Ga Minnesota River Valley gneisses (Bickford et al., 2004). A younger 3663 Ma age for the Nuvvuagittuq tonalite, a ca. 3.37 Ga age from a felsic schist and ca. 2.7 Ga ages from a pegmatite and an amphibolitic gabbro from the sequence are interpreted to reflect metamorphic events that affected the supracrustal sequence. This is corroborated by Nd isotope compositions for the tonalite and pegmatite that are consistent with an origin by melting of ca. 3.8 Ga crust. Nd isotope data from amphibolitic units of the supracrustal sequence are indicative of derivation from a depleted mantle. Vestiges of Paleo-Eoarchean terrains can be found on both the northern/western and southern/eastern margin of the Superior craton and may be remnants of a Paleo-Eoarchean continental platform(s) that were rifted apart during the formation of the Superior Craton.

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REFERENCES CITED

Appel, P.W.U., Fedo, C.M., Moorbath, S., and Myers, J.S., 1998, Recognizable primary volcanic and sedimentary features in a low-strain domain of the highly deformed, oldest known (3.7-3.8 Ga) Greenstone Belt, Isua, West Greenland: Terra Nova, v. 10, p. 57-62.

Banerjee, N.R., Furnes, H., Muehlenbachs, K., Staudigel, H., and de Wit, M 2006, Preservation of approximately 3.4-3.5 Ga microbial biomarkers in pillow lavas and hyaloclastites from the Barberton greenstone belt, South Africa: Earth and Planetary Science Letters, v. 241, p.707-722.

Bennett, V.C., Nutman, A.P. and McCulloch, M.T., 1993, Nd isotopic evidence for transient highly depleted mantle reservoirs in the early history of the Earth: Earth and Planetary Science Letters, v. 119, p. 299-317.

Bickford, M.E., Wooden, J.L. and Bauer, R.L., 2004, New SHRIMP U-Pb zircon ages for the Paleoarchean to Mesoarchean rocks of the Minnesota River Valley: Geological Society of America Abstracts with Programs, v. 36, p. 458.

Blichert-Toft, J., Albarède, F., Rosing, M., Frei, R., and Bridgwater, D., 1999, The Nd and Hf isotopic evolution of the mantle through the Archean: Results from the Isua supracrustals, West Greenland, and from the Birimian terranes of West Africa: Geochimica Cosmochimica Acta, v. 63, p. 3901-3914.

Böhm, C.O., Heaman, L.M., Stern, R.A., Corkery, M.T., and Creaser, R.A., 2003, Nature of the Assean lake ancient crust, Manitoba: a combined SHRIMP-ID-TIMS U-Pb geochronology and Sm-Nd isotope study: Precambrian Research, v. 126, p. 55-94.

Bowring, S.A. and Williams, I.A., 1999, Priscoan (4.00-4.03 Ga) orthogneisses from northwestern Canada: Contributions to Mineralogy and Petrology, v. 134, p. 3-16.

Cates, N., and Mojzsis, S.J., 2007, Pre-3750 Ma supracrustal rocks from the Nuvvuagittuq supracrustal belt, northern Québec. Earth and Planetary Science Letters v. 255, p. 9-21.

Collerson, K.D. and Bridgwater, D., 1979, Metamorphic development of early Archean tonalitic and Tronhjemitic gneisses: Saglek area, Labrador, in Barker, F., ed., Trondhjemites, dacites and related rocks: Amsterdam, Elsevier, p. 205-273.

Collerson, K.D., Campbell, L.M., Weaver, B.L. and Palacz, Z.A., 1991, Evidence for extreme mantle fractionation in early Archaean ultramafic rocks from northern Labrador: Nature v. 349, p. 209-214.

David, J., Parent, M, Stevenson R., Nadeau, P., and Godin, L., 2002, The Porpoise Cove supracrustal sequence, Inukjuak area: A unique example of Paleoarchean crust (ca. 3.8 Ga) in the Superior Province: Ministère des Ressources Naturelles Québec, DV 2002-10.

Davis, D.W., 1996, Provenance and depositional age constraints on sedimentation in the western Superior transect area from U–Pb ages of zircons: Lithoprobe Rep. 53, 18–23.

