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# DEPOSITIONAL ENVIRONMENTS, SEDIMENT DISPERSAL, AND PROVENANCE OF THE EARLY PERMIAN (LEONARDIAN) GLORIETA SANDSTONE, CENTRAL NEW MEXICO

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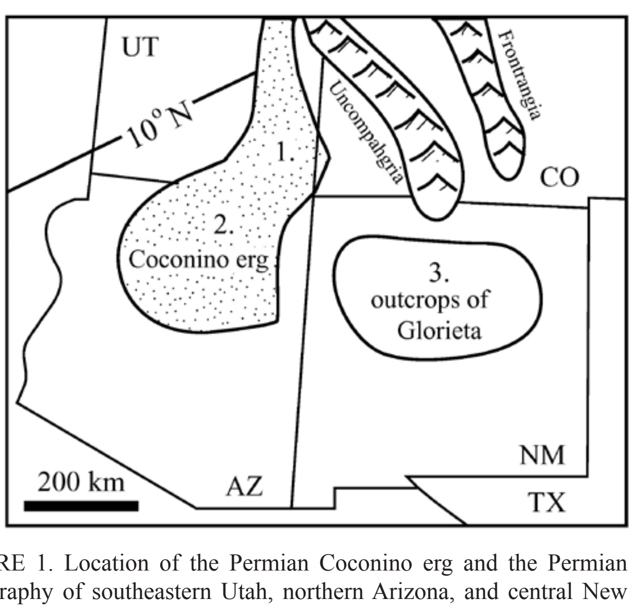
**ABSTRACT**—The Lower Permian (upper Leonardian) Glorieta Sandstone in north-central New Mexico represents an eastward extension of a large erg whose maximum thickness (305 m) and tallest eolian dunes (21 m high) are preserved in the Coconino Sandstone of northern Arizona. Unlike the Coconino dune field, the Glorieta segment of the erg was thinner (<92 m) and was primarily deposited as wind-rippled sand sheets that were locally traversed by dunes less than 7 m high. Crossbed orientations suggest that Glorieta dunes were transported and deposited by northeasterly trade winds, while larger Coconino dunes were driven by northerly and northwesterly onshore winds. Eolian sands of the Glorieta passed southward into a shallow sea where limestones of the San Andres Formation were deposited. Sandstone tongues of the Glorieta extend ~150 km southward into the basal San Andres Formation, and several beds of marine carbonate within the main body of the Glorieta indicate periodic northward transgression of the sea across the erg. Similar U-Pb age populations of detrital zircons in the Glorieta and Coconino suggest similar provenance of the sandstones, which is interpreted to have been deflation of a transcontinental river system whose headwaters were in the Appalachian-Ouachita Orogen and Canadian Shield, with a local source of sediment from the denuded Uncompahgria and Frontrangia uplifts of the Ancestral Rocky Mountains.

## INTRODUCTION

During late Early Permian (late Leonardian) time, a large erg (eolian sand field) corresponding to the Coconino and White Rim sandstones occupied the west-central margin of the Ancestral Rocky Mountains province in what is now northern Arizona and southeastern Utah (Fig. 1; McKee and Oriel, 1967; Peterson, 1980; Blakey 1990, 1996; Middleton et al., 1990). Paleowind directions responsible for dune migration during deposition of the Coconino Sandstone were from the north and northwest (Peterson, 1988), while U-Pb ages of detrital zircons suggest that much of the sand in the Coconino was sourced in the Appalachian Mountains and Canadian Shield, with lesser amounts from the Ancestral Rocky Mountains (Dickinson and Gehrels, 2003).

The Glorieta Sandstone in north-central New Mexico is generally considered to be correlative to the Coconino Sandstone in northern Arizona (Fig. 1; Blakey, 1988, 1990). The Glorieta Sandstone has been interpreted in local areas as either eolian (Milner, 1978) or shallow marine (Kelley, 1972; Lucas et al., 1999), but no detailed, regional sedimentologic studies have been undertaken. As a result, it is unclear how north-central New Mexico fits into late Leonardian paleogeography of the Ancestral Rocky Mountains. In this study, the Glorieta Sandstone was examined at 12 locations in central and northern New Mexico (Fig. 2) with the goals of interpreting depositional environments, sediment dispersal directions from crossbed paleocurrent data, and provenance from U-Pb dates of detrital zircons.



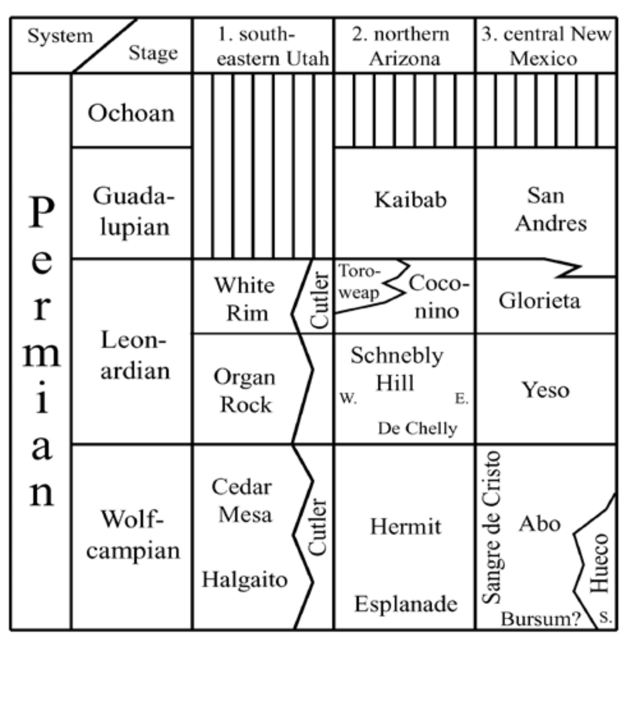




FIGURE 1. Location of the Permian Coconino erg and the Permian stratigraphy of southeastern Utah, northern Arizona, and central New Mexico (Blakey, 1990; Mack and Dinterman, 2002).

## STRATIGRAPHY

Throughout its main outcrop belt in central New Mexico, as well as in the northern Sandia Mountains (Placitas), in the Lucero Mountains, in the southern Nacimiento Mountains, and along the flanks of the Zuni Mountains, the Glorieta is either given formation status or is considered the basal member of the San Andres Formation (Fig. 2; Needham and Bates, 1943; Bachman, 1953; Foster, 1957; Jicha, 1958, Kottlowski, 1963; Kelley, 1972; Nelson, 1986; Colpitts, 1989; Woodward, 1987). In this region, the Glorieta is generally 30 to 90 m thick, although it is anomalously thin (~16 m) at Placitas (Lucas et al., 1999). It consists of cliff-forming, tan to light gray, fine to medium, quartzose sandstone. The Glorieta conformably overlies orange, fine sandstones of the Yeso Formation and is conformably overlain by dolostones or limestones of the San Andres Formation or, in the northeast,

