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FACTS AND HYPOTHESES REGARDING THE MIOCENE–HOLOCENE JEMEZ LINEAMENT, NEW MEXICO, ARIZONA AND COLORADO

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Abstract—The Miocene to Holocene Jemez Lineament (JL) consists of 10 volcanic fields stretching northeast from San Carlos in Arizona to Raton–Clayton in New Mexico and Colorado. We have tabulated data on volcanic style, magma composition, number of volcanic vents, age range, eruptive areas, eruptive volumes, presence of xenoliths and enclaves, significant sites, and significant references for all volcanic fields. The width of each field, based on total extent of mapped volcanic rocks perpendicular to the trend of the lineament, is highly variable; thus, as pointed out by others, magma ascent and volcanism cannot be controlled by a single fault or structure. Volcanic landforms are highly variable from field to field. Spatial-temporal trends are complex, and there is no systematic age progression in either direction; thus, the JL is not a hot-spot trend. There is also no compositional progression along the JL, although intermediate to silicic volcanism and tholeiitic basalts are most common toward the center of the lineament and the cross-cutting Rio Grande rift (RGR). Estimated surface areas and eruptive volumes of each field are highly variable. The Jemez Mountains volcanic field that formed at the intersection of the JL and RGR has erupted three times more volcanic products than all other volcanic fields combined. Recent geophysical studies have highlighted the presence of low-velocity (V_s<4.2 km/s) upper mantle all along the JL and have mapped stark differences in 2008–2010 JL seismicity on either side of the RGR. The only geothermal system along the JL with electrical potential (200–300°C) circulates within the 1.25 Ma Valles caldera (now a National Preserve), but low-temperature geothermal systems occur elsewhere along the JL. Two CO_2 gas fields are found at relatively shallow depths (≤1000 m) toward either end of the JL near the Springerville and Raton–Clayton volcanic fields. He- and C-isotopes indicate the gases are predominately derived from mantle sources. Extractable commodities have been perlite, pumice, sulfur and construction materials, and epithermal gold-silver was mined in the southeastern Jemez Mountains. The Grants uranium district underlies the Mount Taylor volcanic field, but a connection between remobilized (3 to 12 Ma) uranium deposition and magmatism has not yet been firmly established. Many researchers consider the older crustal structure beneath the JL to be a boundary between Proterozoic crustal provinces. Geophysical and geochemical observations support the idea that the Mesoproterozoic ancestry of this feature created fertile mantle lithosphere that has become part of the North American plate. Spacing between JL volcanic fields resembles volcano spacing found along many currently active subduction zones, although evidence for Paleoproterozoic arc-type volcanism is equivocal. Certainly, the alkaline affinity of volcanic rocks along much of the JL does not resemble the dominantly calc-alkaline magmatism of most subduction zones. Recent $^{40}Ar/^{39}Ar$ dating in the Raton–Clayton field indicates that the plate motion signal on time scales less than 1 Ma might constantly be present, but we currently do not have the spatial-temporal resolution to detect that pattern elsewhere along the JL.

INTRODUCTION

More than 60 years have passed since Mayo (1958) defined the Jemez Lineament (JL, aka, Jemez volcanic lineament) as an aid for mineral exploration (Fig. 1, Table 1). Simply stated, the JL is an apparent alignment of 10 volcanic centers stretching from east-central Arizona into the southeast corner of Colorado. The JL appears prominently on maps of late Cenozoic volcanic centers in New Mexico, Arizona and Colorado (Luedke and Smith, 1978a, b), which were designed for evaluations of igneous-related geothermal resources, volcanic hazards, volcano and volcano-tectonic studies and for general knowledge of volcanic rocks. Thereafter, the JL has been mentioned in numerous influential resource-, volcanic-, tectonic-, and seismic-focused papers about northern New Mexico (e.g., Chapin et al., 1978; Aldrich et al., 1981; Goff et al., 1981; Laughlin et al., 1982; Smith and Luedke, 1984; Aldrich, 1986; Spence and Gross, 1990; Magnani et al.,

