**$^{40}\text{Ar}/^{39}\text{Ar}$ ages and geochemical characterization of Cretaceous bentonites in the Nanushuk,
Seabee, Tuluvak, and Schrader Bluff formations, North Slope, Alaska**

Authors

Grant T. Shimer (corresponding)¹, Jeff A. Benowitz², Paul W. Layer^{2,3}, Paul J. McCarthy³,
Catherine L. Hanks³, Marwan Wartes⁴

1. Geology Department, Whitman College, 345 Boyer Avenue, Walla Walla, WA 99362

2. Geophysical Institute, 903 Koyukuk Drive, University of Alaska Fairbanks, Fairbanks, AK,
99775

3. Department of Geosciences, 900 Yukon Drive, University of Alaska Fairbanks, AK, 99775

4. Alaska Division of Geological and Geophysical Surveys, 3354 College Road, Fairbanks, AK
99709

Corresponding author contact information:

Email: shimergt@whitman.edu

Phone: 509-527-5944

Abstract

Diagenetically altered volcanic ash deposits (bentonites) found in Cretaceous terrestrial and marine foreland basin sediments have the potential to be used for chronostratigraphy and subsurface correlation across Alaska's North Slope. Detailed age and geochemical studies of these volcanogenic deposits may also shed light on the tectonic evolution of the Arctic. Though these bentonites have been previously studied, there are few published results for regional bentonite ages and geochemistry due to challenges of dating weathered volcanic ash. We analyzed mineral separates from cored bentonites recovered from wells in the National Petroleum Reserve Alaska. The analyses confirm that an intense period of volcanic ash deposition on Alaska's North Slope began by the late Albian and was persistent throughout the Cenomanian, a period of rapid progradation and aggradation in the Colville basin. These results also add to a sparse record of radioisotopic ages from the Nanushuk Formation. A bentonite preserved in delta plain sediments in the upper Nanushuk Formation dates to 102.6 ± 1.5 Ma (late Albian), while a bentonite near the base of the overlying Seabee Formation was deposited at 98.2 ± 0.8 Ma, in the early Cenomanian. The two ages bracket a major flooding surface at the base of the Seabee Formation near Umiat, Alaska, placing it near the Albian-Cenomanian boundary (100.5 Ma). Several hundred feet up-section, the non-marine Tuluvak Formation contains bentonites with $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 96.7 ± 0.7 to 94.2 ± 0.9 Ma (Cenomanian), several million years older than previously published K-Ar ages and biostratigraphic constraints suggest.

Major and trace element geochemistry of a sub-sample of six bentonites from petroleum exploration wells at Umiat show a range in composition from andesite to rhyolite, with a continental arc source. The bentonites become more felsic from the late Albian (~102 Ma) to late Cenomanian (~94 Ma). A likely source for the bentonites is the Okhotsk-Chukotka Volcanic Belt (OCVB) of eastern Siberia, a continental arc which became active in the Albian and experienced episodes of effusivity throughout the Late Cretaceous. Chronostratigraphically anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ ages coincide with peaks of magmatic activity in the OCVB, suggesting that these anomalously old ages may be due to magmatic contribution of xenocrysts or recycling of detrital minerals from older volcanic events. An alternative explanation for the chronostratigraphically anomalous ages is mixing of bentonites with detrital sediment derived

45 from unroofing and erosion of metamorphic rocks in the Brooks Range, Herald Arch, and
46 Chukotka throughout the mid to Late Cretaceous.

47

1. Introduction

Though biostratigraphy provides broad age control for Late Cretaceous marine and terrestrial clastic sedimentary rocks of the Colville foreland basin on Alaska's North Slope, current age estimates range over millions of years and do not provide precise dates for high-resolution correlation of sedimentary deposits. Cretaceous bentonites are a component of the Colville foreland basin stratigraphy (Mull et al., 2003; Fig. 1) and provide an opportunity for precise radiometric dating. Bentonites are ubiquitous in many outcrops in the Brooks Range foothills, but are particularly prominent at locations such as Umiat Mountain (Anderson and Reynolds, 1966; Lanphere and TAILLEUR, 1983; Houseknecht and Schenk, 2005), and Shale Wall Bluff (Detterman et al., 1963; Fig. 1). Bentonites are also preserved in cores from the numerous legacy wells in the National Petroleum Reserve Alaska (NPR) and other federal drilling sites on the North Slope (i.e., Gubik No. 1 and No. 2 wells; Fig. 1). Despite their ubiquity, there is little published information on the radiometric ages and geochemistry of these bentonites. The purpose of this paper is to report new $^{40}\text{Ar}/^{39}\text{Ar}$ dates from several Cretaceous bentonite deposits, classify the bentonites using bulk X-ray fluorescence spectroscopy (XRF), and evaluate the chronostratigraphic and chemostratigraphic significance of the results in respect to Cretaceous basin evolution in Arctic Alaska.

We analyzed thirteen bentonites from the Nanushuk, Seabee, Tuluvak, and Schrader Bluff formations (Fig. 2). The samples come from one outcrop location and five exploratory wells in the Brooks Range foothills of the central North Slope, Alaska (Fig. 1). We discuss the wide range of ages within the context of known biostratigraphy (Tappan, 1960; Mull et al., 2003; Spicer and Herman, 2010) and regional volcanic events (Moore et al., 1994; Tikhomirov et al., 2012; Akinin et al., 2014). In conjunction with $^{40}\text{Ar}/^{39}\text{Ar}$ age interpretations, we classify magma composition and tectonic origin for six bentonites from the Umiat wells using major and trace element geochemistry.

2. Geologic Background

The North Slope of Alaska includes the northern Brooks Range foothills and the Arctic coastal plain (Schindler, 1988). The bedrock geology of the North Slope consists of Franklinian pre-

Mississippian sedimentary and low grade metamorphic basement and several overlying megasequences: the passive margin Ellesmerian sequence (Mississippian-Triassic); the syn-rift Beaufortian sequence (Jurassic-Early Cretaceous) associated with the opening of the Canada Basin; and the Brookian sequence (Cretaceous-Cenozoic) of sediments derived from uplift and denudation of the Brooks Range, Chukotka, and the Herald Arch, and deposited in the Colville foreland basin (Hubbard et al., 1987; Moore et al., 1994; Bird, 2001; Houseknecht et al., 2012). The Umiat, Square Lake, and Gubik wells in this study are located in the northern foothills of the Brooks Range in folded and faulted Brookian sediments of the foreland basin (Fig. 1). The basin initially formed in response arc-continent collision in the Middle Jurassic to Early Cretaceous, while deformation of the basin fill occurred in the early Cenozoic (Cole et al., 1997; Moore et al., 2004). Early Cretaceous greenschist-facies metamorphism overprints Jurassic metamorphism in the Brooks Range schist belt to the south, evidence for ongoing extension and exhumation in the southern Brooks Range throughout the Cretaceous (Christiansen and Snee, 1994).

We use the 2013 IUGS geologic time scale (Cohen et al., 2013) for numerical ages for Cretaceous age boundaries. The stratigraphic boundaries for the formations in this study were originally determined using megafossil biostratigraphy: most of the Nanushuk Formation falls within the middle Albian or later, as determined by the presence of the ammonite *Gastrolites* sp. (middle Albian) and the bivalve *Inoceramus anglicus* Woods (late Albian), which are similar to species found in Alberta and British Columbia (Imlay, 1961). The Cenomanian is most clearly identified by the presence of *I. dunveganensis*, also known from localities in western Canada (Jones and Gryc, 1960). Supplemental microfossil data, especially foraminifera zones, were also crucial for age differentiation within the Nanushuk Formation and identification of the Albian-Cenomanian boundary (Tappan, 1960; LePain et al., 2009).

We investigated bentonites in the Albian-Cenomanian (113-93.9 Ma) Nanushuk Formation, Cenomanian-Coniacian (100.5-86.3 Ma) Seabee and Tuluvak formations, and Santonian-Maastrichtian (86.3-66.0 Ma) Schrader Bluff Formation (Fig. 2). The Nanushuk Formation is part of a larger Torok-Nanushuk depositional sequence, separated from the Seabee-Tuluvak sequence by a major flooding surface (Decker, 2007). The Torok and Seabee formations are distal shelf to slope and basin mudstones, whereas the Nanushuk and Tuluvak are fluvial, deltaic, and shallow

marine deposits (Mull et al., 2003; LePain et al., 2009). The Schrader Bluff Formation is marine in origin and interdigitates with the alluvial-deltaic Prince Creek Formation (Flores et al., 2007). All of these formations are time-transgressive and generally young towards the east, with the most recent ages for each formation found to the east of the study area (Mull et al., 2003). Though clinoform foreset dips support west to east migration of depositional sequences, there is also a south to north component as determined from paleocurrent data, particularly in the central-eastern Brooks Range foothills (Huffman et al., 1985; Molenaar, 1985; LePain et al., 2009).

Burial and diagenesis can influence both $^{40}\text{Ar}/^{39}\text{Ar}$ ages and geochemical compositions. The maximum burial depth for the Nanushuk Formation near Square Lake No. 1 was approximately 2000 m (Smosna, 1989), and the Nanushuk Formation was thought to have undergone low-temperature, low-pressure diagenesis with a maximum burial temperature of 60 °C (Smosna, 1988; O'Sullivan, 1999). More recent vitrinite reflectance data show a more complicated picture, with up to 8000 ft (2440 m) of denudation in the Square Lake region, and up to 7000 ft (2130 m) in the Umiat and Gubik areas (Houseknecht et al., 2010; Houseknecht et al., 2012). Given the shallow burial, Ar-bearing minerals are not likely to have experienced significant argon-loss.

