Research Article

Revised Maximum Depositional Age for the Ediacaran Browns Hole Formation: Implications for Western Laurentia Neoproterozoic Stratigraphy

Ashley W. Provow,1,2 Dennis L. Newell,2 Carol M. Dehler,2 Alexis K. Ault,2 W. Adolph Yonkee,3 Stuart N. Thomson,4 and Kevin H. Mahan5

1Department of Geosciences, University of Nevada-Las Vegas, Las Vegas, NV 89154, USA
2Department of Geosciences, Utah State University, Logan, UT 84322, USA
3Department of Earth and Environmental Sciences, Weber State University, Ogden, UT 84408, USA
4Department of Geosciences, University of Arizona, Tucson, AZ 85721, USA
5Department of Geological Sciences, University of Colorado Boulder, Boulder, CO 80309, USA

Correspondence should be addressed to Dennis L. Newell; dennis.newell@usu.edu

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Constraining the depositional age of Neoproterozoic stratigraphy in the North American Cordilleran margin informs global connections of major climatic and tectonic events in deep time. Making these correlations is challenging due to a paucity of existing geochronological data and adequate material for absolute age control in key stratigraphic sequences. The late Ediacaran Browns Hole Formation in the Brigham Group of northern Utah, USA, provides a key chronological benchmark on Neoproterozoic stratigraphy. This unit locally comprises <140 m of volcaniclastic rocks with interbedded mafic-volcanic flows that lie within a 3500 m thick package of strata preserving the Cryogenian, Ediacaran, and the lowermost Cambrian history of this area. Prior efforts to constrain the age of the Browns Hole Formation yielded uncertain and conflicting results. Here, we report new laser-ablation-inductively-coupled-mass-spectrometry U-Pb geochronologic data from detrital apatite grains to refine the maximum depositional age of the volcanic member of the Browns Hole Formation to 613 ± 12 Ma (2σ). Apatite crystals are euhedral and pristine and define a single date population, indicating they are likely proximally sourced. These data place new constraints on the timing and tempo of deposition of underlying and overlying units. Owing to unresolved interpretations for the age of underlying Cryogenian stratigraphy, our new date brackets two potential Brigham Group accumulation rate scenarios for ~1400 m of preserved strata: ~38 mm/kyr over ~37 Myr or ~64 mm/kyr over ~22 Myr. These results suggest that the origins of regional unconformities at the base of the Inkom Formation, previously attributed to either the Marinoan or Gaskiers global glaciation events, should be revisited. Our paired sedimentological and geochronology data inform the timing of rift-related magmatism and sedimentation near the western margin of Laurentia.

1. Introduction

Successions of Neoproterozoic strata record dramatic global-scale tectonic, climatic, and biologic events. These include the timing and duration of Laurentia rifting from the Rodinia supercontinent, extensive glaciations that reached equatorial regions, and the proliferation of multicellular life [1–4]. Refining regional and global correlations of Neoproterozoic successions is critical to understanding these events. However, this time period has a limited geologic record that is obscured by structural complexities, extensive cover, metamorphism, and insufficient datable material. This latter point is the case for Neoproterozoic extensional and trailing edge tectonic settings, such as the North American Cordilleran margin, which are dominated by sedimentary input from older sources and generally lack a significant component of syn-depositional magmatic activity [5–10]. Along this margin, Cryogenian through basal Cambrian strata
comprise multiple tectonostratigraphic units that record the evolution from early stages of Rodinian extension and rift-ting, a failed rift, and later rifting and development of a passive margin of a supercontinent or terrane [10–13]. A similar style of fragmentation is interpreted during the separation of South China from India during this time period [14].

A key tectonostratigraphic sequence that preserves a record of the latter stages of rifting along the North American Cordilleran margin is the Brigham Group of southeast Idaho and northern Utah (Figure 1). Within this group is the Neoproterozoic Browns Hole Formation (Fm), which is one of few volcanic-bearing units along the North American

**Figure 1:** Generalized regional geologic map (modified from [10, 74]) showing the distribution of major geological packages in the northeast Utah and southeast Idaho area. This study focuses on the Neoproterozoic strata, which typically represent Cryogenian to Ediacaran aged rocks. In this figure, the early Cambrian units include the Geertsen Canyon and Camelback Mountain formations. The red square in the southeastern part of the map marks the location of the study area and Huntsville sequence (Figure 2(a)), which locally contains the volcanic-bearing strata of the Browns Hole Formation. Correlative strata (Figure 2(b)) near Pocatello, Idaho, denoted by a white square in the northeastern portion, have similar overall stratigraphy, yet the Browns Hole Formation is not present.