DePaolo, D.J., 1981, Neodynium isotopes in the Colorado Front Range and implications for crust formation and mantle evolution in the Proterozoic: Nature, v. 291, p. 193-197

Fedo, C.M. and Moorbath, S., 2005, Conglomerate or deformed and metasomatized tonalite? Geochemical evidence on the origin of the “boulder bed”, Isua greenstone belt, Greenland: Geological Society of America Abstracts with Programs, Vol. 37, No. 7, p. 386.

Furnes, H., Banerjee, N.R., Muehlenbachs, K., Staudigel, H., and de Wit, M., 2004, Early life recorded in Archean pillow lavas: Science, v. 304, p. 578-581.

Henry, P., Stevenson, R.K. and Gariépy, C., 1998. Late Archean mantle composition and crustal growth in the Western Superior Province of Canada: Neodymium and lead isotopic evidence from the Wawa, Quetico and Wabigoon subprovinces. Geochim. Cosmochim. Acta 62, 143-157.

Henry, P., Stevenson, R., Larbi, Y., and Gariépy, C., 2000, Nd isotopic evidence for Early to Late Archean (3.4-2.7 Ga) crustal growth in the Western Superior Province (Ontario, Canada): Tectonophysics, v. 322, p. 135-151.

Hoffman, P.F., 1988, United States of America, the birth of a craton: early Proterozoic assembly and growth of Laurentia: Annual Review of Earth and Planetary Sciences, v. 16, p. 543–603.

Horstwood, M.S.A., Foster, G.L., Parrish, R.R., Noble, S.R., and Nowell, G.M., 2003, Common-Pb corrected in situ U–Pb accessory mineral geochronology by LA-MC-ICP-MS: Journal of Analytical Atomic Spectrometry, v.18, p.837-846.

Hoskin, P.W.O. and Black, L.P., 2000, Metamorphic zircon formation by solid-state recrystallization of protolith igneous zircon: Journal of Metamorphic Geology, v. 18, p. 423–439.

Jacobsen, S.B., 1988, Isotopic constraints on crustal growth and recycling: Earth and Planetary Science Letters, v. 90, p. 315-329.

Komiya, T., Maruyama, S., Masuda, T., Nohda, S., Hayashi, M., Okamoto, K., 1999, Plate tectonics at 3.8-3.7 Ga; field evidence from the Isua accretionary complex, southern West Greenland: Journal of Geology, v. 107, p. 515-554.

Krogh, T.E., 1982, Improved accuracy of U-Pb zircon ages by the creation of more concordant systems using air abrasion technique: Geochimica Cosmochimica Acta, v. 46, p. 637-649.

Ludwig, K.R., 2003, Isoplot/Ex v.3, a geochronological tool kit for Microsoft Excel: Berkeley Geochronology Centre Special Publication no. 4.

Mojzsis, S.J., Arrhenius, G., McKeegan, K.D., Harrison, T.M., Nutman, A.P., Friend, C.R.L., 1996, Evidence for life on Earth before 3800 million years ago: Nature, v. 270, p. 43-45.

Mueller, P.A., Wooden, J., Nutman, A., and Mogk, D., 1998, Early Archean Crust in the Northern Wyoming Province: Evidence from U-Pb Systematics in Detrital Zircons: Precambrian Research, v. 91, p. 295-307.

Mueller, P. A., Wooden, J. L. and Nutman, A. P., 1992, 3.96 Ga zircons from an Archean quartzite, Beartooth Mountains, Montana: Geology, v. 20, p. 327-330.

Nadeau, P., 2003, Structural investigation of the Porpoise Cove area, Northeastern Superior Province, Northern Quebec: M.Sc. thesis, Simon Fraser University, Burnaby, pp. 95.

Nutman, A.P., Allaart, J.H., Bridgwater, D, Dimroth, E., and Rosing, M., 1984, Stratigraphic and geochemical evidence for the depositional environment of the Early Archean Isua supracrustal belt, southern West Greenland: Precambrian Research, v. 25, p. 365-396.

Nutman, A.P., McGregor, V.R., Friend, C.R.L., Bennett, V.C., Kinny, P., 1996, The Itsaq Gneiss Complex of southern West Greenland: the world's most extensive record of early crustal evolution 3900–3600 Ma: Precambrian Research, v. 78, p. 1–39.

Nutman, A.P., Friend, C.R.L. and Bennett, V.C., 2001, Review of the oldest (4400-3600) geological and mineralogical record: Glimpses of the beginning. Episodes, v. 24, p. 93-101.

