RESEARCH ARTICLE

10.1002/2017TC004520

Key Points:

- The Harmony Formation provides a distinctive marker to constrain tectonic models of the western Laurentian margin in Early Paleozoic time
- Detrital zircon U-Pb ages and Hf isotope ratios indicate NW Laurentian origin for the Harmony and subsequent interpreted tectonic passage
- These insights help refine a tectonic model for the Antler orogeny and thus the western Laurentian plate boundary in Early Paleozoic time

Supporting Information:

- Supporting Information S1
- Data Set S1

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Citation:

Linde, G. M., Trexler, J. H., Jr., Cashman, P. H., Gehrels, G. & Dickinson, W. R. (2017). Three-dimensional evolution of the Early Paleozoic western Laurentian margin: New insights from detrital zircon U-Pb geochronology and Hf isotope geochemistry of the Harmony Formation of Nevada. *Tectonics*, 36, 2347–2369. <https://doi.org/10.1002/2017TC004520>

Received 18 FEB 2017

Accepted 10 SEP 2017

Accepted article online 18 SEP 2017

Published online 4 NOV 2017

Three-Dimensional Evolution of the Early Paleozoic Western Laurentian Margin: New Insights From Detrital Zircon U-Pb Geochronology and Hf Isotope Geochemistry of the Harmony Formation of Nevada

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Abstract Uranium-lead (U-Pb) geochronology and Hafnium (Hf) isotope geochemistry of detrital zircons of the Harmony Formation of north central Nevada provide new insights into the tectonic evolution of the Late Paleozoic western Laurentian margin. Using laser-ablation inductively coupled plasma mass spectrometry, 10 arenite samples were analyzed for U-Pb ages, and 8 of these samples were further analyzed for Hf isotope ratios. Three of the sampled units have similar U-Pb age peaks and Hf isotope ratios, including a 1.0–1.4 Ga peak with ϵ_{Hf} values of +12 to −3 and a 2.5–2.7 Ga peak with ϵ_{Hf} values of +7 to −5. The remaining seven samples differ significantly from these three, but are similar to one another; having age peaks of 1.7–1.9 Ga with ϵ_{Hf} of +10 to −20 and age peaks of 2.3–2.7 Ga with ϵ_{Hf} of +6 to −8. The data confirm the subdivision of the Harmony Formation into two petrofacies: quartzose (Harmony A) and feldspathic (Harmony B). The three samples with 1.0–1.4 and 2.5–2.7 Ga peaks are the Harmony A, which originated in the central Laurentian craton. The other seven samples are the Harmony B, which originated in eastern Alberta-western Saskatchewan, north of the Harmony A source. We propose that all Harmony Formation strata were deposited near eastern Alberta and subsequently tectonically interleaved with Roberts Mountains allochthon strata. We interpret that the entire package was tectonically transported south along the western Laurentian margin and then emplaced eastward onto the craton during the Late Devonian-Early Mississippian Antler orogeny.

1. Introduction

Starting in mid-Paleozoic time, the western margin of Laurentia evolved from a passive margin to an active margin characterized by the accretion of allochthonous terranes. Some of these terranes formed along the western margin of Laurentia (e.g., Colpron et al., 2007; Nelson et al., 2006; Nelson & Colpron, 2007). Others have been linked to Baltica, Siberia, or Caledonia, based on faunal characteristics, igneous activity, detrital zircon (DZ) ages, and other criteria (Colpron & Nelson, 2009, and references cited therein). Sharply divergent models to explain these disparate terrane origins include a Scotia-type arc that migrated around the southern margin of Laurentia (Wright & Wyld, 2006), or, alternatively, around its northern margin (Colpron & Nelson, 2009). The kinematics of the Late Paleozoic plate boundary in western Laurentia have been obscured by subsequent Mesozoic and Cenozoic plate boundary tectonism (e.g., Colpron & Nelson, 2009; Dickinson, 2004). In this study, we focus on distinctive rocks *inboard* of the Paleozoic plate boundary to determine the nature and kinematics of the western Laurentian plate margin in mid-Paleozoic time.

In the western U.S., the first Paleozoic tectonism was the Antler orogeny, associated with accretion of the Roberts Mountains allochthon (RMA) and arc volcanism in the Klamath and Northern Sierra terranes (Dickinson, 2004, 2009; Domeier & Torsvik, 2014) (Figure 1). Inboard of the accretionary margin, the first Paleozoic tectonic signal in the western U.S. was latest Devonian formation of the Antler foreland basin (Dickinson et al., 1983) and emplacement of the RMA (Dickinson, 2004; Dickinson, 2009; Domeier & Torsvik, 2014) (Figure 2). The tectonic setting of the Antler orogeny is poorly understood, due to anomalous characteristics of the orogeny and of the allochthon itself (summarized below). However, the Harmony Formation, a fault-bounded, feldspathic arenite, provides a distinctive marker to constrain tectonic models. The Harmony Formation is mapped as the highest structural unit in the imbricated RMA except in the Sonoma Range, where it is found both structurally above and below other RMA thrust blocks (Ferguson et al., 1951; Gilluly, 1967; Hotz & Willden, 1964; Madrid, 1987; Roberts, 1964) (Figure 2). This paper uses new DZ uranium-lead

Figure 1. Map of the western cordillera of North America today, showing terranes and provinces discussed in the text. Map is after Colpron and Nelson (2009).

(U-Pb) ages and hafnium (Hf) isotope ratios of the Harmony Formation to identify its provenance. The provenance constrains its transport history.

The purpose of this paper is to (1) present new U-Pb age and Hf isotope ratios of the Harmony Formation, (2) compare these to signatures of Laurentian basement rocks to determine the provenance of the unit, (3) compare the signatures of passive margin strata with those of the Harmony Formation and the Roberts Mountains allochthon to constrain the origin and relationships of all three, and (4) discuss a transpressive margin system as a tectonic model of deposition and structural emplacement that is compatible with these observations. The result contributes to the tectonic model for the Antler orogeny and the evolution of western Laurentia.

2. Mid-Paleozoic Tectonostratigraphic Framework of Western Laurentia

2.1. Initiation of Active Plate Margin in Western Laurentia

In Alaska and British Columbia, the initiation of subduction and subsequent terrane accretion along the western margin of Laurentia is recorded by Devonian magmatism that propagated southward between 400 Ma and 360 Ma (Colpron & Nelson, 2009, and references therein). Arc magmatism from circa 402–366 Ma occurred in the southern Brooks Range of Alaska (Colpron & Nelson, 2009; McClelland et al., 2006)

Figure 2. Map of north central Nevada, showing sample locations (colored symbols), number of samples in each range, and the traces of the Roberts Mountains and Golconda thrusts. Some RMA rocks crop out to the west of the Golconda thrust in tectonic windows through the allochthon as shown in the schematic, not-to-scale cross section. An inset map of Nevada shows where the larger-scale map is located, as well as the location of the Antler orogenic highlands, the Antler Foreland Basin, and other mountain ranges mentioned in the text. Thrust traces are after Dickinson (2006); Antler highlands and basin are after Poole (1974). (Harmony Formation mapping: Ferguson et al., 1951; Roberts, 1964; Hotz & Willden, 1964; Gilluly, 1967; Madrid, 1987; Doebrich, 1994; Theodore et al., 1994; Jones, 1997a, 1997b).

(Figure 1). Late Middle Devonian (circa 387 Ma) arc magmatism is recorded in the Yukon-Tanana terrane of southeastern Alaska and western British Columbia (Nelson et al., 2006) and in southern British Columbia's Kootenay terrane (Schiarrizza & Preto, 1987) (Figure 1).

In the western U.S., in contrast, continental margin sedimentation continued uninterrupted until the Late Devonian-Early Mississippian Antler orogeny. The final Neoproterozoic rifting that separated the Rodinian supercontinent lasted from circa 570–520 Ma and was followed by the deposition of passive margin sediments through mid-Devonian time (Dickinson, 2009; Poole et al., 1992; Yonkee et al., 2014). The quiescent interval came to an end with the Late Devonian-Early Mississippian Antler orogeny, during which the RMA was emplaced onto the western Laurentian margin (e.g., Dickinson, 2006; Nilsen & Stewart, 1980) (Figures 1 and 2). There was no associated magmatism.

