Sedimentology, U-Pb detrital geochronology, and Hf isotopic analyses from Mississippian-Permian stratigraphy of the Mystic subterrane, Farewell terrane, Alaska

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ABSTRACT

The Farewell terrane of western Alaska is one of the more remote and understudied crustal fragments in the North American Cordillera. Although it is generally accepted that the oldest, Precambrian parts of the Farewell terrane originated along the Arctic margin (i.e., Siberia), the paleogeographic history of the Farewell terrane during much of the middle and late Paleozoic remains unknown. Here, we present new sedimentologic and provenance data from upper Paleozoic clastic strata of the Mystic subterrane, which represents the youngest part of the Farewell terrane. Sedimentary facies consist of high- and low-density sediment-gravity-flow deposits and are interpreted to represent a submarine fan depositional system. Sandstone modal composition trends show a relative abundance of lithic volcanic fragments (~65%) and subordinate occurrences of lithic sedimentary fragments (~15%) and chert (~13%). Laser-ablation–inductively coupled plasma–mass spectrometry analyses of detrital zircons reveal a bulk U-Pb age distribution of Precambrian–Paleozoic grains. U-Pb detrital zircon age spectra from Mississippian strata have a primary peak age between 400 and 325 Ma and secondary peak ages between 480 and 415 Ma and 2000 and 1800 Ma. Devonian–Mississippian zircons exhibit enriched εNd isotopic values (~3 to ~35), whereas Ordovician–Silurian zircons have both enriched (~5 to ~25) and depleted (+5 to +14) εNd values. Age spectra from Permian strata show primary peaks between 320 and 275 Ma and 460 and 415 Ma, with isolated occurrences of Precambrian-age zircons. Pennsylvanian–Permian zircons exhibit depleted εNd values (+2 to +14). Youngest peak ages support a Mississippian–Early Permian maximum depositional age for this part of the Mystic subterrane. Overall, provenance trends reflect primary detrital contributions from arc and recycled orogen source areas, which included both enriched and primitive magmatic sources. New U-Pb and Hf isotope analyses from the Mystic assemblage match most closely with magmatic source areas of the Alexander and Wrangellia terranes. Findings are consistent with a model where the Farewell terrane was proximal to both the Alexander and Wrangellia terranes by Mississippian–Permian time.

INTRODUCTION

The Farewell terrane is a regionally extensive (~40,000 km²), Paleoproterozoic–Jurassic crustal fragment (Decker et al., 1994; Bundtzen et al., 1997; Bradley et al., 2003) that is located in the northernmost part of the North American Cordillera in western Alaska (Fig. 1A). The majority of the terrane crops out along the southern boundary of the Denali fault and northern boundary of the Iditarod–Nixon Fork fault in western Alaska (Fig. 1B). As defined by previous regional studies (Decker et al., 1994; Bundtzen et al., 1997), the Farewell terrane is generally divided into a three-part geologic suite that includes (1) latest Neoproterozoic (?)–Lower Devonian carbonate-dominated strata of the Nixon Fork subterrane, (2) Cambrian–Lower Devonian mixed carbonate and clastic strata of the Dillinger subterrane, and (3) Devonian–Jurassic(? clastic-dominated strata and subordinate volcanic rocks of the Mystic subterrane (Figs. 1B and 1C). Parts of the Dillinger subterrane that outcrop along the Iditarod–Nixon Fork fault were originally referred to as the Minchumina terrane (Jones et al., 1987; Silberling et al., 1994; Patton et al., 1994). The oldest basement rocks in the Farewell terrane are a Paleoproterozoic–Neoproterozoic(? metamorphic complex (Patton and Dutro, 1979; Bradley et al., 2003), which makes up the northernmost exposure of the terrane, just north of the Iditarod–Nixon Fork fault (Fig. 1B).

One of the challenges with beginning to understand the tectonic evolution of accretionary plate margins, such as the North American Cordillera, is deciphering the origin and subsequent paleogeographic history of its constituent terranes prior to accretion. Many of the Paleozoic–Mesozoic terranes that make up the northern North American Cordillera have origins in circum-Arctic regions (e.g., Siberia, Batica), intraoceanic settings in the Panthalassic Ocean, or represent displaced peri-Laurentian crustal fragments that developed in proximity to the western margin of ancestral North America (e.g., Nakleberg et al., 2000; Colpron et al., 2007; Colpron and Nelson, 2009; Miller et al., 2011). The Farewell terrane was initially thought to have originated along the western margin of Laurentia (Coney et al., 1980; Box, 1985; Decker et al., 1994; Plafker and Berg, 1994); however, more recent paleontological studies have made the case for an Arctic-margin origin, which could include localities such as Siberia or northern Laurentia (Soja and Antoshkina, 1997; Blodgett et al., 2002; Dumoulin et al., 2002).

Beyond a proposed Arctic-margin origin, very little is known about the Paleozoic tectonic evolution and paleogeographic history of the Farewell terrane. Occurrences of Siberian floral and faunal assemblages from lower Paleozoic faunas in the Farewell terrane. Occurrences of Siberian floral and faunal assemblages from lower Paleozoic faunas in the Farewell terrane.
strata of the Nixon Fork and Dillinger subterrane have led some to propose that the Farewell terrane originated as a displaced (rifted) fragment of the Siberian platform (Soja and Antoshkina, 1997; Blodgett et al., 2002). Dumoulin et al. (2002) reported faunal occurrences that suggest a paleogeographic link with the Arctic Alaska-Chukotka microplate during the early Paleozoic. Based on previous workers’ suggestion of a Siberian affinity as well as Uralian-aged metamorphism (Bradley et al., 2003), Colpron and Nelson (2009) suggested that the Farewell terrane originated in the “Northwest Passage” (i.e., late Paleozoic Uralian seaway) and remained marginal to Siberia until the Permian, when it was transported to the northwestern Panthalassic Ocean and eventually to the western margin of Laurentia and northern Cordillera.

