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Grand Canyon provenance for orthoquartzite clasts in the lower Miocene of coastal southern California

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ABSTRACT

Orthoquartzite detrital source regions in the Cordilleran interior yield clast populations with distinct spectra of paleomagnetic inclinations and detrital zircon ages that can be used to trace the provenance of gravels deposited along the western margin of the Cordilleran orogen. An inventory of characteristic remnant magnetizations (CRMs) from >700 sample cores from orthoquartzite source regions defines a low-inclination population of Neoproterozoic–Paleozoic age in the Mojave Desert–Death Valley region (and in correlative strata in Sonora, Mexico) and a moderate- to high-inclination population in the 1.1 Ga Shinumo Formation in eastern Grand Canyon. Detrital zircon ages can be used to distinguish Paleoproterozoic to mid-Mesoproterozoic (1.84–1.20 Ga) clasts derived from the central Arizona highlands region from clasts derived from younger sources that contain late Mesoproterozoic zircons (1.20–1.00 Ga). Characteristic paleomagnetic magnetizations were measured in 44 densely cemented orthoquartzite clasts, sampled from lower Miocene portions of the Sespe Formation in the Santa Monica and Santa Ana mountains and from a middle Eocene section in Simi Valley. Miocene Sespe clast inclinations define a bimodal population with modes near 15° and 45°. Eight samples from the steeper Miocene mode for which detrital zircon spectra were obtained all have spectra with peaks at 1.2, 1.4, and 1.7 Ga. One contains Paleozoic and Mesozoic peaks and is probably Jurassic. The remaining seven define a population of clasts with the distinctive combination of moderate to high inclination and a cosmopolitan age spectrum with abundant grains younger than 1.2 Ga. The moderate to high inclinations rule out a Mojave Desert–Death Valley or Sonoran region source population, and the cosmopolitan detrital zircon spectra rule out a central Arizona highlands source population. The Shinumo Formation, presently exposed only within a few hundred meters elevation of the bottom of eastern Grand Canyon, thus remains the only plausible, known source for the moderate- to high-inclination clast population. If so, then the Upper Granite Gorge of the eastern Grand Canyon had been eroded to within a few hundred meters of its current depth by early Miocene time (ca. 20 Ma). Such an unroofing event in the eastern Grand Canyon region is independently confirmed by (U-Th)/He thermochronology. Inclusion of the eastern Grand Canyon region in the Sespe drainage system is also independently supported by detrital zircon age spectra of Sespe sandstones. Collectively, these data define a mid-Tertiary, SW-flowing "Arizona River" drainage system between the rapidly eroding eastern Grand Canyon region and coastal California.

■ INTRODUCTION

Among the most difficult problems in geology is constraining the kilometer-scale erosion kinematics of mountain belts (e.g., Stüwe et al., 1994; House et al., 1998). A celebrated example of the problem, and the subject of vigorous contemporary debate, is the post–100 Ma erosion kinematics of the Colorado Plateau of western North America (e.g., Pederson et al., 2002), and especially of the Grand Canyon region (e.g., Flowers et al., 2008; Karlstrom et al., 2008, 2014; Polyak et al., 2008; Beard et al., 2011; Wernicke, 2011; Flowers and Farley, 2012; Flowers et al., 2015; Lucchitta, 2013; Hill and Polyak, 2014; Darling and Whipple, 2015; Fox et al., 2017; Winn et al., 2017). The erosion problem of the plateaus is particularly well posed. It was a broad cratonic region that lay near sea level for most of Paleozoic and Mesozoic time (e.g., Burchfiel et al., 1992). During the Late Cretaceous–Paleogene Laramide orogeny, the Cordilleran orogen roughly doubled in width. The Colorado Plateau and southern Rocky Mountains thus underwent a transition from residing near sea level, as a retroarc Cordilleran foreland basin during the Late Cretaceous, to a mountain belt residing at elevations of 1–2 km during Paleogene and younger time (e.g., Elston and Young, 1991; Flowers et al., 2008; Huntington et al., 2010; Karlstrom et al., 2014; Hill et al., 2016; Winn et al., 2017). The key challenge posed by this framework lies in using thermochronological information on the unroofing history, and the distribution of sedimentary source regions and corresponding depocenters, to constrain erosion kinematics.

Existing models of erosion kinematics of the region differ mainly in the role they assign to the modern Colorado River (ca. 6 Ma and younger) versus more ancient drainage systems dating back to Laramide time. Despite the lack of consensus, a significant and recent point of agreement, based primarily on thermochronological data, is that a kilometer-scale erosional unroofing event occurred in mid-Tertiary time (ca. 28–18 Ma) in the eastern Grand Canyon region (Fig. 1; Flowers et al., 2008; Lee et al., 2013; Karlstrom et al. 2014; Winn et al., 2017). This unroofing event (described in more detail in the next section)

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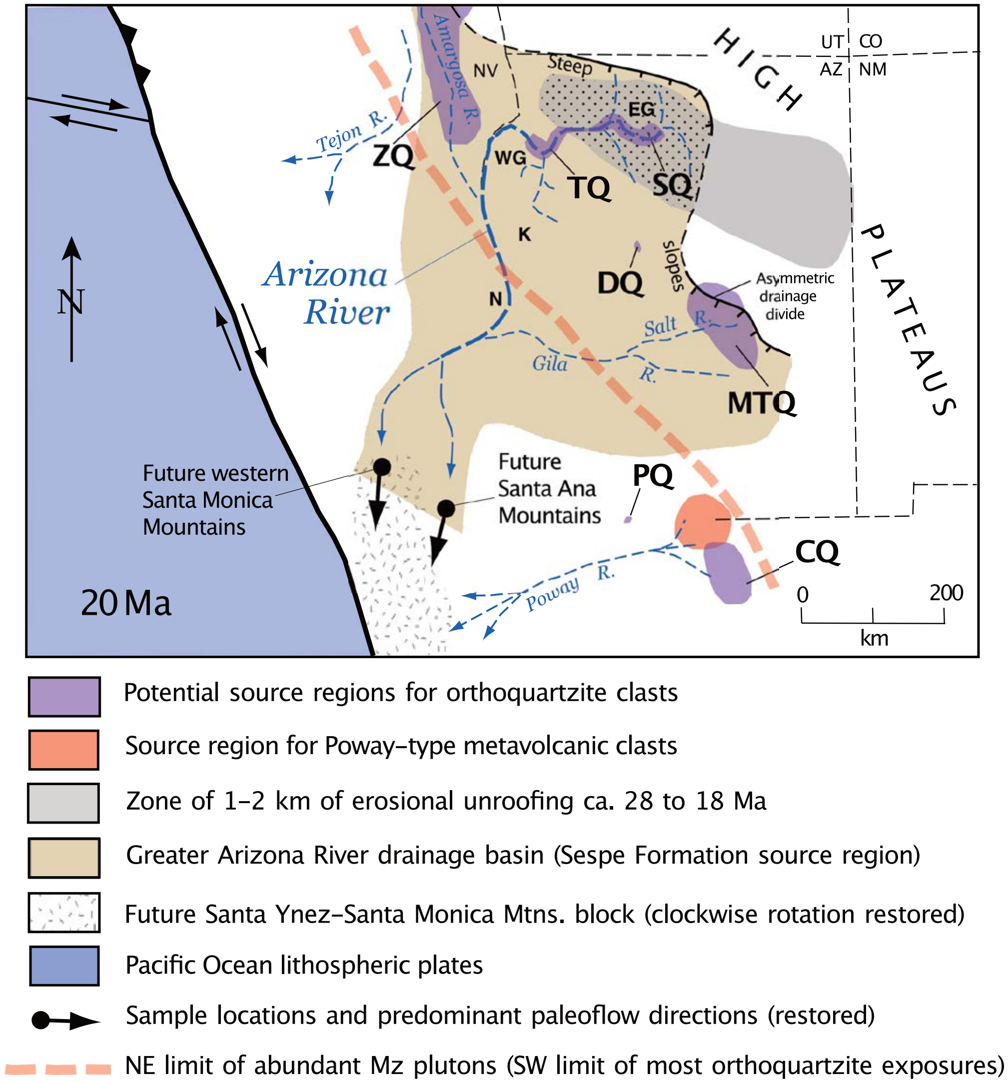

Figure 1. Geologic reconstruction, based on McQuarrie and Wernicke (2005), showing the early Miocene positions of Sespe Formation depocenters in the Santa Monica and Santa Ana mountains with dominant paleoflow directions, and the extent of the Sespe Formation source regions, as inferred by Howard (2000, 2006) and Ingersoll et al. (2018), but including a portion of the southwestern Colorado Plateau, after Wernicke (2011). Stippled area inside zone of 28–18 Ma erosional unroofing delimits 30,000 km² area potentially contributing detritus to the Piuma Member of the Sespe Formation. The four main regions of exposed orthoquartzite (purple) include: (1) Death Valley–Mojave region, with Lower Cambrian Zabriskie Formation (ZQ) and associated Neoproterozoic orthoquartzites; (2) Grand Canyon region, with Shinumo Formation (SQ) of Mesoproterozoic age in eastern Grand Canyon (EG), and quartzitic portions of the Tapeats Formation (TQ) of Cambrian age in western Grand Canyon (WG); (3) central Arizona highlands Paleo- to Mesoproterozoic rocks including Mazatzal, Tonto, and Hess Canyon groups (MTQ) and Del Rio Formation (DQ); (4) Neoproterozoic–Cambrian orthoquartzites (including clasts recycled in Jurassic conglomerates) in the Caborca area of Sonora, Mexico (CQ) and Mesoproterozoic quartzites at Sierra Prieta (PQ) in NW Sonora. Proposed paleorivers discussed in text shown in blue dashed lines. K—Kingman, Arizona; N—Needles, California; AZ—Arizona; CO—Colorado; NM—New Mexico; NV—Nevada; UT—Utah.

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is relatively localized compared with erosion histories of adjacent regions across orogenic strike to the SW and NE, also defined by thermochronological data. To the SE in the Arizona Transition Zone and Mojave-Sonora Desert region, unroofing to near-present levels occurred in Laramide time (ca. 80–40 Ma), with the exception of rocks tectonically exhumed by Tertiary extension (Bryant et al., 1991; Fitzgerald et al., 1991, 2009; Foster et al., 1993; Spotila et al., 1998; Blythe et al., 2000; Mahan et al., 2009). To the NE, in the Colorado Plateau interior, erosional unroofing occurred mainly after 10 Ma, presumably as a result of integration of the Colorado River drainage system at 6 Ma (e.g., Pederson et al., 2002; Flowers et al., 2008; Wernicke, 2011; Hoffman et al., 2011; Kimbrough et al., 2015; Karlstrom et al., 2017; Winn et al., 2017).

Independent of thermochronological data, constraints on erosion kinematics are imposed by the arrival of specific clast types within basins along the flanks, placing a minimum age on the time at which any particular clast type was exposed to erosion. The overall pattern of unroofing thus motivates examination of depocenters along the margins of the Cordillera for evidence of unroofing in the Cordilleran interior, such as migration of drainage divides toward the interior (e.g., Ingersoll et al., 2018). In particular, the mid-Tertiary unroofing event predicts the appearance of eroded detritus from the eastern Grand Canyon region in mid-Oligocene to early Miocene depocenters.

We investigate this hypothesis by applying a new technique that combines paleomagnetic inclination spectra and detrital zircon age spectra of conglomerate clast populations to the gravel fraction of the Sespe Formation, a mid-Tertiary conglomeratic sandstone interval that is broadly distributed throughout coastal southern California (Fig. 2) (Howard, 2000, 2006; Ingersoll et al., 2013, 2018). We focus on the orthoquartzite clast population (as opposed to volcanic, metavolcanic, and metaquartzite clasts also abundant in the Sespe Formation), because it is both ultradurable and its potential sources are widely exposed in the headwater regions of all proposed major paleodrainages tributary to the Sespe basin (Fig. 1). The scope of our study includes characteristic remnant magnetizations (CRMs) from 44 samples from the Sespe orthoquartzite clast population, collected from three well-dated Sespe exposure areas. We compare these data with CRMs of some 700 samples from potential source regions in the Death Valley–Mojave region, the central Arizona highlands, Grand Canyon, and Sonora, Mexico. Our study also includes 936 detrital zircon ages from 12 Sespe orthoquartzite clasts, which we compare to 1870 detrital zircon ages from 23 samples of potential sources.

GEOLOGIC BACKGROUND

Sespe Formation

The modern outcrop distribution of the Sespe Formation (Fig. 2) has been substantially modified by right-lateral shear on the San Andreas fault system and transrotation of the Western Transverse Ranges (e.g., Howard, 1996; Atwater and Stock, 1998). The mid-Tertiary configuration of the Sespe basin can be determined with a high degree of confidence on the basis of palinspastic reconstructions (e.g., Atwater and Stock, 1998; McQuarrie and Wernicke, 2005; Jacobson et al., 2011; Ingersoll et al., 2018), all of which restore the most proximal Sespe depocenters (Santa Monica and Santa Ana mountains) to a position near the modern Colorado River delta (Fig. 1).

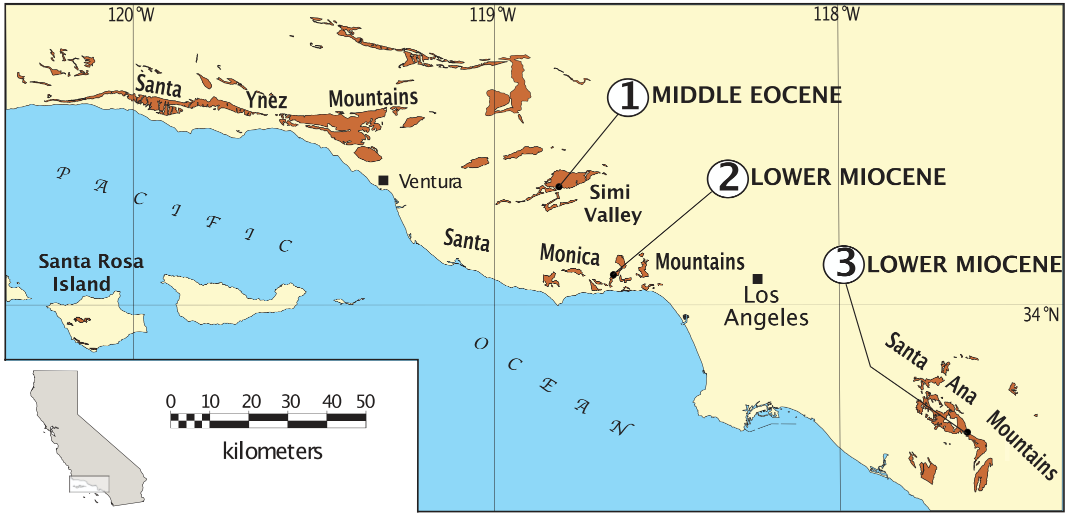

Figure 2. Map showing distribution of surface exposures of early to mid-Tertiary Sespe Formation (reddish-brown shading) in the Los Angeles region (after Lander, 2011) and sample localities (black dots) with Sespe depositional ages, including: 1—View Lane Drive locality in Simi Valley; 2—Piuma Road and Scheuren Road localities in the Santa Monica Mountains; and 3—Red Rock Trail in Limestone Canyon Park and Santiago Canyon Road localities in the Santa Ana Mountains (Tables 2 and 3).