The eastern Klamath, northern Sierran, and Alexander terranes are commonly interpreted as exotic to Laurentia (e.g., Bazard et al., 1995; Beranek et al., 2016, 2013; Colpron & Nelson, 2009; Gehrels et al., 1996; Grove et al., 2008) (Figure 1). These terranes were accreted to the western Laurentian margin during Late Devonian-Early Mississippian time (eastern Klamath and northern Sierra) and mid-Jurassic time (Alexander terrane) (Colpron & Nelson, 2009). Two distinct models have recently been proposed to account for the transport and subsequent accretion onto the western Laurentian margin of these far-traveled terranes: tectonic transport around either the southern or northern margin of Early Paleozoic Laurentia.

Wright and Wyld (2006) proposed a model in which the Alexander, eastern Klamath, and northern Sierran terranes originated near Neoproterozoic Gondwana (Figure 3, Option IIa). These terranes, along with the Roberts Mountains allochthon, were transported in Early Paleozoic time southward along the eastern margin of Laurentia by a strike-slip translation (Wright & Wyld, 2006). Subsequently, a Scotia-style arc developed along the southern Laurentian-northern Gondwanan margins and migrated westward, transporting the terranes to

Figure 3. Contrasting tectonic models proposed to explain the source and tectonic transport of the Roberts Mountains allochthon and Harmony Formation, shown in Early Devonian time. Exotic terranes include the Alexander, Klamath, and Northern Sierran terranes. AX: Alexander Terrane; YR: Yreka Terrane; NS: Northern Sierra Terrane; RMA: Roberts Mountains allochthon (different locations of RMA are as visualized in different tectonic models and are color coded by model); PRA: Peace River Arch; SRA: Salmon River Arch. Map is after Colpron and Nelson (2009). Option I, western equivalent (Poole et al., 1992; Roberts et al., 1958); option IIa, exotic peri-Gondwana (southern transport) (Wright & Wyld, 2006); option IIb, exotic Baltica (northern transport) (Colpron & Nelson, 2009); option III, northwestern Laurentia (Gehrels, Dickinson, Riley, et al., 2000; Suczek, 1977).

the western Laurentian margin (Wright & Wyld, 2006) (Figure 3, Option IIa). The terranes were transported northward along the western margin by slab rollback and subsequently accreted to the margin (Wright & Wyld, 2006) (Figure 3, Option IIa).

A second model explaining the transport and emplacement of these terranes proposed a westward migrating arc north of Laurentia, evolving into a southward propagating transpressional system along western Laurentia (Colpron & Nelson, 2009) (Figure 3, Option IIb). A “Northwest Passage” opened between Laurentia and Siberia in mid-Paleozoic time, within which an arc developed along the northern Laurentian margin in the Early Devonian (Colpron et al., 2007; Colpron & Nelson, 2009) (Figure 3, Option IIb). Terranes were transported from their origin in Baltica to northwestern Laurentia through the “Northwest Passage” by means of the westward migration of the subduction zone (Figure 3, Option IIb). By Middle Devonian time, a sinistral transpressional system developed at the southern end of this arc and gradually propagated southward along western Laurentia and transported the terranes and fragments south along the margin. Colpron and Nelson (2009) inferred that the progressively younger tectonism southward along the Laurentian margin records the southward propagation of the transpressional system. They propose that this fault system could have provided the weakness along which Devonian shortening initiated, resulting in the emplacement of the RMA (Colpron & Nelson, 2009).

2.2. Antler Orogeny

The major structural expression of the Late Devonian–Early Mississippian Antler orogeny in Nevada and Idaho is the emplacement of Cambrian through Devonian oceanic sedimentary rocks, the RMA, eastward onto the western Laurentian craton (e.g., Nilsen, 1977; Nilsen & Stewart, 1980; Poole et al., 1992; Roberts et al., 1958). Strata of the RMA structurally overlie coeval rocks of the western Laurentian passive margin (e.g., Kay, 1951; Madrid, 1987; Roberts et al., 1958; Schuchert, 1923). East of the Antler orogen and west of the Laurentian

craton, sediments shed from the rising Antler highlands filled the Antler foreland basin (Poole, 1974; Trexler et al., 2003) and the Copper Basin of Idaho (Beranek et al., 2016; Link et al., 1996) between Devonian and Early Mississippian time (Figure 2). The age of the Antler orogeny is bracketed between latest Devonian synorogenic strata (e.g., Murphy et al., 1984) and unconformably overlying Late Mississippian successor basin strata (e.g., Trexler et al., 2003). Although the emplacement direction for the RMA is widely cited as “eastward,” the kinematics are not well constrained because no known Antler structures are dated by good biostratigraphic control. Also, in most places, several Late Paleozoic structural events postdate the Antler orogeny and predate rocks of the “Antler overlap sequence” (e.g., McFarlane, 2001; Trexler et al., 2003). So, although Late Paleozoic structures in the region are commonly attributed to the Antler orogeny, their ages are poorly constrained and they may well have formed during one of the subsequent Late Paleozoic events.

The Antler orogeny has several anomalous characteristics, leading to longstanding debate about its tectonic setting (e.g., Burchfiel & Royden, 1991; Nilsen & Stewart, 1980). There are no associated plutonic or high-grade metamorphic rocks. The foreland basin section is anomalously thin and does not contain volcanoclastic rocks. Although many structures have been attributed to the Antler orogeny, none, including outcrops of the “Roberts Mountains thrust” at the base of the allochthon, are demonstrably Late Devonian/Early Mississippian in age. Additional debates center on timing, polarity, or other aspects of specific models (e.g., Burchfiel & Royden, 1991; Ketner et al., 2005; Nilsen & Stewart, 1980).

2.3. Roberts Mountains Allochthon

The RMA is a large, internally complex allochthon. It crops out in north central Nevada between the Roberts Mountains thrust on the east and the Golconda thrust on the west, though some units are exposed in tectonic windows west of the Golconda thrust (Figure 2). Rocks in the RMA range from Cambrian to Devonian in age and include chert, argillite, sandstone, and greenstone. Strata of the allochthon structurally overlie coeval rocks of the western Laurentian passive margin (e.g., Kay, 1951; Madrid, 1987; Roberts et al., 1958; Schuchert, 1923). The metamorphic grade of RMA rocks is generally greenschist facies or lower (Gehrels, Dickinson, Riley, et al., 2000). RMA strata are highly deformed and include imbricated older-over-younger thrust sheets (e.g., Evans & Theodore, 1978; Noble & Finney, 1999; Oldow, 1984). Together, they have been interpreted as a westward thickening subduction assemblage (e.g., Dickinson, 2006; Speed & Sleep, 1982).

A recent study of detrital zircon U-Pb ages and Hf isotope ratios found that the RMA strata comprised two distinct “families” of strata, distinguishable by DZ U-Pb ages and Hf isotope ratios (Linde et al., 2016). These two families are the Ordovician Lower Vinini Formation and the remainder of the Ordovician-Devonian RMA strata. The study proposed that the Lower Vinini Formation sands originated in central Laurentia, and the remainder of the Ordovician through Devonian RMA sands originated in the Peace River Arch region of eastern British Columbia-western Alberta (Figure 1), and that all of the RMA strata were deposited near the Peace River Arch (Linde et al., 2016). It was hypothesized that the strata were subsequently tectonically transported south along the western Laurentian margin and emplaced onto the craton during the Antler orogeny (Linde et al., 2016).

2.4. Harmony Formation

2.4.1. Composition and Age of the Harmony Formation

The Harmony Formation is primarily a distinctive, texturally immature, feldspathic arenite commonly considered to be a part of the RMA. It structurally overlies, or is imbricated with, RMA units (Ferguson et al., 1951; Gilluly, 1967; Hotz & Willden, 1964; Madrid, 1987; Roberts, 1964). The Harmony Formation was first described in Harmony Canyon of the Sonoma Range as a coarse micaceous and feldspathic sandstone (Ferguson et al., 1951) (Figures 2 and 4). Interbedded shale and limestone and graded beds are common (Ferguson et al., 1951; Roberts, 1964) (Figure 4). The initial estimate of the thickness of the Harmony Formation was up to 1,524 m (Ferguson et al., 1951), but because the basal contact of the unit is faulted (Figure 5), its true thickness is unknown (Ferguson et al., 1951, 1952; Hotz & Willden, 1964; Roberts, 1964).