Bradley et al. (2007) reported Precambrian U-Pb detrital zircon ages from isolated lowermost Paleozoic clastic units of the Nixon Fork terrane, which support detrital contributions from Siberian cratonic sources. Permian 40Ar/39Ar plateau ages (284–285 Ma) from Paleoproterozoic–Neoproterozoic (?) basement metamorphic rock of the Farewell terrane have been interpreted to record a late Paleozoic orogenic event (referred to as the “Browns Fork orogen”) coeval with the Pennsylvanian–Permian Uralian orogeny (Bradley et al., 2003). Thick successions of Pennsylvanian–Permian clastic strata in the upper part of the Farewell terrane (i.e., Mystic subterrane) are thought to represent part of the foreland basin associated with the Browns Fork orogen (Bradley et al., 2003). In this study, we present new sedimentologic and provenance data from Mississippian–Early Permian strata of the Mystic subterrane (Fig. 1) that provide initial constraints on the late Paleozoic depositional setting and paleogeographic history of the Farewell terrane. Results from this investigation suggest a revised paleogeographic model wherein the Mystic subterrane was in proximity to the Alexander and Wrangellia terranes during the late Paleozoic.
GEOLOGIC AND STRATIGRAPHIC OVERVIEW

With the exception of regional mapping projects and isolated paleontologic and geochronologic studies (e.g., Reed and Nelson, 1980; Bundtzen et al., 1997; Bradley et al., 2003, 2007; Sunderlin, 2008), very little work has been carried out that focuses on the detailed sedimentology, stratigraphy, and provenance of siliciclastic units from the upper part of the Farewell terrane (Mystic subterrane). Upper Paleozoic rocks of the Mystic subterrane include a Mississippian–Pennsylvanian succession of interbedded sandstone, siltstone, mudstone, and isolated conglomerate (Fig. 1C). These strata have been assigned the map unit “Pzus” from the Talkeetna quadrangle (Reed and Nelson, 1980) and “PDs” from the McGrath quadrangle (Bundtzen et al., 1997). In the northwestern Talkeetna quadrangle, “Pzus” strata are overlain by an additional coarse-grained siliciclastic unit referred to as the Mount Dall conglomerate (Reed and Nelson, 1980). The Mount Dall conglomerate has been interpreted to represent fluvial-deltaic sedimentation (Sunderlin, 2008) with isolated fine-grained intervals containing Permian plant fossils that have a mixed Siberia-Laurentia affinity (Mamay and Reed, 1984; Sunderlin, 2008). Bradley et al. (2003) suggested that the Mount Dall conglomerate represents the foreland basin (referred to as the Dall basin) that was associated with the Pennsylvanian–Permian Browns Fork orogeny.

The most extensive tracts of Mississippian–Permian strata of the Mystic subterrane lie south of the Denali fault in the northwestern Talkeetna quadrangle (Mystic Pass region) and in the east-central part of the McGrath quadrangle (Farewell Lake–Sheep Creek area; Figs. 1B and 2). Isolated occurrences of age-equivalent rocks in the Farewell terrane also crop out in the White Mountains region (southwestern McGrath quadrangle), and in the central part of the Lime Hills quadrangle (Fig. 1B). Although strata in these localities are thought to roughly overlap in age with Mississippian–Permian units in the Mystic Pass and Farewell Lake–Sheep Creek localities, exposures are sparse and very limited. This study focuses on Mississippian–Permian stratigraphy exposed in the northwestern Talkeetna quadrangle (Mystic Pass region) and east-central McGrath quadrangle (Farewell Lake–Sheep Creek area; Fig. 2).

At Shellabarger Pass, the base of the Mystic subterrane is thought to be marked by either a late Early Devonian (Emsian) limestone unit (Gilbert and Bundtzen, 1984; Blodgett and Gilbert, 1992; Blodgett et al., 2002) or a thin (~10 m) nonmarine red-bed sequence consisting of sandstone and conglomerate with coal and fossilized wood and plant debris (Reed and Nelson, 1980). Basal units are overlain by a 125–250-m-thick succession of sandstone, siltstone, and shale that contains late Middle and early Late Devonian fossils (Reed and Nelson, 1980). Up section, there is a 60–90-m-thick reefoid limestone that is overlain by a phosphatic chert unit (informally referred to as the “blackball chert”), which contains radiolarians of Late Devonian age (Famennian; Reed and Nelson, 1980). Pillow basalts are interbedded within some of these strata and have been inferred to be middle or late Paleozoic in age (Reed and Nelson, 1980). Devonian rocks of the Mystic subterrane are overlain by Mississippian–Permian siliciclastic strata (this study) and capped by Permian coarse clastic strata of the Mount Dall conglomerate (Fig. 1C).
The contact between the base of the Mystic subterrane and top of the Dillinger subterrane has been reported as a conformable contact (Bundtzen et al., 1997). However, at the basin scale, debate remains over the fundamental nature of this contact and the stratigraphic interval where it occurs. In the western Talkeetna quadrangle, the basal part of the Mystic subterrane is thought to be locally underlain by carbonate and siliciclastic strata of the Dillinger subterrane. Decker et al. (1994) suggested a Middle Devonian angular unconformity as the contact between the Mystic and Dillinger subterrane; however, local conformable relationships have also been reported in some localities (Blodgett and Gilbert, 1983; Patton et al., 1994). The nature of this contact is not well documented and warrants further investigation.

Strata of Mississippian–Permian age in the Mystic subterrane consist primarily of interbedded sandstone and mudstone, with isolated occurrences of conglomerate. Biostratigraphic age constraints for these units include Upper Mississippian and Middle Pennsylvanian echinoderms and foraminifers (identified by A.K. Armstrong in Reed and Nelson, 1980). Overlying rocks of the Mount Dall conglomerate consist of a >1500-m-thick succession of Permian-aged sandstone, conglomerate, and fossil-leaf–bearing siltstone. A Permian age for the Mount Dall conglomerate is based on occurrences of plant fossils (Zamiopteris) and brachiopods (Mamay and Reed, 1984; Sunderlin, 2008), as well as conglomerate clasts that contain Pennsylvanian to Early Permian conodonts (Bradley et al., 2003). Boulders and cobbles of the Mount Dall conglomerate consist primarily of chert, limestone, and sandstone, all of which may have been sourced from the Dillinger subterrane and older units of the Mystic subterrane (Bradley et al., 2003). A conformable relationship has been proposed for Mississippian–Permian units and overlying strata of the Mount Dall conglomerate based on the similarity in ages between these units (Reed and Nelson, 1980). With the exception of work by Sunderlin (2008) on the Mount Dall conglomerate, there have been no studies to date that have documented detailed sedimentology, stratigraphy, provenance, and structural relationships between the Mississippian–Permian stratigraphic succession and overlying rocks of the Mount Dall conglomerate within the Farewell terrane. The focus of this study is on the Mississippian–Permian clastic strata that underlie basal Devonian strata of the Mystic subterrane and overlie Permian rocks of the Mount Dall conglomerate (Fig. 1C).

