1. Introduction

The timing of and tectonic setting during production of continental crust remain important research issues globally despite decades of post-plate tectonics revolution investigation of these subjects (Armstrong 1968; Moorbath 1975; Reymer and Schubert 1986; Patchett and Chase 2002; Stern and Scholl 2010; Condie et al. 2011; Cawood et al. 2013). Critical questions include the following. What is the relative importance of generation of continental crust in different tectonic settings such as magmatic arcs or extensional domains (e.g. Stern and Scholl 2010)? Both globally and in any one location, is continental crust created continuously or episodically (e.g. Patchett and Chase 2002; Condie et al. 2011)?

The Goochland Terrane is a fault-bounded crustal block in the Appalachian Piedmont Province of Virginia, eastern USA (Figures 1 and 2). Unlike younger neighbouring rocks, geologists commonly interpret the Goochland Terrane as a sliver of Mesoproterozoic continental crust because the oldest exposed rocks in the terrane intruded near the end of this era (Aleinikoff et al., 1996; Owens and Tucker 2003) and because whole-rock Nd isotope model ages for extraction of these intrusive rocks from the depleted mantle are approximately 1400 Ma (Owens and Samson 2004). Nearly all of the Mesoproterozoic and Neoproterozoic rocks in the Goochland Terrane are meta-igneous, affording us an opportunity to investigate the timing and mechanisms of pre-Cambrian growth of this fragment of continental crust.

In this paper, we use whole-rock major and trace element abundances combined with oxygen, samarium-neodymium, lutetium-hafnium, and uranium-lead isotope analyses to elucidate magmatic sources of Goochland Terrane continental crust. We then discuss implications for the magmatic and tectonic development of the Goochland Terrane, especially the mechanisms and timing of addition of continental crust.
2. Geologic setting

Geologists use distinctions originally based on physiography to divide the central and southern Appalachian Orogen into tectonic provinces (Hatcher 1989). From west to east these include the Blue Ridge, Piedmont, and Coastal Plain provinces (Figure 1(b)). The Blue Ridge Province consists of Middle and Late Mesoproterozoic intrusive rocks that formed and were metamorphosed due to amalgamation of Rodinia as well as Neoproterozoic felsic intrusions and sedimentary and volcanic strata formed during breakup of Rodinia (Aleinikoff et al. 1995; Tollo et al. 2004, 2017). The Goochland Terrane is part of the Piedmont Province. Dominated by metasedimentary rocks, the Piedmont Province is the formerly deepest and most outboard part of the Paleozoic Appalachian Orogen currently exposed (Hibbard et al. 2006). The Neoproterozoic and Paleozoic Piedmont Province rocks constitute a collage of fault-bounded terranes of varying origin (Horton et al. 1989). The rocks in some terranes, such as the Westminster Terrane, intruded or were deposited on the eastern edge of Laurentia (Martin et al. 2015). Others, such as the rocks that compose Carolina, intruded and were deposited on the periphery of Gondwana (Pollock et al. 2012). The Goochland Terrane is now located at the north-eastern end of Carolina (Figure 1(a)), and several authors tentatively concluded that like Carolina, the Goochland Terrane may have been exotic to Laurentia prior to possible accretion during the Paleozoic Era (Horton et al. 1989; Rankin 1994; see also Hibbard and Samson 1995; Keppie et al., 1996). The arguments in favour of an exotic origin were differences in structure and stratigraphy between the Goochland Terrane and known Laurentian outliers in the Piedmont Province. In contrast, Farrar (1984), Aleinikoff et al. (1996), and Glover et al. (1997) linked the Goochland Terrane to Laurentia during Proterozoic and Paleozoic time. Overlapping compositions and intrusion ages of latest Mesoproterozoic anorthosite and felsic rocks in the Goochland Terrane and the Blue Ridge Province, as well as the comparable latest Mesoproterozoic metamorphism of the rocks in these two regions, formed the bases for this linkage. Bartholomew and Tollo (2004) likewise concluded that the Goochland Terrane was part of Laurentia during Proterozoic time, placing the Goochland Terrane farther north than its current position during pre-Cambrian time because of similar Neoproterozoic igneous rocks in the Goochland Terrane on the one hand and Pennsylvania, New Jersey, and New York on the other.

The oldest rock units in the Goochland Terrane are a granitic gneiss that intruded at ca. 1050–1010 Ma, called the State Farm Gneiss, and the 1045 ± 10 Ma Montpelier Anorthosite (Figure 2; Table 1; Aleinikoff et al., 1996; Owens and Tucker 2003). The composition of the Montpelier Anorthosite is unusual compared to other massif anorthosite on Earth in three ways: (1) Montpelier is one of the two most potassic anorthosite bodies found on Earth, (2) it is the only known anorthosite in which the barium concentration exceeds the strontium concentration, and (3) quartz is common (Owens and Dymek 2016). A suite of granitic magmas intruded the State Farm Gneiss at ca. 660 to 580 Ma (Owens and Tucker 2003). This suite has not been formally named, so we follow Owens and Tucker (2003) in referring to these rocks as Neoproterozoic granitoid.

The Sabot Amphibolite mostly crops out near the State Farm Gneiss (Figure 2). Dominated by meta-mafic rocks
with minor-intercalated meta-felsic layers, the Sabot Amphibolite nowhere intrudes the State Farm Gneiss; in individual outcrops, the Sabot Amphibolite rests on the Mesoproterozoic gneiss with apparent concordance between the compositional layering in the State Farm Gneiss and the base of the Sabot Amphibolite. Glover (1989) reported a low-angle discordance between the base of the Sabot Amphibolite and compositional layering in the underlying State Farm Gneiss at the regional scale. Weakly deformed pegmatite dikes intruded both the State Farm Gneiss and the Sabot Amphibolite, but the Sabot Amphibolite does not host other intrusions. These outcrop- and map-scale observations lead us to the interpretation that the Sabot Amphibolite is a meta-volcanic sequence deposited atop the State Farm Gneiss (see also Goodwin 1970). The eruption age is difficult to determine both
because meta-basaltic rocks typically yield little zircon related to eruption and also due to the intense deformation and metamorphism of the Sabot Amphibolite. The eruption age was unknown prior to the current study but here we suggest it may be ca. 550 Ma based on U/Pb isotopic dating of zircon from an intercalated felsic layer.

The most widely exposed unit in the Goochland Terrane is the Maidens Gneiss, which consists of intercalated sedimentary and igneous rocks, metamorphosed to granulite facies followed by amphibolite facies (Farrar 1984). The igneous crystallization age of at least some of the Maidens Gneiss igneous rocks is ca. 400 Ma (Owens et al. 2010). There are many small amphibolite bodies within the area nominally mapped as Maidens Gneiss in Figure 2. We take these amphibolite bodies to be structurally dismembered parts of the Sabot Amphibolite based on their locations close to the Sabot Amphibolite and the Mesoproterozoic rocks as well as their compositional similarity to the Sabot Amphibolite.

3. Methods

3.1 Sabot Amphibolite whole-rock major and trace element abundances

Owens collected all samples except sample 1994, which was collected by Samson. Samples 1994 and SA-001 were collected from mafic parts of the Sabot Amphibolite, which also contains minor-intercalated felsic layers at this outcrop. Nd and Sm isotope analyses were performed at Syracuse University following the procedures in Samson et al. (1995) with the modification that the $^{144}$Sm-$^{150}$Nd tracer solution was added to the powdered sample prior to dissolution. The measured $^{143}$Nd/$^{144}$Nd ratio was corrected for mass bias using $^{146}$Nd/$^{144}$Nd = 0.7219. The uncertainty on the initial εNd value of a rock measured in the Syracuse laboratory is better than 0.4 epsilon units (Samson et al. 1995). We used the isotopic compositions of the chondritic uniform reservoir from Bouvier et al. (2008) to calculate εNd values. Depleted mantle values were taken from DePaolo (1981).