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The middle Eocene to lower Miocene Sespe Formation consists predominantly of fluvial to deltaic sandstone and conglomerate, ranging from a few hundred up to 1000 m thick (e.g., Schoellhamer et al., 1981; Howard, 1989, 2000). Although much of the Sespe Formation appears to be Eocene, it also contains an Oligocene to early Miocene component that includes tongues of marine strata. The younger strata have locally been defined as the ca. 27–20 Ma Piuma Member, the upper part of which is paleontologically dated as Hemingfordian in the Santa Monica and Santa Ana mountains (e.g., Lander, 2011, 2013). Compositionally, Sespe sandstones are lithic-poor arkoses derived predominantly from granitic source rocks, with 50% to 95% of detrital zircon ages indicating provenance within the Mesozoic Cordilleran arc, and the remainder derived from various primary and recycled sources of pre–300 Ma grains (Ingersoll et al., 2013, 2018).

Sespe Formation conglomerates are dominated by populations of highly survivable volcanic, metavolcanic, and quartzitic clasts, with smaller populations of less durable rock types (Woodford et al., 1968; Abbott and Peterson, 1978; Belyea and Minch, 1989; Howard, 1989; Minch et al., 1989). The quartzite clast population can be subdivided into orthoquartzites and metaquartzites. Orthoquartzite is defined as an unmetamorphosed quartz arenite with a densely cemented silica matrix (Howard, 2005) and is distinguished from metaquartzite petrographically, due to the destruction of detrital grain boundaries beginning under sub-greenschist to lower greenschist-facies conditions (Wilson, 1973; Howard, 2005). Our focus on orthoquartzite is motivated by two key considerations.

First, crystalline sources tend to be proximal to the coast and consist mainly of feldspathic rock types that are only moderately durable, with the exception of ultradurable metarhyolite, chert, and metaquartzite clasts (e.g., Abbott and Peterson, 1978). It has long been established that orthoquartzite clasts in the Sespe Formation are derived from relatively distant sources within the Cordilleran interior (Howard, 1996, 2000), generally well NE of source regions for clasts of metaquartzites and most crystalline rocks (Fig. 1). Crystalline source regions also occur in the Cordilleran interior, but, given the moderate durability of crystalline clasts (owing to both the mechanical weakness of cleavage and solubility of feldspar), they tend to be eliminated from the gravel fraction during long transport, especially in the presence of ultradurable quartzitic clasts (e.g., Abbott and Peterson, 1978). Fingerprinting of orthoquartzite clasts in the basins thus affords a broad aperture for the observation of erosion kinematics using this approach (Howard, 1989, 2000). Second, one potential Sespe orthoquartzite source, the 1.1 Ga Shinumo Formation, is at present only exposed within a few hundred meters elevation of the bottom of eastern Grand Canyon, in the Upper Granite Gorge area (Fig. 3). Its appearance in the Sespe Formation would therefore constrain the time by which eastern Grand Canyon was in existence, more or less as it is today, greatly limiting the extant range of erosion models.

Orthoquartzite Source Regions

Eastern Grand Canyon is, however, only one of four potential source regions in the Cordilleran interior for orthoquartzite clasts (Fig. 1). The other three include

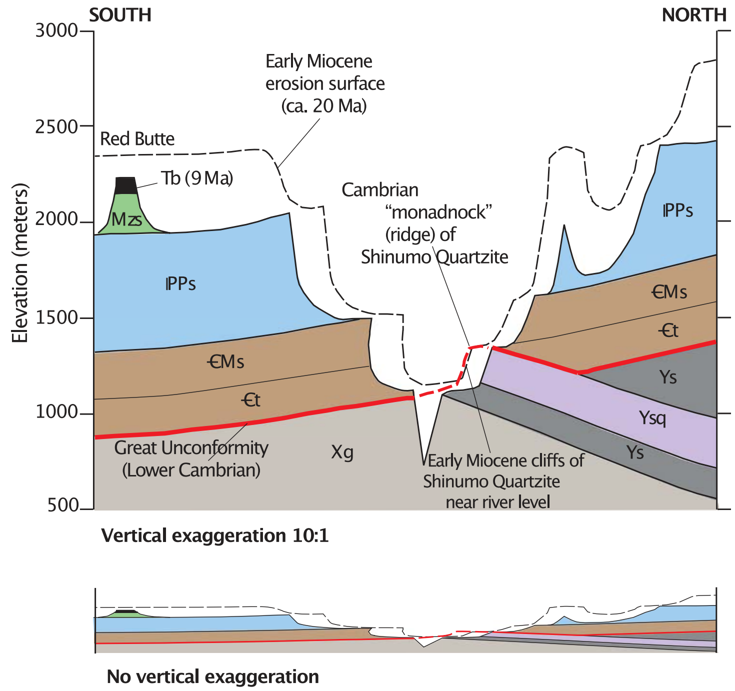

Figure 3. Generalized north-south cross section through the Upper Granite Gorge area of eastern Grand Canyon region showing the disposition of the Shinumo Formation (Ysq) relative to a nominal early Miocene erosion surface. Xg—Paleoproterozoic gneiss; Ys—Mesoproterozoic strata; Ꞓt—Cambrian Tonto Group; ꞒMs—Cambrian through Mississippian strata; ℙPs—Pennsylvanian through Permian strata; Mzs—Mesozoic strata; and Tb—Tertiary basalt.

(1) the Death Valley–Mojave region, which contains Neoproterozoic–Cambrian orthoquartzites (e.g., Stewart et al., 2001; Schoenborn et al., 2012); (2) the central Arizona highlands, which contain late Paleoproterozoic to mid-Mesoproterozoic orthoquartzites (e.g., Doe et al., 2012; Mulder et al., 2017); and (3) the Caborca area of NW Sonora, Mexico, which contains Neoproterozoic–Cambrian orthoquartzites in strata correlative with the Death Valley–Mojave strata (Gehrels and Stewart, 1998; Stewart et al., 2001). In the broader Sonoran region (mainly south of the area shown in Fig. 1), widespread exposures of Jurassic conglomerates (Coyotes Formation and equivalents) contain orthoquartzite clasts of presumed Proterozoic–early Paleozoic age (Stewart and Roldán-Quintana, 1991). In NW Sonora, the only known Mesoproterozoic quartzites, which may or may not be orthoquartzite, occur in a small exposure (6.5 km²) at Sierra Prieta (Fig. 1), where they are intruded by ca. 1.08 Ga anorthosite sills (Izaguirre-Pompa and Iriondo, 2007; Molina-Garza and Izaguirre, 2006).

Various Tertiary paleodrainages have been proposed to connect these potential source regions with mid-Tertiary coastal basins in southern California

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(Howard, 2000, 2006; Ingersoll et al., 2018). These include the Poway (Abbott and Smith, 1989), Amargosa (Howard, 2000), Gila (Howard, 2000), Arizona (Wernicke, 2011), and Tejon (Lechler and Niemi, 2011) paleodrainage systems (Fig. 1).

To distinguish among these source regions, we augment previous studies of orthoquartzite clasts and sources (Howard, 1989, 1996, 2000, 2006) with a novel method, using the combination of paleomagnetic inclination and detrital zircon spectra of orthoquartzite clast populations, to trace provenance (Wernicke et al., 2010, 2012; Wernicke, 2011; Raub, 2013). A key finding from the earlier conglomerate studies was that lowest Sespe sources appear to be dominated by a Gila paleodrainage system, which included (1) Paleoproterozoic orthoquartzites from the central Arizona highlands and (2) metarhyolite clasts derived from southeastern Arizona. The system appears to have evolved by Oligocene time into a more latitudinally extensive system to include a component of metavolcanic and orthoquartzite clasts from the Death Valley–Mojave region (Howard, 2000, 2006).

The Sespe Formation and its Eocene equivalent in the San Diego area, the Poway Group, differ markedly in their clast composition, with the Poway Group being dominated by metarhyolite clasts (Woodford et al., 1968, 1972; Belyea and Minch, 1989). The Poway Group averages 73% quartz porphyry metarhyolite clasts (Bellemin and Merriam, 1958). These "Poway-type" metarhyolite clasts have been texturally and geochemically traced to bedrock sources in the Caborca region of Sonora, Mexico (Fig. 1) (Abbott and Smith, 1989). The Sespe Formation, in contrast, contains a much smaller percentage (<10%) of metarhyolite clasts, which are petrographically and geochemically dissimilar to Poway-type clasts and Sonora metarhyolites, but are similar to Jurassic metarhyolites from the Mount Wrightson Formation of southeastern Arizona (Abbott et al., 1991). These relations are generally interpreted to indicate that the Poway Group and Sespe Formation represent distinct Eocene drainage basins (Woodford et al., 1968, 1972; Kies and Abbott, 1983; Belyea and Minch, 1989; Abbott et al., 1991; Howard, 2000, 2006). Although there may be some overlap of the two source areas (e.g., Ingersoll et al., 2018), transport of significant quantities of Caborca-area orthoquartzites (either Mesoproterozoic Sierra Prieta or Neoproterozoic–Cambrian strata, Fig. 1) in a regional drainage system of any age would also result in a preponderance (≥3:1) of Poway-type clasts relative to the orthoquartzite component, as suggested by the clast composition of the Poway Group. The lack of Sonora-derived metarhyolite clasts in the Sespe drainage basin thus strongly suggests the absence of any significant drainage connection between NW Sonora and the Sespe basin.

Two key attributes have the potential to distinguish between a population of clasts with Shinumo provenance from populations derived from Death Valley–Mojave or central Arizona highlands sources: (1) moderate to high paleomagnetic inclination and (2) the presence of late Mesoproterozoic (1.3–1.0 Ga) or "Grenville-age" detrital zircon. Whereas orthoquartzite populations from the Death Valley–Mojave region generally contain abundant 1.3–1.0 Ga detrital zircons, their CRMs are of low inclination (0°–30°), contrasting them with the Shinumo population. Whereas orthoquartzite populations from the central Arizona highlands may have moderate to high inclinations, they are mostly too old to contain 1.3–1.0 Ga detrital zircons, distinguishing them from the Shinumo population. Therefore, identification of these attributes within a population of Sespe orthoquartzite clasts has the potential to distinguish a Shinumo source from the other sources. If the Shinumo Formation is a Sespe gravel source, it would strengthen the "Arizona River" hypothesis (Wernicke, 2011), independent of low-temperature thermochronometry studies on which it is based (e.g., Flowers et al., 2008, 2015; Wernicke, 2011; Flowers and Farley, 2012, 2013). According to this hypothesis, the mid-Tertiary drainage configuration of the Cordillera included a paleoriver system with headwaters cut near the modern level of erosion of the Upper Granite Gorge area in the eastern Grand Canyon region.

Below, we present paleomagnetic and detrital zircon data from three Sespe clast populations and one potential source rock from the Shinumo Formation, as well as a compilation of existing paleomagnetic and detrital zircon data from the literature. We then compare data from the various source populations with data from Sespe clast populations, focused on the issue of which, if any, of the Sespe clast populations indicate a Shinumo provenance.

Mid-Tertiary (28–18 Ma) Unroofing of the Southwestern Colorado Plateau

As noted above, the primary erosional event in the Cordilleran interior during upper Sespe (Piuma) time occurred within a NW-trending zone, running from the eastern Grand Canyon region through east-central Arizona (Fig. 1), contrasting it with predominantly Laramide unroofing to the SW in the Mojave-Sonoran region and post–10 Ma unroofing to the NE on the Colorado Plateau. In addition to thermochronological data, this event is recorded by kilometer-scale erosion between aggradation of the Eocene to lower Oligocene Chuska Formation and aggradation of the Miocene Bidahochi Formation, whose ages bracket the unroofing event between 26 and 16 Ma (Cather et al., 2008). Numerous thermochronological cooling models indicate ~30 °C of cooling at that time, from ~60 °C prior to 28 Ma (with some interpretations of the data suggesting temperatures as high as 80–90 °C in the Upper Granite Gorge prior to 28 Ma) to <30 °C after 18 Ma (Flowers et al., 2008; Flowers and Farley, 2012; Lee et al., 2013; Karlstrom et al., 2014; Winn et al., 2017).

In the Upper Granite Gorge of eastern Grand Canyon, where the Shinumo Formation is exposed (Fig. 3), the 30 °C (or less) temperatures at the end of the 28–18 Ma erosion event were probably very close to surface temperatures in the SW United States, indicating very little post–18 Ma erosion (Flowers et al., 2008; Flowers and Farley, 2012, 2013; Wernicke, 2011; Karlstrom et al., 2014; Winn et al., 2017). Modern surface temperatures measured throughout the interior of the SW United States (Sass et al., 1994) vary according to

$$T_s(h) = (29 \pm 2)°C + (-8 \pm 1\ °C/km)h, \tag{1}$$

where T_s is surface temperature, and h is elevation above sea level (equation 7 in Wernicke, 2011). Early Miocene surface temperatures were at least 3 °C, and

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perhaps as much as 8 °C, warmer than today (e.g., Huntington et al., 2010). Hence, assuming no erosion, rocks now exposed at a modern elevation of 600 m at the bottom of eastern Grand Canyon would have T_s in the range of 27 °C to 32 °C, depending on the degree of atmospheric cooling since 20 Ma. However, some additional erosion must have occurred after the 28–18 Ma unroofing event. Given a very conservative upper-temperature limit for river-level samples of 40 °C after mid-Tertiary erosion ended (see discussion of error sources for these estimates in Wernicke, 2011, p. 1303–1305) and an early Miocene upper-crustal geothermal gradient of 25 °C/km (based on thermochronometric profiles through tilted fault blocks in the eastern Lake Mead region; e.g., Quigley et al., 2010, and discussion on p. 1295 in Wernicke, 2011), net erosion since 18 Ma would lie in the range:

$$(8-13\ °\mathrm{C}) / (25\ °\mathrm{C/km}) / (1000\ \mathrm{m/km}) = 320-520\ \mathrm{m}, \quad (2)$$

which corresponds to a maximum average regional erosion rate of 18–29 m/m.y.