Nutman, A.P., Friend, C.R.L., Barkera, S. and McGregor, V., 2004, Inventory and assessment of Palaeoarchaean gneiss terrains and detrital zircons in southern West Greenland: Precambrian Research, v. 135, p. 281–314.

O'Neil, J., Maurice, C., Stevenson, R.K., Larocque, J., Cloquet, C., David, J. and Francis, D. 2007. The geology of the 3.8 Ga Nuvvuagittuq (Porpoice Cove) Greenstone Belt, Northeastern Superior Province, Canada; Developments in Precambrian Geology, vol. 15: Earth's Oldest Rocks. Edited by Martin J. Van Kranendonk, R. Hugh Smithies, Vickie C. Bennett. In press.

Percival, J.A. and Skulski, T., 2000, Tectonothermal evolution of the northern Minto Block, Superior Province, Quebec, Canada: Canadian Mineralogist, v. 38, p. 345-378.

Percival, J.A., Mortensen, J.K., Stern, R.A., Card, J.D., and Begin, N.J., 1992, Giant granulite terranes of northeastern Superior Province; the Ashuanipi Complex and Minto Block: Canadian Journal of Earth Sciences, v. 29, p. 2287-2308.

Percival, J.A., Stern, R.A., Skulski, T., Card, K.D., Mortensen, J.K., and Bégin, N.J., 1994, Minto block, Superior province: Missing link in deciphering assembly of the craton at 2.7 Ga: Geology, v. 22, p. 839-842.

Rabeau, O., 2001. Étude de l'évolution du Néodyme dans la croûte continentale du nord-est de la provence du Supérieure, Nunavik, québec. MSc thesis, Université du Québec à Montréal, pp. 76.

Rosing, M., 1999, ^{13}C-depleted carbon microparticles in >3700 Ma seafloor sedimentary rocks from West Greenland: Science, v. 283, p. 674-676.

Rosing, M., Rose, N., Bridgwater, D., and Thomsen, H., 1996, Earliest part of Earth's stratigraphic record: A reappraisal of the >3.7 Ga Isua (Greenland) supracrustal sequence: Geology, v. 24, p. 43-46.

Schidlowski, M., 2001, Carbon isotopes as biogeochemical recorders of life over 3.8 Ga of Earth history: evolution of a concept: Precambrian Research, v. 106, p. 117–134.

Simard, M., Parent, M., David, J., and Sharma, N.M.S., 2003, Géologie de la région de la Rivière Innuksuac (34K et 34L): Ministère des Ressources naturelles, de la Faune et des Parcs, Québec, RG 2002-10.

Skulski, T., Corkery, M.T., Stone, D., Whalen J.B. and Stern, R.A., 2000, Geological and geochronological investigations in the Stull Lake-Edmund Lake greenstone belt and granitoid rocks of the northwestern Superior Province: In Report of Activities 2000, Manitoba Industry, Trade and Mines, Manitoba Geological Survey, p. 117-128.

Stern, R.A., Percival, J.A., Mortensen, J.K., 1994. Geochemical evolution of the Minto block: a 2.7 Ga continental magmatic arc built on the Superior proto-craton: Precambrian Research, v. 65, p. 115-153.

Stevenson, R.K., David, J., and Parent, M., 2006, Crustal Evolution of the Western Minto Block, Northern Superior Province, Canada: Precambrian Research, v. 145, 229-242.

Thurston, P.C., 2002, Autochthonous development of Superior Province greenstone belts?: Precambrian Research, v. 115, p. 11-36.

Tomlinson, K.Y., Hughes, D.J., Thurston, P.C., and Hall, R.P., 1999, Plume magmatism and crustal growth at 2.9–3.0 Ga in the Steep Rock and Lumby Lake area, Western Superior Province: Lithos v. 46, p. 103–136.

Valeriano, C., Machado, N., Simonetti, A., Valladares, C. Seer, H., and Simoes, L.S., 2004, U-Pb geochronology of the southern Brazilia belt (SE-Brazil): sedimentary provenance, Neoprotrozoic orogeny and assembly of West Gondwana: Precambrian Research, v. 130, p. 27-55.

Vavra, G., Schmid, R., and Gebauer, D., 1999, Internal morphology, habit and U-Th-Pb microanalysis of amphibolite to granulite facies zircons: geochronology of the Ivrea Zone (Southern Alps): Contribution to Mineralogy and Petrology, v. 134, p. 380-404.