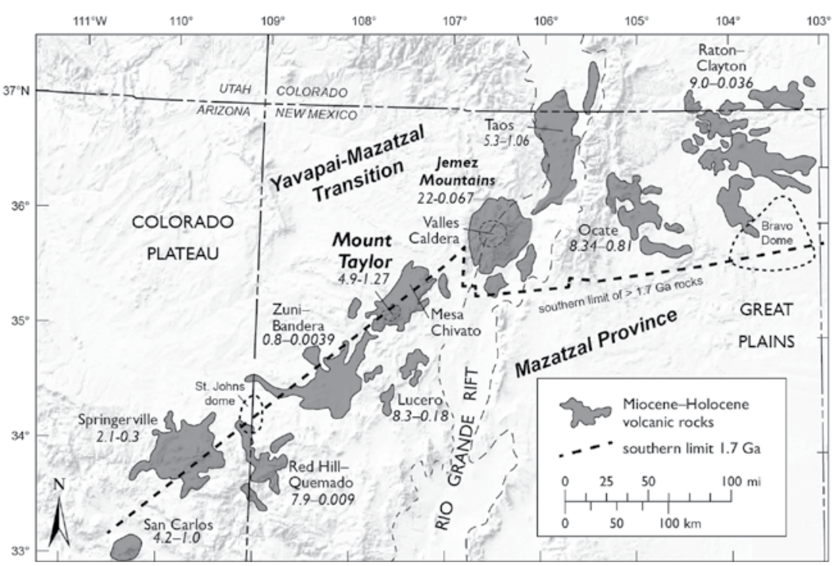

FIGURE 1. Index map showing the locations and age ranges (Table 1) of the volcanic centers that have been used to define the Jemez Lineament. The approximate boundary of the southern limit of 1.7 Ga metamorphic and igneous rocks (dashed line; Grambling et. al., 2015) and the Proterozoic provinces are also shown. The CO_2 gas fields along the JL are shown as dotted lines. The age units are Ma.

TABLE 1. General information on Miocene to Quaternary volcanic fields of the Jemez Lineament, Arizona, New Mexico and Colorado (Fig. 1).