This erosion rate for the bottom of eastern Grand Canyon is in good agreement with the late Tertiary erosional history of the surrounding plateau region based on stratigraphic constraints. Just south of eastern Grand Canyon, the basalt at Red Butte, which lies on an erosion surface 220 m above the surrounding Coconino Plateau, is 9 Ma (Reynolds et al., 1986), indicating an average erosion rate of 24 m/m.y. since then (Fig. 3). East of Grand Canyon, average regional erosion since 16 Ma (i.e., regional unroofing below the basal Bidahochi unconformity) is at most 300–400 m (e.g., figure 15 in Cather et al., 2008), suggesting rates of 19–25 m/m.y., albeit much of the erosion may have been concentrated in the past 6 m.y. at higher rates (Karlstrom et al., 2017).

In the Upper Granite Gorge area, the Shinumo Formation is the most erosionally resistant unit within the gently north-tilted Grand Canyon Supergroup. It is the only stratified unit in eastern Grand Canyon that contains abundant ultradurable orthoquartzite. It eroded into steep, south-facing cuestiform ridges, during both Cenozoic erosion and erosion prior to the Cambrian Sauk transgression, when it formed a series of paleo-islands (Fig. 3). The Cambrian paleo-islands rose 100–200 m above the coastal plain, around which Tonto Group strata, including sandstones of the Tapeats Formation, were deposited in buttress unconformity (Fig. 3; Noble, 1910, 1914; Sharp, 1940; McKee and Resser, 1945; Billingsley et al., 1996; Karlstrom and Timmons, 2012). At present, the Shinumo Formation crops out in a 70-km-long, quasi-linear array of seven exposure areas, with each area 2–5 km long, as measured parallel to the array, mostly on the north side of the modern Colorado River (e.g., figure 3.1 in Hendricks and Stevenson, 2003). The Shinumo Formation is now preserved at elevations as much as 600 m above the modern river level (Billingsley et al., 1996). If our estimate of 300–500 m of post–18 Ma erosion is correct, the Shinumo Formation would have been a highly proximal source of ultradurable, gravel-sized clasts in the high-relief headwaters of a mid-Tertiary Arizona River (Fig. 3).

A second significant source of orthoquartzite in the Grand Canyon region is the Tapeats Formation, but only in the Lower Granite Gorge area of western Grand Canyon (Fig. 1), where it is the oldest exposed stratified unit. In eastern Grand Canyon, exposures of the Tapeats Formation, in contrast to much of the Shinumo Formation, are not densely cemented orthoquartzites (Billingsley et al., 1996). In the Lower Granite Gorge area, however, a large fraction of the Tapeats Formation is "quartzitic and very hard," in contrast to relatively weak sandstones in the remainder (p. 16 in McKee and Resser, 1945).

SAMPLING AND METHODS

We sampled Sespe gravel clasts from the Santa Ana and Santa Monica mountains and from Simi Valley (Fig. 2). We also collected several samples of potential source rocks, in order to reproduce results from extensive existing paleomagnetic and detrital zircon data (Elston and Grommé, 1994; J. Hagstrum, written commun., 2017; Bloch et al., 2006; Mulder et al., 2017), including one sample of the Shinumo Formation and one sample each of the Shinumo and Tapeats formations from the Caltech sample archive (Table 1).

Because dated Sespe sections range broadly in age, from middle Eocene to early Miocene (ca. 48–20 Ma), sample locations (Fig. 2) were restricted to three sections with local paleontological, radiometric, and magnetostratigraphic control of depositional age. They included (1) a middle Eocene section in Simi Valley (exposed along View Lane Drive at the terminus of exit 22A of California Highway 118; Kelly et al., 1991; Kelly and Whistler, 1994; Lander, 2013); (2) the lower Miocene Piuma Member in the Saddle Peak area of the western Santa Monica Mountains (exposed along upper Piuma Road and upper Schueren Road, along and near the range crest) (Lander, 2011, 2013); and (3) correlative lower Miocene strata in the Limestone Canyon Park area of the Santa Ana Mountains (Red Rock Canyon Trail and a nearby road cut through the "marker conglomerate" horizon (Belyea and Minch, 1989) on Santiago Canyon Road (Fig. 2).

In these areas of exposure, in situ paleomagnetic sampling of orthoquartzite clasts in quantity proved to be unfeasible, precluding a conglomerate test. Steep badlands topography along ridge-crest exposures of the Sespe

TABLE 1. COLLECTED SAMPLES FROM GRAND CANYON SOURCES

Sample number	Location: Latitude (°N)	Location: Longitude (°W)	Formation
Grand Canyon, South Kaibab Trail			
IC-1-35*†§	~36°05′30″N	~112°05′20″W	Shinumo
Grand Canyon, Clear Creek Trail			
IC-503-35*	~36°06′20″N	~112°04′50″W	Tapeats
Grand Canyon, River Mile 75			
JG-01-09*	36°03′15.6″N	111°54′03.37″W	Shinumo

*Petrography
†Paleomagnetic analysis
§Detrital zircon analysis

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Formation results in a scarcity of exposed orthoquartzite clasts in outcrops that are both sufficiently indurated and accessible for in situ drilling. Orthoquartzite clasts were mainly sampled from thin, proximal colluvial deposits within a few meters of their Sespe bedrock sources. As discussed further below, the results of Hillhouse (2010) and this study indicate that the CRMs of Sespe orthoquartzite clasts predate weathering, transport, and deposition of the clasts and diagenesis of their sandstone matrix.

A total of 92 Sespe clasts were collected, including 71 from the Miocene sections (30 from Piuma Road, 19 from Schueren Road, and 22 from the Santa Ana Mountains) and 21 from the Eocene section (Table 2). Following petrographic screening (mainly to distinguish orthoquartzites from metaquartzites and other rock types) and assessment of the quality of preserved stratification (often best observed on cut or drilled surfaces; Figs. 4 and S1[1] shows representative examples), 49 samples were selected for paleomagnetic analysis. These included 34 samples from Miocene Sespe sections (17 from Piuma Road, 13 from Schueren Road, and four from the Santa Ana Mountains) and 15 samples from the Eocene Sespe section. All 34 samples from the Miocene Sespe Formation yielded interpretable paleomagnetic data, but only ten of the 15 samples from the Eocene section yielded interpretable data. We therefore report paleomagnetic data for a total of 44 Sespe orthoquartzite clasts (Tables 3 and 4; Table S1[2]).

Our general approach is to compare the distribution of inclinations within clast populations with those of potential source regions, which requires comparison of inclination-only data from the clast populations with three-dimensional paleomagnetic vectors of the source populations. Whereas the latter can be expressed using Fisher statistics, the former cannot, and at present there is no parametric test of statistical distributions applicable to such comparisons (McFadden and Reid, 1982; p. 135 in Fisher et al., 1987). Further, we cannot rigorously define any sort of mean for our clast populations, because as shown below, the clast populations are not normally distributed.

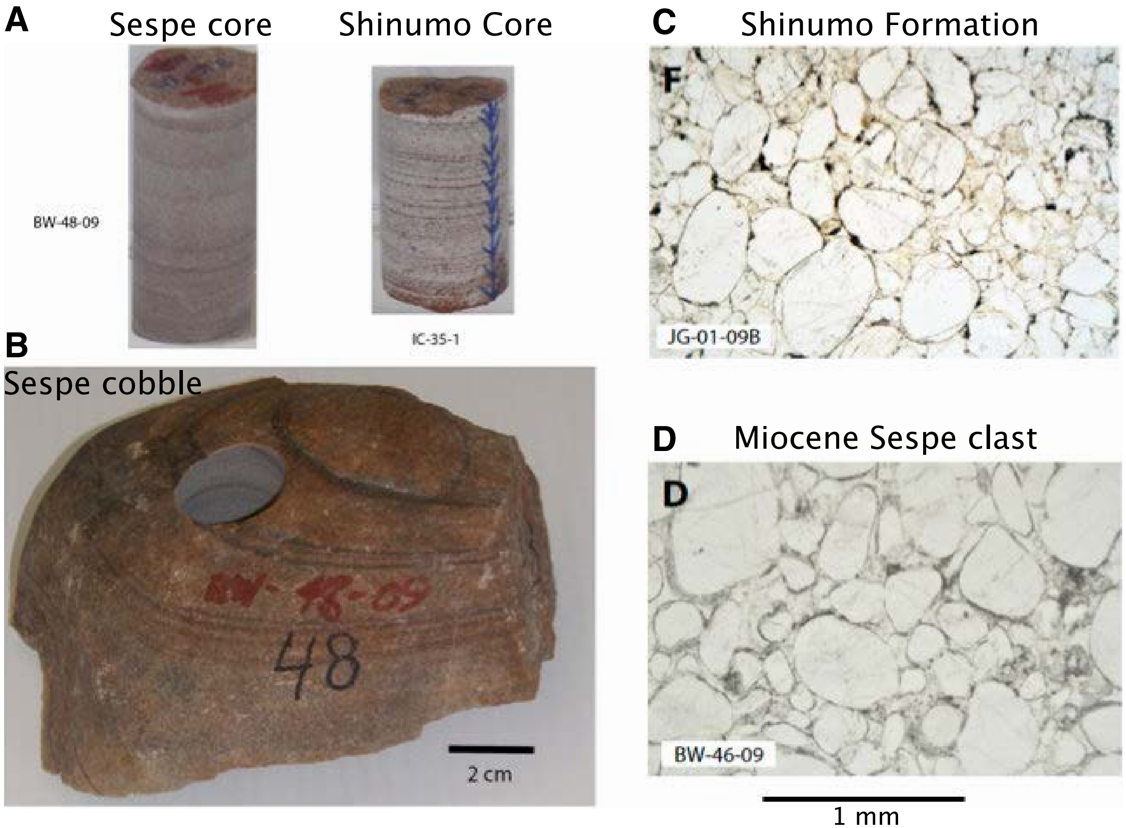

Figure 4. (A) Photograph of paleomagnetic cores of orthoquartzites, drilled perpendicular to bedding, from a Miocene Sespe Formation clast (left) and a bedrock sample of the Shinumo Formation (right). (B) Photograph of a Sespe Formation orthoquartzite cobble showing sedimentary lamination and drill-hole for left-hand paleomagnetic core shown in (A). (C, D) Photomicrographs of orthoquartzites from the Shinumo Formation and a clast from the Miocene Sespe Formation, respectively.

[1]Supplemental Materials. Figure S1. Photographs of seven representative clasts showing stratification. Figure S2. Zijderveld demagnetization plots for all paleomagnetic data. Figure S3. Photomicrographs of selected samples. Figure S4. Relationship between cross-stratification and paleomagnetic inclination in Neoproterozoic–Cambrian and Shinumo strata. Text S1. Discussion of effect of primary structures on paleomagnetic inclination spectra. Please visit https://doi.org/10.1130/GES02111.S1 or access the full-text article on www.gsapubs.org to view the Supplemental File.

[2]Supplemental Tables. Table S1: Columns show sample number, measurement type (alternating field—AF; thermal—TT), field strength (mT) or temperature (T), declination and inclination, publicly available in full form at the MagIC Data Repository (https://earthref.org/MagIC/16684). Table S2: Sheets in Excel file include detrital zircon ages from LaserChron and Apatite to Zircon of Sespe orthoquartzite clasts and Shinumo Formation, publicly available in California Institute of Technology Research Data Repository (https://data.caltech.edu/records/1245). Please visit https://doi.org/10.1130/GES02111.S2 or access the full-text article on www.gsapubs.org to view the Supplemental Tables.

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TABLE 2. COLLECTED SAMPLES OF SESPE CLASTS FROM SOUTHERN CALIFORNIA

Sample number	Location		Age
	Latitude (°N)	Longitude (°W)	
Piuma Road, Malibu			
BW-01-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-02-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-03-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-04-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-05-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-06-09*†§	34°04′13.0″N	118°39′59.86″W	Miocene
BW-07-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-08-09	34°04′13.0″N	118°39′59.86″W	Miocene
BW-16-14*†§	34°4′20.25″N	118°39′29.08″W	Miocene
BW-17-14*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS01*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS02*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS03*	34°4′20.25″N	118°39′29.08″W	Miocene
14LS04*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS05*	34°4′20.25″N	118°39′29.08″W	Miocene
14LS06*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS07*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS08*†§	34°4′20.25″N	118°39′29.08″W	Miocene
14LS09*†§	34°4′20.25″N	118°39′29.08″W	Miocene
14LS10*	34°4′20.25″N	118°39′29.08″W	Miocene
14LS11*†§	34°4′20.25″N	118°39′29.08″W	Miocene
14LS12*†§	34°4′20.25″N	118°39′29.08″W	Miocene
14LS13*	34°4′20.25″N	118°39′29.08″W	Miocene
14LS14*†	34°4′20.25″N	118°39′29.08″W	Miocene
14LS15*†	34°4′12.22″N	118°40′5.59″W	Miocene
14LS16*	34°4′12.22″N	118°40′5.59″W	Miocene
14LS17*	34°4′12.22″N	118°40′5.59″W	Miocene
14LS18*	34°4′12.22″N	118°40′5.59″W	Miocene
14LS19*†	34°4′12.22″N	118°40′5.59″W	Miocene
14LS20*†	34°4′12.22″N	118°40′5.59″W	Miocene
Santiago Canyon Road			
BW-11-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-12-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-13-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-14-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-15-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-16-09*†§	33°42′9.0″N	117°38′31.4″W	Miocene
BW-17-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-18-09*†§	33°42′9.0″N	117°38′31.4″W	Miocene
BW-19-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-20-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-21-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-22-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-23-09	33°42′9.0″N	117°38′31.4″W	Miocene
BW-24-09	33°42′9.0″N	117°38′31.4″W	Miocene
Red Rock Trail, Limestone Canyon Park			
BW-46-09*†§	33°42′10.3″N	117°38′56.65″W	Miocene
BW-47-09	33°42′10.3″N	117°38′56.65″W	Miocene
BW-48-09*†§	33°42′10.3″N	117°38′56.65″W	Miocene
BW-49-09	33°42′10.3″N	117°38′56.65″W	Miocene
BW-50-09	33°42′10.3″N	117°38′56.65″W	Miocene
BW-51-09	33°42′10.3″N	117°38′56.65″W	Miocene
BW-52-09	33°42′10.3″N	117°38′56.65″W	Miocene
BW-53-09	33°42′10.3″N	117°38′56.65″W	Miocene
Schueren Road, Malibu			
15LS01†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS02†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS03†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS04	34°4′42.78″N	118°38′57.60″W	Miocene
15LS05	34°4′42.78″N	118°38′57.60″W	Miocene
15LS06†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS07†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS08†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS09	34°4′42.78″N	118°38′57.60″W	Miocene
15LS10†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS11	34°4′42.78″N	118°38′57.60″W	Miocene
15LS12†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS13†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS14†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS15†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS16†	34°4′42.78″N	118°38′57.60″W	Miocene
15LS17	34°4′49.18″N	118°38′49.61″W	Miocene
15LS18	34°4′49.18″N	118°38′49.61″W	Miocene
15LS19†	34°4′49.18″N	118°38′49.61″W	Miocene
Simi Valley, Ventura County			
16LS01†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS02†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS03†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS04	34°17′9.97″N	118°47′35.11″W	Eocene
16LS05†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS06†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS07†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS08†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS09	34°17′9.97″N	118°47′35.11″W	Eocene
16LS10†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS11	34°17′9.97″N	118°47′35.11″W	Eocene
16LS12†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS13†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS14†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS15	34°17′9.97″N	118°47′35.11″W	Eocene
16LS16†	34°17′9.97″N	118°47′35.11″W	Eocene
16LS17	34°17′9.97″N	118°47′35.11″W	Eocene
16LS18	34°17′9.97″N	118°47′35.11″W	Eocene
16LS19†§	34°17′9.97″N	118°47′35.11″W	Eocene
16LS20†§	34°17′9.97″N	118°47′35.11″W	Eocene
16LS21†	34°17′9.97″N	118°47′35.11″W	Eocene