SEDIMENTOLOGY AND STRATIGRAPHY

Upper Paleozoic siliciclastic units from the northwest quadrant of the Talkeetna quadrangle (Mystic Pass region) and the Farewell Lake–Sheep Creek area in the east-central McGrath quadrangle (Fig. 2) consist of faulted and isoclinally folded successions of interbedded sandstone, siltstone, mudstone, and subordinate conglomerate (Figs. 2 and 3). We present new lithologic and sedimentologic data from these strata, which include the “Pzus” and “PDs” map units from the Talkeetna quadrangle and McGrath quadrangle, respectively (Reed and Nelson, 1980; Bundtzen et al., 1997). Collectively (and informally), we refer to these Mississippian–Permian units as the “Mystic assemblage.” Based on extensive exposure of the Mystic assemblage in the northwest part of the Talkeetna quadrangle, the Mystic Pass field area (Figs. 2B and 3A) is considered the type locality for the Mystic assemblage.

Mystic Assemblage

Lithologic Descriptions

Sandstone, siltstone, and mudstone (subordinate conglomerate). The Mystic assemblage is
characterized by interbedded gray to black, fine- to coarse-grained sandstone, siltstone, mudstone, and subordinate conglomerate (Figs. 3B, 4A, and 4B). Individual beds of fine-grained sandstone, siltstone, and mudstone are thin (5–30 cm thick) and laterally extensive (>100 m; Fig. 4B). Beds of medium- to coarse-grained sandstone range in thickness from 0.5 to 2 m thick (Fig. 4A) and can be traced laterally for >100 m. Nearly all beds are tabular and show little evidence of large-scale erosional scour; however, sole marks are common at the base of sandstone beds. Isolated occurrences of mud rip-up clasts occur near the base of sandstone beds. From base to top, sandstone beds are commonly massive (Sm) to horizontally laminated (Sh) and fine upward into ripple cross-laminated, fined-grained sandstone, massive siltstone, and mudstone (Fm; Figs. 4A and 4B). Flame structures, load structures and convolute laminations are common in sandstone and mudstone beds (Figs. 4B and 4C). Rare occurrences of nondescript bioturbation were observed near the base and top of mudstone and siltstone beds. Conglomerate units occur sporadically throughout the Mystic assemblage and are both matrix and clast supported (Fig. 4). Individual beds are lenticular to tabular, range from 1 to 4.5 m in thickness, and commonly exhibit an upward transition from massive conglomerate (Gm) into fine- to medium-grained massive sandstone (Sm). Pebble clasts are rounded to subrounded and range from 1 to 10 cm in diameter. Conglomerate units are dominated by volcanic and chert pebbles with rare occurrences of sandstone and limestone clasts.
**Interpretation of Depositional Environment**

*Submarine fan system.* The Mystic assemblage exposed in the northwest Talkeetna quadrangle (Mystic Pass) and east-central McGrath quadrangle (Farewell Lake–Sheep Creek area) is interpreted to represent deposition of low- to high-density sediment gravity flows in a submarine fan environment (e.g., Middleton and Hampton, 1973). Upward fining in individual beds of sandstone and conglomerate reflects normal grading and is interpreted to represent event-based sedimentation characterized by initial traction transport that was followed by waning energy conditions, which allowed for final suspension settling of siltstone and mudstone deposits, reflecting suspension fallout during waning flow conditions, which are interpreted to represent Td divisions of the Bouma sequence (Bouma, 1962). All Bouma sequences are interpreted to represent sedimentation of low-density turbidity currents.

We interpret the Mystic assemblage to have been deposited in a basinal (deep-water) submarine fan setting. Thinly bedded tabular sandstone and mudstone units are interpreted as unconfined, low-density flows, which are common in mid- to outer-lobe depositional environments (cf. Mutti and Ricci Lucchi, 1975; Ricci Lucchi, 1975; Mutti, 1977). Thick-bedded conglomerate units, which display tabular to lenticular bedding geometries, are interpreted to represent confined to unconfined, high-density sediment gravity flows that were deposited in more proximal parts of submarine fans (e.g., channel to channel-lobe transition zones). Although we do not rule out a shallow-shelf submarine depositional model for these strata, there is no direct evidence that these strata were directly adjacent or proximal to a shelf-margin depositional system.

**PROVENANCE**

**Compositional Data**

Petrographic and compositional data was obtained for 31 sandstone samples collected from the Mystic assemblage in the northwestern Talkeetna quadrangle (Fig. 5). Standard petrographic thin sections were cut and stained for plagioclase and potassium feldspar. Thin sections were analyzed according to the modified Gazzi-Dickinson point-counting method (Dickinson, 1970; Ingersoll et al., 1984). Modal compositions were determined by identifying 400 grains from each thin section. Point-count parameters and raw point-count data are available in the GSA Data Repository, Appendix 1A, and recalculated data are available in Table 1. Recalculated data are based on procedures defined by Ingersoll et al. (1984) and Dickinson (1985).

1GSA Data Repository Item 2014245, Appendices 1A–3B, is available at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.
The modal composition of the Mystic assemblage is characterized by predominantly lithic fragments (L), with subordinate amounts of quartz (Q) and relatively minor occurrences of feldspar (F) (Q 17%, F 2%, L 81%; Figs. 5A and 5B). The total quartz composition consists primarily of chert (C), with relatively minor amounts of monocrystalline quartz (Qm; Fig. 5C). Feldspar grains are relatively rare, with plagioclase feldspar (P) being more abundant than potassium feldspar (K) grains (Qm 62%, P 23%, K 15%; Fig. 5B). Lithic fragments consist primarily of volcanic types (Lv), with lithic sedimentary grains (Lv) making up a relatively smaller overall occurrence (Lv 81%, Lm 0%, Ls 19%; Figs. 5B and 5C). Lithic volcanic fragments are characterized almost entirely by a mafic to intermediate fine-grained volcanic groundmass (Fig. 5C). Metamorphic fragments are all but absent in these samples (statistically making up less than 1 grain per sample).

