The Permian Monos Formation: Stratigraphic and detrital zircon evidence for Permian Cordilleran arc development along the southwestern margin of Laurentia (northwestern Sonora, Mexico)

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ABSTRACT

The southwestern margin of Laurentia transitioned from a left-lateral transform margin to a convergent margin by middle Permian time, which initiated the development of a subduction zone and subsequent Cordilleran arc along western Laurentia. The displaced Caborca block was translated several hundred kilometers from southern California, USA, to modern Sonora, Mexico, beginning in Pennsylvanian time (ca. 305 Ma). The Monos Formation, a ~600-m-thick assemblage of mixed bioclastic and volcaniclastic units exposed in northwestern Sonora, provides lithostratigraphic, petrographic, and geochronologic evidence for magmatic arc development associated with subduction by middle Permian time (ca. 275 Ma).

The Monos Formation was deposited in a forearc basin adjacent to a magmatic arc forming along the southwestern Laurentian margin. Detrital zircon U-Pb geochronology suggests that Permian volcanic centers were the primary source for the Monos Formation. These grains mixed with far-traveled zircons from both Laurentia and Gondwana. Zircon age spectra in the Monos Formation are dominated by a ca. 274 Ma population that makes up ~65% of all analyzed grains. The remaining 35% of grains range from 3.3 Ga to 0.3 Ma, similar to age spectra from Permian strata deposited in the Paleozoic sequences in the western continental interior. An abundance of Paleozoic through early Neoproterozoic ages suggests that marginal Gondwanan sources from Mexico and Central America also supplied material to the basin.

The Monos Formation was deposited within tropical to subtropical latitudes, yet faunal assemblages are biosiliceous and heterotrophic. The lack of photozoan assemblages suggests that cold-water coastal upwelling combined with sedimentation from the Cordilleran arc and Laurentian continent promoted conditions more suitable for fauna resilient to biogeochemically stressed environments.

We propose that transform faulting and displacement of the Caborca block ceased by middle Permian time and a subduction zone developed along the southwestern margin of Laurentia as early as early Permian time. The Monos basin developed along the leading edge of the continent as a magmatic arc developed, and facies indicate a consistent shoaling trend over the span of deposition.

INTRODUCTION

Subduction initiation and subsequent volcanic arc development are poorly documented and understood in continental settings. Early arc products are commonly overprinted by later intrusions, and deformation may mask the earliest development of sedimentary basins. The Paleozoic southwestern margin of Laurentia transitioned from a passive margin to a transform margin by Pennsylvanian time (Goetz and Dickerson, 1985; Walker, 1988; Dickinson and Lawton, 2001; Saleeby, 2011). By late Paleozoic to earliest Mesozoic time, this margin was restructured again to a convergent system (Dickinson and Lawton, 2001; Barth and Wooden, 2006; Riggs et al., 2010; Arvizu and Iriondo, 2015; Cecil et al., 2018). Reconstructions of this transition have commonly projected subduction initiation at ca. 255 Ma (Dickinson, 2000; Barth and Wooden, 2006; Saleeby, 2011); however, more recent zircon geochronology and whole-rock geochemistry of related plutons (Riggs et al., 2010; Arvizu and Iriondo, 2015; Cecil et al., 2018) and hypabyssal-volcanic clasts from conglomerate in northwestern Sonora, Mexico (Linder, 2013), indicate that subduction-related calc-alkaline magmatism began as early as ca. 275 Ma. The arc was well established by early Mesozoic time and is recorded by arc plutons in the Mojave Desert and Sierra Nevada, USA (Barth and Wooden, 2006; Barth et al., 2012; Cecil et al., 2018), and in northern Sonora (Arvizu et al., 2009; Arvizu and Iriondo, 2015). Late Paleozoic forearc sedimentary successions in southern California, USA, have been inferred to be the earliest volcaniclastic record of Permian Cordilleran magmatism (Carr et al., 1984, 1997; Martin and Walker, 1995; Rains et al., 2012).

The middle Permian Monos Formation (Cooper et al., 1953) is exposed in the northern Sierra del Álamo west of Caborca, Sonora (Fig. 1A), in a small group of hills called the Cerros Monos; this part of Sonora is considered to be part of the Caborca terrane (González-León, 1997; Dickinson and Lawton, 2001; González-León et al., 2009). Preliminary U-Pb dating of detrital zircon from the Monos Formation...
yielded predominantly middle Permian ages and numerous early Permian antecrysts (Riggs et al., 2014; Dobbs et al., 2016), suggesting that the Monos Formation records the onset of Cordilleran arc development as subduction became established. Lithofacies, sandstone petrography, and detrital zircon geochronology help to reconstruct the tectonic history, sedimentary provenance, and basin architecture of the Monos Formation. The Monos Formation was deposited in a shallow-marine basin adjacent to the Permian Cordilleran magmatic arc along southwestern Laurentia. The primary source for the non-biogenic components of the Monos Formation was the volcanic arc, along with minor sourcing from the Laurentian craton and marginal Gondwanan terranes. Permian Cordilleran arc development and a later magmatic hiatus and exhumation and erosion of the arc edifice are supported by Permian plate-motion reconstructions of Panthalassa relative to a fixed North American reference frame (Domeier and Torsvik, 2014). Interpretation of the Monos Formation, therefore, provides insight into the sedimentary response to the inception of arc magmatism and a broader assessment of upper-plate processes during the earliest phases of arc magmatism.

**Geologic Background**

A northwest-striking left-lateral transform plate boundary, termed the California-Coahuila transform by Dickinson (2000), developed along the southwestern margin of Laurentia by Pennsylvanian–Permian time (Walker, 1988; Dickinson and Lawton, 2001) and was inferred to have reached from east-central California to the Ouachita-Marathon collisional belt (north-central Mexico) (Dickinson and Lawton, 2001). This structure is hypothesized to have rearranged southwestern Laurentia by displacing blocks of the North American continent, such as the Caborca block, several hundred kilometers southeast via left-lateral slip (Fig. 1B; Burchfiel and Davis, 1972; Stewart et al., 1984, 1990; Dickinson, 2000; Dickinson and Lawton 2001; Saleeby and Dunne, 2015). Estimates of offset range from 400 to 900 km based on a proposed correlation of miogeoclinal strata in the Caborca block with those in Death Valley or the Mojave Desert in California (Dickinson, 2000; Stevens et al., 2005; Stewart, 2005).

