Neoproterozoic basement history of Wrangel Island and Arctic Chukotka: Integrated insights from zircon U–Pb, O, and Hf isotopic studies

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Chukotka/Wrangel basement geochronology

Abstract: The pre–Cenozoic kinematic and tectonic history of the Arctic Alaska Chukotka terrane (AAC) is not well known. Difficulty in assessing the AAC is predominantly due to lack of comprehensive knowledge about the composition and age of basement of the terrane. During the Mesozoic, the AAC was deformed by crustal shortening, followed by magmatism and extension, with localized high–grade metamorphism and partial melting, which obscured pre–orogenic geological relationships. Zircon U–Pb ages of five granitic and one volcanic sample from greenschist facies rocks on Wrangel Island range between 620±6 Ma and 711±4 Ma, whereas two samples from the migmatitic basement of the Velitkenay massif near the Arctic coast of Chukotka yield 612±7 Ma and 661±11 Ma ages. The age spectrum (0.95–2.0 Ga with peaks at 1.1 Ga, and minor 2.5–2.7 Ga) and trace element geochemistry of inherited detrital zircons in a 703±5 Ma Wrangel Complex granodiorite suggests a Grenville–Sveconorwegian provenance for the pre–700 Ma strata on Wrangel Island and correlation with strata from Arctic Alaska and Pearya. Temporal patterns of inheritance and O–Hf isotopes are consistent with Cryogenian–Ediacaran AAC magmatism in a peripheral/external orogenic setting (i.e., fringing arc on rifted continental margin crust).

Supplementary material: SIMS U–Pb zircon geochronology data, SIMS zircon ¹⁸O/¹⁶O isotopic data, LAICPMS zircon Lu–Hf isotopic data, and zircon cathode-luminescence images are available at

The Arctic Alaska–Chukotka terrane (AAC), forming the Alaskan–Russian margins of the modern Arctic and North Pacific ocean basins, includes Arctic Alaska, Chukotka and the Chukchi Sea shelf, and portions of the East Siberian, Beaufort, and Bering Sea shelves (Fig. 1). The AAC’s enigmatic role in the plate tectonic evolution of the Arctic region is complicated by multiple episodes of deformation and magmatism [e.g., Miller et al., *this volume (a)*], and thus determining its early geologic history poses major analytical challenges. Understanding this history is essential to

46 elucidating paleotectonic associations of the AAC with other basement complexes in
47 the circum-Arctic realm.

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49 The AAC contains numerous exposures of Neoproterozoic metamorphic and igneous
50 basement overlain by a Paleozoic and Mesozoic sedimentary cover sequence (e.g.,
51 Cecile et al., 1991; Patrick and McClelland, 1995; Amato et al., 2009, 2014; Pease et
52 al., 2014; Till et al., 2014a; Akinin et al., 2015). This paper focuses on the
53 geochronology and isotope geochemistry of Neoproterozoic basement rocks from
54 two localities (western Chukotka and Wrangel Island; Figs. 1 and 2) that are
55 exposed ~ 250 km apart in the central part of the AAC. Although in close proximity
56 relative to the AAC's total size (Fig. 1), these exposures have never previously been
57 correlated due to conspicuous differences in metamorphic grade.

58
59 The structure and composition of the AAC has been greatly influenced by crustal
60 shortening followed by extension and voluminous magmatism related to Mesozoic
61 era Pacific margin tectonics (e.g., Patrick, 1988; Miller and Hudson, 1991; Miller et
62 al., 1992; Moore et al., 1994; Hannula et al., 1995; Klemperer et al., 2002; Akinin et
63 al., 2009, 2013). Voluminous magmatism and high-grade metamorphism during the
64 Cretaceous has strongly overprinted Chukotka basement rocks, which are exposed
65 in the Velitkenay massif study area as a culmination of gneissose igneous and
66 metamorphic rocks (Figs. 2 and 3). Wrangel Island, by contrast is a far less
67 deformed basement complex, having experienced coeval greenschist-facies
68 conditions of deformation (Figs. 2 and 3) [Miller et al., 2010, *this volume (b)*; Akinin
69 et al., 2011].

70
71 The robustness of zircon as a geochronometer and geochemical fingerprint medium
72 (e.g., Kröner, 2010) allows us to read and relate the history of basement rocks
73 across this metamorphic gradient, despite the high-grade metamorphism and
74 partial melting of rocks in Chukotka. Integrating the U-Pb geochronology of igneous
75 rocks, oxygen and hafnium isotope geochemistry in zircon, and temporal-based
76 observations about changing character of crustal inheritance in magmas provides
77 insights on the history of Neoproterozoic basement rocks in the AAC terrane. These
78 observations establish potential paleogeographic and tectonic tie-points between
79 the crustal section of Arctic Chukotka and other continental masses in the circum-
80 Arctic.

81
82 Our geochronology-based approach documents several new findings regarding the
83 tectonic history of the Arctic. First, the correlation of the age spectrum of
84 inheritance in a 703 ± 5 Ma pluton from Wrangel Island to detrital zircon signatures
85 from other lithotectonic units in the AAC suggest stratigraphic ties between distant
86 regions of the terrane (Fig. 1). Similar age groups observed in our results, in the
87 pervasively deformed Nome Group metasedimentary rocks on Seward Peninsula
88 (Fig. 1) (previously interpreted as Paleozoic in age based on biostratigraphic
89 data) (Till et al., 2014a, 2014b) and in metasedimentary rocks of the Schist Belt of
90 the central Brooks Range (Fig. 1) (Hoiland et al., *this volume*) indicate that the AAC
91 may contain a regionally correlative, pre-700 Ma stratigraphic interval. Detrital

zircon signatures from these areas have similar age peaks to Grenville-Sveconorwegian-sourced Neoproterozoic strata of the Pearya terrane in the Canadian Arctic (Malone et al., 2014), located on the opposite side of the Amerasia Basin from the AAC (Fig. 1). Analysis of zircon inheritance combined with oxygen and hafnium isotopic compositions of 705-580 Ma (Cryogenian and Ediacaran) metaigneous basement rocks of the AAC establishes that magmatism in the central AAC exhibits temporal trends characteristic of an orogenic setting that generated increasingly primitive magma through time [i.e., external/peripheral orogen (Murphy and Nance, 1991; Collins et al., 2011; Cawood et al., 2016)]. This suggests central AAC became isolated from input of recycled continental detritus, as an offshore arc or an arc established on a rifted ribbon continent, during the period 705–580 Ma. In aggregate, these insights about the AAC’s relationship to other circum-Arctic terranes that comprised the northern margin of Rodinia in late Neoproterozoic time are invaluable for plate-model based paleogeographic and tectonic reconstructions.

GEOLOGICAL SETTING

Arctic Alaska Chukotka Terrane

The southern/southwestern margin of the AAC is juxtaposed against the much older (2.0-3.4 Ga) Kolyma–Omolon block along the Mesozoic-age South Anyui zone (Amato et al., 2015, and references therein) (Figs. 1 and 2). Farther west and north, this boundary of the AAC is projected offshore across the western part of the East Siberian continental shelf to the New Siberian Islands (Franke et al., 2008; Kuzmichev, 2009). The DeLong archipelago has been included as part the AAC based on basement age correlations from Zhokov Island (Fig. 1) (Akinin et al., 2015). Voluminous volcanic deposits of the mid- to Late Cretaceous Okhotsk–Chukotka volcanic belt (OCVB) obscure the South Anyui zone between Chukotka and Kolyma–Omolon east of the 168°E meridian (Fig. 2). The OCVB was generated in several distinct pulses by north dipping subduction beneath the southern continental margin of Arctic Chukotka (Tikhomirov et al., 2008; Akinin and Miller, 2011). Several additional episodes of crustal growth from magmatic addition and subduction accretion have occurred along the margin since Late Cretaceous time (e.g., Hourigan et al., 2009). This obscured boundary of Arctic Chukotka is proposed to cross the Bering shelf south of Saint Lawrence Island and link up with the Angayucham suture zone that delineates the southern boundary of the Alaskan part of the AAC (Churkin and Trexler, 1981; Nokleberg et al., 2000; Amato et al., 2015).