3. Materials and Methods

The wells used in this study were part of the 1944-1953 Naval Petroleum Reserve No. 4 drilling program (Schindler, 1988). Drilling reports from the Umiat (Collins, 1958), Square Lake (Collins, 1959), and Gubik (Robinson, 1958) sites contain lithologic descriptions, well logs, petrography, biostratigraphy, and engineering data (porosity, permeability, oil, gas, and water analyses, and drilling operations). These records were used to locate bentonites for sampling. Samples were recovered from cores stored at the Geologic Materials Center in Eagle River, Alaska. The wells and cores are described in foot increments, so all sample labels include the well site and number code, a depth in feet, and a code for formation. For example, the code UM11-1962S refers to a sample recovered from Umiat No. 11, at 1962 ft (598 m) below the Kelly bushing, within the Seabee Formation. A single outcrop sample (CI-006S) was taken from a bluff along the Colville River southwest of Umiat, Alaska (Fig. 1). The outcrop sample was

recovered at the top of a 180 ft. (55 m)-thick stratigraphic section, ~30 ft. (9 m) above the base of the Seabee Formation, and consisted of unconsolidated grey bentonite with abundant biotite grains. Unlike the outcrop sample, all core samples were lithified. Bentonites in the Seabee, Tuluvak, and Schrader Bluff formations are yellow with no visible mineral phases, whereas the Nanushuk Formation samples are white to grey, with rare biotite present. The outcrop sample from the lower Seabee Formation was biotite-rich.

3.1 $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology

All samples were washed and sieved to produce disaggregated bentonite residue for mineral separation. Separates were magnetically sorted, and datable mineral phases (biotite, sanidine, plagioclase, and volcanic glass) were hand-picked under a binocular microscope. Many of the samples had minimal datable mineral grains. Samples and standards of monitor mineral MMbh-1 (Sampson and Alexander, 1987) with an age of 513.9 Ma (Lanphere and Dalrymple, 2000) were irradiated in the research reactor at McMaster University, Hamilton, Canada, for 20 megawatt-hours.

In the University of Alaska (UAF) geochronology lab, each irradiated sample was loaded into a 2 mm diameter hole in a copper tray, which was then placed into an ultra-high vacuum extraction line. Monitors were fused and samples were heated using a 6-watt argon-ion laser following established techniques (York et al., 1981; Layer et al., 1987; Benowitz et al., 2014). Argon purification was achieved using a liquid nitrogen cold trap and SAES Zr-Al getter at 400 °C. Samples were analyzed in a VG-3600 mass spectrometer. Measured argon isotope values were corrected for system blanks and mass discrimination, as well as calcium, potassium, and chlorine interference reactions (McDougall and Harrison, 1999). System blanks were 2×10^{-16} mol ^{40}Ar and 2×10^{-18} mol ^{40}Ar (10 to 50 times smaller than fraction volumes). Mass discrimination in the system is routinely monitored by running calibrated air shoots and zero-age glass sample measurements on a weekly to monthly basis.

Additional analyses or "runs" were conducted to reduce error on any samples with a sufficient amount of mineral separates. Multiple runs were not possible in most cases due to a limited amount of datable material. We only report second runs for SQ1-257T and GU2-1387T because previous runs were unsuccessful at producing an $^{40}\text{Ar}/^{39}\text{Ar}$ age due to insufficient amount of

material run. We report both runs of UM2-311, and use the combination of the two runs to calculate a higher precision weighted average age.

3.2 Geochemistry

Six samples were selected for X-ray fluorescence (XRF) analysis to determine major and selected trace element compositions. All of the samples are from the Umiat wells. The samples include two bentonites from each of the Nanushuk (UM11-2258N; UM2-311N), Seabee (UM11-1962S; UM11-1682S), and Tuluvak (UM11-391T; UM11-345T) formations. Due to the small volume of material in the cores, we were limited to the <150 µm residues remaining after the crushing and sieving process for $^{40}\text{Ar}/^{39}\text{Ar}$ dating. Five residues from Umiat No. 11 and one from Umiat No. 2 were analyzed. There was sufficient original material from Umiat No. 2 to generate a complete, unsorted "whole rock" sample of the bentonite for comparative purposes.

All samples were processed to create 39.7 mm diameter pressed powder disks. 12.0-15.0 g of bentonite were mixed with 1.2-1.5 g of PXR-200 powder (binding agent). Powder weight was set at 10% of sample weight. To uniformly crush and homogenize the samples, each mixture was placed in a carbon-steel ball mill for 7 minutes, with 10 mL of Vertrel™ as a lubricant. 11.0 g of dried sample was weighed into a steel press and samples were compressed under a pressure of ~4000 kg/cm² (10 tonnes/in²) for 5 minutes.

For trace element analysis, we used the PROtrace 37 routine on a PANalytical Axios XRF in the Advanced Instrumentation Laboratory at UAF. Six standards were run: BE-N (basalt), GSP-1 (granodiorite), JA-2 (andesite), JG-2 (granite), JP-1 (peridotite) and JR-1 (rhyolite). To assess standards, relative percent difference was calculated for barium, cerium, chromium, cobalt, copper, lanthanum, lead, nickel, niobium, scandium, strontium, rubidium, vanadium, yttrium, zinc, zirconium, and titanium oxide using experimental values and preferred values from the GeoReM database (<http://georem.mpch-mainz.gwdg.de/>).

4. Results

4.1 $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology

All $^{40}\text{Ar}/^{39}\text{Ar}$ results were calculated using the constants of Steiger and Jaeger (1977) and are reported to $\pm 1\sigma$ (Table 1). Isochron ages were incalculable for most samples due to homogenous radiogenic content of released and documented argon-loss. Several age spectra depict gas release with an age-stepping-up pattern, which we interpret as argon-loss (see section 5.1). Isochron ages were not determined for samples that exhibited argon-loss. Integrated ages and step-heating age spectra are reported, and are useful for interpreting plateau ages and recognizing argon-loss, recoil, or potential excess argon.

Plateau ages can be determined if three or more consecutive gas fractions are within two standard deviations of each other. The $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages in this study range from ~ 133.1 to ~ 89.3 Ma (Table 1). Of note, the maximum biostratigraphic age range of the sampled strata is 112-85.3 Ma. Most of the analyses have spectra with relatively flat plateaus and errors of ± 1.5 Ma or less, but a single sample from the Schrader Bluff Formation (GU1-736SC) had a plateau error of ± 4.4 Ma. Plateaus were not calculated for samples CI-006S and GU1-2845S. The age from CI-006S is not a true plateau age because the amount of gas released from the more retentive steps only included 45.3% of the total ^{39}Ar release. A plateau was not calculated for GU1-2845S because the mineral separate analyzed was low in potassium, produced an irregular age spectrum, and did not reach the requirements for a plateau age calculation.

We present $^{40}\text{Ar}/^{39}\text{Ar}$ results stratigraphically, from the Nanushuk Formation to the Schrader Bluff Formation. Bentonite samples from the Nanushuk Formation have ages of 110.4 ± 0.9 Ma (UM11-2258N), 102.6 ± 1.5 Ma (UM2-311N#1) and 102.2 ± 1.1 (UM2-311N#2). All of these ages are Albian (113-100.5 Ma), and technically fall within the biostratigraphic framework of the formation.

Bentonite samples from the overlying Seabee Formation have ages of 106.0 ± 0.9 Ma (GU1-2840S), 98.2 ± 0.8 Ma (CI-006S), 131.1 ± 1.3 Ma (UM11-1962S), and 104.1 ± 1.1 Ma (UM11-1682S). These are Hauterivian (~ 131.1 Ma), Albian (~ 106.0 and ~ 104.1 Ma), and Cenomanian (~ 98.2 Ma). The Hauterivian and Albian ages exceed the biostratigraphic age of the Seabee Formation. In addition to these samples, GU1-2845S (also from the Seabee Formation) returned a chronostratigraphically unrealistic $^{40}\text{Ar}/^{39}\text{Ar}$ integrated age of 398.7 ± 43.8 Ma.

Bentonite samples from the Tuluvak Formation, which represents the topset portion of the Seabee-Tuluvak depositional sequence (Houseknecht and Schenk, 2005), have plateau ages of 133.1 ± 2.2 Ma (GU1-1165T), 96.2 ± 2.0 Ma (GU2-1397T), 96.6 ± 1.4 Ma (SQ1-257T), 94.4 ± 0.9 Ma (UM11-345T), and 129.1 ± 1.8 Ma (UM11-391T). The ages are Valanginian (~133.1 Ma), Barremian (~129.1 Ma), and Cenomanian (~96.2, ~96.6, and ~94.4 Ma). The Valanginian and Barremian ages significantly exceed the biostratigraphic age of the Tuluvak Formation.

A single bentonite from the overlying Schrader Bluff Formation has an age of 89.3 ± 4.4 Ma (GU1-736), which falls within the Turonian (89.8-93.9 Ma). The error for this sample is sufficiently large that it may have a Coniacian (89.8-86.3 Ma) or early Cenomanian (100.5-93.9 Ma) age.

4.2 Geochemistry

Most major and trace elements were present in detectable concentrations (Table 2). Only two major elements and oxides (MnO, S) and four trace elements (Ge, Sb, Ta, W) were below detection limits for any of the samples. Of these, only Ta is used in the classification diagrams, because only one sample (UM11-345) had Ta below detection limit. For other elements used in characterization diagrams, SiO₂ compositions range from 58.35-72.02%, whereas K₂O (0.18-1.36%) and Na₂O (2.23-3.30%) are present in much lower concentrations. TiO₂ (1567-7788 ppm), Zr (189-309 ppm), Nb (5.41-12.47 ppm), Ta (1.59-1.79 ppm), Th (20.13-35.39 ppm), Y (15.87-38.56 ppm), and Yb (2.80-5.31 ppm) also show a range of values among samples (Table 2).

We assessed the relative error for elements or oxides used in characterization diagrams by comparing standards to recommended values in the GeoReM database (Table 3). Percent error can be very high for some elements, particularly where published values are at or below detection limits (~1 ppm for most elements). Errors range from 0.0% (compiled recommended values same as measured) up to 399% for elements with very low concentrations. Errors were not large enough to exclude the crucial major and trace elements from use in diagrams.

