Detrital zircon U-Pb geochronology and Hf isotope geochemistry of metasedimentary strata in the southern Brooks Range: Constraints on Neoproterozoic–Cretaceous evolution of Arctic Alaska

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ABSTRACT: Mid-Palaeozoic assembly models for the Arctic Alaska-Chukotka microplate (AACM) predict the presence of cryptic crustal sutures whose exact locations and deformational histories have not been identified in the field. This study presents data on the provenance of poly-deformed and metamorphosed strata in the southern Brooks Range “Schist Belt” of presumed Proterozoic-Devonian depositional age as well as for structurally overlying strata to help elucidate terrane boundaries within the AACM and add new constraints to the palaeogeographic evolution of its constituent parts. Protoliths identified within the Schist Belt support correlations with metasedimentary strata in the Ruby terrane and Seward Peninsula and suggest a (peri-) Baltican origin in late Neoproterozoic–early Palaeozoic time. Proximity to Laurentia is only evident in what are inferred to be post-early-Devonian age strata. In contrast, the North Slope and Apoon terranes originated proximal to Laurentia. The mid-Palaeozoic boundary between these (peri-) Baltic and (peri-) Laurentian terranes once lay between rocks of the Schist/Central Belt and those of the Apoon terrane but is today obscured by severe Mesozoic-Cenozoic deformation. Whether this boundary represents a convergent or transform suture, when exactly it formed, and how it relates to broader Caledonian convergence in the North Atlantic are still unresolved questions.

Supplementary material: Details of the analytical methods and data tables are available at <http://geolsoc.figshare.com/SUP0000>.

The geology of the Arctic Alaska–Chukotka microplate (AACM) is key to refining tectonic reconstructions of the Arctic (e.g. Shephard *et al.* 2013; Miller *et al.* this volume). In particular, current efforts to constrain the timing and kinematics of the Mesozoic-Cenozoic (?) opening of the Amerasia Basin (e.g. Gottlieb *et al.* 2014; Lane *et al.* 2016; Houseknecht & Connor 2016) and also the Palaeozoic-Mesozoic (?) displacement history of “suspect” terranes within the North American Cordillera (e.g. Colpron & Nelson 2009, 2011) require consideration of the pre-Mesozoic geology of the AACM as “piercing points” to help test such models (e.g. Miller *et al.* 2010; Strauss *et al.* 2013). Unfortunately, long-lived and still active Pacific margin deformational belts and

the rift opening of the Amerasia and Eurasia Basins have overprinted the older geology of the AACM, making it difficult to identify the extent and continuation of these Pre-Mesozoic tectonic elements (e.g. Pease 2011)(Fig. 1A-1B). Due to this internal complexity, its vast area, and remote location the geological history of the AACM and its constituent terranes remain poorly constrained, limiting efforts to resolve the earlier tectonic history of the Arctic.

The AACM – which includes the Brooks Range and North Slope of Alaska, the Chukotka Peninsula and Wrangel Island of Arctic Russia, and the adjacent continental shelves of the Beaufort and Chukchi Seas (Fig. 1B) (Miller *et al.* 2006; Amato *et al.* 2009, 2014) – is a composite terrane, formed by tectonic juxtaposition of several previously independent crustal fragments in mid-Palaeozoic time (e.g. Moore *et al.* 1994; Amato *et al.* 2009; Miller *et al.* 2011; Strauss *et al.* 2013, 2017). As in Moore *et al.* (2015) we refer to the entire conjoined block as the “Arctic Alaska–Chukotka microplate” (Fig. 1B), using simply “Arctic Alaska” or “Chukotka” when referring to specific regions of the AACM; and we include the Ruby terrane (Fig. 1C) as belonging to the AACM, despite it having undergone a somewhat divergent (though poorly constrained) displacement history beginning in the Cretaceous (e.g. Moore & Box 2016). During the Mesozoic-Cenozoic (?) opening of the Amerasia Basin, the composite AACM was displaced southward to its present day location (e.g. Grantz *et al.* 1979; Miller *et al.* 2006; Lawver *et al.* 2011) (Fig. 1B), but the timing and kinematics of this opening remain contested (e.g. Grantz *et al.* 2011; Mosher *et al.* 2012; Gottlieb *et al.* 2014; Lane *et al.* 2016; Houseknecht & Connors 2016; Miller *et al.* this volume). Many of these kinematic models require not only translation but also internal deformation of the AACM during Mesozoic-Cenozoic time (e.g. Amato *et al.* 2015; Miller *et al.* this volume; Till 2016), such that the exact definition and the boundaries of the AACM “microplate” or “terrane” are not clear-cut.

Of particular interest to Arctic reconstructions, is the recognition that the far northeastern portion of the AACM (“paraautochthon” of Fig. 1C or “North Slope subterrane” of Strauss *et al.* 2013) formed part of the Laurentian continent in early Palaeozoic time (e.g. Dutro *et al.* 1972; Moore *et al.* 1994; Strauss *et al.* 2013; Cox *et al.* 2015; McClelland *et al.* 2015; Lane *et al.* 2016; Johnson *et al.* 2016), whilst the Seward Peninsula and Chukotka portions of the AACM appear to have been exotic to Laurentia at that time and more likely lay adjacent to Baltica and/or peri-Siberian terranes (e.g. Patrick & McClelland 1995; Amato & Wright 1997; Amato *et al.* 2009; 2014; Till *et al.* 2014a, 2014b; Miller *et al.* 2006, 2010, 2011; Akinin *et al.* 2015; see also Natal’in *et al.* 1999; Blodgett *et al.* 2002; and Dumoulin *et al.* 2002, 2014). Such relations require that an important mid-Palaeozoic “suture” – potentially correlative with that of the Caledonides – exists somewhere within the Brooks Range or under its foreland basin (e.g. Churkin *et al.* 1985; Moore *et al.* 1985, 1992, 1994; Grantz *et al.* 1991; Dumoulin *et al.* 2002; Strauss *et al.* 2013; O’Brien *et al.* 2016). This suture has been suggested to lie just south and west of an enigmatic belt of rocks recording Cambro-Silurian arc volcanism in the Apoon assemblage of the Doonerak window in the Central Brooks Range (e.g. Strauss *et al.* 2017)(Fig. 2). Pre-mid-Palaeozoic strata in the Brooks Range that can be used to better define the location of this suture and its deformational history are scarce; found only in

belts of deformed schists, gneisses, semi-schists and phyllites that are now exposed along the southern flank of the Brooks Range (Fig. 2). Until recently, the protolith ages of these metamorphic rocks, their provenance, and palaeogeographic affinity remained for the most part unknown (e.g. Moore *et al.* 1994; Till *et al.* 2008).

Amongst these metamorphic belts with pre-mid-Palaeozoic stratigraphy is the Schist Belt (Coldfoot terrane of Moore *et al.* 1994), a zone of penetratively deformed high strain rocks along the southern flank of the Brooks Range, whose dominant fabric has been interpreted as related to top-to-the-south shear during regional extension (Christiansen & Snee 1994, Little *et al.* 1994)(Fig. 1C). In most places, the shear zone is structurally overlain and in normal fault contact to the south by a narrow zone of less-resistant, lower grade phyllite (belonging to the “Greywacke-Phyllite belt”) which is in turn structurally overlain by normal fault-bound exposures of the oceanic Angayucham terrane and/or Yukon-Koyukuk Basin (e.g. Carlson 1985; Box 1987; Patton *et al.* 1994, 2009; Christiansen & Snee 1994, Little *et al.* 1994, Oldow *et al.* 1987; Gottschalk & Oldow 1988; Moore *et al.* 1994, 1997a) (Fig. 2). Understanding the palaeogeographic significance of these packages of rocks requires peering beyond their Mesozoic-Cenozoic deformational strain and detrital zircon geochronology offers the most straightforward assessment of protolith and provenance and permits comparison to similar data sets produced elsewhere in the AACM (Amato *et al.* 2009; Strauss *et al.* 2013, 2017; Till *et al.* 2014a; Johnson *et al.* 2016; Gottlieb *et al.* this volume) and the circum-Arctic (e.g. Miller *et al.* 2011; Hadlari *et al.* 2012; Beranek *et al.* 2013a; Andresen *et al.* 2014; Zhang *et al.* 2015)

This study presents U–Pb geochronology and Hf-isotope compositions for detrital zircon from a variety of lithologies and localities within the belts of metasedimentary rock flanking the southern Brooks Range that demonstrate diverse depositional ages and sedimentary provenance for their protolith rocks. This new data set helps correlate the Schist Belt stratigraphy to that of the nearby Ruby and Seward Peninsula terranes, and further refines their early Palaeozoic palaeogeography as likely proximal to the North Atlantic region of the Arctic, receiving detritus of Baltican/Timianian origin. Ultimately, this study supports the hypothesis that the AACM formed by mid-Palaeozoic amalgamation of (peri-) Baltican and (peri-) Laurentian terranes whose boundary within the Brooks Range is suggested to lie somewhere the Central or Schist Belt and the Apoon terrane (Doonerak Window)(Fig. 2).

GEOLOGICAL SETTING

The Brooks Range

The Brooks Range is part of a more extensive belt of Jurassic to Early Cretaceous crustal shortening developed along the northern palaeo- Pacific margin of North America and NE Russia (Fig. 1A). Arc-continent collision in the late Jurassic-Cretaceous was accompanied by development of a northward-propagating fold-thrust belt, formation of a belt of high-pressure, low-temperature metamorphism, emplacement of ophiolite

fragments at highest structural levels, and significant retro-foreland basin subsidence (Roeder & Mull 1978; Mull 1982; Mayfield *et al.* 1988; Pallister *et al.* 1989; Till 1992; Wirth *et al.* 1993; Moore *et al.* 1994, 2015; Patrick 1995)(Fig. 1C). Widespread Aptian-Albian lithospheric extension post-dates convergent structures and structurally exhumed deep levels of the orogen, leading to the development of the Yukon-Koyukuk Basin (YKB) above a structurally thinned metamorphic hinterland (Miller & Hudson 1991; Little *et al.* 1994). The extensional structures related to this event and the sediments deposited in the YKB are in turn overprinted by younger structures related to renewed crustal shortening and strike-slip faulting in the Cenozoic (e.g. Fuis *et al.* 2008).

The Brooks Range consists of a series of structurally stacked lithologic packages (Fig. 1C): 1. (Para-) autochthonous strata of the northeastern Brooks Range, North Slope Basin, and Doonerak Window, which consist of late Palaeozoic-Mesozoic passive margin strata deposited across eroded pre-Mississippian basement rocks and overlain by Cretaceous to Neogene foreland basin sediments; 2. A thick section of Devonian to Mississippian clastic rocks comprising the Endicott Mountains allochthon (EMA); 3. Distal late Palaeozoic-Mesozoic continental margin strata, stacked in a series of rootless allochthons; and 4. The Angayucham terrane which consists of oceanic and ultramafic rocks interpreted to be part of the island arc and intervening ocean floor accreted to Arctic Alaska during Brookian orogenesis (Roeder and Mull, 1978; Pallister *et al.* 1989; Harris, 1995; Patton *et al.* 1994). Deeper structural levels are exposed in the hinterland which have been divided on the basis of strain, metamorphism, and lithology into (from North to South) the Central, Schist, and Greywacke-Phyllite Belts (also known as the Hammond, Coldfoot, and Slate Creek terranes, respectively, of Moore *et al.* 1994, 1997a)(Fig. 1C, Fig. 2), but their pre-deformational positions with respect to each other and to allochthonous and parautochthonous rocks to the north are less clear (e.g. Moore *et al.* 1997b; Till 2016).

Late Palaeozoic-Mesozoic strata within the rootless allochthons of the Brooks Range (i.e. “low grade allochthons” of Fig. 1C and “Endicott Mountains allochthon” of Fig. 2) constitute the sedimentary cover that was stripped off an unknown and possibly now missing basement complex during the Early Cretaceous Brookian orogeny (e.g. Moore *et al.* 1997a). Strata deposited before the mid-Palaeozoic formation of the AACM – possibly having once constituted the missing basement of these allochthonous strata – are thought to be present in the Schist and Central Belts (Dillon *et al.* 1980; Till *et al.* 2008; Amato *et al.* 2014). Although they now comprise separate structural blocks and have previously been referred to as separate terranes (e.g. Moore *et al.* 1994), the Central Belt is thought to represent the stratigraphically lower and possibly more hindward part of the Endicott Mountains allochthon (EMA)(e.g. Moore *et al.* 1997a; Till *et al.* 2008). Where exactly the Schist Belt restores along this former margin with respect to the Central Belt and EMA is less clear (e.g. Vogl *et al.* 2002; Till 2016), but in the simplest scenario the Schist Belt may represent the southernmost contiguous portion of the Central Belt which experienced more extreme penetrative strain during late Cretaceous extension (e.g. Little *et al.* 1994; Law *et al.* 1994; Christiansen & Snee 1994).

The metamorphic hinterland of the Brooks Range orogen is thought to have extended

southward into the Seward Peninsula and the Ruby terrane, which have strata of similar lithology, age, and metamorphic history to those in the southern Brooks Range (e.g. Patton *et al.* 1994; Roeske *et al.* 1995; Amato *et al.* 2009, 2014; Till *et al.* 2014a, 2014b)(Fig. 1C). Similarities of the Schist Belt to the Ruby terrane include the recognition in both of early to middle Palaeozoic carbonate rocks, Devonian orthogneisses, mafic metavolcanic rocks, blueschist-greenschist metamorphic assemblages, Cretaceous K-Ar ages of metamorphic minerals, and the structural position of each terrane with respect to an overthrust oceanic terrane (Patton *et al.* 1994). However, important differences are recognized. For instance, mid-Palaeozoic felsic metavolcanic rocks are locally abundant in the southern Brooks Range but have not been observed in the Ruby terrane (Patton *et al.* 1994). Similarly, mid-Cretaceous granites are abundant in the Ruby terrane but are nowhere recognized in the Brooks Range (Patton *et al.* 1994). Blueschist facies units in the Seward Peninsula have also been correlated to those in the southern Brooks Range, suggesting both regions shared not only their Mesozoic metamorphic history but also their Palaeozoic depositional history (Fig. 1C)(e.g. Till *et al.* 2014a, 2014b and discussions therein).