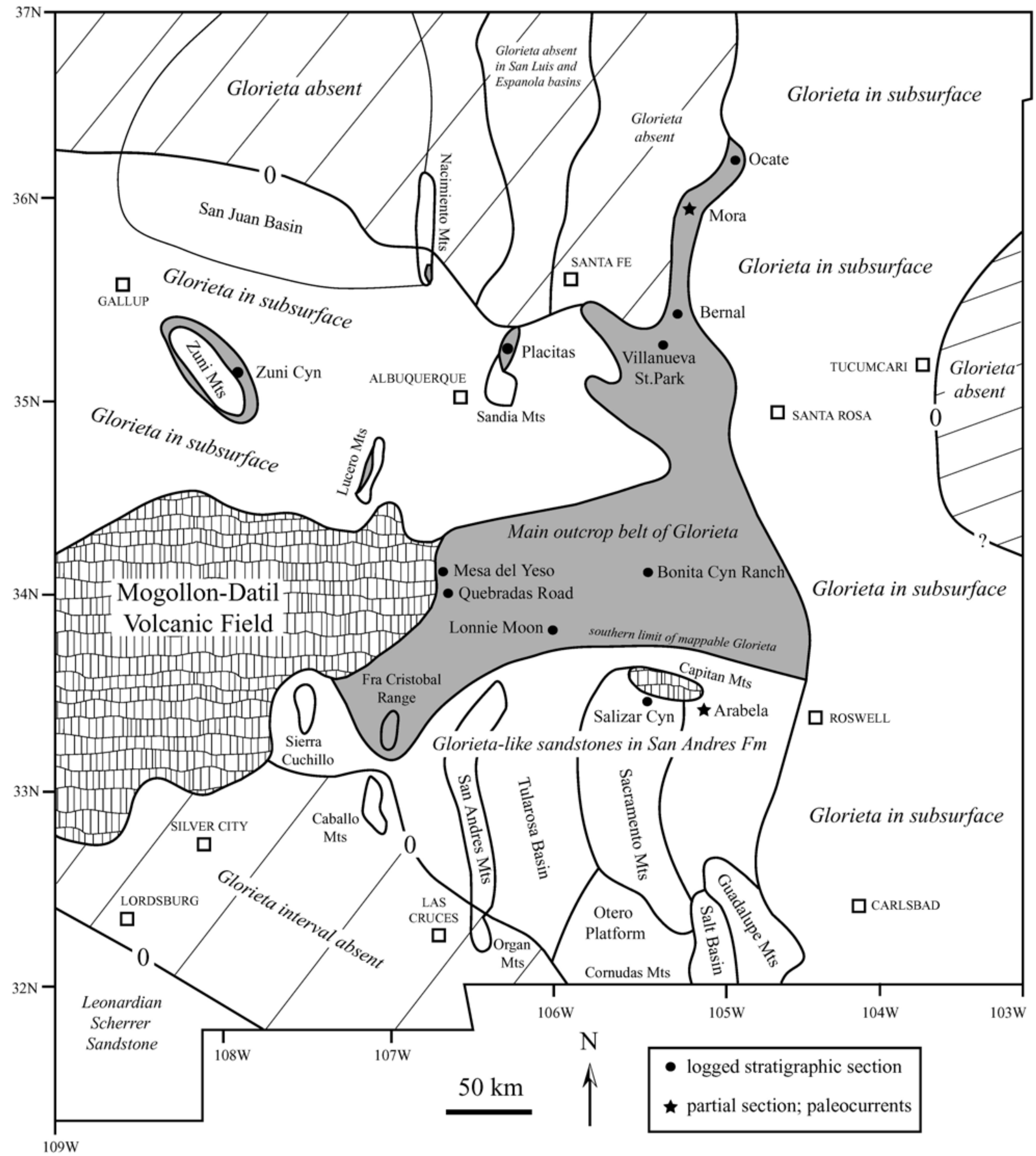



FIGURE 2. Distribution of the Glorieta Sandstone in New Mexico, as well as location of measured sections and additional partial sections where crossbed orientation data were collected. See text for references.

by brown, fine sandstone, siltstone, and gypsum of the Bernal Formation (Fig. 1). In the subsurface of eastern New Mexico the Glorieta is comparable in thickness to outcrops in the main outcrop belt, although it is locally absent east of Tucumcari (Fig. 2; Lloyd, 1949; Baltz, 1965; Foster, 1972; Foster et al., 1972; Roberts et al., 1976; Broadhead, 1987; Broadhead and King, 1988).

To the south of the main outcrop belt, in the Sierra Cuchillo, San Andres Mountains, and Guadalupe Mountains, one or more intervals of Glorieta-like sandstones are present at the Yeso-San Andres contact and within the lower part of the San Andres Formation (Fig. 2; Jahns, 1955; Kottlowski et al., 1956; Kottlowski, 1963; Kelley, 1971). In the northern Sacramento Mountains, Lang (1938), Pray (1961), and Harbour (1970) applied the name Hondo Sandstone Member to a Glorieta-like sandstone tongue in the lower San Andres Formation. Glorieta-like sandstones have also been recognized in the subsurface of the Tularosa Basin and in the northern and central Otero Platform (King and Harder, 1985). However, Glorieta-like sandstones are not present in the subsurface of the southern Otero Platform (King and Harder, 1985), nor in outcrops in the Cornudas Mountains (Nutt and O'Neill, 1998).

The Glorieta is absent throughout most of northwestern and north-central New Mexico, where it was removed by pre- or syn-Triassic erosion (Fig. 2; Baars, 1961; Woodward, 1987). The Glorieta also is absent in southwestern New Mexico (Fig. 2; Kottlowski, 1963), including in the Caballo Mountains (Seager and Mack, 2003) and Organ Mountains (Seager, 1981). In the bootheel of New Mexico, the Scherrer Sandstone occupies the same stratigraphic position as the Glorieta, but its relationship to the Gorieta is unclear (Fig. 2; Gillerman, 1958; Kottlowski, 1963; Zeller, 1965; Zeller and Alper, 1965).

## METHODS

Ten stratigraphic sections of the Glorieta Sandstone were measured and described in terms of rock types, grain size, bedding characteristics, physical and biological sedimentary structures, and fossils (Fig. 2). Crossbed orientations also were collected at the 10 measured sections, as well as at two additional sections (Mora and Arabela), and were plotted on rose diagrams. In addition, three samples were collected for U-Pb dating of detrital zircons; one sample each from the base of the Mesa del Yeso section, the middle of the Bernal section, and the top of the Zuni Canyon Section. Sample preparation followed the technique of Gehrels (2000) and is described in Bauer (2011). A total of 276 zircon grains were dated (Bauer, 2011) and the data compiled on relative age-probability plots at the LaserChron Center, University of Arizona (Gehrels, 2000).

## LITHOFACIES AND DEPOSITIONAL ENVIRONMENTS

The Glorieta Sandstone is separated into seven lithofacies (four sandstone, one shale, and two carbonate) (Figs. 3–5). The four sandstone lithofacies constitute 95.5% of the total exposed thickness of the logged sections.

### Rippled Sandstone

The most abundant lithofacies (44% of total exposed thickness) is rippled, well-sorted, fine to very fine sandstone. It consists of two varieties, which exist in intervals that range from less than a meter to 12 m thick. The more common variety has irregular, wavy beds a few centimeters thick, whereas the less common variety consists of centimeter-scale, light and dark layers of very fine and fine sandstone referred to as "pin-stripe" laminae (Fryberger and Schenk, 1988). Both varieties have few foresets and often coarsen upward, although the latter feature is locally difficult to detect because of uniform grain size. Ripple marks have gently sinuous, bifurcating crests with high ripple index (~50) due primarily to low amplitude (<1 cm).