5. Discussion

5.1 A Note on geologic ages and the utility of age spectra

There are five $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages in this study that we interpret to constrain geological ages based on the plateau characteristics and biostratigraphically determined formation boundaries (Tappan, 1960; Jones and Gryc, 1960; Imlay, 1961) and age limits of the Nanushuk, Seabee, and Tuluvak formations (Mull et al., 2003; Cohen et al., 2013). Four of the five samples have plateau ages (UM11-345T; SQ1-257T; GU2-1397T; UM2-311N) and a fifth (CI-006S) has sufficiently low error and flat release to be considered reliable. In the following discussion we interpret these ages. In section 5.3 we suggest that the remaining chronostratigraphically anomalous ages are relevant in the broader context of regional tectonic activity and Brookian sedimentation.

Several of the age spectra show argon-loss, characterized by an age stepping-up pattern, including UM2-311N, CI-006S, and GU2-1397T (Fig. 3A and 3C; Fig. 4C). These spectra show lower than average ages in less retentive, low-temperature steps. This is likely due to loss of argon, which can occur around crystal edges through diffusion (McDougall and Harrison, 1999). Biotite is prone to argon-loss through diffusion along cleavage planes (Harrison et al., 1985), and argon-loss was only detected in biotite samples (UM2-311N #1, UM2-311N #2, CI-006S, GU2-1397T). Despite argon-loss, the spectra from UM2-311N and GU2-1397T meet the plateau age criteria of three or more consecutive gas fractions that represent >50% of total gas release and have mean standard weighted deviations <2.5 (Table 1).

5.2 Albian-Cenomanian Bentonites: Nanushuk, Seabee, and Tuluvak formations

We discuss these results from a stratigraphic perspective, beginning with the Nanushuk Formation. The first published radiometric age from the Nanushuk Formation is from sample UM2-311N in Umiat No. 2 (this paper). As discussed in section 3.1, there were two runs of UM2-311N (Fig. 3A and 3B). The plateau ages from both runs of UM2-311N are within error (Table 1). UM2-311N #1 produced a plateau age of 102.6 ± 1.5 Ma; UM2-311N #2 produced an age of 102.2 ± 1.1 Ma. A weighted mean age of these two runs is 102.4 ± 1.0 Ma, placing the delta plain deposits of the upper Nanushuk Formation in the late Albian. In the Umiat area, the lower Nanushuk Formation is thought to be mid-Albian in age based on plant fossils (Spicer and Herman, 2010) and ammonites (Imlay, 1961). A late Albian $^{40}\text{Ar}/^{39}\text{Ar}$ age from the upper

Nanushuk Formation is consistent with the biostratigraphy, as the underlying sediments of the lower Nanushuk are separated from the upper Nanushuk Formation by a major flooding surface and over 100 m of shale (Shimer et al., 2014; Fig. 8). The results presented here represent a more precise age for the thick, time-transgressive upper Nanushuk Formation at this location, and demonstrate the utility of $^{40}\text{Ar}/^{39}\text{Ar}$ dating of bentonites when sample sizes are sufficient.

CI-006S is the only sample from the Seabee Formation with an $^{40}\text{Ar}/^{39}\text{Ar}$ age that we consider to constrain the geologic age of the formation. This sample, just above the base of the Seabee Formation, has an age of 98.2 ± 0.8 Ma, placing it in the early Cenomanian. Approximately 25 km to the northeast, a major flooding surface lies between the upper Nanushuk Formation and Shale Wall facies of the Seabee Formation (Houseknecht and Schenk, 2005; Shimer et al., 2014). We interpret a similar flooding surface at the Colville site. Together with the new ages from the upper Nanushuk Formation, the lower Seabee Formation age brackets the timing of the preceding transgression to between 102.4 ± 1.0 Ma (Albian) and 98.2 ± 0.8 Ma (Cenomanian). This age is near the Albian-Cenomanian boundary (100.5 Ma), and is older than the major transgression associated with the late Cenomanian-early Turonian anoxic event (Prokoph et al., 2001). Consequently, it may be related to late Albian transgressions recognized worldwide (Hancock and Kauffman, 1978).

There are three Tuluvak Formation ages in this study from three different sites, and all have ages significantly older (Cenomanian) than the Turonian age commonly cited for the Tuluvak Formation (Lanphere and TAILLEUR, 1983; Mull et al., 2003). The samples, from west to east, are SQ1-257T (Fig. 4B: 96.6 ± 1.4), UM11-345T (Fig. 4A: 94.4 ± 0.9 Ma), and GU2-1397T (Fig. 4C: 96.2 ± 2.0 Ma). The youngest age (UM11-345T) is stratigraphically lowest (Fig. 5), raising questions either the significance of the ages (discussed below) or the assumptions about stratigraphic order. Stratigraphically, these inconsistent results may suggest northward prograding terrestrial topset clinoforms, or simply the result of a series of thin deposits separated by sequence-bounding unconformities in a low accommodation setting (Houseknecht, pers. comm.) .

For these three Tuluvak Formation bentonites, each of the dated samples were from a different mineral phases, which might explain the apparent stratigraphic discrepancy. UM11-345T is a

suspected sanidine or high-K plagioclase age, based on the Ca/K ratio (>1), whereas GU1-1397T is a biotite age, and SQ1-257T is a volcanic glass age (Fig. 4). Plagioclase has lower argon retention than biotite (McDougall and Harrison, 1999), which makes the young age from UM11-345T the most likely outlier. Volcanic glass, like that found in SQ1-257T, can be prone to argon-loss during devitrification (McDougall and Harrison, 1999), but SQ1-257T has a flat release spectrum with no evidence for argon-loss (Fig. 4B). Unaltered glass shards have also been dated using K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ near Ocean Point in the overlying Prince Creek Formation, and these also show little argon-loss (Conrad et al., 1990; Flaig, 2010; Flaig et al., 2014). Though older than expected, all three Tuluvak Formation ages have robust age spectra and fall within biostratigraphic constraints (100.5-93.9 Ma), which suggests that the Tuluvak Formation is older than previously thought in the study area. Considering the discussion below, further investigation of Tuluvak Formation geochronology is needed to better assess regional age variability in respect to both stratigraphic architecture and analytical error.

5.3 Chronostratigraphically Anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ Ages

We obtained six anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ ages (Fig. 6 and Fig. 7) that meet the criteria for good plateau ages, yet are not geologically realistic considering the known biostratigraphic boundaries (Tappan, 1960), chronologic age constraints (Mull et al., 2003; Cohen et al., 2013), published $^{40}\text{Ar}/^{39}\text{Ar}$ and K-Ar ages (Lanphere and Tailleur, 1983; Conrad et al., 1990), and other ages obtained in this study. All of the chronostratigraphically anomalous ages discussed in this section are significantly older than the known biostratigraphy, but there is no overall trend based on analytical procedure, stratigraphy, or depositional environment that can be used to explain the older ages.

To discuss our interpretations of these samples we divide our discussion the time periods in which the ages fall. These chronostratigraphically anomalous ages are Albian (113-100.5 Ma; Fig. 6A-C), Barremian (125-129.4 Ma; Fig. 7A), Hauterivian (129.4-132.9 Ma; Fig. 7B), and Valanginian (132.9-139.8 Ma; Fig. 7C).

There are three anomalous Albian $^{40}\text{Ar}/^{39}\text{Ar}$ ages (GU1-2840S, 106.0 ± 0.9 Ma; UM11-1682S, 104.1 ± 1.1 Ma; and UM11-2258N, 110.4 ± 0.9 Ma). Both UM11-1682S (Fig. 6A) and GU1-2840S (Fig. 6C) are from the Seabee Formation. Based on the date obtained from CI-006S

(Table 1), known biostratigraphic constraints for Cenomanian (100.5-93.9 Ma) Seabee Formation in the Umiat area (Tappan, 1960; Mull et al., 2003), and age established boundaries (Cohen et al., 2013), these ages are as much as 10 myr too old.

UM11-2258N is the second bentonite dated from the Nanushuk Formation. Though still within the Albian (113-100.5 Ma), UM11-2258N returns an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 110.4 ± 1.0 Ma, approximately 8 myr older than the two runs of UM2-311N (weighted average 102.4 ± 1.0 Ma), which is stratigraphically lower in the Nanushuk Formation (Fig. 8). As with the other chronostratigraphically anomalous Albian samples, we can find no analytical, stratigraphic, or sedimentological explanation for the older ages. Like UM2-311N, UM11-2258N experienced minor argon-loss, but this is accounted for in the $^{40}\text{Ar}/^{39}\text{Ar}$ plateau (Fig. 3A and 3B; Fig. 6B; Table 1). Both samples are from non-marine to marginal marine delta plain sediments near the top of the Nanushuk Formation, so we do not suspect differences in sorting or diagenesis. Though Umiat No. 11 well is heavily faulted, previously described faults are not located at these depths (Collins, 1958). Furthermore, stratigraphic sections of the Nanushuk Formation for Umiat No. 2 and Umiat No. 11 (Fig. 8) are very similar, which suggests that there is not significant faulting throughout the interval.

Lacking analytical or depositional explanations for these older than expected, chronostratigraphically anomalous ages, especially for UM11-2258N, we suggest several alternative hypotheses related to bentonite contamination during eruption, transport, and deposition: the incorporation of xenocrysts from the magmatic source areas, or bentonite mixing with exhumed detrital minerals from igneous or metamorphic sources, either from the Brooks Range, as in Toro et al. (1998), or from distant sources in Chukotka. It would only take a few older high potassium grains to bias the $^{40}\text{Ar}/^{39}\text{Ar}$ results, as our processing method reduced most of the bentonites to a very small sample of datable grains that we assumed were air fall phenocrysts. If there was some detrital mixing, grains could be incorporated into the bentonites during deposition in marine or lacustrine environments dominated by suspension settling of fine particles (such as micas), or perhaps mixed in through bioturbation or soft sediment deformation. In a related study from the North Slope, researchers noted age bias in a devitrified tephra in the informally named “Pebble Shale”, an age bias caused by the presence of detrital muscovite. The

364 muscovite caused a return in chronostratigraphically anomalous ages similar to bedrock ages in
365 the source areas (Benowitz and Layer, 2014).