Name, SW to NE	Volcanic Style	Magma Compositions	Number of Vents	Age Range	Eruptive Area (km^2)[1]	Estimated Volume (km^3)	Xenoliths & Enclaves	Significant Locations	References
San Carlos	Monogenetic flows, maar and diatreme	Basanite, alkali basalt and basaltic trachyandesite	3+	4.2 to 1 Ma?	≤50	≤1.5	Lherzolite	Peridot Mesa, Soda Springs vent and Black Hill	Wohletz, 1978; Holloway and Cross, 1978; Luedke and Smith, 1978a
Springerville	Monogenetic	Alkali basalt	405	2.1 to 0.3 Ma	3000	90	None reported	Cerro Hueco, Twin Knolls	Baldridge et al., 1989; Condit and Connor, 1996
(Mount Baldy)	(Alkalic shield)	(Trachyandesite to trachyte; minor basalt)	(20+)	(about 9 to 8 Ma)	335	33.5	(None reported)	(Mount Baldy and Mount Ord)	(Baldridge et al., 1989, p. 194)
Red Hill–Quemado	Monogenetic, some maars	Basanite, alkali basalt, basaltic andesite	40+	7.9 to 0.0094 Ma	≤200	≤6	None reported	Zuni Salt Lake (maar), Red Hill cone and maar	Crumpler and Aubele, 1990; McIntosh and Cather, 1994; Horning et al., 1996; Onken and Formen, 2017
Zuni–Bandera	Monogentic	Basanite, alkali to tholeiitic basalt; basaltic andesite	100+	0.8 to 0.0039 Ma	2460	75	Lherzolite	Bandera cone, El Calderon, McCarty's vent and flow	Laughlin et al., 1972; Luedke and Smith, 1978a; Laughlin et al., 1994; McIntosh, 1974
Lucero	Monogentic	Basanite, alkali basalt and tholeiitic basalt	25±	8.3 to 0.18 Ma	≤40	≤1.2	Lherzolite - granulite	Cerro Verde, Cerro Colorado, Black Mesa and Alkali Buttes	Baldridge et al., 1987, 1989; Zimmerer, person. comm., 2016
Mt. Taylor including Grants Ridge, Mesa Chivato and Puerco Necks	Stratovolcano of intermingled domes, flows, flanking monogenetic vents and plugs, and isolated domes	Basanite, basalt, trachybasalt, trachyandesite, trachydacite, trachyte, alkali rhyolite	300±	4.9 to 1.27 Ma	1700	≤120	Lherzolite, gabbro, norite, anorthosite, trochtolite, plus many mafic enclaves in andesitic–dacitic rocks	Mount Taylor, La Mosca, Grants Ridge, Cerro Chivato, Cerro Pelon, Cerro Redondo, Cerro Colorado, Cerro Aguila	Hunt, 1938; Lipman and Moench, 1972; Crumpler, 1980a, 1980b; Christiansen et al., 1983; Perry et al, 1990; Keating and Valentine, 1998; Goff et al., 2013, 2019, 2020
Jemez Mountains, Valles/Toledo calderas, Cerros del Rio and Santa Ana basaltic fields	Dome and flow complexes with intermingled lavas and tuffs, surmounted by two calderas with associated ignimbrites; flanked by three peripheral basaltic fields with some maars	Minor nephenlinite-basanite, alkali and tholeiitic basalt, trachyandesite, andesite, trachydacite, dacite, rhyolite and high-silica rhyolite	400 - 450	22 to 0.067 Ma	4400	2320	Granite, granulite, and other pre-caldera rocks in ignimbrites; mafic enclaves in andesitic-dacitic rocks	Valles caldera, Redondo Peak, Tschicoma Peak, Polvadera Peak, St. Peter's Dome, San Diego Canyon, Santa Clara Canyon, Battleship Rock, Cerro del Medio, Cerros del Rio, Montoso maar and Santa Ana Mesa	Smith and Bailey, 1968; Smith et al.,1970; Aubele, 1978; Eichelberger and Koch, 1979; Gardner et al., 1986; Self et al., 1986; Goff et al., 1989; Dunker et al., 1991; Gardner et al., 2010; Goff et al., 2011; Wolff et al., 2005, 2011; Kelley et al., 2013; Zimmerer et al., 2016
Taos Plateau	Broad basaltic shields, minor pillow basalt and maars with intermingled domes; very minor tuffs	Alkalic and tholeiitic basalts, basaltic trachyandesite, andesite, trachydacite, dacite and alkali rhyolite	150+	5.3 to 1.06 Ma	7000	420	None reported	Rio Grande Gorge, San Antonio Mtn, Taos Plateau, Black Mesa, Red Hill, No Agua dome, Cerro Chilfo, Cerro Negro	Lipman, 1969; Lipman and Mehnert, 1975; Dungan, 1987; Dungan et al., 1989; Thompson et al., 2015
Ocate (Mora)	Monogenetic flows and minor domes	Alkali basalt, minor andesite and dacite	45-50	8.34 to 0.81 Ma	1460	35	None reported	Wagon Mound fissure, Agua Fria, Cerrito Pelon, Cerro del Oro	Dungan et al., 1989; Nielsen and Dungan, 1985; O'Neill and Mehnert, 1988; Aubele and Crumpler, 2001
Raton–Clayton including Capulin	Monogenetic flows and cones of various ages intermingled with intermediate to silicic domes and flows	Basanite, alkali and tholeiitic basalt, trachybasalt, basaltic trachyandesite, trachyandesite and dacite; rare nephelinite	150+	9.0 to 0.036 Ma	3225 [2]	100	None reported	Capulin Mountain, Sierra Grande, Don Carlos Hills, Rabbit Ear Mountain, Carr Mountain	Stormer, 1972a, 1972b; Dungan et al., 1989; Aubele and Crumpler, 2001; Nereson et al., 2013; Zimmerer, 2019

[1]Measured from Luedke and Smith (1978a) or from sources cited in references column; estimated error ±20%.
[2]This area is the eruptive area from maps and GPS; see text.

2004, 2005; Schmandt and Humphreys, 2010; Nereson et al., 2013; Sosa et al., 2014; Channer et al., 2015). The objectives of this paper are to (1) describe the basic characteristics, (2) list some facts about the volcanic evolution, and (3) discuss some hypotheses about the origin and structure of the JL in order to place our new findings at Mount Taylor in a broader context (Goff et al., 2019, 2020).