**Cenozoic basin fill**

**Early Cambrian-Neoproterozoic**

**Archean/Paleoproterozoic**

**Triassic-Jurassic**

**Paleozoic**

**Thrust trace**

**Inferred thrust trace**

**Normal fault**

**20 km**
Cordilleran margin. Specifically, the volcanic member of the Browns Hole Fm serves as a distinctive stratigraphic marker unit with the potential to provide geochronologic control to improve regional stratigraphic correlations and insights into rifting [10, 15–18]. To date, however, volcanic and volcanioclastic materials from this unit have yielded sparse, uncertain, and conflicting geochronologic results (Figures 2 and 3) [12, 13, 19]. Hornblende from a volcanic clast in an agglomerate, located ~40 m above the base of the contact between the Browns Hole and underlying Mutual formations near Huntsville, UT, yields a maximum depositional age of 580 ± 14 Ma (2σ) from total \(^{40}\text{Ar}/^{39}\text{Ar}\) gas analysis [12, 13]. But, apatite grains from a thin basalt flow located ~30 m above the agglomerate at this location have an isotope dilution-thermal ionization mass spectrometry (ID-TIMS) U-Pb date of 609 ± 25 Ma (2σ) (Figures 2 and 3) [19]. Although these dates overlap at the tails of their uncertainties, the juxtaposition of the younger maximum depositional age stratigraphically well below the older mean absolute age is problematic for interpreting the timing of deposition for the Browns Hole volcanioclastics and related strata.
Here, we provide multiscale characterization of the Browns Hole Fm near Huntsville, UT (Figure 1) and leverage the volcanic member to refine the maximum depositional age of this unit. Dateable detrital zircon grains are absent in this part of the unit, but pristine detrital apatite grains are abundant that we target with U-Pb analyses via laser-ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) and cathodoluminescence (CL) imaging. We explore implications of this new depositional age for regional sequence stratigraphy including sediment accumulation rates of the Brigham Group, the timing of Laurentian rifting in this region, and connections between regional unconformities and previously inferred global glaciations.

2. Tectonic-Climatic Framework and Regional Neoproterozoic Stratigraphy

Throughout the Neoproterozoic, Laurentia was subject to a myriad of tectonic and climatic processes. By the early Neoproterozoic, ca. 900 Ma, most known continental land masses assembled around Laurentia [1, 4, 20] to form the supercontinent Rodinia [4, 20]. Break-up of Rodinia likely initiated after 800 Ma, with peaks in extension-related magmatic activity between 780 and 750 Ma [4, 21–23], and the evidence for regional extension in western Laurentia includes the Gunbarrel magmatic event ca. 780 Ma in what is now western Montana and western Wyoming [6, 23–25]. Laurentia fully separated from the Amazonia and Rio de la Plata cratons [4, 26] and the supercontinent Gondwana by ca. 540 Ma [22, 27–29]. Evidence in the rock record indicates two global glaciations occurred broadly contemporaneously with these tectonic events [30–32]. These include the Sturtian glaciation from ~720 to 660 Ma [33–36] and the subsequent Marinoan glaciation between ~650 and 635 Ma [37–39]. Climatic and tectonic processes likely worked in concert to supply sediment and create accommodation space necessary to create thick packages of Neoproterozoic sedimentary rocks, including the volcaniclastic rocks that are the focus of this study.

Neoproterozoic strata preserved along the western margin of Laurentia comprise diamictite, shale, and other siliciclastic rocks, as well as carbonates, volcaniclastic rocks, and minor volcanic intervals. Yonkee et al. [10] identified four distinct stages of Neoproterozoic to early Cambrian tectonostratigraphic evolution that are represented in rocks exposed throughout southeast Idaho and northern and central Utah: (1) initial basin development; (2) early rifting and volcanism; (3) failed rifting, broad subsidence, and deposition of thick siliciclastic stratigraphic packages, including the Brigham Group; and (4) final rifting and transition to a passive margin. Although initial basin development (Stage 1) is important for the overall story of the region, this work bears on units deposited during early rifting (Stage 2), failed rifting and broad subsidence (Stage 3), and the final “rift-to-drift” transition (Stage 4).

2.1. Diamictite and Heterolithic Strata (Stage 2—Early Rift)

In the Wasatch and Bear River ranges of northern Utah, early rift units include the Perry Canyon, Maple Canyon, and Kelley Canyon formations. The Perry Canyon Fm,
which directly overlies Proterozoic crystalline basement, is highly variable in thickness and facies. This package of strata near Huntsville, UT (hereafter referred to informally as the Huntsville sequence) also contains <120 m thick volcanoclastic diamictite and siltstone that directly overlies Proterozoic crystalline basement [40]. The diamictite is over lain by <460 m of the informal greywacke member of the Perry Canyon Fm [41]. Above this is the Maple Canyon Fm, which comprises <450 m of fluvial olive-green argillite, fine-grained green arkose with interbedded conglomerate, and quartzite [42]. Finally, the overlying Kelley Canyon Fm includes <600 m of marine argillite, fine-grain sandstone, and minor carbonate (Figure 2(a)) [43]. Several interpretations exist for the depositional environment of these units including a glacier-fed sedimentary basin, eustatic sea level rise after deglaciation, and thermal or tectonic subsidence [43, 44].