The oldest dated meta-igneous basement rocks in the AAC are the 968±5 Ma Ernie Lake orthogneiss, located in the southern Brooks Range (BR, Fig. 1)(Amato et al., 2014). The next oldest are ca. 870 Ma granitic orthogneisses and meta-volcanic rocks exposed on Seward Peninsula (Amato et al., 2009; 2014)(SP, Fig. 1). These isolated exposures are succeeded, from 750 Ma and continuing to the start of the Phanerozoic, by punctuated magmatism with age gaps no greater than 30 Ma (Amato et al., 2014)(Fig. 1b). Although magmatism recurred in a regular and

episodic manner after 750 Ma, the age interval(s) of magmatism are regionally variable (Fig. 1b). The oldest documented ages of Neoproterozoic magmatism in the non-Alaskan part of the AAC are much younger than the oldest ages reported for Arctic Alaska (Fig. 1b). On Wrangel Island, Neoproterozoic magmatism is as old as 700 Ma (Cecile et al., 1991), which is the oldest magmatism documented in the basement of the AAC outside of Arctic Alaska. In the westernmost part of the AAC, crustal xenoliths in Neogene lavas derived from basement rocks underlying Zhokov Island in the DeLong archipelago range from 660–600 Ma (Fig. 1)(Akinin et al., 2015). In eastern Chukotka (Koolen metamorphic complex), a broad range of ages from 670–565 Ma are documented (K, Fig. 1)(Natal'in et al., 1999; Amato et al., 2009; 2014). Farther west in mainland Chukotka, Neoproterozoic magmatism had not been reported prior to this study.

Study Areas

Wrangel Island is an isolated, 7600 km² landmass surrounded by the vast East Siberian continental shelf of the Arctic Ocean (Fig. 1). Located 140 km north of the Arctic coastline of Chukotka, ~400 km south of the shelf-slope break and nearly 1000 km east of the nearest of the New Siberian and DeLong Islands (Fig. 1), Wrangel Island is a key locality for studying the composition and geologic history of crust that makes up the northern flank of the AAC. The Precambrian basement of Wrangel Island (the Wrangel Complex) is exposed in the core of an E-W trending anticlinorium in the center of the island, unconformably overlain by Paleozoic and Mesozoic sedimentary cover (Figs. 2 and 3a) (Kos'ko et al., 1993). Wrangel Complex basement rocks are felsic to intermediate volcanic rocks, volcanoclastic rocks, slate/phyllite, minor grey and black slate, quartzite, conglomerate, and very minor mafic volcanic rocks, with quartz-feldspar porphyry, gabbro, diabase, felsite dikes and sills and small granitic plutons (Kos'ko et al., 1993). Previous geochronologic studies documented a Cryogenian age for volcanic and granitic rocks in Wrangel Complex (700 and 630 Ma; Cecile et al., 1991) whereas microfossils indicate middle Riphean (pre-Cryogenian) and latest Proterozoic–Early Cambrian age strata (Kos'ko et al. 1993 and references in *Russian* therein). In the study area, Devonian(?), Carboniferous, Permian and Triassic strata unconformably overlie the Wrangel Complex in a sequence that transitions from locally-sourced terrigenous shallow water clastic and carbonate shelf deposits in Late Paleozoic time to deep water siliciclastic strata in the Triassic (Fig. 3a)(Miller et al., 2010).

The stratigraphic section of Chukotka is similar to Wrangel Island, but, unlike Wrangel Island, it was the site of voluminous Cretaceous magmatism and high-grade metamorphism at deeper crustal levels (Figs. 2 and 3b)(Miller et al., 2009; Akinin et al., 2011). Most of western Chukotka is covered by Mesozoic age rocks that consist of Triassic to Early Cretaceous deep water sedimentary rocks intruded by mid-Cretaceous plutonic rocks and/or unconformably overlain by Late Cretaceous volcanic rocks of the Okhotsk-Chukotka Volcanic Belt (Fig. 2)(Miller and Verzhbitsky, 2009). The Velitkenay (massif) complex (Figs. 2 and 3b) is one of few areas in Chukotka where basement rocks are exposed. Here they crop out as a

migmatitic gneiss complex that experienced mid-Cretaceous age peak metamorphism, based on several lines of evidence. Meter-scale undeformed granite pegmatite and aplite dikes are intruded parallel to the dominant planar foliation in the study area. Cretaceous plutons in the Velitkenay complex span ~ 105–100 Ma (Akinin et al., 2012), coeval with widespread magmatism in Chukotka spanning ~ 118–100 Ma (Fig. 2) (Miller et al., 2009). Biotite and hornblende ^{40}Ar – ^{39}Ar ages from Velitkenay range from 100–95 Ma and exhibit well defined step heating plateau ages, signifying relatively fast cooling from high-grade conditions at that time [Miller et al., this volume (b)]. This mid-Cretaceous age deformation, metamorphism, and plutonism has obscured the original basement–cover relationships that are still observable on Wrangel Island (Fig. 3).

METHODS

Overview and Sample Preparation

Our investigation of Arctic Chukotka and Wrangel Island basement is built around analyses of nine igneous rock samples that yielded zircons with Neoproterozoic crystallization ages (Table 1). Several analytical techniques are integrated in this study: Secondary ion mass spectrometry (SIMS)–based zircon U–Pb geochronology (8 samples), zircon trace element geochemistry (3 samples), and zircon $^{18}\text{O}/^{16}\text{O}$ isotopic analyses (3 samples), and laser ablation–inductively coupled plasma mass spectrometry (LA–ICPMS) zircon Lu–Hf isotopic analyses (4 samples). Collectively, these methods are used to fingerprint and correlate the age and geochemistry of Neoproterozoic magmatism and crustal inheritance that characterizes the basement of Arctic Chukotka and Wrangel Island.

Six samples were collected on Wrangel Island during 2006 and two samples were collected from the Velitkenay complex in Arctic Chukotka during 2011. An additional sample used in the study is a previously dated orthogneiss from the Koolen Lake region on the Chukotka Peninsula (sample G31)(Fig. 1), for which a zircon U–Pb crystallization age of 574 ± 9 Ma (lower concordia intercept, 2σ) has been reported (Amato et al., 2014).

Five samples, all from Wrangel Island, are granite to granodiorite composition plutonic rocks (Table 1). Four of the five were collected from the Wrangel Complex basement: Sample VP06–35a is a granitic xenolith enclave enclosed in a potassium-feldspar augen gneiss VP06–35b (Fig. 4a), and samples ELM06WR28e and ELM06WR29 are from a weakly foliated granodiorite pluton. The fifth sample (VP06–36b) is a granitic cobble from conglomerate that unconformably overlies the Wrangel Complex (Figs. 3a, 4b). The sixth Wrangel Island sample (VP06–36a) is a quartz-rich meta-volcanic rock (Fig. 4c).

The remaining three samples, all from Chukotka, have experienced sufficiently high metamorphic grade during deformation that intrusive or extrusive proto-lithology cannot be determined (Table 1). Two of the samples are from orthogneiss outcrops:

Sample G31 from Koolen Lake and a fine-grained, quartzofeldspathic, biotite-bearing orthogneiss sample (11EGC21) from the Velitkenay complex. The third sample (11EGC36) is a leucogranite from the migmatite zone in the Velitkenay complex (Figs. 3b and 4d).

Zircon aliquots were handpicked from purified mineral separates produced by standard crushing, grinding, sieving, hydrodynamic, density and magnetic separation techniques on 0.25-2 kg of sample material. Zircons were mounted in epoxy, polished to expose crystal interiors, photographed under reflected light and imaged in a scanning electron microscope using a cathode luminescence (CL) detector. For SIMS work, zircon mounts were gold coated for conductivity, but gold coating was removed by light polishing and cleaning for LA-ICPMS analyses.

SIMS Zircon U-Pb geochronology

Prior studies of Wrangel Island basement rocks were carried out using thermal ionization mass spectrometry (TIMS) dating of bulk zircon separates (e.g., Cecile et al., 1991). SIMS Zircon geochronology guided by CL imaging allowed pre- and syn-magmatic growth domains in zircon crystals to be distinguished and analyzed. SIMS U-Th-Pb geochronology was carried out on zircon separates from eight igneous samples (six from Wrangel Island and two from Velitkenay) in order to obtain an igneous crystallization age for each sample and, in selected samples with dateable inherited zircon domains, to investigate ages of inherited material.