**Figure 1.** (A) Inset map of the state of Sonora, Mexico, and location of the Sierra del Álamo, northwestern Sonora (red star). (B) Regional tectonic map of the western United States and northwestern Mexico, modified from Dickinson (2000), Dickinson and Lawton (2001), Karlstrom et al. (2003), Dickinson (2004), and Whitmeyer and Karlstrom (2007). Approximate plate motion of Panthalassa with Laurentia fixed (ca. 270 Ma) is from Domeier and Torsvik (2014). Location of the paleo-equator is from Scotese and Langford (1995). CCT—California-Coahuila transform fault, from Dickinson and Lawton (2001).
The initiation of subduction and subsequent development of a magmatic arc began by Permian–Triassic time (ca. 275–235 Ma; Miller et al., 1995; Barth et al., 1997; Barth and Wooden, 2006; Arvizu et al., 2009; Riggs et al., 2010, 2014; Arvizu and Iriondo, 2015; Cecil et al., 2018; Fig. 1B). Middle Permian (ca. 270–260 Ma) volcaniclastic strata of Holland Camp, a part of the El Paso terrane in the northern Mojave Desert, are interpreted as the earliest detrital record of volcanic arc development in the Mojave region (Burchfiel and Davis, 1972; Rains et al., 2012; McDonald, 2016). Paleozoic and Mesozoic rocks exposed in the Cerros Monos (Sierra del Álamo, Sonora) include the Permian Monos Formation and overlying Permian–Jurassic El Antimoni Group (Fig. 2). The El Antimoni Group, a ~3.4-km-thick succession of shallow-marine to terrestrial units, includes Triassic conglomerate containing Permian–Triassic hypabyssal-volcanic clasts and abundant Permo-Triassic zircons, suggesting proximity and continued active volcanism in the Cordilleran arc throughout Triassic time (González-León et al., 2005; Linder, 2013; Riggs et al., 2014).

Southern Laurentia straddled or was just north of the equator in Permian time, placing the Caborca block in tropical to subtropical latitudes, although the precise location is still a matter of debate (Scotese and Langford, 1995; Domeier and Torsvik, 2014; Tomezoli et al., 2018; Muttoni and Kent, 2019; Correia and Murphy, 2020; Kent and Muttoni, 2020; Fig. 1B). Isotopic data suggest that climate gradually warmed throughout the Permian (Kiehl and Shields, 2005). Dry northeasterly trade winds from the Laurentian midcontinent likely reached the region (Parrish and Peterson, 1988).

The Monos Formation

The base of the Monos Formation (Fig. 2) is not exposed but is presumed to be in unconformable contact with late Paleozoic strata exposed in western and central Sonora (Stanley and González-León, 1995; González-León, 1997). The Monos Formation is folded into a south-plunging anticline that is faulted and intruded by microdiorite dikes. The Permian–Jurassic El Antimoni Group disconformably overlies the Monos Formation in the southern part of the Cerros Monos and is in fault contact with the formation in the southeast (González-León et al., 2005; Fig. 2).

The lower part of the Monos Formation consists of non-fossiliferous volcaniclastic sandstone interbedded with siliceous mudstone. The remainder of the formation is predominantly carbonate rocks that contain abundant types of brachiopods, pelecypods, scaphopods, corals, fusulinids, cephalopods, sponges, and gastropods. Biostratigraphic age constraint from the uppermost fossiliferous strata of the Monos Formation was determined via the identification of fusulinid (Parafusulina antimonioensis Dunbar), ammonite (Waagenoceras dieneri Boese), and brachiopod (Waagenoconcha montpeliereensis (Girty)) index fossils indicating a Wordian (middle Permian, 268.8–265.1 Ma; Cohen et al., 2013) age (Cooper and Arellano, 1946; Cooper et al., 1953, 1965; Buitrón-Sánchez et al., 2007). Giant Parafusulina antimonioensis Dunbar is physically similar to fusulinids found in northern California (Parafusulina californica [von Staff]), which led some workers to argue that Parafusulina is a guide fossil for North American exotic terranes that were displaced during late Paleozoic and Mesozoic times (Ross and Ross, 1983; Stewart et al., 1990; González-León and Stanley, 1993; Calmus et al., 1997; González-León, 1997; Vachard et al., 2000). Others, however, have noted that Parafusulina antimonioensis Dunbar is also reported proximal to the Caborca block’s current position adjacent to the American craton (e.g., Glass Mountains, Texas, USA; Kobayashi, 1957; Ross, 1963), suggesting that translation of the Caborca block...
ceased by the latest Paleozoic (Buitrón-Sánchez et al., 2007). Moreover, Guadalupian (273–259 Ma) conodonts from fossiliferous strata of the Monos Formation more strongly resemble those from the Guadalupian Delaware Basin in western Texas and New Mexico than those from Permian basins from distal terranes (Lara-Peña and Navas-Parejo, 2018a, 2018b). This suggests that the translation of the Caborca block occurred during Pennsylvanian–Permian time and had ceased by the middle to late Permian. This reconstruction implies that the Monos Formation is proximal to its original depositional location and records a period of time postdating the relocation of the Caborca block (Stone and Stevens, 1988; Buitrón-Sánchez et al., 2007; Lawton et al., 2017).

**METHODS**

**Field Methods**

The Monos Formation has been informally divided into four units based on assemblages of lithofacies (Dobbs et al., 2016; Dobbs, 2017; Fig. 3). Three stratigraphic sections (sections 1–3) were measured at decimeter scale in the northern and northeastern portions of the mapping area, where the Monos Formation is well exposed and has minimal structural complications, intrusive contacts, or diagenetic alteration (Fig. 2). Twenty-eight (28) sandstone horizons were sampled throughout the Monos Formation for detrital zircon analysis. Of those, only six samples yielded zircons. The low zircon yield is attributed to the predominance of carbonate and silty lithologies in the Monos Formation. Units 1 and 2 were sampled at a higher density due to their greater abundance of siliciclastic material and lack of biostratigraphic constraint. Approximately 10 kg was collected per sample within the coarsest-grained outcrops that were not intruded or altered. Thin sections were cut from 20 samples of the Monos Formation for petrographic and detrital mode characterization. Point counts were performed on 11 of these using the Gazzi-Dickinson method (Dickinson, 1970; Ingersoll et al., 1984) in order to determine the tectonic provenance of the samples.

**Detrital Zircon U-Pb Geochronology and Geochemistry**

Zircon grains were separated from sandstone samples using standard crushing and heavy-liquid separation techniques (e.g., Gehrels, 2000) at Northern Arizona University (Flagstaff, Arizona, USA). Samples MDZ4 and MDZ9 were annealed and chemically abraded to eliminate portions of zircon grains that have experienced lead loss (cf. Mattinson, 2005; Riggs et al., 2014). U-Th-Pb analysis of detrital zircon was done at the University of California–Santa Barbara (UCSB) laser-ablation split-stream (LASS) petrochronology lab. This lab uses a Nu Instruments Plasma HR MC-ICPMS (high-resolution multi-collector inductively coupled plasma mass spectrometer) for isotopic analysis. This mass spectrometer is coupled to a Photon Machines 193 nm Excimer laser with a HeLex sample cell that delivered a 25 μm beam for U-Pb analyses. Analyses were calibrated using zircon reference materials 91500 (1082.4 ± 0.4 Ma; Wiedenbeck et al., 1995) and GJ-1 (601.9 ± 0.7 Ma; Horstwood et al., 2016) as primary standards between every eight unknown samples. Plešovice zircon (337.71 ± 0.37 Ma; Sláma et al., 2008) was used as the secondary standard and was analyzed at the beginning and end of each run; measured Plešovice ages were within 1%–2% of the accepted isotope-dilution thermal ionization mass spectrometry (ID-TIMS) value (337.13 ± 0.37 Ma; Sláma et al., 2008). Data were reduced using the lolite 2.31 software package in WaveMetrics Igor Pro. Graphical representations of the data were created using IsoPlot (http://www.bgc.org/isoplots etc/isoplots.html) and detritalPy software (Sharmann et al., 2018). Due to precision cutoffs in different isotopic systems, 207Pb/206Pb ages are reported for grains older than ca. 1.2 Ga, while 207Pb/206U ages are reported for grains younger than ca. 1.2 Ga. Grains exhibiting >10% discordance or >5% reverse discordance were discarded; discordance was determined using 207Pb/206Pb versus 206Pb for grains <1.2 Ga, and 207Pb/206Pb versus 206Pb for grains >1.2 Ga. In total, the Monos Formation yielded 638 concordant grains (88% of all analyzed grains).