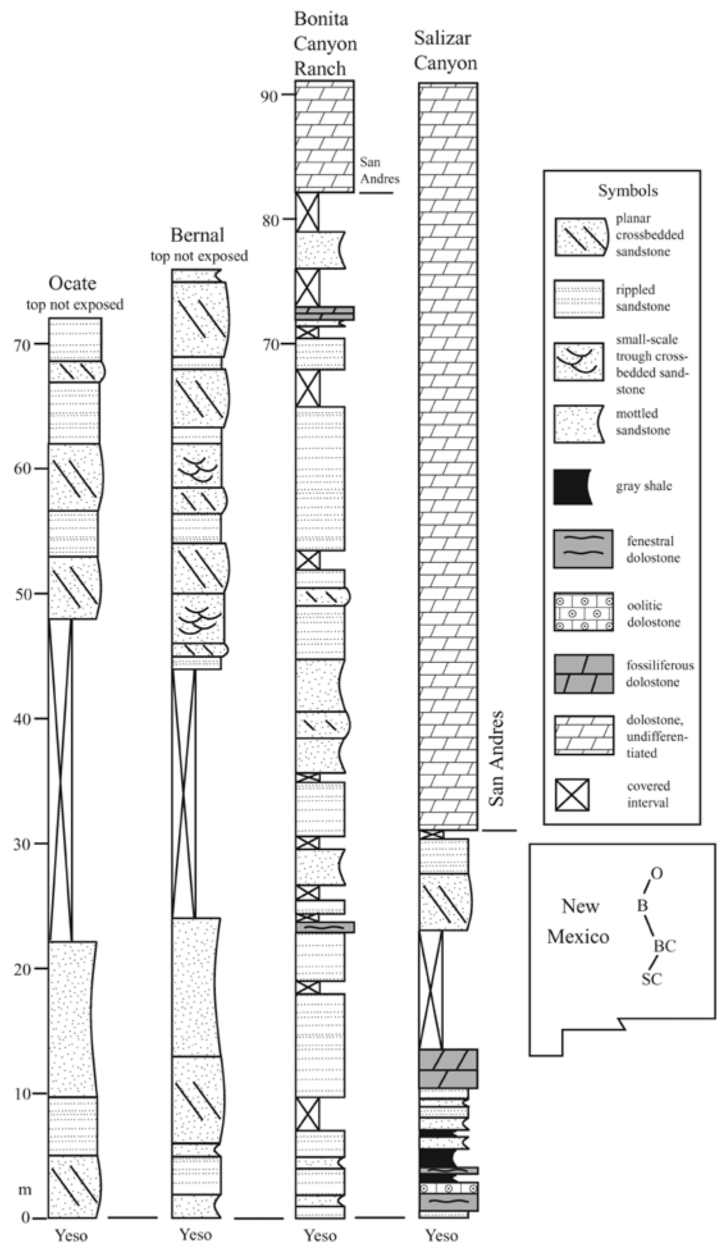



FIGURE 3. Measured sections and lithofacies at Ocate, Bernal, Bonita Canyon Ranch, and Salizar Canyon.

The rippled sandstone lithofacies has features diagnostic of wind ripples that have been created in the laboratory and observed in modern eolian deposits (Hunter, 1977; Fryberger and Schenk, 1988; Anderson and Bunas, 1993). Upward-coarsening and the paucity of foresets are related to the fact that the dominant process in the formation of wind ripples is saltation, with subordinate avalanching of grains down the lee side of ripple bedforms. Finer grains are suspended by collisions between the saltating grains, while coarser grains are not suspended by collisions, but instead "creep" along the bed surface. This process, plus sheltering of some of the finer grains in the troughs result in upward-coarsening of the ripples and the grain-size variation of "pin-stripe" laminae (Hunter, 1977; Fryberger and Schenk, 1988; Anderson and Bunas, 1993). Thin (<2 m) intervals of rippled sandstone interbedded with planar crossbedded sandstone may represent interdune deposits (Kocurek, 1981), whereas thicker intervals are more likely to represent eolian sand sheets peripheral to or independent of dune fields (Kocurek and Nielson, 1986).

### Planar Crossbedded Sandstone

Present at each of the logged sections and comprising 28% of the total exposed thickness is planar crossbedded, well-sorted, fine or medium sandstone. Individual crossbed sets range from 1 to 6.5 m thick, with the majority ≤3 m. Crossbeds exist as individual sets interbedded with other lithofacies, or as cosets in intervals up to 25 m thick. In a few cases, root traces up to 15 cm long are present in the upper parts of beds, and foreset surfaces have eolian ripple marks whose crests are oriented perpendicular to the dip of the foreset. Paleocurrent data locally display considerable scatter, but overall are strongly unidirectional to the southwest (Fig. 6).

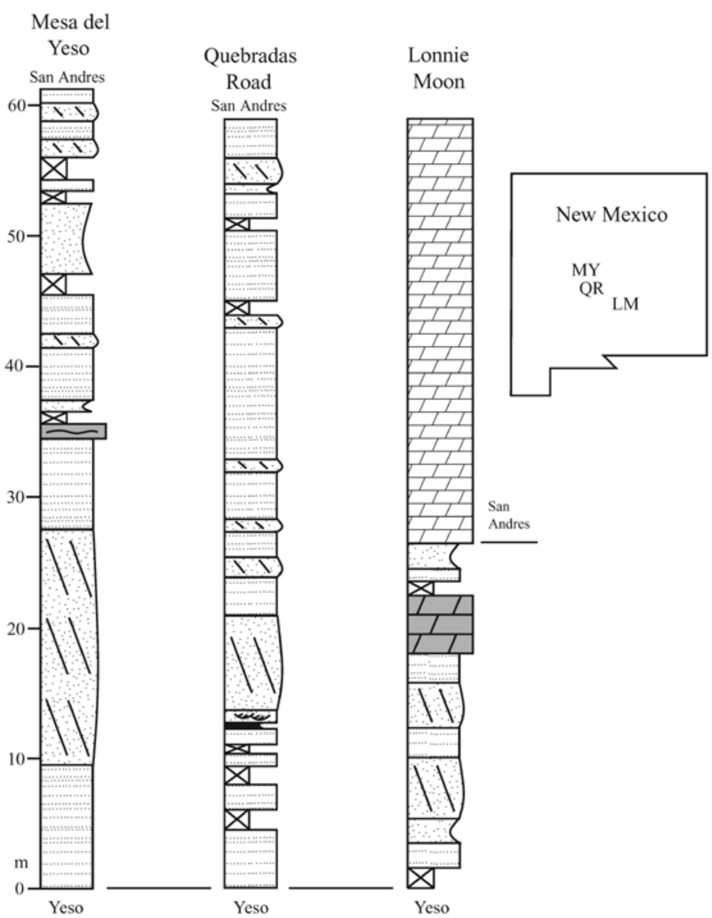



FIGURE 4. Measured sections and lithofacies at Mesa del Yeso, Quebradas Road, and Lonnie Moon. See Figure 3 for lithofacies symbols.