Schist Belt

The Schist Belt of the south-central Brooks Range, where the majority of the samples for this study were collected, is a zone of high-strain, penetratively deformed rocks up to 25 km wide that extends for 600 km along strike of the southern flank of the Brooks Range (Till *et al.* 2008) (Fig. 2). Schist Belt units are polymetamorphosed and contain well-developed foliations, one of which formed during high-pressure metamorphism and another which overprints it at greenschist-facies conditions (Dillon *et al.* 1986; Vogl 2003; Till *et al.* 2008 and references therein). The most detailed studies to date of the structural fabrics of the Schist Belt (Little *et al.* 1994; Law *et al.* 1994) found that the younger overprinted fabric (i.e. the greenschist facies event) was the most pervasive, forming the dominant foliation, S_D , in the Schist Belt. This dominant foliation is associated with N-S-stretching and top-down-to-the-south extensional shear (Little *et al.* 1994; Law *et al.* 1994; Christiansen & Snee 1994). Elsewhere, however, the dominant foliation in the Schist Belt has been interpreted as coeval with high-pressure metamorphism and thrusting (Till & Snee 1995; Till *et al.* 1988; Gottschalk & Oldow 1988; Till *et al.* 2008). This interpretation is challenged by the relative age relationships between older fabrics and metamorphism and the overprinting greenschist-facies dominant foliation in the Schist Belt (Little *et al.* 1994; Law *et al.* 1994; Christiansen 1989; Christiansen & Snee 1994; Dinklage 1998).

The northern boundary of the Schist Belt (i.e. contact with the Central Belt) has been interpreted by some workers as a zone of thrust faulting (e.g. Patrick *et al.* 1994; Moore *et al.* 1997a; Till & Snee 1995; Till *et al.* 2008; Till 2016) and by others as a strain gradient where the younger dominant foliation formed during top-to-the-south extension dies out (Little *et al.* 1994; Law *et al.* 1994; Christiansen & Snee 1994), but it may be a combination of both of these features. The lesser carbonate units and more abundant relict blueschist facies mineral assemblages in the Schist Belt (e.g. Dusel-Bacon *et al.* 1989, Till *et al.* 2008) help distinguish it from the Central Belt to the north, as does the

complete transposition in the Schist Belt of depositional features that are relict in the Central Belt (Moore *et al.* 1994; Till *et al.* 2008) (Fig. 2). However, the exact boundary between these two belts has not been mapped in sufficient detail along much of the range to definitively characterize the nature of the contact between the Schist Belt and Central Belt. The southern boundary of the Schist Belt is a series of top-to-the-south normal sense shear zones that excise section and juxtapose rocks of the lower grade Greywacke-Phyllite Belt structurally above those of the Schist Belt (Little *et al.* 1994; Moore *et al.* 1997a; Till *et al.* 2008).

The dominant lithologies in the Schist Belt include polydeformed and metamorphosed pelitic and quartzose schists (e.g. Quartz-Mica-Schist unit), metavolcanic rocks, calcareous schists and marble and occasional orthogneiss bodies (e.g. Brosge & Reiser 1971, Nelson & Grybeck 1980, Dillon *et al.* 1986; Little *et al.* 1994; Vogl 2003; Till *et al.* 2008)(Fig. 2). Strain in the Schist Belt is sufficiently high that all original stratigraphic units are isoclinally folded and severely transposed. Despite this metaclastic rocks of the Schist Belt are commonly divided into 1) a structurally lower unit of calcareous metasedimentary rocks exposed in structural windows and along the north side of the belt intruded in places by Neoproterozoic or Devonian orthogneisses, and 2) an upper unit of quartz-mica schist which encloses Devonian volcanic and volcanoclastic sequences (Moore *et al.* 1997a, 1997b). The protoliths of the metasedimentary rocks of the Schist Belt are thought to range in depositional age from late Proterozoic to Late Devonian (e.g. Till *et al.* 2008; Moore *et al.* 1997a), as demonstrated by rare fossil data and limited igneous and detrital U–Pb zircon results (e.g. Dillon *et al.* 1980; Hitzman *et al.* 1986; Moore *et al.* 1997b; Till *et al.* 2008). The only detrital zircon data published from the southern Brooks Range is that of Moore *et al.* (1997b): a quartzite within the Marion Creek schist, thought to be correlative with the quartz-mica-schist unit (e.g. Brosge & Reiser 1964), that yielded rounded (n=16) Precambrian zircon with ages of 3.0, 2.0 to 1.8, and 1.5 to 1.4 Ga and a population of euhedral (n=11) zircon with ages from 370 to 360 Ma. Because of the dearth of age constraints and its extensive metamorphic overprinting, the provenance and extent of early Palaeozoic and older strata in the southern Brooks Range remains for the most part unconstrained.

Cosmos Hills

The Cosmos Hills is an E-W trending doubly plunging anticlinal fold structure south of the Brooks Range (Hitzman *et al.* 1982). High strain metamorphic rocks are exposed in the Cosmos Hills below a low angle normal-sense fault zone that juxtaposes the Angayucham Terrane and overlying sediments of the Yukon-Koyukuk Basin. Cosmos Hills footwall rocks are lithologically correlative with, and thought to form part of, the Schist Belt (Fig. 2) (Hitzman *et al.* 1982; 1986; Christiansen & Snee 1994) and so are without distinction in places as Schist Belt.

Samples of metamorphic rocks from the Cosmos Hills in this study were described by Christiansen (1989; see also Christiansen & Snee 1994). They exhibit a prominent, subhorizontal, locally mylonitic, dominant metamorphic foliation, S_D , defined by greenschist facies mineral assemblages. The dominant foliation isoclinally folds and

transposes compositional layering and at least one earlier metamorphic foliation. The grade of metamorphism associated with S_D is higher in the Cosmos Hills than in the adjacent Schist Belt or previously described Wiseman quadrangle region of the Brooks Range in that biotite + garnet assemblages characterize penetratively deformed rocks exposed at the deepest structural levels of the Cosmos Hills anticline (Christiansen & Snee 1994).

Greywacke-Phyllite Belt

A package of metagreywacke and phyllite with quartzose and chert-rich detritus and limited carbonate rocks is found between the Schist Belt and the Angayucham terrane (Moore *et al.* 1994; Murphy & Patton 1988; Patton *et al.* 1994; Till *et al.* 2008)(Fig. 2). This belt of rocks is heterogeneous and in places difficult to differentiate from adjacent belts of rock (Till *et al.* 2008). It is known as the Slate Creek subterrane (Moore *et al.* 1994), Phyllite belt (Patton 1973; Till *et al.* 2008), Phyllite-greywacke belt (Till 2016), southern Coldfoot terrane (Jones *et al.* 1987), metagraywacke belt (Gottschalk *et al.* 1998), metagraywacke and phyllite belt (Dillon *et al.* 1986), and is perhaps similar to parts of the Venetie terrane (Jones *et al.* 1987). Structural thickness estimates range from 1-3 km (Hitzman *et al.* 1986; Howell *et al.* 1986; Gottschalk 1987). Locally, metagabbro and diabase of unknown age have intrusive and extrusive relationships within this belt (Nelson & Grybeck 1980; Dillon *et al.* 1986; Dillon 1989; Karl *et al.* 1989; Moore *et al.* 1994; Patton *et al.* 1994). Age constraints include late Palaeozoic (possibly Mississippian to Triassic) radiolarian chert layers (Dillon *et al.* 1986), Devonian palynomorphs/flora (Gottschalk 1987; 1998), Devonian plant fossils and spores (Gottschalk 1987; Brosgé & Reiser 2000; Moore *et al.* 1994), and mid-Palaeozoic (likely late Silurian to early Mississippian) megafossils and conodonts (Patton *et al.* 1968; Oliver *et al.* 1975; Till *et al.* 2008). But these fossils are mainly from fault-bound slivers along the margins and thus their inclusion within the Greywacke-Phyllite Belt is suspect. Though the Greywacke-Phyllite Belt of the Brooks Range is correlated to a lithologically and structurally similar belt of rocks in the Ruby terrane (e.g. Patton *et al.* 1994; 2009), the original palaeogeographic relationship of these packages of rocks is unclear because models vary widely for the displacement history of the Ruby terrane with respect to Arctic Alaska (e.g. Moore & Box 2016).

Yukon-Koyukuk Basin

The Yukon-Koyukuk Basin (YKB) is a successor basin formed within the hinterland of the Brooks Range fold-thrust belt (e.g. Patton *et al.* 2009) thought to be the result of a significant episode of extension within the Brooks Range (e.g. Miller & Hudson 1991). The major pulse of submarine deposition in the Yukon-Koyukuk basin is ~113 Ma (Patton *et al.* 2009), though deposition may have begun as early as Barremian (131-126 Ma) (Patton *et al.* 1994; Patton & Box, 1989). The YKB consists of deltaic, submarine fan, and marginal shelf and slope deposits whose basal units have mollusks and ammonites reported of Albian age (113-100 Ma) and granite clasts with 113 Ma cooling ages, whereas some nonmarine fluvial deposits are thought to be Late Cretaceous (Cenomanian and Turonian? ~100-90 Ma) based on plant and mollusk fossils (Patton *et*

al. 2009 and references therein). The YKB samples included in this study structurally overlie the Cosmos Hills and provide a constraint on the structural evolution of the southern flank of the Brooks Range.

METHODS

Twelve samples of variably metamorphosed metapelite from the Schist Belt, two samples of sandstone from the structurally overlying Greywacke-Phyllite Belt, and two samples from the overlying Yukon-Koyukuk basin sediments were sampled from various localities in or near the southern central Brooks Range (Fig. 2; Table 1) and studied for U–Pb detrital zircon (DZ) geochronology. A subset of five samples was further assessed for zircon Hf isotope compositions. Six samples are from the Cosmos Hills (Fig. 2B), one sample is from the Ruby Terrane, and the remaining samples are from the Wiseman quadrangle (Fig. 2A). Many of the samples were taken from archived material from areas with an established structural framework (Hitzman *et al.* 1982; Hitzman 1983; Christiansen & Snee 1994; Little *et al.* 1994; Law *et al.* 1994; Table 1).

Zircon was extracted by conventional bulk separation methods (i.e. mechanical, magnetic, and density), mounted in epoxy and sectioned to expose core-and-rim relationships. Mounted zircon were imaged for cathodoluminescence textural assessment by scanning electron microscopy (SEM) and then subjected to *in situ* isotopic analysis by secondary ion mass spectrometry (SIMS) and/or laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS). A detailed explanation of analytical methods and instrumentation can be found in supplementary materials S1.

Our visualizations and interpretations are based on relatively concordant and precise subsets of each of the U–Pb detrital zircon data sets. Specifically, we include only analyses with 1) <10% uncertainty in interpreted age, 2) 90–105% concordance (calculated as ratio of $^{206}\text{Pb}/^{238}\text{U}$ age to $^{206}\text{Pb}/^{207}\text{Pb}$ age when >600 Ma, otherwise not determined), and 3) Uranium concentrations < 1500 ppm. Further, we consider zircons exhibiting apparently high levels of common Pb (i.e. ^{204}Pb well above the limit of detection) to be of suspect age significance, even where a common Pb correction has been applied. Because of our inability to precisely measure low concentrations of ^{204}Pb correction by LA-ICP-MS, no common Pb correction is applied to U–Pb data collected at Stockholm University and UC-Santa Cruz. A common Pb correction was made for the data collected at the Arizona LaserChron as well as at the NordSIM facility by using the measured $^{206}\text{Pb}/^{204}\text{Pb}$ ratio and the interpreted initial Pb ratio according to the model of Stacey & Kramers (1975). We assign a cutoff of 800 Ma (corresponding to a natural gap in our data set) above which we preferred $^{207}\text{Pb}/^{206}\text{Pb}$ ages and below which we preferred $^{206}\text{Pb}/^{238}\text{U}$ ages as our “best” or “interpreted” age for each analysis. Weighted mean and concordia plots were generated using Isoplot (Ludwig 2008). Errors are reported at 2-sigma unless otherwise reported. Where long-term instrumental reproducibility data is unavailable we suggest an additional ~2% uncertainty as observed in published interlaboratory U–Pb geochronology comparisons (e.g. Kosler *et al.* 2013).

Visual and statistical comparison methods are used to assess detrital zircon provenance. In doing so, we plot and assess each sample individually rather than as sums or averages to provide a more accurate representation of the natural variance. Although combining detrital zircon results from several samples collected within the same unit/region/craton produces apparently higher confidence as to the presence or absence of age populations (as a result of the larger sample size), such averaging tends to over-smooth natural variability and may lead to erroneous visual and statistical interpretations such as when applying the Kolmogorov-Smirnov (K-S) test for similarity (e.g. Press *et al.* 1986; Guynn & Gehrels 2010). To assess Early Palaeozoic palaeogeography (i.e. from which continent these rocks were receiving sediment at the time of deposition) we choose to highlight the >850 Ma age fraction of each sample because these appear to be the most stable provenance patterns for Baltica and Laurentia from their separation in the Neoproterozoic until their re-assembly in the mid-Palaeozoic.