Initially mapped as one lithostratigraphic formation (e.g., Doebrich, 1994; Gilluly, 1967; Roberts, 1951), the Harmony Formation actually comprises two distinct units (Gehrels, Dickinson, Riley, et al., 2000). Designated as petrofacies, these can be distinguished by DZ spectra and, with less clarity, by composition. The generally more quartzose, more texturally mature petrofacies is known as “Harmony A” and the more feldspathic, more texturally immature petrofacies as “Harmony B” (Gehrels, Dickinson, Riley, et al., 2000).

Figure 4. The Harmony Formation in Little Cottonwood Canyon, Galena Range, Nevada. (left) The 1 m red bar is next to a 1 m trekking pole. (right) The scale totals 15 cm. Photos by Gwen Linde.

The distinctive Harmony B crops out everywhere the Harmony Formation is found, while the Harmony A has been described only in Little Cottonwood Canyon of the Galena Range (Figure 2).

Although the depositional age of the Harmony Formation has been interpreted variously as Cambrian, Devonian, and Devonian-Mississippian, it is considered herein to be Cambrian. No fossils have been found in the Harmony A, and its age control is only by association with the Harmony B. Late Cambrian fossils of North American affinity have been identified in Harmony B outcrops in the Osgood Mountains and the Hot Springs Range (Hotz & Willden, 1964). Clasts and olistoliths of the Harmony B have been observed in the Devonian Scott Canyon Formation, near the mouth of Galena Canyon in the Galena Range (Doebrich, 1994; Theodore et al., 1994; T. Theodore, personal communication, 2015), and confirmed by our fieldwork (Figure 6). The only datum supporting a Devonian or younger age is a single conodont element interpreted as Devonian (Jones, 1997a, 1997b). Weighing the unrepeatable nature of the single occurrence against the other evidence presented, we consider the Harmony B to be Cambrian. Analytical results of this study pertain to the age of the Harmony and are discussed below.

2.4.2. Location and Structural Position of the Harmony Formation

The Harmony B crops out in several ranges in north central Nevada (Figure 2); its basal contact is most commonly structural (Figure 5). In the Hot Springs Range, the Harmony B rests on the Cambrian Paradise Valley Chert with a contact that has been interpreted as depositional (Hotz & Willden, 1964), structural (Madrid, 1987), or obscured (Jones, 1997a) (Figures 2 and 5). In the other locations where the Harmony Formation is exposed, it is mapped above the uppermost thrust in the RMA. An exception is in the Sonoma Range, where it is found *within* the RMA, both structurally above and structurally below the Valmy Formation (Ferguson et al., 1951; Gilluly, 1967; Hotz & Willden, 1964; Madrid, 1987; Roberts, 1964) (Figures 2 and 5).

Harmony A outcrops have been found only in the Galena Range (Figure 2). Initial workers interpreted the contact between the Harmony A and B in Little Cottonwood Canyon as sharp, but concordant, and proposed that the two petrofacies recorded two successive submarine fan deposits (Dickinson & Gehrels, 2000).

Upper Paleozoic rocks of the Antler overlap sequence unconformably overlie the Harmony Formation. In the Sonoma Range, the Pennsylvanian-Permian Antler Peak Limestone was deposited on the Harmony Formation (Ferguson et al., 1951). In the Galena Range, the conglomeratic Pennsylvanian Battle Formation was deposited on the Harmony Formation (Roberts, 1964). Similar relationships are found in Idaho, where the Sun Valley Group overlap sequence unconformably overlies RMA-equivalent strata (Link et al., 1996).

The Harmony Formation must have been exposed to erosion as part of the advancing RMA by Mississippian time, because clasts of the distinctive Harmony B occur in Mississippian strata of the Antler foreland basin.

Figure 5. Tectonostratigraphic diagram of units of the Roberts Mountains allochthon in north central Nevada mountain ranges in which the Harmony Formation crops out, showing locations of DZ samples. Units are shown in their physical, structurally superimposed, and not chronostratigraphic order. Units are shown from west to east (left to right) as indicated. Most units are internally disrupted with multiple imbricate thrusts not shown on this chart. Autochthonous units are shown with crosshatching. RMT: Roberts Mountains Thrust, as mapped by Ferguson et al. (1951), Hotz and Willden (1964), Wrucke (1974), Madrid (1987), and Theodore et al. (1994).

Figure 6. The Devonian Scott Canyon Formation with Harmony Formation clast inclusion, in Galena Canyon, Galena Range, Nevada. The red arrow points to the Harmony Formation clast. Photo by Gwen Linde.

These Harmony Formation clasts are exposed in strata in the Diamond Mountains (Harbaugh, 1980), the Adobe Range (Ketner & Ross, 1990), Independence Mountains, Piñon Range, East Range, and Fish Creek Mountains (Ketner, 1998) (Figure 2).

3. Approach and Methodology

3.1. The Harmony Formation—A Distinctive Kinematic Marker

The long-recognized paradox of the immature Harmony Formation rocks in the highest, presumably farthest-traveled, structural slice in the RMA presents an opportunity to test models for the tectonic setting of the Antler orogeny. The readily identifiable Harmony B feldspathic arenite is a good kinematic marker for motion along the western margin of Laurentia. It is texturally and compositionally immature, documenting minimal sedimentary transport between erosion and deposition. The sedimentology of the Harmony Formation provides an additional constraint on its tectonic history: sole marks, graded beds, plane beds, and interbedded marine shales, as well as facies T_A , T_B , and T_C of the classic Bouma (1963) sequence establish that these rocks are turbidites deposited on inner and middle submarine fans (Suczek,

1977). Our approach to resolving the Harmony paradox is to use new U-Pb ages and hafnium Hf isotope analyses of the Harmony Formation to identify the provenance of the DZ grains and then use this to constrain the mode and kinematics of transport of Harmony Formation rocks along the Laurentian margin. This section starts with a review of the contrasting models for the origin and transport of RMA rocks, and the debate about the source of the Harmony Formation. A summary of the geologically distinct crustal provinces that could be source terranes follows. We then document the analytical methods used to acquire DZ data in this study.

3.1.1. Provenance of the Harmony Formation

The provenance of the arkosic Harmony sediment is controversial. Some workers proposed derivation from a western landmass, because they could not identify a feldspathic granitoid source to the east (Ferguson et al., 1951; Ketner, 1977). The Salmon River Arch of western Idaho was cited by some workers as the Harmony source, based on lithological similarities between granitic rocks there and the Harmony B (Rowell et al., 1979; Schweickert & Snyder, 1981; Stewart & Suczek, 1977; Suczek, 1977) (Figures 1 and 3). Early U-Pb age analyses of detrital zircons suggested that the source was south of the present Harmony outcrops (Smith & Gehrels, 1994; Wallin, 1990). Subsequently, more extensive U-Pb analyses of detrital zircons led to the proposal that the Harmony A originated in central or southern Laurentia, and the Harmony B was derived from northern Laurentia (Gehrels, Dickinson, Riley, et al., 2000) (Figure 3). Later workers suggested that the Harmony Formation was an exotic accreted terrane, based on the absence of local source terranes (Jones-Crafford, 2008; Ketner et al., 2005).