3.2 Sabot Amphibolite whole-rock neodymium isotopes

Owens collected all samples except sample 1994, which was collected by Samson. Samples 1994 and SA-001 were collected from mafic parts of the Sabot Amphibolite, which also contains minor-intercalated felsic layers at this outcrop. Nd and Sm isotope analyses were performed at Syracuse University following the procedures in Samson et al. (1995) with the modification that the $^{144}$Sm-$^{150}$Nd tracer solution was added to the powdered sample prior to dissolution. The measured $^{143}$Nd/$^{144}$Nd ratio was corrected for mass bias using $^{146}$Nd/$^{144}$Nd = 0.7219. The uncertainty on the initial εNd value of a rock measured in the Syracuse laboratory is better than 0.4 epsilon units (Samson et al. 1995). We used the isotopic compositions of the chondritic uniform reservoir from Bouvier et al. (2008) to calculate εNd values. Depleted mantle values were taken from DePaolo (1981).

3.3 Zircon sample collection and processing

Martin collected all samples except Mont1, which was collected by Owens. We separated and mounted zircon in the laboratories in the Department of Geology at the University of Maryland. Zircon was isolated from each sample by first crushing the rock using a mortar and pestle, then removing the silt- and clay-size grains by hand-panning in water, removing magnetic grains with a Frantz magnetic barrier separator, and removing less-dense grains in methylene iodide. We picked zircon crystals by hand, then cast the grains in an epoxy disk together with shards or loose grains of standard zircon crystals. After the epoxy hardened, we ground and polished the disks by hand to expose the interiors of the grains. Prior to isotope analysis, the grains were...
imaged using backscattered electrons and cathodoluminescence in the JEOL JXA-8900R electron probe microanalyzer at the University of Maryland. The images were used to avoid multiple cathodoluminescence zones, cracks, and inclusions during selection of spots for isotope analysis. We report all uncertainties at the 2-sigma level.

### 3.4 Oxygen isotopes in spots in zircon

After polishing and imaging the zircon grains, we analysed oxygen isotopes in spots in the zircon using the CAMECA IMS 1280 secondary ion mass spectrometer (SIMS) housed in the WiscSIMS laboratory at the University of Wisconsin Madison. The analyses were conducted over two consecutive days. We followed the mount preparation and analytical procedures described in Kita et al. (2009), Valley and Kita (2009), and Wang et al. (2014). The primary beam consisted of Cs ions focused to a beam diameter of 15 μm. Secondary ions for 16O, 18O, and 16O1H were collected simultaneously by three Faraday cup detectors. Ratios of 16O/16O (hereafter called OH/O) in the sample zircon were corrected for background by subtracting the measured ratio in bracketing standard zircon, which was assumed to be anhydrous (Wang et al. 2014). Zircon standard KIM-5, with a δ18O value of 5.09 ± 0.12‰ VSMOW (Valley 2003), was included in the epoxy mounts with the Goochland Terrane zircon. We corrected for instrumental mass fractionation by bracketing four analyses of spots in KIM-5 before and after every 15 analyses of spots in Goochland Terrane zircon. Reported precisions are the two standard deviation values around the means of the measured δ18O values of the eight bracketing analyses of KIM-5 for each block of 15 Goochland Terrane analyses. We report and discuss all results relative to VSMOW.

### 3.5 Uranium/lead isotope dating of spots in zircon from Sabot Amphibolite sample sabot1

Zircon from a felsic layer in the Sabot Amphibolite, extracted from sample Sabot1, was analysed at the Arizona LaserChron Center at the University of Arizona to determine the eruption age of this meta-volcanic unit (Figure 2; Table 1). Gehrels and Pecha (2014) discussed the analytical methods in detail; here we provide a brief summary. Material was ablated from the polished surface of zircon crystals using a Photon Machines Analyte G2 Excimer laser with a beam diameter of 10 μm. Despite declining analytical precision with decreasing laser beam diameter, we used a 10 μm-diameter spot to permit targeting distinct cores or rims with minimal overlap of different cathodoluminescence domains and to avoid inclusions and cracks. A flow of helium carried the ablated zircon into the plasma source of a Nu high resolution inductively coupled plasma mass spectrometer, which was fitted with a flight tube wide enough to measure uranium, thorium, and lead isotopes simultaneously. All measurements were made in static mode using discrete dynode ion counters. Each analysis comprised three parts: background measurement via a 15-second integration on peaks with the laser off, fifteen 1-second integrations with the laser firing, and a 30-second delay to purge the previous sample. The depths of the resulting ablation pits were approximately 15 μm.

Data reduction was performed offline using an Excel program written at the Arizona LaserChron Center. A common lead correction was applied to the 204Pb signal assuming an initial lead composition from Stacey and Kramers (1975). We corrected for mass fractionation of isotopes during analysis using bracketing analyses of standard zircon ‘Sri Lanka,’ which has an accepted age of 563.5 ± 3.2 Ma (Gehrels et al. 2008). We calculated concentrations of uranium and thorium in the sample zircon based on analyses of the Sri Lanka standard zircon, which contains approximately 518 ppm uranium and 68 ppm thorium.

### 3.6 Uranium/lead and hafnium isotopes in spots in zircon from other samples

U/Pb and Hf isotopes in zircon from all samples except Sabot 1 were analysed in the GeoAnalytical Lab at Washington State University using the laser ablation split stream technique (Table 1). Analyses took place during two separate sessions on two consecutive days. We followed the measurement and data reduction procedures detailed in Fisher et al. (2014), which we briefly summarize here. The polished surface of each zircon grain was ablated using a NewWave 213 nm Nd:YAG laser. The laser diameter was 40 μm for samples 511001, 511005, and Mont1 and 30 μm for samples 511004 and 512001. We also used a 30 μm beam for 10 spots in zircon from sample 511001 and six spots in zircon from sample 511005 (marked in Table S2). The laser spots were centred on the pits created by the oxygen isotope analyses. The ablated zircon was carried out of the sample cell in a stream of He, which was then split into two separate paths. One stream entered a ThermoScientific Element2 single collector high resolution inductively coupled plasma mass spectrometer for measurement of uranium and lead isotopes. The second stream passed into a ThermoScientific Neptune multi-collector inductively coupled plasma mass spectrometer for measurement of lutetium, ytterbium, and hafnium isotopes. Analyses of shards or loose grains of at least two different zircon standards bracketed every 12 analyses of unknown zircon.
Each analysis consisted of background measurement on peaks for 30 s with the laser off followed by 65 s of ablation. The first 6 s of ablation were marked by increasing signal intensities, so the first 6 s of data were not included in the U/Pb date calculation. The next 30 s of data were used to calculate the U/Pb dates. U/Pb data were reduced offline using an in-house Excel program. Repeated measurements of reference zircon R33 and Temora yielded concordia ages of 420 ± 5 and 422 ± 6 Ma, respectively (Table S2, Figure S1), within uncertainty of the accepted crystallization ages (Black et al. 2004). HF data were reduced offline using Iolite software following the methods described by Fisher et al. (2014). We removed from further consideration any Hf analysis with a duration of less than 20 s. Repeated measurements of reference zircon FC1, R33, and Temora produced mean 176Hf/177Hf ratios of 0.282163 ± 53, 0.282754 ± 37, and 0.282663 ± 41, respectively (Table S2), all of which are within uncertainty of the accepted values (Woodhead and Hergt 2005; Fisher et al. 2014).