Whitehouse, M.J., Kamber, B. and Moorbath, S., 1999, Age significance of U–Th–Pb zircon data from early Archaean rocks of west Greenland—a reassessment based on combined ion-microprobe and imaging studies: Chemical Geology, v. 160, p. 201-224.

Whitehouse, M.J. and Kamber, B., 2005, Assigning Dates to Thin Gneissic Veins in High-Grade Metamorphic Terranes: A Cautionary Tale from Akilia, Southwest Greenland: Journal of Petrology, v. 46, p. 291-318.

Wilde, S.A., Valley, J.W., Peck, W.H. and Graham, C.M., 2001, Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Ga ago: Nature, v. 409, p. 175-178.

Wardle, R.J., James, D., Scott, D.J. and Hall, J., 2002, The southeastern Churchill Province: synthesis of a Paleoproterozoic transpressional Orogen: Canadian Journal of Earth Sciences, v. 39, p. 639–663.

FIGURE CAPTIONS

Figure 1. (A) The Superior Province showing location of inset (B) of Minto Block. Numbers 1 to 6 refer to localities discussed in text. 1) Isua greenstone belt and Itsaq Gneiss complex, Greenland; 2) Assean Lake, Manitoba; 3) Northern Superior Superterrane; Ontario 4) Minnesota River Valley gneisses; 5) Beartooth Mountains; Wyoming Craton 6) Acasta Gneiss, Slave Province, Northwest Territories. (B) Geological subdivisions of the Minto Block (after Percival et al., 1994) in the Northeast Superior Province and location of the Nuvvuagittuq supracrustal sequence and figure 2 (square) in the Inukjuak domain.

Figure 2. Geological map of the Nuvvuagittuq supracrustal sequence modified after David et al. (2002) and O'Neil et al. (2007) with U-Pb geochronology sample localities (stars).

Figure 3. Photographs of Nuvvuagittuq lithologies: (A) massive amphibolites with quartz veins; color variation due to variable cummingtonite content; (B) detail of massive amphibolite; (C) beige colored cummingtonite-anthophyllite amphibolite (right) with gabbro (left); (D) garnet-bearing biotite + cummingtonite + anthophyllite + quartz amphibolite; (E) rusty-colored ultramafic sill in background with gabbro sill in foreground; (F) banded iron formation; (G) banded iron formation with quartz-rich silica oxide facies (white unit); (H) conglomeratic unit POR29 with quartz clasts, black amphibolitic clasts and grey biotite schist clasts; (I) Nuvvuagittuq supracrustal sequence showing conglomeratic unit (Cgl) clearly visible to the left, alternating with silicate-rich iron formation, and boudinaged amphibolite unit to the right; (J) detail of dated felsic schist POR1091f; (K) sampled outcrop showing felsic schist folded with amphibolite and conglomeratic unit (Cgl); and (L) tectonized tonalite POR23.

Figure 4. U-Pb Concordia diagrams for (A) Plagioclase + Quartz + Biotite schist POR1091f; (B) Plagioclase + Quartz + Biotite schist POR134; (C) Conglomerate POR29: (D) Mylonitized Tonalite

POR23; and (E) Pegmatite POR1091G and amphibolitic gabbro POR30. Data-point error ellipses are 2σ.

Figure 5. Age versus Nd isotope compositions of selected units from the Nuvvuagittuq supracrustal sequence compared with data from Southern West Greenland (squares) and Labrador (rectangle). Greenland and Labrador data from Bennett et al., 1993; Blichert-Toft et al., 1999; Collerson et al., 1991. DM: Depleted Mantle; CHUR: Chondritic Uniform Reservoir.

Figure 6. Major Precambrian elements from North America and Greenland (modified from Hoffman, 1989) showing remnants of oldest terrains (3.6 Ga to 4.0 Ga) located in the geological provinces of Superior, Nain, Wyoming and Slave: (1) Beartooth Mountains, Montana; (2) Assean Lake, Manitoba; (3) Northern Superior Superterrane, Ontario; (4) Nuvvuagittuq, Québec; (5) Minnesota River Valley gneisses; (6) Saglek-Hebron area, Northern Labrador; (7) Itsaq Gneiss complex, Greenland; (8) Acasta Gneiss, Slave Province, Northwest Territories. See text for discussion. BO: Baffin Orogen; KO: Ketillidian Orogen; NO: Nagssugtoqidian Orogen; NQO: New Quebec Orogen; PO: Penokean Orogen; TOG: Torngat Orogen; TO: Talston Orogen; THO: Trans-Hudson Orogen; WO: Wopmay Orogen; GP: Grenville Province; HP: Hearne Province; RP: Rae Province; MCR: Mid-Continent-Rift; NQU: New Quebec Ungava.