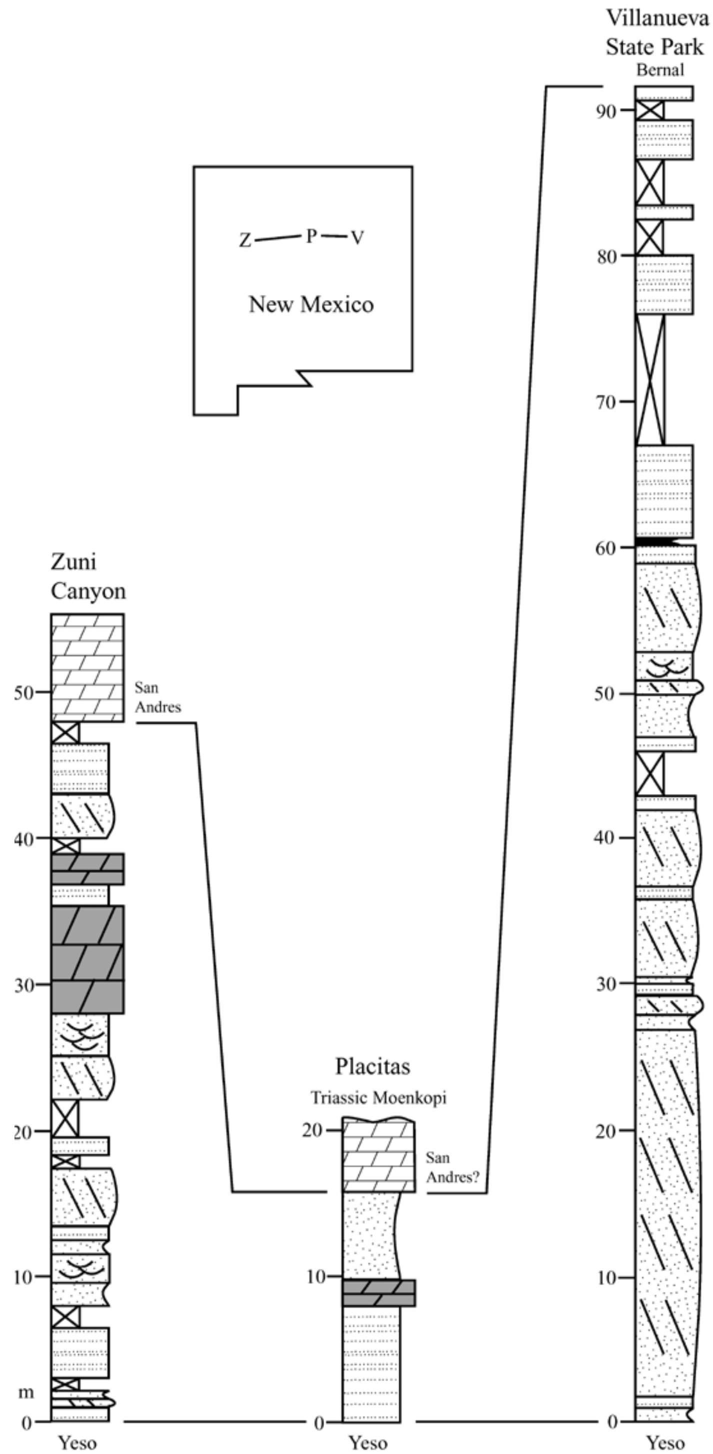



FIGURE 5. Measured sections and lithofacies at Zuni Canyon, Placitas, and Villanueva State Park. See Figure 3 for lithofacies symbols.

Thick stratigraphic intervals of well-sorted, fine or medium sandstone with relatively large-scale (≥2 m) crossbeds most likely represent deposition as eolian dunes (Kocurek, 1981, 1991). Consistent with this interpretation is the close stratigraphic association of the crossbedded sandstones with eolian-rippled sandstones, including ripple marks created by wind blowing across the slip face of the dune. Crossbed orientations indicate that northeasterly winds were primarily responsible for dune migration.

### Mottled Sandstone

Beds of mottled, well-sorted, fine sandstone comprise 20% of the exposed sections and range from individual beds 0.5 m thick to multi-bed intervals up to 12 m thick. Mottling is expressed as irregular patches of gray, yellow and pink. In addition, the mottled beds commonly display an irregular, rough surface that is recessed with respect to beds above and below. In a few cases, root traces a few millimeters wide and 5 to 10 cm long are present, along with small (0.5 cm wide, 3 cm long), straight to sinuous feeding burrows. Most beds, however, are internally structureless.

The lack of primary sedimentary structures inhibits interpretation of the depositional environment of the mottled sandstones. However, the grain size and sorting of the mottled sandstones are similar to those of rippled and crossbedded sandstones, suggesting the mottled sandstones are eolian as well. Those few beds with root traces and burrows show that primary sedimentary structures were destroyed by pedoturbation and bioturbation. A modern analog is heavily bioturbated sand interbedded with wind-rippled sand in the eolian sand sheet marginal to the Great Sand Dunes, Colorado (Fryberger et al., 1979). Those mottled beds in the Glorieta that are internally structureless may have been completely homogenized by post-depositional processes. Alternatively, the structureless beds could have been deposited from suspension, although this process, which produces loess, is generally restricted to silt-sized sediment (Pye, 1995).

### Small-Scale Trough-Crossbedded Sandstone

The least abundant of the sandstone lithofacies (3.5% of exposed sections) consists of well-sorted, fine and medium sandstone with trough crossbeds in sets 0.2 to 0.5 m thick. The beds are present at Bernal (Fig. 3), Quebradas Road (Fig. 4), Villanueva State Park, and Zuni Canyon (Fig. 5) in intervals 0.5 to 4 m thick. The beds are laterally continuous within the limits of the outcrop (10–50 m) and paleocurrents within each interval tend to be unidirectional. Locally, the trough crossbeds are interbedded with thin (<20 cm) intervals of climbing ripple cross-laminae.

The small-scale trough crossbedded lithofacies is interpreted to have been deposited by shallow streams, in which small dunes and climbing current ripples were the dominant bedforms. The absence of evidence of channel morphology and lateral accretion sets suggest that the streams were broad and perhaps braided. The streams may have existed in interdune areas (Langford and Chan, 1989), or periodically flowed across eolian sand sheets.

### Gray Shale

Rare beds (<1% of exposed sections) of gray, silty shale are present at Salizar Canyon (Fig. 3), Quebradas Road (Fig. 4), and Villanueva State Park (Fig. 5). The majority of beds are less than 0.5 m thick, although one bed at Salizar Canyon is 1.5 m thick. Several of the thinner beds grade laterally into planar crossbedded sandstones and are thought to represent interdune ponds. In contrast, the thicker shale bed may be marine, because it is interbedded with marine dolostone, although no marine fossils are present in the shale.