366 Xenocrysts with older $^{40}\text{Ar}/^{39}\text{Ar}$ ages could also be a source of bias for the $^{40}\text{Ar}/^{39}\text{Ar}$ age of a
367 sample. Older ages are possible in igneous systems where previously cooled xenocrysts can be
368 incorporated into magma without outgassing preexisting radiogenic argon or re-equilibrating
369 with atmospheric argon before eruption (Gillespie et al., 1982). A potential source region with
370 appropriately aged eruptive events is the Okhotsk-Chukotka Volcanic Belt (OCVB) in eastern
371 Siberia (Fig. 9). The OCVB has five main episodes of volcanic activity, with the oldest episode
372 occurring at ~108-98 Ma (Tikhomirov et al., 2012; Akinin et al., 2014). Both the
373 chronostratigraphically anomalous and geologically reasonable $^{40}\text{Ar}/^{39}\text{Ar}$ ages from this study
374 cluster around peaks of activity in the OCVB (Fig. 10).

375 As for the detrital bedrock grain contamination hypothesis, the potential sources for detrital
376 biotite and muscovite in the Nanushuk and Seabee formations are the ancestral Brooks Range,
377 Chukotka, and the Herald Arch. The dominant source areas for Brookian sediments appear to
378 have shifted over time, from Chukotka to the Brooks Range, based on detrital zircon provenance
379 (Lease et al., 2014). Metamorphic rocks in the Brooks Range would be the most likely source of
380 Albian-age detrital micas. Blueschist facies metamorphic rocks from the southern Brooks Range
381 (Fig. 9) were exhumed from the Late Jurassic to middle Albian. White micas from these
382 deposits have $^{40}\text{Ar}/^{39}\text{Ar}$ ages as old as ~120 Ma (Till, 1992), and metamorphic events at ~110-
383 105 Ma (Till and Patrick, 1991) and ~105-103 Ma (Vogl et al., 2002) have also been recognized.
384 Additional studies report $^{40}\text{Ar}/^{39}\text{Ar}$ ages from white mica in metamorphic rocks that indicate
385 remobilization around ~108 Ma (Till and Snee, 1995). Since Aptian-Albian chloritoid and
386 muscovite from blueschist facies metamorphic rocks are found in the Nanushuk Formation (Till,
387 1992), these detrital sediments may be the source of relict mica ages found in mixed-age
388 bentonites.

389 We have identified potential xenocryst or detrital material sources above, but cannot directly link
390 these potential contaminants to our chronostratigraphically anomalously old dates. If there is a
391 detrital component, the contamination is not thought to completely overprint the volcanic
392 $^{40}\text{Ar}/^{39}\text{Ar}$ ages, nor highly distort the XRF geochemistry. Unfortunately, there is no definitive
393 sedimentary evidence of mixing in the bentonites, and we have no comparative geochemistry of

individual mineral grains. We suggest this potential source of contamination in the interest of motivating future studies of detrital $^{40}\text{Ar}/^{39}\text{Ar}$ ages on the North Slope. Given the documentation of anomalous old bentonite $^{40}\text{Ar}/^{39}\text{Ar}$ ages and the likelihood of xenolith or detrital contributions being responsible for these anomalous old bentonite ages, it is possible many or all of our bentonite age determinations should be considered maximum depositional ages. However, in many cases these bentonite ages approximate the actually age of deposition based on corroborating data.

To further expand on the xenocryst/detrital contamination hypotheses that may be responsible for the older than expected, chronostratigraphically anomalous ages, three of the samples in this group have even older Early Cretaceous ages. These include UM11-391T (129.1 ± 1.8 Ma; Barremian), UM11-1962S (131.2 ± 1.3 Ma; Hauterivian) and GU1-1165T (133.1 ± 2.2 Ma; Valanginian). All three ages are from biotite separates. The spectra (Fig. 7A-C) do not display “saddle” shapes or stepping-down age patterns, which are possible indications of excess argon in retentive high temperature sites (Harrison, 1990). Biotite grains with excess argon are also known to produce flat release patterns if the excess argon is uniformly distributed throughout the mica (McDougall and Harrison, 1999). We interpret excess argon as a possible explanation for these chronostratigraphically anomalous old biotite ages. Alternatively, these ages may reflect a xenolith component, and this interpretation is consistent with the timing of a minor peak in volcanic activity during the Barremian-Hauterivian in the OCVB (Akinin and Miller, 2011; Fig. 10). Again, an alternative explanation for the chronostratigraphically anomalous ages is the mixing of the volcanogenic material with a detrital component from the Brooks Range schist belt (Fig. 9), which experienced greenschist-facies metamorphism during the Early Cretaceous (Christiansen and Snee, 1994). Without additional analyses of individual mineral grains it is difficult to evaluate these hypotheses. Nevertheless, the Tuluvak and Seabee formation bentonites were recovered from different terrestrial and marine sediments, so it is highly unlikely that the same degree of detrital contamination would occur in these settings.

5.4 Unknown Phases

GU1-736SC, from the Schrader Bluff Formation, has an $^{40}\text{Ar}/^{39}\text{Ar}$ release pattern that meets the criteria for a plateau age (Table 1; Fig. 11). However, that $^{40}\text{Ar}/^{39}\text{Ar}$ age (89.3 ± 4.4 Ma) places

the sample at the Turonian-Coniacian boundary, significantly older than the Santonian-Maastrichtian (83.6-66 Ma) age for the formation (Mull et al., 2003). We do not interpret the $^{40}\text{Ar}/^{39}\text{Ar}$ age to reflect the depositional age of the strata. We also do not propose an age revision for the Schrader Bluff Formation, because the mineral separate dated had a Ca/K ratio > 500, and it is difficult to obtain an accurate and precise age on low potassium mineral separates (Fig. 11A). This also means the sample is not a pure plagioclase separate, and without additional mineralogical information it is not possible to place the age in a geologic context.

GU1-2845S is unique among all the analyses, in that no $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age or even semi-precise calculated age was possible. The age spectrum is very irregular and clearly does not form a plateau (Fig. 11B). In this case the cause is clear: the Ca/K ratio is >1000. The inferred plagioclase separates from this sample may have been unexpected calcite or aragonite crystals, and the integrated age of 398.7 ± 43.8 Ma is geologically meaningless. We include the spectra to contrast with the chronostratigraphically anomalous ages that we interpret as geologically significant.

5.5 Geochemistry

We classified the bentonites using traditional igneous geochemistry diagrams in order to determine whether there was significant variation in bentonite composition through time. Many variables influence bentonite geochemistry, including original magmatic variation, wind and water sorting of the volcanic ash by density, particle size, and shape, and chemical alteration through surface weathering and burial diagenesis. We use immobile trace element geochemistry to minimize diagenetic factors and focus on the Umiat wells to track changes from the Albian Nanushuk Formation (UM11-2258N; UM2-311N) through the Cenomanian Seabee (UM11-1962S; UM11-1682S) and Tuluvak formations (UM11-391T; UM11-345T) at a single location. The bentonites in the Umiat wells are bracketed by $^{40}\text{Ar}/^{39}\text{Ar}$ ages from UM2-311N (102.4 ± 1.0 Ma) and UM11-345T (94.4 ± 0.9 Ma), though the 94.4 Ma age is older than expected given the biostratigraphic control, probably due to reduced argon retention in plagioclase (see discussion above). These dates allow us to characterize changes in bentonite composition from the late Albian (113-100.5 Ma) through the entire Cenomanian, if error margins are considered (100.5-93.9 Ma).

One of the restricting factors in the geochemical study was sample size. Most of the available bentonite was used in the $^{40}\text{Ar}/^{39}\text{Ar}$ dating process for many of the samples. We used the $<150\text{ }\mu\text{m}$ size fraction (residue) from each of the $^{40}\text{Ar}/^{39}\text{Ar}$ samples in this study to produce XRF disks. There was concern that the size fraction would influence results through the removal of larger phenocrysts. There was sufficient “whole rock” sample from UM2-311N to compare bulk compositions of the unsieved bentonite to the $<150\text{ }\mu\text{m}$ fraction. A cross-plot of trace element geochemistry from the whole rock and $<150\text{ }\mu\text{m}$ residue shows a correlation, though not a definitive one (Fig. 12A). We cautiously interpret this result as evidence for moderate geochemical fractionation through size grading, and infer that the XRF results from the pressed disks are largely representative of the bulk geochemistry of the bentonites. In contrast, a comparison of the trace element geochemistry of the $<150\text{ }\mu\text{m}$ size fractions from two different Nanushuk Formation bentonites shows no correlation between the two samples (Fig. 12B)

The only existing X-ray diffraction study of the Umiat bentonites found evidence for both montmorillonite and beidellite, which are both smectites (Anderson and Reynolds, 1966). Smectites are known to undergo Si loss (relative to Al), in a transition to illite through burial and diagenesis (Land et al., 1997). It is, therefore, important to compare major element geochemistry to trace element concentrations to properly characterize a bentonite or a mineral phase within a bentonite. Various studies of bentonites emphasize different immobile elements, such as Nb, Zr, Cr, Ni, V, La, Ce, Nd, and Y (Christidis, 2001), as well as Ti and Th (Kiipli et al., 2008). These are generally high field strength elements or large ion lithophiles (Fanti, 2009).

We plotted the major element data on a traditional total-alkali-silica (TAS: Fig. 13). The TAS diagram shows that samples from the Nanushuk (UM2-311), Seabee (UM11-1962; UM11-1682) and Tuluvak (UM11-391) formations plot as distinct groups in the andesite and dacite fields (Fig. 13A). UM11-2258 plots closest to the dacitic Seabee Formation bentonite. Temporarily ignoring UM11-2258, there is a trend of increasingly felsic (silica-rich) volcanic ash deposits from the Albian (UM2-311: $102.4 \pm 1.0\text{ Ma}$) to the end of the Cenomanian (UM11-354T: $94.4 \pm 0.9\text{ Ma}$) that may relate to evolution of Arctic arc magmatism throughout the mid to Late Cretaceous.