RESULTS

Zircon U–Pb geochronology and petrography

Zircons from the majority of samples are relatively small (20–100 microns in length) and vary in color from pink to clear, reflecting a fine grain detrital origin, and, in some samples, the zircons are coated with graphite. Zircons from the Yukon-Koyukuk Basin and Greywacke-Phyllite Belt samples are larger, but still diverse, reflecting a detrital origin. Metamorphic zircon overgrowths characterized by distinct worm-like and/or non-oscillatory CL textures mantling oscillatory-zoned detrital cores are recognized in several of the samples from the Schist Belt and the Cosmos Hills (Fig. 3).

In situ U–Pb analyses of detrital zircon from sixteen samples (a total of 1269 analyses: N=16 n=1269) are presented in supplementary materials S2. Excessive discordance in some of the detrital zircon U–Pb data set is attributed to mixing with younger metamorphic overgrowths and/or Pb-bearing inclusions (Fig. 3). Despite discordance, 890 detrital zircon (DZ) ages that are ≤10% discordant, ≤5% reversely discordant, and have <10% age uncertainty, and whose uranium concentrations are <1500 ppm were obtained and are visualized using kernel density estimation (KDE) plots with a consistent bandwidth of 10 Myr to permit comparison between results of varying precision (Fig. 4). In addition to detrital populations, sixty-seven concordant ages were obtained from metamorphic overgrowths identified by their CL texture and by their especially low Th/U ratios (Fig. 5A) in at least 5 different samples that indicate a common timing of zircon growth across the localities studied here. The U–Pb data from zircon overgrowths where assessed in combination with detrital cores is presented here, but as an adequate discussion of their petrogenesis and significance is beyond the scope of this paper they will be discussed separately in a forthcoming companion paper. The following discussion of our results is organized by both sample locality (e.g. Cosmos Hills, Wiseman quadrangle) and tectonic domain (e.g. Schist Belt, Greywacke-Phyllite Belt).

Metapelites of the Cosmos Hills (four samples)

All of these samples were collected from beneath the Cosmos Hills detachment fault from metapelites exhibiting a common upper greenschist facies metamorphism (see Hitzman *et al.* 1982; Christiansen 1989).

Sample (**1988 PCBR**) **CH-107** is a garnet-quartz-mica schist (Christiansen 1989), mapped as the Kogoluktuk Schist (Hitzman *et al.* 1982) collected from the deepest exposed structural levels in the Cosmos Hills (Fig. 2B) (Christiansen & Snee 1994). The dominant foliation in this sample is defined by large phengite-rich domains and quartz seams. Chlorite appears to be statically replacing garnet in much of the sample. This sample was collected near a garnet-hornblende-bearing metabasite body (CH-107b of Christiansen 1989) that preserves the potentially earlier garnet-hornblende assemblage (discussed above). Zircon from this sample is typically anhedral to subhedral and displays the greatest degree of metamorphic overgrowth of any of the samples studied. These metamorphic overgrowths are ubiquitous on all zircon grains recovered and are often >50 microns thick. There is slight CL zoning in some rims. Zircon separates were analyzed in two different sessions (Fig. 4), with the LaserChron session focused on resolving metamorphic overgrowths and the Stockholm session targeting relict cores for detrital provenance. Results from the two sessions differ slightly but within expected behavior for LA-ICPMS comparisons (e.g. Kosler *et al.* 2013), and thus detrital data from both sessions are discussed in combination hereafter. *Detrital zircon data:* 42 concordant core ages passing our selection criteria were obtained out of 162 total “core” analyses (Fig. 4). Forty percent of those grains are Cambrian to Devonian in age (n=17) with the remainder mostly Proterozoic ages: 405-530 (n=17), 600-620 (n=2), 950-1180 (n=9), 1235-1500 (n=9), 1610 (n=1), 1760-1850 (n=3), and 2800 Ma (n=1) (Fig. 4). Discordance is difficult to assess in the Silurian and younger grains which may be spuriously young due to mixing with overgrowths. *Zircon overgrowths:* A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of the 35 youngest concordant metamorphic overgrowth ages from the LaserChron session is 114.02 ± 0.67 Ma, consistent with their concordia age of 114.1 ± 2.2 Ma. The weighted mean age of the 19 youngest Stockholm session ages, 111.3 ± 1.6 Ma, is consistent with a concordia age of the most pronounced cluster (n=13), 110.0 ± 1.3 Ma. Several concordant analyses in the 130-145 Ma range may coincide with an earlier metamorphism. Metamorphic overgrowths in this sample have low Th/U ratios, with the lowest Th/U ratios corresponding to the most concordant U–Pb ages (Fig. 5A).

Sample (**ELM**) **CH-1D** is a garnet biotite muscovite chlorite schist from the Kogoluktuk schist unit (Hitzman *et al.* 1982) collected from near the contact with the Kogoluktuk orthogneiss (Fig. 2B). Zircons from this sample display thin metamorphic zircon overgrowths around detrital populations. *Detrital zircon data:* 66 acceptable ages were obtained from 82 analyses. Detrital populations recognized are 410-590 (n=11), 650-790 (n=12), 950-1910 (n=40), 2500-2730 (n=3) Ma (Fig. 4). The 3 youngest apparently concordant grains provide a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 418 ± 5 Ma (MSWD=0.44) as a possible maximum depositional age constraint. Two of those grains, however, have high uranium concentrations (>1000 ppm) and the third may yield a spuriously younger age due to mixing with young metamorphic overgrowth, limiting our confidence in this youngest population. As such, the next youngest set of concordant grains, which yield a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 458 ± 9 (2 σ , internal errors only, MSWD=1.0, n=4),

offers a more robust constraint on maximum depositional age. *Zircon overgrowths*: Only one rim was acceptably resolved from this sample, giving a $^{206}\text{Pb}/^{238}\text{U}$ age of 112.8 Ma.

Sample (**PCBR**) **87-6B** is a muscovite schist mapped as Kogoluktuk schist (Hitzman *et al.* 1982; Christiansen 1989). *Detrital zircon data*: Zircon yield was low for this sample, resulting in 25 acceptable concordant ages out of 52 analyses (Fig. 4). Thirty-two percent of the zircon (n=8) are <850 Ma and exhibit three age peaks of 404, 494, 585 Ma. The older ages define small peaks at 1000, 1160, 1275, 1515, 1661, 1815, 1895, 1976, 2700 Ma. The youngest grains, which define the 408 Ma peak (n=2), have high uranium (~1000 ppm), which lowers the confidence in using these as a constraint on depositional age. The next youngest age cluster (n=3) at 480 Ma provides a more conservative upper limit on depositional age. *Zircon overgrowths*: No overgrowths were resolved by LA-ICPMS analysis, but CL images show bright rims on most grains consistent with zircon from samples with resolved metamorphic overgrowths.

Sample (**PCBR**) **87-1E** is a muscovite schist, mapped within the Anirak Schist (Hitzman *et al.* 1982; Christiansen 1989). Small sample size resulted in a poor zircon yield. Only 5 acceptable ages were obtained out of 15 analyses because of pronounced discordance. *Detrital zircon data*: The concordant grains yield ages of 450, 455, 1045, 1330, and 1900 Ma (Fig. 4), indicating a detrital protolith with an Ordovician maximum depositional age of ~453 Ma. *Metamorphic overgrowths*: A linear fit of the discordant data (n=10) intercepts concordia in the range of 60-145 Ma, but estimated errors and MSWD's from such fits are very high. Thus, whereas discordance in this sample is consistent with being due to mixing with metamorphic overgrowths of a similar age as seen in other samples, the data is not sufficient to resolve its precise timing.

Yukon-Koyukuk sediments of the Cosmos Hills (two samples)

Two samples, (1988 PCBR) CH-118 and (PCBR) 87-9, form part of the elastic-detrital sedimentary succession of the Yukon-Koyukuk Basin exposed in the hanging wall above the Cosmos Hills detachment fault. They are themselves weakly metamorphosed (chlorite-grade) and slightly deformed.

Sample (**1988 PCBR**) **CH-118** is a foliated mica-rich quartzite/greywacke from the southern margin and upper structural levels of the Cosmos Hills (Fig. 2B), juxtaposed above the "Anirak" quartz mica schist (Hitzman *et al.* 1982; Christiansen 1989). Zircon from this sample range from 30-200 microns in length. Most exhibit oscillatory zoning but a few display bright non-oscillatory zoned rims. This sample yielded 106 acceptable ages and is characterized by detrital populations in the age ranges ~90 Ma (n=5), 360-450 Ma (n=16), 620-780 Ma (n=16), 950-1260 Ma (n=27), 1320-1800 Ma (n=40), and single grains at 146 Ma and 2700 Ma (Fig. 4). Prominent age probability peaks exist at 89, 385, 424, 678, 1044, 1239, 1357, 1445, and 1635 Ma. The youngest population (n=5) yields a weighted mean $^{206}\text{Pb}/^{238}\text{Pb}$ age of 89.0 +/- 2.3 Ma (2 σ , internal errors only, MSWD=1.17), providing a Late Cretaceous maximum depositional age constraint for this sample (Fig. 5B). A single young outlier yields an age of 146 Ma (Fig. 5B). The second youngest population (n=8) yields a weighted mean $^{206}\text{Pb}/^{238}\text{Pb}$ age of 379 +/- 4 Ma (2 σ ,

internal errors only, MSWD=1.3). All Mesozoic grains have Th/U ratios between 0.4-1.1, typical of igneous zircon and in strong contrast to the low Th/U values seen in metamorphic overgrowths of the other samples (Fig. 5A).

Sample (**PCBR**) **87-9** is a pebble conglomerate with abundant phyllite clasts collected above the Cosmos Hills near the southern structural contact with overlying Angayucham terrane units (Fig. 2B). Good zircon yield resulted in 112 acceptably concordant ages out of 138 analyses. This sample is characterized by detrital populations in the age ranges of 86-90 Ma (n=2), 337 Ma (n=1), 374-490 Ma (n=11), 565-740 Ma (n=16), 930-1830 Ma (n=77), and the remaining ages (n=5) are dispersed between 1960-2110 and 2790-2980 Ma (Fig. 4). The most dominant age probability peaks occur at 635 and 1650 Ma, and sub-peaks are recognized at 424, 488, 741, 1070, 1240, 1370, 1450, and 1550 Ma (Fig. 4). Two concordant Cretaceous U–Pb ages yield a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 87.8 ± 3.3 (2 σ , internal errors only, MSWD=0.75)(Fig. 5C) defined by normal Th/U values (Fig. 5A), consistent with those of sample (1988 PCBR) CH-118. A young Palaeozoic grain at 337 Ma is concordant and has typical zircon chemistry. The next youngest Palaeozoic population (n=3) is 385 ± 11 Ma (2 σ , internal errors only, MSWD=1.03).

Schist Belt metapelites from the John River, Wiseman quadrangle (four samples)

All of these samples were collected along the John River. Though they display some compositional heterogeneity, these samples exhibit a common greenschist facies mineral assemblage.

Sample **13-CH-BR03b** is a chlorite white mica schist sampled adjacent to the John River and mapped as part of the Ambler sequence of Dillon et al, (1986). This sample was analyzed in two sessions with two different instruments (SIMS n=12/14; LA-ICMPS n=30/87). Relative age spectra are consistent across both sessions and so are combined in discussion. All but 2 of the 42 acceptably concordant ages fall between 1000-1800 Ma and define peaks at 1145, 1332, and 1482 Ma. Most grains (n=36) fall between 1000-1500 Ma (Fig. 4). Two Archean grains of 2675-2775 Ma are present. A conservative upper limit is given at 1040 Ma as defined by the weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of the 3 to 7 youngest grains.

Sample **13-CH-BR04/A** is a composite of two different samples collected at the same locality (BR04 and BR04A) of a micaceous quartz-rich greywacke exposed along the John River and mapped as part of the Ambler sequence of Dillon *et al.* (1986)(Fig. 2A). Both samples were crushed separately, analyzed separately, and are displayed separately but discussed in combination. Geochronological data from this sample is difficult to interpret. Ubiquitous metamorphic overgrowths on small zircon produced extreme discordance in most analyses (spot size and ablation time constraints led to unavoidable age domain overlap), for which a more aggressive trimming approach was needed to recover data from single age domains (see supplementary materials S1). *Detrital zircon data*: BR04 yielded 46 acceptably concordant ages out of 67 analyses (Fig. 4). All but four ages fall between 950-1750 Ma, which includes prominent peaks at 1038, 1226,

1621, and 1722 Ma. Two Archean grains and two zircon between 1800-2000 Ma are the oldest grains identified. A robust maximum depositional age is obtained from the group of the four youngest concordant ages, whose weighted mean $^{207}\text{Pb}/^{206}\text{U}$ age is 968 Ma. BR04A yielded 62 acceptable concordant ages from 93 analyses. All but five ages fall between 900-1750 Ma, which includes prominent peaks at 1066, 1275, 1451, and 1667 Ma. The remaining 5 grains yield ages of 1830-1925 Ma (n=3) and ~2640 Ma (n=2). *Metamorphic overgrowths:* In an attempt to characterize the young component defined by this rims, the isotopic data from mixed analyses (typically the beginning or end of an analysis) was used to define a mixing array. Utilizing this approach, the youngest resolved zircon rim component identified was a 3.5 second integration window at the start of analysis BR04A-41, which gave a concordant $^{206}\text{Pb}/^{238}\text{Pb}$ age of 119 +/- 17 Ma. Consistent with this young age, a fit to the 85 youngest resolved rim components yielded a lower intercept age of 118 +/- 54 Ma.