DATA REPOSITORY MATERIAL

Figure A1. International and in-house U-Pb zircons standards measured during the period of this study. (A) U-Pb TIMS analyses for z6266 zircon standard (560.2 ± 1.1 Ma) agree within error of the age (559.0 ± 0.2 Ma) determined by Stern and Amelin (2003). (B) An in-house U-Pb zircon standard (KL01125) analyzed by TIMS (2760.2 ± 1.2 MA) agrees within error with U-Pb analyses by LA-MC-ICP-MS (C; 2759.8 ± 1.4 Ma).

Figure A2. Cathodeluminescence photographs of representative zircons from dated units. Zircons from (A) Plagioclase + Quartz + Biotite schist POR1091f, (B) Plagioclase + Quartz + Biotite schist POR134, (C) conglomerate POR29, and (D) mylonitized tonalite POR23 all show prominent primary oscillatory (OZP) or sector zoning (SZP) and secondary structures such as partial trangressive recrystallization (TRS) or secondary overgrowths (OGS). Zircons from (B) felsic schist POR134 and (E) amphibolitic gabbro POR30 lack distinct primary magmatic features and have a diffuse/mottled appearance and/or radial-sector zoning (RZS) suggestive of metamorphic growth (Hanchar and Miller, 1993).

ADDITIONAL REFERENCES RELATED TO DATA REPOSITORY

Hanchar, J.M., and Miller, C.F., 1993, Zircon zonation patterns as revealed by cathodoluminescence and backscattered electron images; implications for interpretation of complex crustal histories: Chemical Geology, v. 110, p. 1-13.

Stern, R.A., Amelin, Y., 2003, Assessment of errors in SIMS zircon U–Pb geochronology using a natural zircon standard and NIST SRM 610 glass. Chemical Geolology, v. 197, p. 111-142.