FACTS REGARDING VOLCANISM ALONG THE JEMEZ LINEAMENT

Fact 1: Length of JL

When measured from the southeast edge of the San Carlos field (Peridot Mesa) to the northeast edge of the coalesced vent areas of the Raton–Clayton field, the JL is about 800 km long (Fig. 1; Luedke and Smith, 1978a, b). This length does not include long lava flows that extend into Oklahoma. The JL is not perfectly linear and has a pronounced right-stepping dogleg where it crosses the Rio Grande rift (RGR; Chapin et al., 1978). The JL west of the RGR follows the southeastern margin of the Colorado Plateau and the JL to the east cuts across the High Plains province.

Fact 2: Width of JL

The width of the JL is highly variable depending on the position and extent of coalesced vent and eruptive areas at each volcanic field (Fig. 1). For example, the maximum width based on total extent of mapped volcanic rocks perpendicular to the SW-NE trend of the JL is at the Springerville volcanic field (about 110 km wide) and at the Raton–Clayton field (about 125 km wide; Luedke and Smith, 1978a, b). However, the latter field has several pulses of volcanism and considerable open space among the various vents and flows (Stormer, 1972a; Dungan et al., 1989), thus an exact width of the field is equivocal. By comparison, the minimum width of the JL is at San Carlos (about 35 km wide).

Fact 3: The JL is not controlled by a single fault or set of structures

It should be clear from examination of Figure 1 and the remarks above that volcanic eruptions along the JL do not emanate from a single long fault, a narrow band of semi-parallel faults, an extended series of en-echelon faults, or a single deep-seated structure or zone of crustal weakness of northeast trend. JL volcanic fields and eruptive loci are too broad and too irregularly spaced to originate from a single structure or set of structures. We will elaborate on these points further in the "Proterozoic ancestry" section of this paper. Nonetheless, a few noteworthy but relatively short northeast-trending structures are found within the JL, such as (from southwest to northeast) the Cuates graben (N25E) in the Mount Taylor field (Goff et al., 2019), the El Malpais graben (N25E) in the Zuni-Bandera field, and the Jemez and Embudo fault zones (N55E) in the Jemez Mountains field (Aldrich, 1986). The Cuates graben contains 2.2 Ma monogenetic mafic vents and fault-controlled fissure vents (Goff et al., 2019), and the El Malpais graben is filled with the 3900-year-old McCartys and three older lava flows (Channer et al., 2015). The implied intersection of the Jemez and Embudo fault zones occurs beneath the Valles-Toledo caldera complex and associated rhyolitic volcanism.

Fact 4: The JL has a wide variety of volcanic landforms

The JL displays most if not all major volcanic landforms and structures (Table 1; Aldrich et al., 1981; Crumpler and Aubele, 2001; Crumpler, 2001): scoria cones, pahoehoe and a'a lava flows, lava shields, lava domes, blocky lava flows, extensive ignimbrites, volcanic necks and plugs, a large composite cone (Mount Taylor), many maar volcanoes, a few pillow basalts, the Valles-Toledo caldera complex, and two recognized diatreme (Wohletz et al., 1978; Goff et al., 2019). The only other stratovolcano along the JL is Sierra Grande, which is located in the center of the Raton–Clayton field and is primarily composed of two-pyroxene andesite flows that erupted 3.8±0.2 to 2.77±0.11 Ma (Stroud, 1997; Nereson et al., 2013). Sierra Grande, which is 10 km in diameter and has relief of 573 m above the Great Plains, is small in comparison to Mount Taylor but is a prominent landmark in this area. Truly, the variety of volcanic landforms along the JL contributes to New Mexico's oft-cited moniker as the "Volcano State."

Fact 5: Spatial-temporal patterns of volcanism are complex

The overall timing of volcanism along the JL from San Carlos to Raton–Clayton has no trend; thus the JL is not a hot spot track (Fig. 1; Luedke and Smith, 1978a; Aldrich, 1986), as stated by some researchers (e.g., Suppe et al., 1975; Morgan and Morgan, 2007). The youngest mafic rocks have erupted in the Zuni-Bandera volcanic field (McCartys and Bandera flows, about 4 and 10 ka, respectively; Laughlin et al., 1994; Dunbar and Phillips, 2004), and the youngest silicic eruptions formed in the Jemez Mountains volcanic field (East Fork Member rhyolites within Valles caldera at about 73–68 ka; Zimmerer et al., 2016). By comparison, the San Carlos field at the southwest end of the JL is 4.2 to 1.0 Ma (Wohletz, 1978; Holloway and Cross, 1978), whereas the youngest flow in the Raton–Clayton field is 36.6±6.0 ka (Zimmerer, 2019). Thus, the central JL has younger eruptions than the ends to the east and the west.