### Fenestral Dolostone

Thin (0.3–0.5 m), rare (<1% of exposed sections) beds of tan to light gray dolostone with millimeter-scale vugs (fenestral fabric) are present at Salizar Canyon, Bonita Canyon Ranch, and at Mesa del Yeso (Figs. 3, 4). In addition to fenestral fabric, the beds display some combination of wavy laminae, desiccation cracks, and brecciation on bed tops. At Bonita Canyon Ranch,

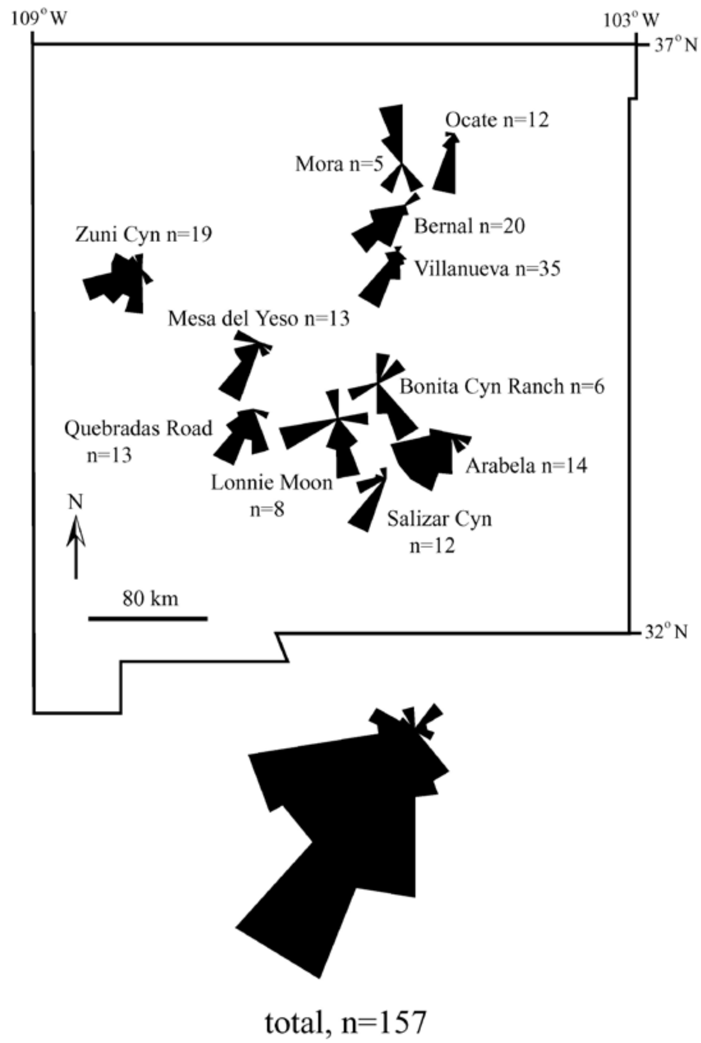



FIGURE 6. Crossbed orientations from ten measured sections, as well as two other partial sections (Mora and Arabela).

the fenestral dolostone bed also has centimeter-scale laminae and lenses of well-sorted, fine sandstone.

The fenestral dolostone lithofacies has features diagnostic of high intertidal and supratidal environments of modern carbonate tidal flats (Shinn, 1983). Wavy laminae probably represent stromatolites, whereas fenestral fabric is interpreted to have formed by desiccation or escape of gas bubbles. Brecciation may have resulted from desiccation and/or storm currents. The thin sandstone layers and lenses may represent eolian wind ripples spreading across the supratidal flat, or marine sand washed into the intertidal or supratidal zones by storms or high tides.

### Fossiliferous Dolostone

Beds of dark gray, massive dolostone 1 to 8 m thick are present at Salizar Canyon, Bonita Canyon Ranch, Lonnie Moon, Placitas, and Zuni Canyon (Figs. 4–6), but make up about 4% of the total exposed thickness of the Glorieta Sandstone. Several beds contain dolomitized brachiopods, corals, and/or moulds of shells of unknown affinity. In addition, petrographic analysis of one bed at Salazar Canyon revealed dolomitized ooids and scattered shell fragments (Fig. 3).

Dolomitization has apparently destroyed most of the texture of the original limestones, inhibiting interpretation of depositional environment. However, those beds with brachiopods or corals suggest deposition in a shallow-marine environment. Moreover, ooids are common in high-energy marine environments, including tidal deltas (Evans and Bush, 1969), carbonate beaches (Inden and Moore, 1983), and shallow-marine bars and shoals (Ball, 1967).

## SANDSTONE PETROLOGY AND U-Pb AGES OF DETRITAL ZIRCONS

Sandstones of the Glorieta are quartarenites (>95% quartz), making them difficult to interpret in terms of provenance (Milner, 1978). In order to shed light on the possible source of the sand, 276 U-Pb ages of detrital zircons were analyzed from three Glorieta sandstones, following the technique of Gehrels (2000). The zircon ages from the Glorieta fall into six distinct populations (Fig. 7; Bauer, 2011). The two most abundant populations, each of which constitutes about 30% of the total grains, include grains ranging from 300 to 499 Ma and from 1000 to 1299 Ma. Making up just less than 20% of the total grains is a population that ranges from 1600 to 1799 Ma, whereas Archean grains (>2500 Ma) constitute almost 17% of the grains. Smaller populations include grains from 1800 to 2199 Ma (~12%) and grains from 1300 to 1499 Ma (~5%).