We calculated the ratio 176Hf/177Hf at the interpreted time of zircon crystallization using the measured and corrected 176Hf/177Hf and 176Lu/177Hf values and the 176Lu decay constant of 1.867 × 10−11 year−1 (Scherer et al. 2001; Soderlund et al. 2004). Epsilon Hf values were calculated using the isotopic compositions of the chondritic uniform reservoir given by Bouvier et al. (2008). The depleted mantle reservoir for hafnium follows a linear evolution starting with separation from the chondritic uniform reservoir at 3800 Ma to an εHf value of +16 today (Vervoort et al. 2014). Present-day values for the depleted mantle reservoir are 176Hf/177Hf = 0.283238 and 176Lu/177Hf = 0.039755 (Goodge et al. 2017).

3.7 Uranium/lead data analysis and presentation

We prepared concordia and weighted mean diagrams using Isoplot version 4.15 (Ludwig 2008). The weighted means were calculated with weighting according to the square of the internal uncertainties. The total uncertainty on the determination of the eruption age of the protolith for sample Sabot1 was calculated by quadratic addition of the measurement (internal) and systematic (external) uncertainties.

3.8 Oxygen isotopes in multi-grain quartz separates

In this study, we compare the oxygen isotope composition of coexisting quartz and zircon to test for post-intrusive or post-eruptive isotopic alteration of the quartz. We isolated quartz from the portion of each sample that floated in methylene iodide during zircon separation. These low-density grains were mostly quartz and plagioclase, which are difficult to distinguish optically. To make the quartz and plagioclase appear different under the stereographic microscope, we bathed the separate in concentrated tetra-fluoroboric acid at room temperature for 10 min, which did not affect the appearance of the quartz but caused an opaque white crust to grow on the surfaces of the plagioclase. We then picked quartz grains by hand. Sample Mont1 did not yield sufficient quartz for analysis.

We analysed oxygen isotopes in approximately 1–2 mg of quartz from each of the four remaining samples at the University of Wisconsin Madison using the rapid heating, defocused beam method for laser fluorination described by Spicuzza et al. (1998). This technique utilizes a defocused CO2 laser beam to heat the quartz aliquot in an atmosphere of BrF5. Oxygen isotopes were analysed with a Finnigan MAT 251 mass spectrometer. Repeated analyses of the UWG-2 garnet standard before and after the analyses of the Goochland Terrane quartz were used to correct the raw quartz analyses using the accepted UWG-2 δ18O value of 5.8‰ VSMOW (Valley et al. 1995). The two standard deviation value around the mean of the eight analyses of UWG-2 throughout the single day of analysis of Goochland Terrane quartz was ±0.17‰. We take this value as the uncertainty of the quartz O isotope measurements (Table S3). We report and discuss all results relative to VSMOW.

4. Results

4.1 Geochemical classification and tectonic setting

Figure 3 shows thin section photomicrographs of each sample. Table S1 contains element abundances for the Sabot Amphibolite samples and Figure 4 shows that the mafic Sabot Amphibolite samples are mostly basalt and basaltic andesite (57–42 weight per cent SiO2), whereas the felsic sample is rhyolite (75 weight per cent SiO2). Owens and Tucker (2003) found that the composition of the State Farm Gneiss ranges from granite to quartz monzodiorite (77–56 weight per cent SiO2) and the compositions of the Neoproterozoic granitoid range from granite to syenite (77–60 weight per cent SiO2).

We used the discrimination schemes described in Agrawal et al. (2008), Verma and Verma (2013), and Verma et al. (2013) to help provide information about the tectonic setting during generation of these magmas. Most of the felsic and intermediate samples from the State Farm Gneiss plot in the within-plate field (combined continental rift and ocean island field), although in some discrimination diagrams some samples fall in the combined island arc and continental arc field (Tables S4, S5; Figure 5). Most Neoproterozoic granitoid samples, both felsic and intermediate, plot in the combined continental rift and ocean island field (Tables S6, S7; Figure 5). The mafic samples from
the Sabot Amphibolite mostly fall in the mid-ocean ridge field; a few samples plot in the continental rift field (Table S8; Figure 5(c)). Figure 5 shows only three of the many possible discrimination diagrams; tables S4-S8 tabulate the results of all the discriminations.

4.2 Sample Sabot1 field relations and uranium/lead isotopic dates in spots in zircon

Near the location of samples 512001 and Sabot1, multiple felsic layers are intercalated with the more abundant mafic layers typical of the Sabot Amphibolite (Figure 6). The felsic layers have thicknesses of 0.3–4 m whereas the mafic layers are 3–5 m thick. At least some of the felsic layers appear to be composites made up of thinner felsic layers with thicknesses of 5–10 cm (Figure 6(b)). The contacts between the mafic and felsic layers are parallel to the foliation in the rocks; the felsic layers nowhere cut across the mafic layers. These contacts are sharp and planar at the mesoscopic scale seen in the field (Figure 6(a, b)), but the contacts undulate at the microscopic scale in thin section (Figure 6(c)). At the microscopic scale, some contacts show interfingering of layers with the two compositions (Figure 6(d)).

The eruption age of the volcanic protolith of the Sabot Amphibolite was unknown prior to the current study, so our work on this unit first focused on determining the age of volcanic eruption via U/Pb dating of spots in zircon obtained from a felsic layer (sample Sabot1). Cathodoluminescence images of zircon revealed
Figure 4. Classification diagram for the Sabot Amphibolite samples. Each square represents one of the whole-rock analyses in Table S1. Element abundances were normalized to a 100% total, volatile-free. The Catoctin Formation fields, shown here for comparison to the Sabot Amphibolite, include the compositions of most extrusive rocks and related dikes in Espenshade (1986), Burton et al. (1995), Badger and Sinha (2004), Southworth et al. (2006), and Badger et al. (2010). Classification scheme from Le Maitre (2002).

Figure 5. Tectonic discrimination diagrams for analyses of whole rocks. (a) Discrimination of tectonic setting for the felsic samples from the State Farm Gneiss and Neoproterozoic granitoid using combined immobile major and trace elements (Verma et al. 2013). Six of the 11 State Farm Gneiss samples plot in the within-plate field (combined continental rift and ocean island field). Discriminant functions given in Table S4. (b) Discrimination for the intermediate samples from the State Farm Gneiss and Neoproterozoic granitoid using only major elements (Verma and Verma 2013). Discriminant functions given in Table S5. (c) Discrimination for the mafic samples from the Sabot Amphibolite using only trace elements (Agrawal et al. 2008). Discriminant functions given in Table S8. The discriminant function values are dimensionless.
complex zoning in most grains, with most displaying different cathodoluminescence responses from core and rim domains (Figure 7). In some grains, inclusions are more abundant in cores than rims (Figure 7(a)). Figure 8 and Table S9 present the results of 35 spot analyses in 23 grains from sample Sabot1. Figure 9 shows that all analyses older than ca. 500 Ma came from zircon core domains and that grain cores mostly have lower U/Th ratios than grain rims. Figure 9 also shows that analyses that yielded \(^{206}\text{Pb}/^{238}\text{U}\) dates younger than ca. 453 Ma ranged to higher U/Th ratios and had a higher mean U/Th ratio compared to older analyses.

4.3 Whole-rock neodymium isotopes

Table S10 presents the results of the Sm-Nd isotope analyses of the Sabot Amphibolite whole-rock samples. Figure 10 shows these data as well as the whole-rock Nd isotope data from the older Proterozoic units in the Goochland Terrane (Owens and Samson 2004). The \(\varepsilon\text{Nd}\) values for the Sabot Amphibolite at the time of eruption (550 Ma) ranged from +6.3 to +1.5. The two most positive Sabot samples fall within the range of values for the depleted mantle at 550 Ma, considering uncertainties. All the Sabot Amphibolite analyses yielded initial values more positive than the initial \(\varepsilon\text{Nd}\) values for the older Proterozoic rocks. Initial \(\varepsilon\text{Nd}\) values for the 1040 Ma Montpelier Anorthosite, 1030 Ma State Farm Gneiss, and 600 Ma Neoproterozoic granitoid were +1.1 to +0.7, +1.2 to −0.2, and +0.6 to −0.3, respectively. None of the initial \(\varepsilon\text{Nd}\) values from these older Proterozoic units were near depleted mantle values at the time of intrusion.