3.1.2. Tectonic Models for Derivation of the Harmony Formation

Three general models have been invoked to explain the provenance, deposition, and transport of the RMA and the Harmony Formation (Figure 3). Option I is that the oceanic sediments of the RMA/Harmony are the coeval, deep marine equivalent of the autochthonous passive margin strata of eastern Nevada (e.g., Burchfiel & Davis, 1972; Poole et al., 1992; Roberts et al., 1958) (Figure 3, Option I). This model is based upon the similar ages of the autochthonous and allochthonous strata, and the relative physical proximity of the proposed basin of deposition, directly offshore of the final position of the allochthonous strata. The two versions of the second model call for an exotic, extra-Laurentian origin for the RMA/Harmony, and subsequent tectonic transport to offshore of the emplacement location. Option IIa proposes a peri-Gondwanan origin for the RMA/Harmony with subsequent transport around the southern margin of Laurentia (Wright & Wyld, 2006) (Figure 3, Option IIa). Option IIb invokes origin in Baltica and subsequent tectonic transport around the northern margin of Laurentia (Colpron & Nelson, 2009) (Figure 3, Option IIb). Both options are based on similarity of DZ ages and geologic history of the RMA/Harmony to those of other terranes interpreted to be exotic and far traveled. Option III calls for a distant, though still Laurentian, origin for the RMA/Harmony and *sedimentary* transport prior to emplacement (Figure 3, Option III). The Peace River Arch region of western Canada is a possible source, based on similarity of detrital zircon U-Pb ages (Gehrels, Dickinson, Riley, et al., 2000). Finally, some workers suggested the Salmon River Arch region of Idaho as a potential source, based on interpreted and lithologic similarities between these strata and the Harmony Formation (Suczek, 1977) (Figures 1 and 3).

3.2. Crustal Provinces in the North American Craton

The North American craton contains several Proterozoic and Archean crustal provinces that are geologically distinct source terranes for the strata deposited on the rifted western margin of Laurentia (e.g., Gehrels et al., 2011, and references cited therein) (Figure 7).

The Yavapai-Mazatzal Province (1.8–1.6 Ga) extends across central North America (Hoffman, 1989; Whitmeyer & Karlstrom, 2007) (Figure 7). The granite-rhyolite province (1.48–1.34 Ga) crops out within the Yavapai-Mazatzal Province (Anderson & Morrison, 1992; Bickford et al., 1986). The Yavapai-Mazatzal Province is bounded on the north and northwest by the Trans-Hudson orogenic terrane (Whitmeyer & Karlstrom, 2007) (2.0–1.8 Ga) and Archean rocks (> 2.5 Ga) of the Wyoming and Superior provinces (Figure 7), on the east and southeast by the terranes of the Grenville orogen (1.2–1.0 Ga) (Hoffman, 1989), and on the west by the Mojavia terrane (> 2.0–2.4 Ga with 1.7–1.8 Ga arcs) (Nelson et al., 2011; Whitmeyer & Karlstrom, 2007).

Detrital-zircon data demonstrate that sources for the upper Neoproterozoic-early Cambrian western Laurentian passive margin changed between upper Neoproterozoic and Lower Cambrian time (Linde et al., 2014a, and references cited therein). A significant sediment source for the upper Neoproterozoic

Figure 7. Locations of crustal provinces in North America that are potential source terranes for western Laurentian sediments. The location of Transcontinental Arch is after Sloss (1988); WY: Wyoming province; HE: Hearn province; SU: Superior province; CORD: Cordilleran; APP: Appalachian; OA: Ouachita-Marathon. Figure is after Gehrels et al. (2011) and compiled from Bickford et al. (1986), Hoffman (1989), Ross (1991), Burchfiel et al., (1992), Anderson and Morrison (1992), Bickford and Anderson (1993), Van Schmus et al. (1993), Villeneuve et al. (1993), Dickinson and Lawton (2001), and Dickinson and Gehrels (2009).

Laurentian passive margin strata from the northwest U.S. to Sonora, Mexico, was the 1.2–1.0 Ga Grenville orogen of southern and eastern North America (e.g., Gehrels & Pecha, 2014; Lawton et al., 2010; Linde et al., 2014a; Rainbird et al., 1997, 2012; Yonkee et al., 2014) (Figure 7). In contrast, the 1.8–1.6 Ga Yavapai-Mazatzal and 1.48–1.34 Ga midcontinent granite-rhyolite provinces are the dominant sediment sources of many strata higher in the passive-margin section (e.g., Gehrels & Pecha, 2014; Lawton et al., 2010; Linde et al., 2014a; Yonkee et al., 2014) (Figure 7).

3.3. Analytical Methods

This study reports U-Pb and Hf isotope DZ analyses from 10 arenite samples of the Harmony Formation (Figures 2 and 5 and Table 1). Zircons from six of these samples were previously analyzed by isotope dilution-thermal ionization mass spectrometry (ID-TIMS) (Gehrels, Dickinson, Riley, et al., 2000; Smith & Gehrels, 1994) (Table 1). This study reports ~200 additional U-Pb analyses and ~50 new Hf isotope analyses of zircons from these same six samples by laser ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS) (Figure 5 and Table 1). In addition, four additional arenite samples were used for ~100 LA-ICP-MS U-Pb zircon analyses. These samples have been described in Linde et al. (2013). For two of these samples, ~25 of the grains were further analyzed by LA-ICP-MS for Hf isotope ratios.

Table 1
Locations of Samples Analyzed in This Study Referenced to UTM Locations

Sample location UTM: NAD 83 (11°N)	Easting	Northing	Sample from	Harmony A/B	Hf?
Elbow Canyon <i>Sonoma Range</i>	441026	4514450	Gehrels et al. (2000)	B	Yes
Harmony Canyon <i>Sonoma Range</i>	446225	4533064	Gehrels et al. (2000)	B	Yes
Kluncy Canyon <i>Sonoma Range</i>	0447927	4531085	Linde et al. (2013)	B	No
Gough's Canyon <i>Osgood Mountains</i>	0471723	4556266	Linde et al. (2013)	B	No
Little Cottonwood Canyon #10 <i>Galena Range</i>	0490510	4495682	Linde et al. (2013)	A	Yes
Little Cottonwood Canyon #9 <i>Galena Range</i>	0491232	4495602	Linde et al. (2013)	B	Yes
Little Cottonwood Canyon #4 <i>Galena Range</i>	490623	4494977	Gehrels et al. (2000)	B	Yes
Little Cottonwood Canyon #3 <i>Galena Range</i>	490459	4494916	Gehrels et al. (2000)	B	Yes
Little Cottonwood Canyon #2 <i>Galena Range</i>	490529	4495008	Gehrels et al. (2000)	A	Yes
Little Cottonwood Canyon #1 <i>Galena Range</i>	490600	4494792	Smith and Gehrels (1994)	A	Yes

The reanalysis of the original six arenite samples incorporated new methodology. Grains were selected randomly rather than by color or morphology. The sample size was increased to 200 grains for the U-Pb ages, to obtain a more statistically significant sample. Hf isotope analyses were incorporated. The additional sampling was designed to broaden the geographic scope into new drainages in the Sonoma Range and the Osgood Mountains and to test the finding of two distinct petrofacies in the Galena Range (Gehrels, Dickinson, Riley, et al., 2000).

Zircons were separated and analyzed at the University of Arizona LaserChron facility using standard techniques to yield a best age distribution reflective of the true distribution of detrital-zircon ages in each sample (Gehrels & Pecha, 2014). Zircon grains were selected from all areas of the sample mount and not biasing selection by size, crystal shape, or color. Grains with fractures, inclusions, or zonation were excluded. This is a different grain selection procedure than was used in the earlier study. In the ID-TIMS study, zircon crystals were selected based on color and morphology (Gehrels, Dickinson, Riley, et al., 2000). In the current study, the analyses were positioned in the zircon cores identified by cathodoluminescence imaging to reduce the chance of analyzing overgrowths that might be compromised by Pb loss. Hf analyses were not conducted on every arenite sample due to cost constraints. Hf analyses were positioned on top of the U-Pb-ablated pits to ensure that Hf isotope data were collected from the same growth domain as the U-Pb ages. The data were collected over multiple years. LA-ICP-MS analyses in 2009 used a New Wave UP193 HE excimer laser, and in 2013–2014, the analyses were performed using a Photon Machines Analyte G2 excimer laser. Fewer grains were analyzed for these samples in 2009; reduced cost and increased speed of analyses since that time have made analyses of greater numbers of grains more affordable and practical. In all analyses, the laser was connected to a Nu Plasma multicollector high-resolution ICP-MS, using methods described in Gehrels and Pecha (2014). For both U-Pb and Hf analyses, we used a beam diameter of 35 μm ; for a few very small zircon grains we used a beam diameter of 30 μm . Metadata for U-Pb-Th analyses are in Appendix 1.