Sample **ELM-13-BR22/A/B** is a composite of three different hand samples (i.e. BR22, BR22A, BR22B) collected near each other at the same outcrop along the John River and presumed to belong to the same unit (Fig. 2A). Each sample was crushed separately and analyzed separately, but because zircon yield was low and zircon textures and ages are consistent between the three samples, they are combined in all geochronological plots and data tables and discussion hereafter (spot identification numbers in the U–Pb table preserve their original sample association). Predominantly, this is an albite-chlorite-epidote-titanite-calcite-quartz-graphite-white mica schist with incipient growth of biotite. A likely protolith is volcanic-rich lithic sandstone or intermediate/mafic volcanic rock. Analyses were performed with high spatial resolution at the NordSIM facility because of the small size and inclusion rich character of metamorphic overgrowths. CL images show that the overgrowths do not exhibit oscillatory zoning, but instead a “wormy” appearance due to the interference of solid inclusion trails on luminescence (Fig. 3). *Detrital zircon data:* Acceptably concordant detrital cores (n=39) display a spread from 477-760, with a peak at 622 Ma; and from 900-1750, with small peaks at 1180 and 1620 Ma; and a single age at 2690 Ma (Fig. 4). The youngest grain is not within error of any others and so may have experienced some mixing or Pb-loss, thus the most robust constraint on maximum depositional age for this sample is obtained from the second and third concordant ages as <571 Ma. *Metamorphic overgrowths:* Seven overgrowths were dated that define two age clusters centered at ~115 (n=4) and ~145 Ma (n=3). The two youngest ages overlap with concordia and provide a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 113.4 +/- 1.3 Ma. Uranium concentrations of overgrowths vary between 190 and 660 ppm with one at 1660 ppm. Th/U ratios vary between 0.008 and 0.027 with one at 0.079 (Fig. 5A).

Sample **13-CH-BR09 (MS4)** is a chlorite white mica schist sampled along the John River and presumed to belong to the quartz mica schist unit of the Schist Belt (correlative with that mapped to the east by Little et al 1994; Law *et al.* 1994, discussed above)(Fig. 2A). Twenty concordant SIMS analyses (out of 38 total) display a range of ages from 640-670 Ma (n=2), 980-1725 Ma (n=17), and a single grain at 1935 Ma (Fig. 4). Eleven other grains yielded apparently Neoproterozoic ages but were excluded due to discordance or high uranium concentration.

Schist Belt metapelites near North Fork Koyukuk River, Wiseman quadrangle (three samples)

These three samples were collected from west of the North Fork Koyukuk River (see geological mapping and structural context as provided in Little *et al.* 1994)(Fig. 2A). They exhibit a common greenschist facies metamorphic overprint.

Sample **90-LL-71** is a garnet quartz chloritoid white mica chlorite schist belonging to the quartz mica schist (QMS) unit of Little *et al.* (1994), and mapped specifically as graphitic quartz mica schist (Pzsq) subunit or else locally graphitic micaceous albite calc-schist (Pzcb) subunit of QMS. Of 94 total analyses 92 passed our filtering criteria. Most of the grains are younger than 850 Ma (n=67), whereas the remainder (n=25) are Proterozoic or late Archean (Fig. 4). Post-850 Ma grains exhibit probability peaks at 377 (n=27), 420 (n=40), 464 (n=20), 564 (n=4), 619 (n=1), 692 (n=1), whereas populations of older grains center on 966, 1080, 1460, 1840, 2530, 2617, and 2704 Ma. The youngest 4 grains produce a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 368 ± 3 Ma (2σ , internal errors only, MSWD=0.79), which provides a maximum depositional age constraint for this sample.

Sample **90-LL-33** is an epidote titanite albite schist belonging generally to the quartz mica schist (QMS) unit of Little *et al.* (1994) and specifically the graphitic quartz mica schist (Pzsq) subunit of QMS (Fig. 2A). Mineralogically it contains abundant chlorite, white mica, biotite, albite, epidote/clinozoisite, titanite. All of the analyzed zircon grains (n=10) from this sample are within error of Devonian, but the low zircon yield makes it difficult to assess whether older populations exist that were not effectively sampled here (Fig. 4). Seven grains cluster on concordia with a weighted mean of 390.6 ± 3.4 (2σ , internal error only, MSWD=1.3) providing a reasonable estimate of the age peak, whereas two younger grains indicate either a younger depositional age of ~375 Ma or Pb-loss. An older concordant 433 Ma grain may indicate magmatic inheritance or detrital mixing, depending on the nature of the protolith (i.e. volcanic or sedimentary, respectively).

Sample **89 BRLM-98** is garnet albite mica schist also belonging generally to the quartz mica schist (QMS) unit of Little *et al.* (1994), and specifically to the graphitic schist with minor quartzite (Pzsq) subunit of QMS (Fig. 2A). Mineralogically, it contains abundant albite, white mica, chlorite, and relict garnet. Only a fist-size fragment of rock was available for crushing, which resulted in poor zircon yield. Out of 30 total analyses 23 acceptable ages were obtained, which are characterized by detrital age clusters of 355-445 (n=17), 550 (n=1), 970-995 (n=3), and 1325-1385 (n=2)(Fig. 4). The dominant probability peaks occur at 377 and 430 Ma.

Metapsammities of the Greywacke-Phyllite Belt, Brooks Range and Ruby terrane, Dalton Highway localities (two samples)

Two samples from the Greywacke-Phyllite Belt collected at two locations along the Dalton Highway – one belonging to the Brooks Range and the other to the Ruby structural high – produce similar detrital zircon populations in different abundances.

Sample **AK07 SA2A** is from a deformed phyllite with thick beds of fine sandstone within it, sampled from fresh-quarried pieces just north of “Gobblers Knob” (mile marker 132 on the Dalton Highway) on the northwest margin of the Ruby structural high (phyllite map unit Pzp of Patton & Miller 1973). Good zircon yield and low discordance resulted in 98 acceptable age analyses out of 103, which are dominated by Precambrian ages (n=79) (Fig. 4) that produces significant peaks at ~1900 Ma (n=18) and ~1100 Ma. The remainder of the zircon are Palaeozoic (n=19) and range in age from 358 to 470 Ma with two young grains of ~320 Ma (Fig. 4).

Sample **AK07 SA4** is located further to the north, in the Brooks Range, just southeast of Coldfoot, up Slate Creek (Fig. 2A), and is a black to silvery slate, multiply deformed with fine-grained grey sandstone to thin brownish sand-siltstone (Slate Creek phyllonite of Moore *et al.* 1997a). This sample produced 102 acceptable ages out of 105 analyses. It is predominantly characterized by Palaeozoic age zircon (n=64) ranging in age from 332 to 499 Ma with single outliers at 254 Ma and 641 Ma. Approximately 38 of the zircon are Precambrian in age with a main peak at 1943 Ma (Fig. 4).

Hf-isotopes in zircon

Hf isotope compositions (n=121) were determined for a subset of the samples (90-LL-71, 90-LL-33, CH-1D, CH-107, and (1988 PCBR) CH-118) and are presented in figure 6 (see supplementary materials S3). The zircons analyzed exhibit a wide range of ϵHf_T compositions, varying between +20 and -20 with two outliers plotting more negative than -20 (see Fig. 6A). Temporally, ϵHf_T starts from near CHUR at ~1800 Ma and trends steeply upwards through time toward more juvenile compositions, hovering near depleted mantle values by ~1350 Ma. Compositions then become progressively more evolved towards 900 Ma. Grains from 800 to 550 Ma are CHUR-like, becoming slightly more evolved with time, but trends are poorly resolved (Fig. 6B). Palaeozoic zircon compositions begin near depleted mantle at the start of the Cambrian but then trend steadily along a crustal evolution line until ~400 Ma, at which point the compositions trail off very steeply towards highly negative (evolved) values at ~360 Ma (Fig. 6B). ϵHf_T compositions for sample 90-LL-33 (n=7) all cluster around ~4.5. A single grain assessed from the ~90 Ma population (sample (1988 PCBR) CH-118) yields a $\epsilon\text{Hf}_T = 5.2$ (Fig. 6A). Analyses of metamorphic overgrowths on samples CH-1D and CH-107 are not presented here as their discussion is beyond the scope of this paper.

DISCUSSION – DETRITAL ZIRCONS

Detrital signatures and maximum depositional ages (MDA) of metasedimentary packages

Detrital zircon results from samples from the Schist Belt studied here indicate that a diversity of protoliths are present within the Schist Belt (Fig. 4). These protoliths exhibit a range of sedimentary source areas that are presumed to reflect variations in depositional

age and/or location and that are recognizably distinct from the signatures of samples collected from the adjacent Yukon-Koyukuk Basin and Greywacke-Phyllite Belt. In total, we recognize six distinct “signatures” in this study that characterize different depositional systems/protoliths. These are described from oldest to youngest as inferred from their youngest included populations and related to strata elsewhere within the AACM exhibiting similar provenance.

A major limitation of this data set is the low-n sample size of many of the Schist Belt samples for which archived aliquots were small and for which zircon large enough to extract may not have been originally abundant in their fine grain protoliths. The requisite number of analyses in a study depends widely on the questions being asked (Vermeesch *et al.* 2016). General shape of the age distribution may be adequately characterized with only a few dozen grains (Avdeed *et al.* 2011). At least 117 analyses would be needed to be 95% confident of having not missed *any* population having a >5% relative abundance (Vermeesch 2004). The probability of missing just *one* particular population of interest is given by $P=(1-f)^n$, where f is the frequency of that population and n is the sample size (Dodson *et al.* 1988). This suggests that only 59 grains must be analyzed to reduce the probability to <5% of missing the youngest age population having >5% relative abundance (Dodson *et al.* 1988). Eight of the samples analyzed here are of sufficient size to determine maximum depositional age. The remaining samples are interpreted by comparing their general age distribution, lithology, and location to the more robust signatures recognized here and elsewhere within the AACM. Inclusion of these low-n samples into Signature Groups is non-unique and may change with further analyses.

Consideration of the youngest zircon age or population within a sample can help place an upper age constraint on the timing of deposition (e.g. Dickinson & Gehrels 2009), but this cannot be assumed to reflect the actual depositional age of strata without additional geologic constraints. Quite often detrital zircon maximum depositional ages (MDA) are as much a reflection of tectonic setting and drainage patterns as they of age (e.g. Cawood *et al.* 2012). Active tectonic environments, for example, typically contain zircon ages close to depositional age, whereas passive margin environments show a moderate to large time gap between the youngest zircon age and the depositional age (Cawood *et al.* 2012). While the tectonic setting of the metasedimentary rocks presented here is unknown, it is suspected deposition within the AACM would have been in or near an active continental margin at various times (e.g. Neoproterozoic, Devonian, etc., as indicated by igneous bodies of those ages, e.g. Amato *et al.* 2014), during which times we might expect to have youngest detrital zircon populations that are close to the depositional age.

Recognizing subsequently reworked detrital signatures wherein there is no new input of younger zircon sources (e.g. Gottlieb *et al.* 2014) becomes a major challenge and necessitates consideration of additional geological constraints. Further, the general difference in relative abundance across differing age populations may not be a robust indicator of variations in catchment area or proximity but instead may be geological (e.g. hydraulic sorting, as in Malusà *et al.* 2016) or analytical (e.g. selection bias, as in Kosler *et al.* 2013) or else in our case may reflect some preferential metamorphic overgrowth or replacement process affecting various populations differently. It is with these caveats that we group samples into characteristic signature groups of which we infer diverse

depositional ages and settings.

Signature A – Early Neoproterozoic (Tonian) MDA

Signature A is defined by samples BR04, BR04A, and BR03b (Fig. 4), which are distinct from the remainder of samples studied in that they yield no Phanerozoic grains, exhibiting instead a range of ages from ~1.0-1.8 Ga and a small Archean population. The abundance of ~1.0 Ga grains places an early Neoproterozoic upper bound on depositional age. Signature A samples are potentially correlative along strike to calcareous schist units elsewhere within the Brooks Range that host igneous bodies of Neoproterozoic age (Till *et al.* 2008). In the western Brooks Range these igneous bodies are dated as ~750 Ma (Karl *et al.* 1989; Amato *et al.* 2014), whereas the nearby Ernie Lake and Sixtymile plutons (30 km and 10 km west of sample BR03b, respectively) yielded early Neoproterozoic ages (~970 Ma; Amato *et al.* 2014; Dillon *et al.* 1979, 1980)(Fig. 2A). The Ernie Lake pluton is interpreted as intrusive into marble at some places (Amato *et al.* 2014). Such relationships suggest that Signature A samples in this study may have a Neoproterozoic depositional age (i.e. and thus represent non-crystalline Precambrian “basement”), but the lack of intrusive constraints where our samples were collected combined with the local structural complexity leaves us unable to definitely place a lower bound on the depositional age for metasedimentary rocks characterized by this detrital zircon signature.

Signature B – Late Neoproterozoic (Ediacaran) MDA

Signature B is defined by the previously reported “Neoproterozoic theme” samples of Till *et al.* (2014a) from the Seward Peninsula and samples BR22/A/B and MS4 of this study (Fig. 4), which are characterized by a similar distribution of pre-850 Ma ages as observed in signature A but with the addition of an abundance of grains in the 550-760 Ma range and rare grains of Cambrian and/or Ordovician age. Because of structural complexity the stratigraphic relationship between signature A and signature B samples is unconstrained. These two signatures could have been derived from widely different source areas and then structurally accreted during either Brookian orogenesis or older (Caledonian?) deformation, or they could represent a semi-conformable stratigraphic package.

Signature C – Ordovician-Devonian (?) MDA

Signature C is best defined by sample 1D, and characterizes samples 6B, 1E, and CH-107 (Fig. 4). These samples show a spread of ages from 1.0-2.0 Ga and an abundance of Neoproterozoic and Palaeozoic age zircon. This signature comprises heterogeneous age spectra whose youngest ages are suspect of Pb loss or else analytical mixing with metamorphic overgrowths. Because of the small number of concordant analyses (n=67, 5, 33, and 47, respectively), the small sample size available, their fine grain pelitic character, and the abundance of metamorphic alteration/overgrowth in these samples, their similarity cannot be robustly assessed. They are discussed as one signature based on their geographic proximity and mapped lithological similarity within the Cosmos Hills. The abundance of Palaeozoic age zircon populations distinguishes Signature C from

Signatures A and B. This signature is constrained to be no older than Ordovician. Though speculative, the absence of a strong Devonian peak in this signature where such ages are common in post-mid-Devonian strata regionally (e.g. Gottlieb *et al.* 2014; Moore *et al.* 2015) suggests a possibly pre-mid-Devonian depositional age.