4.4 Uranium/lead and hafnium isotopes in spots in zircon

Tables S2 and S3 and Figures 11 and 12 present the results of the laser ablation split stream analyses of spots in zircon. Figure 13 shows images of zircon grains from each sample with \(\delta^{18}\text{O}\) values, U/Pb dates, and \(\varepsilon\text{Hf}\) values from the indicated spots.

All analysed zircon from Montpelier Anorthosite sample Mont1 had low uranium contents, from 53 to 10 ppm (see also Aleinikoff et al., 1996). Consequently, the \(^{207}\text{Pb}\) signals during many of our analyses were insufficient to reliably determine crystallization dates, and we successfully determined dates from only five spots. All five analyses are
strongly discordant, with Proterozoic $^{206}\text{Pb}/^{207}\text{Pb}$ dates and ca. 300 Ma $^{206}\text{Pb}/^{238}\text{U}$ dates (Figure 11(a); Tables S2, S3). εHf values at 1040 Ma, the approximate time of intrusion, ranged from +12.9 to +1.4 (Tables S2, S3; Figure 12). The three most positive εHf values were near the depleted mantle value at 1040 Ma.

All but three analyses from State Farm Gneiss sample 511005 are concordant; the three discordant analyses plot only slightly off concordia (Tables S2, S3; Figure 11(b)). Most of the analyses give dates between 1050 and 1000 Ma, near the published intrusion age of ca. 1050–1010 Ma (Owens and Tucker 2003). Spot 30 in the core of grain big_20 is concordant and, at ca. 990 Ma, slightly younger than most other analyses. The U concentration and U/Th ratio of this grain is much higher than all other analyses from this sample. We interpret this crystal to be of metamorphic origin. We use 1030 Ma to calculate εHf values from all other spots; these values ranged from +2.7 to −1.5 at this time (Tables S2, S3; Figure 12). These near-chondritic values were about 10 εHf units less than the depleted mantle value at 1030 Ma.

The U/Pb isotope data from Neoproterozoic granitoid sample 511004 define two age groups (Tables S2, S3; Figure 11(c)). The first group consists of the two analyses

Figure 7. Images of four zircon crystals from Sabot Amphibolite felsic sample Sabot1. Grain numbers correspond to those in Table S9. The left-hand column contains backscattered electron images and the right-hand column contains cathodoluminescence images. Circles represent 10 μm-diameter laser ablation spots that yielded the $^{206}\text{Pb}/^{238}\text{U}$ date shown.
from the core of grain 22, which plot near concordia and have \(^{206}\text{Pb}/^{207}\text{Pb}\) dates of ca. 1130 and 1108 Ma. The U concentrations of these two spots are lower than the concentrations in all other analysed grains from this sample. We interpret the core of grain 22 to be a xenocryst inherited from Late Mesoproterozoic igneous rocks such as the State Farm Gneiss. Owens and Tucker (2003) also argued for inheritance of Mesoproterozoic zircon in this granitoid body. The Montpelier Anorthosite is not a viable source for the zircon xenocryst because the sample 511004 grain 22 core U concentrations of 368 and 190 ppm are much higher than the U concentrations in zircon from the Montpelier Anorthosite. In contrast, the grain 22 core U contents are similar to those in zircon from the State Farm Gneiss. We use an age of 1035 Ma to calculate the \(\varepsilon\)Hf values for these two spots. The \(\varepsilon\)Hf values at 1035 Ma were +1.7 and +1.2, consistent with derivation from the State Farm Gneiss or a similar source. The second age group consists of all other analyses from this sample. Many of these analyses are concordant with dates between 680 and 620 Ma. Discordant analyses plot slightly off concordia with most \(^{206}\text{Pb}/^{207}\text{Pb}\) dates between 700 and 620 Ma. Owens and Tucker (2003) interpreted the \(^{206}\text{Pb}/^{207}\text{Pb}\) date of ca. 654 Ma from a single concordant zircon analysis from this granitoid body to date intrusion. Using this previous dating as a guide, we interpret all Group 2 analyses to result from zircon crystallization at ca. 654 Ma followed by minor lead loss from many of the grains. \(\varepsilon\)Hf values of these analyses at 650 Ma ranged from +6.0 to +0.8 (Figure 12).

The U/Pb isotope data from Neoproterozoic granitoid sample 511001 similarly define two age groups (Tables S2, S3; Figure 11(d)). The first group includes the three concordant analyses from three separate grains that yielded \(^{206}\text{Pb}/^{207}\text{Pb}\) dates between 1091 and 1005 Ma. This group also contains the discordant analyses that yielded \(^{206}\text{Pb}/^{207}\text{Pb}\) dates between 1062 and 730 Ma and \(^{206}\text{Pb}/^{238}\text{U}\) dates between 696 and 577 Ma. Analyses of spots in the cores of grains big_5 and big_12 also are included in this group because outer parts of these grains have \(^{206}\text{Pb}/^{207}\text{Pb}\) dates of 827 and 922 Ma, respectively. Similarly, we include analyses big_9_15, big_11_21, and big_14_27 in this group because neighbouring spots in the sample ca. 1000 Ma to result from inheritance of Mesoproterozoic zircon and the lower intercept of 600 + 7/-9 Ma to be the time of granitoid intrusion. Guided by this previous dating, we interpret Group 1 zircon to be xenocrysts inherited from Late Mesoproterozoic Goochland Terrane rocks. Nine of the 21 analyses in Group 1 have U contents between 60 and 22 ppm. These low concentrations overlap those of zircon in the Montpelier Anorthosite but are lower than the concentrations in zircon from the State Farm Gneiss, Neoproterozoic

**Figure 8.** (a) and (b) Concordia plots for 10 μm-diameter zircon U/Pb analyses from Sabot Amphibolite felsic sample Sabot1. Pb* refers to radiogenic lead. In (a), analyses coloured black are interpreted to result from lead loss and/or mixture of igneous and metamorphic zircon during laser ablation. (c) Weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) date (horizontal green bar) for the 12 low U/Th, concordant analyses of zircon cores (vertical red bars) that we use to determine the eruption age of the volcanic protolith of the Sabot Amphibolite.
granitoid sample 511004, and the other zircon grains in sample 511001. We conclude that these low U grains were inherited from the Montpelier Anorthosite or a similar source whereas the higher U xenocrysts were inherited from the State Farm Gneiss or a similar source. εHf values of the Group 1 analyses at 1035 Ma ranged between +11.7 and +2.5, consistent with derivation from the Montpelier Anorthosite. εHf values more positive than +2.7 do not overlap the εHf values of State Farm Gneiss zircon. The remaining analyses comprise Age Group 2. These analyses are concordant or slightly discordant and have ^{206}Pb/^{207}Pb dates between 694 and 556 Ma. Based on

Figure 9. U/Th ratio versus ^{206}Pb/^{238}U date for all spot analyses of zircon from Sabot Amphibolite sample Sabot1. All analyses older than ca. 500 Ma came from grain cores. Analyses younger than ca. 453 Ma produced a wider range in U/Th ratio as well as a higher mean U/Th ratio than older analyses. Grain rims mostly have higher U/Th ratios than cores. Uncertainties are not shown for clarity but are included in Table S9. Pb* refers to radiogenic lead.