Zircon U-Pb ages for samples VP06-35a, VP06-35b, VP06-36a, and VP06-36b were measured using the Cameca IMS 1270 ion microprobe in the NordSIMS facility at Stockholm University (methodology of Whitehouse et al., 1999). The instrument was tuned to extract a -4 nA O_2^- primary beam from the oxygen source, which was used together with Kohler mode illumination of a ca. 33 mm beam aperture to evenly sputter a $25 \times 50 \mu\text{m}$ ellipsoid on polished zircon surfaces. A 30 eV energy window was used at 5600 mass resolution to separate $^{206}\text{Pb}^+$ from molecular interferences. A single collector electron multiplier was used in ion counting mode to measure secondary ion beam intensities for masses $^{196}[\text{Zr}_2\text{O}]^+$, $^{196}[\text{HfO}]^+$, $^{204}\text{Pb}^+$, $^{204}[\text{background}]^+$, $^{206}\text{Pb}^+$, $^{207}\text{Pb}^+$, $^{208}\text{Pb}^+$, $^{232}\text{Th}^+$, $^{238}\text{U}^+$, $^{248}[\text{ThO}]^+$, $^{254}[\text{UO}]^+$, and $^{270}[\text{UO}_2]^+$. U/Pb ratio calibration was based on analyses of the Geostandards zircon 91500. Data reduction was performed using NordAge (Whitehouse et al., 1999). NordSIMS analyses focused on determining the crystallization age of zircons in various samples and avoided pre-magmatic inclusion domains. CL images of grains analyzed at the NordSIMS facility are included as supplementary material.

Zircon U-Pb ages and selected trace element data for samples ELM06WR28e, ELM06WR29, 11EGC21, and 11EGC36 were measured using the Sensitive High Resolution Ion Micro Probe- Reverse Geometry (SHRIMP-RG) at the Stanford USGS Micro Analysis Center (SUMAC) at Stanford University using standard laboratory procedures for polished epoxy grain mounts. The SHRIMP-RG instrument was tuned to extract a -5 nA O_2^- primary beam from the oxygen source, which was

focused through a 100 μ m diameter Kohler aperture to create a 25–30 μ m sputter pit on polished zircon surfaces. Immediately prior to data acquisition for a given spot, the primary beam was rastered for two minutes to remove gold from the intended analytical spot. During data acquisition, sputtered secondary ions were accelerated into the mass spectrometer and relevant masses were measured on a single collector electron multiplier in ion counting mode. Run table acquisition parameters for mass stations were calibrated on MAD–Green zircon, with secondary tuning parameters adjusted for best peak shape and mass resolution of $\sim$ 7000 for $^{206}\text{Pb}^+$. Each analysis included 4–5 cycles of measurement of mass stations $^{89}\text{Y}^+$, $^{139}\text{La}^+$, $^{140}\text{Ce}^+$, $^{146}\text{Nd}^+$, $^{147}\text{Sm}^+$, $^{153}\text{Eu}^+$, $^{155}\text{Gd}^+$, $^{179}[\text{DyO}]^+$, $^{182}[\text{ErO}]^+$, $^{188}[\text{YbO}]^+$, $^{196}[\text{Zr}_2\text{O}]^+$, $^{196}[\text{HfO}]^+$, $^{204}\text{Pb}^+$, $^{204}[\text{background}]$, $^{206}\text{Pb}^+$, $^{207}\text{Pb}^+$, $^{208}\text{Pb}^+$, $^{232}\text{Th}^+$, $^{238}\text{U}^+$, $^{248}[\text{ThO}]^+$, $^{254}[\text{UO}]^+$, $^{270}[\text{UO}_2]^+$ with varying count times, resulting in each analysis requiring 12–20 minutes depending on acquisition parameters. Calibration of Pb/U and U/UO carried out using method of Ireland and Williams (2003) using R33 (419 Ma, Black et al., 2004) as primary standard. Over five sessions, 2-sigma errors of the weighted mean of standard Pb/U calibrations were 0.6%, 0.6%, 0.9%, 1.0% and 1.9%. Trace element concentrations were obtained from ratios of [trace element]/ $^{196}[\text{Zr}_2\text{O}]^+$ normalized to MAD–Green as concentration standard (Barth and Wooden, 2010). Raw data were reduced to ratios, concentrations, and ages using Squid2 (Ludwig, 2012).

For results from both laboratories, concordia and weighted mean plots of reduced U–Pb data were generated with Isoplot 3.75 (Ludwig, 2012) and DensityMap.R (Sircombe, 2007) using uncorrected and corrected for $^{204}\text{Pb}(\text{c})$ ratios of $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{238}\text{U}/^{206}\text{Pb}$ for concordia diagrams, and ^{207}Pb corrected $^{206}\text{Pb}/^{238}\text{U}$ ages for weighted mean calculations. DensityMap.R was used to visually recognize concordance of age clusters (Sircombe, 2007). Common lead corrections were applied using a modern-day average terrestrial common Pb composition, i.e., $^{207}\text{Pb}/^{206}\text{Pb} = 0.83$ (Stacey and Kramers, 1975), where significant ^{204}Pb counts were recorded and is assumed to represent surface contamination.

SIMS Oxygen Isotopic Ratios in Zircon

$^{18}\text{O}/^{16}\text{O}$ ratios of zircons from three samples were analyzed on the Cameca 1270 IMS at the UCLA SIMS Laboratory to gain insight on their petrogenesis. Analyses were carried out in routine fashion using methods described by Trail et al. (2007), operating in multi-collection mode with a CS^+ primary beam focused on a 15 μ m sputter pit. Both Velitkenay massif samples were analyzed in addition to the Koolen Lake orthogneiss (sample G31)(Fig. 1). For sample 11EGC36, which yielded few large zircons in the original mineral separation, the oxygen work was done on zircons previously dated using SHRIMP–RG, that were plucked from the original U–Pb mounts, remounted, polished, and imaged using standard mount preparation methods in a fresh epoxy matrix for oxygen work. For the other two samples, fresh zircons were mounted from original mineral separates. Oxygen data was collected using R33 (Black et al., 2004) as a primary standard and 91500 (Wiedenbeck et al.,

2004) as a secondary standard. Data are reported as $\delta^{18}\text{O}$ (VSMOW) results, which is the $^{18}\text{O}/^{16}\text{O}$ ratio ($\pm 2\text{SD}$) relative to Vienna Standard Mean Ocean Water measured *per mil* (Valley, 2003). Cited precisions given for unknowns are calculated as the geometric mean of the standard reproducibility and the analytical uncertainty during analysis of the unknown. CL images of grains analyzed for oxygen isotopes are included as supplementary material.

Laser Ablation Lu–Hf Isotopic Ratios in Zircon

Laser ablation Lu–Hf zircon isotopic analyses were carried out at the Washington State University Geoanalytical Laboratory on a granodiorite sample from Wrangel Island (ELM06WR29), the two Velitkenay samples, and the Koolen Lake orthogneiss sample to gain further information about their petrogenesis. Relatively large zircons that had been previously analyzed during SIMS U–Pb and/or oxygen work were targeted for Lu–Hf analysis, with a few zircons from each sample that exhibited similar CL appearance to primary targets, but no prior SIMS analyses, included as secondary targets (see Supplementary Materials). Target domains in zircons were ablated with a New Wave 213 nm Nd:YAG laser using a 40 μm diameter circular laser spot. Data acquisition and reduction protocol followed were as described in section 2.2.2 of Fisher et al. (2014) using Mudtank zircon as the primary standard ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282507$) and R33 (Fisher et al., 2014) and 91500 (Fisher et al., 2014) as secondary standards (Supplementary Table 3). $^{176}\text{Hf}/^{177}\text{Hf}_{\text{initial}}$ isotopic ratios and $\epsilon\text{Hf}_{\text{initial}}$ results were calculated using U–Pb ages of 661 Ma, 611 Ma, and 574 Ma for the respective samples, $\lambda = 1.867 \times 10^{-11}/\text{yr}^{-1}$, present day $^{176}\text{Lu}/^{177}\text{Hf}_{\text{CHUR}} = 0.0336$ and present day $^{176}\text{Hf}/^{177}\text{Hf}_{\text{CHUR}} = 0.282785$ (CHUR=chondritic uniform reservoir). A correction factor of 1.00011248 was applied to the measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratios to obtain corrected results. Sample averages are reported as $\epsilon\text{Hf}_{\text{initial}}$ mean values ($\pm 2\text{SD}$) relative to CHUR. CL images of grains analyzed for hafnium isotopes are included as supplementary material.