Maximum depositional ages (MDAs) were calculated for each of the samples, assuming that the youngest cluster of grains approximately defines the age of deposition. Here MDAs were calculated using the YC01s+1 method from Dickinson and Gehrels (2009a). We inferred that the largest number of youngest grains whose weighted average yields a mean square weighted deviation (MSWD) very close to 1.0 provides the best estimate of an MDA (Wendt and Carl, 1991).
RESULTS

Facies and Stratigraphic Observations

Previous work by Cooper et al. (1953) briefly described the Monos Formation as a >600-m-thick succession of red siltstone, sandstone, and fossiliferous sandstone, but these authors did not suggest any further subdivision of the formation. We divide the Monos Formation into four informal units (Figs. 2–3) based on assemblages of lithofacies defined in outcrop (Table S1). Unit 1 is the only wholly siliciclastic unit within the Monos Formation. Units 2 through 4 are predominantly bioclastic; differences among these three units are based on allochem proportion. In general, unit 2 is mudstone dominated, unit 3 is wackestone-packstone dominated, and unit 4 is grainstone dominated (Fig. 3; Table S1). The predominant facies for each unit are: medium- to coarse-grained volcanioclastic and feldspathic sandstone interbedded within siliceous, laminated mudstone in unit 1; planar-laminated to wavy-laminated or massive, sponge spicule–rich mudstone and wackestone in unit 2; planar- and wavy-laminated to massive, spicle-bryozone-echinoderm-rich wackestone and packstone in unit 3; and massive and echinoderm-rich, bioclastic grainstone in unit 4 (Fig. 3; Table S1).

Petrographic Descriptions

Detailed thin-section descriptions from a representative suite of volcanioclastic to calcareous Monos lithologies are given in Table S2 (footnote 1). These descriptions include: grain size, sorting, and rounding; matrix and cement composition; and volcanic, siliciclastic, and bioclastic components, including monomineralic grains and lithic fragment types. The suite of samples becomes increasingly calcareous upsection while volcanioclastic lithologies are confined to the lower two units of the formation.

The main monomineralic components are plagioclase feldspar, minor amounts of potassium feldspar, and rare monocrystalline quartz. Monocrystalline quartz becomes more common in the upper calcareous units. Dense minerals such as zircon, unidentified opaque minerals, and iron oxides are rare.

The dominant lithic components are volcanic with rare sedimentary fragments that are more common in the upper calcareous units. The majority of the volcanic lithics are undifferentiated volcanic epiclasts and occasional volcanic lithic fragments with microlitic and lathwork textures. Rare vitric volcanic lithic fragments contain vesicles. Interstitial components include a complex mixture of carbonate sparite cement, microporphyry, micrite, and feldspathic-cherty epimatrix composed of altered volcanic debris. In the calcareous units, micrite is more common in the lower mud-rich units, while sparite and microsparite cement dominates the interstitial component in the coarser units upsection. The main bioclastic components are sponge spicules, echinoderms, molluscs, and brachiopods.

Diagenetic alteration of feldspar and volcanic lithic fragments to clay minerals, carbonates, and quartz is commonplace in the lower volcanioclastic sandstones.

Sandstone Detrital Modes

Samples with principal grain sizes larger than silt were point counted and analyzed following the Gazzi-Dickinson point counting method (n = 300; Dickinson, 1970; Ingersoll et al., 1984). Point-count results and recalculated detrital modes for tectonic provenance interpretation are reported in Table S3 (footnote 1) and Table 1, respectively. Modal compositions for sandstones in lower units are dominated by feldspar and lithic fragments with secondary amounts of quartz, transitioning to more quartz in the bioclast-rich upper units (Fig. 4). Plagioclase crystal habits are commonly euhedral to twinned, although alteration to clays minerals commonly obscures the twinning. Potassium feldspar grains occasionally display plagioclase cores, which suggests that the potassium feldspar component has a secondary, diagenetic origin. One sample contains abundant altered volcanic lithic fragments with lathwork and microlitic textures while the other samples are dominated by undifferentiated altered volcanic lithic fragments (Figs. 6A–6D). Bioclasts include echinoderms, sponge spicules, and bryozoans.

Interpretation: Unit 1 records marine sedimentation in the form of gravity-flow deposits interpreted as turbidites based on the vertical succession of sedimentary structures (e.g., Bouma, 1962). Facies 1 contains Bouma sequence intervals Tc–Tm and Tm along with erosive contacts of basal sandstones into mudstone, which is consistent with turbidity-current origins. The predominance of volcanic lithic fragments along with feldspar-bearing, altered pseudomatrix suggests a volcanic provenance.
Figure 4. Modal Qt-F-L (Qt—total quartz; F—feldspar; L—lithic rock fragment) and Qm-F-Lt (Qm—monocrystalline quartz; Lt—total lithic rock fragments) compositional ternary plots for the Monos Formation, following Dickinson et al. (1983) and Marasgia and Ingersoll (1992). Recalculated parameters are given in Table 1. Brown squares—unit 1; green triangles—unit 2; cyan circles—unit 4; yellow star—mean composition of all samples. Larger numbered symbols are unit-specific, averaged compositions. Red arrows are compositional trends moving upsection. Tectonic compositional fields: CI—craton interior; TC—transitional craton; BU—basement uplift; RO—recycled orogen; DA—dissected arc; TA—transitional arc; UA—undissected arc; CONT.—continental arc; INTRA.—intra-oceanic arc; MO—mixed orogen; QR—quartzose recycled; TR—transitional recycled; LR—lithic recycled.