Figure 10. Whole-rock initial εNd values for the Goochland Terrane Proterozoic units. Only the Sabot Amphibolite protolith had initial εNd values near depleted mantle values at the time of intrusion or eruption. The Catoctin Formation meta-basalt field encompasses values calculated at 570 Ma in Badger et al. (2010). The Blue Ridge meta-anorthosite field is the analysis of Roseland Anorthosite sample H-22–81 calculated at 1040 Ma (Pettingill et al. 1984). The Blue Ridge meta-granitoid field includes five analyses of 1018 Ma granitoid in Fullagar et al. (1997) and two analyses of 1056 and 1019 Ma granitoid in Fisher et al. (2010), all calculated at 1030 Ma. The three fields for Blue Ridge Province rocks are shown for comparison to the values from the Goochland Terrane rocks.
the previous dating, we interpret Group 2 analyses to be zircon that crystallized at ca. 600 Ma. The εHf values at 600 Ma were +0.7 to −1.4 (Figure 12).

Only one analysis from sample 512001, spot 18 in grain 14, yielded concordant dates near the Sabot Amphibolite protolith eruption age of 552 ± 11 Ma (Tables S2, S3; Figure 11(e), 13(i); see Section 5.1 for discussion of eruption age). At 90 ppm, this spot has the lowest U concentration of any spot measured in sample 512001. We interpret this part of grain 14 to be magmatic zircon that crystallized near the time of eruption of the Sabot Amphibolite protolith. The εHf value of this spot at 550 Ma was +8.8. The other analyses from sample 512001 fall into two groups as follows: (1) Discordant analyses with 206Pb/207Pb dates between ca. 626 and 557 Ma; and (2) Concordant analyses with dates between about 500 and 450 Ma. All but one spot in Group 1 have U/Th ratios between 6.9 and 1.1, whereas U/Th ratios in Group 2 zircon range as high as 37.6. We interpret Group 1 to be zircon that crystallized near the time of protolith eruption and then lost lead during ca. 450 Ma metamorphism. We take Group 2 to be metamorphic zircon (see Section 5.1 for discussion of the age of metamorphic zircon growth). εHf values for Group 1 zircon were +12.1 to +9.0 at 550 Ma.

### 4.5 Oxygen isotopes in zircon and quartz

Tables S3 and S11 and Figure 14 present our oxygen isotope data. Figure 13 shows images of zircon crystals from each sample along with oxygen isotope values for the indicated spots. There is a trend of lower δ18O values in zircon from the three progressively younger groups of rocks (State Farm Gneiss, Neoproterozoic granitoid, Sabot Amphibolite; Figure 14).
\( \delta^{18}O \) values from Montpelier Anorthosite sample Mont1 zircon occupy the range +8.4 to +7.6‰ (Tables S3, S11). This sample did not yield sufficient quartz for analysis.

\( \delta^{18}O \) values from State Farm Gneiss sample 511005 zircon range from +9.7 to +6.4‰ (Tables S3, S11). Two U/Pb isotopic analyses from this sample were more than 20% discordant (analyses big_7_9 and big_21_31) and spot big_20_30 was located in a metamorphic grain (Table S3; Figure 11(b)). Excluding the values from these three spots as well as values from spots without U/Pb isotope data, most \( \delta^{18}O \) values from sample 511005 zircon fall in the range +8.3 to +7.2‰ (Figure 14). There are also three isotopically heavier outliers at +9.7, +9.0, and +8.7‰. The \( \delta^{18}O \) value for the multi-grain quartz separate from this sample was +10.0‰ (Table S3).

Analyses of zircon from Neoproterozoic granitoid sample 511004 yielded \( \delta^{18}O \) values between +6.7 and +1.8‰ (Tables S3, S11). Removing xenocrystic grain 22, analyses more than 20% discordant, and spots without U/Pb isotope data leaves a still-large span from +6.4 to +1.8‰. The multi-grain quartz separate yielded a \( \delta^{18}O \) value of +5.4‰ (Table S3).

Analyses of zircon from Neoproterozoic granitoid sample 511001 yielded \( \delta^{18}O \) values between +8.3 and +6.2‰ (Tables S3, S11). Removing xenocrysts, analyses more than 20% discordant, and analyses without U/Pb isotope data from the same spot leaves a range between +6.8 and +6.2‰ (Figure 14). The \( \delta^{18}O \) value for the multi-grain quartz separate from this sample was +10.0‰ (Table S3).

All except three spots in zircon from Sabot Amphibolite felsic sample 512001 yielded \( \delta^{18}O \) values between +9.5 and +7.2‰ (Tables S3, S11). The three exceptions are spots 18, 18b, and 18e in the interior of grain 14, which gave values of +5.6, +5.3, and +5.1‰, respectively (Tables S3, S11; Figure 13(i)). The multi-grain quartz separate from sample 512001 yielded a \( \delta^{18}O \) value of +12.3‰ (Table S3).

5. Discussion

### 5.1 Eruption age of the Sabot Amphibolite protolith

Before discussing the zircon U-Pb isotope data, we consider the following three endmember interpretations of the contacts between the felsic and mafic layers in the Sabot Amphibolite. (I) The two rock types belong to entirely separate units, and all contacts between the felsic and mafic layers are now and have always been high strain zones. None of the contacts were ever intrusive or depositional. (II) The felsic layers were originally...
Figure 13. Backscattered electron (left column) and cathodoluminescence (right column) images of some zircon grains from each of the Goochland Terrane samples excluding sample Sabot1. The circle indicates the analysis spot, which had a diameter of 15 μm for O and 30 or 40 μm for U/Pb and Hf isotopes. The date is the $^{206}\text{Pb}/^{207}\text{Pb}$ date.
intrusions into the mafic volcanic rocks. The contacts between the felsic and mafic layers are parallel to the tectonic foliation within the layers due to post-intrusion deformation. (iii) The felsic layers are volcanic rocks erupted at broadly the same time as the mafic rocks; the contacts between the felsic and mafic layers were originally depositional. Elements of these three end-member options can be combined into other alternatives. For example, a high strain zone can overprint an original depositional or intrusive contact. Another alternative is that option II is correct for some felsic layers and option III for others.

All the Proterozoic rocks in the Goochland Terrane were metamorphosed and deformed pervasively and intensely, mostly obscuring original intrusive and depositional contacts. Consequently, we cannot rule out high strain zone or intrusive contacts between all the felsic and mafic layers in the Sabot Amphibolite. However, we discount option I for the following reasons. (A) The rocks near the contacts do not appear more highly strained than the rocks distal to the contacts. (B) The contacts are sharp in outcrop; at this scale, the different compositions are not intermingled as often occurs in high strain zones. The interfingering of the two compositions at the millimetre scale, however, could have resulted from deformation focused on the contact. (C) There is no independent evidence for the extreme interpretation that the felsic and mafic layers formed entirely separately with no original relationship between them.

We prefer the interpretation that these were originally depositional, not intrusive, contacts between felsic and mafic volcanic rocks for the following reasons. (A) Some of the metre-scale felsic layers appear to have been built from successive centimetre- to decimetre-scale felsic layers, which is expected for deposition of volcanic rocks but may not be as common in dikes or sills. (B) The microscopic interfingering of felsic and mafic layers is consistent with deposition of pulses of mafic and felsic lavas. Accordingly, throughout the remainder of the article, we use the interpretation that the protoliths of these rocks were interbedded felsic and mafic volcanic strata that erupted at broadly the same time. If our interpretation is incorrect and the dated felsic layer actually was an intrusion into the mafic rocks, then the crystallization age of the felsic layer is a minimum age for eruption of the mafic volcanic rocks.