Signature D – mid-to-late-Devonian MDA

Signature D is defined by samples 90-LL-71 (this study)(Fig. 4), which is characterized by a clear abundance of Devonian age zircon in addition to an abundance of Ordovician-Silurian age grains. Samples 90-LL-33 and BRLM-98 (this study) of smaller sample size are also assigned to this signature by the similar age populations recognized and the similar Devonian maximum depositional age. This signature is so dominated by Palaeozoic age detritus that it is difficult to assess pre-Devonian sources. However, from our limited data set it appears that Signature D differs from previously discussed Signatures A-C in its Precambrian distribution. Specifically signature D exhibits apparent maxima at ~1.0, ~1.8, and ~2.6 Ga. Sample 90-LL-33, which yielded a nearly unimodal zircon population, may be meta-volcanic and might therefore support an along-strike correlation with Devonian felsic volcanic units further west (i.e. Ambler sequence in Hitzman *et al.* 1982) as suggested by Dillon *et al.* (1986). In this case, sample 90-LL-33 provides a single but direct constraint on the isotopic character of local volcanism, with an age of 390.6 ± 3.4 Ma (2-sigma, internal error only, $n=7$, MSWD=1.3) and $\epsilon\text{Hf}_T = 3.9 \pm 1.9$ (2-sigma, internal error only, $n=5$, MSWD=0.52). Such an isotopic composition and age is consistent with previously published whole rock ϵNd_T values from Ambler volcanics (Nelson *et al.*, 1993). The Devonian age of the youngest detrital zircon populations and of possibly intercalated volcanic rocks constrain the depositional age of Signature D as being mid-to-late Devonian or younger. The youngest 4 grains produce a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 368 ± 3 Ma (2 σ , internal errors only, MSWD=0.79). The juxtaposition of mid-Devonian volcanics with supra-CHUR values (sample 90-LL-33) with metasedimentary units containing late Devonian detrital zircon with sub-CHUR compositions (90-LL-71) may reflect structural layering that obscures subtle differences in depositional age and thus of improper grouping.

Greywacke-Phyllite Belt – Carboniferous (?) MDA

Two samples from the Greywacke-Phyllite Belt, one from the Brooks Range and the other from the Ruby terrane, SA4 ($n=102/105$) and SA2A ($n=100/103$), respectively, are distinct from any of the other signatures reported (Fig. 4). Depending on how one restores the Ruby terrane (see Moore & Box 2016 and references therein), these samples may have originated at widely separated localities. Both samples contain similar age populations, but of varying abundances. Though they contain an abundance of Palaeozoic age zircon similar to other young samples, Greywacke-Phyllite Belt samples yield a negligible amount of Neoproterozoic age zircon and exhibit a pre-850 Ma distribution that does not show the 1.0-1.8 Ga of Signatures A-C, but instead contains an abundance of ~1.1 Ga and ~1.9 Ga detritus. The youngest true population of zircon common to both samples constrains depositional age to post-middle Devonian, but a younger depositional age is suggested by Carboniferous ages in both samples ($n=7$) and a single latest Permian

age. Precambrian ages are far more abundant in the Ruby terrane sample, which may suggest that the Greywacke-Phyllite belt rocks structurally overlying the Ruby terrane were closer to cratonic source areas during time of deposition. Alternatively, if deposition were diachronous then the variation in abundance may reflect temporal changes to catchment areas common to both terranes.

Yukon-Koyukuk Basin – Late Cretaceous MDA

Two samples from the Yukon-Koyukuk Basin, (1988 PCBR) CH-118 (n=103/112) and (PCBR) 87-9 (n=120/138) (Fig. 4), contain a prominent ~90 Ma peak not observed in the other signatures, reflecting a much younger depositional age. The youngest Palaeozoic populations are observed at 379 and 385 Ma. Other important populations defining this signature include 410-450, ~470, and 600-740 Ma, with a spread from 1.0-1.8 Ga that rises to a distinct maxima at ~1.65 Ga, with then a few Archean age zircon. The pre-850 Ma distribution from these samples shows a very close similarity to the distributions of our signatures A-C, but most closely resembles signature C. Aside from the relatively small ~89 Ma population and two outliers at 146 and 337 Ma, no other post-Devonian zircon are recognized. A Late Cretaceous maximum depositional age is provided by the data.

Similarities of DZ signatures between the southern Brooks Range, Seward Peninsula and Ruby Terrane of the AACM

Data from this study supports previous correlations of metamorphic units across the Ruby and Seward terranes with those of the southern Brooks Range (e.g. Till *et al.* 2014a, 2014b) and is consistent with a presumed shared lower Palaeozoic continental margin setting across these southern (present day coordinates) terranes of the AACM (e.g. Dumoulin *et al.* 2002).

Signature A

Detrital zircon data from three quartzites of the Ruby terrane exhibit a similar provenance as our Signature A samples, though with few grains older than 1.6 Ga (Bradley *et al.* 2007; note that their Minook “complex” sample is also considered Ruby terrane in Till *et al.* 2007) (Figs. S4 and S5A-B). As in the Brooks Range Schist Belt, extreme strain and metamorphism of the Ruby terrane complicates stratigraphic interpretations (e.g. Roeske *et al.* 1995). The three Ruby terrane samples were previously interpreted as Devonian (Dover 1994) or Ordovician (Harris 1984) by association with nearby fossil-bearing carbonates, however the detrital zircon spectra without any Phanerozoic grains were interpreted as consistent with a Neoproterozoic to Cambrian depositional age and therefore implying a complex interlayering of Neoproterozoic to Devonian strata or else a previously unrecognized unconformity (Bradley *et al.* 2007). Similar structural complexity may exist in the southern Brooks Range as suggested by contradictory age constraints obtained from correlative calcareous schist units across the range (i.e. intrusive Late Proterozoic orthogneiss further west in contrast to mid-Palaeozoic conodonts recovered to the east) (e.g. Till *et al.* 2008).

Till *et al.* (2014a) reported four samples from the Nome Complex of Seward Peninsula that exhibit a similar provenance to our Signature A, which they refer to as their “Mesoproterozoic theme”. Samples of this character were interpreted as Devonian or younger because of an apparently interbedded relationship to sediments that have yielded middle Devonian zircon. As such, Till *et al.* (2014a) suggests they were deposited in small basins eroding local fault scarps, thereby alternating between proximal and distal sources. However, extreme structural complexity in the Nome Complex makes recognizing interbedding versus tectonic interlayering relationships difficult, allowing the possibility that the Mesoproterozoic theme units of the Till *et al.* (2014a) study are of Neoproterozoic depositional age structurally interleaved within younger sediments as in the Ruby terrane. The same ambiguity exists in the Schist Belt where sedimentary structures have been completely transposed. Regardless of whether the samples studied were deposited in the Neoproterozoic or simply reworked exclusively from underlying Neoproterozoic lithologies at a later time, they provide a constraint as to the signature of Neoproterozoic strata in the southern AACM and are thus useful for comparison to Neoproterozoic deposits elsewhere (Fig. 7A-7B).

Signature B

An important mafic, calcareous, and feldspathic schist unit (i.e. “Casadepaga” schist) on the Seward Peninsula, as well as some other potentially correlative samples there, is similar to our signature B and contains an abundance of Neoproterozoic age zircon together with a spread of Mesoproterozoic ages (the “Neoproterozoic theme” of Till *et al.* 2014a; see also Amato *et al.* 2009). Mesoproterozoic grains appear to be more abundant in Schist Belt samples than in those of the Seward Peninsula. The Casadepaga schist is suggested to have been deposited during a time of active mafic magmatism (Till *et al.* 2014b). Rare young zircon grains suggest a possibly Ordovician maximum depositional age for the Casadepaga schist (Till *et al.* 2014a). Alternatively, Miller *et al.* (this volume) correlate the detrital signature of the Casadepaga schist to ~550 Ma sandstones and shales in the Polar Ural Timanides and suggest a late Neoproterozoic age (i.e. that the Casadepaga schist represents the “pre-Paleozoic Timanian ‘basement’ of the Seward Peninsula”).

Signature C

A similar signature is noted in the Seward Peninsula by both Amato *et al.* (2009) and Till *et al.* (2014a) and combined within the “Early Palaeozoic theme” of Till *et al.* (2014a). Seven of these “Early Palaeozoic theme” samples (3 from the “Mixed unit” and 4 from the “calcareous metasiliceous unit”) and one from higher grade units potentially correlative with those (KM12, Amato *et al.* 2009) contain abundant Cambrian-earliest Devonian age zircon with subdued Neoproterozoic age populations (Fig. 7). Mesoproterozoic grains appear to be more abundant in Schist Belt samples than in those of the Seward Peninsula.

Signature D

Signature D from the quartz mica schist is similar to that reported from the correlative Marion Creek schist of Moore *et al.* (1997b). Though Brosge & Reiser (1964) had considered this unit to be the metamorphosed equivalent of the Late Devonian Hunt Fork Shale, Moore *et al.* (1997b) identified a population of Late Devonian zircon grains considered too young for Hunt Fork and therefore suggested its protolith to more likely be the widespread Upper Devonian-Mississippian fluvial-deltaic strata (i.e. “Kanayut”) of the Endicott Group of the Endicott Mountain allochthon (EMA=“Northern Thrust Assemblages” in Fig. 2; a stack of Upper Devonian through Lower Cretaceous sedimentary rocks imbricated and displaced hundreds of kilometers northward during the Cretaceous Brookian orogeny, as in Mayfield *et al.* 1988; Moore *et al.* 1994). Age data presented here confirms the younger maximum depositional age as in Moore *et al.* (1997b), but the recognition of highly evolved ϵ_{Hf} ratios in zircons of this population raise the possibility of these being derived from the younger and more radiogenic granites of the North Slope and/or Yukon (Fig. 6). Thus, whereas Moore *et al.* (1997b) considered Kanayut conglomerate as the most likely protolith, our data suggests Kekiktuk conglomerate as a possible protolith, which would have implications for the structural position of the Schist Belt within the Brooks Range. Specifically, if the quartz mica schist of the Schist Belt is metamorphosed Kekiktuk, then the Schist Belt may be considered parautochthonous with respect to the Mesozoic-Cenozoic fold-thrust belt.

A similar signature is present in rocks from the Seward Peninsula and is undifferentiated within the “Early Palaeozoic theme” of Till *et al.* (2014a). Two samples from Till *et al.* (2014a), a calcareous metasiliceous unit and a pelitic schist unit, contain a clear Devonian age zircon population with subdued older populations, similar to our signature D samples. A strong Devonian peak is recognized in sample BM7 from Amato *et al.* (2009) but it is different from our Signature D samples in its having an abundance of ~550 Ma age zircon.

Greywacke-Phyllite Belt

Bradley *et al.* (2007) reported data on a greywacke of the Rampart Group in the southern Ruby terrane (Tanana quadrangle; sample 03ATi47b) correlative with the Greywacke-Phyllite Belt further north. They obtained 34 concordant ages (of 69 analyses) defining similar populations to our signature E, including 1912-1953 (n=3), 909-1217 (n=8), and 351-504 Ma (n=14), with a single young concordant grain at 260 Ma. Ignoring discordance, their data yields 68% Palaeozoic ages compared to 61% in our sample AK07 SA4. The consistency of age populations including young outliers between our samples and that of Bradley *et al.* (2007) is consistent with the interpretation that the Greywacke-Phyllite Belt of the Brooks Range is correlative to that of the Ruby terrane (e.g. Patton *et al.* 1994; map units Pzp and Pzg of Patton & Miller 1973; Patton *et al.* 2009).

Yukon-Koyukuk Basin

No similar spectra have been reported in the AACM, but the zircon spectra identified

here together with the lithology of clasts within the samples is consistent with local derivation (e.g. Dillon & Smiley 1984; Murphy *et al.* 1989; Patton *et al.* 1994) from the metamorphic units of the southern Brooks Range (this study) with the addition of late Cretaceous volcanic sources (e.g. Wilson *et al.* 2015, and references therein). Interlayered tuffs within late Cretaceous units ~100 km to the west of the Cosmos Hills gave an ~85 Ma K–Ar biotite age and a dacite tuff ~70 km to the south yielded an ~87 Ma K–Ar biotite age (Patton & Miller 1968; Patton *et al.* 1968; recalculated in Wilson *et al.* 2015). ϵHf_T of ~5 (relative to ~17 for depleted mantle) of the ~90 Ma population reported herein suggests a minor component of crustal assimilation into whatever mantle-derived magmas were generated at this time. A zircon ϵHf_T composition of +5 corresponds to a whole rock ϵNd_T composition of +2 (using recalculated initial array of Vervoort *et al.* 2011: $\epsilon\text{Hf}_T = 1.50 * \epsilon\text{Nd}_T + 1.57$, and assuming zircon as the primary reservoir of Hf in crustal rocks, e.g. Hoskin and Schaltegger 2003). Such a composition is consistent with published whole rock Nd compositions of ~+1 ϵHf_T for the ~90 Ma calc-alkaline series igneous bodies of the eastern Yukon-Koyukuk Basin (Arth *et al.* 1989a), but differs from the ~115 Ma igneous bodies studied in the western portion of the Yukon-Koyukuk Basin or within the Ruby terrane where compositions trend closer to -9 ϵNd_i (Arth *et al.* 1989a; Arth *et al.* 1989b; Moll-Stalcup & Arth 1989; Roeske *et al.* 2015). Alternatively, a more distal source of late Cretaceous zircon might be provided by the voluminous Okhotsk-Chukotka volcanic belt (OCVB) in Russia, where volcanism at ~85 Ma exhibits a range of ϵNd_T from +1 to -4 (Pease *et al.* this volume). The general age distribution is most similar to our Signature C, consistent with the interpretation of having been locally sourced from the erosion of Brooks Range lithologies

Broader source regions for detrital zircon populations

Mesoproterozoic and older age zircon, Baltican-type (found in Signatures A, B, C)

The distribution of pre-Neoproterozoic age detritus in signatures A, B, and C (Fig. 4) is relatively consistent, differing noticeably only in Signature D and in the Greywacke-Phyllite Belt samples (Fig. 7A). No crystalline basement rocks of Mesoproterozoic age or older have been identified within the AACM (e.g. Amato *et al.* 2014), but the presence of basement of this age has been inferred at depth based on radiogenic isotope studies of younger igneous rocks exhibiting highly evolved compositions (e.g. Nelson *et al.* 1993; Amato *et al.* 2014). Whether the evolved isotopic compositions reflect assimilation of underlying crystalline basement or assimilation of sedimentary detritus eroded from cratonic regions of this age remains unknown, but the latter scenario seems more likely.