TABLE 1. RECALCULATED* POINT-COUNT DATA FROM THE MONOS FORMATION

| Sample | Qt-F-L | Qm-F-Lt | Qp-Lv-Ls | Qm-K-P | Lvv-Lvmi-Lvl |
|--------|--------|---------|----------|--------|--------------|
|        | Qt (%) | F (%)   | L (%)    | Qm (%) | F (%)        | Lt (%)  | Qp (%) | Lv (%) | Ls (%) | Qm (%) | K (%) | P (%) | Lvv (%) | Lvmi (%) | Lvl (%) |
| Mts1   | 0      | 45      | 55       | 0      | 45           | 55      | 0      | 100    | 0      | 100    | 0     | 0     | 100     | N/A      | 62      | 4      | 34      |
| Mts2   | 3      | 94      | 3        | 1      | 94           | 5       | 29     | 71     | 0      | 1      | 98    | N/A    | 89      | 0      | 11      |
| Mts3   | 1      | 64      | 35       | 1      | 64           | 35      | 1      | 99     | 0      | 1      | 33    | N/A    | 86      | 0      | 13      |
| Mts4   | 1      | 84      | 15       | 1      | 84           | 15      | 0      | 100    | 0      | 1      | 6     | N/A    | 86      | 0      | 13      |
| Mts5   | 2      | 75      | 23       | 0      | 75           | 25      | 6      | 94     | 0      | 1      | 33    | N/A    | 86      | 0      | 13      |
| Mts6   | 0      | 70      | 30       | 0      | 70           | 30      | 0      | 100    | 0      | 0      | 11    | 98    | N/A      | 86      | 0      | 11      |
| Mts7   | 2      | 43      | 54       | 0      | 43           | 57      | 4      | 93     | 3      | 0      | 13    | 87    | N/A      | 86      | 0      | 13      |
| Mts8   | 1      | 59      | 41       | 0      | 59           | 41      | 1      | 96     | 3      | 0      | 34    | 66    | N/A      | 86      | 0      | 13      |
| Mts12  | 6      | 39      | 54       | 1      | 39           | 60      | 0      | 91     | 0      | 3      | 13    | 85    | N/A      | 86      | 0      | 13      |
| Mts19  | 68     | 13      | 19       | 0      | 13           | 87      | 78     | 18     | 4      | 0      | 48    | 52    | N/A      | 86      | 0      | 13      |
| Mts20  | 40     | 9       | 51       | 24     | 9            | 67      | 24     | 57     | 19     | 72     | 9     | 19    | N/A      | 86      | 0      | 13      |
| Average| 11     | 54      | 35       | 3      | 54           | 43      | 14     | 83     | 3      | 7      | 18    | 75    | 76      | 2      | 22      |

Notes: Qt—total quartz (= Qp + Qm + Qpc; Qp—polycrystalline quartz; Qm—monocrystalline quartz; Qpc—polycrystalline chert lithic); F—feldspar; L—lithic rock fragment; Lt—total lithics (= L + Qpc); Lv—volcanic lithic fragment; Ls—sedimentary lithic fragment; K—potassium feldspar; P—plagioclase feldspar; Lvv—vitrinous volcanic lithic fragment; Lvmi—microlitic volcanic lithic fragment; Lvl—lathework volcanic lithic fragment. Lvv-Lvmi-Lvl data are unavailable (N/A) if volcanic grain type was not identifiable. See Table S3 (text footnote 1) for point-count data.

*QtFL%Q = 100(Qt/(Qt+F+L)), QtFL%F = 100(F/(Qt+F+L)), QtFL%L = 100(L/(Qt+F+L)), QmFLt%Qm = 100(Qm/(Qm+F+Lt)), QmFLt%F = 100(F/(Qm+F+Lt)), QmFLt%Lt = 100(Lt/(Qm+F+Lt)). QpLvLs%Qp = 100(Qp/(Qp+Lv+Ls)), QpLvLs%Lv = 100(Lv/(Qp+Lv+Ls)), QpLvLs%Ls = 100(Ls/(Qp+Lv+Ls)), QmK-P%Qm = 100(Qm/(Qm+K+P)), QmK-P%K = 100(K/(Qm+K+P)), QmK-P%P = 100(P/(Qm+K+P)). LvvLvmiLvl%Lvv = 100(Lvv/(Lvv+Lvmi+Lvl)), LvvLvmiLvl%Lvmi = 100(Lvmi/(Lvv+Lvmi+Lvl)), LvvLvmiLvl%Lvl = 100(Lvl/(Lvv+Lvmi+Lvl)).
Moreover, the abundance of microlitic and lathwork volcanic fragments implies proximity to a volcanic center, which sourced volcaniclastic turbidity currents into the basin.

Unit 2

Unit 2 consists of sponge spicule–rich mudstone (F3) and wackestone (F4; Figs. 5C–5D; Table S1 [footnote 1]). Individual beds range from ~10 cm to >1 m thick. Other facies making up the unit include rare, laminated siltstone (F5) and laminated bioclastic wackestone, packstone, and grainstone that locally exhibit cross laminae, cross bedding, and zones of brachiopod-rich rudstone (F6; Fig. 5C). Facies 5 is typically interbedded in sharp contact with F3. No bioclasts were noted in F5. Facies 6 occurs throughout the upper half of section 2 (Fig. 2). Like in F3, the most identifiable bioclasts in F6 are sponge spicules, followed by echinoderm and brachiopod fragments. Brachiopod bioclasts are typically fragmented, are large (>1 cm), and occur either solitarily within wackestones-packstones or in brachiopod rudstones (Figs. 5C–5D). Planar laminations are common and cross laminations, cross bedding, and syn-sedimentary deformation occur in three 20–100-cm-thick beds.