There are two plausible interpretations of the age significance of the U-Th-Pb isotope data from sample Sabot1. (1) The old cores with low U/Th ratios crystallized from the magma that formed the volcanic protolith whereas both the rims and young grains with high U/Th ratios grew during metamorphism. (2) The old, low U/Th cores were inherited from older Proterozoic igneous rocks in the
Goochland Terrane whereas the high U/Th rims and young grains crystallized from the magma that formed the volcanic protolith. We prefer the first option for the following four reasons. First, the U/Th ratio in igneous zircon typically is lower than in metamorphic zircon (Hoskin and Schaltegger 2003). Using option one, the low U/Th analyses correspond to igneous zircon and the high U/Th analyses to metamorphic zircon, in accordance with the conclusions of Hoskin and Schaltegger (2003, and references therein). In contrast, following option two, the high U/Th analyses came from igneous zircon. Second, our measured U/Th ratios in igneous zircon from the Montpelier Anorthosite, State Farm Gneiss, and Neoproterozoic granitoid range from 1.2 to 7.6 (Table S2; see also Owens and Tucker 2003). The U/Th ratios for the Sabot1 core analyses we use to date eruption range from 0.8 to 4.3; three analyses yielded U/Th ratios less than 1.2 (Table S9). It is difficult to explain the fact that 3 of these 12 Sabot Amphibolite zircon cores have U/Th ratios less than 1.2, but none of the 108 analyses of zircon from the older Proterozoic rocks yielded a U/Th ratio less than 1.2 if these Sabot Amphibolite grains were inherited from the older igneous rocks. Third, the Montpelier Anorthosite and State Farm Gneiss contain zircon that crystallized at ca. 1050–1010 Ma (Table S3; Owens and Tucker 2003). Both studied Neoproterozoic granitoid bodies also bear Late Mesoproterozoic zircon xenocrysts in addition to ca. 650–600 Ma zircon (Table S3; Owens and Tucker 2003). If the volcanic protolith of the Sabot Amphibolite inherited zircon from these sources, we would expect Mesoproterozoic and/or 650–600 Ma zircon in the Sabot Amphibolite as well, but we found no zircon of these ages in sample Sabot1. Nearly complete resetting of conjectural older Proterozoic xenocrystic zircon in the Sabot Amphibolite by loss of nearly all previously incorporated lead is not a compelling argument because there is no obvious mechanism to reset older Proterozoic zircon in the Sabot Amphibolite but not in the underlying State Farm Gneiss or Neoproterozoic granitoid. Finally, fourth, option two implies little or no growth of zircon during metamorphism to amphibolite and possibly granulite facies. However, we might expect metamorphic zircon crystallization in the Sabot Amphibolite felsic unit because zircon grew during metamorphism at these conditions in felsic rocks in other locales (e.g. Ustaomer et al. 2012; He et al. 2016; Tollo et al. 2017).

Accepting interpretations III and I, we take the eruption age of the volcanic protolith of the Sabot Amphibolite felsic unit to be the weighted mean of the 12 old, low U/Th core analyses that are at least 98% concordant (Figure 8(c)). This age is 552 ± 11 Ma (MSWD = 0.67). This study did not aim to date the metamorphism of the Sabot Amphibolite, and consequently the study’s design produced interpretations of the age(s) of metamorphism that are less well determined than the protolith eruption age. Nonetheless, we tentatively interpret the cluster of seven high U/Th analyses that are at least 97% concordant between ca. 451–437 Ma to date growth of metamorphic zircon (Table S9). We also suggest that the following analyses may date growth of zircon during metamorphism: (A) The two high U/Th rim analyses at ca. 425–423 Ma that are 96–97% concordant, and (B) The two high U/Th rim analyses at ca. 415–414 Ma that are 96–98% concordant. The discordant dates that fall between these age groups resulted from lead loss and/or mixing of age domains during laser ablation. The discordant dates at ca. 414 and 400 Ma resulted from lead loss.

5.2 Interpretation of zircon spot O isotope measurements

All U/Pb isotopic analyses from Montpelier Anorthosite sample Mont1 were very discordant (Table S3; Figure 11(a)). Accordingly, we interpret this sample’s zircon δ18O values of +8.4 to +7.6‰ to result from alteration after intrusion.

We interpret the main span of values from +8.3 to +7.2‰ in State Farm Gneiss sample 511005 zircon to be the result of the original magmatic crystallization of zircon. The three more positive outliers at +9.7, +9.0, and +8.7‰ may have resulted from unrecognized alteration of the zircon.

For zircon from Neoproterozoic granitoid sample 511004, plotting the OH/O ratio versus the δ18O of the concordant, non-xenocrystic analyses reveals a linear trend, with higher OH/O ratios corresponding to less positive δ18O values (Figure S2). This trend indicates that alteration of zircon from this sample resulted in addition of water and a decrease of δ18O values. We therefore take the eight most positive of these δ18O values as the best indicator of values during zircon crystallization in the sample 511004 magma (Figure S2). These eight values range between +6.4 and +5.9‰ (Figure 14).

Analyses of zircon from Neoproterozoic granitoid sample 511001 yielded a narrow range of δ18O values, from +6.8 to +6.2‰. We interpret these values to have been set during crystallization of zircon in the sample 511001 magma.

Spot 18 was the only analysis in any grain from sample 512001 of the Sabot Amphibolite that produced concordant U/Pb isotopic dates near the eruption age of the Sabot Amphibolite protolith; all other analyses were discordant or came from younger, metamorphic zircon (Table S3; Figure 11(e)). Accordingly, we interpret the δ18O values from spots 18, 18b, and 18e to be the result of zircon crystallization in the Sabot Amphibolite...
protolith magma (Figure 14). These values overlap the range of $\delta^{18}$O values from zircon equilibrated with a mantle composition (Valley et al. 1998). We interpret the $\delta^{18}$O values from all other spots to result from alteration or growth of metamorphic zircon. Spot 18c in the core of grain 14 had a $\delta^{18}$O value of +7.5‰ whereas neighbouring spots 18, 18b, and 18e in the interior of the grain gave $\delta^{18}$O values between +5.6 and +5.1‰ (Figure. 13(i) and 13(j)). Alteration of the O isotopic composition at spot 18c may have been more effective than at spots 18, 18b, and 18e because of higher U concentration at spot 18c, as indicated by the dark cathodoluminescence response at spot 18c compared to the generally bright response at spots 18, 18b, and 18e.

5.3 Evolution of Goochland Terrane continental crust

Our hafnium and uranium/lead isotope analyses in spots in zircon suggest the tectonic evolution illustrated in Figure 15. This interpretation is supported by our zircon oxygen isotope, whole-rock neodymium isotope, and whole-rock major and trace element abundance data.

Initial separation of Goochland Terrane crust from the depleted mantle occurred prior to intrusion of the Montpelier Anorthosite and State Farm Gneiss (Figure 12, 15(a)). Although the age of this first crust formation is poorly known, in Figure 12 we show it near 1400 Ma because this is the Nd isotope model age for separation of State Farm Gneiss whole rocks from the depleted

Figure 15. Illustration of the interpreted magmatic evolution of the Goochland Terrane during Proterozoic time. Only the State Farm Gneiss protolith magma was derived entirely from melting of preexisting crust. All other units were the result of mixing melt from the depleted mantle with preexisting crust.
mantle (Owens and Samson 2004). Note, however, that the depleted mantle model age is not necessarily the actual time of derivation of continental crust from the mantle; the model age can be compromised by many processes including derivation from mantle with an isotopic composition different from that of the model depleted mantle, as well as mixing preexisting continental crust with mantle melt. We do not ascribe tectonic significance to the model age because of this uncertainty. No one has identified pre-latest Mesoproterozoic rocks or xenocrysts in the Goochland Terrane, so the tectonic setting during this initial crust formation event is unknown.

All our chemical evidence is consistent with derivation of the magma that became the State Farm Gneiss from this initial Goochland Terrane crust with no additional sources required. Both the small spread of zircon spot initial εHf values (Figure 12) and the small range of whole-rock initial εNd values (Figure 10) point to extraction from a single source of melt. These values are much less positive than latest Mesoproterozoic depleted mantle values, indicating that the melt source was preexisting crust (Figure 12). Likewise, the zircon spot O isotope values between +8.3 and +7.2‰ are consistent with derivation from preexisting crust. In contrast, the magmas that became the four other Proterozoic units in the Goochland Terrane probably were produced by mixing melt derived from the depleted mantle with preexisting crust.