We use immobile element concentrations and ratios to characterize magmatic source material (Winchester and Floyd, 1977; Pearce et al., 1984; Pearce, 1996; Hastie et al., 2007) and infer tectonic setting (Pearce, 1983; Pearce and Peate, 1995). The trace element studies are designed with various proxies for assessing mafic to felsic composition (Zr/TiO_2) and alkalinity (Nb/Y). Increasingly high Zr/TiO_2 ratios show a transition from mafic (basaltic) to felsic (rhyolitic) magmas, whereas increasing Nb/Y shows an increase in alkalinity (Pearce, 1996). Trace element plots of the Umiat bentonites show a pattern of increasingly felsic compositions through the Albian-Cenomanian (Fig. 13B). The results are consistent with the TAS diagram (Fig. 13A), except that each sample plots as slightly more felsic than shown in the TAS diagram, with Tuluvak Formation samples plotting nearly in the rhyolite field. Since the increase in SiO_2 content appears to be systematic over time (with the exception of UM11-2258N), the bentonite geochemistry may be an indicator of magmatic system evolution throughout the Albian-Cenomanian. Si is often lost through burial and diagenesis (Land et al., 1997), and the probable loss of Si in these bentonites emphasizes the utility of immobile element diagrams where Zr serves as a proxy for original Si content.

Using the Rb vs Y+Nb classification system for plutonic rocks (Pearce et al., 1984), all of the Umiat bentonites plot as volcanic arc magmas (Fig. 14A). Volcanic arcs are likely to produce explosive eruptions with volcanic ash. Given the shift in values for the residue and whole rock samples from UM-311N, we also plotted results on a second diagram to further classify the bentonites. In the Th/Yb vs Ta/Yb plot (Fig. 14B), all of the samples plot in the continental arc field. These results help to narrow the potential source areas for the Umiat bentonites. The Early Cretaceous plutons of the Yukon-Koyukuk basin, to the south side of the Brooks Range (Fig. 9), are tholeiitic to shoshonitic, likely sourced from a volcanic island arc (Box and Patton, 1989; Miller, 1989). In contrast, the OCVB was part of an active continental margin with a range of basaltic-andesitic-dacitic-rhyolitic compositions during the Albian-Cenomanian (Tikhomirov et al., 2006; Belyi, 2008). Furthermore, the timing of the Umiat bentonites compares well to periods of volcanic activity in the OCVB (Fig. 10), particularly with events in the northern OCVB, including the Western Okhotsk and Anadyr segments (Fig. 9; Tikhomirov et al., 2006; Akinin and Miller, 2011; Akinin et al., 2014). Consequently, as suggested by previous workers (e.g., Bergman, 2006), the OCVB appears to be the most likely source for bentonite deposits in the Albian-Cenomanian Brookian sediments of Alaska's North Slope.

6. Conclusions

$^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages from bentonites at four sites in the Nanushuk, Seabee, and Tuluvak formations of northern Alaska's Colville foreland basin range from late Albian (102.4 ± 1.0 Ma) to late Cenomanian (94.4 ± 0.9 Ma). These ages are consistent with stratigraphic positions, and add radiometric data points to the existing biostratigraphic framework. This assemblage of ages includes the first radiometric dates from the upper Nanushuk Formation. The upper Nanushuk Formation age of 102.4 ± 1.0 Ma and a 98.2 ± 0.8 Ma age from the base of the overlying Seabee Formation bracket a major flooding surface very close to the Albian-Cenomanian boundary (100.5 Ma).

Six of the bentonites analyzed in this study returned chronostratigraphically anomalous dates that meet the criteria for a plateau age determination. These samples produced small amounts of datable mineral grains, and the anomalous ages were all older than the stratigraphic positions. Given the other chronostratigraphic constraints, bentonite radiometric ages that are anomalously old should be considered maximum depositional ages. To explain the chronostratigraphically anomalous ages, we suggest that these samples may be contaminated with older xenocrysts during eruption, or with detrital mineral grains during transport and deposition. The likely presence of older inherited grains heavily bias the final $^{40}\text{Ar}/^{39}\text{Ar}$ ages. If this hypothesis is correct, apparent phenocrysts in mineral-poor bentonites may actually be inherited minerals, and we encourage future researchers to compare the geochemistry of individual grains to identify possible variation due to the contribution of xenocrysts or detrital grains. Unfortunately this is difficult with the small bentonite samples from these cores, which is why we were not able to conduct these analyses. If detrital contamination can be proven for the foreland basin bentonites, chronostratigraphically anomalous ages may be an effective means of studying unroofing and erosion of Brookian source areas in the Brooks Range, Chukotka, and the Herald Arch.

The combination of $^{40}\text{Ar}/^{39}\text{Ar}$ ages and geochemical data from bentonites at Umiat indicate that the likely source for Albian-Cenomanian tephras on the North Slope is the Okhotsk-Chukotka Volcanic Belt (OCVB) in eastern Siberia. Albian to Cenomanian dates from the bentonites compare well with known periods of effusive volcanic activity in the OCVB, and though no

direct correlation is possible with the existing data set, the continental arc origin of the andesitic to rhyolitic Brookian bentonites also points towards Chukotka. The Umiat bentonites become increasingly felsic over time, indicating a change in magmatic composition in the source area. These results are promising, and suggest a more intensive study of Cretaceous bentonite samples from cores should be useful for studying Albian-Cenomanian volcanic activity in the Arctic.

7. Acknowledgements

Funding for the $^{40}\text{Ar}/^{39}\text{Ar}$ analyses came from the U.S. Department of Energy contract DE-FC26-08NT005641. Additional funding for $^{40}\text{Ar}/^{39}\text{Ar}$ dating came from the Geological Society of America (GSA) and the UAF Geology BP Scholarship Program. Samples were collected with the help of Ken Papp of the Alaska Geologic Materials Center in Eagle River, Alaska. Special thanks to Dr. Ken Severin of the Advanced Instrumentation Laboratory (AIL) at UAF for instrumental time, and to Dr. Maciej Sliwinski (AIL) for assistance with data processing and interpretation. Steven Bergman of Shell provided access to data for context and comparison. This manuscript also benefited from the helpful suggestions of the manuscript reviewers, especially David Houseknecht of the U.S. Geological Survey.

8. References

- Akinin, V.V., Layer, P., Benowitz, J., and Ntaflos, T., 2014, Age and composition of final stage of volcanism in Okhotsk-Chukotka volcanic belt: an example from the Ola Plateau (Okhotsk segment): ICAM VI: Proceedings, St. Petersburg. VSEGEI, p. 19-60.
- Akinin, V.V., and Miller, E.L., 2011, Evolution of calc-alkaline magmas of the Okhotsk-Chukotka Volcanic Belt: Petrology, v. 19, p. 237-277.
- Anderson, D.M., and Reynolds, R.C., 1966, Umiat Bentonite: An Unusual Montmorillonite from Umiat Alaska: American Mineralogist, v. 51, 1443-1455.
- Belyi, V.F., 2008, Problems of geological and isotopic age of the Okhotsk-Chukotka Volcanogenic Belt (OCVB) : Stratigraphy and Geological Correlation, v. 16, p. 639-649.

568 Benowitz, J.A., and Layer, P.W., 2014, $^{40}\text{Ar}/^{39}\text{Ar}$ step heat analysis of core from the N. Kalikpik
 569 Test Well #1: Data Report #426, Alaska Geologic Materials Center, State of Alaska
 570 Department of Natural Resources, 7 p.

571 Benowitz, J.A., Layer, P.W., and Vanlaningham, S., 2014, Persistent long-term (c. 24 Ma)
 572 exhumation in the Eastern Alaska Range constrained by stacked thermochronology:
 573 Geological Society, London, Special Publications, v. 378, p. 225-243.

574 Bergman, S., 2006, North Alaska Upper Cretaceous Tephra: Eurasian or North American Source
 575 Calderas? Abstract, 102nd Annual Meeting of the Cordilleran Section, GSA, Anchorage,
 576 Alaska.

577 Bird, K.J., 2001, Alaska: a twenty-first century petroleum province, *in* M.W. Downey, J.C.
 578 Threet, and W.A. Morgan, eds., *Petroleum Provinces of the Twenty-First Century*:
 579 American Association of Petroleum Geologists Memoir 74, p. 137-165.

580 Box, S.F., and Patton, W.W., 1989, Igneous history of the Koyukuk Terrane, Western Alaska:
 581 Constraints on the origin, evolution, and ultimate collision of an accreted island arc
 582 terrane: *Journal of Geophysical Research*, v. 94, p. 15843-15867.

583 Christidis, G.E., 2001, Geochemical correlation of bentonites from Milos Island, Aegean,
 584 Greece: *Clay Minerals*, v. 35, p. 295-306.

585 Christiansen, P.P., and Snee, L.W., 1994, Structure, metamorphism, and geochronology of the
 586 Cosmos Hills and Ruby Ridge, Brooks Range schist belt, Alaska: *Tectonics*, v. 13, p.
 587 193-213.

588 Cohen, K.M., Finney, S., and Gibbard, P.L., 2013, *International Chronostratigraphic Chart*:
 589 International Commission on Stratigraphy.
 590 <http://www.stratigraphy.org/ICSchart/ChronostratChart2013-01.pdf>.
 591

592 Cole, F., Bird, K.J., Toro, J., Roure, F., O'Sullivan, P.B., Pawlewicz, M., and Howell, D.G.,
 593 1997, An integrated model for the tectonic development of the frontal Brooks Range and
 594 Colville Basin 250 km west of the Trans-Alaska Crustal Transect: *Journal of Geophysical*
 595 *Research*, v. 102, p. 20,685-20,708.

596 Collins, F.R., 1958, Test Wells, Umiat Area, Alaska: U.S. Geological Survey Professional Paper
597 305-B, p. 71-206.

598 Collins, F.R., 1959, Test Wells, Square Lake and Wolf Creek Areas, Alaska: U.S. Geological
599 Survey Professional Paper 305-H, p. 423-484.

600 Conrad, J.E., McKee, E.H., and Turrin, B.D., 1990, Age of tephra beds at the Ocean Point
601 dinosaur locality, North Slope, Alaska, based on K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ analyses: Evolution
602 of Sedimentary Basins-North Slope Basin, U.S. Geological Survey Bulletin 1990, 14 p.