Analytical results are displayed graphically on normalized-probability plots and Hf-evolution diagrams for visual comparison among zircon populations (Figures 8–10). Hf-evolution diagrams display epsilon Hf (ϵHf_t) values at the time of zircon crystallization (Figures 8–10). For U-Pb analyses, measured ion intensities from the Nu HR ICP-MS are imported into a data reduction program, “agecalc,” which reduces data, calculates ages, applies corrections and filters, and creates data tables, concordia diagrams, histograms, and normalized-probability plots (Gehrels & Pecha, 2014). For Hf analyses, a data-reduction program, “hfcalc,” reduces data, calculates Hf ratios, applies corrections, and creates data tables and Hf-evolution charts (Gehrels & Pecha, 2014). We rejected U-Pb analyses for which uncertainties are greater than 10%, discordance is greater than 20%, and reverse discordance is greater than 5%. For Hf analyses, we applied a 2 sigma filter (Gehrels & Pecha, 2014).

4. Results—DZ Analyses of the Harmony Formation

4.1. Harmony A

4.1.1. DZ U-Pb Age Spectra and Hf Isotope Ratios of Harmony A

The three Harmony A samples, LCC #1, #2, and #10, have similar U-Pb ages and Hf isotope ratios, which differ significantly from those of Harmony B (Figure 8). The Harmony A samples have a major U-Pb age peak at circa 1.0–1.2 Ga, which includes 45–62% of the analyses from each sample (Figure 8b). Smaller age populations

Figure 8. (a) Hf isotope data and (b) U-Pb ages for Harmony Formation. U-Pb dates were run for all sample grains; approximately one fourth of these grains were analyzed for Hf isotopes. The lower graph shows the normalized probability plots of U-Pb ages. The subdivision of Harmony A and B is indicated. The number of grains analyzed for U-Pb ages and Hf isotopes is shown. The upper graph shows epsilon Hf values at the time of zircon crystallization ($\epsilon\text{Hf}_{(t)}$) versus age for each sample. The average measurement uncertainty for all Hf analyses on this chart in plus-minus epsilon units is shown in the upper right at the 2σ level. Reference lines on the Hf plot are as follows: Depleted mantle (DM) is calculated using $^{176}\text{Hf}/^{177}\text{Hf} = 0.283225$; $^{176}\text{Lu}/^{177}\text{Hf} = 0.038513$ (Vervoort & Blichert-Toft, 1999); chondritic uniform reservoir, CHUR, is calculated using $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ (Bouvier, Vervoort, & Patchett, 2008).

from circa 1.2–1.5 Ga, circa 1.6–1.8 Ga, and circa 2.5–2.8 Ga each comprise up to 15% of the zircons (Figure 8b). The LCC#1 sample also has a small ($n = 7$) peak at circa 685 Ma, consisting of 4% of the total analyses (Figure 8b). The major age peak at circa 1.0–1.2 Ga yielded $\epsilon\text{Hf}_{(t)}$ ratios of +10 to −8 (Figure 8a). The age populations of 1.2–1.5 Ga yielded $\epsilon\text{Hf}_{(t)}$ ratios of +12 to −1, the age populations of 1.6–1.8 Ga yielded $\epsilon\text{Hf}_{(t)}$ ratios of +10 to −13, and those of 2.5–2.8 Ga yielded $\epsilon\text{Hf}_{(t)}$ ratios of +7 to −2 (Figure 8a).

4.1.2. Provenance of Harmony A

The U-Pb ages of the Harmony A detrital zircons are compatible with provenance in the central Laurentian craton. We can account for these age spectra as follows: the 1.0–1.2 Ga grains are consistent with provenance in the Grenville orogen (Bickford & Anderson, 1993; Hoffman, 1989; Van Schmus et al., 1993) (Figures 7 and 9b),

Figure 9. (a) Hf isotope data and (b) U-Pb ages for Harmony A samples and selected western Laurentian passive margin strata, showing the similarities between the U-Pb ages and Hf isotope analyses of the Harmony A and these passive margin strata. Data from the upper Neoproterozoic Mutual Formation and Caddy Canyon Quartzite are from Gehrels and Pecha (2014). Colored age bars that correspond to Laurentian basement terrane ages are superimposed over the U-Pb ages on the normalized probability plots. The ages are from references cited in Figure 3. Colored Hf isotope range areas that correspond to the same Laurentian terranes are shown on the (upper) Hf evolution diagram (Grenville: Bickford et al., 2010; Mueller et al., 2008; Granite-rhyolite province: Goodge & Vervoort, 2006; Mueller et al., 2008; Yavapai-Mazatzal: Holm et al., 2013; Archean: Rohr et al., 2008; Rohr et al., 2010; and Idaho: Gaschnig et al., 2013). Diagrams and symbols are as in Figure 9.

and the 1.3–1.5 Ga grains are similar to those with origin in the granite-rhyolite province of central Laurentia (Anderson & Morrison, 1992; Bickford et al., 1986; Bickford & Anderson, 1993; Hoffman, 1989; Van Schmus et al., 1993) (Figures 7 and 9b). The 1.6–1.8 Ga grains are consistent with provenance in the Yavapai-Mazatzal terranes and the 2.5–2.8 Ga grains are comparable with those originating in the Archean craton (Bickford et al., 1986; Hoffman, 1989; Ross, 1991; Van Schmus et al., 1993) (Figures 7 and 9b). The seven zircons from 673 to 716 Ma in sample LCC#1 sample have a potential source in Idaho. Several workers have reported zircon ages that fall within this range in igneous suites in Idaho: 684 ± 4 Ma and 685 ± 7 Ma (Link et al., 2017; Lund et al., 2003); 667 ± 5 Ma, 717 ± 4 Ma, and 709 ± 5 Ma (Fanning & Link, 2004); and 680–706 Ma (Durk et al., 2007; Link et al., 2017). The grains in the Harmony A arenites could have been derived from the igneous terranes or from recycled sediments originally derived from these terranes.

The Hf isotope ratios of the Harmony A samples are also similar to those in potential source terranes in central Laurentia. The 1.0–1.2 Ga grains have juvenile to intermediate values ($\epsilon_{\text{Hf}(t)} + 10$ to -8), similar to those of the

Figure 10. (a) Hf isotope data and (b) U-Pb ages for Harmony B samples and select Laurentian passive margin strata, showing the similarities between the U-Pb ages and Hf isotope analyses of the Harmony B and these passive margin strata. Colored age bars that correspond to Peace River Arch region and Swift Current anorogenic province basement terrane ages are superimposed over the U-Pb ages on the normalized probability plots. Data from the Horsethief Creek Group and the Hamill Group are from Gehrels and Pecha (2014). U-Pb analyses of the Addy Quartzite are from Linde et al. (2014b). Hf-isotope analyses of the Addy Quartzite are Linde's unpublished work. Diagrams and symbols are as in Figure .

Grenville orogen (Bickford et al., 2010; Mueller et al., 2008), and the 1.4–1.5 Ga grains have juvenile to moderately juvenile values ($\epsilon\text{Hf}_{(t)} + 10$ to $+2.5$), compatible with the granitoids of the mid-Laurentian craton (Goodge & Vervoort, 2006) (Figures 7 and 9a). The 1.6–1.8 Ga grains have juvenile to intermediate values ($\epsilon\text{Hf}_{(t)} + 10$ to -6), similar to the Yavapai orogenic terrane (Holm et al., 2013). The 2.5–2.8 Ga grains have moderately juvenile to intermediate values ($\epsilon\text{Hf}_{(t)} + 7$ to -2), compatible with those in northern Greenland and Arctic Canada of the northeast Canadian shield (Rohr et al., 2008, 2010) (Figures 7 and 9a). The 673–716 Ma grains have intermediate to evolved values ($\epsilon\text{Hf}_{(t)} + 1.9$ to -5.3), which is within the range of 630–730 Ma zircons in the Idaho batholith, interpreted as inherited from the Windermere Supergroup volcanics (Gaschnig et al., 2013).