The discordancy of the U/Pb dates from all the Montpelier Anorthosite zircon complicates interpretation of both the U-Pb and Lu-Hf isotope data from this sample’s zircon (Figure 11(a); Tables S2, S3). Two interpretations of these data are plausible. (A) The analysed zircon crystallized during anorthosite intrusion at 1045 ± 10 Ma and lost Pb during metamorphism at ca. 300 Ma. The array of initial εHf values in zircon spots that stretch from depleted mantle values to +1.4 resulted from mixing of a melt from the depleted mantle with preexisting Goochland Terrane crust (Figure 12; Table S3). (B) The analysed zircon crystallized during metamorphism at ca. 300 Ma.

Arguments in favour of an igneous origin include the following. (1) Option A is consistent with the interpretation by Aleinikoff et al. (1996) that the ca. 1045 Ma igneous zircon lost a small amount of Pb during metamorphism at ca. 300 Ma. These authors discussed mechanisms for Pb loss from Montpelier Anorthosite zircon despite low U contents. (2) An igneous origin also is consistent with the fact that these authors did not find metamorphic zircon with an age of ca. 300 Ma. (3) Interpretation A provides an explanation for the result that the $^{206}\text{Pb} / ^{207}\text{Pb}$ dates are older than the $^{206}\text{Pb} / ^{238}\text{U}$ dates in every grain. (4) The interpretation that mixing of melt from the depleted mantle with preexisting crust produced the large spread of initial εHf values is identical to one explanation for the abundance of quartz as well as the high K and Ba concentrations in the Montpelier Anorthosite (Owens and Dymek 2016). (5) The ends of the range of the zircon spot εHf values at 1040 Ma match the known values of the possible mixing components, the depleted mantle and the initial Goochland Terrane crust. If the spread of εHf values at 1040 Ma actually resulted from assigning an incorrect crystallization age to the zircon, we might expect to calculate εHf values much more or less positive than these two components.

On the other hand, points that support a metamorphic origin include the following. (1) Metamorphic crystallization at ca. 300 Ma obviates the need to call for nearly complete Pb loss from the zircon despite low U concentrations. (2) Interpretation B resolves the apparent conflict between the presence of a large range in zircon spot εHf values but little spread in whole-rock εNd values at 1040 Ma.

We cannot rule out either possibility. However, we favour interpretation A because it is most consistent with the previously published data and interpretations. If an igneous origin is correct, one speculative way to explain the large spread in zircon spot εHf values but not in whole-rock εNd values at 1040 Ma is zircon crystallization both before and during assimilation of the crustal rocks into the magma. After complete assimilation, the crustal component dominated the Nd isotopic composition of the magma as a whole.

Neoproterozoic granitoid samples 511004 and 511001 both contain zircon xenocrysts derived from the State Farm Gneiss or crust of similar age. Sample 511001 additionally contains zircon xenocrysts inherited from the Montpelier Anorthosite or similar rocks. The Montpelier Anorthosite is geographically closer in map view to the granitoid body at location 511001 than it is to the body at location 511004 (Figure 2). This proximity may explain the presence of zircon derived from the Montpelier Anorthosite in sample 511001 but not sample 511004. Zircon in the Neoproterozoic granitoid bodies had εHf values between +6.0 and −1.4 at the time of intrusion, whereas the State Farm Gneiss whole-rock is estimated to have had an εHf value near −5 at 650–600 Ma (Figure 12). Although inadvertent mixing of different isotopic domains in zircon during laser ablation may have produced part of the large spread of initial εHf values in zircon from sample 511004, none of the initial values of zircon from either Neoproterozoic granitoid body was near the estimated State Farm Gneiss whole-rock εHf value at 650–600 Ma. The more positive initial εHf values of the Neoproterozoic granitoid zircon require mixing of the State Farm Gneiss with a source more enriched in $^{176}\text{Hf}$.
produce the Neoproterozoic granitoid magma. We infer that this more enriched source was melt from the depleted mantle (Figure 12, 15(d)). Whole-rock initial εNd values near zero (Figure 10) similarly point to derivation from a source enriched in 143Nd with the latest Mesoproterozoic rocks, as also inferred by Owens and Samson (2004). Because of the lower initial values, our zircon Hf isotope data suggest a greater contribution of latest Mesoproterozoic granitoid bodies range from +6.8 to +5.9‰, consistent with at least partial derivation of the Neoproterozoic granitoid magmas from preexisting crust.

The range of initial εHf values of spots in zircon from Sabot Amphibolite felsic sample 512001 from near depleted mantle values to +8.8 indicates derivation of melt dominantly from the depleted mantle, with a contribution from the older Goochland Terrane rocks (Figure 12, 15(e)). Whole-rock initial εNd values show a similar spread from depleted mantle values to +1.5, likewise revealing mixing of these two sources (Figure 10). O isotope values in the zircon in the two Neoproterozoic granitoid bodies range from +6.8 to +5.9‰, consistent with at least partial derivation of the Neoproterozoic granitoid magmas from preexisting crust.

The range of initial εHf values of spots in zircon from Sabot Amphibolite felsic sample 512001 from near depleted mantle values to +8.8 indicates derivation of melt dominantly from the depleted mantle, with a contribution from the older Goochland Terrane rocks (Figure 12, 15(e)). Whole-rock initial εNd values show a similar spread from depleted mantle values to +1.5, likewise revealing mixing of these two sources (Figure 10). O isotope values in the zircon in the two Neoproterozoic granitoid bodies range from +6.8 to +5.9‰, consistent with at least partial derivation of the Neoproterozoic granitoid magmas from preexisting crust.

The tectonic setting for all of this latest Mesoproterozoic and Neoproterozoic magmatism may have been extensional. The tectonic environment for the oldest magmatism is the most poorly known. However, like other latest Mesoproterozoic anorthosite and related granitoid in eastern North America, Bartholomew and Hatcher (2010) and McLelland et al. (2010) inferred that the Montpelier Anorthosite and State Farm Gneiss magmas formed during extension related to Grenvillian post-orogenic collapse. Tectonic discrimination using trace elements from the mafic parts of the Sabot Amphibolite indicates a mid-ocean ridge affinity (Figure 5(c); Table S8), consistent with derivation mostly from the depleted mantle (Salters and Stracke 2004).

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Accepting the igneous interpretation for the origin of the Montpelier Anorthosite zircon, punctuated addition of juvenile continental crust to the Goochland Terrane spanned at least 500 M.y and perhaps 850 M.y or longer. Production of the melts that became the Montpelier Anorthosite at ca. 1045 Ma, the Neoproterozoic granitoid between ca. 660 and 580 Ma, and the Sabot Amphibolite protolith at ca. 550 Ma all involved a component of melt from the depleted mantle. All this magmatism was caused by processes operating during the terminal stages of the Grenville orogeny followed by continental breakup. Further back in time, initial Goochland Terrane crust was extracted from the depleted mantle prior to intrusion of the Montpelier Anorthosite and State Farm Gneiss magmas, possibly near 1400 Ma. Thus, the Proterozoic magmas of the Goochland Terrane are evidence for extension-related, punctuated addition of juvenile crust to this continental sliver over at least 500 M.y.

Although the additions of juvenile crust spanned a long time, the cumulative new continental crust produced during latest Mesoproterozoic and Neoproterozoic time appears to have been volumetrically minor because the combined modern spatial extents of the Montpelier Anorthosite, Neoproterozoic granitoid, and Sabot Amphibolite are small compared to the map extents of the State Farm Gneiss (Figure 2). Based on the relatively large area of exposed State Farm Gneiss, the volumetrically most important crust-forming event during Proterozoic time was the initial formation of Goochland Terrane continental crust prior to latest Mesoproterozoic time (Figure 15(a)). This early event could have been even more volumetrically important than suggested by the geologic map in Figure 2 if rocks like the State Farm Gneiss underlie parts of the Maidens Gneiss. Unfortunately, the tectonic setting for the initial formation of Goochland Terrane continental crust is unknown (Figure 15(a)). However, our data indicate that addition of juvenile crust during ca. 1050–550 Ma extension was volumetrically minor, consistent with the conclusions of Stern and Scholl (2010) for the modern
Earth. Our data also highlight the episodic nature of such additions.