Table 1. U-Pb zircons analyses (TIMS-ID) for Nuvvuagittuq lithologies

Num.	Weight (mg)	U (ppm)	Th/U	PbCom (pg)	204Pb/206Pb	208Pb/206Pb	206Pb/238U	±1σ	207Pb/235U	±1σ	207Pb/206Pb	±1σ	207Pb/206Pb Age (Ma)	±2σ
POR1091G - late pegmatite dyke; UTM 339752E 6463032N; L.I. 815 ± 530 Ma, U.I. 2686 ± 12 Ma														
1	0.006	3086	15.52	324.3	1672	4.989	0.4583	0.65	11.247	0.66	0.17798	0.07	2634.1	2.4
2	0.002	4683	16.83	249.2	1206	4.965	0.5007	0.47	12.618	0.47	0.18278	0.07	2678.3	2.3
3	0.008	2232	14.20	42.4	13030	3.999	0.5213	0.32	13.199	0.32	0.18360	0.10	2685.6	3.2
4	0.005	2309	13.27	14.0	26928	3.795	0.5173	0.38	13.116	0.38	0.18391	0.06	2688.4	2.0
POR30 - amphibolitic gabbro; UTM 339940E 6463086N; age 2692.7 ± 2.8 Ma														
1	0.001	86	0.30	3.3	1391	0.085	0.5188	0.25	13.185	0.26	0.18433	0.10	2692.2	3.2
2	0.0007	69	0.32	3.9	878	0.091	0.5197	0.38	13.223	0.37	0.18455	0.17	2694.1	5.7
POR23 - mylonitized tonalite; UTM 339479E 6465022N; L.I. 875 ± 500 Ma, U.I. 3658.5 ± 2.5 Ma														
1	0.002	74	0.837	6	1128	0.227	0.7529	0.31	35.056	0.33	0.335283	0.10	3641.4	2.9
2	0.001	131	0.821	5.4	1541	0.229	0.7363	0.44	34.241	0.47	0.337263	0.15	3650.4	4.7
3	0.002	127	0.888	2.9	4223	0.239	0.7605	0.23	35.508	0.24	0.338625	0.05	3656.6	1.6
4	0.006	79	0.848	8.4	2430	0.229	0.7597	0.26	35.520	0.27	0.339083	0.06	3658.6	1.9
POR1091f plagioclase-quartz-biotite schist; UTM 339855E 6463017N; L.I. 2227 ± 200 Ma, U.I. 3779 ± 21 Ma														
1	0.002	130	0.207	3.2	2657	0.065	0.6771	0.22	29.937	0.25	0.32065	0.08	3572.9	2.6
2	0.002	227	0.381	22.9	909	0.112	0.7144	0.21	33.412	0.23	0.33921	0.09	3659.2	2.8
3	0.002	256	0.269	33.4	537	0.081	0.7100	0.33	33.262	0.39	0.33978	0.18	3661.8	5.4
4	0.001	256	0.285	32.4	554	0.082	0.7171	0.35	33.603	0.41	0.33985	0.17	3662.1	5.2
5	0.002	177	0.507	6.8	2443	0.143	0.7355	0.20	35.087	0.22	0.34597	0.07	3689.3	2.2
6	0.001	227	0.368	7.3	2909	0.109	0.7352	0.34	35.270	0.34	0.34793	0.08	3698.0	2.4
7	0.003	411	0.449	1.7	27856	0.129	0.7307	0.15	35.090	0.17	0.34831	0.06	3699.6	2.0
8	0.001	331	0.447	4.1	7604	0.125	0.7518	0.17	36.355	0.18	0.35071	0.05	3710.1	1.4
9	0.003	224	0.494	8.8	3703	0.137	0.7614	0.19	36.838	0.20	0.35092	0.05	3711.0	1.4
10	0.003	65	0.453	4.8	1962	0.127	0.7497	0.32	36.303	0.34	0.35100	0.09	3711.3	2.7
11	0.001	273	0.476	3.5	3728	0.132	0.7534	0.19	36.557	0.21	0.35192	0.06	3715.3	1.8
12	0.004	273	0.374	5.7	9086	0.104	0.7561	0.24	36.738	0.25	0.35238	0.07	3717.3	2.2
13	0.002	68	0.551	1.6	4054	0.158	0.7494	0.24	36.596	0.26	0.35417	0.10	3725.0	2.9
14	0.002	119	0.398	9.5	1235	0.110	0.7692	0.25	37.774	0.26	0.35615	0.07	3733.5	2.0
15	0.001	140	0.511	4.5	1526	0.143	0.7668	0.31	38.123	0.32	0.36058	0.13	3752.3	3.8
POR134 plagioclase-quartz-biotite schist; UTM 339620E 6464220N; L.I. 400 ±100 Ma; U.I. 3366.4±2.3 Ma														
1	0.001	82.4	0.878	3.8	5031	0.245	0.6684	0.156	25.769	0.171	0.27960	0.048	3360.7	1.5
4	0.001	87.2	1.049	2.6	2155	0.289	0.6809	0.286	26.307	0.292	0.28020	0.07	3364.0	2.2
5	0.002	86.5	0.954	7.6	997	0.262	0.6816	0.272	26.252	0.279	0.27935	0.074	3359.3	2.3
7	0.001	84.5	1.106	3.2	4001	0.305	0.6787	0.164	26.250	0.178	0.28053	0.044	3365.9	1.4
8	0.001	90.4	0.926	3.8	1030	0.257	0.6739	0.343	25.764	0.356	0.27727	0.063	3347.6	2.0
9	0.001	91.6	1.071	6.3	829.7	0.295	0.6787	0.308	26.242	0.321	0.28043	0.055	3365.3	1.7
10	0.001	101.9	0.997	3.6	1228	0.277	0.6740	0.303	25.650	0.313	0.27600	0.052	3340.4	1.6

L.I.: Lower intercept; U.I.: Upper intercept; UTM coordinates from NAD27 Grid.