The Neoproterozoic strata in southeast Idaho that make up the Pocatello Fm share many lithologic and paleoenvironmental affinities with the Huntsville sequence. The three members from base to top are the Bannock Volcanic Member (base not exposed), Scout Mountain Member, and the informal upper member. The Bannock Volcanic Member (200–450 m) grades into the Scout Mountain Member and consists of altered basalts overlain by a lower diamictite inter-bedded with rhyolitic pyroclastic beds. Above this, the 700 m thick Scout Mountain Member contains interbedded arkosic and quartzite, cobbled conglomerate, and an upper diamictite with minor dolostone. This grades into the informal upper member and, after the last carbonate, comprises a thick package of argillite and quartzite of the upper member (<300 m, Figure 2(b)) [45].

The Huntsville sequence and Pocatello Fm have yielded zircon U-Pb dates aimed as constraining their depositional ages (Figure 2). Based on detrital zircon U-Pb geochronology, volcanoclastic rocks in the Perry Canyon Fm in the Huntsville sequence have maximum depositional ages of 703 ± 12 Ma (2σ) for the glacial diamictite member and 667 ± 10 Ma (2σ) for the greywacke member (Zpgw) [10, 40]. Fanning and Link [46] constrained the maximum depositional age of the Pocatello Fm to ca. 667 Ma using sensitive, high-resolution ion-microprobe (SHRIMP) analyses of detrital zircons isolated from a bed interpreted as a reworked fallout tuff. More recent work interpreted this bed as a siltstone, and zircon U-Pb geochronology via chemical abrasion-thermal ionization mass spectrometry (CA-ID-TIMS) yielded a similar date (~675 Ma), but this maximum depositional age need not be that close to the actual depositional age [47]. Based on the stratigraphic correlation of the upper diamictite and dolostone at this location to other dated end-Marinoan cap carbonates [45], an alternative interpretation is that the depositional age is younger than ~635 Ma [47].

2.2. Lower Brigham Group (Stage 3—Failed Rift). The Brigham Group is a regionally extensive siliciclastic sequence of fluvial, shallow marine, and offshore marine strata that overlie the Kelly Canyon Fm in Utah and the Pocatello Fm in Idaho [10, 15, 42]. At the base of the Brigham Group is the Caddy Canyon Fm, <760 m of quartzite and siltstone deposited in a nearshore marine, braided delta, or braided fluvial system. Unconformably, overlying the Caddy Canyon Fm is the Inkom Fm (<150 m), which typically comprises siltstone and shale deposited in a low-energy marine environment [10, 18, 42, 48]. In places, the base of the Inkom Fm contains 60 m deep paleovalleys incised into the Caddy Canyon Fm that are filled locally with pebble conglomerate. These paleovalleys are enigmatic and were previously interpreted as a regional unconformity and sequence boundary related to sea level fall and erosion during Marinoan glaciation [18]. The final unit representing Stage 3 in this area is the overlying Mutual Fm, >360 m of regionally extensive pink to red fluvial and shallow marine quartzite [10, 16, 18, 42, 48].

2.3. Upper Brigham Group (Stage 4—Late Rifting). In northern Utah, Stage 4 begins with the Browns Hole Fm, which locally overlies the Mutual Fm and is divided into two informal members: the lower volcanic member and the upper terra cotta quartzite member. The only known exposure of volcanic member is in the Huntsville sequence and it is composed of <140 m of dark red to purple volcanoclastic sedimentary rocks and mafic-volcanic rocks [19]. The overlying terra cotta quartzite member is <45 m of “terra cotta-colored,” hematite-cemented quartz arenite, interpreted as fluvial, lacustrine, or eolian facies [49]. The Browns Hole Fm is not present in southeast Idaho (Pocatello area) and stratigraphically equivalent rocks may be expressed as different facies north of the Huntsville sequence (e.g., High Creek, Utah).

Within the Huntsville sequence, the Browns Hole Fm is overlain by the Geertsen Canyon Fm, ~550 m of shallow marine to fluvial, conglomeratic sandstones [16]. Here, the lower contact is covered but is identified by the appearance of less mature and coarser feldspathic sandstones compared to the underlying terra cotta quartzite member. The upper member of Geertsen Canyon Fm, ~480 m above the contact with the Browns Hole Fm, likely contains the Precambrian-Cambrian boundary. Upwards in the section, this unit contains Skolithos, occasional trilobite tracks, abundant fusoidal structures, and sparse phosphatic fragments, indicating that this part of the section could be as young as the Middle Cambrian [42, 50, 51]. In Idaho, the base of the Camelback Mountain Fm, stratigraphically correlative with Geertsen Canyon Fm, exhibits ~60 m deep paleovalleys cut into the Mutual Fm, and this major unconformity is interpreted as a regional sequence boundary [10, 18, 49].