RESULTS

Neoproterozoic Magmatism Geochronology

U–Pb geochronology of the Wrangel Island igneous samples produced unambiguous results (Figure 5 and Table 1). Three distinctive age groups are apparent from the data. The meta-volcanic sample (VP06–36a), both granodiorite samples (ELM06WR28e, ELM06WR29), and the granitic xenolith (VP06–35a) from the augen gneiss all yield ^{207}Pb –corrected $^{206}\text{Pb}/^{238}\text{U}$ weighted average ages ($\pm 2\text{s}$) in the range 697.3 ± 5.0 Ma to 711.4 ± 4.2 Ma (Fig. 5). The granitic clast (VP06–36b) from the Devonian(?) conglomerate is slightly younger (673.3 ± 4.2 Ma)(Fig. 5). The augen gneiss (VP06–35b) that contained the granitic xenolith is considerably younger, yielding an age of 619.8 ± 6.2 Ma.

In the Velitkenay massif, zircon ages from the orthogneiss sample (11EGC21) exhibit more discordance than any of the Wrangel results and define a discordia chord (Fig. 6a). Interpretation based on additional geological evidence beyond geochronology data is required to constrain crystallization age and the timing of Pb loss leading to discordance. This sample exhibits a well-developed gneissose foliation, indicating its protolith was metamorphosed and deformed under high-grade conditions. Although metamorphism complicates protolith determination, the consistency of zircon rare earth element spectra from analysis to analysis suggests zircons crystallized from a common igneous source (e.g., Hoskin and Schaltegger, 2003)(Fig. 6b). Oxygen isotope ratios of the zircons, discussed in a following section, also support the interpretation of a single igneous protolith. Deformation and metamorphism of the sample make determination of volcanic versus plutonic origin difficult, but the abundance of quartzofeldspathic minerals indicate an intermediate to felsic protolith.

Two approaches to determine upper and lower concordia intercept ages for the sample data (representing the ages of crystallization and Pb loss, respectively) are evaluated. The first approach involves calculating a crystallization age based only on the analytical data (independent of other geologic evidence), and suggests crystallization at 681 ± 16 Ma and Pb loss at 148 ± 28 Ma (Fig. 6a). The alternative approach, fixing the lower intercept at 102 ± 4 Ma (the age range of Mesozoic magmatism independently documented in Velitkenay, Akinin et al., 2012) yields a younger (but equivalent within uncertainties) crystallization age of 661 ± 11 Ma (Fig. 6c). If the results are calculated without fixing the lower intercept, but excluding the two most discordant analyses, an intermediate result is obtained which exhibits a lower intercept age of 114 ± 57 Ma (within error of the age of magmatism in the Velitkenay complex) and a crystallization age of 673 ± 18 Ma (Fig. 6a). In evaluating each of these results, the result calculated with the fixed lower intercept is determined to be the most robust (Fig. 6c). The occurrence of widespread plutonism and regional metamorphism is compelling geologic evidence for a Pb loss event in mid-Cretaceous time, and the first (data-only, Fig. 6a) result yields a lower intercept age that is older than the independently constrained timing of magmatism by at least 15 Ma. Conversely, fixing the lower intercept to the age of the thermo-magmatic event and including all the data yields a slightly younger, more precise, and still statistically sound result (Fig. 6c). It is notable that the crystallization age determined for this sample is younger than the 702 ± 12 Ma range of ages for the oldest samples in the Wrangel Complex but overlaps the dated 673 ± 4 Ma granitic cobble from Wrangel Island.

Sample 11EGC36 from the Velitkenay massif was collected from a leucocratic phase that is widespread in the core of the metamorphic complex and commonly contains schlieren of foliated and partially melted gneiss (Fig. 4d). Although this sample was collected from an exposure of largely undeformed leucocratic granite that cuts bodies of 105 Ma foliated granite (Akinin et al., 2012), all zircons analyzed yield Neoproterozoic ages. Like orthogneiss sample 11EGC21, the zircons from this sample exhibit consistently similar rare earth element concentrations (Fig. 6d) and

oxygen isotope signatures (discussed in a following section), suggesting they were derived from a common igneous source. Unlike sample 11EGC21, most of the analyses of zircons from this sample yield concordant results, with 16 of 20 analyses clustering to define a concordia age at 611.4 ± 5.7 Ma (Fig. 6e). The remaining four (more discordant) results plot scatter near a chord with a calculated upper intercept at 609 ± 13 Ma and a fixed lower intercept at 102 ± 4 Ma (Fig. 6e). The 16 data points used in the concordia age calculation yields a ^{207}Pb corrected $^{206}\text{Pb}/^{238}\text{U}$ weighted mean age of 612.3 ± 7.3 Ma with an asymmetric skew towards younger ages in the population, likely reflecting some Pb loss (Fig. 6f). It is notable that the age of this zircon population is within error of the age determined for the augen gneiss from Wrangel (VP06–35b)(Fig. 5). Unlike that sample, which also contained inheritance of >660 Ma zircons, none of the zircons from 11EGC36 yielded ages older than this youngest age group.

Detrital Zircon Inheritance

Most zircons in granodiorite sample ELM06WR29 contain inherited domains clearly recognizable in CL images (Fig. 7). After determining the crystallization age of this sample by analyzing rim domains, additional zircons were mounted for analysis of core ages. The geochronology results span a range of predominantly Mesoproterozoic ages and nearly concordant isotope ratios (Fig. 7). About 50% of the inherited zircons from ELM06WR29 are <1.3 Ga and 90% are <1.8 Ga (Fig. 6). Although several analyses yielded slightly discordant (>-5 to $<+10\%$) Pb/U and Pb/Pb ages <1 Ga, the uncertainties (2σ) of these results all range as old as 1 Ga. As shown in the histogram and probability density plot inset (Fig. 6b), whether the data set only includes the most concordant analyses or includes some discordant ones as well, the overall spectra are remarkably similar.

Evaluating inheritance results from sample ELM06WR29 requires making some assumptions about the physical mechanism by which the pre-magmatic (i.e., inherited) zircons were incorporated and preserved in the granodiorite host rock. The numerous age peaks observed between 1.0–1.8 Ga suggest this intrusion incorporated crustal sources that contained a well-mixed population of zircons, such as expected from siliciclastic sedimentary rocks. The relative lack of discordance indicates pre-magmatic zircons did not experience significant high-grade conditions of metamorphism as would be expected at depth in the crust where the granodiorite magma was generated. Thus, we suggest the inherited zircons were incorporated at supracrustal depths and the spectrum of inherited, predominantly concordant ages is a proxy for the detrital zircon signature of the strata the pluton intruded.

The youngest inherited ages from ELM06WR29 were used to constrain the depositional age of the youngest strata intruded by the pluton. Because the youngest U–Pb ages exhibit variable degrees of discordance, both $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages were examined in this process. The three youngest, reliably concordant ($<+5\%$ and $>-5\%$ discordance) $^{206}\text{Pb}/^{238}\text{U}$ results (965 ± 42 Ma, 984 ± 54 Ma, and 994 ± 82

Ma) yield a weighted mean age of 975 ± 31 Ma (MSWD=0.3). This result excludes three ages that are >+5% discordant (963 ± 86 Ma, +12% discordant; 808 ± 16 Ma, 21% discordant; 852 ± 108 Ma, 30% discordant) and one age <-5% discordant (981 ± 28 , -10% discordant). Of those four, it is notable the two ages with lesser discordance do overlap in uncertainty with the weighted mean result. The three youngest $^{207}\text{Pb}/^{206}\text{Pb}$ results (899 ± 102 Ma, 969 ± 48 Ma, and 994 ± 46 Ma) yielded a nearly identical weighted mean age of 975 ± 32 Ma (MSWD=1.3). These results, coupled with the age obtained for pluton crystallization, indicate the youngest supracrustal strata intruded by the pluton were deposited between 703 ± 5 Ma and 975 ± 32 Ma.