*Petrography
†Paleomagnetic analysis
§Detrital zircon analysis

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TABLE 3. SUMMARY OF ANALYSES PERFORMED ON SESPE CLAST SAMPLES

Location	Number collected	Stratified orthoquartzites	Interpretable paleomagnetic vector	Zircon analyses
Miocene Sespe				
Santa Monica Mountains				
Piuma Road	30	17	17	6
Schueren Road	19	13	13	0
Santa Ana Mountains				
Limestone Canyon Park	8	2	2	2
Santiago Canyon Road	14	2	2	2
Eocene Sespe				
Simi Valley Landfill	21	15	10	2
Total	92	49	44	12

TABLE 4. SUMMARY OF PALEOMAGNETIC RESULTS

Clast	Inclination (°)	Maximum angular deviation	Peak temperature (°C)	Location	
				Latitude (°N)	Longitude (°W)
South Kaibab Trail, Grand Canyon					
IC-1-35†	54.9	7.7	672	36.0917	112.0889
Bolero Lookout–Santiago Cyn Road					
BW16-09†	17.1	4.2	660	33.702500	117.642056
BW18-09†	27	6.4	672	33.702500	117.642056
Red Rock Trail, Limestone Canyon Park					
BW46-09†	51.6	4.9	500	33.702861	117.649069
BW48-09†	55	9.8	672	33.702861	117.649069
Saddle Peak–Piuma Road					
BW0609†	56.6	3.8	672	34.070278	118.666628
14LS01*	53.1	2.2	640–680	34.072292	118.658078
14LS02	17.6	12.7	650	34.072292	118.658078
14LS04*	21.2	7.6	650–660	34.072292	118.658078
14LS06	5.5	8.9	650	34.072292	118.658078
14LS07	7.0	2.2	670	34.072292	118.658078
14LS08*†	42.3	1.3	670–680	34.072292	118.658078
14LS09*†	68.7	1.7	670	34.072292	118.658078
14LS11†	43.6	5.3	660	34.072292	118.658078
14LS12†	43.4	1.7	660	34.072292	118.658078
14LS14	48.2	13.0	650	34.072292	118.658078
14LS15*	38.8	12.1	615–630	34.070061	118.668219
14LS17*	23.5	2.5	575–585	34.070061	118.668219
14LS19	13.6	9.0	555	34.070061	118.668219
14LS20	13.2	6.4	650	34.070061	118.668219
14BW16*†	58.5	2.2	680	34.072292	118.658078
14BW17	43.8	11.9	565	34.072292	118.658078
Saddle Peak–Schueren Road					
15LS01*	17.9	7.0	650	34.078550	118.649333
15LS02*	4.3	2.8	615–650	34.078550	118.649333
15LS03	24.9	4.9	670	34.078550	118.649333
15LS06	77.2	7.5	600	34.078550	118.649333
15LS07	6.2	7.2	500	34.078550	118.649333
15LS08	36.6	1.6	660	34.078550	118.649333
15LS10	64.6	4.2	575	34.078550	118.649333
15LS12*	45.8	3.3	400–450	34.078550	118.649333
15LS13	14.4	1.3	580	34.078550	118.649333
15LS14	17.5	5.9	500	34.078550	118.649333
15LS15	40.1	7.0	515	34.078550	118.649333
15LS16	15.0	5.3	350	34.078550	118.649333
15LS19	3.2	1.8	585	34.080328	118.647114
Simi Valley					
16LS01*	12.8	1.2	670	34.286103	118.793086
16LS02	45.6	3.0	515	34.286103	118.793086
16LS06	18.7	8.0	660	34.286103	118.793086
16LS08*	50.0	0.7	640	34.286103	118.793086
16LS09	1.9	0.9	670	34.286103	118.793086
16LS12	0.4	7.1	545	34.286103	118.793086
16LS13	16.4	5.6	575	34.286103	118.793086
16LS16	31.2	4.2	650	34.286103	118.793086
16LS19*†	41.5	2.2	640	34.286103	118.793086
16LS20†	44.8	4.8	650	34.286103	118.793086

*Multiple cores
†Zircon analysis

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Following paleomagnetic analysis, detrital zircon spectra were determined for a subset of 12 of the 44 Sespe clast samples. This subset was selected based on quality of paleomagnetic data (good orientation statistics and demagnetization temperatures suggestive, in most cases, of hematite as the carrier phase) and included two samples with low inclination and ten samples with moderate to high inclination. Of the ten with moderate to high inclination, eight were from the Miocene Sespe, and two were from the Eocene. The two samples with low inclination were both from the Miocene Sespe, from the road cut on Santiago Canyon Road (Table 4).

Paleomagnetic Analysis

All selected Sespe orthoquartzite clasts and the Shinumo sample were cut along their bedding planes with a non-magnetic brass blade, and then they were cored in the laboratory using an electric drill with a nonmagnetic bit. Sample cores were soaked in dilute HCl for up to 36 h to remove any possible fluid-related magnetic signatures and then stored in a magnetically shielded room.

Demagnetization and paleomagnetic measurements were carried out at the California Institute of Technology Paleomagnetics Laboratory using 2G™ Enterprises rock magnetometers with three-axis DC SQuID sensors with sensitivities of 2×10^{-13} Am^2 per axis, using a RAPID automatic sample changer. Details of the equipment and demagnetization procedures are described in Kirschvink et al. (2008). After measuring the natural remnant magnetization (NRM), we used five alternating field (AF) steps of 2–10 mT to remove viscous components of multi-domain magnetite and other soft magnetic components. To thermally demagnetize our samples, we heated them in a magnetically shielded ASC furnace in steps of 5–50 °C, from 0 °C up to a maximum of 710 °C to constrain the CRM. Magnetization components were defined by least squares using the principal component analysis technique of Kirschvink (1980) and software of Jones (2002).

Detrital Zircon Analysis

Mineral separations and U-Pb isotopic analyses were performed for a total of 13 samples, 12 from Sespe clasts and one from the Shinumo Formation. Six of these samples, including four samples from the Santa Ana Mountains, one sample from the Santa Monica Mountains, and one sample of Shinumo Formation (Tables 1 and 2) were separated and analyzed by Apatite to Zircon, Inc., using standard separation techniques and laser ablation–inductively coupled mass spectrometry. Analysis and preparation of zircon age data followed procedures described in Moore et al. (2015). For the seven additional samples, including five from the Santa Monica Mountains and two from the Simi Valley area, zircon extractions, using standard techniques, were performed at the California Institute of Technology and the University of Arizona. U-Pb analyses were performed at the University of Arizona Laserchron Center. Zircon grains were mounted in epoxy with Sri Lanka, FC-1, and R33 primary standards. The epoxy mount was sanded down to 20 µm, polished, and imaged with a Hitachi 3400N scanning electron microscope (SEM). Laboratory procedures for U-Pb isotopic analyses and screening for discordant grains follow methods described in Gehrels et al. (2006, 2008) and Gehrels and Pecha (2014).

RESULTS

Paleomagnetic Data

Demagnetization data for all samples are summarized in Table 4 and presented in complete form in Table S1 (footnote 2). Demagnetization plots for all samples are shown in Figure S2 (footnote 1). Representative demagnetizations of Sespe cobbles, including two from Miocene (Figs. 5A and 5B) and two from Eocene (Figs. 5C and 5D) sections, show well-preserved, high-temperature CRMs of moderate to high inclination. Measured remnant magnetizations of the sample suite have intensities ranging from 10^{-9} to 10^{-6} Am^2, well above instrument sensitivity of 10^{-13} Am^2. Up to five steps of alternating field (AF) demagnetization in 20 mT increments up to 100 mT generally had little effect on remanence, indicating magnetite is not a significant carrier. Characteristic directions in most samples are defined by multiple demagnetization steps ranging from 590 to 670 °C, suggesting that hematite is the main carrier of magnetization in these samples. This observation is consistent with petrographic evidence that samples typically contain pigmentary hematite, which imparts their characteristic red and red-purple hues (Fig. S1 [footnote 1]). However, in 15 of the 44 samples with interpretable data, the carrier phases were magnetite or other lower-temperature phases. Maximum angular deviations (MADs) calculated from principal component analysis average ~5° in our sample set (Table 4).

Distributions of paleomagnetic inclination from the Sespe clast populations, plotted in Figure 6 in 4° bins, show that both Miocene and Eocene populations exhibit bimodal distributions with maxima near 15° and 45° and minima near 30° (Fig. 6). The Miocene population, however, has a stronger peak near 45°, and the Eocene population has a stronger peak near 15°, although the latter population includes only ten samples. For the data set as a whole, only three of 44 samples lie in the three bins between 24° and 36°. By comparison, the three bins between 12° and 24° contain 13 samples, and the three bins between 36° and 48° contain 11 samples.

In addition to the new data, we compiled existing paleomagnetic data from possible source regions (references provided in Table 5), which we present as (1) directions from individual, demagnetized sample cores, corrected for bedding tilt (Fig. 7) and (2) histograms showing spectra of inclinations (Fig. 7). The compilation is limited to Neoproterozoic–Cambrian strata from the Death Valley–Mojave region, the Caborca region, the Shinumo and Tapeats formations in Grand Canyon, and the Tapeats Formation and equivalents in the

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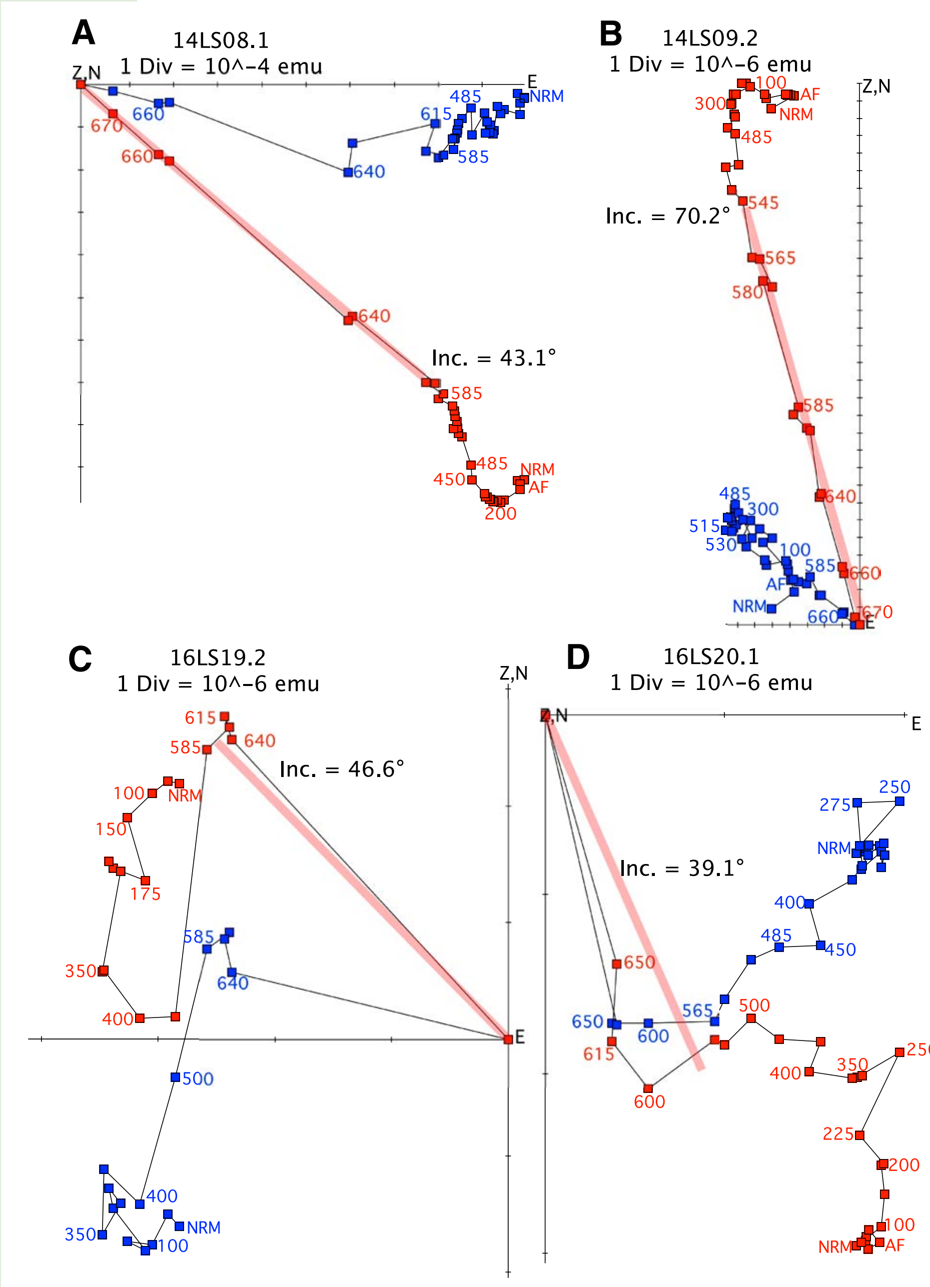

Figure 5. Zijderfeld plots showing thermal demagnetization histories of samples of orthoquartzite clasts from Miocene (A, B) and Eocene (C, D) Sespe Formation conglomerates. Detrital zircon spectra were determined for all four samples, as annotated on Figure 8.