Coeval passive margin sedimentary units of western Laurentia have U-Pb ages and Hf isotope ratios similar to those of Harmony A (Figure 9). We compared our Harmony A data to those of western Laurentian passive margin units for which both DZ U-Pb ages and Hf isotope ratios are available. The U-Pb ages and Hf isotope ratios of the upper Neoproterozoic-Lower Cambrian Mutual Formation and Caddy Canyon Quartzite are similar to those of the Harmony A (Figure 10). The Mutual Formation and Caddy Canyon Quartzite are interpreted as originating in central Laurentia prior to the uplift of the Transcontinental Arch (Gehrels & Pecha, 2014;

Figure 11. Map of western Canada showing the Cordilleran accreted terranes, the Cordilleran passive margin, and the basement provinces of the Canadian Shield. The Peace River Arch and Rimbey arc regions are outlined by red dashed lines. The Swift Current anorogenic province (SCAP) is highlighted in yellow. The location of the Hamill Group, Horsethief Creek Group, and Addy Quartzite samples are shown. Sample locations are shown by colored diamonds; Horsethief Creek Group (HC) and Hamill Group (HA) are from Gehrels and Pecha (2014); Addy Quartzite (AD) is from Linde et al. (2014b). Map is after Gehrels and Ross (1998); the basement provinces are compiled from Collerson et al. (1988), Hoffman (1989), Ross (1991), Villeneuve et al. (1993), and Peterson et al. (2015).

Linde et al., 2014a; Yonkee et al., 2014) (Figure 9). In summary, we conclude that the sediments comprising the Harmony A, the Mutual Formation, and the Caddy Canyon Quartzite were derived from sources in the central Laurentian craton prior to the uplift of the Transcontinental Arch.

4.2. Harmony B

4.2.1. DZ U-Pb Age Spectra and Hf Isotope Ratios of Harmony B

The seven Harmony B samples, LCC #3, #4, and #9, and Kluncy, Gough's, Harmony, and Elbow Canyons, revealed a major U-Pb age peak at circa 1.7–1.9 Ga, comprising 60–65% of the analyses (Figure 8b). The samples also yield smaller age populations at circa 2.4–2.8 Ga, representative of 20% of the sample (Figure 8b). The circa 1.7–1.9 Ga grains yielded $\epsilon_{\text{Hf}(t)}$ ratios of +1 to –20 (Figure 8a). The age populations of 2.4–2.7 Ga yielded $\epsilon_{\text{Hf}(t)}$ ratios of +5 to –5 and the age populations of 2.7–2.8 Ga yielded $\epsilon_{\text{Hf}(t)}$ ratios of +7 to –9 (Figure 8a).

4.2.2. Provenance of Harmony B

The Harmony B arenites are composed of detrital zircons that have potential sources in igneous terranes of eastern Alberta and western Saskatchewan. The Harmony B primary age populations of 1.75–1.9 Ga are within the zircon ages of two igneous provinces: the 1.73–1.77 Ga Swift Current anorogenic province of the Kivalliq Igneous Suite of western Saskatchewan (Collerson et al., 1988; Peterson et al., 2015) and the 1.79–1.86 Ga Rimbey arc of eastern Alberta (Gehrels & Ross, 1998; Villeneuve et al., 1993) (Figures 10b and 11). The Harmony B age populations of ca. 2.4–2.7 Ga and 2.7–2.8 Ga are similar to those of Archean terranes in the same region (Figures 10b and 11). There are no Hf isotope analyses for the igneous provinces of eastern Alberta and western Saskatchewan; therefore, no direct comparison between the Hf isotope ratios of the

Harmony B samples and these potential source terranes is possible. We compared our Harmony B data with analyses of western Laurentian passive margin units for which both U-Pb ages and Hf isotopes are available. The Harmony B grains have a range of U-Pb ages and $\varepsilon\text{Hf}_{(t)}$ values similar to those of the Horsethief Creek and Hamill Groups and the Addy Quartzite, passive margin units interpreted to have originated in eastern Alberta-western Saskatchewan (Gehrels & Pecha, 2014; Linde et al., 2013) (Figure 11).

Other potential source terranes for the Harmony B are the Belt Supergroup, the Trans-Hudson orogen, and the Yavapai-Mazatzal terrane. Another possibility is that the Harmony Formation shared a source with the remaining strata of the Roberts Mountains allochthon. The Belt Supergroup and Harmony B sediments share many detrital zircon age populations, but approximately 20% of the Belt Supergroup zircons are younger than the youngest (circa 1.75 Ga) Harmony B zircons (Link et al., 2013). If the Belt Supergroup was the source of the Harmony B, some of these 1.4–1.7 Ga zircons should occur in the Harmony B, but they do not. Harmony B sediments also share many age populations with the Trans-Hudson orogen. However, the 1.8–2.0 Ga Trans-Hudson orogen (Whitmeyer & Karlstrom, 2007) could not have supplied the 1.75–1.8 Ga zircons which comprise nearly 35% of the Harmony B. The Yavapai-Mazatzal terrane also shares Paleoproterozoic age populations with the Harmony B. If the Yavapai-Mazatzal terrane was the source of the Harmony B, the Harmony B should contain grains of the age of the other major age provinces in central Laurentia, including the Grenville orogenic terrane and the granite-rhyolite magmatic province. These latter are located within the Yavapai-Mazatzal terrane (Figure 7). It is not likely that river systems would flow across the Yavapai-Mazatzal terrane and transport only grains of Paleoproterozoic age, without also entraining younger Grenville or granite-rhyolite province grains. Therefore, it is unlikely that the Harmony B originated in the Belt Supergroup, the Trans-Hudson orogen, or the Yavapai-Mazatzal terrane.

The RMA strata and the Harmony B share some age populations, but significant dissimilarities preclude a common source. The RMA strata have major age populations from 1.80 to 1.95 Ga, with no grains less than circa 1.8 Ga (Linde et al., 2016); however, the Harmony B has a significant age population less than circa 1.8 Ga. Hence, the Harmony B sediments were not derived from the same source as the RMA strata, or from the RMA itself.

The detrital zircon U-Pb age spectra and Hf isotope ratios of the Harmony B are consistent with origin in eastern Alberta and western Saskatchewan (Figures 10 and 11). This region was exposed throughout Cambrian time and submerged from Ordovician through Mississippian time (Cant, 1988; Cant & O'Connell, 1988; Kent, 1994). The provenance of the sediments of the Harmony B, the Horsethief Creek Group, the Hamill Group, and the Addy Quartzite therefore appears to be the Rimbey arc of eastern Alberta and the Swift Current anorogenic province of western Saskatchewan (Figures 10 and 11).

4.3. Considerations for the Age of the Harmony Formation

The fossil content of the Harmony B and our direct correlation of the U-Pb ages and Hf isotope data of Harmony A samples with other fossil-controlled sequences corroborate the Cambrian age of the Harmony Formation and make a Devonian or later age unlikely. (1) The Rimbey arc and Swift Current anorogenic province—the most likely sources of the zircons in the Harmony B—were submerged, and the site of shallow shelf carbonate deposition, during the Devonian (Cant, 1988; Cant & O'Connell, 1988; Kent, 1994) (Figures 12c–12e). This is incompatible with a Mississippian or Devonian depositional age for the Harmony. (2) Although no fossils occur in the Harmony A, the similarity of Harmony A U-Pb detrital zircon age spectra and Hf isotope ratios with those of late Neoproterozoic–Early Cambrian passive margin units, such as the Mutual Formation and the Caddy Canyon Quartzite, strongly suggests deposition at the same time as these units and before the uplift of the Transcontinental Arch. Both of these observations corroborate an Early Cambrian, rather than Devonian, age for the Harmony.