Apparent matrix in these samples is made up of chert and euhedral plagioclase feldspar as seen in thin section. Faint boundaries within the matrix are rarely delineated by iron oxide rims, suggesting that the interstitial material is a pseudomatrix or epimatrix made up of altered volcanic lithic fragments. Monomineralic grains within unit 2 include common plagioclase feldspar, potassium feldspar, minor quartz, and rare heavy minerals (Table S2 [Footnote 1]). Undifferentiated volcanic lithic fragments are abundant, and sedimentary lithic fragments are rare. Bioclasts are common and include echinoderms, bryozoans, and molluscs. Plagioclase grains are mostly altered to clay minerals, although twinning is still visible. Potassium feldspar grains typically contain plagioclase cores, suggesting diagenetic alteration of the plagioclase grains. Spicule-rich mudstone is predominantly micrite with rare subangular monocrystalline quartz (Fig. 5E). Bioclasts include common sponge spicules and rare foraminifera, radiolari, gastropods, and ostracods. Authigenic quartz commonly replaces micrite matrix and bioclasts. Grains are subrounded to locally...
Figure 6. Photomicrographs of the Monos Formation, with scale bars as shown. (A) Photomicrograph under plane-polarized light of sample M1 (zircon sample MDZ1) containing opaque volcanic-glass fragments (Lv). Note the irregular shape, vesicles (red V) infilled with microsparite cement (CaC), and euhedral plagioclase feldspars (P) both solitary and within the glass fragments. (B) Photomicrograph of sample M3 (zircon sample MDZ3) with polarized light, exhibiting a large potassium feldspar (K) and plagioclase feldspars differentially weathered to clay (P) within a feldspathic-cherty matrix (M) that is likely altered volcanic debris. (C) Photomicrograph with polars crossed of sample M5, which contains both microlitic (Lvmi) and sedimentary (Ls) lithic fragments, a possible bryozoan fragment (Bry), and monocrystalline quartz (Qm). (D) Photomicrograph with polars crossed of sample M5, which contains a bryozoan bioclast (Bry), sedimentary lithic fragment (Ls), and subrounded volcanic lithic fragments (Lv), within a feldspathic-cherty matrix (M) that is likely altered volcanic debris. (E) Photomicrograph in plane-polarized light of sample M10 (zircon sample MDZ3), a spicule-rich (Sp) wackestone-packstone containing a large solitary echinoderm (Ech) bioclast. (F) Photomicrograph with polars crossed of sample M15, a fossiliferous packstone exhibiting highly fragmented echinoderm bioclasts (Ech) and subrounded to subangular dark volcanic lithic fragments (Lv). (G) Photomicrograph with polars crossed of sample M17, a fossiliferous grainstone containing a rounded lithic fragment that is likely siltstone (Ls?), echinoderm bioclasts (Ech) surrounded by calcite overgrowths, a bryozoan bioclast (Bry), and a possible foraminifera bioclast (Foram). (H) Photomicrograph in plane-polarized light of sample M18, an echinoderm-rich fossiliferous grainstone containing echinoderm bioclasts (Ech) with well-developed calcite overgrowths, and a large brachiopod (mollusc?) bioclast (Brach(?)).
subangular. Intergranular pores are rarely infilled by microsparite. Other samples are spicule-rich sparse biomicrites with predominant micrite and common microsparite. Bioclasts include predominant sponge spicules, minor echinoderms, and rare radiolarians, gastropods, and ostracods. Detrital grains interspersed within the micrite matrix include rare monocrystalline quartz and silt. Grains are moderately to locally well sorted with subrounded to locally subangular shapes. Authigenic quartz commonly replaces matrix and minor bioclasts. Intragranular bioclast pores are commonly infilled with micrite and rarely with microsparite.

Interpretation: Deposition of unit 2 likely occurred in a low-energy, shallow-marine environment with normal salinity. Mudstone-wackestone deposition represents normal sedimentation, while rudstones represent influence from storm action. Observed sedimentary and biogenic structures and bioclasts suggest deposition within sub- to intertidal regions.

Unit 3

In outcrop, unit 3 consists primarily of spicule-bryozoan-echinoderm-rich wackestone and packstone (F7) (Table S1 [footnote 1]). Sponge spicules and bryozoan fragments are locally abundant, along with minor amounts of echinoderms and brachiopods. Laminae are defined by alternating sand-sized, allochem-rich layers that in places normally grade into allochem-poor layers. Other lithofacies include very fine-grained laminated sandstone (F8) and bioclastic packstone-grainstone (F9). Facies 8 forms thin, planar-laminated beds in sharp contact with beds of F7. Facies 9 is similar to F7 but with an increase in bioclasts. This facies exhibits planar and wavy laminae, although rare massive beds are also present.

Matrix material includes spicule-rich, sparse biomicrite and echinoderm-rich, poorly washed biomicrite. Bioclasts include predominant echinoderms, common sponge spicules, and minor amounts of brachiopods. Detrital components include rare siltstone, chert, and peloidal to undifferentiated volcanic lithic fragments, and subrounded monocrystalline quartz grains (Fig. 6F). Grains are poorly sorted and subrounded, with large (>50 mm) echinoderms within micrite. Authigenic quartz commonly replaces matrix and bioclasts.

Interpretation: The general increase in allochem proportion and grain size suggests that unit 3 was deposited in a shallower setting than unit 2. Water depth was likely shallow marine, within the subtidal to intertidal ranges. The interbedded planar- to wavy-laminated bioclastic wackestone and packstone are interpreted as deposits reworked by storms.

Unit 4

Unit 4 consists of facies 10: massive and echinoderm-rich, bioclastic grainstone (Table S1 [footnote 1]). Identified bioclasts include echinoderms, bryozoans, and brachiopods in outcrop, and ostracods and foraminifera in thin section. The grains are moderately to well sorted, ranging from coarse sand to small pebbles in size. Rare beds of moderately well-sorted coarse sandstone beds with rounded sedimentary chert and mudstone lithic fragments are also present.

In thin section, unit 4 samples are made up of bryozoan-rich packed biomicrite containing micrite and microsparite (Figs. 6G–6H). Bioclasts include predominant bryozoans, abundant sponge spicules, minor amounts of echinoderms and molluscs, and rare amounts of ostracods. Detrital grains include chert, siltstone, undifferentiated volcaniclastic lithic fragments, and siliceous mudstone lithic fragments; polycrystalline and monocrystalline quartz grains are common. Grains are moderately sorted and subrounded. Micrite rims surrounding bioclasts and micrite infills of intergranular pores are common. Complex, wavy to linear patterns within the sample may represent the preferential orientation of bryozoan bioclasts. Echinoderm-rich, sorted to rounded biosparites are predominantly grain supported with common to minor amounts of micrite and microsparite.

Interpretation: Unit 4 represents the deposition of bioclastic-sand shoals and banks above fair-weather wave base. Wave and tidal reworking winnowed all fine-grained material and destroyed any internal structure, yielding its massive appearance. Increased allochem proportions and grain size suggest a general shoaling from unit 3.

Detrital Zircon U-Pb Geochronology and Maximum Depositional Ages

A total of 638 zircons from six samples of the Monos Formation were analyzed for detrital zircon U-Pb geochronology (Fig. 7; Table S4 [footnote 1]). In all cases, detrital zircon age spectra are dominated by a single Permian population (65% of all analyzed zircons) with a peak age of 274 Ma (Fig. 8). The remaining 35% of grains range from 3300 to 300 Ma and have multiple peak ages at ca. 2730 Ma, 1770 Ma, 1690 Ma, 1520 Ma, 1065 Ma, 610 Ma, 430 Ma, and 340 Ma (Fig. 8). Calculated maximum depositional ages (MDAs) range between 269 ± 3 and 258 ± 3 Ma (Table 2; Fig. S1 [footnote 1]). The lower four samples yield Ordovician MDAs within uncertainties (269 ± 3 to 263 ± 3 Ma), which is in agreement with the index fossils from the middle to upper fossiliferous units of the Monos Formation (Fig. 2; Cooper et al., 1953, 1965; Buitrón-Sánchez et al., 2007; Lara-Peña and Navas-Parejo, 2018a, 2018b). This, combined with the volcaniclastic nature of the lower portions of the formation and the observation that the majority of analyzed grains make up the youngest peak age, suggests that our calculated ages for these strata are representative of depositional ages. MDAs are generally good indicators of depositional age in arc-related settings, where zircon is likely crystalized and deposited contemporaneously with the infilling of the basin (e.g., Dickinson and Gehrels, 2009a). The uppermost sample (MDZ9) yielded a Capitanian–Wuchiapingian MDA of 258 ± 3 Ma. Biostratigraphic analysis of the Monos Formation has not identified index fossils this young. This sample, however, was taken toward the uppermost portion of the formation in a region where index fossils have not been identified. We argue therefore that the slightly younger age from our analysis is within reason for a depositional age for the uppermost Monos Formation. All samples are either the same age within uncertainties or young upsection. The basal and uppermost samples from the Monos
Formation constrain the timing of deposition to ca. 269 ± 3 to 258 ± 3 Ma.