Table 2. LA-MA-ICPMS U-Pb data for Nuvvuagittuq lithologies

	208Pb/206Pb	±1σ	206Pb/238U	±1σ	207Pb/235U	±1σ	207Pb/206Pb	±1σ	207Pb/206Pb age (Ma)	±2σ
POR1091f - plagioclase-qartz-biotite schist; UTM 339855E 6463017N										
1	0.1099	0.9	0.6551	0.8	29.449	1.0	0.32602	0.19	3598.5	5.8
2	0.1472	0.9	0.7155	0.6	33.748	0.6	0.34210	0.19	3672.2	5.7
3	0.1592	1.5	0.7465	0.6	35.774	0.7	0.34756	0.07	3696.3	2.2
4	0.1498	1.1	0.7403	1.0	35.743	1.1	0.35018	0.16	3707.8	4.7
5	0.1367	1.1	0.7594	0.8	37.154	1.0	0.35486	0.06	3728.0	1.7
6	0.1683	1.2	0.7584	0.9	37.147	1.0	0.35523	0.11	3729.6	3.3
7	0.1545	1.1	0.7579	0.7	37.228	0.8	0.35623	0.08	3733.9	2.5
8	0.2376	1.7	0.7406	0.8	36.416	1.0	0.35661	0.13	3735.5	3.9
POR23 - mylonitised tonalite; UTM 339479E 6465022N; L.I. 0 ± 200, U.I. 3661.1 ± 4.1 Ma										
1	0.169	1.4	0.7434	1.4	34.618	1.4	0.33775	0.10	3652.6	3.2
2	0.225	1.6	0.7575	1.4	35.297	1.4	0.33793	0.10	3653.4	3.0
3	0.248	1.6	0.7507	1.4	34.980	1.4	0.33795	0.10	3653.5	3.1
4	0.205	1.5	0.7620	1.5	35.547	1.5	0.33835	0.10	3655.3	3.0
5	0.234	1.4	0.7466	1.4	34.972	1.4	0.33974	0.10	3661.6	2.9
6	0.197	1.4	0.7686	1.4	36.026	1.4	0.33993	0.10	3662.5	3.1
7	0.262	1.4	0.7477	1.4	35.053	1.4	0.34001	0.10	3662.8	2.9
8	0.233	1.6	0.7523	1.4	35.296	1.4	0.34026	0.10	3663.9	3.0
9	0.257	1.8	0.7570	1.3	35.537	1.3	0.34048	0.10	3664.9	2.9
10	0.264	1.7	0.7503	1.4	35.222	1.5	0.34048	0.10	3664.9	3.0
11	0.238	1.4	0.7528	1.4	35.362	1.4	0.34071	0.10	3665.9	3.1
12	0.236	1.4	0.7587	1.3	35.644	1.4	0.34074	0.10	3666.1	3.0
13	0.260	1.4	0.7661	1.4	36.001	1.4	0.34084	0.10	3666.5	3.1
POR29 - polymictic conglomerate; UTM 339869E 6463025N ; 1) L.I. 0 ± 250, U.I 3649 ± 10 Ma, 2) L.I. 2099 ± 130, U.I. 3787 ± 25										
1	0.055	12.5	0.5463	0.8	14.293	0.8	0.18974	0.27	2739.9	8.7
2	0.052	1.7	0.5525	0.7	16.371	0.9	0.21490	0.38	2942.9	12.3
3	1.490	0.9	0.5676	0.8	17.964	2.1	0.22955	0.22	3048.9	7.0
4	0.089	4.7	0.5966	0.9	20.205	1.2	0.24562	0.35	3156.8	11.1
5	0.113	1.4	0.5944	0.7	20.874	0.9	0.25468	0.41	3214.1	12.9
6	0.059	4.2	0.6399	0.9	23.088	1.8	0.26169	1.04	3256.9	32.7
7	0.048	1.4	0.5479	1.4	19.781	2.6	0.26182	0.75	3257.7	23.8
8	0.120	1.5	0.6360	1.0	23.185	1.4	0.26440	0.32	3273.1	9.9
9	0.190	2.4	0.6061	1.2	24.900	2.1	0.29795	0.99	3459.6	30.7
10	0.128	2.5	0.5968	1.7	25.002	2.2	0.30386	0.64	3490.0	19.7
11	0.017	0.8	0.7066	0.9	29.909	0.9	0.30698	0.10	3505.8	3.2
12	0.184	1.5	0.6856	0.8	30.894	1.0	0.32682	0.11	3602.2	3.5
13	0.214	2.3	0.6826	1.6	30.790	2.2	0.32715	0.66	3603.8	20.4
14	0.185	0.9	0.7504	0.8	34.589	1.2	0.33430	0.12	3636.9	3.8
15	0.163	0.9	0.7220	0.8	33.324	0.8	0.33476	0.31	3639.0	9.6
16	0.088	1.1	0.7527	0.8	34.745	0.9	0.33478	0.05	3639.1	1.6
17	0.138	1.0	0.7567	0.9	35.192	1.6	0.33731	0.07	3650.6	2.0
18	0.210	0.7	0.6871	0.7	32.039	0.9	0.33820	0.06	3654.6	1.8
19	0.248	0.9	0.7545	1.0	35.277	1.0	0.33912	0.05	3658.8	1.5
20	0.144	0.9	0.7551	0.9	35.308	1.0	0.33913	0.05	3658.9	1.6
21	0.230	0.7	0.7415	0.8	34.723	0.9	0.33963	0.08	3661.1	2.4
22	0.262	0.9	0.7373	1.1	35.028	1.4	0.34459	0.25	3683.2	7.5
23	0.213	0.7	0.7384	1.1	35.741	1.4	0.35105	0.20	3711.5	6.1
24	0.201	1.4	0.7684	0.7	37.964	0.8	0.35833	0.14	3742.8	4.3
25	0.207	0.9	0.6708	1.0	33.247	1.0	0.35946	0.06	3747.6	1.7
26	0.165	0.8	0.7590	0.9	37.626	1.0	0.35954	0.06	3747.9	1.7
27	0.217	0.8	0.7605	0.9	37.709	0.9	0.35962	0.05	3748.3	1.6
28	0.220	1.9	0.7631	0.8	38.105	1.0	0.36215	0.10	3758.9	3.1
29	0.203	2.3	0.7841	0.8	39.223	1.0	0.36282	0.06	3761.7	1.8
30	0.304	2.5	0.7029	0.7	35.364	1.1	0.36488	0.20	3770.3	6.1