603 Decker, P.L., 2007, Brookian sequence stratigraphic correlations, Umiat field to Milne Point
604 field, west-central North Slope, Alaska: Alaska Division of Geological and Geophysical
605 Surveys Preliminary Interpretive Report 2007-2, 21 p.

606 Detterman, R.L., Bickel, R.S., and Gryc, G., 1963, Geology of the Chandler River Region,
607 Alaska: U.S. Geological Survey Professional Paper 303-E, p. 223-324.

608 Fanti, F., 2009, Bentonite chemical features as proxy of late Cretaceous provenance changes: A
609 case study from the Western Interior Basin of Canada: *Sedimentary Geology*, v. 217, p.
610 112-127.

611 Flaig, P.P., 2010, Depositional environments of the Late Cretaceous (Maastrichtian) dinosaur-
612 bearing Prince Creek Formation: Colville River region, North Slope, Alaska. Ph.D.
613 Dissertation, University of Alaska, 311 p.

614 Flaig, P.P., Fiorillo, A.R., and McCarthy, P.J., 2014, Dinosaur-bearing hyperconcentrated flows
615 of Cretaceous Arctic Alaska: Recurring catastrophic event beds on a distal paleopolar
616 coastal plain: *Palaios*, v. 29, p. 594-611.

617
618 Flores, R.M., Myers, M.D., Houseknecht, D.W., Stricker, G.D., Brizzolara, D.W., Ryherd, T.J.,
619 and Takahashi, K.I., 2007, Stratigraphy and facies of Cretaceous Schrader Bluff and
620 Prince Creek formations in Colville River bluffs, North Slope, Alaska: U.S. Geological
621 Survey Professional Paper 1748, 52 p.

622 Gillespie, A.R., Huneke, J.C., and Wasserburg, G.J., 1982, An assessment of ^{40}Ar - ^{39}Ar dating of
623 incompletely degassed xenoliths: *Journal of Geophysical Research*, v.87, p. 9247-9257.

624 Hancock, J.M., and Kauffman, E.G., 1978, The great transgressions of the Late Cretaceous:
625 Journal of the Geological Society, v. 136, p. 175-186.

626 Harrison, T.M., 1990, Some observations on the interpretation of feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ results:
627 Chemical Geology (Isotope Geoscience Section), v. 80, p. 219-229.

628 Harrison, T.M., Duncan, I., and McDougall, I., 1985, Diffusion of ^{40}Ar in biotite: Temperature,
629 pressure, and compositional effects: Geochimica et Cosmochimica Acta, v. 49, p. 2461-
630 2468.

631 Hastie, A.R., Kerr, A.C., Pearce, J.A., and Mitchell, S.F., 2007, Classification of altered volcanic
632 island arc rocks using immobile trace elements: Development of the Th-Co
633 discrimination diagram: Journal of Petrology, v. 48, p. 2341-2357.

634 Houseknecht, D.W., and Schenk, C.J., 2005, Sedimentology and sequence stratigraphy of the
635 Cretaceous Nanushuk, Seabee, and Tuluva formations exposed on Umiat Mountain,
636 North-Central Alaska: U.S. Geological Survey Professional Paper 1709-B, 18 p.

637 Houseknecht, D.W., Bird, K.J., and O'Sullivan, P., 2010, Constraining the age and magnitude of
638 uplift in the northern National Petroleum Reserve in Alaska (NPPRA) – apatite fission-
639 track analysis of samples from three wells: U.S. Geological Survey Professional Paper
640 1784-A, 30 p.

641 Houseknecht, D.W., Burns, W.M., and Bird, K.J., 2012, Thermal maturation history of Arctic
642 Alaska and the southern Canada Basin, in N.B. Harris and K.E. Peters, eds., Analyzing
643 the Thermal History of Sedimentary Basins: SEPM Society for Sedimentary Geology
644 Special Publication 11, Tulsa, OK, p 199-219.

645 Hubbard, R.J., Edrich, S.P., and Rattey, P., 1987, Geologic evolution and hydrocarbon habitat of
646 the 'Arctic Alaska microplate': Marine and Petroleum Geology, v. 4, p. 2-34.

647 Huffman Jr., A.C., Ahlbrandt, T.S., Pasternack, I., Stricker, G.D., and Fox, J.E., 1985,
648 Depositional and sedimentologic factors affecting the reservoir potential of the
649 Cretaceous Nanushuk Group, central North Slope, Alaska, in A.C. Huffman, ed.,

650 Geology of the Nanushuk Group and Related Rocks, North Slope, Alaska: U.S.
 651 Geological Survey Bulletin 1614, p. 61-74.

652 Imlay, R.W., 1961, Characteristic Lower Cretaceous Megafossils from Northern Alaska: U.S.
 653 Geological Survey Professional Paper 335, 74 p., 20 plates.

654 Jones, D.L. and Gryc, G., 1960, Upper Cretaceous Pelycypods of the Genus *Inoceramus* from
 655 Northern Alaska: U.S. Geological Survey Professional Paper 334-E, p. 149-163.

656 Kiipli, T., Jeppsson, L., Kallaste, T., and Söderlund, U., 2008, Correlation of Silurian bentonites
 657 from Gotland and the eastern Baltic using sanidine phenocryst composition, and
 658 biostratigraphical consequences: Journal of the Geological Society, London, v. 15, p.,
 659 211-220.

660 Land, L.S., Mack, L.E., Milliken, K.L., and Lynch, F.L., 1997, Burial diagenesis of argillaceous
 661 sediment, south Texas Gulf of Mexico sedimentary basin: a reexamination: Geological
 662 Society of America Bulletin, v., 109, p. 1-15.

663 Lanphere, M.A., and Dalrymple, G.B., 2000, First-principles calibration of ^{38}Ar tracers:
 664 Implications for the ages of $^{40}\text{Ar}/^{39}\text{Ar}$ fluence monitors: U.S. Geological Survey
 665 Professional Paper 1621, 10 p.

666 Lanphere, M.A. and TAILLEUR, I.L., 1983, K-Ar ages of bentonites in the Seabee Formation,
 667 Northern Alaska: A Late Cretaceous (Turonian) time-scale point: Cretaceous Research, v.
 668 4, 361-370.

669 Layer, P.W., Hall, C.M., and York, D., 1987, The derivation of $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra of single
 670 grains of hornblende and biotite by laser step heating: Geophysical Research Letters, v.
 671 14, p. 757-760.

672 Lease, R.O., Houseknecht, D.W., and Kylander-Clark, A.R.C., 2014, Detrital zircon constraints
 673 on Arctic Alaska foreland basin evolution: Brookian chronostratigraphy, clastic
 674 progradation, provenance, and orogenic exhumation: Abstract, 2014 GSA Annual
 675 Meeting, Vancouver, British Columbia.

676 LePain, D.L., McCarthy, P.J., and Kirkham, R., 2009, Sedimentology, stacking patterns, and
 677 depositional systems in the middle Albian-Cenomanian Nanushuk Formation in outcrop,
 678 Central North Slope, Alaska: Alaska Division of Geological and Geophysical Surveys
 679 Report on Investigations 2009-1, 86 p.

680 McDougall I., and Harrison, T.M., 1999, Geochronology and the Thermochronology by the
 681 $^{40}\text{Ar}/^{39}\text{Ar}$ Method: Oxford University Press, New York. 269 p.

682 Miller, T.P., 1989, Contrasting plutonic rock suites of the Yukon-Koyukuk basin and the Ruby
 683 Geanticline, Alaska: Journal of Geophysical Research, v. 94, p. 15969-15987.

684 Molenaar, C. M., 1985, Subsurface correlations and depositional history of the Nanushuk Group
 685 and related strata, North Slope, Alaska, *in* A. C. Huffman, ed., Geology of the Nanushuk
 686 Group and Related Rocks, North Slope, Alaska: U.S. Geological Survey Bulletin 1614, p.
 687 37-60.

688 Moore, G.W., Drummond, K.J., Teraoka, Y., McCoy, F.W., Swint-Iki, T.R., and Gartner, A.L.,
 689 2000, Geologic Map of the Circum-Pacific Region, Arctic Sheet: U.S. Geological Society
 690 Map CP-48.

691 Moore, T.E., Wallace, W.K., Bird, K.J., Karl, S.M., Mull, C.G., and Dillon, J.T., 1994, Geology
 692 of Northern Alaska, *in* The Geology of North America, v. G-1, The Geology of Alaska:
 693 Geological Society of America, p. 49-140.

694 Moore, T.E., Potter, C.J., O'Sullivan, P.B., Shelton, K.L., and Underwood, M.B., 2004, Two
 695 stages of deformation and fluid migration in the west-central Brooks Range fold and
 696 thrust belt, northern Alaska, *in* R.Swenson, F. Roure, and J.W. Granath, eds.,
 697 Deformation, fluid flow, and reservoir appraisal in foreland fold and thrust belts: AAPG
 698 Hedberg Series, no. 1, p. 157-186.

699 Mull, C.G., Houseknecht, D.W., and Bird, K.J., 2003, Revised Cretaceous and Tertiary
 700 stratigraphic nomenclature in the Colville Basin, northern Alaska: U.S. Geological
 701 Survey Professional Paper 173, 51 p.

702 Mull, C.G., Houseknecht, D.W., Pessel., G.H., and Garrity, C.P., 2004, Geologic Map of the
703 Umiat Quadrangle, Alaska: U.S. Geological Survey Scientific Investigations Map 2817-
704 A, scale 1:250,000.

705 O'Sullivan, P.B., 1999, Thermochronology, denudation, and variations in paleosurface
706 temperature: a case study from the North Slope foreland basin, Alaska: Basin Research,
707 v. 11, p. 191-204.

708 Pearce, J.A., 1983, Role of the sub-continental lithosphere in magma genesis at active
709 continental margins, *in* C.J. Hawkesworth and M.J. Norry, eds., Continental Basalt
710 Mantle Xenoliths: Nantwich, Shiva Publications, p. 230-249.