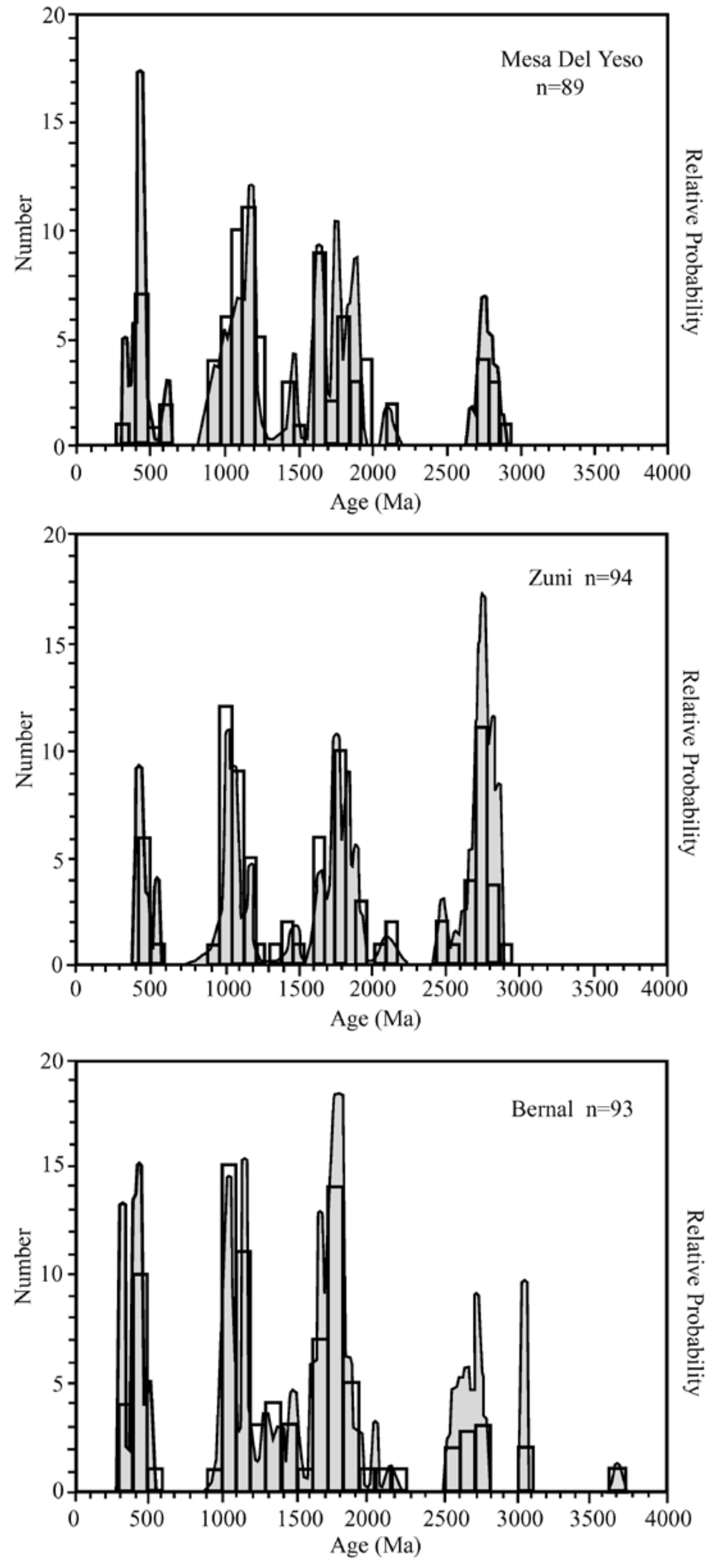



FIGURE 7. Relative probability age plots of U-Pb dates of detrital zircons from the Glorieta Sandstone at Mesa del Yeso, Bernal, and Zuni Canyon.

## DISCUSSION

### Depositional Environments and Sediment Dispersal

The predominance of eolian lithofacies in the Glorieta Sandstone suggests that it represents an eastern continuation of the Coconino erg, although several differences between the two

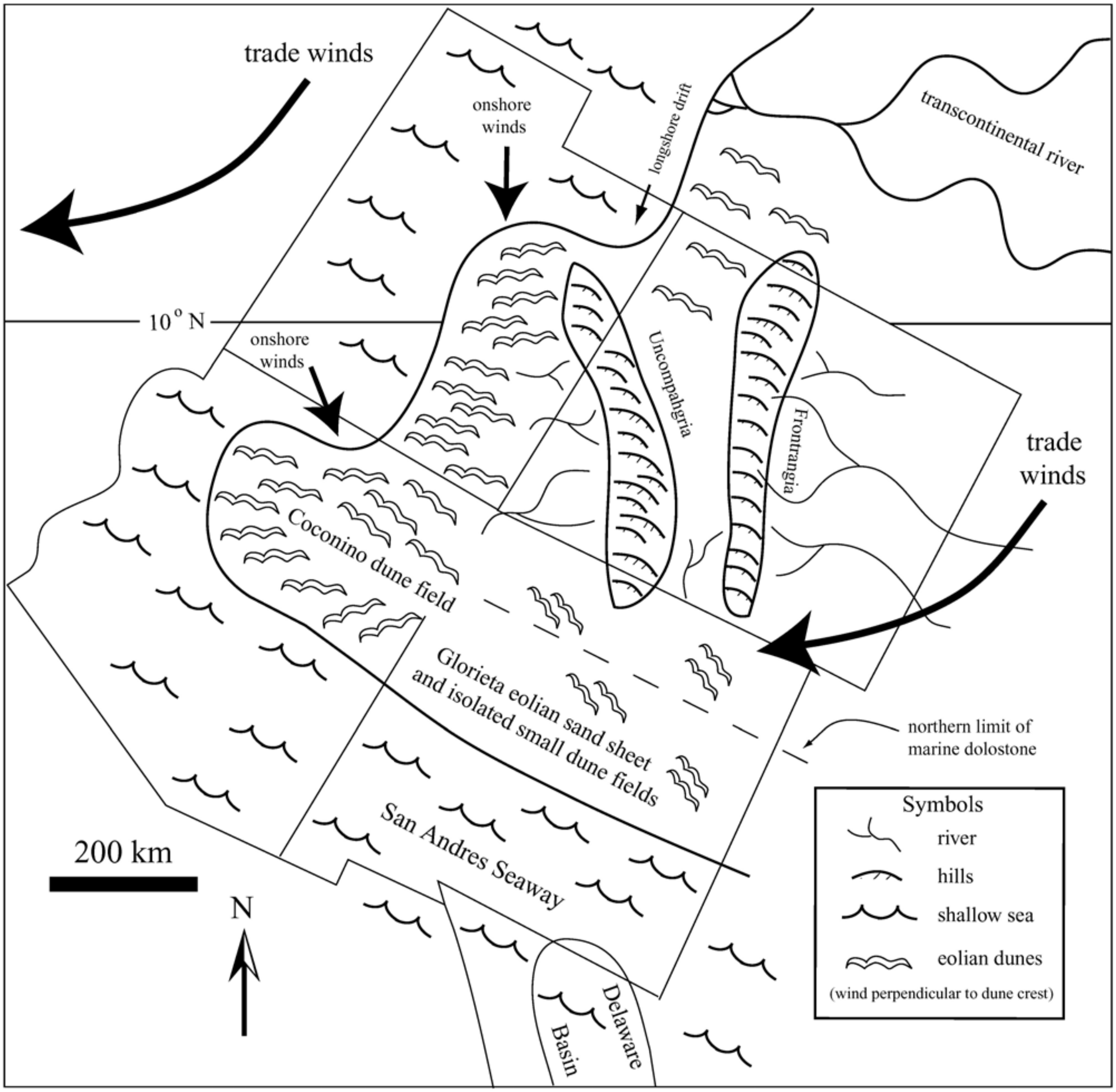



FIGURE 8. Paleogeographic map during deposition of the Leonardian Coconino, White Rim, and Glorieta sandstones in the Four Corner states. The states have been rotated to fit Leonardian paleolatitudinal orientation (Peterson, 1988).

regions of the erg are significant (Fig. 8). The Coconino Sandstone in northern Arizona is more than twice as thick as the Glorieta Sandstone in New Mexico. The Coconino is thickest in the eastern Grand Canyon (180 m) and along the Mogollon Rim near Sedona (305 m) (Blakey and Middleton, 1998), whereas maximum thickness of the Glorieta throughout its main outcrop belt is just over 90 m. The Coconino was primarily deposited as dunes, which, based on the thickness of individual crossbed sets, were up to 21 m high (Blakey and Middleton, 1998). In contrast, deposition of the Glorieta took place primarily as wind-rippled sand sheets that were only periodically traversed by dunes less than 7 m high. Crossbed orientations further indicate that the Coconino sand was transported and deposited by northerly and northwesterly winds (Peterson, 1988), suggesting that much of the sand was blown onto the erg by onshore winds (Fig. 8). The northeasterly winds responsible for dune migration in the Glorieta (Fig. 6) are more consistent with regional trade winds.