3. Multiscale Characterization of the Browns Hole Formation

We describe the macro-to-microscale characteristics of the lower volcanic and the upper terra cotta quartzite members near Huntsville, UT (Figure 1), using field and petrographic observations. The contact between the Mutual Fm quartzites and the overlying volcanic member of the Browns Hole is gradational and interpreted as the stratigraphically lowest occurrence of volcanoclastic materials, locally marked by a distinctive, laterally discontinuous light purple-grey siltstone.
band (Figure 3, “lithic tuff”). Above the contact, the volcanic member of the Browns Hole Fm consists of ~83 m of interbedded dark purple to red, hematite-cemented volcaniclastic siltstone, sandstone, and agglomerate that collectively coarsen upward and include basalt flows. Fine-grained beds contain locally convolute laminations, indicating soft-sediment deformation. Sandstone beds also contain long, flat intraclasts of finer-grained material aligned approximately parallel to bedding. Agglomerates are composed of rounded to subrounded clasts of primarily vesicular basalt and porphyritic trachyandesite. Pebble- to cobble-sized clasts are supported by a matrix composed of coarse-sand-sized quartz grains with very irregular and pitted edges that are rimmed with fibrous clay minerals (Fig. S1a). Quartz grains are supported in a matrix composed of coarse-sand-sized quartz grains and volcaniclastic grains of the same composition as the larger volcanic clasts (Figure 3).

Petrographically, the distinctive thin siltstone band at the base of the Browns Hole Fm is composed of angular quartz grains with very irregular and pitted edges that are rimmed with fibrous clay minerals (Fig. S1a). Quartz grains are supported in a matrix composed of coarse-sand-sized quartz grains and volcaniclastic grains of the same composition as the larger volcanic clasts (Figure 3).

Volcanic clasts are a dominant component of much of the volcanic member and range from fine-grained sand to cobbles (Fig. S1d). In the thin section, these clasts, which are generally porphyritic andesite, contain fine-grained sericite and clay minerals that have replaced grains interpreted as feldspars (Fig. S1e). Few preserved feldspar grains are found in the matrix outside of volcanic clasts and, where present, are degraded with fine-grained sericite and clays in pore spaces (Fig. S1f). Minor accessory minerals include titanite and barite.

The top of this unit is marked by a poorly exposed and laterally discontinuous basalt flow. Based on the last up-section appearance of basalt clasts, the flow is inferred to be ~10 m thick (Figure 3). Compositionally, this flow classifies as a basanite, a silica-undersaturated basalt [19, 52], but for consistency with prior literature, we will use the term basalt. The base of the basalt, where exposed, is in contact with a coarse-grained, volcaniclastic sandstone with minor siltstone. The contact is also characterized by irregular, 0.5–1.0 cm thick quartz and zeolite (likely analcime) veins lined with specular hematite. These veins interfinger with the basalt and siltstone along the basal contact. The basalt horizon exhibits vesicular zones, <0.3 m in diameter, with sub-cm-scale vesicles filled with quartz, plagioclase, and specular hematite. The basalt surrounding these vesicular zones is massive with an aphantic groundmass dominantly composed of calcium plagioclase, observed petrographically and via SEM-EDS, and lesser pyroxene and irregular patches of clay and hematite. Locally, the basalt is cut by mm thick epidote veins and epidote-coated minor slip surfaces.

The contact between the basalt and overlying terra cotta quartzite is covered throughout the study area. Based on the last occurrence of basalt clasts and the estimated upper contact with the overlying Geersten Canyon Fm, the terra cotta quartzite member is inferred to be ~50–60 m thick. This unit is a medium- to coarse-grained quartzite with silica and hematite cement. Locally, sedimentary features include faint crossbedding and symmetric ripples (Figure 3).

4. Apatite U-Pb Geochronology

4.1. Sample and Methods. To improve constraints on the depositional age of the volcanic member of the Browns Hole Fm, we sampled a medium- to coarse-grained volcaniclastic sandstone collected ~10 m above the contact between the Mutual Fm and the volcanic member of the Browns Hole Fm (sample AP17_66MF-9; 41.32746°N, 111.73554°W). This sample contains volcanic clasts ranging in size from coarse sand to pebble. U-, Th-, and Pb-bearing accessory phases were recovered using standard crushing, sieving, density, and magnetic separation techniques, taking care to avoid clasts. The sample yielded too few zircon grains for U-Pb analysis. However, it contained abundant high-quality apatite grains that we targeted for U-Pb dating after determining that these were detrital in origin (see Section 5.1). We note that this sample does not include apatite cement in the matrix as observed lower in the section.

Representative polished chips of the sample were examined with the FEI Quanta FEG 650 scanning electron microscope at the Utah State University Microscopy Core Facility to determine the lithologic context of the apatite grains. An Oxford EDS detector was used to make composite element maps (P and Ca) to aid in the location of apatite grains. Spot EDS analysis and SEM back-scattered electron (BSE) images were used to identify chemical differences indicating overgrowths or chemical alteration rims on apatite crystals.