711 Pearce, J.A., 1996, A user's guide to basalt discrimination diagrams, *in* D.A., Wyman, (ed.),
712 Trace Element Geochemistry of Volcanic Rocks: Applications for Massive Sulfide
713 Exploration: Geological Association of Canada, Short Course Notes, v. 12, p. 79-113.

714 Pearce, J.A., and Peate, D.W., 1995, Tectonic implications of the composition of volcanic arc
715 magmas: Annual Review of Earth and Planetary Sciences, v. 23, p. 251-286.

716 Pearce, J.A., Harris, N.B.W., and Tindle, A.G., 1984, Trace element discrimination diagrams for
717 the tectonic interpretation of granitic rocks: Journal of Petrology, v. 25, 956-983.

718 Prokoph, A., Villeneuve, M., Agterberg, F.P., and Rachold, V., 2001, Geochronology and
719 calibration of global Milankovitch cyclicity at the Cenomanian-Turonian boundary:
720 Geology, v. 29, p. 523-526.

721 Robinson, F.M., 1958, Test Wells, Gubik Area, Alaska: U.S. Geological Survey Professional
722 Paper 305-C, p. 207-264.

723 Sampson, S.D., and Alexander, E.C., 1987, Calibration of the interlaboratory $^{40}\text{Ar}/^{39}\text{Ar}$ dating
724 standard, MMhb1: Chemical Geology, v. 66, p. 27-34.

725 Schindler, J.F., 1988, History of exploration in the National Petroleum Reserve in Alaska, with
726 emphasis on the period from 1975-1982, *in* G. Gryc, ed., Geology and Exploration of the
727 National Petroleum Reserve in Alaska, 1974-1982: U.S. Geological Survey Professional
728 Paper 1399, p. 13-76.

729 Shimer, G.T., McCarthy, P.J, and Hanks, C.L., 2014, Sedimentology, stratigraphy, and reservoir
730 properties of an unconventional reservoir in the Cretaceous Nanushuk Formation at
731 Umiat Field, North Slope, Alaska: AAPG Bulletin, v. 98, p. 631-661.

732 Smosna, R., 1988, Low-temperature, low-pressure diagenesis of Cretaceous sandstones, Alaskan
733 North Slope: Journal of Sedimentary Petrology, v. 58, p. 644-655.

734 Smosna, R., 1989, Compaction law for Cretaceous Sandstones of Alaska's North Slope: Journal
735 of Sedimentary Petrology, v. 59, p. 572-584.

736 Spicer, R.A., and Herman, A.B., 2010, The Late Cretaceous environment of the Arctic: A
737 quantitative reassessment based on plant fossils: Palaeogeography, Palaeoclimatology,
738 Palaeoecology, v. 295, p. 423-442.

739 Steiger, R.H., and Jaeger, E., 1977, Subcommittee on geochronology: Convention on the use of
740 decay constants in geo and cosmochemistry: Earth and Planetary Science Letters, v. 36,
741 p. 359-362.

742 Tappan, H.N., 1960, Cretaceous biostratigraphy of northern Alaska: AAPG Bulletin, v. 44, p.
743 273-297.

744 Tikhomirov, P.L., Akinin, V.V., Ispolatov, V.O., Alexander, P., Cherpanova, I.Y.
745 Cherpanova, and Zagoskin, V.V., 2006, The Okhotsk-Chukotka Volcanic Belt: Age of
746 its northern part according to new Ar-Ar and U-Pb geochronological data: Stratigraphy
747 and geological Correlation, 2006, v. 15, p. 524-537.

748 Tikhomirov, P.L., Kalinina, E.A., Moriguti, T., Makishima, A., Kobayashi, K., Cherpanova,
749 I.Yu., and Nakamura, E., 2012, The Cretaceous Okhotsk-Chukotka Volcanic Belt (NE
750 Russia): Geology, geochronology, magma output ranges, and implications on the genesis
751 of silicic LIPs: Journal of Volcanology and Geothermal Research, v., 221-222, p. 14-32.

752 Till, A.B., 1992, Detrital blueschist-facies metamorphic mineral assemblages in Early
753 Cretaceous sediments of the foreland basin of the Brooks Range, Alaska, and
754 implications for orogenic evolution: Tectonics, v. 11, p. 1207-1223.

755 Till, A.B. and Patrick, B.E., 1991, Ar-Ar evidence for a 110-105 a amphibolite facies overprint
756 on blueschist in the south-central Brooks Range, Alaska: Geological Society of America
757 Abstracts Programs, 23, A436.

758 Till, A.B., and Snee, L.W., 1995, $^{40}\text{Ar}/^{39}\text{Ar}$ evidence that formation of blueschists in continental
759 crust was synchronous with foreland fold and thrust belt deformation, western Brooks
760 Range, Alaska: *Journal of Metamorphic Geology*, v. 13, p. 41-60.

761 Toro, J., Cole, F., and Meier, J., 1998, $^{40}\text{Ar}/^{39}\text{Ar}$ ages of detrital minerals in Lower Cretaceous
762 rocks of the Okpikruak Foramtion: evidence for Upper Paleozoic metamorphic rocks in
763 the Koyukuk arc, in Gray, J.E., and Riehle, J.R., *Geologic Studies in Alaska by the U.S.*
764 *Geological Survey*, 1996: U.S. Geological Survey Professional Paper 1595, p. 169-182.

765
766 Vogl, J.J., Calvert, A.T., and Gans, P.P., 2002, Mechanisms and timing of exhumation of
767 collision-related metamorphic rocks, southern Brooks Range, Alaska: insights from
768 $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology: *Tectonics*, v. 21, doi 10.1029/2000TC001270.

769 Winchester, J.A., and Floyd, P.A., 1977, Geochemical discrimination of different magma series
770 and their differentiation products using immobile elements: *Chemical Geology*, v. 20, p.
771 325-343.

772 York, D., Hall, C.M., Yanase, Y., Hanes, J.A., and Kenyon, W.J., 1981, $^{40}\text{Ar}/^{39}\text{Ar}$ dating of
773 terrestrial minerals with a continuous laser: *Geophysical Research Letters*, v. 8, p. 1136-
774 1138.

775

Figure 1 Geologic Map of the Study Area

The study sites are located in the central Brooks Range foothills along the Colville River, near the border of the National Petroleum Reserve – Alaska. The Brooks Range foothills contain fold and thrust structures, including anticline-syncline pairs, that expose the Brookian foreland basin deposits at the surface (Geologic map modified from Mull et al., 2004). The inset map in the lower right corner shows the location of the study area along the Colville River, as well as locations mentioned in the text: Umiat (UM) and Shale Wall Bluff (SW). Inset map modified from Decker (2007).

Figure 2 Brookian Stratigraphy of the North Slope

A) Bentonites in this study are taken from the Tuluvak, Seabee, and Nanushuk Formations, which are highlighted in with a bold black line in the stratigraphic section chart (modified from Mull et al., 2003). These formations range from Albian-Coniacian in age. A single sample from the base of the Schrader Bluff Formation is also included. The Schrader Bluff Formation is Santonian to Maastrichtian in age. B) The sample legend depicts the approximate stratigraphic position of each of the samples, as well as the well location. Bold, underlined samples are discussed in section 5.1 and 5.2.

Figure 3 $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectra, Nanushuk and Seabee Formations

A-B) The age spectra from UM2-311N have flat release spectra that meet the criteria for a plateau age. Grey boxes indicate steps selected for plateau ages. A step-up release pattern indicates argon-loss. C) Sample CI-006 technically does not meet the criteria for a plateau age, despite having a flat release spectrum.

Figure 4 $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectra, Tuluvak Formation

The age spectra from all three samples presented here have flat release spectra that meet the criteria for plateau ages. Grey boxes indicate steps selected for plateau ages.

Figure 5 Tuluvak Formation Stratigraphy

Stratigraphic sections from four of the wells in this study illustrate the thickness and lithology of the Tuluvak Formation, as well as the depths of tephra samples. Wells are “hung” on the base of the lowest sandstones in the Tuluvak Formation. Depths from the surface are listed in both feet (grey) and meters (black), with samples labeled based on depth in feet. The plateau age of each sample is listed next to the sample number.

Figure 6 Chronostratigraphically Anomalous Albian $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectra

The age spectra in this figure have anomalous Albian $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages that are not in line with their stratigraphic positions. The Seabee and Tuluvak formations are Cenomanian and younger. UM11-2258N is considered too old based on its relative stratigraphic position in comparison to UM2-311N.

Figure 7 Chronostratigraphically Anomalous Valanginian-Barremian $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectra

The age spectra in this figure have anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages that **are not in line with** their stratigraphic positions. The three samples here are from the Seabee and Tuluvak

formations, which are Cenomanian and younger. The Valanginian-Barremian dates are unrealistic, and indicate excess argon.

Figure 8 Nanushuk and Seabee Formation Stratigraphy

Stratigraphic sections from three of the wells in this study illustrate the thickness and lithology of the Nanushuk and Seabee formations, as well as the relative depths of bentonite samples. Depths from the surface are listed in both feet (grey) and meters (black), with samples labeled based on depth in feet. The plateau age of each sample is listed next to the sample number.

Figure 9 Cretaceous Igneous and Sedimentary Units in Siberia, Alaska, and Canada

The Okhotsk-Chukotka Volcanic Belt (OCVB) is a potential source for Brookian tephra deposits. Both the OCVB and metamorphic rocks in the southern Brooks Range could be sources of detrital sediment or xenoliths responsible for anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ ages. The regions depicted include the western Okhotsk (W/O) and Anadyr (AN) segments of the OCVB, and the Yukon-Koyukuk terrane (Y-K). Modified from Moore et al. (2000) with additional data from Moore et al. (1994).

Figure 10 $^{40}\text{Ar}/^{39}\text{Ar}$ Plateau Ages Versus Okhotsk-Chukotka Volcanic Belt Ages

When plotted against a histogram of radiometric ages for the Okhotsk-Chukotka Volcanic Belt (OCVB), the results of this study (dots) fall near peaks in activity. Modified from Akinin and Miller (2011).