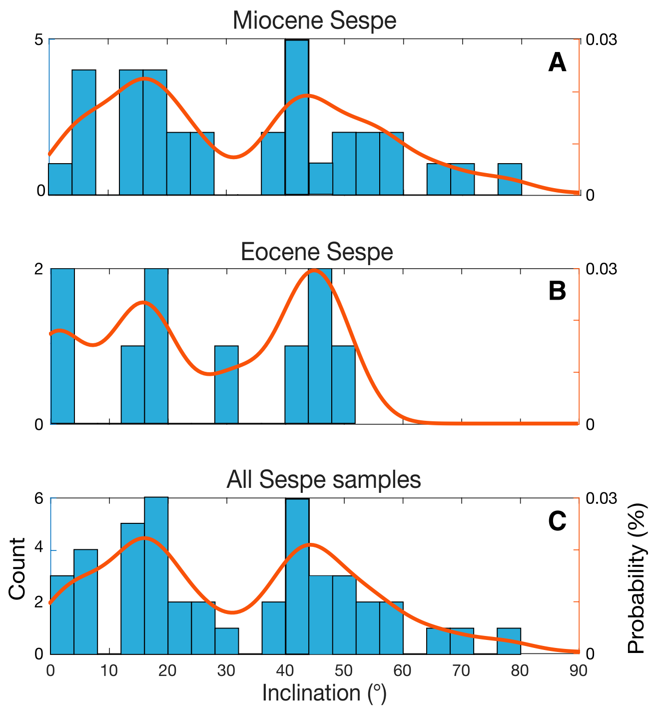

Figure 6. Histograms and population density functions (PDFs) of paleomagnetic inclinations measured in clasts of the Miocene (A) and Eocene (B) Sespe Formation, shown as a sum in (C).

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TABLE 5. REFERENCES FOR PREVIOUS DETRITAL ZIRCON AND PALEOMAGNETIC DATA

Figure	Sample or formation	Reference
Detrital zircon data		
8A	Tapeats 2	Gehrels et al., 2011
8B	Shinumo TO1-75-5	Bloch et al., 2006
8C	Shinumo TO1-75-2z	Bloch et al., 2006
8B	Shinumo TO1-75-4	Bloch et al., 2006
8F	Shinumo TO1-76-2	Bloch et al., 2006
8G	Shinumo TO1-76-3	Bloch et al., 2006
8H	Shinumo Basal Gravel LC-16-76-5	Mulder et al., 2017
8U	Zabriskie Quartzite	Stewart et al., 2001
8V	Upper Stirling NR9S	Schoenborn et al., 2012
8W	Troy Formation	Stewart et al., 2001; Mulder et al., 2017
8X	Dripping Springs Formation	Stewart et al., 2001; Mulder et al., 2017
8Y	Del Rio Quartzite	Spencer et al., 2016
8Z	Blackjack	Doe et al., 2012
8AA	Yankee Joe	Doe et al., 2012
8AB	White Ledges	Doe et al., 2012
8AC	Morrison Formation	Dickinson and Gehrels, 2008
8AD	Morrison Formation	Dickinson and Gehrels, 2008

Figure	Sample or formation	Reference	Maximum demagnetization temperature (°C)
Paleomagnetic data			
9A	Tapeats, Grand Canyon	Elston and Bressler, 1977	500–590
9B	Tapeats, central Arizona	Elston and Bressler, 1977	Undetermined
9C	Zabriskie Formation	Gillett and Van Alstine, 1979	640
9D	Wood Canyon Formation (red-purple mudstones only)	Gillett and Van Alstine, 1979 (Figs. 3F and 4)	640
9E	Rainstorm, all locations	Minguez et al., 2015; Van Alstine and Gillett, 1979	500–610
9F	Rainstorm, Nopah Range	Minguez et al., 2015	500–610
9G	Rainstorm, Winters Pass Hills	Minguez et al., 2015	500–610
9H	Rainstorm, Desert Range	Van Alstine and Gillett, 1979	650
9I	Neoproterozoic–Cambrian, Caborca Region	Molina-Garza and Geissman, 1999	355–660 (average 530)
9J	Lower Shinumo	Elston and Grommé, 1994	550
9K	Middle Shinumo (Pole 4)	Elston and Grommé, 1994	500–620
9L	Upper Shinumo	Elston and Grommé, 1994	650
9M	All above Shinumo	Elston and Grommé, 1994	See above

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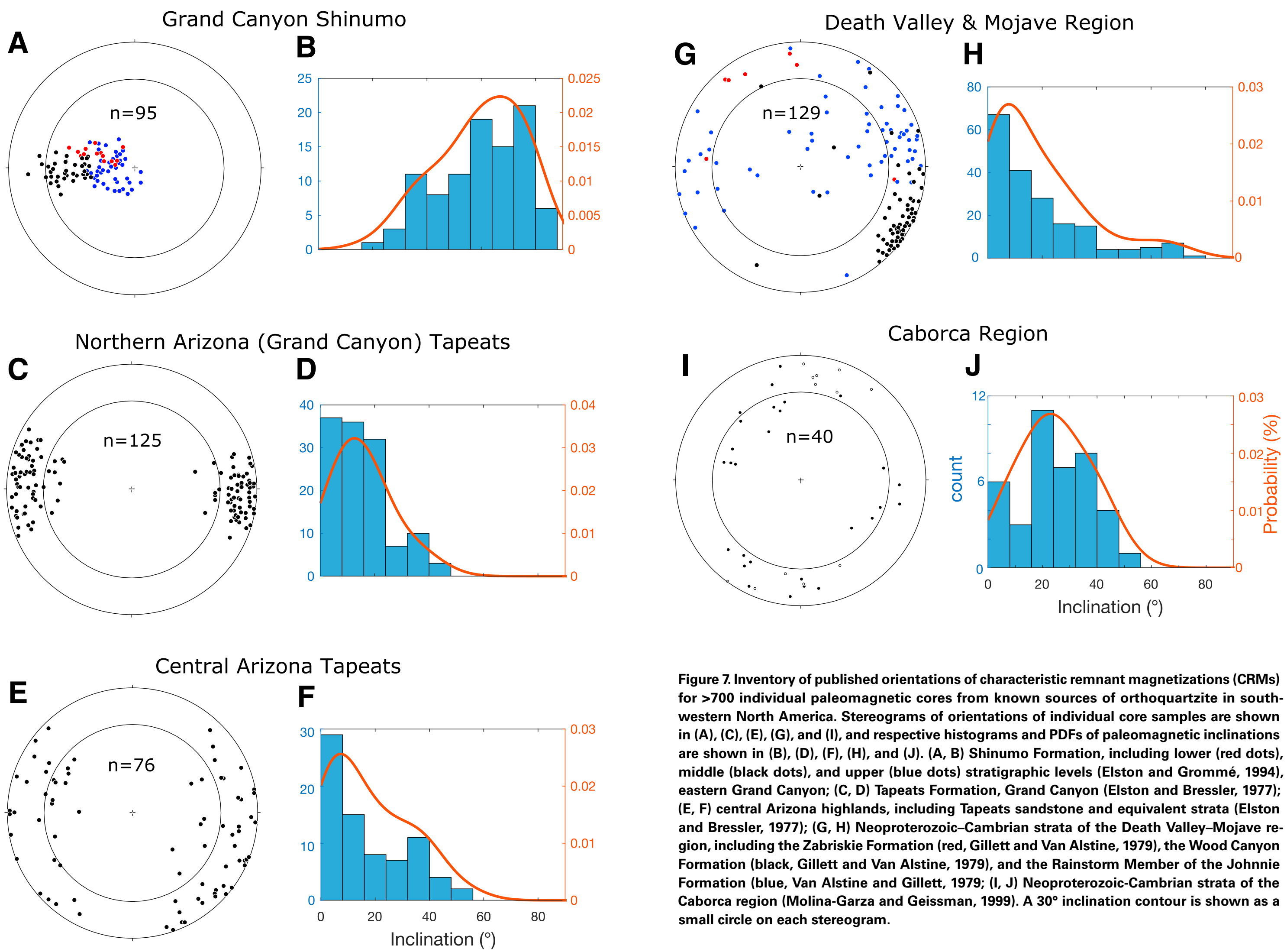

Figure 7. Inventory of published orientations of characteristic remnant magnetizations (CRMs) for >700 individual paleomagnetic cores from known sources of orthoquartzite in southwestern North America. Stereograms of orientations of individual core samples are shown in (A), (C), (E), (G), and (I), and respective histograms and PDFs of paleomagnetic inclinations are shown in (B), (D), (F), (H), and (J). (A, B) Shinumo Formation, including lower (red dots), middle (black dots), and upper (blue dots) stratigraphic levels (Elston and Grommé, 1994), eastern Grand Canyon; (C, D) Tapeats Formation, Grand Canyon (Elston and Bressler, 1977); (E, F) central Arizona highlands, including Tapeats sandstone and equivalent strata (Elston and Bressler, 1977); (G, H) Neoproterozoic–Cambrian strata of the Death Valley–Mojave region, including the Zabriskie Formation (red, Gillett and Van Alstine, 1979), the Wood Canyon Formation (black, Gillett and Van Alstine, 1979), and the Rainstorm Member of the Johnnie Formation (blue, Van Alstine and Gillett, 1979; (I, J) Neoproterozoic-Cambrian strata of the Caborca region (Molina-Garza and Geissman, 1999). A 30° inclination contour is shown as a small circle on each stereogram.

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central Arizona highlands. The only published paleomagnetic study on Proterozoic strata in the central Arizona highlands was the measurement of the NRM of Mesoproterozoic strata of the Apache Group (Pioneer Shale), which did not differ significantly from the modern field (Runcorn, 1964). Diabase sills that intruded the Apache Group at 1.1 Ga yield moderate inclinations (Harlan, 1993), as expected for late Mesoproterozoic time (e.g., Meert and Stuckey, 2002; Evans et al., 2016). Although we might expect moderate inclinations for central Arizona orthoquartzites, at present there is no basis to assume any particular distribution of inclinations from a population of Proterozoic clasts derived from the central Arizona highlands.

Because any given clast population represents a regional mixture of individual pebbles and cobbles from disparate sources, clast magnetizations are best compared with regional populations of magnetizations from individual paleomagnetic cores, as opposed to, for example, any particular site mean. In this form, a ready comparison can be made between a clast population and source populations according to some defined area. The Shinumo data (Figs. 7A and 7B) show well-grouped, moderate to high inclination, with only a few measurements (three of 95) below 30°. The Tapeats Formation cores in Grand Canyon (Figs. 7C and 7D) are shallowly inclined and well grouped into an east-west orientation. The Tapeats and related strata in the central Arizona highlands (Figs. 7E and 7F) are also mostly of low inclination but far more scattered in declination, likely due in part to their more complex thermal and tectonic history. The Death Valley–Mojave region data (Figs. 7G and 7H) are also generally of low inclination and fairly diverse in declination. These data generally reflect a period of long residence of SW Laurentia at low paleolatitude in Neoproterozoic–Paleozoic time, not returning to higher paleolatitudes until the Jurassic. In sum, the extant data from potential source populations show broadly unimodal, shallow inclination spectra, except for the Shinumo Formation, which shows a moderate- to high-inclination spectrum.

Detrital Zircon Data

Detrital zircon age spectra of orthoquartzites from both potential sources and the Sespe Formation, including new data presented here and a compilation of published data (Table 5), are presented in Figure 8. Representative spectra from sources in Grand Canyon, including the Shinumo Formation and Tapeats Formation, are shown in the left-hand column (Figs. 8A–8H), which includes sample IC-1-35 obtained for this study (Fig. 8E). Representative spectra from potential sources in the Death Valley–Mojave region (Figs. 8U and 8V) and central Arizona highlands (Figs. 8W–8AA) are shown in the right-hand column. Also shown in the right-hand column, for reasons discussed in detail below, are representative spectra from the Westwater Canyon Member of the Upper Jurassic Morrison Formation, which appears to be a source for one of the Sespe clasts. Representative spectra from the Death Valley region include the Zabriskie and upper Stirling Formations, and from the Arizona region include the Troy, Dripping Springs, Del Rio, Blackjack, Yankee Joe, and White Ledges Formations. Samples in the center column include ten clasts from the Miocene Sespe Formation and two clasts from the Eocene Sespe. As noted above, of the ten Miocene Sespe samples, eight have moderate to high paleomagnetic inclinations, and two have low paleomagnetic inclinations. As noted above, the low-inclination samples (Figs. 8Q and 8R) were both collected from the same outcrop of “marker conglomerate” at the base of the Sespe along Santiago Canyon Road in the Santa Ana Mountains. The two clasts from Eocene Sespe both have moderate inclination. Analytical data for the 13 samples analyzed for this study are presented in Table S2.

The most prominent observation regarding the source spectra is that Grand Canyon and Death Valley sources both have multimodal (“cosmopolitan”) spectra, with discernable peaks near 1.2, 1.4, and 1.7 Ga. In contrast, the central Arizona highlands sources tend to have unimodal or bimodal spectra and include small numbers of pre–2.0 Ga grains. The only central Arizona highlands source with a Grenville-age peak is the Troy Quartzite, which features a strong peak at 1.26 Ga and a broad distribution of older ages, with a much weaker peak at 1.48 Ga (Fig. 8W). The only other source with any Grenville component is the Dripping Springs Formation, which contains a few ages (<5%) younger than 1.3 Ga, associated with a broad peak at 1.4 Ga. The youngest zircons in the Dripping Springs and Troy formations are 1.23 and 1.20 Ga, respectively. Depositional ages of the other central Arizona orthoquartzite bodies are too old to contain Grenville-aged zircons and tend to be strongly unimodal at 1.7 Ga. Therefore, either (1) strong unimodality or (2) absence of pre–1.20 Ga Grenville-aged zircons, discriminate central Arizona sources from both Death Valley–Mojave region sources and Grand Canyon sources.

The data from the 12 Sespe clasts fall into two basic groups, which include nine samples with cosmopolitan spectra (Figs. 8I–8Q) and three with strongly unimodal spectra (Figs. 8R–8T). The cosmopolitan spectra tend to have three modes near 1.2 Ga, 1.4 Ga, and 1.7 Ga and minor amounts of pre–2.0 Ga grains. Although the modes are variable in detail, they are mostly subequal, with the exception of sample BW4809, in which Grenville-age grains are much less abundant than in other cosmopolitan samples. The three samples with unimodal spectra all have peaks near 1.7 Ga and a few pre–2.0 Ga grains.