Selected trace element geochemical concentrations and ratios (Hf, Th/U, Y/Yb, and Hf/Yb) of inherited zircons in ELM06WR29 are shown in Figure 7c. As a simple test of the central tendency and variability of these results through time, we calculated an arithmetic mean and standard deviation of the data at ten discrete 100 Myr intervals, beginning with the interval 875–975 Ma (Fig. 7c). Several age measurements overlap multiple age bins due to analytical uncertainty, so each data point is included in a single bin, determined by the best estimate of the age value (e.g., spot WR29(C)–20.1 yielded an age of 1474.9 ± 33.9 Ma and is included in the bin 1375–1475). Because ≤ 2 grains fall in any bin older than 1775–1875 Ma, those intervals and their data are excluded from the calculations (Fig. 7c). The arithmetic mean and standard deviation of syn-magmatic results from the granodiorite ($n=33$) are shown for comparison in background (Fig 7c). The binned results exhibit increasing mean Hf and Hf/Yb, and decreasing mean Th/U from 1.5 to 1.0 Ga (Fig. 7c). A reversal in the trends of mean Hf, Y/Yb, and Th/U is observed in the youngest (875–975 Ma) bin ($n=3$). Relative to the syn-magmatic results, the inheritance results exhibit lower average Y/Yb and Th/U ratios, but less pronounced deviation in Hf and Hf/Yb (Fig. 7c).

Oxygen Isotopes

Forty two (of 45) R33 standard analyses yielded an uncorrected average $\delta^{18}\text{O}$ value of $+6.05\text{‰} \pm 0.36\text{‰}$ (all uncertainties reported as 2 standard deviations). Three R33 analyses were discarded because they yielded $\delta^{18}\text{O}$ values around $+8\text{‰}$. All results were divided by a correction factor of ~ 1.09 so that the average $\delta^{18}\text{O}$ value of R33 analyses equaled the published value of $+5.55\text{‰}$ (Black et al., 2004). This resulted in a marginally low average value (but still within 2 st. dev. error envelope) for the experimental value of zircon standard 91500 relative to the published value ($+9.75\text{‰} \pm 0.34\text{‰}$ versus 10.07, Wiedenbeck et al., 2004). For 91500 and the unknowns, the uncertainty reported is twice the standard deviation of the population of individual results. The population means for unknowns are reported in Table 1.

The 661 ± 11 Ma orthogneiss sample from the Velitkenay complex (sample 11EGC21) yielded results ranging $+4.49\text{‰} \pm 0.10\text{‰}$ to $+6.19\text{‰} \pm 0.11\text{‰}$ (mean value

+5.87‰ ± 1.32‰, n=11)(Fig. 8). If the three lightest results are excluded, the uncertainty decreases by a factor of >2.5, yielding a result of +5.95‰ ± 0.48‰. The 612±7 Ma inherited zircon population in the undeformed leucocratic granite (11EGC36) yielded a slightly lighter and less scattered set of results ranging +4.28‰ ± 0.12‰ to +5.18‰ ± 0.10‰ (mean value +4.85‰ ± 0.50‰, n=12)(Fig. 8). Excluding the lightest result yields a slightly more precise result of +4.90‰ ± 0.32‰. For the 574±9 Ma Koolen Lake orthogneiss (sample G31 from Amato et al., 2014), the results range +5.73‰ ± 0.08‰ to +6.24‰ ± 0.14‰, (mean value of +6.02‰ ± 0.29‰, n=16), which overlaps the result from the Velitkenay orthogneiss but is heavier in ¹⁸O (beyond the uncertainty limits) than the 612±7 Ma zircon population (Fig. 8). Likewise, results from the older two zircon populations are within the +5.3‰ ± 0.3‰ range documented for zircon in equilibrium with mantle (Valley, 2003), whereas the Koolen Lake population is slightly heavier (Fig. 8). However, given the Koolen orthogneiss contains inherited zircons that yield ²⁰⁷Pb/²⁰⁶Pb ages as old as 1.7 Ga (Amato et al., 2014), it stands to reason that some crustal component is contributing to the magma geochemistry of this sample. Although both core and rim domains were analyzed in three zircons from sample G31, the variation between zircon growth zones is <0.5‰, albeit in each of the three grains, rims exhibited heavier results than the cores.

Hafnium Isotopes

The following results were obtained for secondary standards (mean ± 2 std. dev. results of solution MC-ICPMS analyses of Fisher et al. (2014) shown in parentheses): R33, +7.1 ± 0.9 (+8.0 ± 0.7); and 91500, +6.3 ± 1.3 (+6.9 ± 0.4). The population means for unknowns are reported in Table 1.

The granodiorite sample from Wrangel Island (sample ELM06WR29) was the most problematic sample to analyze, as four of the eight zircon grains analyzed were drilled through by the laser, and yielded less than 30 seconds of data for each analysis. The results of the remaining four analyses ranged from -4.7 ± 1.1 to -2.9 ± 1.2 (mean value of -3.6 ± 1.5), which are the least radiogenic (i.e., most negative in εHf_{initial}) results observed in the study (Fig. 9). For sample 11EGC21, results ranged +3.0 ± 1.3 to +5.3 ± 1.1 (mean value of +4.2 ± 1.3, n=10)(Fig. 9). The inherited zircons in sample 11EGC36 yielded a slightly more radiogenic and scattered set of results ranging +6.7 ± 1.4 to +10.2 ± 1.1 (mean value of +7.7 ± 2.7, n=8)(Fig. 9). For the Koolen Lake orthogneiss (sample G31 from Amato et al., 2014), results range +7.6 ± 1.4 to +11.5 ± 1.0 (mean value of +9.6 ± 2.4, n=10)(Fig. 9). Thus, the least and most radiogenic ratios are found in the oldest and youngest zircon populations respectively (Fig. 9). In aggregate, these data demonstrate Hf isotopic composition of these central AAC magmas increasingly trend towards depleted mantle-like signatures (Vervoort and Bilchert-Toft, 1999) over 125 Myr in Neoproterozoic time (Fig 9).

DISCUSSION

Crustal inheritance patterns in AAC magmatic rocks

Crustal inheritance in granitic magmas, and how this inheritance changes through time, can be determined by zircon geochronology and isotope geochemistry. Results from this study indicate several distinctive styles of inheritance. The inheritance that occurred early in the time span of magmatism [703±5 Ma granodiorite sampled from the Wrangel Complex (sample ELM06WR29)] are likely detrital zircons from supracrustal strata that contaminated the magma during emplacement. 80 Myr later, recycling of igneous (rather than metasedimentary) basement in 620±6 Ma potassium feldspar bearing augen gneiss from the Wrangel Complex (sample VP06-35b) is indicated by inheritance of a 711±4 Ma granitic xenolith (VP06-35a)(Fig. 4a) and xenocrystic zircons that yield mostly concordant ages between 680-710 Ma (Fig. 5). In far eastern Chukotka, Amato et al. (2014) documented two Neoproterozoic episodes of crustal recycling: Leucosome generation during partial melting of the Neoproterozoic Koolen Lake paragneiss basement at 666 ± 5 Ma (zircon U-Pb) and rare ca. 1.7 Ga zircons in the 574 ± 9 Ma Koolen Lake orthogneiss. 612±7 Ma zircons in undeformed leucogranite sample 11EGC36 are evidence of a much younger episode of basement recycling during mid-Cretaceous peak metamorphism and migmatite generation in the Velitkenay massif. Gneiss enclaves in Cretaceous leucogranite are further evidence of incomplete assimilation of basement (Fig. 4d) In sum, these observations provide evidence for multiple, and temporally discrete episodes of reworking of crustal and crustal-derived material in the Arctic Chukotka basement during Neoproterozoic magmatism, followed by an episode of remobilization of these older rocks during a mid-Cretaceous tectonothermal event 0.5 Gyr later.

Correlation of Wrangel inheritance data across AAC

The age spectrum of inherited zircons in sample ELM06WR29 (Fig. 7) is most simply interpreted as a detrital zircon (DZ) age spectrum of assimilated sedimentary rocks. This allows DZ geochronology-based paleogeographic correlations of Wrangel Complex metasedimentary basement to other parts of the AAC (Fig 10). Pervasively deformed metamorphic rocks of the Nome Complex on the Seward Peninsula (Till et al., 2014a,b) and central Brooks Range Schist Belt strata (Hoiland et al., this volume) also contain predominantly 0.9 to 2.1 Ga DZ with a Late Mesoproterozoic age peak (Fig. 10).

On Seward Peninsula (Fig. 1), this predominantly Mesoproterozoic age spectrum [classified as “Mesoproterozoic theme” strata by Till et al. (2014a,b)] has been reported in four samples from two different map units that are interpreted as Paleozoic age (as young as Devonian), despite the absence of Paleozoic age zircons in any of those samples (Till et al., 2014b). Age assignments are based on fossil data from surrounding rocks and stratigraphic interpretations about the deformed sequences (Till et al. 2014b).