Zircons of Mesoproterozoic to early Neoproterozoic age in the Arctic may have been eroded from a northward continuation of the Grenville and/or Sveconorwegian orogens (e.g. Lorenz *et al.* 2012; Zhang *et al.* 2015, 2016) or northward-displaced fragments of the Appalachians and British Caledonides (e.g. Harland 1997; Pettersson *et al.* 2010; Kirkland *et al.* 2007, 2008, 2011), or from magmatic rocks associated with the Valhalla orogeny (Cawood *et al.* 2010, 2015). However, since Mesoproterozoic age zircon were widely dispersed across both Laurentia and Baltica by erosion of these Mesoproterozoic-

age orogens both prior to breakup of Rodinia (Cawood *et al.* 2007; Bingen *et al.* 2011) and then again during the late Neoproterozoic breakup and formation of passive margins (Heaman *et al.* 1992; Cawood *et al.* 2007; Dewing *et al.* 2008; Hadlari *et al.* 2014; Malone *et al.* 2014), these signatures may represent second or third cycle reworking of such zircons. Spencer *et al.* (2017) recognize differences in the temporal trends of Hf-isotope compositions of post-1200 Ma zircons associated with each of these orogens. Specifically, they found the ϵHf -time trajectory of the Sveconorwegian to follow a trend of $\sim 0.01 \text{ } \epsilon\text{Hf} / \text{Ma}$, the Valhalla to follow a trend of $\sim 0.02 \text{ } \epsilon\text{Hf} / \text{Ma}$, and the Grenville to follow a notably steeper trend of ~ 0.04 . The 1400-900 Ma ϵHf -time trajectory of Signature C in this study (samples CH-1D and CH-107 of the Cosmos Hills) appears more consistent with either a Sveconorwegian or Valhalla orogen source (Fig. 6), but its supra-CHUR values imply Sveconorwegian (Fig. 6). On the other hand, the steeper Mesoproterozoic trajectory of Signature D (sample 90-LL-71) ending in sub-CHUR values appears more consistent with a Grenville orogen source. More detrital zircon Hf-isotope data is needed to robustly apply this test, but we suggest that future studies make use of such trends to better differentiate these otherwise similar-age provinces widely recognized within the sedimentary record.

Mesoproterozoic and older age zircon, Laurentian-Type (found in Signature D and Greywacke-Phyllite Belt)

Signature D and the Greywacke-Phyllite Belt samples (Fig. 4) differ from the other signatures in their pre-850 Ma distribution (Fig. 7A). These two signatures contain a relative abundance of Grenville age ($\sim 1.1 \text{ Ga}$) and Laurentian craton age ($\sim 1.9 \text{ Ga}$) detritus which could have been sourced either directly from the NE Laurentian craton or indirectly by erosion of Laurentian lineage sediments (e.g. Lane & Gehrels 2014). Regardless, these samples indicate a likely derivation from Laurentian sources, which is consistent with their presumed depositional ages (post-early Devonian) being younger than the accretion of the southern Brooks Range terrane(s) to the northern margin of Laurentia (e.g. Dumoulin *et al.* 2002; Strauss *et al.* 2013).

Neoproterozoic zircon (found in Signatures A, B, C)

Potential sources for Neoproterozoic age zircon include: 1) Magmatism associated with the assembly and accretion of the Timanian microcontinent to Baltica: Novaya Zemlya (635-710 Ma, Pease & Scott 2009; Corfu *et al.* 2010), Timanides (730-550 Ma; Gee & Pease 2004; Corfu *et al.* 2010; Kuznetsov *et al.* 2007; 2010; Puchkov 1997; Lorenz *et al.* 2008, 2013). Locally, magmatism may have persisted through latest Cambrian time (Pease & Scott 2009; Lorenz *et al.* 2007). 2) Timanian-age magmatic rocks within the AACM itself: Chukotka (644 Ma, Natal'in *et al.* 1999; 574, 656, 666 Ma, Amato *et al.* 2014), Wrangel island (699 and 633 Ma, Cecile *et al.* 1991; Kos'ko *et al.* 1993; 706-707, Gottlieb *et al.* this volume), Seward Peninsula (681 and 676 Ma, Patrick & McClelland 1995; 539, 562, 565, 687 and 663 Ma, Amato *et al.* 2009; 678, Amato & Wright 1998; 669, 666 Ma, Amato *et al.* 2014), Southwestern Brooks Range (750 Ma, Karl *et al.* 1989; 742 Ma, Amato *et al.* 2014; 705 Ma, Karl & Aleinikoff 1990). ϵHf_T values for Neoproterozoic age zircon in this study are mostly supra-CHUR, consistent with zircon

reported from sources in the Timanide terranes (Kuznetsov *et al.* 2010), Chukotka (Gottlieb *et al.* this study), and as inferred for the now missing “Crockerland” source terrane north of the Franklinian Basin (Anfinson *et al.* 2012).

Source areas for Timanian age zircon in the Arctic may have become increasingly restricted through early Palaeozoic time as the Timanide basement was buried by Devonian shelf strata concurrent with broad uplift and local rifting in the interior of the Baltican shield (Nikishin *et al.* 1996; Miller *et al.* 2011). In the Polar Urals and also near St Petersburg Neoproterozoic zircon are most abundant in the basal units above the Timanian unconformity but are reduced in the lower Devonian (Miller *et al.* 2011). On Novaya Zemlya Neoproterozoic detritus is abundant in Cambrian and Ordovician sediments eroding Timanide basement, but then is replaced in Silurian-Devonian time by Sveconorwegian age material eroded from Caledonian uplifts (Lorenz *et al.* 2013). This pattern suggests our Signature B samples are consistent with Cambro-Ordovician signatures associated with Baltica, and offers a possible explanation for the dampening of this Neoproterozoic signal in our Signatures C and D whose maximum depositional ages are younger (Fig. 4). However, since the true depositional ages and stratigraphic relationships of our Signatures A-D remain unknown due to structural complexity and metamorphism, correlation to Palaeozoic successions in Baltica is highly speculative.

Cambrian-Silurian zircon (found in Signatures C, D and G-P Belt)

The nearest Cambrian to Silurian age magmatic and associated sedimentary strata are found in the Doonerak fenster within the Brooks Range where the Apoon assemblage has yielded igneous and detrital populations within that range having juvenile ϵHf_T compositions and is thought to represent a fragment of a peri-Laurentian Taconic-Caledonian arc system (570-420 Ma Strauss *et al.* 2017)(Fig. 6B). The next nearest known source of zircons of this age is within the Chukchi Borderland where dredged orthogneisses suggest a continuation of Cambrian-Silurian arc magmatism (530-420 Ma; Brumley *et al.* 2015; O’Brien & Miller 2014; O’Brien *et al.* 2016). The Chukchi borderland likely restores adjacent to the northern parts of the Caledonian orogen recognized in Pearya, northern Norway, and Greenland (Brumley *et al.* 2015; O’Brien *et al.* 2016). Subduction related volcanism is recorded within the Pearya terrane at various times (c. 480-440 Ma; Trettin 1987, McClelland *et al.* 2012). While most previous models had considered Pearya exotic to Laurentia (e.g. Trettin 1987; Colpron & Nelson 2009), Hadlari *et al.* (2014) proposed that the Pearya terrane was incompletely separated from Laurentia by a failed Neoproterozoic rift, and thus it remained in a peri-Laurentian position until subsequent collisional events closed the intervening basins. Further east, Caledonian age zircon source areas are recognized, such as in East Greenland, Svalbard, New England, western Newfoundland, and the British Isles (e.g. Kalsbeek *et al.* 2008; Rehnström 2010; Pettersson *et al.* 2009, van Staal and Barr 2012).

Devonian zircon (found in Signatures C, D and G-P Belt)

Local sources for Devonian age zircon in the AACM (Fig. 4) are abundant and include bimodal volcanic rocks within the Schist Belt (375-385 Ma; Raterman *et al.* 2006,

McClelland *et al.* 2006), orthogneiss bodies within the Schist and Central Belts and in the adjacent Ruby and Seward Peninsula terranes (385-400 Ma; Aleinikoff *et al.* 1993, Patton *et al.* 1987, Toro *et al.* 2002; Amato *et al.* 2009, 2014; potentially some as young as 360-365 Ma in Dillon *et al.* 1980, Patton *et al.* 1987), orthogneiss and granite bodies present beneath the North Slope and in the northeastern Brooks Range (360-375 Dillon *et al.* 1987; Bird *et al.* 1978; Mortensen & Bell 1991; Lane *et al.* 1993; Lane 2007; T. Moore, unpublished data) and in Chukotka (350-365 Ma; Natal'in *et al.* 1999; Amato *et al.* 2014; Lane *et al.* 2015; Luchitskaya *et al.* 2015). Beyond the AACM Devonian rocks have been recognized widely, such as in Pearya (390 & 370 Ma, Trettin 1987; Trettin *et al.* 1992).

Detrital zircon data from this study confirm the importance of Devonian igneous bodies as sources for contemporaneous basins and younger sediments across northern Alaska and provide some additional insight into the character and age of that magmatism. Specifically, combining detrital zircon data from all samples helps characterize the timing of Devonian igneous activity, defining probability maxima at ~365, ~380, and ~420 Ma, suggesting that the Devonian was characterized by punctuated magmatic events. Hf data for Devonian zircons (this study) indicate progressively more evolved (sub-CHUR) ϵHf_t values through time (Fig. 6B), which requires progressively more assimilation of crustal material into the source regions. The final pulse of magmatism at ~365 Ma are possibly pure S-type crustal melts (approaching $\epsilon\text{Hf}_t = -20$). The age and isotopic composition of this late evolved population is most similar to granites of the northeastern Brooks Range and Yukon (e.g. Lane 2007; Nelson *et al.* 1993).

The southern Brooks Range terranes: peri-Baltican versus peri-Laurentian?

The southwestern subterrane of the AACM have been considered to have a non-Laurentian origin on the basis of palaeontological affinities and magmatic ages atypical of Laurentia (e.g. Dumoulin *et al.* 2002; Amato *et al.* 2014), but non-unique interpretations of such datasets exist permitting either peri-Baltican (e.g. Miller *et al.* 2011) or peri-Laurentian origins (e.g. Hadlari *et al.* 2014) and several ideas on how they may have evolved through time (e.g. Colpron & Nelson 2011).

Because there are no pre-Neoproterozoic rocks identified in the AACM (e.g. Amato *et al.* 2014) characterization of its basement relies on inferences from sedimentary and isotopic signatures of cover sequences. A major challenge in metamorphic terranes such as the Schist Belt is that transposition fabrics have obscured the sedimentary structures and/or facies information which otherwise may have helped to assess sedimentary transport, distances, directions, and recycling. If the detrital zircons were derived from sedimentary units shed locally from underlying basement, then the signatures discussed here would be directly indicative of the basement of the terrane. However, since the zircons may have been transported long-distances in a sedimentary system and/or by sedimentary recycling, it is plausible that their provenance area is unrelated to the basement on which they are now found (e.g. Moore *et al.* 1997c). Thus, we cannot say anything definitive about the pre-Neoproterozoic basement terranes of the AACM. Instead, we can focus on constraining their palaeogeographic associations during early Palaeozoic time, prior to

their accretion to form the AACM, and make inferences as to their origin. We note that while matching detrital zircon affinities to a cratonic source area does not require palaeogeographic proximity, it does require one or more of the following: 1) sedimentary communication with the craton (i.e. not separated by an ocean basin or a drainage divide); 2) erosion of allochthonous fragments of the craton; and/or, 3) reworking of sediments formerly shed from the craton; and thus in context is useful for discrimination of cratonic affinity.