5. Discussion—Paleogeographic and Tectonic Implications of Harmony Formation Sources

5.1. Different Provenance for “Harmony A” and “Harmony B”

Harmony A and B are two distinct units, with different provenance. We interpret that the sediments of Harmony A originated in the central Laurentian craton, before the uplift of the Transcontinental Arch, and were deposited as turbidites in a basin off the western Laurentian margin in late Neoproterozoic or Early

Figure 12. Paleogeographic maps of Laurentia for Early Cambrian through Mississippian time (Blakey, 2013). White lines show the approximate position of the paleoequator. Blue wavy lines show approximate sediment transport pathways of units discussed. The Transcontinental Arch (Sloss, 1988) and Peace River Arch (Ross, 1991) are superimposed. (a) Early Cambrian time. The Harmony A is derived from the central Laurentia, before the uplift of the Transcontinental Arch. The Horsethief Creek Group and Hamill Group are deposited near southeast British Columbia. (b) Late Cambrian Time. The Transcontinental Arch has been uplifted. The Harmony B is shed from the Rimbey arc of eastern Alberta and the Swift Current Anorogenic Province of western Saskatchewan. (c) Middle Devonian time. An arc has moved westward around the north edge of Laurentia, from northern Baltica to the western margin of northern Laurentia, and a sinistral transpressional fault system has developed along the western margin of Laurentia. The Harmony A and B and RMA strata are tectonically transported south along the margin by this fault system. (d) Late Devonian time. Shortening and development of an accretionary prism has initiated along much of the western margin of Laurentia, emplacing the Harmony Formation and RMA strata onto the craton. (e) Early Mississippian time. The Antler orogeny has uplifted the accretionary prism of the Harmony Formation and RMA strata into a highland on the western Laurentian margin.

Cambrian time (Figure 12a). In contrast, the sediments of Harmony B probably originated in eastern Alberta-western Saskatchewan and were deposited as turbidites in a basin off the northwestern Laurentian margin in Late Cambrian time (Figure 12b). It is likely that passive-margin arenites such as the Horsethief Creek Group, Hamill Group, and Addy Quartzite in southern British Columbia and northeastern Washington also originated in eastern Alberta-western Saskatchewan (Figures 12a and 12b). These units are the nearshore and shallow shelf equivalents of the deeper marine Harmony B (Gehrels & Ross, 1998; Lindsey & Gaylord, 1992). From Ordovician through Devonian time, other RMA strata were deposited in the region (Linde et al., 2016).

The *nature* of the contact between Harmony A and Harmony B is unclear. The new DZ data presented here substantiate the relative ages—Harmony A is pre-Transcontinental Arch, that is, no younger than Early Cambrian, while Harmony B is Late Cambrian, based on trilobites. The initial workers interpreted the contact between the Harmony A and B in Little Cottonwood Canyon as sharp but concordant and proposed that the two petrofacies recorded two successive submarine fan deposits (Dickinson & Gehrels, 2000). Fieldwork for this study indicated that the contact can be located within a few meters and traced in the field.

Sedimentary structures show that both sections are upright, with Harmony B on top of Harmony A. However, the contact is not well enough exposed to determine whether it is depositional or faulted.

Harmony A and B cannot be redefined as different formations at this time, because they cannot always be distinguished in the field and are therefore not mappable separately. Some exposures, especially the coarser grained feldspathic beds, are definitive. However, quartz arenite and carbonate beds occur in both A and B and are not uniquely identifiable in the field except where feldspathic arenite is present.

5.2. Tectonic Models for Origin of Harmony Formation, Reexamined

Some of the proposed models for the origin and transport of the Harmony Formation can now be excluded as reasonable explanations. Option I proposed that the RMA and Harmony Formation were the offshore, deep marine, equivalents of coeval autochthonous passive margin strata (e.g., Burchfiel & Davis, 1972; Poole et al., 1992; Roberts et al., 1958) (Figure 3, Option I). The new analyses presented here document that the source of the Harmony A sediments is the central Laurentian craton, but the source of the Harmony B sediments is not. Furthermore, it is now known that the rest of the RMA strata, exclusive of the lower Vinini Formation, also do not share DZ age populations or a source with coeval sediments of the passive margin (Linde et al., 2016). In short, the RMA and Harmony Formation were *not* the offshore, deep marine equivalents of coeval autochthonous passive margin strata.

Option II suggested that the RMA and Harmony Formation were tectonically transported to the western Laurentian margin along with known exotic terranes such as the Alexander terrane (Colpron & Nelson, 2009; Wright & Wyld, 2006) (Figure 3, Options IIa and IIb). However, the new analyses show that the Harmony Formation has age populations and Hf isotope ratios similar to western Laurentian sources. The Harmony B also contains trilobite faunas of North American affinity, precluding an exotic origin (Hotz & Willden, 1964).

Option III invoked a source in western Laurentia, in the Peace River Arch for the RMA, and the Salmon River Arch for the Harmony Formation (Gehrels, Dickinson, Riley, et al., 2000; Suczek, 1977) (Figure 3, Option III). The Salmon River Arch can now be ruled out, because it lacks potential source regions with appropriate DZ age populations. More recent work has shown that the Salmon River Arch is an uplifted belt of Mesoproterozoic Yellowjacket Formation intruded by 650 Ma plutons (Lewis et al., 2012; Lund et al., 2010). However, the Peace River Arch region is a good candidate as a source for RMA sediments, and the Rimbey arc—Swift Current anorogenic province to the southeast of the Peace River Arch is the likely source for Harmony B sediments.

5.3. Implications for the Paleozoic Evolution of the Western Laurentian Margin

The Harmony Formation's provenance can be used to refine tectonic models for the Antler orogeny. In particular, the distinctive feldspathic arenite of Harmony B provides a useful kinematic marker. Because both the Harmony Formation and the geologic record of the Antler orogeny are confined to Nevada and Idaho, the conclusions drawn from them applies to the "American" portion of the western Laurentian margin. It supplements existing literature on tectonic evolution in Canada and the Arctic, extending it southward geographically.

5.3.1. Initial Sinistral Motion on the Active Margin

As the U.S. portion of western Laurentia evolved from a passive to an active margin, southward propagating sinistral slip predated significant shortening as evidenced by the DZ ages of the RMA, including the Harmony Formation. The DZ data are consistent with two interpretations: (1) Rocks of the Harmony B were sourced in eastern Alberta-western Saskatchewan. (2) The Ordovician-Devonian arenites in the RMA (with the exception of the lower Vinini Formation, which originated in central Laurentia), were sourced in the Peace River arch region (Linde et al., 2016). These all had to be transported south along the Laurentian margin to reach their current positions (Figure 12d). Our interpretation is that *Tectonic* transport is the most likely scenario, considering the sedimentary characteristics of RMA rocks, including the Harmony Formation. In comparison to coeval sands in the autochthonous continental margin section, the less mature texture of RMA arenites rules out longshore transport, which would entail significant reworking (Linde et al., 2016). The textural immaturity of the Harmony B further supports the tectonic transport interpretation, because it includes grains (e.g., mica and feldspar) that would not survive extensive reworking or sedimentary transport.

5.3.2. Age of Initiation of the Active Margin

The initiation of our proposed sinistral slip regime along the western Laurentian margin is constrained to Late Devonian time. The RMA contains rocks as young as the Middle Devonian Slaven Chert (Gilluly & Gates, 1965), so it could not have been displaced southward until after this time. The initiation of the Antler foreland basin in central Nevada, recorded by the arrival of clastic sediments from the west, occurred in latest Devonian–Early Mississippian time (Poole, 1974; Trexler et al., 2003).

5.3.3. Shortening Inboard of the Continental Margin

The record of the Antler orogeny inboard of the active plate margin is as follows: (1) formation of a foreland basin containing generally west derived sediments, followed by (2) folds and thrust faults that document generally east directed shortening. Although the Antler foreland basin deposits record a newly uplifted sediment source to the west by latest Devonian time, *faults* that accommodate shortening across the continental margin have not been documented before mid-Mississippian time in what is now north central Nevada. East vergent folds and east directed imbricate thrust faults duplicate the Antler foreland basin section in the Adobe Range and the Snake Mountains (McFarlane, 2001; Trexler et al., 2003). These structures predate the Late Mississippian (Chesterian) rocks that overlie them along an angular unconformity. The distribution and age of the known thrust faults may be an artifact of preservation. Much of the area west of the trace of the RMT is covered by the Golconda allochthon or by Cenozoic volcanic rocks (Figure 2). In addition, Pennsylvanian and Permian structures are well documented in north central Nevada (e.g., Trexler et al., 2004) and may have obliterated Antler-age structures. Finally, as noted above, no present exposures of the mapped Roberts Mountains thrust can be constrained to Late Devonian–Early Mississippian in age, and many have unequivocally been active (and possibly reactivated?) since that time.