**Summary of Modal Compositional Trends**

Compositional data from the Mystic assemblage suggest that detritus was derived from a source rich in volcanic rocks. Source areas also contributed chert and sedimentary rock fragments, which are reflected as minor detrital contributions. Comparing our compositional data with the provenance fields of Dickinson et al. (1983), the majority of the Mystic assemblage plots with sandstone derived from undisected arc sources (~84% of total samples), with minor contributions from recycled orogen sources (~16% of total samples; Fig. 5A). It is worth noting that while lithic volcanic fragments make up a majority of overall framework grains (~65%) in the Mystic assemblage, feldspar is only a minor component (<3%) in these samples (Fig. 5B). This could be explained by a combination of several scenarios such as breakdown/weathering of detrital feldspar during sedimentary transport or diagenetic breakdown/dissolution of in situ feldspar after deposition. In summary, the sandstone compositional data from the Mystic assemblage indicate primary contributions from arc sources and secondary contributions from a recycled orogen source.

**U-Pb Detrital Zircon Geochronology**

**Methods**

Zircon separates were obtained from four medium-grained sandstone samples (MYS-01, MYS-03, MYS-05, and FSC-02) from the lower and upper parts of the Mystic assemblage in the western Alaska Range (for detailed locations of samples in the Talkeetna and McGrath quadrangles, see Table 2 and Figure 2). Individual zircon crystals were extracted by traditional methods of crushing, grinding, Wilfley table, heavy liquid, and a Frantz magnetic separation. A large split of these grains (typically thousands of grains) was incorporated into a 2.54 cm (1 in.) epoxy mount together with fragments of the Sri Lanka (SL) zircon standard. The mounts were sanded down to a depth of ~20 μm, polished, imaged, and cleaned prior to isotopic analysis. A detailed description of analytical methods, as well as data and concordia plots, is available in the GSA Data Repository, Appendix 2A (see footnote 1).

**U-Pb Age Distribution**

The Mystic assemblage of the western Alaska Range shows numerous occurrences of concordant Precambrian–Paleozoic zircons (Fig. 6), with Paleozoic (Pz) age grains being more abundant than Precambrian (Pc) grains.

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**TABLE 1. RECALCULATED MODAL PERCENTAGES OF SANDSTONE FROM THE MYSTIC ASSEMBLAGE**

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<td>93</td>
<td>1</td>
<td>1</td>
<td>98</td>
<td>50</td>
</tr>
<tr>
<td>01 DZ</td>
<td>11</td>
<td>2</td>
<td>88</td>
<td>6</td>
<td>2</td>
<td>92</td>
<td>77</td>
</tr>
<tr>
<td>03 DZ</td>
<td>10</td>
<td>1</td>
<td>89</td>
<td>5</td>
<td>1</td>
<td>94</td>
<td>79</td>
</tr>
<tr>
<td>05 DZ</td>
<td>19</td>
<td>3</td>
<td>78</td>
<td>7</td>
<td>3</td>
<td>90</td>
<td>69</td>
</tr>
</tbody>
</table>

**Note:** Q — quartz; F — feldspar; L — lithics; Qm — monocrystalline quartz; Lm — lithic metamorphic; Lt — lithic sedimentary.

**TABLE 2. LIST OF LOCATIONS AND DESCRIPTIONS OF DETRITAL ZIRCON SAMPLES**

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Location</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>MYS 01</td>
<td>Mystic Pass</td>
<td>62°38.495</td>
<td>152°29.788</td>
<td>Fine- to medium-grained sandstone</td>
</tr>
<tr>
<td>MYS 05</td>
<td>Mystic Pass</td>
<td>62°39.573</td>
<td>152°31.560</td>
<td>Fine- to medium-grained sandstone</td>
</tr>
<tr>
<td>MYS 03</td>
<td>Mystic Pass</td>
<td>62°38.216</td>
<td>152°29.777</td>
<td>Fine-grained sandstone</td>
</tr>
<tr>
<td>FSC 02</td>
<td>Sheep Creek</td>
<td>62°26.141</td>
<td>153°53.018</td>
<td>Fine- to medium-grained sandstone</td>
</tr>
</tbody>
</table>
The majority of Paleozoic age grains have ages between 480 and 415 Ma (peak occurrences between 425 and 450 Ma), 400 and 325 Ma (peak occurrences between 335 and 380 Ma), and 320–275 Ma (peak occurrences between 280 and 300 Ma) (Fig. 6). Precambrian peak ages are primarily Paleoproterozoic and fall between 1.8 and 2.0 Ga (14%). There are also isolated occurrences of Neoproterozoic (5%), Mesoproterozoic (7%), and Archean age grains (5%) (Fig. 6). Based on age spectra and populations of youngest grains, we have subdivided the Mystic assemblage into a lower member and an upper member.

**Lower Mystic assemblage.** Two samples from the lower part of the Mystic assemblage (MYS-01 and MYS-05) contain a range of Paleozoic- and Precambrian-age detrital zircons. A primary Paleozoic peak age for sample MYS-01 occurs at 349 Ma, while secondary Paleozoic and Precambrian peaks occur at 435 Ma and 1872 Ma (Fig. 6). The three youngest overlapping concordant grains are Mississippian (Serpukhovian) in age (324 ± 2 Ma, 325 ± 7 Ma, and 325 ± 12 Ma). Sample MYS-05 has a primary Paleozoic peak at 335 Ma, with secondary Paleozoic and Precambrian peaks at 380, 450, and 1870 Ma (Fig. 6). The three youngest overlapping concordant grains from this sample are Pennsylvanian (Bashkirian–Moscovian) in age (307 ± 7 Ma, 313 ± 3 Ma, and 314 ± 3 Ma).

**Upper Mystic assemblage.** Two samples from the upper part of the Mystic assemblage (FSC-02 and MYS-03) also contain both Paleozoic- and Precambrian-age detrital zircons. A primary Paleozoic peak age for sample FSC-02 occurs at 440 Ma, while a secondary Paleozoic peak occurs at 282 Ma (Fig. 6). The three youngest overlapping concordant grains in this sample are Early Permian (Artinskian) in age (278 ± 4 Ma, 280 ± 5 Ma, and 281 ± 6 Ma). Sample MYS-03 has a primary Paleozoic peak at 298 Ma, with a secondary Paleozoic peak at 425 Ma (Fig. 6). The three youngest overlapping concordant grains from this sample are Middle Permian (Rodian–Wordian) in age (266 ± 4 Ma, 266 ± 7 Ma, and 271 ± 6 Ma).