Figure 11 $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectra of Unknown Phases

The two samples depicted here are from unknown mineral phases, as evidenced by high Ca/K ratios. Sample GU1-736T is a possible calcic plagioclase, but without further mineralogical data the age is geologically uninterpretable. GU1-2845S is from likely calcite separates.

Figure 12 Nanushuk Formation Geochemistry

Bivariate plots comparing tephra geochemistry from Nanushuk Formation samples, with labeled circles depicting elemental results from different sample analyses. Grey circles represent elements used in Figure 14. A) The correlation between the $< 150\ \mu\text{m}$ and whole rock trace element geochemistry from UM2-311 is relatively strong, with a slope approaching 1.0 and a relatively high R^2 value. Based on this diagram we infer minor geochemical fractionation based on particle size. B) The wide scatter of data points for UM2-311 and UM11-2258 trace element geochemistry show weak correlation between these samples from the Nanushuk Formation.

Figure 13 Igneous Classification Diagrams

A) Total-Alkali-Silica (TAS) diagram of the Umiat bentonites. The Umiat bentonites become increasingly felsic with time, with the exception of UM11-2258, which may be contaminated with detrital sediment. B) A similar trend of increasingly felsic composition appears in the trace element classification diagram, where Zr/TiO_2 serves as a proxy for SiO_2 content. Some SiO_2 loss is expected, and the trace element diagram does classify the tephras as more felsic than the TAS diagram.

865

866 **Figure 14 Volcanic Arc Classification Diagrams**

867 A) The likely source of the Umiat bentonites was a volcanic arc. B) More specifically, the
868 Umiat tephtras are geochemically similar to igneous rocks in continental arc settings. The
869 classification diagrams are based on Pearce et al. (1984).

870

871

Figure 1

Figure 2

B

W Sample Legend E

Well Formation	SQ1	UM11	UM2	GU1	GU2
Schrader Bluff (SC)				736	
Tuluvak (T)	<u>257</u>	<u>345</u> 391		1165	<u>1397</u>
Seabee (S)	<u>CI-006</u>	1682 1982		2840 2845	
Nanushuk (N)		2258	<u>311</u>		
<div> <div>terrestrial/ shallow marine</div> <div>shallow marine/marine shelf/ marine slope</div> <div>deep marine basin</div> </div>					

Figure 3

A Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13 **Nanushuk Fm. Bentonites**

- UM11-2258
- UM2-311r
- ◻ UM2-311w "whole rock"

Seabee Fm. Bentonites

- ▲ UM11-1682
- △ UM11-1962

Tuluvak Fm. Bentonites

- UM11-345
- UM11-391

Figure 14 **Nanushuk Fm. Bentonites**

■ UM11-2258

□ UM2-311r

■ UM2-311w "whole rock"

Seabee Fm. Bentonites

▲ UM11-1682

△ UM11-1962

Tuluva Fm. Bentonites

○ UM11-345

● UM11-391

1 **Table 1 ⁴⁰Ar/³⁹Ar Results and Plateau Information**

Sample	Formation	Phase	Integrated Age	Plateau Age	Isochron Age	Fractions	% ³⁹ Ar release	MSWD
GU1-736SC	Schrader Bluff	plagioclase?	93.6±6.5	89.3 ± 4.4	87.6 ± 3.6	8 of 11	87.2	0.27
GU1-1165T	Tuluvak	biotite	158.1±5.6	133.1 ± 2.2		7 of 16	57.4	1.8
GU1-2840S	Seabee	biotite	104.8±0.9	106.0 ± 0.9		13 of 16	94.0	0.57
GU1-2845S	Seabee	calcite	398.7±43.8	None		--	--	--
GU2-1397T	Tuluvak	biotite	96.2±2.5	96.2 ± 2.0		4 of 13	96.5	0.68
SQ1-257T	Tuluvak	glass	96.7±1.5	96.6 ± 1.4	95.3 ± 2.8	10 of 12	99.6	0.25
CI-006S	Seabee	biotite	96.7±0.7	98.2 ± 0.8		4 of 11	45.3	-
UM2-311N #1*	Nanushuk	biotite	97.4 ± 1.5	102.6 ± 1.5		4 of 10	60.2	0.33
UM2-311N #2*	Nanushuk	biotite	98.3±1.6	102.2 ± 1.1		6 of 10	72.6	0.64
UM11-345T	Tuluvak	sanidine?	94.2±0.9	94.4 ± 0.9	94.2 ± 1.1	8 of 10	97.2	0.27
UM11-391T	Tuluvak	biotite	123.5±2.0	129.1 ± 1.8	125.6 ± 2.0	11 of 16	59.3	1.23
UM11-1682S	Seabee	biotite	100.3±1.1	104.1 ± 1.1		11 of 16	82.0	0.51
UM11-1962S	Seabee	biotite	137.3±3.5	131.2 ± 1.3	131.3 ± 2.4	10 of 16	55.5	0.36
UM11-2258N	Nanushuk	biotite	106.5±1.0	110.4 ± 0.9		11 of 16	79.0	0.67

2
3
4

1 **Table 2 XRF Major and Trace Element Analysis Results**
2

Oxides (wt. %)	UM11-345	UM11-391	UM11-1682	UM11-1962	UM11-2258	UM2-311	UM2-311w
SiO ₂	68.80	72.02	63.03	61.27	64.49	58.35	57.67
Al ₂ O ₃	19.67	18.71	20.24	21.40	21.73	22.62	23.37
Fe ₂ O ₃	4.96	4.02	4.70	5.48	3.63	4.07	4.23
MgO	2.49	2.54	2.41	2.93	2.37	2.85	2.81
CaO	0.72	0.51	4.18	1.36	0.43	1.88	3.58
MnO	---	---	0.09	0.02	0.01	0.02	0.03
Na ₂ O	3.30	3.07	3.27	3.15	2.94	2.48	2.23
K ₂ O	0.21	0.18	0.92	1.03	1.36	1.03	1.09
TiO ₂	0.23	0.16	0.53	0.51	0.37	0.79	0.77
P ₂ O ₅	0.01	0.01	0.08	0.23	0.07	0.17	0.18
Cl	0.01	0.01	0.02	0.03	0.02	0.04	0.03
S	0.04	---	0.26	0.13	0.01	---	0.01
Trace Elements (ppm)	UM11-345	UM11-391	UM11-1682	UM11-1962	UM11-2258	UM2-311	UM2-311w
Ba	737.74	382.91	994.69	790.43	1317.81	1137.21	1092.27
Ce	109.16	116.55	119.18	74.57	92.15	114.96	91.59
Co	3.65	2.05	4.10	5.97	3.34	10.16	9.73
Cr	2.83	3.51	8.88	21.21	2.95	108.79	106.94
Cs	1.29	2.08	11.48	5.43	4.67	1.29	2.77
Cu	4.78	4.39	10.20	8.08	14.56	14.81	15.28
Ga	27.93	27.93	22.85	24.85	28.08	29.57	28.50
Hf	7.28	7.00	5.87	4.44	10.23	6.45	5.85
La	48.78	57.82	68.52	43.85	45.56	57.56	43.54
Mn	21.06	18.07	660.78	159.27	60.14	147.21	218.77
Mo	1.23	0.48	4.13	2.19	2.40	1.06	1.59
Nb	12.47	7.85	8.93	6.83	11.77	5.41	5.88
Nd	49.98	51.20	58.09	29.93	36.61	45.32	32.04
Ni	12.25	2.41	4.64	7.54	10.23	32.13	30.19
Pb	30.89	36.57	24.63	28.77	51.78	46.37	41.24
Rb	9.56	9.54	46.48	37.31	47.52	39.91	39.33
Sc	14.49	10.75	14.86	8.10	12.50	12.40	10.65
Sm	9.91	5.96	7.74	2.40	3.73	1.44	3.36
Sn	7.19	10.10	7.04	8.25	9.98	9.53	9.19
Sr	331.84	218.60	361.31	356.59	255.87	22.09	22.15
Ta	---	1.59	1.62	1.79	1.73	1.44	2.50
Th	25.50	25.51	29.91	20.13	31.61	35.99	35.39
TiO₂	2343.50	1567.15	5291.84	5097.02	3725.19	7855.94	7788.39
U	12.02	9.91	11.78	8.53	12.39	13.90	10.51
V	10.67	4.31	28.80	49.63	26.22	76.80	73.13
Y	38.56	44.18	37.33	15.87	36.08	18.51	16.30
Yb	2.83	5.31	3.68	3.30	4.65	2.80	3.86
Zn	77.54	42.74	105.41	88.00	91.96	80.96	70.30

1 **Table 3 XRF Trace Element Standards**
2

Sample	TiO ₂	Nb	Rb	Ta	Th	Y	Yb	Zr
JA-2 measured:	7130	9	72	<DL	5	18	1.0	118
• compiled:	6700	9.8	68	0.61	4.7	18	1.6	119
• % difference:		-8.0%	5.9%	---	6.4%	0.0%	-37.5%	-0.8%
BE-N measured:	30143	106	55	8.0	11	30	1.0	303
• compiled:	26100	100	47	5.5	11	30	1.8	265
• % difference:		-6.0%	17.0%	45.5%	0.0%	0.0%	-44.4%	14.3%
GSP-1 measured:	5079	20	206	4.0	82	21	3	420
• compiled:	6500	27.9	254	---	106	26	---	530
• % difference:		-28.3%	-18.9%	---	-22.6%	-19.2%	---	-20.8%
JG-2 measured:	449	14	302	<DL	32	85	10	93
• compiled:	440	15	297	1.9	29.7	89	8.3	97
• % difference:		-6.7%	1.7%	---	7.7%	-4.5%	20.5%	-4.1%
JR-1 measured:	1019	16	262	3.0	29	45	6	100
• compiled:	1100	15.5	257	1.9	26.5	46	4.6	102
• % difference:		3.2%	1.9%	57.9%	9.4%	-2.2%	30.4%	-2.0%
JP-1 measured:	32	1	1	6	2	<DL	8	5
• compiled:	17	1.2	0.8	0.02	0.2	1	0.02	6
• % difference:	88.0%	69.4%	25.0%	299%	90.0%	---	399.0%	-16.7%

3
4