Local imbrication of the Harmony Formation with the Valmy Formation of the RMA (Ferguson et al., 1951; Gilluly, 1967; Hotz & Willden, 1964; Madrid, 1987; Roberts, 1964) suggests that the Harmony was structurally stacked with the RMA prior to emplacement onto the continental margin. Most of the sinistral displacement, and possibly much of the structural stacking also, appears to have predated the emplacement of the RMA into its present place on the continental margin.

6. Conclusions

The new DZ U–Pb age spectra and Hf isotope analyses presented here support a previous suggestion (Gehrels, Dickinson, Riley, et al., 2000) that rocks of the Harmony Formation were derived from western Laurentia. These data *rule out* several previously proposed models for the tectonic derivation of the Harmony Formation, and hence for the associated tectonic evolution of the western Laurentian margin. They record development of the Nevada part of the margin in unprecedented detail.

Our new data also confirm that the rocks mapped as Harmony Formation include two different sedimentary units (Gehrels, Dickinson, Riley, et al., 2000). They establish that the sediments of Harmony A originated in the central Laurentian craton, before the uplift of the Transcontinental Arch, and were deposited as turbidites in a basin off the western Laurentian margin in late Neoproterozoic or Early Cambrian time (Figure 12a). In contrast, the sediments of Harmony B originated in eastern Alberta–western Saskatchewan and were deposited as turbidites in a basin off the northwestern Laurentian margin in Late Cambrian time (Figure 12b). The presence of both units structurally high in the imbricated RMA requires complex structural stacking during transport and/or emplacement.

The two-part provenance of the two-member Harmony Formation is analogous to that of the Ordovician–Devonian rocks of the RMA. DZ analyses have shown that the lower Vinini Formation sediments were derived from the central Laurentian craton after the uplift of the Transcontinental Arch (Linde et al., 2016). The arenites in the rest of the RMA have a source in the Peace River Arch region of eastern British Columbia–western Alberta (Linde et al., 2016).

The rocks of the Harmony Formation and RMA strata were *tectonically* transported south along the Laurentian margin to reach their current position. The textural and compositional immaturity of Harmony B precludes sedimentary transport by longshore drift. Similarly, RMA strata deposited in northwestern Laurentia were subsequently tectonically transported south along the western Laurentian margin and emplaced onto the craton during the Antler orogeny (Linde et al., 2016).

The “American” western margin of Laurentia was therefore one of sequential sinistral slip and transpression in Late Devonian-Early Mississippian time. The present locations of Harmony and other RMA rocks relative to their source areas record significant sinistral displacement along the Laurentian margin prior to emplacement. The sinistral motion started no earlier than mid-Devonian time. Both timing and kinematics are consistent with the continued propagation of an active plate boundary southward along the Laurentian margin in Devonian time, as proposed by Colpron and Nelson (2009).

Many of the “anomalous” characteristics of the Antler orogeny are not so anomalous when considered as part of a transpressive margin system. The absence of Antler-aged plutons, volcanic rocks, and volcanoclastic sediments in the foreland basin all indicate that there was not enough subduction associated with the Antler orogeny to generate magmatism. The absence of high-grade metamorphic rocks is consistent with crustal-scale strike-slip motion, and with relatively minor crustal-scale vertical motion associated with the plate boundary. The anomalously thin foreland basin section suggests comparatively minor crustal loading by the advancing allochthon. In conclusion, the Antler orogeny may be better explained by sinistral transpression than by the traditional margin-perpendicular-shortening model.

Appendix A: Metadata for U-Th-Pb Analyses

Appendix 1 provides the analytical parameters for all analyses described in this paper.

<i>Laboratory and sample preparation</i>	
Laboratory name	Arizona LaserChron Center, University of Arizona
Sample type/mineral	Detrital zircons
Sample preparation	Conventional mineral separation, 1 inch epoxy, surface polished to 9 μm
Imaging	CL, Hitachi 3000N, Gatan Chroma CL
<i>Laser ablation system</i>	
Make, Model & type	Photon Machines Analyte G2
Ablation cell & volume	HelEx cell, dual volume
Laser wavelength (nm)	193 nm
Pulse width (ns)	<4 ns
Fluence ($\text{J}\cdot\text{cm}^{-2}$)	~5 $\text{J}\cdot\text{cm}^{-2}$
Repetition rate (Hz)	7 Hz
Ablation duration (secs)	~15 secs
Ablation pit depth / ablation rate	12 μm pit depth, measured using an optical microscope and BSE/ SEM images
Spot diameter (μm) nominal/actual	35 μm (30 μm for a few small grains)
Sampling mode	Static spot ablation
Carrier gas	100% He in the cell, Ar make-up gas combined using a Y-piece 80% along the sample transport line to the torch.
Cell carrier gas flow (l/min)	MFC1 = 0.1 LPM (inner cell) MFC2 = 0.3 LPM (outer cell)
<i>ICP-MS Instrument</i>	
Make, Model & type	Nu Instruments, Nu Plasma HR #068, MC-ICP-MS
Sample introduction	Ablation aerosol
RF power (W)	1300W
Make-up gas flow	Argon controlled by MFC set at 1/02 LPM
Detection system	mixed Faraday-multiple ion counting array
Masses measured	202, 204, 206, 207, 208, 232, 238
Integration time per peak	1 sec for each isotope
Total integration time per output datapoint (secs)	~1.02 secs
‘Sensitivity’ as useful yield	0.6% U (#ions detected/#atoms sampled)*100
IC Dead time (ns)	12.0 ns (IC0)
<i>Processing</i>	
Gas blank	15 second on-peak zero subtracted
Calibration strategy	Sri Lanka zircon (UofA crystal #2) as primary standard
Reference Material info	Calibration from Gehrels, Valencia, and Ruiz, (2008)
Data processing package used / Correction for LIEF	Agecalc, described by Gehrels et al. (2008) and Gehrels and Pecha (2014)
Mass discrimination	207Pb/206Pb and 206Pb/238U normalised to reference material

Table . (continued)

Laboratory and sample preparation	
Common-Pb correction, composition and uncertainty	204-based common-Pb correction applied to the data. Composition assumed from Stacey and Kramers (1975).
Uncertainty level & propagation	Uncertainties of isotope ratios and ages are reported in DR Table 1 at 1-sigma. Age uncertainties are reported with only internal (measurement) uncertainties and also with internal and external (systematic) uncertainties combined by quadratic addition.
Quality control / Validation	SL-2 analyses are included in DR Table 1.

Acknowledgments

We would like to acknowledge the essential role that Bill Dickinson played in all of our Harmony Formation studies. He had thought much about the difficult puzzle of the Harmony over several decades and was always enthusiastic about hearing new ideas and sharing his perspective. Bill was an incomparable mentor and model for all of us, and his loss to the entire geoscience community is profound. We also acknowledge the contributions and guidance of Jim Trexler. As Linde's doctoral advisor, Jim was always open to new ideas and methods. He was a source of sound guidance and support, keeping us grounded in the reality of the rocks in the field. For both of these gentlemen, *requiem aeternam dona ei, Domine, et requiescat in pace*. Additionally, we thank the researchers and staff at the Arizona LaserChron center, especially Mark Pecha and Nicole Giesler and volunteer laboratory assistants Dan Sturmer of Shell Oil Company and University of Nevada, Reno graduate students Connor Newman and Kyle Basler-Reeder for critical help in Tucson. This research was funded under Arizona LaserChron National Science Foundation (NSF) grant EAR-1338583 to Gehrels. The original collecting effort for these samples was supported by NSF grants EAR-9116000 and EAR-9416933 to Gehrels and Dickinson. LaserChron analyses of some samples was supported by NSF grant EAR-0510915 to Trexler and Cashman. Linde also thanks the following organizations for generous scholarship support of her graduate work: The Raytheon Corporation Student Veterans Scholarship, the Rocky Mountain Association of Geologists Veterans Memorial Scholarship, the Nevada Petroleum and Geothermal Society, the Graduate Student Association of the University of Nevada Reno, and the Cordilleran Section of the Geological Society of America. Data supporting these analyses and conclusions can be found in the supporting information.

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