The Glorieta part of the erg was bordered to the south by a shallow sea, represented by limestones of the lower San Andres Formation (Fig. 8). The sea probably also existed south of the Coconino dune field, but its exact position is poorly constrained. Tongues of Glorieta-like sand extended up to 150 km southward during lowstands of the San Andres seaway. This is evident at Salizar Canyon (Fig. 3) and at Arabela (Fig. 2), where the Glorieta tongues are interpreted as eolian. However, Glorieta tongues have not been examined in this study south of the northern Sacramento Mountains, where their depositional environments are unknown. The Glorieta also has evidence in the form of interbedded peritidal and shallow-marine carbonate at Salizar Canyon, Bonita Canyon Ranch, Mesa del Yeso, Lonnie Moon, Zuni Canyon and Placitas (Figs. 3–5) for periodic marine transgressions that extended northward a minimum of 175 km into the main body of the Glorieta. Interbedding of marine and eolian beds, including limestones, also exists where the upper part of the Coconino intertongues with the Toroweap Formation in northwestern Arizona (Rawson and Turner-Peterson, 1980; Turner, 1990). Marine transgressions are not evident, however, in the main part of the Coconino dune field in northeastern Arizona (Blakey, 1990).

Several of the differences between the Coconino and Glorieta, such as thickness, dune abundance, and dune height, are best explained by a greater supply of sand to the Coconino than to the Glorieta. Eolian sand sheets, which were common in the Glorieta, generally are considered to be associated with low sand supply, although stabilization of the sand by vegetation or a high water table may have been important as well (Kocurek and Nielson, 1986). Greater sand supply to the Coconino may have been related to a combination of terrestrial and longshore sources of sand. Greater subsidence beneath the Coconino dune field also may have been a factor in the different thicknesses of the two formations (Blakey, 1988, 1990). The transgressions and regressions during Glorieta deposition may be glacial-eustatic in origin, based on studies that suggest continental glaciers on southern Pangea persisted into and through Leonardian time (Crowley and Baum, 1992; Frances, 1994; Fielding et al., 2008). The large geographic range of the sea-level changes in the Glorieta may have been enhanced by a very low depositional slope. Absence of marine tongues in the Coconino in northeastern Arizona may reflect the fact that the outcrops are north of the maximum extent of northward marine transgression and/or that the rate of progradation of the Coconino dune field exceeded the rate of sea-level rise.

## Provenance of the Sand

The provenance of the sand in the Glorieta can be constrained by comparison of its detrital zircon age populations with: (1) ages and locations of crystalline bedrock in North America (Fig. 9A), (2) detrital zircon ages from the Coconino Sandstone (Dickinson and Gehrels, 2003), and (3) detrital zircon ages of Guadalupian-age, deep-sea sandstones deposited in the Delaware Basin of southeastern New Mexico and west Texas (Soreghan and Soreghan, 2013). The zircon age populations in the Glorieta are nearly identical to those in the Coconino analyzed by Dickinson and Gehrels (2003), suggesting similar provenance of the sands. Dickinson and Gehrels (2003) suggested that the majority of the sand in the Coconino was derived from a transcontinental river system which headed in the Appalachian Orogen and on the Canadian Shield, and that a secondary source of sand came from the Ancestral Rocky Mountains. This model also seems applicable to the Glorieta, with minor modifications discussed below (Fig. 9B).

Dickinson and Gehrels (2003) suggested that Archean (>2500 Ma) and Paleoproterozoic (1800–2199 Ma) zircon grains in the Coconino, and by comparison in the Glorieta, were derived from the Superior and Trans-Hudson provinces (Fig. 9A). Although some of these grains may have been recycled out of lower Paleozoic passive-margin and cratonic sandstones, the large size of this population (33%, Coconino; 28% Glorieta) suggests that the majority of the grains are first-cycle in origin.

Dickinson and Gehrels (2003) further argued that three zircon age populations in the Coconino (300–499 Ma, 500–799 Ma, 1000–1299 Ma) are best explained by derivation from the Appalachian Orogen. The youngest and oldest of the three populations are well represented in the Glorieta (31%, 300–499 Ma; 30%, 1000–1299 Ma), whereas the population of intermediate age (500–799 Ma) constitutes only 1.3% of the total zircon grains in the Glorieta. Dickinson and Gehrels (2003) suggested that the Paleozoic population (300–499 Ma) was derived from the Taconic, Acadian, and Alleghenian provinces, whereas grains from 500 to 747 Ma were potentially sourced in the Carolina and Avalon accreted terranes (Fig. 9A). Guadalupian-age deep-sea sandstones in the Delaware Basin also have a significant population of zircons between 300 and 749 Ma, but Soreghan and Soreghan (2013) pointed out that these grains also could have been sourced from the Yucatan and Maya accreted terranes in the Ouachita-Marathon Orogen (Fig. 9A; Weber et al., 2008; Martens et al., 2010). Consequently, a potential Ouachita source for the Glorieta is incorporated into the drainage model presented here (Fig. 9B). It is unlikely, however, that Paleozoic and Neoproterozoic zircon grains in the Glorieta were transported northward from crystalline rocks in the Coahuilla, Oaxaquia, Acatlan, and Mixteca terranes of northern Mexico, because of the intervening San Andres seaway (Fig. 9).

There are three other potential sources of latest Proterozoic and Paleozoic zircon grains in the Coconino and Glorieta. The Wichita Mountains of Oklahoma have granites dated between 530 and 533 Ma (Fig. 9A; Hames et al., 1995; Wright et al., 1996; Hogan et al., 2000). However, Soreghan and Soreghan (2013) indicated that the Wichita Mountains were onlapped by late Early Permian time, presumably before deposition of the Glorieta and Coconino. There are also plutons in southern Colorado that range in age from 664 to 457 Ma (McMillan and McLemore, 2004), but they are small and widely scattered and were probably not a significant source of zircons to the Coconino and Glorieta. Finally, the Franklinian Orogen in northern North America supplied grains ranging from 370 to 450 Ma to its foreland basin (Patchett et al. 2004; Anfinson et al., 2012) and may have supplied zircons to the Coconino and Glorieta via the tributary drainage that crossed the Canadian Shield (Fig. 9B).

The oldest zircon population (1000–1299 Ma) considered by Dickinson and Gehrels (2003) to be of Appalachian provenance corresponds to the Grenville province (Fig. 9A). These grains could have come from exposures of Grenville basement in the core of the Appalachian Orogen, or were recycled from upper Proterozoic continental-rift sandstones (e.g., Ocoee Group), which were originally derived from Grenville basement and subsequently deformed during Appalachian orogenesis.

Grenville-age crust is also present in the Chiapas, Yucatan, and Maya accreted terranes of the Ouachita Orogen (Cameron et al., 2004; Weber et al., 2008; Martens et al., 2010). The Keweenawan province also has rocks in the age range of the Grenville province. However, the Keweenawan province was an unlikely source of zircons to the Glorieta, because the southern arms were buried by Paleozoic sediments and the ultramafic rocks of the province probably contain little or no zircons.