Interestingly, the increasing number of $^{40}Ar/^{39}Ar$ dates now available at individual volcanic fields reveals patterns indicating migration of activity toward the center of certain fields (Ocate, Raton–Clayton, Jemez; Olmstead and McIntosh, 2004; Nereson et al., 2013; Kelley et al., 2013, respectively). Recently published $^{40}Ar/^{39}Ar$ dates for nine vents in the Raton–Clayton volcanic field (368.2±7.3 ka to 36.6±6.0 ka) were combined with previously published dates to document a general pattern of eastward migration during the last 1.3 Ma that is not recognized in the older volcanic rocks in this field (Zimmerer, 2019).

Fact 6: The JL has no compositional progression

There is also no compositional progression of volcanism along the JL. Most volcanic products are alkaline to slightly alkaline (see Table 1 and references therein). Highly silica-undersaturated rocks erupted toward either end of the JL; nepheline-bearing trachytes are found in the Mount Baldy area of the Springerville field (Baldridge et al., 1989), and feldspathoidal lavas are observed in the Raton–Clayton field (Stormer, 1972b; Dungan et al., 1989). Some older nephelinite lavas were erupted in the southern Jemez Mountains volcanic field (Wolff et al., 2005), and a single nephelinite dike was discovered south of the Mount Taylor field (Goff et al., 2019). Some calc-alkaline rocks were produced toward the center of the JL at Mount Taylor (Goff et al., 2019; 2020), the Jemez Mountains volcanic field (Tschicoma Formation; Goff et al., 1989; Rowe et al., 2007; Kelley et al., 2013) and in the Taos field (Dungan, 1987; Dungan et al., 1989). Alkalic mafic rocks such as basanite are common within most of the JL but tholeiitic basalts are most common in the Jemez Mountains and Taos fields (Gardner et al., 1986; Wolff et al., 2005; Kelley et al., 2013; Dungan et al., 1989).

Fact 7: Surface areas of individual fields

The surface areas of most JL volcanic fields have been estimated previously (Table 1, see references) or by the current authors using available maps and other resources. For example, the area of Mount Baldy eruptions, sometimes included with the Springerville field, is calculated at 335 km^2 from the map in Baldridge et al. (1989, fig. 8). Some reported areas are unusually large. For example, the reported area of the Raton–Clayton field is 19,400 km^2 (Stormer, 1972b; Dungan et al., 1989, p. 474; Aubele and Crumpler, 2001), but examination of various maps (i.e., Luedke and Smith, 1978a) shows considerable open space between eruptive units. Most recently, Nereson et al. (2013) used GIS to calculate the total land-surface area covered by lava flows in the Raton–Clayton field to be 3225.5 km^2 and in the Ocate field to be 1457.6 km^2 (±20%). That being the case, we calculate the total surface area of eruptive products along the JL to be 23,870 km^2 (Table 1) with an estimated error of ±20%.

Fact 8: Erupted volumes of monogenetic fields

The erupted volume of most JL fields has not been calculated except for the Jemez Mountains and Mount Taylor, which have relatively large effusions of intermediate to silicic products (see below). The calculation of the volume of eruptive products is important in the assessment of volcanic hazards through the evaluation of the length of time, rates, and explosive behavior associated with past eruptions. These calculations are also valuable in estimating the flux of gases released during the formation of volcanic fields, as described later in this paper. Indeed, volumes are difficult to calculate because in most cases thickness is highly variable due to topography, and drill hole thicknesses over the given areas are not widely determined. For monogenetic lava fields, we have estimated an average thickness of 30 m based on our observations of variable lava flow thicknesses at many of these fields. In truth, most flows are not 30 m thick over broad areas, but this estimate accommodates increased thicknesses in ravines, vents and cones. Thus, the estimated volume of the Springerville field, considered to be one of the three largest monogenetic fields in the United States, is 90 km^3. For the Taos volcanic field, we have raised the average thickness to 60 m to accommodate canyon exposures along the Rio Grande gorge and the thicknesses of several large intermediate composition domes. The resulting estimate is 420 km^3. This estimate is not unreasonable, because Dungan et al. (1989) have calculated the volume of the extensive Servilleta Basalt component of this field at 200 km^3. Thus, the total erupted volume of (primarily) monogenetic fields is estimated at 880 km^3 with an estimated error of ±20%.