Three of the four samples are assigned to the calcareous metasiliceous unit (Dcs, Fig. 10) and are interleaved with other metasedimentary rocks that contain Middle Devonian age zircons (Till et al., 2014a,b). They report Dcs “Mesoproterozoic theme” samples (Fig. 10) were deposited in a restricted sub-basin associated with the formation of the Aurora Creek zinc deposit. An Early Devonian maximum age is inferred from the youngest detrital zircons in strata that host the deposit (Till et al., 2014b), although none of the Dcs “Mesoproterozoic theme” samples have Paleozoic zircons (Fig. 10). In their model, “Mesoproterozoic theme” Dcs strata were locally derived from associated fault scarps that recycled basement strata into the sub-basin during Paleozoic time, yet contain no Paleozoic zircons.

Strata with the “Mesoproterozoic theme” are also observed in another Nome Group unit (DOx, “Mixed Unit”), which exhibits a broad range of protolith ages with uncertain stratigraphic relationships to one another (Till et al., 2014a,b). That “Mesoproterozoic theme” sample comes from a white metaquartzite layer in a marble-rich, structurally lower part the mixed unit (DOx) that consists of Ordovician through Devonian or younger strata (Till et al., 2014a,b).

In the central Brooks Range, two Schist Belt samples collected along the John River (near Ernie Lake, Fig. 1) also yield predominantly “Mesoproterozoic theme” spectra (Fig. 10) and are structurally interlayered with dated or inferred Devonian age rocks (Hoiland et al., *this volume*). Like the Nome Group strata on Seward Peninsula, constraining the depositional age of these Schist Belt strata based on stratigraphic relationships is complicated by deformation (Hoiland et al., *this volume*). Hoiland et al. (*this volume*) interpret them as more likely Neoproterozoic age than Devonian, or at least as having been recycled from local Neoproterozoic basement.

The similar age spectrum of inherited zircon in 703 ± 5 Ma granodiorite from Wrangel Island (Fig. 10), and the absence of DZ ages < 700 Ma in Nome Group and Schist Belt “Mesoproterozoic theme” strata (Fig. 10) suggest either the interpretation of Paleozoic depositional ages is incorrect or these strata are recycled from Neoproterozoic strata in the local basement. The first case is plausible if these “Mesoproterozoic theme-bearing” samples are Neoproterozoic, not Devonian, but are structurally juxtaposed with Paleozoic rocks containing Devonian DZ. In the western Brooks Range (Fig. 1), structural geometries supporting this interpretation have been documented in at least two localities. Out-of-sequence structural juxtaposition of Proterozoic and Paleozoic strata is exposed in the western Brooks Range at Mount Angayukaqsaq (Fig. 1) (Till et al., 1988). Nearby, 705 ± 35 Ma granite and its Proterozoic country rock comprise the structurally lowest component of the (Brookian) Schist Belt in the Kallarichuk Hills (Fig. 1), interpreted as stratigraphic basement exposed in structural windows and lateral ramps, and as out-of-sequence thrusts (Karl and Aleinikoff, 1989). Till et al. (2014b) also describe Neoproterozoic and suspected Neoproterozoic metasedimentary rocks in the southwestern Brooks Range that “...yielded detrital zircon much like the Mesoproterozoic theme (Till and Dumoulin, unpublished data).” In the alternative case, if Seward Peninsula (and/or Schist Belt) units are indeed Paleozoic in age as

suggested by Till et al. (2014a,b), their detrital zircon signatures indicate they were recycled from strata that correlate to the Wrangel inheritance. In either case, the documentation of these correlative age spectra in several locations across the AAC represents the first evidence of a regionally extensive stratigraphic interval of inferred Neoproterozoic age (Fig. 10).

Circum-Arctic correlation of AAC Neoproterozoic strata

Correlation of AAC crust to elsewhere in the circum-Arctic region is complicated by the inaccessibility of the continental shelves (e.g., Barents Shelf, Marengo et al., 2013; Chukchi Shelf, Klemperer et al., 2002; Northwind Ridge, Grantz et al., 1998; Chukchi Borderland, Brumley et al., 2014). Despite this, several new insights are available from the Wrangel inheritance data and contextually linked Schist Belt and Nome Group “Mesoproterozoic theme” data. Across the circum-Arctic region, Neoproterozoic strata with DZ ages similar to Wrangel inheritance and the “Mesoproterozoic theme” are common, but in contrast to these AAC strata, many of those strata exhibit age peaks older than 1.6 Ga (e.g., Fig. 10 in Malone et al., 2014; Fig. 3 in Zhang et al., 2015). Provenance comparisons between the AAC and Pearya terrane (Malone et al., 2014) establish key similarities and contrasts in Neoproterozoic stratigraphy on opposite flanks of the Amerasia Basin (Figs. 1 and 10).

The depositional age range proposed for this Neoproterozoic AAC sequence is broadly coeval to a ca. 1030–710 Ma period of sedimentation and orogeny (Valhalla orogen) that has been documented in numerous Arctic and North Atlantic Caledonide terranes (e.g., Cawood et al., 2010). The AAC age spectra are remarkably similar to strata from the Pearya terrane (Malone et al., 2014) (Fig. 10), which have been correlated as part of the Valhalla orogen (Cawood et al., 2016). The Grenville–Sveconorwegian orogenic belts that formed on the margins of Baltica and Laurentia during the assembly of Rodinia have been suggested as provenance of those Pearya strata and similar DZ signatures in the Valhalla orogen (Fig. 11) (Kirkland et al., 2007; Malone et al., 2014).

Trace element data from inherited detrital zircons in the 703±5 Ma granodiorite sample provide independent evidence that AAC strata have some orogenic belt provenance. Increasing Hf and Hf/Yb, coupled with and decreasing Th/U observed from 1.3 to 1.0 Ga (Fig. 7c) indicate progressively younger zircons in the inherited population formed from more evolved/fractionated magmas relative to the older magmas (e.g., Barth and Wooden, 2010; Claiborne et al., 2010). The detrital zircon geochemical trends record a prolonged interval characterized by an increase in “crustal recycling” signatures (Fig. 7c), consistent with Late Mesoproterozoic assembly of Rodinia (e.g., Spencer et al., 2015), further strengthening ties between the AAC and these sources (Fig. 11).

The 975±32 Ma youngest zircon age population in our data overlaps the timing of magmatism that has been dated in the Brooks Range (Amato et al., 2014), Pearya

(Malone, 2012; Malone et al., 2014), the Northwest terrane of Svalbard (Pettersson et al., 2009), the Farewell terrane of central Alaska (Bradley et al., 2014), the Kalak Nappe Complex in Finnmark (Kirkland et al., 2006), and Central Taimyr (Vernikovskiy et al., 2011)(Fig. 1). Malone et al. (2014), building on work of Cawood et al. (2010) and Kirkland et al. (2011) suggested a paleogeographic model in which arc magmatism along the Rodinian margin in the time-span 980 to 920 Ma affected parts of the margins of Siberia, Laurentia and Baltica (Fig. 11). The few 0.9–1.0 Ga zircon ages in AAC strata suggest a linkage to these magmatic sources (whether by primary proximity or sediment recycling), further strengthening an association with the Valhalla orogen. However, in the Pearya Terrane, the stratigraphic interval that bears remarkably similar spectra (Succession IIA of Malone et al., 2014) also contains strata dominated by a ca. 970 Ma age peak (Fig. 10) (Succession IIB of Malone et al., 2014), as do numerous other Valhalla orogen strata (Cawood et al., 2007; Bingen et al., 2011). At present, no strata that also exhibit a primary age peak at 970 Ma have been documented in the AAC. This is a distinct difference between the AAC and the stratigraphic records in these other locales.

Neoproterozoic magmatism in central AAC: A ribbon continent setting?

The integrated U–Pb, oxygen, and hafnium isotopic data from 710–580 Ma igneous zircons studied across the AAC suggest that the magmas generated during late Neoproterozoic time reflect consistent, volumetrically important additions of isotopically juvenile (i.e., primitive mantle–derived) material to a more evolved crustal infrastructure during this interval.

One of the oldest samples from this study, the 703±5 Ma inheritance-rich granodiorite on Wrangel Island, preserves numerous detrital zircons as cores (Fig. 7), signifying that early on there were supracrustal assimilants incorporated in the magmas. Field relationships on Wrangel Island suggest assimilation of crustal material as late as 620±6 Ma (Fig. 4b), as does the presence of 1.7 Ga zircons in the 574±9 Ma Koolen Lake orthogneiss and 666±5 Ma leucosomes in Koolen Lake paragneiss in far eastern Chukotka (Amato et al., 2014).