Two other zircon populations in the Coconino and Glorieta are best explained by derivation from basement-cored uplifts of the Ancestral Rocky Mountains. The large population from 1600 to 1799 Ma (19%, Coconino; 18%, Glorieta) matches the Yavapai-Mazatzal province, whereas the smaller population from 1300 to 1499 Ma (12%, Coconino; 5%, Glorieta) corresponds to the Granite-Rhyolite province (Fig. 9A). Crystalline rocks of these two provinces would have occupied the core of the ranges of the Ancestral Rocky Mountains. By the time of deposition of the Glorieta, however, all of the uplifts of the Ancestral Rocky Mountains, with the possible exception of Uncompahgria and Frontrangia, had been onlapped by post-tectonic sediment and were no longer sources of sediment (Mack and Dinterman, 2002; Soreghan and Soreghan, 2013). Although onlap of Uncompahgria may have been underway by Leonardian time (Soreghan and Soreghan, 2013), the range was still supplying arkosic detritus to the uppermost Cutler Formation, which is coeval to the White Rim and Coconino Sandstones (Fig. 1; Blakey, 1990). Uncompahgria was a likely source for Yavapai-Mazatzal-age and Granite-Rhyolite-age zircons in the Coconino, based on its proximity to the Coconino dune field. Given the geographic location of the Glorieta and the northeasterly winds responsible for deposition of Glorieta dune sand, it is likely that Frontrangia was the source of its 1600 to 1799 Ma and 1300 to 1499 Ma zircon grains. There is no stratigraphic evidence, however, for the timing of onlap of Frontrangia. Alternatively, zircons in the age range of 1600 to 1799 Ma and 1300 to 1499 Ma could have been recycled from Pennsylvanian and Lower Permian syntectonic sandstones and conglomerates that were derived from the Ancestral Rocky Mountains before their burial. This is unlikely in New Mexico, however, because there is no evidence for widespread erosional unconformities between any of the Pennsylvanian and Permian formations.

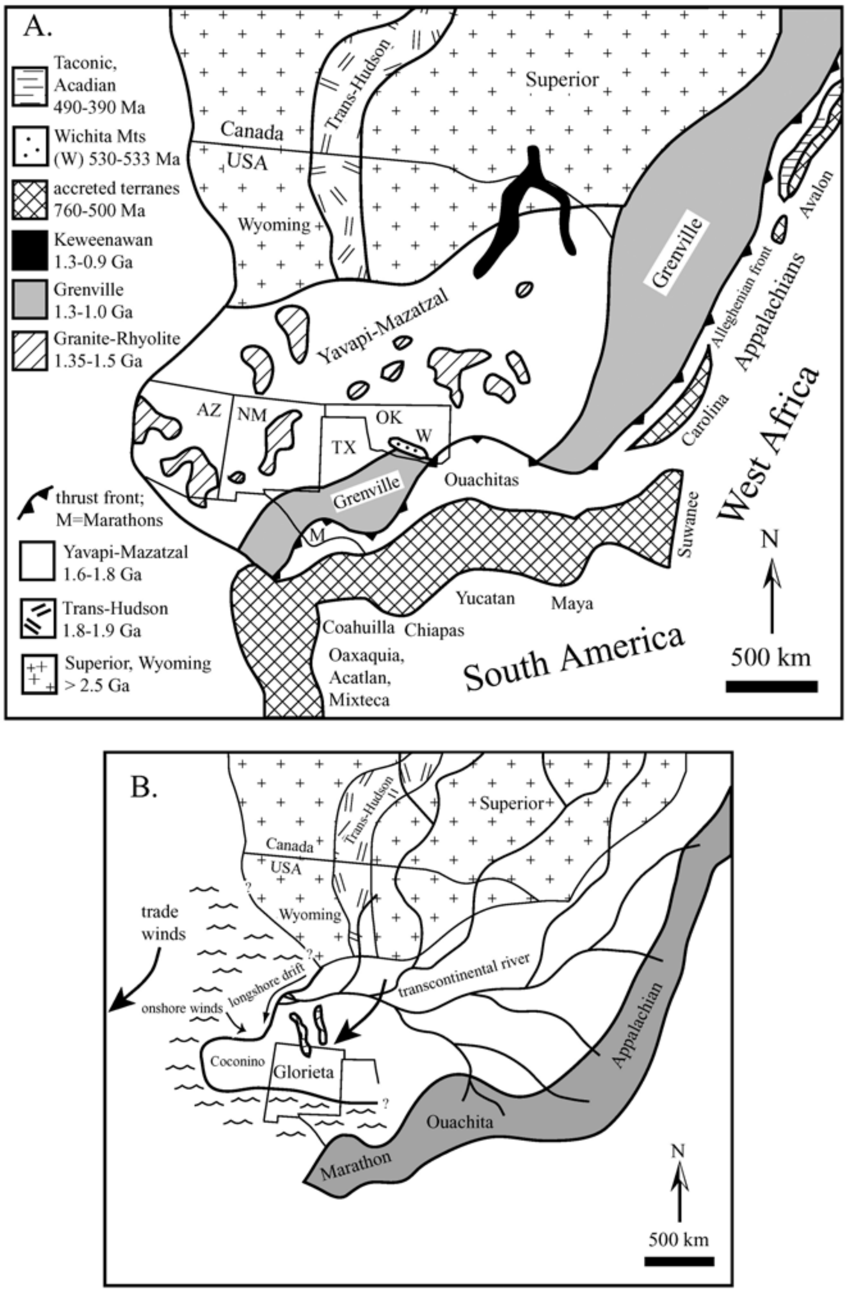



FIGURE 9. A. Distribution of crystalline basement rocks in North America, adapted from Soreghan and Soreghan (2013). W=Wichita Mountains, M=Marathon Mountains, AZ=Arizona, NM=New Mexico, TX=Texas, OK=Oklahoma. B. Schematic model of a transcontinental river system that supplied sand to the Glorieta and Coconino sandstones, based on Dickinson and Gehrels (2003) and Soreghan and Soreghan (2013). The model also assumes that two denuded, but not completely covered ranges of the Ancestral Rocky Mountains, Uncompahgria (western range) and Frontrangia (eastern range), supplied sand to the Coconino and Glorieta sandstones.

Two conditions are implicit for a transcontinental river system to be the source of sand in the Coconino and Glorieta. First, the river system needed enough discharge to flow several thousand kilometers across the continent. This is possible if the headwaters were in the Appalachian-Ouachita Mountains, where high elevation would have promoted large annual precipitation. The Canadian Shield also could have supplied significant discharge to the transcontinental river system, because of its position within a temperate, mid-latitude paleoclimatic belt. The second condition is that upon approaching its terminus, the Permian transcontinental river system needed to periodically experience very low or no discharge, so that strong regional winds could deflate sand from the river. Consistent with this idea is the fact that paleoclimate in the American Southwest became progressively drier through the Permian and was arid by Leonardian time (Mack, 2003). Moreover, the trade winds, suggested by Glorieta paleocurrent data, should have been strong and persistent enough to move large volumes of sand from the river.

A modern example of the type of river discussed above is the Rio Grande in Colorado and New Mexico, whose discharge is largely supplied by snowmelt and heavy rainfall in the headwater mountains of southwestern Colorado and northernmost New Mexico, but which flows through a region of arid climate (Mack and Leeder, 1998). There is historical evidence in southern New Mexico that the Rio Grande periodically went dry (Mack and Leeder, 1998), as well as evidence in the Holocene record from the same region for times when windblown sand created dunes and sand sheets that spread across the floodplain and onto adjacent terraces (Mack et al., 2011). A small coastal dune field also exists downwind of the modern Rio Grande delta, although in this location the climate is more humid than in New Mexico.

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