■ DISCUSSION

Potential Drivers of Shallow-Water Heterozoan and Spiculitic Facies Assemblages within the Monos Formation

The Monos basin was likely at tropical to subtropical latitudes, and thus photozoan assemblages (e.g., corals, foraminifera, molluscs; Nelson, 1988) would be expected given the shallow-water setting of the Monos Formation; yet, mixed heterozoan and sponge-spicule faunal assemblages dominate the succession. The abundance of these assemblages reflects environmental conditions (e.g., increased salinity, high nutrient influx, cool water temperatures, increased water turbidity) that inhibited phototroph development. A potential mechanism for transportation of nutrient-rich and cool water into shallow regions is coastal upwelling, the process in which denser, cooler, and nutrient-rich water replaces warm surface water. Several lines of evidence support this. The paleogeographic setting of western Laurentia adjacent to the Panthalassan Ocean suggests that enhanced onshore wave and swell propagation were common, promoting oceanic upwelling. Paleoclimate simulations by Kiehl and Shields (2005) showed that while late Paleozoic ocean temperatures were substantially warmer than modern ocean temperatures, equatorial ocean temperatures did not vary much from present-day tropical water temperatures. Kiehl and Shields (2005) attributed this to cool-water upwelling in the eastern-tropical Panthalassan Ocean. Permian siltstone and phosphatic deposits from the northern Rocky Mountains (northwestern North America) have been interpreted as recording marine upwelling along the western margin of Laurentia (Sheldon, 1964; Parrish, 1982; Carroll et al., 1998). Bioclasts that make up the carbonate units of the Monos Formation include large proportions of echinoderms, bryozoans, and molluscs. The mineralogical structures of these fauna may preferentially precipitate in cool-water conditions (Bone and James, 1993). Additionally, the lack of phototrophs and predominance of heterotrophs and aphotic fauna are essential aspects of cool-water carbonate communities (Nelson, 1988).

An alternative source for a high nutrient influx during the deposition of the Monos Formation is the delivery of biogeochemically reactive nutrients via eolian transport of dust into the Paleozoic seas of southwestern Laurentia. The transport of large quantities of iron-rich dust into Late Pennsylvanian western seas is recorded in eolian mudrock from the Horseshoe atoll (western Texas; Sur et al., 2015). Large-scale loess-erg systems across the western Laurentian interior persisted well into Mesozoic...
times, which may have provided ample wind-blown nutrient-rich detritus that enhanced bioavailability of nutrients essential to primary production (Lawton et al., 2018). While not definitively of eolian origin, rounded to subrounded quartz grains within the upper units of the Monos Formation are consistent with the physical abrasion expected in an eolian setting.

Facies Interpretations and Depositional History

The Monos Formation represents deposition in a marine environment that shallowed from deep-water turbidites to carbonate shoals. Unit 1 records volcaniclastic marine sedimentation in the form of turbidity currents. This environment contrasts with uppermost Proterozoic through Permian shallow-water miogeoclinal rocks exposed in central and western Sonora (Stewart et al., 1990; Fig. 1), suggesting that the Monos Formation records tec-tonism and subsidence in the area related to the inception of subduction and magmatism. The primary source for unit 1 was the nascent Cordilleran arc (Fig. 4). Mean-modal values of units 1 and 2 plot in the undissected to transitional arc field (Dickinson and Suczek, 1979; Dickinson, 1982, 1986; Dickinson et al., 1983; Marsaglia and Ingersoll, 1992; Fig. 4; Table 1; Table S3 [footnote 1]). We consider a more likely provenance to be an undissected arc, and attribute the transitional arc classification to diagenetic alteration of volcanic lithic fragments producing an abundance of epimatrix and pseudomatrix (e.g., Dickinson, 1970), which artificially shifts the sandstone compositional mode from more lithic rich (undissected arc field) to more feldspar rich (transitional arc field).

Bed thickness, grain size, and grain-to-mud ratios that consistently increase upsection throughout units 2 and 3 indicate continued shoaling. Units 2–4 represent deposition within an open carbonate-ramp system as indicated by (1) the gradual increase in allochem proportions relative to micrite from units 2 through 4, representing an increase in energy, (2) the presence of rudstones indicative of potential storm deposits within units 2 and 3, which suggests deposition above storm wave base, (3) the occurrence of massive bioclastic grainstone deposits that represent carbonate shoals common in modern-day examples of carbonate ramps (Gammon and James, 2001), and (4) transport of shallow-water bioclasts (e.g., large, fragmented brachiopods) into deeper water (cf. Flügel, 2004).

Provenance of the Monos Formation

Sandstone Petrography

The predominance of both volcanic lithic fragments with microlitic and lathwork textures (Fig. 6A) and plagioclase feldspar and the paucity of quartz in units 1 and 2 is consistent with derivation from intermediate volcanism associated with a magmatic arc (Fig. 4). Mean-modal values of units 1 and 2 plot in the undissected to transitional arc field (Dickinson and Suczek, 1979; Dickinson, 1982, 1986; Dickinson et al., 1983; Marsaglia and Ingersoll, 1992; Fig. 4; Table 1; Table S3 [footnote 1]). We consider a more likely provenance to be an undissected arc, and attribute the transitional arc classification to diagenetic alteration of volcanic lithic fragments producing an abundance of epimatrix and pseudomatrix (e.g., Dickinson, 1970), which artificially shifts the sandstone compositional mode from more lithic rich (undissected arc field) to more feldspar rich (transitional arc field).

Units 3 and 4 are dominantly carbonate with rare clastic interbeds. Unit 4 sandstone contains >25% mono- and polycrystalline quartz grains, in strong contrast to units 1 and 2, and plots in the recycled orogen field (Fig. 4). Volcanic-lithic fragments are overall less abundant. Matrix composition is dominantly sparite and micrite, and plagioclase feldspars are also much less common, suggesting a minimal contribution of tuffaceous detritus compared to units 1 and 2. These observations suggest that the upper Monos Formation, while still in a depocenter proximal to the arc, was also influenced by detrital input from the continent.

Volcanic arcs within tropical zones may produce bathymetric highlands that penetrate the photic zone and can catalyze carbonate development (e.g., Scholl et al., 1985). The predominance of carbonate in the Monos Formation, especially in the upper part, suggests that arc edifice development was primarily subaqueous. Forearc basins, however, may act as depocenters well after convergence-related subsidence has terminated or shifted geographically (e.g., Great Valley of California; Dickinson, 1995). Alternatively, carbonate development may have occurred after volcanism ceased or during lulls in volcanism. The younging upsection between unit 1 and unit 4, however, and the presence of volcanic lithic fragments within the upper units of the Monos Formation suggest that regional volcanism was at least intermittently ongoing.