**Maximum depositional age.** Although age constraints are sparse for much of the Mystic subterrane, Upper Mississippian–Middle Pennsylvanian marine fossils have been used to constrain the age of rocks that make up the Mystic assemblage (Reed and Nelson, 1980). While it has been shown that the youngest concordant detrital age in a sample can be a statistically robust and valid approach to constrain maximum depositional ages (Dickinson and Gehrels, 2009), here we rely on the youngest graphical age peak as a conservative, first-order constraint on the maximum depositional age of
each sample. Based on the youngest peak ages from the Mystic assemblage, we report an early Middle Mississippian maximum depositional age for the lower part of the assemblage and an Early Permian maximum depositional age for the upper part of the assemblage.

Bradley et al. (2007) reported zircon ages from two detrital samples as well as U-Pb thermal ionization mass spectrometry (TIMS) ages from a 1-m-thick ash-fall tuff from the Mystic subterrane. One of the detrital samples was collected from map unit “Pzus” (Reed and Nelson, 1980) located near Surprise Glacier, just a few kilometers north of Mystic Pass. This sample yielded an interpreted Mississippian maximum depositional age based on the youngest concordant grain age of 335 Ma (Bradley et al., 2007). The other two samples were also collected from map unit “Pzus,” near Pingston Creek, several kilometers west of Mystic Pass. The detrital zircon sample yielded a youngest concordant grain age of 293 Ma; however, the age of the underlying tuff deposit is interpreted to be Late Triassic, at 223 Ma (Bradley et al., 2007).

**HF Isotopic Analyses of Detrital Zircon**

**Methods**

HF isotopic data were collected from three out of the four sandstone samples (MYS-03, MY5-05, and FSC-02) for which there were U-Pb age data. The HF analyses account for ~53% of zircon grains analyzed for U-Pb geochronology and represent both the lower (MYS-05) and upper (MYS-03 and FSC-02) parts of the Mystic assemblage. These analyses were conducted at the Arizona LaserChron Center (ALC) at the University of Arizona. A description of the analytical methods and the HF analytical results can be found in the GSA Data Repository, Appendix 3A (see footnote 1).

**HF Isotopic Distribution**

HF isotopic data from Paleozoic and Precambrian zircons of the Mystic assemblage show a wide range of both positive and negative εHf values (+15 to –32; Fig. 6). Approximately 65% of εHf values are reported from Paleozoic-age zircon grains, and the remaining ~35% of εHf values are from Precambrian grains. The εHf values are reported in the context of the four primary U-Pb detrital zircon age clusters from the Mystic assemblage (275–320 Ma, 325–400 Ma, 415–480 Ma, and 1800–2000 Ma).

Pennsylvania–Early Permian detrital zircons with ages that cluster between 275 and 320 Ma are dominated almost entirely by positive εHf values (+2 to +14), whereas Middle Devonian–Mississippian detrital zircons with ages that cluster between 325 and 400 Ma are characterized primarily by negative εHf values (–3 to –43; Fig. 6). Ordovician–Silurian zircons that cluster between 415 and 480 Ma show a range of both positive (+5 to +15) and negative εHf values (–3 to –25; Fig. 6). Precambrian detrital zircons, including those that fall in the 1800–2000 Ma age range, typically have εHf values that range between –10 and +10 (Fig. 6).

**Lower Mystic assemblage.** One sample from the lower part of the Mystic assemblage (MYS-05) contains Ordovician–Silurian and Devonian–Mississippian detrital zircons with negative εHf values. Ordovician–Silurian zircon εHf values range from –3 to –12, whereas Devonian–Mississippian zircon values range from –10 to –43 (Fig. 6). Pervasive occurrences of negative εHf values from Paleozoic detrital zircons suggest they were derived from enriched, continental magmatic source areas. Proterozoic- and Archean-age zircon clusters have εHf values that range from –10 to +8 and –10 to +5, respectively (Fig. 6), and suggest a combination of magmatic sources that included depleted mantle reservoirs and more enriched continental sources.

**Upper Mystic assemblage.** Two samples from the upper part of the Mystic assemblage (FSC-02 and MYS-03) contain clusters of Ordovician–Silurian detrital zircons that have both positive and negative εHf values ranging from –26 to + 15 (Fig. 6). Both samples also contain Pennsylvania–Early Permian zircons with entirely positive εHf values (+1 to +12; Fig. 6). Occurrences of positive εHf values for clusters of Pennsylvania–Early Permian zircons suggest depleted mantle reservoir sources, whereas Ordovician–Silurian detrital zircons were likely sourced from a combination of depleted mantle and enriched continental sources. Samples MY5-03 and FSC-02 contain clusters of Paleoproterozoic-age zircons that have both positive and negative εHf values (+1 to –15). FSC-02 has clusters of Mesoproterozoic and Neoproterozoic zircons with positive and negative εHf values that range from +10 to +5 (Fig. 6). Precambrian εHf values suggest a combination of magmatic sources that included enriched continental sources and depleted mantle reservoir sources.

**DISCUSSION**

Comparison of U-Pb ages and Hf isotopic values of detrital zircons from the Mystic assemblage with ages and geochemistry from possible magmatic source areas provides a powerful provenance approach to better understand the late Paleozoic tectonic history of the Farewell terrane. Low U-Th ratios (values <8) from detrital zircon grains within the Mystic assemblage suggest that the majority of grains were derived from igneous rather than metamorphic source areas. A summary and plot of U-Th data are available in GSA Data Repository, Appendix 2A (see footnote 1).

**Comparison with Potential Source Areas**

Late Paleozoic paleogeographic models for the Farewell terrane include restored locations along the northwestern margin of Laurentia (e.g., Plafker and Berg, 1994), marginal to Siberia in the Uralian seaway (Colpron and Nelson, 2009), or more generally, as part of a zone of convergence that extended from the Urals to offshore the Cordilleran margin of North America (Bradley et al., 2003). The wide range of magmatic source areas along these regions, a detrital zircon provenance investigation should ideally involve a comprehensive examination of the entire Cordillera and all Arctic margin crustal provinces. Such an approach is beyond the scope of this study given the limited number of samples. However, we do provide a general summary of potential Paleozoic magmatic sources from northwestern Laurentia and from Arctic terranes that experienced magmatism during the middle and late Paleozoic (Fig. 7). This summary focuses on magmatic events that overlap with prominent detrital zircon age populations from the Mystic assemblage in the Ordovician–Silurian (peak ages at 425, 435, and 440 Ma), Devonian–Mississippian (peak ages at 335, 349, and 380 Ma), and Pennsylvania–Permian (peak ages at 282 and 298 Ma) (Fig. 6).