L.I.: Lower intercept; U.I.: Upper intercept; UTM coordinates from NAD27 Grid.

Table 3. Sm-Nd data for selected Nuvvuagittuk lithologies

Sample	Age (Ga)	Rock type	Nd ppm	Sm ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	2s error	εNd(t)	TDM	UTM Easting	UTM Northing
POR1091f-1	3.82	qz-plag-bi schist	13.9	2.27	0.0988	0.510176	0.000010	0.3	3.9	339855	6463017
POR1091f-2	3.82	qz-plag-bi schist	19.7	3.05	0.0934	0.510085	0.000011	1.2	3.9	339855	6463017
POR28-2	3.82	amphibolite	7.23	2.08	0.1740	0.512178	0.000013	2.3		339855	6463017
POR26	3.82	amphibolite	12.1	2.96	0.1472	0.511426	0.000009	0.8	3.9	339670	6464152
POR16	3.82	amphibolite	11.5	2.92	0.1533	0.511512	0.000010	-0.5		340785	6465596
POR30	3.82	gabbro	4.48	1.36	0.1832	0.512426	0.000016	2.6		339940	6463086
POR23	3.66	marginal tonalite	21.6	3.66	0.1025	0.510263	0.000008	-1.8	3.9	339479	6465022
repeat	3.66	marginal tonalite	21.1	3.56	0.1017	0.510271	0.000014	-1.3	3.9	339479	6465022
WP-59	3.66	marginal tonalite	25.5	4.09	0.0968	0.510140	0.000008	-1.5	3.9	339538	6464578
WP-104	3.66	marginal tonalite	36.4	4.81	0.0799	0.509737	0.000008	-1.4	3.9	339457	6464791
WP-69b	3.66	central tonalite	23.3	4.22	0.1096	0.510497	0.000009	-0.6	3.9	340667	6463335
WP-67	3.66	central tonalite	14.1	2.20	0.0943	0.510117	0.000009	-0.8	3.8	340584	6463849
POR1091G2	2.7	pegmatite	7.96	1.61	0.1218	0.510854	0.000007	-8.8	3.8	339752	6463032
POR1091G2	2.7	pegmatite	8.45	1.73	0.1238	0.510887	0.000008	-8.9	3.8	339752	6463032

UTM coordinates from NAD27 Grid; Model ages (TDM) are calculated assuming a linear depleted mantle with a present day value of eNd = +10 and 147Sm/144Nd = 0.2136 (Jacobsen, 1988).