Ambiguity arises in defining a “Laurentian” provenance signature due to possible reworking of massive and widespread Rodinia-age clastic packages that were eroded from both Laurentian and Baltican sources within the Grenville orogen (Rainbird *et al.* 1992, 1997, 2012). These Mesoproterozoic packages blanketed Laurentia and provide a source for Baltican signatures despite Baltica having been rifted away in late Neoproterozoic time. Lane & Gehrels (2014), expanding upon work by Hadlari *et al.* (2012), proposed four detrital zircon “lineages” as characterizing late Neoproterozoic to early Palaeozoic sediments deposited on northwestern Laurentia: A) sediments sourced from “Sequence A” (i.e. Wernecke Supergroup and Hornby Bay Group, 1.6-2.1 Ga and 2.2-2.8 Ga), B) sediments sourced from “Sequence B” (i.e. Mackenzie Mountains and Shaler Supergroups, dominated by 1.0-1.25 Ga, 1.25-1.6 Ga, and 1.6-1.75 Ga populations), C) sediments sourced directly from the Laurentian craton (i.e. 1.75-2.2 Ga, 2.2-2.4 Ga, 2.5-2.8 Ga), and D) sediments representing mixtures of sources A-C. Thus, whereas a key characteristic of Laurentia commonly referenced in provenance studies is a noted gap of magmatism between 1.49–1.61 Ga (e.g. Van Schmus *et al.* 1993), both the persistence of sedimentary cover containing zircons of these ages (as in the Shaler supergroup) in Northern Laurentia (e.g. Lane & Gehrels 2014) and the presence of non-North American-aged crystalline rocks within the Grenville orogen of eastern Canada (e.g. Pinware terrane, Heaman *et al.* 2004) complicates the use of such criteria. The “Sequence B” type Laurentian signature (e.g. Rainbird *et al.* 1997) is perhaps distinguished from true Baltican signatures only by the relative abundance of Palaeoproterozoic and Archean grains (>1.6 Ga grains comprise less than 10% in most Baltican/AACM sediments whilst they comprise greater than 10% in all of the subsequent Laurentian signatures)(Fig. 7).

Comparison of (para-) autochthonous strata of both Baltica and Laurentia from their separation in the late Neoproterozoic until their reassembly in the mid-Palaeozoic (Fig. 7) provides a baseline against which to compare allochthonous terranes (e.g. Hadlari *et al.* 2012). The most unambiguously lower Palaeozoic strata (and thus the most robust constraint on early Palaeozoic palaeogeographic affinity) from this study are the Signature A and B strata (Fig. 4), but because Signature C exhibits a similar pre-mid-Palaeozoic age distribution (Fig. 7A) it is also used in our comparisons. The signature of Laurentia is represented here by considering the detrital zircon spectra of unambiguously Laurentian strata of Upper Neoproterozoic to Lower Devonian age in: 1) The Franklinian Basin (Anfinson *et al.* 2011; Beranek *et al.* 2013a; Hadlari *et al.* 2014; Kirkland *et al.* 2009), 2) The Mackenzie Mountains and NW Laurentia (Leslie 2009; Gehrels & Pecha 2014; Hadlari *et al.* 2012), 3) The NW Cordilleran passive margin (data set of Gehrels and Pecha, 2014), 4) The eastern Laurentian margin in Newfoundland (Cawood &

Nemchin 2001); and 5) The North Slope subterrane of northeast Arctic Alaska which although displaced is confirmed to be Laurentian by trilobite and conodont fossil assemblages (Macdonald *et al.* 2009; Strauss *et al.* 2013; Lane *et al.* 2015; Johnson *et al.* 2016). The late Neoproterozoic-early Devonian signature of Baltica is represented by various autochthonous and parautochthonous strata in: Norway (Kirkland *et al.* 2011; Gee *et al.* 2014; Andresen *et al.* 2014; Zhang *et al.* 2015, 2016), Estonia (Isozaki *et al.* 2014; Polvedere *et al.* 2014), and the Russian Urals (Miller *et al.* 2011). Data from the non-Laurentian portions of the AACM are shown for comparison: Seward Peninsula (Amato *et al.* 2009; Till *et al.* 2014a), Ruby terrane (Bradley *et al.* 2007), Chukotka (Gottlieb *et al.* this volume). Also shown for comparison: sample 12JT22 of Clarence River group, northeastern Brooks Range (Johnson *et al.* 2016) and sample J1426 of Central Belt unit Trembley Creek phyllite (Strauss *et al.* 2017).

From rifting in the Neoproterozoic until re-assembly in mid-Palaeozoic time, the >850 Ma spectra remains relatively consistent despite differing degrees of early Palaeozoic magmatic inputs into the system, which enables us to compare our samples of unknown age to the persistent signatures of both Baltica and Laurentia (Fig. 7). Signatures A-C are consistent with samples from the Seward Peninsula, Ruby terrane, Chukotka, and Baltica and are notably different from Laurentian baselines (Fig. 7). Our only samples whose detrital spectra appear consistent with a Laurentian lineage are the samples of Greywacke-Phyllite Belt and signature D of the Schist Belt (Fig. 7), which contain abundant grains from 1.8-2.5 Ga. The abundance of upper Palaeozoic grains in these signatures implies post-Romanzof/Caledonian deposition and therefore is consistent with renewed association to Laurentia by mid-Devonian time.

The earliest rift deposits associated with the breakup of Rodinia and the opening of the Iapetus Ocean (i.e. base of “Assemblage 4” of Cawood *et al.* 2007: e.g. Grampian Group of the Dalradian Supergroup in Scotland, Lyell Land and Nathorst Land groups of the Eleanor Bay Supergroup of East Greenland, Murchisonfjorden Group of Svalbard) exhibit detrital signatures (e.g. Slama *et al.* 2011; Cawood *et al.* 2003, 2007) similar to the early Palaeozoic signatures attributed above to Baltica and AACM rocks in that they contain very few Archean age grains. These deposits are siliciclastic-dominated and variations in their distribution, thickness, and facies appear to be structurally controlled (Cawood *et al.* 2007 and references therein). These early rift deposits may be sourced from Baltica directly or else were reworked from underlying Baltica-derived early Neoproterozoic successions (e.g. Moine Supergroup, Krummedal succession) having a similar age distribution (Cawood *et al.* 2007) and presumably exposed in rift shoulders. While offering a potential source for the zircon signature identified in early Palaeozoic strata in the AACM, these early Baltica-derived rift deposits would have been buried by subsequent development of the passive margin around Laurentia. In contrast, those deposits appear to have continued to be reworked in Baltica, such as reported by Zhang *et al.* (2015, 2016) for northerly-derived sediments (present day coordinates) deposited in the foreland basin to the Timanides (see their sample STP 1), implying continued reworking at least in latest Ediacaran-Cambrian time. A Laurentian origin may still be possible for AACM signatures if tectonism and subduction continued outboard of Laurentia where correlative Meso- and early Neoproterozoic sedimentary packages may

have been rifted away as part of ribbons or microcontinents such as inferred for the Precordillera (Thomas & Astini 1996; Dalziel 1997) and proposed for Pearya (Hadlari *et al.* 2014; Malone *et al.* 2014). Such an outboard arc system may provide a mechanism for continued reworking of Baltican signatures in a peri-Laurentian realm. Note that the “Cambrian of northern Laurentia type II” of Hadlari *et al.* (2012) and the Franklinian and deep basin signatures of Hadlari *et al.* (2014) all contain a significant amount of Archean age zircon, in strong contrast to its scarcity in Pearya and the AACM (e.g. Malone *et al.* 2014; Hadlari *et al.* 2014). Whether such rifted fragment(s) would have remained proximal to Laurentia (as in Hadlari *et al.* 2014; Malone *et al.* 2014) or been displaced long distances and associated with other continents (e.g. Timanides) before their return should be the focus of future studies.

In summary, integrating a variety of data sets is necessary to distinguish Baltican versus Laurentian crustal fragments. In particular, previous studies have highlighted the following distinguishing characteristics: 1) The lack of 1.49–1.62 Ga igneous ages within Laurentia (e.g. Van Schmus *et al.* 1993), but as discussed above use of this criteria is complicated by the persistence of detrital grains of that age in Neoproterozoic strata and by igneous bodies of that age in Grenvillian allochthons emplaced onto Laurentia; 2) The paucity of 0.75–0.55 Ga igneous activity within Laurentia (i.e. Crockerland/Arctida/Timanian-age magmatism), however this criteria is also compromised by recognition of similar ages within Laurentia and by the possible persistence of peri-Laurentian arcs of this age (e.g. Hadlari *et al.* 2014); 3) the widespread recognition of Mesoproterozoic Nd model ages uncommon to Laurentia (e.g. Nelson *et al.* 1993; Amato *et al.* 2014), except that such data offers non-unique solutions; and 4) The presence of fauna of Siberian-Alaskan endemic form in early Palaeozoic carbonates (e.g. Dumoulin *et al.* 2002), however use of this criteria is complicated by palaeogeographic ambiguity as well as the fact that the nature of the basement beneath the carbonate platform remains unconstrained (e.g. Dumoulin *et al.* 2014). As discussed above we have suggested two additional features that may help differentiate Baltican from Laurentian origin terranes: 5) Distinctions in the crustal reworking trajectories (i.e. zircon Hf-isotope trends) of Sveconorwegian versus Grenville versus Valhalla orogens (e.g. Spencer *et al.* 2017) can help differentiate the original source areas of Mesoproterozoic age detritus; and 6) Subaerially exposed Archean zircon sources (whether primary or recycled) were abundant on Laurentia (e.g. Cawood & Nemchin 2001), but not so on Baltica, during their separation in the late Neoproterozoic until reunification in the mid-Palaeozoic (e.g. Cawood *et al.* 2007). In consideration of these combined data sets, a Baltican origin for the southwestern AACM, though not necessarily unique, remains permissible and more likely.

Relation to other exotic terranes of the Cordillera

While a review and comparison to Cordilleran terranes (e.g. Alexander, Doonerak, Farewell, Klamaths, etc.)(Fig. 8) is beyond the scope of this paper, we note that a growing body of data points to their shared origin in the high Arctic (e.g. Colpron & Nelson 2011; Beranek *et al.* 2013b; Bradley *et al.* 2014). The timing and mechanisms of their dispersal remains poorly constrained, but U-Pb and Hf zircon fingerprints have

proven a powerful tool to unraveling this complex history (e.g. Beranek *et al.* 2013b; Tochilin *et al.* 2014; Malone *et al.* 2014; White *et al.* 2015; Malkowski and Hampton, 2014; Pecha *et al.* 2016; Strauss *et al.* 2017). We note that the great size of the AACM and its central location between the Cordilleran terranes (Fig. 8) and their presumed origins in the Atlantic Caledonides and Baltica suggests that many of those terranes would have interacted with and/or been in communication with the AACM in the early Palaeozoic and ultimately may have themselves rifted away from the AACM (e.g. Devonian-Mississippian rifting, Miller *et al.* 2011; Ordovician rifting, Till *et al.* 2014b).

Locating mid-Palaeozoic sutures within the Brooks Range

Recognizing pre-Devonian strata of (peri-) Baltican affinity in the southern Brooks Range narrows the domain within the AACM wherein might lie a mid-Palaeozoic suture between Laurentian and Baltican crustal fragments (Fig. 8), and thus may be of use as piercing points in tectonic reconstructions (e.g. O'Brien *et al.* 2016; Miller *et al.* this volume)(Fig. 8). It appears clear that Artic Alaska is composed of three distinct pre-Devonian crustal provinces (from north to south, present day coordinates): 1) a now displaced fragment of northern Laurentia that underlies the northeastern Brooks Range and presumably also the North Slope Basin (the “North Slope subterrane”, e.g. Strauss *et al.* 2013); 2) a perhaps only partially preserved peri-Laurentian island arc system evidenced by igneous and volcanoclastic packages in the Doonerak Window and Whale Mountain allochthon (the “Apoon arc subterrane”, e.g. Strauss *et al.* 2017; Johnson *et al.* 2016); and 3) a microcontinent of unknown size and formation age (herein called the “southern Brooks Range subterrane” or simply “southern Arctic Alaska”) that potentially belonged to the microcontinent that collided with Baltica during the late Neoproterozoic-Cambrian Timanide orogeny and then also to have collided with Laurentia during the Early-Middle Devonian Romanzof orogeny.

Because the present day contact of southern Brooks Range subterrane strata with peri-Laurentian Apoon arc subterrane strata on the south flank of the Doonerak Window is an Early Cretaceous thrust and because the rocks within the Doonerak Window are themselves displaced by Brookian duplexing, the actual mid-Palaeozoic lithospheric tectonic boundary must lie somewhere to the south once Brookian allochthons are restored (e.g. Moore *et al.* 1997a; Strauss *et al.* 2017). Moore *et al.* (1997a) argued that the EMA and Central Belt are linked as hanging wall units and that they must be restored south of the Doonerak Window. Whether the Schist Belt then restores north or south of the EMA/Central Belt remains unconstrained (e.g. Moore *et al.* 1997b) but our data suggests that pre-Devonian strata are abundant in both the Schist and Central Belt and so they may be correlative except for having undergone divergent Mesozoic metamorphic histories (e.g. Till 2016). Regardless, all of the allochthonous strata of the Brooks Range would have been deposited on top of the southern Brooks Range subterrane, which when Brookian structures are restored implies a significant pre-Cretaceous geographic area for the southern Brooks Range subterrane. As it would have been larger than what remains today (e.g. Mayfield *et al.* 1993), some of the terrane’s basement must have been subducted into the mantle in the Early Cretaceous or else subsequently displaced to other parts of Alaska. A further complexity exists in that if early Palaeozoic strata of the

southern Brooks Range terrane were emplaced onto the Laurentian margin as rootless thrusts during mid-Palaeozoic orogenesis, then their original basements could have translated or rifted away in late Palaeozoic time to become allochthons of the Cordillera (e.g. Miller *et al.* 2011).