Apatite U-Pb analyses were conducted at the University of Arizona LaserChron Center. Approximately 250 grains were poured and arranged on double-sided adhesive tape to avoid bias that may be introduced during individual grain selection. Grains were mounted in epoxy and polished to their approximate midsection and cleaned using similar methods to zircon preparation [53]. A subset (n = 100) of apatite grains were analyzed for U and Pb using a Nu Plasma high-resolution multicollector LA-ICP-MS. Laser ablation utilized a Photon Machines Analyte G2 excimer laser with a spot diameter of 30 μm for 15 seconds with 8 mJ output energy and pulse frequency of 7 Hz, creating an ablation pit approximately 15–20 μm deep. Apatite standards used during analysis were the McClure Mountain (523.5 Ma) and Madagascar (486.6 Ma)apatite [54]. Apatite U-Pb data were corrected for common Pb using an iterative 207Pb common Pb correction after the Stacey and Kramers [55] common Pb model. Final data reduction and calculations for Tera-Wasserburg concordia were completed using Isoplot [56] and RadialPlotter software [57]. Errors are reported at 2σ.

Following U-Pb analyses, CL imaging of analyzed apatite grains was conducted using a Gatan ChromaCL2 detector on a Hitachi 3400N SEM at the University of Arizona.
LaserChron Center. Apatite rarely shows zonation in CL or BSE images, so these methods were not used prior to U-Pb analyses for choosing spot locations. Also, apatite U-Pb dates record cooling below 500°C [58], so even if some magmatic zoning is present, all zones should record the same cooling age. In this study, the CL images provided information on authigenic mineral overgrowths or the presence of mineral inclusions that may have interfered with spot analyses rendering those dates unreliable.

4.2. Results. Imaging and chemical mapping (Ca, P) using SEM and SEM-EDS revealed that apatite grains are found only in the matrix and are not present within volcanic lithics (Figure 4). Apatite crystals are euhedral to subhedral with only minimal rounding of tips. Note the pristine tip of the lower right grain, indicated with a red arrow. (b) Apatite grain with a faint irregular rim, outlined in orange, that is characterized by a heavily pitted texture. Observe the infilling of some cracks by light grey to white minerals (containing elements with higher atomic numbers), and pits with dark grey minerals (containing elements that have lower atomic numbers). (c) Pitted texture that is common in many analyzed in situ grains. This texture is different from the pitted texture in (b) in that it is not coupled with a change in material density. (d) A SEM CL image of an apatite grain showing oscillatory zoning that is common in grains of igneous origin. (e) An example of a large central inclusion found in several apatite grains in this sample. (f) Optical, cross-polar image of apatite separates. Note the smaller grains are largely intact, whereas larger grains tend to be missing at least one tip. Also note the center right grain, indicated with a red arrow, containing the large central inclusion, similar to what is seen in (e).

Figure 4: SEM BSE (a–c, e), SEM CL (d), and optical (f) images of apatite grains. (a) Example of in situ apatite grains; yellow circles indicate locations of grains. Note that all apatite grains shown in this image are intact except for some cracking and are euhedral to subhedral with only minimal rounding of tips. Note the pristine tip of the lower right grain, indicated with a red arrow. (b) Apatite grain with a faint irregular rim, outlined in orange, that is characterized by a heavily pitted texture. Observe the infilling of some cracks by light grey to white minerals (containing elements with higher atomic numbers), and pits with dark grey minerals (containing elements that have lower atomic numbers). (c) Pitted texture that is common in many analyzed in situ grains. This texture is different from the pitted texture in (b) in that it is not coupled with a change in material density. (d) A SEM CL image of an apatite grain showing oscillatory zoning that is common in grains of igneous origin. (e) An example of a large central inclusion found in several apatite grains in this sample. (f) Optical, cross-polar image of apatite separates. Note the smaller grains are largely intact, whereas larger grains tend to be missing at least one tip. Also note the center right grain, indicated with a red arrow, containing the large central inclusion, similar to what is seen in (e).
<50 μm long, are fully intact and unfractured (Figure 4(a)). BSE imaging and EDS analysis show that almost all apatite grains lack rims or overgrowths. Only one observed grain exhibits a faint thin irregular rim in BSE with no discernable chemical differences using EDS (Figure 4(b)). Most in situ grains show a pitted texture along grain boundaries that extends <5 μm into the grain (Figure 4(c)).

CL images of separated apatite crystals obtained after U-Pb spot analysis support SEM observations from in-situ grains (Figure 4(d)). Smaller apatite crystals have euhedral intact, faceted, or slightly rounded tips; larger grains generally have at least one tip missing. Based on SEM observations described above, the missing and rounded tips are the result of mineral separation techniques rather than transport induced. CL images show concentric zoning with respect to trace element concentration [59], which is common in magmatic grains [60]. Grains do not show any evidence of overgrowths in CL. Several grains have micron-scale mineral inclusions (Figures 4(e) and 4(f)). Based on these morphologies and textures, we suggest apatite crystals are proximally sourced detrital grains that have experienced minimal transport and postburial physical or chemical alteration.