Precambrian-age zircons are present in the Mystic assemblage and are primarily Paleoproterozoic, with peak ages of ca. 1870 Ma (Fig. 6). Given the relatively small percentages of these ages, it is likely that these grains have been recycled from sedimentary or metasedimentary sources. It is important to note that 2000–1800 Ma igneous zircons occur in both the Laurentian and Siberian cratons (as well as others) and do not represent a unique population for provenance comparisons (Prokopiev et al., 2008; Condie et al., 2009; Safonova et al., 2010).

**Uralian Source Areas**

Uralian magmatism was nearly continuous along the Arctic margin throughout much of the Paleozoic, but it has been divided into pulses that occurred at 460–420, 415–395, 365–355, 345–330, 320–315, and 290–250 Ma (Fershtater et al., 2007). Ordovician–Silurian grains from the Mystic assemblage overlap with magmatic events associated with early arc activity in the Uralides, which peaked between 460 and 420 Ma (Fershtater et al., 2007). Plutonic and volcanic rocks of this age are most abundant in the Tagil megazone and Platinum belt of the northwest.
Urals (Fershtater et al., 2007). Magmatism during this time was characterized by mafic to ultramafic primary melts associated with an island-arc setting (Fershtater et al., 2007).

Devonian–Mississippian subduction-related granites were emplaced during two episodes of magmatism at ca. 370–350 Ma and 335–315 Ma and yielded relatively primitive ranges of isotopic signatures ($\varepsilon_{\text{Nd}}$: −4.1 to +4.1 and −0.9 to +6.6, respectively; Bea et al., 2002). The timing of these magmatic episodes corresponds with Mississippian age populations in the Mystic assemblage; however, the isotopic signatures do not match well with the primarily enriched isotopes ($\varepsilon_{\text{Hf}}$: −10 to −30) observed in the detrital zircons.

Subduction processes are thought to have ceased in the Urals by the end of the Pennsylvanian (Bea et al., 2002; Brown et al., 2006). However, the generation of intracontinental granite bodies continued through the Permian and transitioned from initial intermediate to juvenile magmatism ($\varepsilon_{\text{Nd}}$: −2.2 to +6.4) to more enriched magmatism ($\varepsilon_{\text{Nd}}$: −12 to +5) during the latest Permian (Bea et al., 2002). Finally, it should be noted that much of the Pennsylvanian in the Urals was characterized by a lull in magmatism from 315 to 300 Ma (Fershtater et al., 2007), which overlaps, in part, with one of the primary age populations from the upper Mystic assemblage.

Peri-Laurentian Source Areas

Potential peri-Laurentian sources for the Mystic assemblage include rocks from Stikinia and Yukon-Tanana terrane and to a lesser extent, Quesnellia and the Slide Mountain assemblage. Devonian to Mississippian peak ages from the Mystic assemblage overlap with magmatism of the Middle to Late Devonian Ecstall cycle (390–365 Ma), Late Devonian–earliest Mississippian Finlayson cycle (365–357 Ma), Early Mississippian Wolverine cycle (357–342 Ma), and Late Mississippian Little Salmon cycle (342–314 Ma), which are associated with Yukon-Tanana and related terranes (Colpron et al., 2006; Piercey et al., 2006; Nelson et al., 2006). Middle to Late Devonian felsic igneous rocks have been reported from parts of the Endicott and Tracy Arm assemblages of the Yukon-Tanana terrane in southeast Alaska and British Columbia (McClelland et al., 1991; Gehrels et al., 1992; Currie, 1994; Gehrels, 2001) and from as far north as east-central Alaska and western Yukon (Dusel-Bacon et al., 2004, 2006). Although less voluminous, Devonian magmatism also occurs in continental margin rocks of western Laurentia in the Selwyn basin, Cassiar terrane, and the western Kootenay terrane of southeast British Columbia (Nelson et al., 2006; Paradis et al., 2006).

**Figure 7.** Summary of U-Pb detrital zircon ages and $\varepsilon_{\text{Hf}}$ isotope composition from the Mystic assemblage presented in the context of Cordilleran and circum-Arctic magmatic source areas. Refer to text for a summary of previous studies that helped define age ranges for potential magmatic source areas (DGS—Donjek Glacier suite, BGS—Barnard Glacier suite). Note that the lower Mystic assemblage (MA) also includes 25 zircon ages from age-equivalent strata reported by Bradley et al. (2007). For clarity, 1σ error bars of $\varepsilon_{\text{Hf}}$ values are not shown here but have been included in Figure 6. YTT—Yukon-Tanana terrane; CHUR—chondritic uniform reservoir.
Latest Devonian–Early Mississippian intermediate to mafic igneous rocks record a hallmark episode of magmatism in the Yukon-Tanana terrane in British Columbia, the Yukon, and parts of east-central Alaska (e.g., Dusel-Bacon and Aleinikoff, 1985; Aleinikoff et al., 1986; Mortensen, 1990; McClelland et al., 1991; Gehrels et al., 1992; Johnston et al., 1996; Alldrick, 2001; Szumigala et al., 2002; Day et al., 2003; Colpron et al., 2006; Dusel-Bacon et al., 2004, 2006; Mihalynuk et al., 2006; Murphy et al., 2006; Nelson et al., 2006; Roots et al., 2006). Late Devonian–Mississippian magmatism has also been recorded in the Queensel, Slide Mountain, and Stikine terranes in southwestern Yukon and western British Columbia (e.g., Mortensen, 1990; Currie, 1994; Gunning et al., 1994; Greig and Gehrels, 1995; Johnston et al., 1996; Gunning et al., 2006). Isolated occurrences of Early Mississippian metavolcanic and metaplutonic rocks are present within the Endicott Arm assemblage from southeast Alaska (Gehrels and Kapp, 1998).

Devonian–Mississippian felsic to intermediate plutonic and volcanic rocks occur in the Iskut and Scud River areas and Senemof Hills of Queenella (Gunning et al., 1994; Brown et al., 1996; Logan et al., 2000; Logan, 2004; Simard and Devine, 2003; Simard, 2003; Gunning et al., 2006). Some of the Early Mississippian plutonic complexes in northern Stikinia are thought to record relatively primitive arc magmatism (Logan et al., 2000; Logan, 2004). Middle to Late Mississippian-age igneous rocks are rare in the northern parts of the Yukon-Tanana terrane, as only isolated occurrences have been reported in association with the Fortymile River assemblage of east-central Alaska (Dusel-Bacon et al., 2006). Much of the Late Mississippian magmatism along the northern Cordillera is recorded in the southeastern parts of the Yukon-Tanana terrane in mafic to felsic plutonic rocks of the Wolf Lake–Jennings River and Glenlyon areas of northern British Columbia and southern Yukon (Nelson and Friedman, 2004; Colpron et al., 2006; Mihalynuk et al., 2006; Roots et al., 2006). Late Mississippian intermediate to felsic igneous rocks have also been reported from northwestern Stikinia (Mihalynuk et al., 1994; Gunning et al., 2006).