Schematic palaeotectonic history (Fig. 9)

A simplified Phanerozoic 2-D schematic tectonic history of the southern Brooks Range terrane is depicted in Figure 9 to illustrate possible associations between the terrane and cratonic regions with respect to detrital zircon samples in this study. **Signature A** of the Schist Belt may have been deposited within marginal rift grabens flanking Baltica associated with the breakup of the supercontinent Rodinia in late Neoproterozoic time. It is interpreted to have sourced remnant Sveconorwegian highlands and their associated clastic aprons, and/or the Baltica craton, and potentially near contemporaneous rift-related volcanic rocks. Alternatively, a peri-Laurentian origin (i.e. correlative with Pearya, as in Hadlari *et al.* 2014; Malone *et al.* 2014) is possible if the Sveconorwegian/Baltica zircon was sourced by reworking of Mesoproterozoic clastics that were elsewhere buried at this time in northern Laurentia but may have been differentially exposed and eroding in a peri-Laurentian realm. **Signature B** of the Schist Belt may be latest Precambrian and sourcing zircons from the accretionary terranes of the Timanide microcontinent. **Signature C** of the Cosmos Hills may represent erosion of strata with Signature A or B affinities but with the addition of early Palaeozoic igneous sources such as the Doonerak, Chukotka, or similar arc systems that served to close the Iapetus Ocean. Arc magmatism ceased due to collision (i.e. Romanzof) in late Silurian-Devonian time (Moore *et al.* 1994; Lane 2007). Subduction was re-established in middle Devonian time on the proto-Pacific outboard margin of the newly formed Laurussian continent (e.g. Dillon *et al.* 1980; Rubin *et al.* 1990) which induced backarc extension (Rateman *et al.* 2006). **Signature D** of the Schist Belt was likely deposited on the south-facing passive margin that formed as this new Devonian arc system pulled away from the continental margin. It would have received detritus from the north from what were likely predominantly Laurentian source areas in the remnant orogenic highlands. Backarc spreading formed the Angayucham ocean (e.g. Pallister *et al.* 1989). **Signature G** of the Greywacke-Phyllite Belt would have been deposited sometime after the formation of this south-facing passive margin and ocean, possibly continuing into Triassic time, receiving primarily (peri-) Laurentian detritus. Around this time, a major tectonic transition occurs as the Angayucham ocean begins to close and invert, ultimately generating the Brookian orogeny. Although the polarity of subduction during the Mesozoic Brookian orogeny is still debated (e.g. Miller & Hudson, 1991; Moore *et al.* 1994; Till, 2016), in Late Cretaceous time a Pacific subduction system retreats southward (Akinin *et al.* 2009; Miller *et al.* this volume). Extension in the Brooks Range is matched by rapid subsidence in the Yukon-Koyukuk basin where sediments are sourced from rift flank shoulders of proximal subterranean and the contemporaneous volcanism associated with arc retreat and/or flareup. **Signature Y** samples were deposited into this successor basin and represent erosion of uplifted and exhumed strata of the southern Brooks Range Schist Belt and/or Central Belt.

CONCLUSIONS

U–Pb detrital zircon data from this study identify a range of probable Neoproterozoic to Devonian age protoliths within the polydeformed and metamorphosed southern Brooks Range Schist Belt (Fig. 4). Specifically, the Schist Belt and Cosmos Hills samples reveal four different signatures: 1) metasediments having an early Neoproterozoic (Tonian) maximum depositional age (Signature A) exhibiting a range of detrital zircon ages from ~1.0–1.8 Ga and a small Archean population; 2) metasediments having a late Neoproterozoic (Ediacaran) maximum depositional age (Signature B) characterized by a range of detrital zircon ages from ~1.0–1.8 Ga, a small Archean population, and a 550–720 Ma population; 3) metasediments having an Ordovician–Devonian (?) maximum depositional age (Signature C) with a spread from 1.0–2.0 Ga and an abundance of Neoproterozoic and Cambrian to early Devonian (?) age zircon; 4) metasediments having a mid-to-late Devonian maximum depositional age (Signature D) characterized by an abundance of Devonian age and Ordovician–Silurian age zircon, with apparent maxima at ~1.0, ~1.8, and ~2.6 Ga. These therefore characterize the pre-Mississippian basement rocks involved in Brookian orogenesis that now form the Central and Schist Belts of the southern Brooks Range (Fig. 8). These new data sets and those published previously for the adjacent Seward and Ruby terranes support previously proposed correlations that link pre-Mississippian rocks across this broader portion of northern Alaska (Till *et al.* 2014a; Amato *et al.* 2009; Bradley *et al.* 2007)(Fig. S4) and to broader Chukotka (Gottlieb *et al.* this volume). Greywacke–Phyllite Belt samples of presumed post-middle Devonian, though possibly Triassic or Carboniferous maximum depositional age, are characterized by an abundance of Palaeozoic age zircon and a pre-850 Ma spectra marked by peaks of ~1.1 Ga and ~1.9 Ga age detritus. Sediments of the Yukon–Koyukuk Basin of presumed Late Cretaceous age have peaks at ~90, ~380, 410–450, ~470, and 600–740 Ma, with a spread from 1.0–1.8 Ga that rises to a distinct maxima at ~1.65 Ga, with then a few Archean age zircon (Fig. 4). These U–Pb age constraints from detrital zircon expand our information base on pre-Devonian rocks in the Schist Belt of the Brooks Range and together with Hf isotope geochemistry is able to place additional constraints on the origin and evolution of this portion of the AACM, supporting a non-Laurentian early Palaeozoic position for the southern Brooks Range Schist and Central Belts that are inferred to have once been in sedimentary communication with (if not in close proximity to) the Timanides and Baltica, but which later exhibits a renewed association to Laurentia by at least mid-Devonian time (Fig. 8, Fig. 9).

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Figure Captions

Fig. 1. (a) Index map showing present day tectonic configuration of the northern circum-Pacific orogenic belts (modified from Dickinson, 2009, which was after Dickinson *et al.* 1986). Mercator projection has pole at 25°N, 15°E. A, Alaska; C, Chukotka; G, Greenland; J, Sea of Japan; O, Sea of Okhotsk; T, Taiwan (adjacent island). (b) Index map of the Pacific Arctic (adapted from Miller *et al.* 2011) showing locations of prominent Neoproterozoic to Mesozoic orogenic belts with legend colors indicating dominant deformational age. C, Caledonides; C, Chukchi Borderland; CB, Canada Basin; E, Ellesmerianides; LR, Lomonosov Ridge; NZ, Novaya Zemlya; SAZ, South Anyui Suture Zone; SV, Svalbard; SZ, Severnaya Zemlya, T, Timanides. (c) Generalized geology and structural features of the Alaska portion of the Arctic Alaska–Chukotka microplate (AACM) (modified from Moore *et al.* 2015). Intrusive rocks are not shown. Edge of continental shelf in Arctic Ocean depicted at 200 m isobath. Terrane abbreviations: AM, Angayucham; KY, Koyukuk; FW, Farewell; RT, Ruby; YTT, Yukon-Tanana; KB, Koyukuk Basin. Original map sources: Moore *et al.* (2000), Grantz *et al.* (2011), Nokleberg *et al.* (2000), Patton *et al.* (2009).

Fig. 2. Tectonic domain and indicator mineral map of the southern Brooks Range (modified from Till *et al.* 2008) with generalized cross section (modified from Toro *et al.* 2002) showing location and context of geologic maps shown in A and B: (a) Geology of Wiseman quadrangle region between John River and the Dalton highway (modified from Dillon *et al.* 1986; Wilson *et al.* 2015), showing all Schist Belt and one Greywacke-Phyllite Belt sample locations. (b) Simplified geologic map of Ambler district and Cosmos Hills, showing Cosmos Hills sample locations (adapted from Selby *et al.* 2009, Till *et al.* 2008). Extensional detachment faults are interpreted between the Schist Belt and Greywacke-Phyllite Belt as well as between the Greywacke-Phyllite Belt and Angayucham Belt (as in Christiansen 1989; Christiansen & Snee 1994; Little *et al.* 1994; Law *et al.* 1994). Terrane and unit lithologies denoted by color as in legend at bottom. Purple numbers indicate U-Pb zircon ages for orthogneiss as in Dillon *et al.* 1979, 1980; Aleinikoff *et al.* 1993; Raterman *et al.* 2006; McClelland *et al.* written communication in Till *et al.* 2008). Sample localities shown from this study are numbered sequentially (as in Table 1): 1, (1988 PGBR) CH-118; 2, (PGBR) 87-9; 3, (PGBR) 87-1E; 4, (1988

PCBR) CH-107; 5, (PCBR) 87-6B; 6, (ELM) CH-1D; 7, 13-CH-BR03b; 8, 13-CH-BR04/A; 9, ELM-13-BR22/A/B; 10, 13-CH-BR09 (MS4); 11, 90-LL-71; 12, 90-LL-33; 13, 89 BRLM-98; 14, AK07 SA-4; sample AK07 SA-2A is located beyond map area to the south along the Dalton highway.

Fig. 3. Secondary electron (left) and color cathodoluminescence (right) images of representative zircon from sample ELM-13-BR22/A/B from the John River of the Wiseman quadrangle showing relict oscillatory-zoned cores of detrital origin overgrown by inclusion-rich zircon rims. Each zircon image was digitally cropped from the original data set, resized to a common scale, and aligned together for convenience. U-Pb ages (Ma) assessed at the NordSIM facility, Stockholm, Sweden, are annotated adjacent the SIMS ablation pits.

Fig. 4. Kernel density estimation (KDE) plots of detrital zircon results from this study stacked in relation to inferred structural position as depicted at left. Superscripts on sample names indicate the laboratories (listed at top) in which each sample was studied. Shaded fields and labels indicate important age domains discussed in text: 80-100 Ma, 350-400 Ma, 410-530 Ma, 550-740 Ma, 900-1800 Ma, 1800-2000 Ma. Diagram was constructed in R with the use of Provenance (Vermeesch *et al.* 2016).

Fig. 5. (a) Zircon U-Pb age plotted against Th/U from Phanerozoic zircon combined from all samples, showing a notable difference in chemistry between metamorphic overgrowths (~110 Ma) and all detrital grains (including the ~90 Ma population identified within the two youngest samples). Grey fields drawn by eye highlight the concordant groupings. Th/U values of < 0.1 are a common characteristic of metamorphic zircon (e.g. Rubatto, 2002). **(b-c)** Concordia diagrams of zircon U-Pb isotopic data for the youngest detrital population identified in Koyukuk basin sediments structurally juxtaposed against the southern flank of the Cosmos Hills. **(b)** Sample (1988 PCBR) CH-118 is a pelitic sandstone deformed and metamorphosed at chlorite grade conditions. Inset shows weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age results. Red and blue ellipses indicate data collected during two different sessions at the Laserchron facility. **(c)** Sample (PCBR) 87-9 is a slaty phyllite, with inset showing weighted mean results. Grey ellipses indicate discordant data that was not used in final age calculations. Diagram was constructed with the use of Isoplot (Ludwig, 2008). Error ellipses are shown at 2-sigma. MSWD, mean square of weighted deviates.

Fig. 6. $\epsilon\text{Hf}(t)$ vs. U-Pb age of detrital and metamorphic zircon from this study compared to reference data sets; **(a)** plot scaled from 0 to 3000 Ma (a few analyses from highly discordant zircons are not shown) with reference to data from Spencer *et al.* (2017); **(b)** scaled sub-plot highlighting Palaeozoic-late Neoproterozoic age populations with reference to data from Gottlieb *et al.* (this volume), Strauss *et al.* (2017). DM, depleted mantle (Vervoort & Blichert-Toft, 1999), CHUR, chondritic uniform reservoir (Bouvier *et al.* 2008); black arrow shows interpreted crustal evolution trajectory assuming present day $^{176}\text{Lu}/^{177}\text{Hf} = 0.0115$ (Vervoort & Patchett, 1996; Vervoort *et al.* 1999).

Fig. 7. Cumulative probability plots of detrital zircon results from reference samples of Laurentia, Baltica, and the AACM, compared to data from the southern Brooks Range. The persistence of unique cratonic signatures on Baltica and Laurentia from the time of Rodinia rifting to Pangean assembly permits the comparisons of >850 Ma spectra without knowing the precise depositional age. Only detrital ages that are <10% discordant, ≤5% reversely discordant and have <10% age uncertainty, were used. Plots include only ages >850 Ma ($^{207}\text{Pb}/^{206}\text{Pb}$ ages were used) and include measured age uncertainty. Data sources are indicated in plots. The red dotted line highlights an apparent difference in relative abundance of >2.0 Ga age zircon in early Palaeozoic sediments of Baltica versus Laurentia. Such >2.0 Ga ages never comprise less than 8% of the total spectra for Laurentian signatures considered here, whereas such high abundance is rare in Baltican sediments.

Fig. 8. Index map showing location of Arctic Alaska-Chukotka microplate (AACM) in relation to other key regions of the Arctic and Cordillera (adapted from Miller *et al.* 2011, with “Arctic”-affinity terranes of the Cordillera as in Colpron & Nelson, 2009; 2011; Gasser, 2013). References for the location and age of basement rock types in the Arctic: Pearya (Arctic Canada) (Hadlari *et al.* 2012, 2014); Chukchi Borderland (Brumley, 2014; Brumley *et al.* 2015; O’Brien & Miller, 2014); North Slope (Strauss *et al.* 2013); Arctic Chukotka, Russia and Alaska (Amato *et al.* 2009; 2014; Miller *et al.* 2006 and 2011). A, Alexander terrane; C, Chukchi Borderland; CB, Canada Basin; E, Ellesmerian deformation front; F, Farewell terrane; K, eastern Klamath terranes; KB, Kilbuk terrane; LR, Lomonosov Ridge; NZ, Novaya Zemlya; O, Okanagan terrane; S, northern Sierra terranes; SAZ, South Anyui Suture Zone; SV, Svalbard; SZ, Severnaya Zemlya.

Fig. 9. Simplified Phanerozoic 2-D schematic tectonic history of the southern Brooks Range terrane illustrating possible palaeogeographic settings for detrital zircon samples (this study) in context of the broader geologic history of the AACM (schematic style after Collins *et al.*, 2011). Line of section varies between time slices to approximately parallel the prevailing transport direction. Consequently, the possibly adjoined Baltica craton is out of the plane of view in Phanerozoic time slices. See text for details. B, Baltica; L, Laurentia; sAA, southern Arctic Alaska terrane(s).

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