A total of 100 grains were analyzed using LA-ICP-MS. Of these, 95 yielded reasonable 207Pb common Pb-corrected dates and errors. These 95 dates range from 458 ± 126 Ma to 1140 ± 205 Ma (2σ), with a weighted mean of 613 ± 13 Ma (2σ without outlier rejection; Table S1). Individual analyses and corresponding error ellipses uncorrected for common Pb (excluding two detrital grain dates > 1 Ga) define a discordia line on a Tera-Wasserburg diagram (Figure 5(a)). The discordia line is anchored at the upper intercept using a 207Pb/206Pb common Pb value for 610 Ma of 0.878 from the Stacey and Kramers [55] model and is linearly regressed through the datapoints. The lower intercept of the discordia line yields a date of 613 ± 10 Ma (2σ) (Figure 5(a)), which is the age of the dominant population [61]. Nine grains fall off and to the right of discordia, suggesting they experienced Pb loss (Figure 5(a)).

Owing to low U concentration (~1 ppm), individual common-Pb-corrected apatite grain U-Pb analyses have high uncertainties. Thus, to assess the different age populations and minimum age peak with greater precision, we used the program RadialPlotter [57]. Analysis of the 95 individual dates reveals a maximum of four age peaks of 524 ± 50 Ma, 621 ± 12 Ma, 770 ± 74 Ma, and 1137 ± 204 Ma (2σ; Fig. S2a). The youngest peak is controlled by two grains with moderate precision and high amounts of radiogenic Pb. One of these grains (grain 67; Table S1) has downhole laser ablation data that is progressively older towards the rim of the grain. This is potentially an artifact of downhole fractionation. However, this grain is also one of nine that likely experienced Pb loss and was removed from the dataset for these reasons. Eliminating grain 67 deconvolves the data into three peaks: 617 ± 11 Ma, 767 ± 76 Ma, and 1137 ± 204 Ma (2σ; Fig. S2b). The new youngest peak now corresponds to ~94% of the data within 2σ. The second young grain was not removed from

Figure 5: Apatite U-Pb results. (a) Tera-Wasserburg diagram anchored through common 207Pb/206Pb value for 610 Ma of 0.878 [55]. The black line is anchored through the 207Pb/206Pb value for 610 Ma and regressed through the data with the lower intercept indicating when the time when the apatite crystal became a closed system with respect to the decay of U and retention of Pb. Ellipses shaded in red, purple, and blue are grains with potential Pb loss; red is more likely to have experienced Pb-loss, purple is less likely, and blue are the least likely grains to have experienced Pb-loss. (b) RadialPlotter “abanico” diagram showing the results of the minimum peak algorithm for 207Pb common Pb-corrected U-Pb ages and associated kernel density estimate (KDE) age spectrum. Color scale depicts relative amount of radiogenic Pb. Precision (x-axis) is determined by dividing the 1 − σ error of each grain by their calculated date (t). More precise ages plot to the right, whereas less precise ages will plot to the left.
the dataset because it has a higher uncertainty that overlaps with the new youngest peak of 617 ± 11 Ma (2σ). Using the minimum peak algorithm in RadialPlotter [57], which is based on the Galbraith algorithm that assesses the youngest age that a dataset can realistically have [62], yields a minimum age of 613.5 ± 11.8 Ma (2σ; Figure 5(b)), which we round to 613 ± 12 Ma and report as our maximum depositional age. The ~767 Ma peak is based on five grains (55, 77, 88, 47, and 100; Table S1) and is strongly influenced by one grain with relatively high uncertainty. These are likely grains from an older igneous source. Some of these grains show evidence of significant transport-induced rounding and others do not (Fig. S3). The oldest peak (~1137 Ma) is based on two grains that show significant transport induced rounding in CL imaging (Fig. S3), and these detrital grains likely originated from a much older igneous source.

5. Discussion

5.1. Revised Depositional Age. Based on our new detrital apatite U-Pb geochronology, the revised maximum depositional age of the volcanic member of the Browns Hole Fm is 613 ± 12 Ma (2σ). Although this is a maximum depositional age, we argue that it may be close to the actual depositional age for several reasons. First, the apatite grains associated with this date are euhedral to subhedral and do not show evidence of significant reworking, rounding, or disaggregation during transport. Apatite grains are nearly intact and any observed grain-fracturing is in situ and due to compaction. This is consistent with the intact and faceted grains that are missing tips in the mineral separate. Second, apatite crystals are found solely within the matrix and not within volcanic lithics (Figure 4), indicating that they were not derived from well lithified materials supplying lithics to the unit. Also, the apatite grains are not associated with alteration textures in the sample matrix (e.g., [63]) and lack evidence for an authigenic origin. Collectively, these textural observations suggest that the apatite grains are detrital in origin with a proximal and likely un lithified source. For the depositional age to be significantly younger (i.e., outside of uncertainty) than ~613 Ma requires that the apatite grains be sourced from a nearby and very easily weathered material that was preserved on the landscape for >10 Myr and then moved with low transport energy to preserve the pristine character of the matrix-hosted grains. Potential older source materials that fit this character are not preserved in the Huntsville exposures of the Brown Hole Fm or the underlying Mutual Fm, which is composed of 100’s meters of highly lithified and mature siliciclastic rocks and an unlikely source of pristine detrital apatite grains.