Grains of Pennsylvanian–Permian age overlap with the timing of the Klinkit magmatic cycle (314–269 Ma) associated with the Yukon-Tanana terrane (Colpron et al., 2006; Dusel-Bacon et al., 2006; Nelson et al., 2006; Roots et al., 2006). Pennsylvanian igneous rocks have been documented in the Stikine River region of Stikinia (Mihalynuk et al., 1994; Gunning et al., 1994, 2006), and Late Pennsylvanian–age felsic volcanic rocks (308 Ma) are present in the eastern part of Stikinia (Diakow and Rogers, 1998). However, volcanism in the Stikine terrane is thought to have ceased by latest Carboniferous time (Gunning et al., 2006). Pennsylvanian igneous rocks from the western margin of Stikinia exhibit enriched mid-ocean-ridge basalt (E-MORB) geochemical signatures (Logan, 2004) and are interpreted to have developed within an intra-arc or back-arc rift setting. Permain igneous rocks have been reported from the Taku terrane in southeast Alaska (Gehrels, 2002) and exhibit E-MORB to enriched plume-plate geochemical signatures (Stowell et al., 2000). Geochemical analyses from the Klinkit Group show that volcanic rocks have very minor crustal inheritance and reflect primitive (εNd:+6.7 to +7.4) isotopic signatures (Simard et al., 2003).

The age of the Klinkit Group overlaps with Pennsylvanian–Early Permian ages in detrital zircon populations from the Mystic assemblage. However, Klinkit cycle magmatism is primarily recorded by mafic volcaniclastic rocks, which are unlikely to be a significant zircon source. Consequently, most of the Klinkit magmatic cycle actually corresponds to a decline (or near absence) in the frequency of Late Pennsylvanian–Early Permian igneous ages in the Yukon-Tanana terrane (Nelson et al., 2006).

**Insular Terrane (Wrangellia and Alexander) Source Areas**

Previous work has demonstrated that the Wrangellia and Alexander terranes have been proximal to each other since late Paleozoic time (Gardner et al., 1988), and together formed a composite terrane that has been referred to as the Insular terrane (Monger et al., 1982). Ordovician–Silurian detrital zircons from the Mystic assemblage overlap closely in age with igneous rocks that are associated with the Descon and Klakas arcs of the Alexander terrane in southeast Alaska (Gehrels and Saleeby, 1987; Gehrels, 1990; Gehrels et al., 1996). Cambrian–Ordovician volcanic rocks also occur in the Donjek assemblage in the Saint Elias Mountains of SW Yukon (Dodds and Campbell, 1988; Mihalynuk et al., 1993; Beranek et al., 2012). Ordovician–Silurian igneous rocks of the Alexander terrane show relatively juvenile εNd values (+3.3 to +12; Samson et al., 1989; Gehrels et al., 1996; Cecil et al., 2011). This age population and isotopic signature have also been reported from detrital zircon data sets from middle to late Paleozoic-age rocks in the Banks Island assemblage and Alexander terrane (e.g., Beranek et al., 2012; Tochilin et al., 2014).

Isolated occurrences of Devonian-age igneous rocks have been reported from the Sicker Group of southern Wrangellia (Muller, 1980; Brandon et al., 1986; Dodds and Campbell, 1988; Andrew et al., 1991; Parrish and McNicol, 1992; Sluggett, 2003). More recent geochronologic studies from the Sicker Group on Vancouver Island make the case for magmatic pulses during the Late Devonian–Mississippian (370–339 Ma) and Pennsylvanian–Permian (reported ages of ca. 295 Ma, 309 Ma, and 313.5 Ma; Ruks and Mortensen, 2007; Ruks et al., 2010). Additionally, studies by Israel et al. (2014) documented a Late Devonian age (363.5 Ma) for gabbros from both Wrangellia and the Alexander terranes, as well as a 353.8 Ma age for a felsic tuff from the Station Creek Formation on Wrangellia. These recent findings indicate that igneous source rocks for most of the Devonian–Mississippian detrital zircon ages in the Mystic assemblage are present in the Wrangellia terrane and perhaps the Alexander terrane as well.

Pennsylvanian–Permian plutonic suites (320–285 Ma) have been reported from the Skolai arc and Strela metamorphics of Wrangellia (Nokleberg et al., 1986; Aleinikoff et al., 1988; Dodds and Campbell, 1988; Gardner et al., 1988; Beard and Barker, 1989; Plafker et al., 1989). Recent studies of magmatic rocks in the Alexander and Wrangellia terranes refined the timing of magmatism for the Barnard Glacier suite as ca. 301–307 Ma and that of the Donjek glacier suite to be 284–291 Ma (Beranek et al., 2014). Pennsylvania arc-related igneous rocks of the Skolai group have positive εNd values (εNd:+5.2 to +7.3; εHf:+10.6 to +11.4) suggestive of more depleted mantle sources (Greene et al., 2009). Geochemical values from the Barnard and Donjek Glacier suites (εNd:−0.5 to +6.8 and +2.1 to +6.1, respectively) suggest mixing of both depleted and slightly more enriched mantle sources (Beranek et al., 2014). Late Pennsylvanian–Permian magmatism and primarily juvenile isotopic signatures are both compatible with detrital zircon trends from the Mystic assemblage.

**Up-Section Trends in Provenance**

Detrital zircon ages from the lower parts of the Mystic assemblage cluster in two distinct Paleozoic age groups that include Ordovician–Silurian ages (peak ages at 435 and 450 Ma) and Devonian–Mississippian ages (peak ages at 335, 348, 380 Ma; Fig. 6). Ordovician–Silurian detrital zircons overlap closely with plutonic and volcanic source areas that are associated with the Descon and Klakas arcs of the Alexander terrane in southeast Alaska. Late Devonian–Early Mississippian detrital zircon ages overlap with plutonic and volcanic source areas from both Insular (Alexander and Wrangellia)
and peri-Laurentian (Yukon-Tanana, Stikinia, and Quesnellia) terranes. However, none of the peri-Laurentian terranes includes compatible magmatic sources for zircon populations of Ordovician–Silurian age present in the Mystic assemblage. The timing of magmatism in the Urals also overlaps, albeit somewhat variably, with both age populations but consists of Devonian–Mississippian igneous rocks that are far less enriched than isotopic signatures observed in the Mystic assemblage.