The three samples of AAC basement we analyzed for oxygen isotopes, ranging in age from 661±11 Ma to 574±9 Ma, exhibit mantle–like $\delta^{18}\text{O}$ results (Fig. 8). When compared to the global record of oxygen isotopes in zircon since 2 Ga, the results from this study are at the light end of the spectrum, consistent with limited recycling of high $\delta^{18}\text{O}$ sediment (Figs. 8, 11)(Valley, 2003). Thus the oxygen isotope results support 661±11 Ma and younger magmatism in a mantle–input dominated system (i.e., low volume of assimilated material), or mixing of mantle–derived magmas with crustal materials that had mantle–like oxygen isotope ratios.

The hafnium isotopic compositions of the four samples analyzed exhibit a near vertical (i.e., becoming increasingly positive) trend over the time span of magmatism towards more radiogenic values (Fig. 9), also suggesting an

environment with decreasing input of evolved material during magmatism (Fig. 11). Increasingly radiogenic Hf isotopic compositions of arc crust may indicate generation of isotopically juvenile mantle-derived melts in a continental arc setting undergoing extension (Mišković and Schaltegger, 2008), subduction-driven removal of arc lithosphere (Collins et al. 2011), and/or isolation from evolved sedimentary inputs due to slab retreat and consequent oceanward arc migration (Collins, 2002; Collins and Richards, 2008). From a global plate tectonic setting perspective, this trend is characteristic of “peripheral” orogenic systems (Murphy and Nance, 1991), such as circum-Pacific continental margins since the Mesozoic (Collins et al., 2011). As a well-documented example, the geochemical and tectonic evolution of the Tasmanides reflects multiple episodes of juvenile crust formation during the Phanerozoic in a retreating arc tectonic setting (Kemp et al., 2009; Collins et al., 2011).

Given these similarities, an extensional arc tectonic setting for arc magmatism across the interval 710- 580 Ma is supported by O and Hf isotope geochemistry of zircon and patterns of inheritance over the duration of Neoproterozoic magmatism in the central AAC (Fig. 11). The increasingly juvenile contributions to the bulk composition of the crust suggests that the AAC might have formed as part of a ribbon continent that separated from a larger continental landmass by extension/spreading in an oceanic backarc region (Fig. 11)(Malone et al., 2014). Over the long duration of Neoproterozoic magmatism in the AAC, the geochemical signals of relatively primitive (i.e., mantle-derived, including accreted oceanic crust) versus relatively evolved (i.e., continent-derived) components become progressively more primitive, best exemplified by the steep vertical trend in $\epsilon\text{Hf}_{\text{initial}}$ (Fig. 9). The mantle-like isotopic values in zircons as old as 661 ± 11 Ma suggest relatively limited volumes of high $\delta^{18}\text{O}$ crust available for assimilation for most of the duration of magmatism, which may indicate oceanward arc migration occurred early in the time span of magmatism.

Implications

Although it is beyond the intended scope of this paper (establishing piercing points for Mesozoic Arctic plate tectonic reconstructions), these new data also provide insight into Rodinian continental reconstructions. Most importantly, the “peripheral/external fingerprint” of Cryogenian–Ediacaran magmatism underscores the need to understand the paleogeographic implications of inheritance in the 703 ± 5 Ma granodiorite and correlative spectra elsewhere in the AAC. Understanding the derivation and crustal sources of those spectra, in conjunction with more comprehensive geochemical and petrological studies of magmatism in Neoproterozoic basement of the AAC, could better specify the AAC’s position along the margin of Rodinia, and elucidate the paleotectonic relationships between the AAC and the Timanian and Caledonian orogenic belts.

CONCLUSIONS

This paper provides new insight into the Neoproterozoic paleotectonic evolution of the central part of the Arctic Alaska Chukotka microplate, for which little data on its basement rock units previously existed. These findings and associated interpretations allow more robust comparisons with displaced terranes and long-lived continental margins in the circum-Arctic region.

(1) Although Cretaceous high-grade metamorphism and deformation had obscured the original lithology of basement in the Velitkenay massif of Arctic Chukotka, integrating zircon U–Pb geochronology, trace element geochemistry and O and Hf isotopic results suggest the area was a locus of magmatism at 661 ± 11 Ma and 612 ± 7 Ma.

(2) The 661 ± 11 Ma and 612 ± 7 Ma ages from Arctic Chukotka are similar to previously published TIMS ages (Ceclie et al., 1991). Our newly published SIMS ages for magmatism in the Wrangel Complex on Wrangel Island suggest a shared Neoproterozoic history of these basement complexes.

(3) Inheritance of pre-magmatic zircon is observed at ca. 700–580 Ma and again at ca. 100 Ma. This occurs as the incorporation of supracrustal (i.e. detrital) zircons in a 703 ± 5 Ma granodiorite intrusions, as 711 ± 4 Ma granitic xenoliths in 620 ± 6 Ma intrusions, and as the inheritance of 612 ± 7 Ma zircons in ca. 100 Ma leucogranites associated with peak metamorphism in the Velitkenay complex.

(4) The age spectrum of supracrustal zircon inheritance in 703 ± 5 Ma granodiorite from Wrangel Island could have been derived from end Mesoproterozoic orogen source regions of northern Rodinia (e.g., Grenville–Sveconorwegian belts). These spectra have also been observed in the Brooks Range (Hoiland et al., this volume) and rocks mapped as Paleozoic in the Nome Complex on Seward Peninsula (Till et al., 2014a; 2014b). Correlation of this DZ signature represents the first evidence of a regionally extensive Neoproterozoic age stratigraphic interval across the AAC.

(5) Ca. 710–580 Ma O–Hf isotopic evolution of central AAC crust indicates a temporal trend toward more primitive magmatic character, consistent with the AAC basement being formed and modified in an “external” or “peripheral” orogenic setting (Fig. 11)(Collins et al., 2011; Cawood et al., 2016).

FIGURE CAPTIONS

Figure 1. **(a)** Circum-Arctic topography and bathymetry in polar projected image (Jakobsson et al., 2012) showing the location of the Arctic Alaska Chukotka terrane (AAC), delineated along its northern flank by the shelf to slope break (thin dashes) and across continental and marine shelves by geological and geophysical constraints discussed in text (thick dashes). Dashed white polygon shows location of Figure 2. AAC regions: BR, Brooks Range; K, Koolen dome; NSI, New Siberian Islands; SP, Seward Peninsula; V, Velitkenay massif; W, Wrangel Island; ZH, Zhokov Island of DeLong archipelago. Continental shelf areas: Be, Bering Sea Shelf; Bf, Beaufort Sea Shelf; Ch, Chukchi Sea Shelf. Numbered locations in Brooks Range: 1, Ernie Lake; 2, Mount Angayukaqraq; 3, Kallarichuk Hills. Other regions referred to in text: F, Farewell Terrane; KNC, Kalak Nappe Complex; K-O, Kolyma-Omolon block; P, Pearya Terrane; SAZ, South Anyui zone (cross hatched area, Amato et al., 2015); SV, Svalbard; T, Central Taimyr. Italicized labels indicate areas with geochronology data relevant to parts of this study. **(b)** Age range of magmatism ($\pm 2\sigma$ error bars) based on zircon U-Pb reported in various parts of the AAC, compiled from Amato et al. (2014), Akinin et al. (2015) and results from this study. Shaded region in inset highlights age range of results presented in this manuscript.

Figure 2. Regional study area map of central Chukotka and Wrangel Island from Miller and Verzhbitsky (2009), highlighting the relatively limited aerial distribution of pre-Mesozoic outcrop and pervasive distribution of Early to mid-Cretaceous plutonic and Late Cretaceous Okhotsk-Chukotka Volcanic Belt rocks across Arctic Chukotka, that are absent from Wrangel Island. Ages of plutons in Velitkenay massif from Akinin et al. (2012).

Figure 3. Geological maps and generalized stratigraphic columns showing sample locations for each study area. **(a)** Kichshnikov River area of Wrangel Island (simplified from Miller et al., 2010); **(b)** Velitkenay massif in Arctic Chukotka based on previously unpublished mapping from 2011 field season.