Our new 613 ± 12 Ma (2σ) date overlaps within error with the previously reported U-Pb apatite date of 609 ± 25 Ma (2σ) from the overlying basalt flow (Verdel, 2009) within the volcanic member of the Browns Hole Fm. This basalt date is based on the weighted mean of two multi-grain ID-TIMS analyses (n = 5; n = 8), one of which was slightly discordant, resulting in the relatively large error. Overlapping errors aside, the mean dates for these two samples are different and consistent with the stratigraphic position of the two units (i.e., the detrital grains are located ~70 m down section from the basalt unit). The general consistency of these dates, and volcanogenic origin of the sediments, suggests that a similar period of volcanic activity may have been responsible for the detrital apatite grains and basalt flow.

The earlier reported maximum depositional age for this interval of 580 ± 14 Ma (2σ) was obtained using total ⁴⁰Ar/³⁹Ar analysis of hornblende hosted in a detrital volcanic clast [12, 13]. This analytical method assumes a closed system with respect to Ar, and any Ar loss due to postdepositional reheating or diagenesis yields anomalously young dates [64]. Modern step-heating methods are necessary to identify potential Ar loss and yield a reliable plateau age that is representative of mineral formation or cooling through its closure temperature [65, 66]. For these reasons, we cannot interpret this ⁴⁰Ar/³⁹Ar date in the context of the deposition of the Browns Hole Fm.

In Sections 5.2 and 5.3, we explore the broader implications of the revised maximum depositional age for the Browns Hole Fm. We acknowledge that these interpretations are based largely on the data from one sample, and future studies are encouraged to corroborate our results.

5.2. Stratigraphic Implications. The new maximum depositional age of 613 ± 12 Ma for the volcanic member of the Browns Hole Fm has several implications for the timing and duration of deposition of regional stratigraphy. Importantly, we interpret this as close to the actual depositional age, and it corroborates, with improved resolution, the 609 ± 25 Ma basalt age, thus providing a critical benchmark in the Ediacaran stratigraphy. Our new date decreases the amount of time available for the deposition of ~1400 m of Browns Hole Fm and lower Brigham Group by ~33 Myr (Figure 6(a)). If the Caddy Canyon Fm was deposited ca. 650 Ma [10], and the paleovalleys at the base of the Inkom Fm were formed at the end of Marinoan glaciation [18]; then, ~1400 m of strata was deposited over ~37 Myr, from ca. 650 Ma to 613 Ma, or ~38 mm/kyr (Figure 6(a)). However, it is likely that the pink dolostones overlying glacial diamictites of the Pocatello and Perry Canyon formations relate to Marinoan glaciation [45, 67] and the Caddy Canyon Fm was deposited closer to ca. 635 Ma, and thus, the ~1400 m of sediment was deposited in ~22 Myr from ca. 635 Ma to 613 Ma, or ~64 mm/kyr (Figure 6(b)). We acknowledge that these estimates are based on the preserved thicknesses of sediments and do not account for postdepositional compaction or deformation. Also, these are average sedimentation rates, which do not consider periods of rapid sedimentation or nondeposition.

Our revision to the Browns Hole Fm maximum depositional age (from ~580 to 613 Ma) also impacts regional correlations and interpretations of unconformities with overlying stratigraphy. Specifically, the unconformity between the volcanic member and overlying terra cotta quartzite member of the Browns Hole Fm may represent more time than previously recognized. Likewise, the older depositional age of the Browns Hole Fm may also increase
This scenario requires that the paleovalleys at the base end of Marinoan glaciation in western Laurentia [45, 47, 49].

In this scenario, the cause of the paleovalleys in the Inkom is presently unknown. The Inkom Fm, previously interpreted as a signature of global glaciation. Prior study correlates these Inkom Fm paleovalleys to sea level fluctuations during and following Marinoan glaciation [10, 18]. However, other work correlates exposures of the upper diamictites and local pink dolostones of the Pocatello and Perry Canyon formations that are located stratigraphically well below the Inkom Fm with the end of Marinoan glaciation in western Laurentia [45, 47, 67]. This scenario requires that the paleovalleys at the base of the Inkom Fm are of a different origin and possibly related to another younger glacial advance and associated sea level drop. Our new geochronologic data precludes presently known glacial events, such as the Gaskiers glaciation at ca. 580 Ma [68], as the cause of these features. Thus, Inkom Fm paleovalleys may reflect a currently unknown glacial advance or some other tectonic or geomorphic control, but additional work is needed to evaluate the timing and cause of valley formation.