Detrital Zircon Geochronology

Approximately 65% of all analyzed zircon grains are Permian (Fig. 8). We attribute these grains to the Cordilleran arc developing across southwestern Laurentian at that time. Recent detailed U-Pb zircon geochronology of nearby plutonic suites (e.g., the Los Tanques pluton and Sierra Pinta complex) as well as volcanic clasts from the Permio-Triassic succession of the El Antimonio Group yielded ages from ca. 275 to ca. 235 Ma, suggesting a local source for Permian grains from the Monos Formation (Arvizu et al., 2009; Riggs et al., 2009, 2010; Lindner, 2012; Arvizu and Iriondo, 2015). Given the proximity of the Los Tanques and Sierra Pinta complexes and the good overlap in documented ages, the nascent Cordilleran arc was a likely source.

The remaining ~35% of grains from the Monos Formation have ages ranging from Archean to Pennsylvanian. The age spectrum is a reasonable
match for Laurentian continental sources, with key age peaks from the Appalachian terranes (eastern North America; 470–280 Ma; Becker et al., 2005, 2006, and references therein), the Grenville orogeny (eastern to central North America; 900–1300 Ma; Moecher and Samson, 2006), the Midcontinent granite–rhyolite province (central United States; 1300–1500 Ma; Van Schmus et al., 1996), the Yavapai and Mazatzal terranes (southwestern United States; 1650–1800 Ma; Whitmeyer and Karlstrom, 2007), and the Canadian shield provinces (1800–2000 Ma; Whitmeyer and Karlstrom, 2007; Fig. 8). The Archean through late Paleozoic age spectrum of the Monos Formation is also a plausible match to detrital zircon spectra obtained from Permian strata of the Grand Canyon sequence (i.e., Hermit, Coconino, Toroweap, and Kaibab Formations, Arizona, USA; Gehrels et al., 2011; Fig. 9). Statistical comparisons (i.e., Saylor and Sundell, 2016) to the Paleozoic Grand Canyon succession show moderate similarities between the spectra (similarity = 0.88, likeness = 0.63, kernel density estimation [KDE] cross correlation = 0.44; Table 3). The weak KDE cross-correlation result between these spectra is likely a result of the variance in abundance between middle Paleozoic through early Neoproterozoic grains from these data sets. The Monos Formation contains an anomalously high amount of Paleozoic through early Neoproterozoic grains, with peaks at 342, 428, and 610 Ma (Fig. 8). Peak ages at 342 and 428 Ma could be derived from Appalachian plutonism and metamorphism associated with the Taconic and Acadian orogenies (Hatcher et al., 1989; Drake et al., 1989; Osberg et al., 1989; Bradley et al., 2000). However, the spectrum age peak at 610 Ma (n = 25) is outside of the accepted range for early Taconic magmatic origins (i.e., 500–550 Ma; Drake et al., 1989) and suggests that these grains derive from an alternate source. Coeval Permian eolian strata deposited across Oklahoma, Texas, and Kansas, USA, have been interpreted to likewise be sourced by proximal sources south and southeast of the Ouachita suture (e.g., ca. 420 Ma, Maya–Yucatan terrane; Soreghan and Soreghan, 2013); increased burial of the Ancestral Rocky Mountains and the increase in monsoonal westerlies may have hampered sourcing from far-eastern Laurentian terranes (Soreghan et al., 2018).

D detrital zircon data from the Permian Basin (western Texas) display Paleozoic–early Neoproterozoic age spectra similar to that of the Monos Formation (Liu and Stockli, 2019), with Neoproterozoic, Silurian, and Mississippian peaks (Fig. 9). While the Mississippian–Silurian age peaks overlap with ages of Appalachian magmatism, Liu and Stockli (2019) attributed their derivation to magmatic arcs in Gondwanan or peri-Gondwanan terranes of Mexico and Central America based on their closer proximity and the greater abundance of euhedral zircon grains in these age populations. Moreover, the 591–624 Ma age range is rare in Appalachian foreland clastic successions, whereas grain ages in that range overlap with the well-defined ages for peri-Gondwanan terranes in central Mexico (500–750 Ma; Krogh et al., 1993; Lopez et al., 2001; Keppie et al., 2006; Weber et al., 2006, 2007, 2008; Martens et al., 2010). The similar high abundance of these ages within the Monos Formation suggests that partial derivation from Gondwanan or peri-Gondwanan terranes is likely and that the Monos Formation received sediment from both the interior of Laurentia and the marginal terranes from Mexico and Central America. Similarity statistics between the Monos Formation and the Permian Basin strata are comparable to those between the Monos Formation and the Grand Canyon units (Table 3). The key age peaks of the Monos Formation suggest that while most of the material

### Figure 9

Histogram (bin size = 50 m.y.) and kernel density estimate plots of the >300 Ma Monos Formation spectrum compared to potential source regions: Paleozoic Grand Canyon sequence (Arizona, USA) and Permian Basin (western Texas, USA), compiled from Gehrels et al. (2011) and Liu and Stockli (2019), respectively.

### Table 3. Statistical Comparisons of the Monos Formation (Sonora, Mexico) Detrital Zircon Spectrum to Compiled Datasets from the Grand Canyon Paleozoic Sequence (Arizona, USA) and the Permian Basin (Western Texas, USA)

| Test | Grand Canyon sequence | Permian Basin |
|------|-----------------------|--------------|
| Similarity | 0.88 | 0.86 |
| Likeness | 0.63 | 0.62 |
| Cross correlation coefficient | 0.44 | 0.39 |

*Notes: These calculations were performed using DZstats (Saylor and Sundell, 2016) and detritalPy (Sharman et al., 2018) software. Source region data from the Grand Canyon and Permian Basin sequence from Gehrels et al. (2011); Leary et al. (2020) and references therein.*

Permian basin data are from Liu and Stockli (2019).
sourcing the Monos was the incipient magmatic arc, some detritus was also derived from both the interior of Laurentia and peri-Gondwana terranes.

Alternatively, the Archean–Paleozoic signature of the Monos Formation could be attributed to recycling of local Proterozoic–Paleozoic sedimentary rocks and metamorphic basement during the emplacement of the young Cordilleran arc and related crustal inflation. González-León et al. (2009) argued that the detrital zircon spectra from the Triassic–Lower Jurassic El Antimonio and Barranca Groups reflect basement recycling of local material due to the lack of a 1.8 Ga Mojave province signature and presence of a Proterozoic spectrum similar to that of local basement. Recycling of local basement influencing the detrital zircon spectrum of forearc sediments has also been noted in younger strata where preservation is likely better (e.g., Eocene strata of the Talara Andean forearc basin, northwestern Peru; Hessler and Fildani, 2015). However, Sonoran Paleozoic miogeoclinal strata are likely sourced from Proterozoic igneous and metamorphic basement terranes of the southwestern United States and Sonora (Gehrels and Stewart, 1998) and do not have significant populations of zircons that are common in the Monos spectrum (e.g., 600 Ma and 1100 Ma age peaks; Fig. 8). We therefore argue that local basement recycling was not the only source for the >300 Ma spectrum in the Monos Formation and suggest that sourcing from distal regions from Laurentia and Gondwana is more consistent with our detrital zircon data.