5.4 Alteration of the oxygen isotope compositions of quartz in Goochland Terrane rocks

To assess whether the O isotope compositions of quartz in the State Farm Gneiss, Neoproterozoic granitoid, and Sabot Amphibolite were altered after cooling following intrusion or eruption, we calculated a temperature for each sample using the quartz-zircon O isotope thermometer as formulated by Valley et al. (2003). The thermometer requires equilibrium of O isotopes in quartz and zircon to yield a valid temperature. If a geologist inputs values from quartz and zircon that are not in isotopic equilibrium, the user still can mathematically operate the equation that underlies the thermometer as long as the $\delta^{18}\text{O}$ value of the quartz is more positive than that of the zircon. In the case of isotopic disequilibrium, however, the output would have no physical meaning; it would not be a temperature at which equilibrium was attained between the two minerals. For each sample, we used the mean $\delta^{18}\text{O}$ value of the spots in igneous, unaltered zircon that we use for the interpretations in section 5.3. The results are given in Table S12.

Only the quartz and zircon from Neoproterozoic granitoid sample 511001 yielded a temperature consistent with little post-cooling alteration of the quartz O isotope composition (Lackey et al. 2008). This sample produced a narrow range of zircon $\delta^{18}\text{O}$ values, from +6.8 to +6.2‰, indicating little alteration of the O isotope composition of zircon as well as quartz in this unit.

In contrast, the calculated temperature for State Farm Gneiss sample 511005 is 1240°C, hotter than granite intrusion temperatures (Lackey et al. 2008). The quartz in Neoproterozoic granitoid sample 511004 has a less positive $\delta^{18}\text{O}$ value than the zircon, which renders the thermometer inoperable. Finally, the calculated temperature for Sabot Amphibolite sample 512001 is 340°C, colder than expected for quartz that equilibrated with zircon during post-eruptive cooling (Lackey et al. 2008). The too-hot and too-cold temperatures from the State Farm Gneiss and the Sabot Amphibolite, respectively, and especially the less positive $\delta^{18}\text{O}$ value in quartz than in zircon from Neoproterozoic granitoid sample 511004, indicate that the O isotope values of quartz were altered to be out of equilibrium with those of co-existing zircon. Because quartz and zircon O isotopes are not in equilibrium in each of the three samples, the thermometer equation can yield or fail to yield a result mathematically (Table S12), but the outputs have no physical meaning. They are not temperatures of equilibrium between quartz and zircon. Instead, the calculations demonstrate post-intrusion or post-eruption alteration of the O isotope compositions of quartz in these three units. These three samples span the range of Proterozoic magmatic ages in the Goochland Terrane, suggesting that alteration of the O isotope composition of quartz may be widespread in the Proterozoic rocks of the terrane. If correct, the O isotope compositions of many of the Proterozoic whole rocks likewise may have been modified by processes such as metamorphism that occurred after intrusion or eruption. The extent of alteration of the Proterozoic rocks after intrusion or eruption thus must be assessed prior to interpreting whole-rock O isotope measurements.

5.5 Implications for the Proterozoic tectonic affinity of the Goochland Terrane

Interpretation that the Goochland Terrane was part of or near Laurentia during the Proterozoic Eon relies on geological links between these two pieces of continental crust. To ascribe a Laurentian affinity to the Goochland Terrane, geologists commonly point to similarities between the Proterozoic intrusive rocks of the Goochland Terrane and those of the nearby Blue Ridge Province (Aleinikoff et al., 1996; Farrar 1984; Glover et al. 1997; Owens and Samson 2004). This correlation mostly works well for the latest Mesoproterozoic rocks. The intrusion age and the unusual chemistry and mineralogy of the Montpelier Anorthosite match those of the Roseland Anorthosite in the Blue Ridge Province (Owens and Dymek 2016), and the intrusion age and chemistry of the State Farm Gneiss are similar to many granitic bodies in the Blue Ridge Province (Tollo et al. 2006, 2017). Whole-rock initial $\varepsilon$Nd values also overlap for the Montpelier and Roseland anorthosite bodies (Figure 10). However, the seven published analyses of ca. 1050–1010 Ma Blue Ridge meta-granitoid all have more negative initial $\varepsilon$Nd values than the State Farm Gneiss, suggesting sources with different Nd isotopic compositions (Figure 10).

Another similarity is that rocks in both the Goochland Terrane and the Blue Ridge Province were metamorphosed at ca. 1000 Ma (Aleinikoff et al., 1996; Southworth et al. 2010; Tollo et al. 2017; our analysis of State Farm Gneiss zircon grain big_20_30).

Likewise, the compositions of the Neoproterozoic igneous rocks in the Goochland Terrane are similar to those in the Blue Ridge Province. Neoproterozoic granitoid in both regions has alkaline to mildly peralkaline, A-type compositions (Owens and Tucker 2003; Tollo et al. 2004). Sabot Amphibolite mafic and felsic rocks also have similar compositions to those in the Catoctin Formation in the Blue Ridge Province to the west (Figure
6. Conclusions

Our element abundance and isotope data lead to the following conclusions.

(1) Uranium/lead isotope dating of spots in zircon from a felsic layer intercalated with the dominant mafic rocks of the Sabot Amphibolite indicate that eruption of the protolith occurred at 552 ± 11 Ma.

(2) Goochland Terrane crust initially formed prior to latest Mesoproterozoic intrusion of the Montpelier Anorthosite and State Farm Gneiss. Extraction of the initial Goochland Terrane crust from the depleted mantle might have occurred near 1400 Ma, which is the Nd isotope model age for extraction of State Farm Gneiss whole rocks from the depleted mantle.

(3) The magma that became the State Farm Gneiss protolith could have been sourced solely from partial melting of this initial Goochland Terrane crust.

(4) In contrast, the magma that became the Montpelier Anorthosite may have been a mixture of the initial Goochland Terrane crust and melt extracted from the depleted mantle. This addition of juvenile continental crust may have occurred in an extensional setting during orogenic collapse at the end of the Grenville Orogeny.

(5) The magmas that became the Neoproterozoic granitoid similarly were a mixture of melt from the depleted mantle and latest Mesoproterozoic rocks such as the Montpelier Anorthosite and the State Farm Gneiss. This addition of juvenile continental crust probably took place during early rifting of the Goochland Terrane.

(6) The mafic Sabot Amphibolite protolith had trace element abundances similar to those of mid-ocean ridge basalt and was derived dominantly from partial melting of the depleted mantle during continental rifting.

(7) In summary, the Proterozoic rocks of the Goochland Terrane represent punctuated addition of juvenile crust to this continental sliver between ca. 1050 and 550 Ma. This juvenile crust formed in extensional tectonic regimes.

(8) After intrusion or eruption, the O isotopic compositions of quartz in the State Farm Gneiss, Neoproterozoic granitoid, and Sabot Amphibolite were altered.

Highlights for Martin et al. Goochland paper

- The Goochland Terrane is a fault-bounded block in the central Appalachian Piedmont Province.
- The volcanic protolith of the Sabot Amphibolite may have erupted at 552±11 Ma.
• Juvenile continental crust was added before 1050 Ma and at 1045, 600–580, and 550 Ma.
• Terminal Mesoproterozoic and Neoproterozoic crust addition in extensional settings.
• Goochland Neoproterozoic magmatism was 10 M.y. younger than in the Blue Ridge Province.

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