Our new data also have implications for the timing and expression of rifting in western Laurentia. Previously reported geochronology and paleomagnetic data indicate that continental land masses attached to the western region of Laurentia began rifting after ~720 Ma and were completely separated sometime after 600 Ma [4, 10]. The appearance of proximally sourced volcanic material and a

| Neoproterozoic carbonate | Cap carbonate | Volcanic rocks | Glacial Diamictite |
|-------------------------|--------------|----------------|-------------------|

Figure 6: Two possible Wheeler diagrams modified and revised from Balgord et al. [40] and Yonkee et al. [10] with relevant representative regional stratigraphic sections. The major Cryogenian glaciations, Sturtian and Marinoan, are indicated by a green shaded bar spanning both diagrams. As the contact between the terra cotta quartzite and volcanic members of the Browns Hole Formation is covered in the study area, we label the contact as a “potential unconformity.” Important geochronology discussed in the text is denoted in red: (a) This research, (b) Crittenden and Wallace [13] and Christie-Blick and Levy [49], and (c) Verdel [19]. (a) Revised stratigraphy if the paleovalleys at the base of the Inkom Formation correspond to sea level fluctuations related to Marinoan glaciation. (b) Revised stratigraphy if upper carbonates and diamictites in the Perry Canyon and Pocatello formations are related to Marinoan glaciation. In this scenario, the cause of the paleovalleys in the Inkom is presently unknown.

5.3. Implications for Rodinian Paleoenvironment and Rifting. Revising the depositional age of the Browns Hole Fm impacts interpretation of paleovalleys at the base of the Inkom Fm, previously interpreted as a signature of global glaciation. Prior study correlates these Inkom Fm paleovalleys to sea-level fluctuations during and following Marinoan glaciation [10, 18, 37–40]. However, other work correlates exposures of the upper diamictites and local pink dolostones of the Pocatello and Perry Canyon formations that are located stratigraphically well below the Inkom Fm with the end of Marinoan glaciation in western Laurentia [45, 47, 67]. This scenario requires that the paleovalleys at the base of the Inkom Fm are of a different origin and possibly related to another younger glacial advance and associated sea level drop. Our new geochronologic data precludes presently known glacial events, such as the Gaskiers glaciation at ca. 580 Ma [68], as the cause of these features. Thus, Inkom Fm paleovalleys may reflect a currently unknown glacial advance or some other tectonic or geomorphic control, but additional work is needed to evaluate the timing and cause of valley formation.
basalt flow within the Browns Hole Fm implies that mafic magmatism, possibility associated with rifting, was active ca. 613 Ma. Also, the basanitic composition of the basalt flow capping the volcanic member of the Browns Hole Fm [19] is consistent with this interpretation. Basanite, an alkali basalt, is common in continental rift and continental and oceanic mantle plume settings [69]. For example, basanite is frequently found in locations along the East African Rift, such as the Turkana Rift volcanoes [70, 71]. Additional regional evidence of mafic magmatism at this time includes a 601 ± 27 Ma (2σ) gabbroic sill (Ramshorn gabbro) intruded into the lower Clayton Mine Quartzite (<667 Ma) in central Idaho [72]. Browns Hole Fm data overlap within analytical error of this sill date, suggesting that rift-related (and possibly mantle plume) mafic magmatism occurred regionally during this time. Evidence of rift-related volcanism at ca. 569 Ma is also preserved in the Hamill Group of the southeastern Canadian Cordillera [73]. Collectively, these data may represent some of the last episodes rift-related volcanism before the transition to a passive margin.

6. Conclusions

Our work centered on the Browns Hole Fm reveals that detrital apatite U-Pb geochronology is a useful alternative or supplement to detrital zircon U-Pb geochronology to determine the maximum depositional age of a sedimentary rock. It is particularly valuable in cases where a statistically significant population of zircon grains cannot be obtained, zircons are not present at all, or if challenges in isolating young populations emerge. In our example, detrital apatite textural characteristics and U-Pb data from the Browns Hole Fm are used to refine the chronology of regional Neoproterozoic stratigraphy and inform timing of rifting of western Laurentia. This demonstrates the importance of incorporating morphology as well as the in-situ context of detrital apatite grains when interpreting U-Pb dates. Our data also bear on the significance of deep paleovalleys in the Inkom Fm, previously linked to the Marinoan glaciation. The revised Browns Hole Fm maximum depositional age precludes this interpretation, as well as linkages to other younger glaciations such as the Gaskiers and indicates that more work is needed to evaluate the timing and paleoclimatic significance of these difficult to date Neoproterozoic sedimentary sequences and geomorphologic features.

Data Availability

All data used in this manuscript are provided in the main text and supplemental materials.

Conflicts of Interest

There are no conflicts of interest to disclose related to this research.

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Supplementary Materials

In support of this manuscript, supplemental materials are provided that include three figures (Figures S1–S3) and one data table (Table S1). The data table provides the U-Pb analytical data and calculated dates for the analyzed apatite grains. (Supplementary Materials)

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