Figure 4. Field photos and thin section images (cross-polarized light) of selected samples used in this study. **(a)** Augen gneiss sample VP06-35b with granitic enclave (outlined) VP06-35a from Wrangel Complex; **(b)** Granitic cobble VP06-36b (outlined) in Devonian (?) conglomerate that sit unconformably on the Wrangel Complex; **(c)** Meta-volcanic sample VP06-36a from Wrangel Complex; **(d)** Cretaceous leucogranite sample 11EGC36 with gneissose basement enclaves exposed in Velitkenay complex.

Figure 5. SIMS geochronology plots of crystallization ages for samples from Wrangel Island. First column shows Tera-Waserberg style concordia results, plotted using *Isoplot 3.75* (Ludwig, 2012) with error ellipses shown at 68.3% (1σ) confidence. Second column shows density distribution mapping of data (Sircombe, 2007) for data from each sample shown in first column, illustrating concordance of

inheritance (e.g., sample VP06–35b) Contour intervals on density distributions at 67% and 95% confidence. Inset boxes for samples VP06–35b, VP06–36b, and ELM06WR29 in first column show area of density distribution plot. Third column shows ^{207}Pb corrected $^{206}\text{Pb}/^{238}\text{U}$ weighted average ages calculated using *Isoplot* 3.75 (Ludwig, 2012), with individual data point error bars shown at 2σ . Results shown for sample ELM06WR29 are from analysis of zircons on mount ELM4 (whereas data shown in Fig. 7 is from mounts ELM4 and ESG24). Asterisk (*) designates ^{204}Pb –based correction applied to isotopic ratios. MSWD, Mean Square of Weighted Deviates.

Figure 6. SIMS zircon geochronology and trace element geochemistry results from Velitkenay massif sample, Arctic Chukotka. **(a–c)** Biotite bearing gneiss sample 11EGC21; **(a)** Conventional concordia plot of ^{204}Pb corrected results, created with *Isoplot* 3.75 (Ludwig, 2012); **(b)** Chondrite normalized rare earth element concentrations of individual zircon analyses, showing an overall consistent and typical igneous pattern (e.g., Hoskin and Schaltegger, 2003) suggesting an orthogneiss classification is appropriate for this sample; **(c)** Tera–Waserberg plot of data using fixed lower intercept based on the timing of peak metamorphism in the Velitkenay massif; **(d–f)** Undeformed leucogranite sample 11EGC36 with Neoproterozoic age inherited zircons; **(d)** Chondrite normalized rare earth element concentrations of individual zircon analyses, showing a similar pattern to sample 11EGC21, also suggesting these zircons were derived from an igneous protolith; **(e)** Tera–Waserberg concordia plot of uncorrected for common Pb results created with *Isoplot* 3.75 (Ludwig, 2012), showing calculated concordia and upper intercept results of inheritance in this sample; **(f)** ^{207}Pb corrected $^{206}\text{Pb}/^{238}\text{U}$ ages, showing weighted average result for inheritance that exhibits some indication of Pb loss in the younger part of the population. MSWD, Mean Square of Weighted Deviates.

Figure 7. SIMS (SHRIMP-RG) inheritance results from granodiorite sample ELM06WR29 from Wrangel Island. **(a)** ^{204}Pb corrected Tera–Waserberg concordia plot of inheritance data plotted with *Isoplot* 3.75 (Ludwig, 2012), symbolized according to discordance. Error crosses are 1σ . Cathode luminescence images of selected inherited zircon domains from this sample that yielded concordant results. Dotted circle shows sputter pit location (25 μm diameter). Age uncertainty shown at 2σ ; **(b)** Relative probability and histogram plot of ^{204}Pb corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages of inherited domains in zircons, created with *Isoplot* 3.75 (Ludwig, 2012). Filled relative probability curve and histogram shows all data and outline shows relative probability curve for concordance–filtered data; **(c)** Plots of trace element data from inherited domains ($n=105$) versus ^{204}Pb corrected $^{207}\text{Pb}/^{206}\text{Pb}$ age results. Ellipses plot trace element concentration or ratio versus best estimate of measured age (e.g. 1223 ± 19 Ma plotted at 1223 Ma). Diamond symbols and error bars plot arithmetic mean and standard deviation of ellipse data in 100 Myr age bins, starting at 925 ± 50 Ma (see text). Horizontal line and gray shaded region plots arithmetic mean and standard deviation of syn–magmatic results from the sample.

Figure 8. Oxygen isotope results plotted as $\delta^{18}\text{O}$ versus protolith crystallization age from Velitkenay samples and Koolen orthogneiss sample G31 (Amato et al., 2014). Mantle zircon range ($\pm 2\sigma$) from Valley (2003) shown for comparison.

Figure 9. Hf isotope results plotted as $\epsilon\text{Hf}_{\text{initial}}$ versus protolith age for Wrangel sample ELM06WR29, both Velitkenay samples and Koolen orthogneiss sample G31. Depleted mantle $\epsilon\text{Hf}_{\text{initial}}$ trajectory calculated using isotope ratios reported in Vervoort and Patchett (1996). CHUR, Chondritic uniform reservoir.

Figure 10. Normalized probability plot (Gehrels, 2010a; top) and cumulative probability plot (Gehrels, 2010b; bottom) of inheritance ages in Wrangel Complex sample ELM06WR29, compared to correlative AAC strata and similar age strata from Pearya terrane. Grey shaded region in normalized probability plot is 980–920 Ma. Sources of detrital zircon data: Nome Group, Till et al. (2014b); Pearya, Malone et al. (2014); Brooks Range Schist Belt, samples 13-CH-BR03b and 13-CH-BR04/A from Hoiland et al. (*this volume*).

Figure 11. Hypothetical tectonic evolution of the AAC during Neoproterozoic time based on data from this study and references in text. Arrows above surface show direction of inferred sediment transport. **(a)** Delivery of sediments with Grenville–Sveconorwegian sources to Rodinian margin; **(b)** Development of Tonian arc magmatism along Rodinian margin, inboard of Arctic Chukotka/Wrangel Island crust; **(c)** Establishment of Cryogenian arc on Arctic Chukotka/Wrangel Island crust, outboard of Tonian arc trend; **(d)** Arc built on rifted fragment of Rodinian margin, isolated from craton-sourced sediments by back-arc basin.

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Table 1. Zircon geochronology and isotope geochemistry

Sample	Rock Type	Location	Lat	Long	Age (Ma)	±2s (Ma)	Age type	MSWD	d ¹⁸ O (‰)	±2SD (‰)	eHf _{initial}
Plutonic											
VP06-35a	granitic xenolith	Wrangel Island	71.0934	-179.2454	711.4	4.2	²⁰⁶ Pb*/ ²³⁸ U	1.1	---	---	---
ELM06WR29†	deformed granodiorite	Wrangel Island	71.1148	-179.3972	703.4	5.0	²⁰⁶ Pb*/ ²³⁸ U	1.1	---	---	-3.6
ELM06WR28e	deformed granodiorite	Wrangel Island	71.1112	-179.3877	697.3	5.0	²⁰⁶ Pb*/ ²³⁸ U	1.7	---	---	---
VP06-36b	granitic clast in Paleozoic cgl	Wrangel Island	71.1020	-179.2440	673.3	4.2	²⁰⁶ Pb*/ ²³⁸ U	1.1	---	---	---
VP06-35b	augen gneiss	Wrangel Island	71.0934	-179.2454	619.8	6.2	²⁰⁶ Pb*/ ²³⁸ U	2.5	---	---	---
Volcanic											
VP06-36a	Felsic meta-volcanic rock	Wrangel Island	71.1020	-179.2440	702.0	3.7	²⁰⁶ Pb*/ ²³⁸ U	1.3	---	---	---
Indeterminate (igneous) protolith											
11EGC21	biotite gneiss	Velitkenay	69.2307	177.2225	661	11	fixed lower int.	1.3	5.87	1.32	4.2
11EGC36	Inheritance in leucogranite	Velitkenay	69.2831	176.9151	612.3	7.3	²⁰⁶ Pb*/ ²³⁸ U	0.9	4.85	0.50	7.7
G31‡	orthogneiss	Koolen Dome	65.9472	-171.1667	574	9	lower int.	1.8	6.02	0.29	9.6

cgl., conglomerate; int., concordia intercept

WGS84 datum for geographical locations

* ²⁰⁷Pb corrected

† Hf mean excludes 4 outlier datapoints shown in Figure 9

‡ Age data from Amato et al. (2014)

±2SD

1.5

1.3

2.7

2.4