The upper units of the Mystic assemblage contain two primary Paleozoic age clusters of detrital zircons that include Ordovician–Silurian (peak ages at 440 and 425 Ma) and Pennsylvanian–Early Permian (peak ages at 298 and 282 Ma) age populations. The Late Pennsylvanian–Early Permian populations are perhaps the most diagnostic trend from the Mystic assemblage. With the exception of the Wrangellia and Alexander terranes, there is a general paucity of known magmatic ages from all of the source regions that are compatible with the range of Pennsylvanian–Permian ages in the Mystic assemblage. This population of detrital zircon ages, as well as their depleted isotopic composition, overlaps closely with sources in the Wrangellia and Alexander terranes. Additionally, a comparison with Permian-age units from the Alexander terrane and the associated Banks Island assemblage (Tochilin et al., 2014) reveals similar trends in both age populations and $\varepsilon_{Hf}$ values to that of the upper Mystic assemblage (Fig. 8).

Although magmatism documented in the Urals and peri-Laurentian terranes is correlative with some of the Paleozoic detrital zircon populations observed in the Mystic assemblage, only the Wrangellia and Alexander sources can account for all of these trends. Thus, provenance trends from the Mystic assemblage are best explained by a scenario in which the Farewell terrane was receiving primarily arc-derived detritus from exhumed parts of the Wrangellia and Alexander terranes during Pennsylvanian–Permian time and perhaps as early as the Mississippian.

**Tectonic and Paleogeographic Implications**

Given the size and scale of the Farewell terrane and the fact that we have only just begun to understand the basic geologic framework of western Alaska, it is difficult at this stage to confidently summarize in detail the Neoproterozoic to late Paleozoic paleogeographic history of the Farewell terrane. However, this study serves as a small contribution and offers a revised interpretation of the tectonic history and paleogeography of the Farewell terrane. Our data are consistent with a model where the Farewell terrane was in the Panthalassic Ocean (along the outboard margin of the Slide Mountain Ocean) and adjacent to components of the Insular terrane during the Mississippian–Pennsylvanian (Fig. 9). This is also somewhat consistent with the hypothesis put forth by Bradley et al. (2003) for the Browns Fork orogen. While the precise location and basin setting of the Farewell terrane during this time are debatable, it was most likely proximal to, and receiving detrital contributions from, the Wrangellia and Alexander terranes.

As previously mentioned, the Browns Fork orogeny recognized by Bradley et al. (2003) was based on a ca. 285 Ma metamorphic age in the Kuskokwim Mountains (proposed hinterland) and the occurrence of a thick succession of Pennsylvanian–Permian coarse-grained clastics comprising the Dall basin (proposed foreland basin) in the Mystic subterrane. The Mystic assemblage and Mount Dall conglomerate may partially overlap in age; however, their stratigraphic relationships and/or structural juxtaposition have not been verified. Beranek et al.
Pennsylvanian–Early Permian paleogeographic map showing a reinterpreted location of the Farewell terrane (Mystic subterrane) (FW/MYS) in the Panthalassic Ocean and proximal to peri-Laurentian terranes along the western margin of the Slide Mountain Ocean. Paleogeographic map is modified from Nelson et al. (2013). Dashed FW/MYS depicts a previously proposed late Paleozoic paleogeographic locality along the Siberian margin (Colpron and Nelson, 2009). New provenance data from the Mystic assemblage support a revised paleogeographic model where the Farewell terrane was in a basinal setting that was proximal to parts of the Alexander (AX) and Wrangellia (WR) terranes during Mississippian–Early Permian time (depicted by solid FW/MYS). ST—Stikinia, YT—Yukon-Tanana, QN—Quesnellia, EK—eastern Klamath terranes, NS—northern Sierra terranes, AA—Arctic Alaska, OM—Omulevka, SIB—Siberia, KAZ—Kazakhstania, BAR—Barentsia, BAL—Baltica, LAU—Laurentia, SAM—South America, AFR—Africa, ARB—Arabia.

Mystic assemblage overlap with magmatic source areas from the Wrangellia, Alexander, and Yukon-Tanana terranes and in part with Stikinia and Quesnellia. Pennsylvanian–Permian zircon populations in the upper part of the Mystic assemblage overlap in age with magmatic sources from the Wrangellia and Alexander terranes, which also correspond to times of minimal zircon production from Yukon-Tanana and related terranes. Maximum depositional ages from the Mystic assemblage overlap in part with previously reported Mississippian–Pennsylvanian fossil age ranges for the Mystic assemblage and extend the upper age limit of these strata to at least the Early Permian. Provenance trends from the Mystic assemblage are best explained by a paleogeographic model wherein the Farewell terrane was in the Panthalassic Ocean (along the outboard margin of the Slide Mountain Ocean) proximal to, and receiving detritus from, arc and recycled orogen sources of the Alexander and Wrangellia terranes during Mississippian–Early Permian time.

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SUMMARY AND CONCLUSIONS
Mississippian–Early Permian clastic strata of the Mystic assemblage were deposited in a basin (deep-water) submarine fan setting that was characterized by both high- and low-density sediment gravity flows. Modal composition trends from the Mystic assemblage support a model wherein sediment was derived from both arc and recycled orogen source areas. Detrital zircon trends from the lower parts of the Mystic assemblage primarily consist of Devonian–Mississippian grains that exhibit negative εHf values. Secondary age populations include Ordovician–Silurian (416–460 Ma) ages with negative εHf values and Paleoproterozoic-age detrital zircons (1.8–2.0 Ga), which exhibit a range of positive to negative εHf values. The upper part of the Mystic assemblage consists of Pennsylvanian–Permian and Ordovician–Silurian detrital zircon grains with sparse occurrences of Paleoproterozoic–Mesoproterozoic grains. Pennsylvanian–Permian detrital zircons exhibit entirely positive εHf values, while Ordovician–Silurian and Precambrian grains have a range of positive to negative εHf values.

Ordovician–Silurian detrital zircons are found throughout the Mystic assemblage and overlap closely with magmatic sources in the Alexander terrane. Devonian–Mississippian detrital zircon ages in the lower part of the


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