Three of the nine cosmopolitan spectra also contain a small but significant fraction (~5%) of Paleozoic and Mesozoic grains. The Paleozoic grains in sample LS1114 average 331 Ma, and a single Mesozoic age is 153.0 ± 2.8 (1 sigma) Ma (Fig. 8L). In sample BW4809, six Paleozoic grains define a tightly clustered unimodal peak at 485 Ma, and there are no Mesozoic grains (Fig. 8P). In sample BW1609, five Mesozoic grains cluster tightly near 168 Ma, and a single Paleozoic grain is 510 ± 10 Ma (Fig. 8Q).

We observe a general distinction in detrital zircon spectra between the Miocene and Eocene Sespe clast populations. In the Miocene population, nine of ten spectra contain abundant Grenville-aged zircons, with eight of these nine having a well-defined peak. All nine samples contain grains younger than 1.20 Ga in their populations. The one remaining sample is unimodal with a 1.7 Ga peak. In contrast to the cosmopolitan spectra, the Eocene Sespe clasts are both unimodal with 1.7 Ga peaks.

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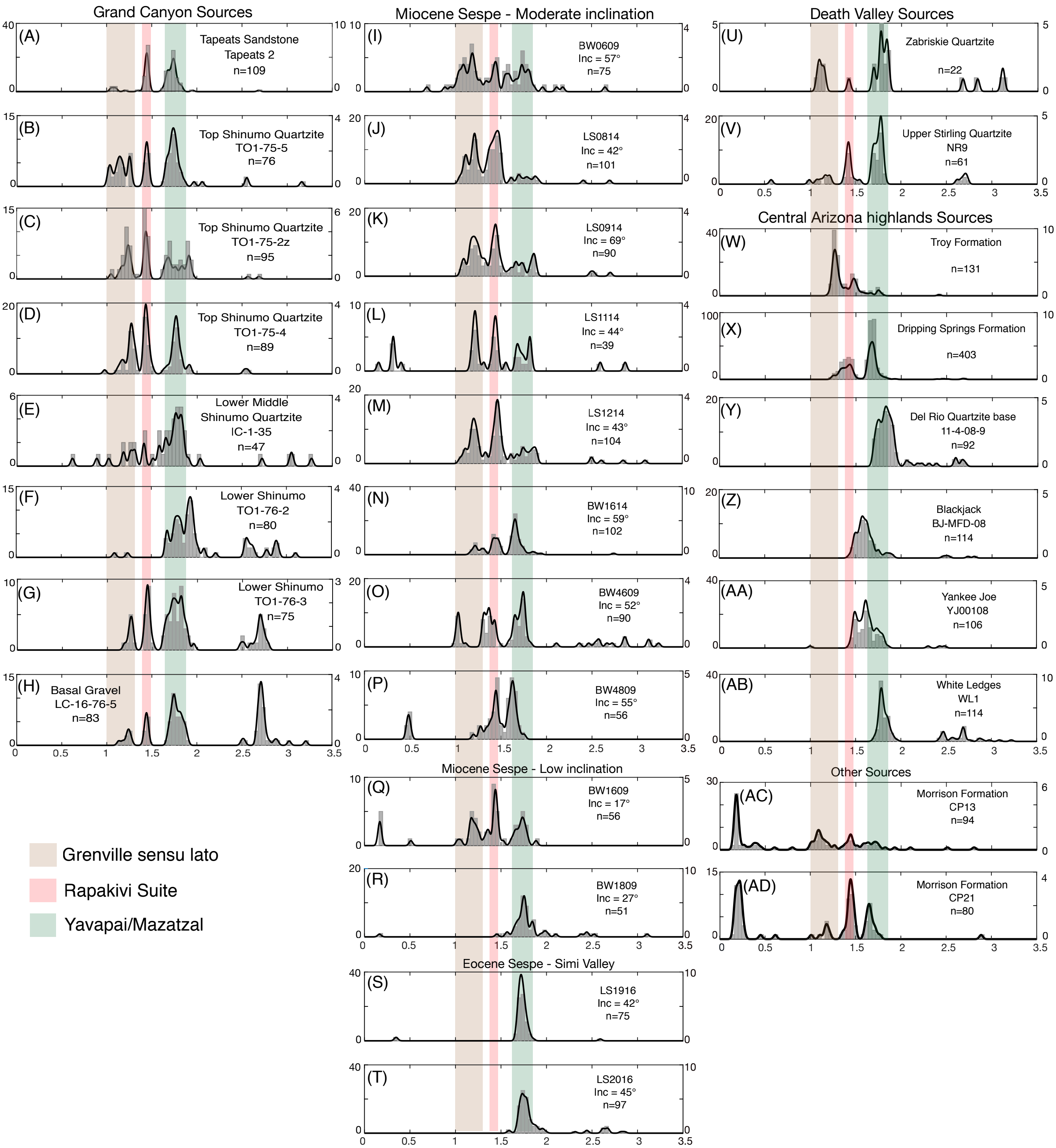

Figure 8. Detrital zircon spectra of potential sources and Sespe clasts. Potential Grand Canyon sources in the left column include the Tapeats Formation (A) and the Shinumo Formation (B–H). The center column includes Miocene Sespe clasts with moderate inclinations (I–P), Miocene Sespe clasts with low inclination (Q, R), and Eocene Sespe clasts with moderate inclination (S, T). The right column shows Death Valley sources including the Zabriskie Quartzite (U) and Upper Stirling Quartzite (V), and central Arizona highland sources including the Troy Quartzite (three samples) (W), the Dripping Springs Formation (three samples) (X), the Del Rio Quartzite base (Y), the Blackjack (Z), Yankee Joe (AA), and White Ledges (AB). We also include two samples of the Morrison Formation (AC) and (AD). Data sources are listed in Table 4.

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DISCUSSION

Paleomagnetic Inclination Analysis

Comparison of Sespe Formation Clasts and Sandstone Matrix

Paleomagnetic data from Piuma Member sandstones, collected in the same area that we collected orthoquartzite clasts along Piuma Road, have a tilt-corrected mean inclination of 39°± 6° (α_{95}) (Hillhouse, 2010). The CRM is carried by elongate, authigenic hematite that grew along cleavage planes within detrital biotite (Hillhouse, 2010). Because orthoquartzite clasts are generally devoid of detrital micas (Figs. 4C and 4D) and other soluble phases, it is highly unlikely that the clasts carry this magnetization.

Further, in unmetamorphosed red beds in general, the permeabilities of ultradurable clasts, such as orthoquartzite and metarhyolite (<10^{-4} darcy), are at least three orders of magnitude lower than those of their porous sandstone matrix (0.1–1 darcy: e.g., table 2.2 in Freeze and Cherry, 1979). This, in turn, suggests a strong contrast between clasts and matrix in exposure to diagenetic pore fluids. Thus, the elimination and replacement of the predepositional CRM in orthoquartzite clasts with an early Miocene magnetization, similar to that of the Sespe sandstone matrix, are unlikely. We also note that, whereas the clast CRMs are of high coercivity and unblocking temperature, peak temperatures of the Sespe Formation are generally well below 150 °C, based on maximum burial depths of 5000 m in the Saddle Peak area (e.g., section D–D′ of Dibblee, 1993) and 3000 m in the northern Santa Ana Mountains (e.g., section F–F′ of Schoellhamer et al., 1981). These clasts, therefore, tend to retain their original CRMs during transport, deposition, and diagenesis in the shallow crust, especially if those magnetizations are of high coercivity and unblocking temperature (e.g., Pan and Symons, 1993; Hodych and Buchan, 1994).

Comparison of Sespe Clasts to Possible Sources

Histograms of inclination data from each potential source formation are plotted at a uniform scale for comparison with histograms from clasts in the Sespe Formation at a suitably expanded vertical scale (Fig. 9). An important assumption in any comparison of Sespe clasts to source data is that the latter are representative of the source region as a whole. In other words, we assume it is unlikely that the inclination distribution of 188 randomly sampled orthoquartzites in the Death Valley–Mojave region would differ significantly from the 188 samples shown in the left-hand column of Figure 9. The fact that distributions from individual samples and formations are, without exception, similar to the overall distribution, suggests that the extant data set is representative of the region. There are probably sources where moderate to high inclinations are recorded by Death Valley–Mojave orthoquartzites, for example, by remagnetization in the contact aureoles of Mesozoic or Tertiary intrusions. But, such sources, if present, would occupy only a small fraction of the very extensive drainage area of Sespe gravels, and so they would be unlikely to influence the inclination distribution of the clast population as a whole.

With respect to sources in Figure 9, the low-inclination population of clasts from the Miocene and Eocene Sespe Formation could only have been derived from sources in the left-hand column, which includes Neoproterozoic–Cambrian formations in the Death Valley–Mojave region, the Tapeats Formation (both in the central Arizona highlands and in Grand Canyon), or (improbably) from Neoproterozoic–Cambrian strata of the Caborca region. The moderate- to high-inclination population of clasts, however, could not have been derived from the Neoproterozoic–Cambrian source populations, and requires either a Shinumo Formation source, shown in the upper right-hand portion of Figure 9, or some other unidentified source with similar paleomagnetic characteristics. Such a source could plausibly be Mesoproterozoic or Paleoproterozoic orthoquartzites in the central Arizona highlands, where as noted above, paleomagnetic data are lacking, or less plausibly from NW Sonora. Summations of the low-inclination distributions (from the Tapeats Formation and the Death Valley–Mojave region) and the moderate- to high-inclination distributions (Shinumo Formation) each define two unimodal distributions (Fig. 10). A comparison of these distributions with the distribution of the Miocene Sespe clast population suggests that neither source alone could produce the bimodal clast distribution, but a combination of the two sources could.

Cumulative distribution functions (CDFs) from the Miocene and Eocene Sespe clast populations are compared to those from each of the three source regions in Figure 11. Distributions from the Death Valley–Mojave region, both as individual formations (including the Rainstorm Member of the Johnnie Formation, the Wood Canyon Formation, and the Zabriskie Formation), and as a whole, lie well to the left (low-inclination side) of the Miocene Sespe distribution, and somewhat to the left of the Eocene Sespe distribution (Fig. 11A). Distributions from the Grand Canyon region lie either well to the left (Tapeats Formation) or well to the right (Shinumo Formation) of both Miocene and Eocene Sespe distributions (Fig. 11B). A distribution from the central Arizona highlands region (Tapeats Formation) lies to the left of the Sespe distributions (Fig. 11C).

The comparisons in Figures 11A–11C appear to exclude the Death Valley–Mojave region as a sole source for the Miocene and Eocene Sespe distributions. However, because the central Arizona highlands region may contain sources with moderate to high inclinations, it cannot be ruled out as a source for either the Miocene or Eocene Sespe clast distributions. Linear combinations of the two Grand Canyon sources (Tapeats and Shinumo Formations as end members) compare well with the Miocene Sespe clast distribution for a broad range of mixtures (Fig. 11D). For Shinumo fractions ranging from ~30%–60%, Kolmogorov-Smirnov tests yield p-values of 0.05 or greater (Fig. 12), indicating that the derivation of Sespe clasts from this range of mixtures cannot be ruled out at 95% confidence. There is a strong maximum value of p for these mixtures of $p = 0.34$ for a Shinumo fraction of 35%–40%. The same comparison of Grand Canyon sources and Eocene Sespe clasts is not as strong. For these mixtures, p-values of 0.05 or greater are restricted to Shinumo fractions of ~10%–15%, with a maximum of only $p = 0.07$. These comparisons suggest that

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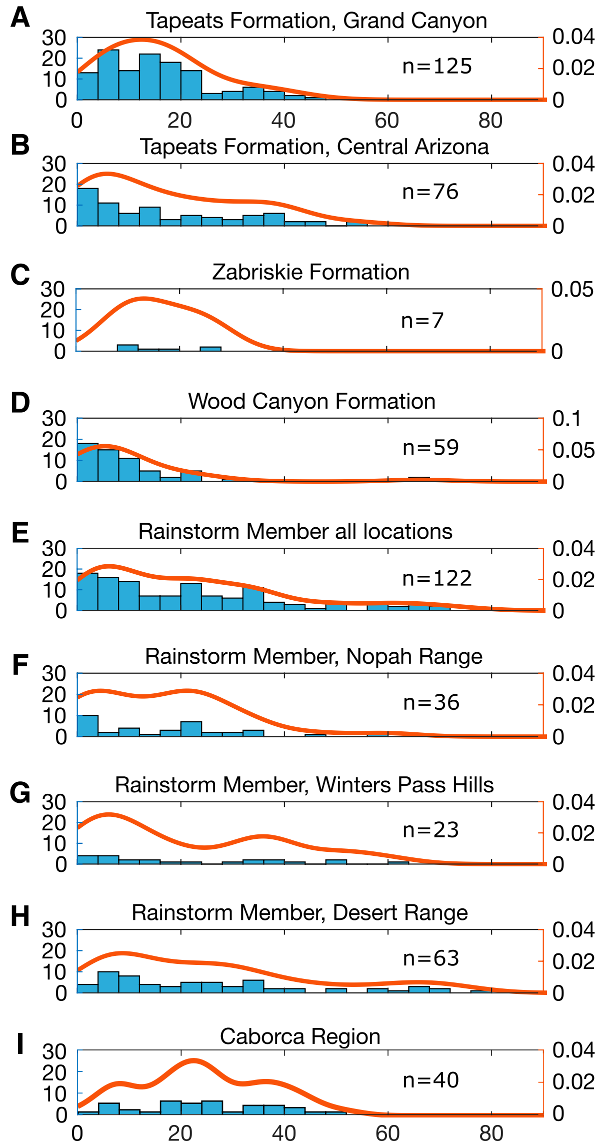