Relating the Monos Formation to Regional Tectonics

Based on Permian ages of calc-alkaline plutons in the Los Tanques–Sierra Pinta region, Permian volcanic clasts in El Antimonio Group conglomerates, and ca. 269–258 Ma maximum depositional ages of the Monos Formation, we argue that if the California-Coahuila fault crossed northern Sonora, transcurrent movement transitioned to transpression or convergence and subsequent subduction by ca. 280 Ma. This transition is supported by the change from oblique to convergent Panthalassan plate directions relative to a fixed North American reference frame during the late Permian based on global plate reconstructions by Domeier and Torsvik (2014). From ca. 319 to 280 Ma, the Panthalassan plate motion was obliquely convergent to parallel to the western Laurentian continent (Domeier and Torsvik, 2014). This corresponds to the left-lateral offset of the California-Coahuila fault (Fig. 10A). At ca. 279 Ma, the Panthalassan plate vector rotated ~25° counterclockwise (Domeier and Torsvik, 2014) and became approximately perpendicular to the fault. The change in plate motion may signal that the collision of the Caborca block along the Panthalassa-Pangea margin may have acted as a locus for the initial tectonic transition to a convergent margin and subsequent Cordilleran development by ca. 275 Ma (Fig. 10B). The plate motion maintained this vector until ca. 270 Ma and then later reconfigured into a direction more obliquely convergent during latest Permian times (Fig. 10C).

The shift in plate motions from semi-parallel to perpendicular to the continental margin matches with the earliest onset of subduction-related volcanic activity and later volcanioclastic contribution to the Monos Formation. Deposition of the Monos Formation took place along the margin of Laurentia, close to the only exposed arc-related plutons in Sonora (Arvizu et al., 2009; Riggs et al., 2010; Arvizu and Iriondo, 2015). Assuming the Monos Formation is autochthonous or para-autochthonous, this places the depocenter between the trench axis and magmatic arc (i.e., forearc setting). Given the limited outcrop extent of this unit, however, approximate basin size and location remain speculative. The youngest ages of arc-related plutons in the Caborca region that are likely related to the Monos Formation are ca. 280 Ma, which corresponds to a more oblique vector of Panthalassan plate motion. This suggests that arc development may have diminished significantly soon after 260 Ma due to changes in relative motion. Continued transpression may have caused persistent uplift of the region as evidenced by an unconformity between the Monos Formation and the overlying El Antimonio Group. Upper Permian to Lower Triassic rocks of the basal El Antimonio Group are interpreted as fluvial to shallow-marine units related to unroofing of Permian arc volcanoes (Linder, 2013). Moreover, Mesozoic sedimentary basins throughout northeastern Mexico (e.g., Dickinson and Lawton, 2001; Rubio-Cisneros and Lawton, 2011; Peryam et al., 2012) and the southwestern United States contain uncommon Permian zircon grains (Riggs et al., 2014, 2016). We argue that continued transpression along the margin caused the exhumation and transport of Permian material into these basins (Fig. 10C). Continent-derived zircon grains within the Monos Formation require explanation: either (1) the Monos Formation was deposited on the continent side of the arc in a retroarc foreland setting, (2) the incipient volcanic arc did not develop a robust topographic barrier that would have limited supply of craton-derived detritus, or (3) older strata that carried the continent-derived zircons were uplifted during arc development.

CONCLUSIONS

Stratigraphic facies analysis in conjunction with detrital zircon U-Pb geochronology of the Monos Formation yields information on the timing and style of Cordilleran arc development in southwestern Laurentia. Turbiditic facies that transition to a carbonate ramp, together with fossil assemblages, represent a shoaling-upward succession. Cold-water coastal upwelling and arc and continental sedimentation likely influenced the heterozoan faunal assemblage that makes up the Monos Formation.

Detrital zircon geochronology indicates the Monos Formation is middle Permian in age, corresponding with biostratigraphic constraints. A middle Permian age coincides with emplacement of Sonoran calc-alkaline plutons (ca. 275–260 Ma) interpreted as the earliest record of Cordilleran-arc magmatism. Sediment pathways from the Laurentian continent and marginal Gondwanan terranes into the Monos Formation are demonstrated by the presence of Archean through late Paleozoic zircons. These grains likely have origins from recycled Paleozoic cover from the Laurentian midcontinent and from tectonic terranes defining the northern margin of Gondwana. The volcanioclastic nature of
Figure 10. Tectonic reconstruction of the southwestern margin of Laurentia during the Permian. Red arrows show the plate motion directions of Panthalassa after Domeier and Torsvik (2014) relative to a fixed North American reference frame. Paleogeographic interpretations are from Leary et al. (2020) and references therein. CCT—California-Coahuila transform; O-M—Ouachita-Marathon; CAB—Caborca block; Fm.—Formation. (A) Initial truncation of Caborca block in the latest Pennsylvanian (Stevens et al., 2005). Panthalassa plate directions promote initial translation of the Caborca block toward its final position in northwestern Sonora and Baja, Mexico. (B) Orthogonal Panthalassa plate directions promote subduction zone development, arc magmatism, and subsequent deposition of the Monos Formation. The Cordillera was likely not a large topographic barrier to sediments from the Laurentian continent based on the presence of recycled Paleozoic sediments making up part of the Monos Formation’s detrital zircon spectrum. (C) Panthalassa plate direction returns to a more transpressional orientation. Arc magmatism may have dramatically slowed; sediment from exhumation and erosion of the arc edifice is recycled into Mesozoic basins in the southwestern United States and northwestern to north-central Mexico.
the Monos Formation, Permian zircons, and geographic location between coeval arc plutons and the inferred locus of southwestern Laurentian subduction initiation suggest deposition in a forearc setting.

ACKNOWLEDGMENTS

This research was funded to Riggs in part by the U.S. National Science Foundation (EAR 1322035). The Geological Society of America provided to Dobbs additional funding through a Graduate Student Research Grant. Northern Arizona University School of Earth and Sustainability supported this project through a Ronald G. Blakey Scholarship and Graduate Student and Pioneer Natural Resources awards. Funds for the publication fee were generously provided by the Stanford School of Earth, Energy and Environmental Sciences. Fruitful discussions with Ryan Leary and Pilar Navas-Parejo are very much appreciated. Field assistants Eric McDonald and Liz Johnson provided pivotal assistance to the research project, and we thank Andrew Kyland-er-Clark (UCSB LASS lab) for help in zircon analysis. Reviews by Tim Lawton, an anonymous reviewer, and Associate Editor Andrea Fildani greatly improved the clarity of the manuscript.

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The Monos Formation: Evidence for Permian Cordilleran arc development

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Dobbs et al. | The Monos Formation: Evidence for Permian Cordilleran arc development

536
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