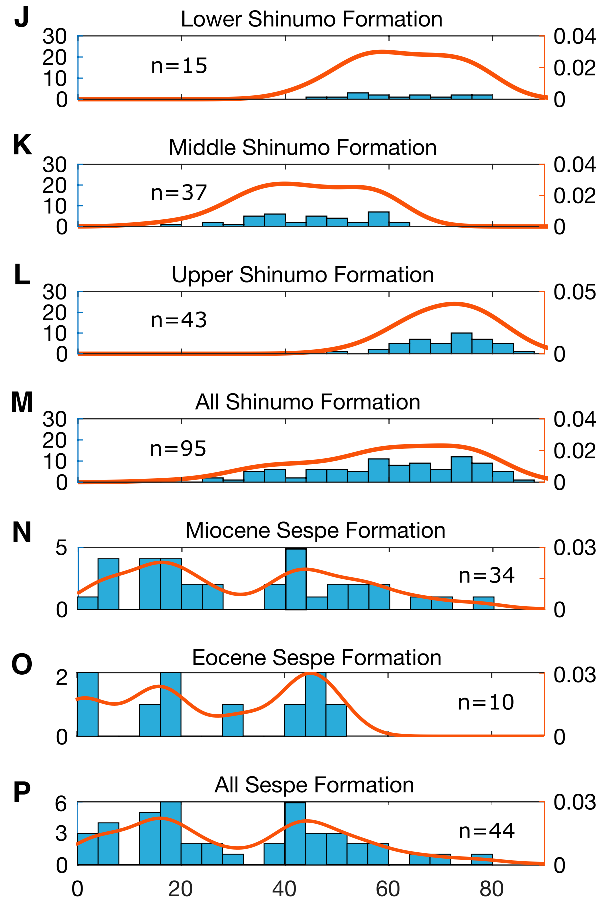

Figure 9. Histograms of paleomagnetic inclinations from potential sources, plotted at a uniform scale, and from clasts in the Sespe Formation, plotted at a suitably expanded scale. Potential sources from Grand Canyon include the Tapeats Formation (A) and the Shinumo Formation (J–M). Potential sources from the central Arizona highlands include the Tapeats Formation (B). Potential Death Valley sources include the Zabriskie Formation (C), Wood Canyon Formation (D), Rainstorm Member of the Johnnie Formation (E–H). Potential sources from Caborca include Ediacaran–Cambrian strata, mainly the El Arpa, Caborca, Clemente, Papalote, and Cerro Prieto formations (I). Paleomagnetic inclinations were measured in this study from the Miocene Sespe Formation (N) and the Eocene Sespe Formation (O), shown also as a sum (P). Paleomagnetic inclinations of the Rainstorm Member from the Nopah Range and Winters Pass Hills were measured after thermal demagnetization of 500–610 °C (Minguez et al., 2015), and inclinations of the Rainstorm Member from the Desert Range were demagnetized to 650 °C (Van Alstine and Gillett, 1979). Directions from the Wood Canyon (red-purple mudstones only) and Zabriskie Formations, both in the Desert Range, were measured after thermal demagnetization to 640 °C (figures 3F and 4 in Gillett and Van Alstine, 1979). Paleomagnetic inclinations from the Tapeats Formation in the central Arizona highlands and in Grand Canyon were measured after thermal demagnetization at temperatures of 500–590 °C (Elston and Bressler, 1977). Inclinations from the lower Shinumo Formation were measured after demagnetization at 550 °C, from the middle Shinumo at 500–620 °C (data referred to as "Pole 4"), and from the upper Shinumo at 500–620 °C (Elston and Grommé, 1994). Inclinations from clasts in the Miocene Sespe Formation (M) and clasts in the Eocene (N) are from this study, plotted also as a sum (O). Data sources are listed in Table 4.

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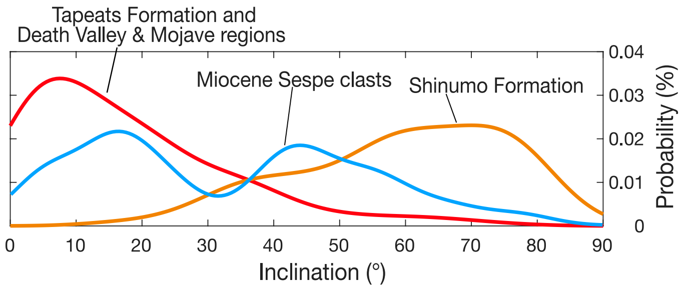

Figure 10. Comparisons of probability density functions (PDFs) of paleomagnetic inclination data from Miocene Sespe orthoquartzite clasts (blue curve), a summed population of Tapeats Formation, from both Grand Canyon and central Arizona highlands, and formations from the Death Valley–Mojave regions (red curve) and the Shinumo Formation (gold curve).

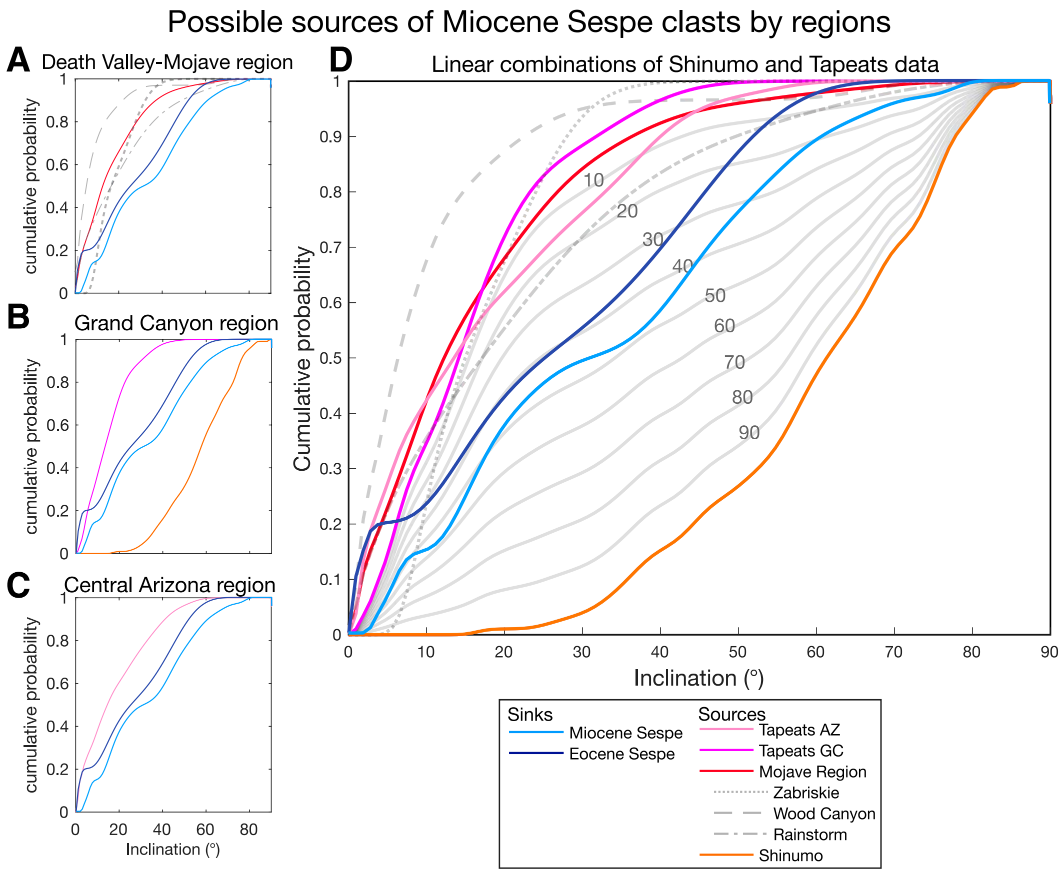

Figure 11. Comparisons of cumulative distribution functions (CDFs) of inclination data from Sespe orthoquartzite clasts (blue-hued curves) and possible sources (red-hued curves), including (A) Death Valley–Mojave region sources, (B) Grand Canyon region sources, and (C) central Arizona highlands sources. The three gray curves in (A) are summed to yield an average for the Death Valley–Mojave region (red). (D) Summary plot showing linear mixtures of Tapeats Formation from Grand Canyon and Shinumo Formations as end members, contoured in 10% increments (solid gray curves). AZ—Arizona; GC—Grand Canyon.

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a sole Grand Canyon source comprising a mixture of Tapeats and Shinumo Formation clasts is a viable explanation of the inclination distributions for the Sespe clast populations and is particularly strong for the Miocene population. As noted earlier, ultradurable orthoquartzites in the Tapeats Formation are not exposed in eastern Grand Canyon but are characteristic of western Grand Canyon exposures. Therefore, a roughly equal mixture of Tapeats and Shinumo clasts implies that the source areas included both the Upper Granite Gorge of eastern Grand Canyon and the Lower Granite Gorge of western Grand Canyon.

These comparisons, of course, may be equally well explained with mixtures that include components from both Death Valley–Mojave and central Arizona highlands sources, either with or without a very small contribution from Sonoran sources. Death Valley–Mojave sources cannot be distinguished from the Tapeats Formation in Grand Canyon, and Proterozoic sources from the central Arizona highlands may have moderate to high inclinations, and thus be indistinguishable from the Shinumo Formation. The key to distinguishing a Shinumo contribution to the Sespe clast population thus lies in a simple test that distinguishes the Shinumo Formation from orthoquartzites in the central Arizona highlands, using detrital zircon age spectra.

Detrital Zircon Analysis

Here, we apply the detrital zircon test to our analysis of populations of paleomagnetic inclinations, in order to discriminate source regions, both for individual clasts and for the population of clasts as a whole in the Piuma Member and Eocene Sespe populations (Table 6).

In this analysis, it is important to first consider the three orthoquartzite clast samples containing small but significant populations of Paleozoic and Mesozoic grains (samples LS1114, BW4809, and BW1609; Figs. 8L, 8P, and 8Q). These data raise the question of whether those grains are detrital components of the orthoquartzite, or whether they are "allochthonous" and incorporated upon or into the clast during weathering and transport.

Sample LS1114 (Fig. 8L), from the Piuma Member, has a unique detrital zircon spectrum relative to all other samples, and its source is therefore quite uncertain. Based on comparison with the extensive detrital zircon data set from Mesozoic sandstones on the Colorado Plateau (Dickinson and Gehrels, 2008), its most likely source is the Upper Jurassic Morrison Formation (Table 6). Similar to the Morrison, LS1114 has a moderate paleomagnetic inclination, scarcity of grains between 0.5 and 1 Ga in its detrital zircon spectrum, and is a well-indurated, light pinkish-gray, medium- to coarse-grained orthoquartzite. Although the Mesozoic peak in the Sespe spectrum is not as prominent as in the two Morrison spectra, the ratio of Mesozoic to Proterozoic grains is more similar between LS1114 and CP21, from the Morrison, than it is between the two Morrison samples.

In contrast to LS1114, we interpret the Paleozoic grains in samples BW4809 and BW1609 (Figs. 8P and 8Q) to be allochthonous. Both samples were collected from the Miocene Sespe in the Santiago Canyon road cut. Their detrital zircon spectra are a poor match for any known Paleozoic or Mesozoic sandstone in having a small, single Paleozoic mode. Further, clasts from this outcrop exhibit petrographic evidence for the extensive development of silica glaze on the clast surface, beneath which thin films of allochthonous grains are adhered to the clast exterior and narrow fractures in the clast interior that also locally contain allochthonous grains (Fig. 13). Both of these clasts are densely cemented, purple-hued orthoquartzites that are a poor lithologic match for even the most densely cemented late Paleozoic or Mesozoic sandstones

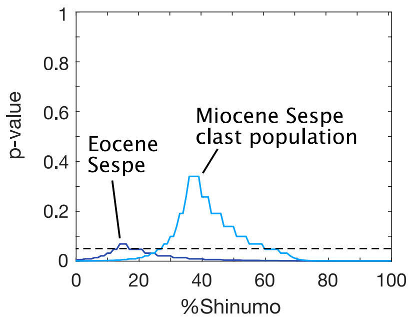

Figure 12. P-values for comparisons of cumulative distribution functions (CDFs) from Figure 11, including (A) Eocene Sespe and (B) Miocene Sespe inclination populations and mixtures of Tapeats Formation from eastern Grand Canyon (right end members) and Shinumo Formation (left end members) inclination populations.

TABLE 6. SUMMARY OF RESULTS FOR SAMPLES WITH PALEOMAGNETIC AND DETRITAL ZIRCON DATA

Sample and location	Paleomagnetic inclination (°)	Grenville detrital zircon peak?	Interpreted source region
Miocene Sespe (moderate and high inclination)			
Piuma Road			
LS1114	44	Yes	Morrison Formation
BW0609	57	Yes	Grand Canyon (Shinumo)
LS0814	42	Yes	Grand Canyon (Shinumo)
LS0914	69	Yes	Grand Canyon (Shinumo)
LS1214	43	Yes	Grand Canyon (Shinumo)
BW1614	59	Yes	Grand Canyon (Shinumo)
Red Rock Trail, Limestone Canyon Park			
BW4609*	52	Yes	Grand Canyon (Shinumo)
BW4809	55	Yes	Grand Canyon (Shinumo)
Miocene Sespe (low inclination)			
Santiago Canyon Road			
BW1609	17	Yes	Death Valley
BW1809	27	No	Central Arizona highlands
Eocene Sespe (moderate inclination)			
Simi Valley			
LS1916	42	No	Central Arizona highlands
LS2016	45	No	Central Arizona highlands

*Characteristic magnetization is carried by magnetite, which has not been observed in the extant database for Shinumo.

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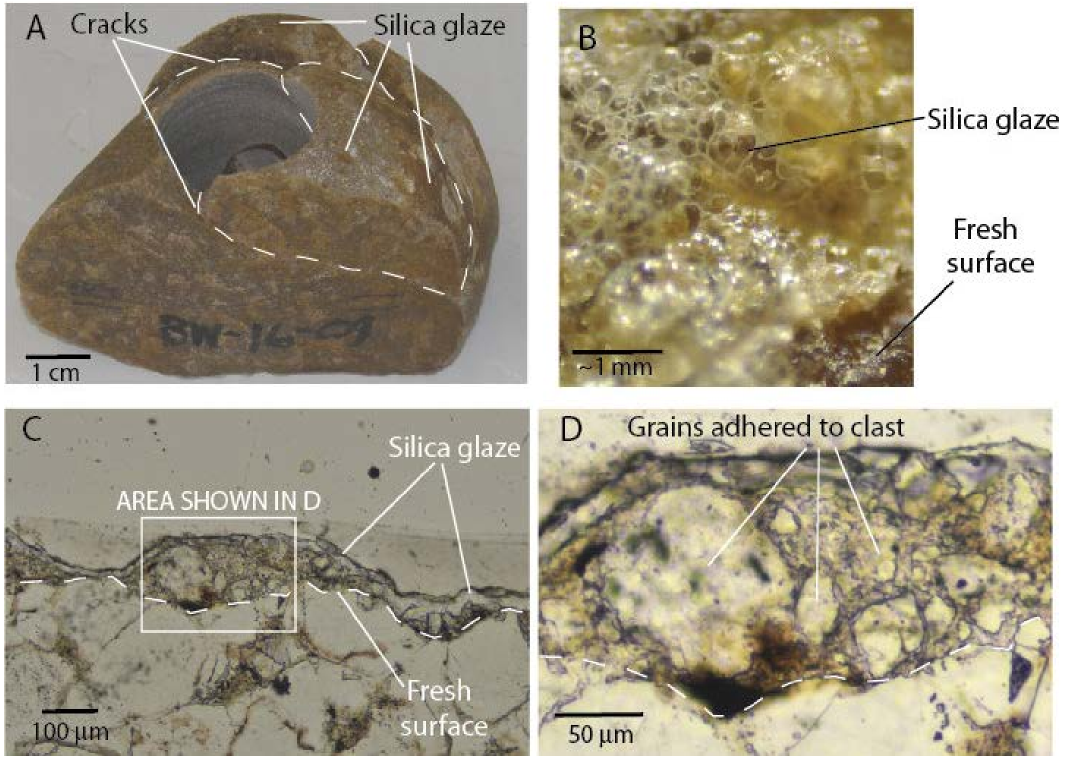

Figure 13. Images of silica glaze on a Sespe orthoquartzite clast from road cut on Santiago Canyon Road. (A) Photograph showing light-brown weathering patches of silica glaze on clast exterior and the location of cracks within the sample that locally contain detrital material external to the clast; (B) photograph showing small-scale mammillary texture of silica glaze in reflected light; (C, D) photomicrographs of thin sections cut normal to clast exterior showing silica glaze in cross section, which includes external grains adhered to the clast, in transmitted light.

in the potential source regions. These samples both have low inclination but contrasting detrital zircon spectra (Figs. 8Q and 8R). The unimodal spectrum of BW1809 (Fig. 8R) indicates that it was derived from the central Arizona highlands, suggesting that the inclination distribution of central Arizona orthoquartzites may include shallowly inclined samples. Sample BW1609 (Fig. 8Q), which has a strong Grenville-age peak, is probably derived from the Death Valley–Mojave region, based on its inclination, densely cemented grains, and purple hue (Table 6). This, of course, assumes that its small population of Mesozoic grains is allochthonous.

The two Eocene Sespe clasts with moderate inclinations both have unimodal peaks at 1.7 Ga and a smattering of Archean grains, indicating derivation from the central Arizona highlands (Figs. 8S and 8T; Table 6).

The remaining seven samples were all collected from the Piuma Member (five from the Piuma Road section and two from the Red Rock Trail section) and have both moderate to high inclination and relatively broad Grenville-age zircon peaks. Among known potential sources, these characteristics restrict this population to a Shinumo Formation source, among known sources. As noted above, the Troy Quartzite at the top of the Apache Group is the only Proterozoic orthoquartzite in the central Arizona highlands to contain appreciable Grenville-age zircons (Fig. 8W versus Figs. 8X–8AD) and therefore could be a potential source. However, the Troy data are dominated by an early Grenville peak near 1.26 Ga, with no grains younger than 1.20 Ga, and very weak peaks near 1.4 and 1.7 Ga. In contrast, Miocene Sespe clasts and the Shinumo Formation are both characterized by broader Grenville peaks (including many grains between 1.0 and 1.20 Ga) and much stronger peaks at 1.4 and 1.7 Ga. A K-S test comparing the Troy data (Fig. 8W) with Miocene Sespe clasts LS0814 and LS1214 (Figs. 8J and 8M) yields p-values of 2.1×10^{-5} and 3.5×10^{-4}, respectively, ruling out derivation of sands in the Troy Formation and sands in the Miocene Sespe clasts from the same source. Therefore, extant data from the Apache Group do not provide a compelling match for orthoquartzite clasts in the Miocene Sespe Formation.

Interpretive Complications

We consider here three important issues in interpreting the Shinumo Formation as the bedrock source for the moderately inclined mode of orthoquartzite

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clasts in the Miocene Sespe Formation. These include (1) primary structures within source formations, such as cross-stratification, and their influence on the inclination spectra of clast populations; (2) recycling of clasts from gravel sources that are intermediate in age between the Shinumo and Sespe Formations, which may compromise the interpretation of a Shinumo source for Miocene Sespe clasts; and (3) buried or now-eroded sources for the clasts outside of the eastern Grand Canyon region.

Primary Structure

Orthoquartzites in the southwestern United States are substantially compacted after deposition, commonly cross-stratified, and locally contain paleoliquefaction structures. An analysis of the potential effects of primary structures on paleomagnetic inclination spectra is provided in Supplemental Text S1 and Figure S4 (footnote 1). Our analysis suggests that primary structures, especially cross-stratification, may have a measurable effect on the distribution of paleomagnetic inclinations in any given sample population. Relationships between the measured orientations of foresets and of paleomagnetic inclinations in potential source regions indicate that the difference between low-inclination and moderate- to high-inclination populations would be augmented to some degree by this effect. Depending on the volume fraction of foreset laminations sampled by the clast population, such augmentation would be in the range of 0° to 15°, which serves to slightly enhance the distinction between the two populations, rather than obscure it.

Recycling of Clasts

An additional complication in any provenance study is the possibility of recycling of clasts from secondary sources. It is possible that a significant fraction of Sespe gravel clasts is derived from conglomeratic strata that are intermediate in age between the time of exposure of their bedrock source and the time of Sespe deposition (e.g., Dickinson, 2008). As noted above, in the case of the Shinumo bedrock source region, extensive thermochronometric data demonstrate that unroofing of the Upper Granite Gorge in the eastern Grand Canyon region, which includes all known exposures of the Shinumo Formation, did not occur before ca. 28–18 Ma (Flowers et al., 2008; Flowers and Farley, 2012; Lee et al., 2013; Karlstrom et al., 2014; Winn et al., 2017). Therefore, assuming lower Miocene Sespe orthoquartzites are indeed derived, in part, from the Shinumo Formation, the possibility of clast recycling does not alter the conclusion that sedimentary transport from Upper Granite Gorge bedrock sources to coastal California occurred between ca. 28 and 20 Ma.

There is also the possibility that the clasts are recycled from conglomeratic strata that contain orthoquartzite detritus, either derived from the Shinumo Formation or from an unknown source with similar paleomagnetic and detrital zircon characteristics. Because the Shinumo Formation was buried in Cambrian time, and remained so until the Oligocene, any pre-Oligocene recycling path must have begun prior to Cambrian burial. For example, Shinumo clasts could have been eroded into Neoproterozoic rift basins in the Death Valley region and then supplied to the Sespe Formation via an Amargosa paleoriver. Other potential recycled sources include the Jurassic cobble and boulder conglomerates of the Coyotes Formation near Hermosillo, Mexico, and possible equivalents exposed as far north as the Caborca area, but these are unlikely as Sespe sources, as noted above. These and other recycling histories, although possible, thus require postulation of either distant or unknown reservoirs of orthoquartzite clasts that would somehow overwhelm extant, broadly exposed reservoirs in their contributions to the Miocene Sespe basin.

Buried or Now-Eroded Sources

As in any provenance study, it is possible that an unknown source, either eroded away since 20 Ma or buried beneath the extensive alluvial deposits in the Basin and Range region, could have provided a clast population with any combination of the paleomagnetic and detrital zircon characteristics needed to explain the Sespe clast data. Nearly all of the moderate- to high-inclination clast population in the Piuma Road section has Shinumo characteristics (seven out of the eight measured clasts, or 88%). Our results agree well with the observation (described above in Introduction and Geologic Setting) that the Shinumo Formation lies within the only known region in the Cordilleran interior that underwent kilometer-scale erosional denudation during Piuma time (ca. 28–18 Ma). In other words, the Shinumo Formation is apparently the dominant source for the moderate- to high-inclination clast population. In contrast, the hypothesis that Piuma orthoquartzite clasts are substantially derived from the central Arizona highlands can be rejected at a high level of confidence, because eight out of eight clasts (Fig. 14) failed the detrital zircon test for central Arizona highlands sources. Deriving the Piuma orthoquartzite clast population from now-eroded or -buried sources in the Mojave region is clearly possible. However, it is inconsistent with the Laramide unroofing history of the region (80–40 Ma, versus the ca. 20 Ma depositional age), which suggests a fairly stable landscape from 40 to ca. 20 Ma (e.g., Spotila et al., 1998). In sum, we interpret our results to support the hypothesis stated in the Introduction, that the mid-Tertiary, rapid unroofing event in the eastern Grand Canyon source region is reflected in an abundance of eastern Grand Canyon orthoquartzite clasts in coeval basins of coastal southern California.

Detrital Zircon Spectra in Sespe Sandstone

In modern Colorado River sands, 20% of the detrital zircon population ranges in age from 300 to 900 Ma, reflecting the dominant contribution of Permian through Jurassic aeolianites widely exposed throughout the Colorado River drainage basin (Kimbrough et al., 2015). The Arizona River drainage proposed

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here (Fig. 1) and in Wernicke (2011) includes part of the southwestern margin of the Colorado Plateau that, in turn, contains part of the region of 28–18 Ma erosion (stippled region in Fig. 1). The area of the plateau included within the Arizona River drainage is nominally 30,000 km^2 (Fig. 1), which is ~6% of the area of the modern Colorado River drainage basin that includes the Colorado Plateau and environs (~500,000 km^2, Table 1 in Kimbrough et al., 2015). Thus, if the modern Colorado River drainage was limited to headwaters in the eastern Grand Canyon region, the expected contribution of 300–900 Ma zircon grains would be (0.06) (0.20) = 0.012, or ~1% of the population. Detrital zircon age determinations from 22 samples of the Sespe Formation (including 1378 total grains) yielded a contribution of 0.7% of 300–900 Ma detrital zircons (Spafford, 2010; Table 1 in Ingersoll et al., 2013), in reasonable agreement with the expected ratio. This 300–900 Ma population could be derived entirely from the Mojave-Sonora region, entirely from the Grand Canyon region, or most likely from some combination of the two. In other words, the sandstone detrital zircon data are insufficient to discriminate between Mojave-Sonora and Grand Canyon sources for the 300–900 Ma detrital zircon component, contrary to the conclusion of Ingersoll et al. (2013) that the data indicate no drainage link between southern California river deltas and the Grand Canyon region during Sespe time.

CONCLUSION

As summarized in Table 6 and Figure 14, our results show that combined intraclast paleomagnetic inclination and detrital zircon data provide significant new insights into the provenance of Sespe clast populations that cannot be derived from either data set alone. The eight moderate- to high-inclination clasts from the Miocene Sespe for which we obtained detrital zircon spectra uniformly contain Grenville-age detrital zircon peaks (Fig. 14), ruling out both the Death Valley–Mojave and central Arizona highlands regions as source populations. With the exception of LS1114, which appears to be Jurassic, we interpret them all as being derived from the Shinumo Formation (Figs. 3 and 14). The two Miocene Sespe clasts that have low inclination were both collected from the Santiago Canyon Road locality, from the basal conglomerate of the lower Miocene Sespe Formation. Given that one yielded a unimodal detrital zircon peak at 1.7 Ga and the other a cosmopolitan spectrum, the central Arizona highlands and Death Valley–Mojave region both appear to be possible sources for the broader Miocene orthoquartzite population (Howard, 1996). The two Eocene Sespe clasts with moderate paleomagnetic inclinations yielded unimodal zircon age spectra with peaks at 1.7 Ga, indicating derivation from the central Arizona highlands. Clearly, more data will be required to further test the hypothesis that the Eocene Sespe is predominantly sourced from the central Arizona highlands (e.g., Howard, 2000, 2006). It is noteworthy, however, that the outcome of moderate inclination plus a unimodal 1.7 Ga peak observed in the Eocene Sespe was not observed in any of the ten Miocene Sespe samples. Therefore, regardless of how one interprets these data in terms of provenance, they have clear potential to identify and characterize contrasting clast populations (Fig. 14).

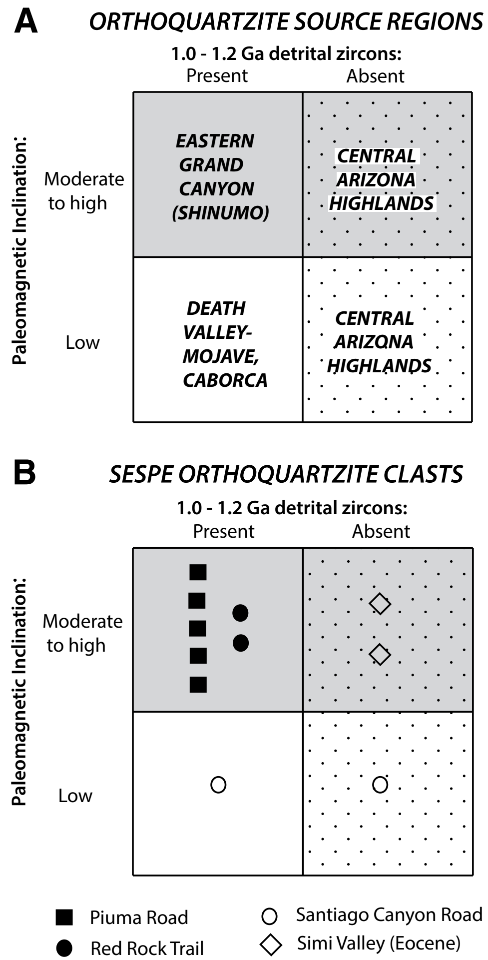

Figure 14. Matrices summarizing research outcomes of paleomagnetic and detrital zircon data for (A) orthoquartzite detrital source regions and (B) pre-Mesozoic orthoquartzite clasts in which both paleomagentic inclination and detrital zircon data were obtained (n = 11), keyed to sample collection locality.

Because all seven of the moderate- to high-inclination Miocene Sespe clasts of pre-Mesozoic age contain post–1.2 Ga zircons, it is likely that most or all of the total population of moderate- to high-inclination clasts (19 of 34 samples, or 56%) have similar characteristics. Therefore, if our interpretation is correct that these characteristics indicate a Shinumo source, it places an important constraint on the erosion kinematics of the post-Laramide Cordillera.

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Because the only known exposures of the Shinumo Formation lie within a few hundred meters elevation of the bottom of eastern Grand Canyon, our interpretation supports the existence of a mid-Tertiary drainage connection, or Arizona River, between high-relief, eroding uplands in the eastern Grand Canyon region and the coast. Further, it is highly unlikely that a SW-flowing Arizona River running near the bottom of eastern Grand Canyon would have "jumped" out of Grand Canyon before reaching the coast. Assuming it did not, the only plausible course would have run through an existing western Grand Canyon, as also implied by a roughly equal mixture of ultradurable Tapeats (exposed only in western Grand Canyon) and Shinumo clasts suggested by the simple linear mixing models of the Piuma inclination spectra. Our results thus provide independent support for models that suggest western Grand Canyon was carved to within a few hundred meters of its current depth no later than 20 Ma, and perhaps as early as Late Cretaceous/Paleocene time, based on thermochronological evidence (e.g., Flowers et al., 2008; Wernicke, 2011; Flowers and Farley, 2012).

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