Fact 9: Erupted volumes of Mount Taylor and Jemez Mountains volcanic fields

Our estimate for the volume of the Mount Taylor stratovolcano is about 85 km^3 (Goff et al., 2020). Perry et al. (1990) estimated a volume of between 25 to 30 km^3 for monogenetic mafic lavas on surrounding mesas including Mesa Chivato. Recent detailed mapping on southwest Mesa Chivato (Goff et al., 2019) has recognized a 4-km-wide northeast-trending graben (Cuates graben) along the axis of Mesa Chivato that presumably contains previously unaccounted for ponded lavas. Additionally, Mesa Chivato contains a few trachyandesite to trachyte domes and flows (Crumpler, 1980a, b; Goff et al., 2019). The current authors generously estimate the volume of Grants Ridge Rhyolite center, associated mafic lavas, and the many volcanic necks along the Rio Puerco at ≤2 km^3. Thus, the total erupted volume of the Mount Taylor field is probably around 120 km^3.

Gardner (1985, p. 150) first calculated the volume of the main portion of the Jemez Mountains volcanic field at 2100 km^3. This total included 1000 km^3 for the "original" Keres group, 500 km^3 for the now-obsolete Polvadera Group (presently part of the Keres Group, see stratigraphic revisions in Goff et al., 2011, and Kelley et al., 2013), and 600 km^3 for the Tewa Group, mostly consisting of the Valles/Toledo calderas and the Bandelier Tuff. The volume of the Tewa Group, particularly the Bandelier Tuff, has since been raised to 800 km^3 because of geothermal drilling intercepts through intracaldera tuffs acquired in the 1980s and early 1990s (Goff, 2010; Goff et al., 2011). The volume of three monogenetic lava fields peripheral to the main Jemez Mountains field (El Alto, Cerros del Rio and Santa Ana Mesa) is estimated at ≤20 km^3 from maps and from the assumptions discussed above, insignificant when compared to the volume of the main Jemez Mountains field. Thus, the revised estimate for the Jemez Mountains volcanic field is 2320 km^3 or nearly 3 times greater than the combined erupted volume of all other volcanic fields along the JL.

Fact 10: JL P- and S-wave velocity anomalies

Several seismic experiments, including 2-D refraction and reflection lines and 3-D teleseismic arrays, have been deployed across the JL to assess the thermal state of the underlying mantle. These experiments include CD-ROM (Continental Dynamics–Rocky Mountains; Dueker et al., 2001; Yuan and Dueker, 2005; Zurek and Dueker, 2005, Magnani et al., 2004, 2005), LA RISTRA (Colorado Plateau–Rio Grande rift–Great Plains Seismic Transect; Gao et al., 2004; West et al., 2004; Wilson et al., 2005), and the Earthscope USArray transportable array (TA), which had a station spacing of 70 km across the United States. Over the years, several investigators have inverted various combinations of body and surface wave and receiver function information from all of these datasets to image the mantle and crust across the RGR and along the JL. Spence and Gross (1990) were the first to recognize low-velocity mantle beneath the JL using teleseismic data; these researchers were surprised to find that the mantle velocity anomaly beneath the JL was more robust than the signal from the RGR.

P- and S-wave travel time delays observed along the northwest-oriented LA RISTRA line were used by Gao et al. (2004) and West et al. (2004) to image an approximately 200-km-wide low-velocity zone centered on the RGR that extends to depths greater than 200 km, with S-wave velocities (V_s) as low as 4.2 km/s. This value is 6-7% below the global average and is indicative of partial melt. Wilson et al. (2005) used receiver functions to demonstrate that the small-scale convection associated with the RGR is restricted to the upper mantle. In contrast, the S-wave velocities beneath the JL where the LA RISTRA line passes across Mount Taylor (4.3–4.4 km/s at 70 km) are not as low as those beneath the RGR along this particular transect (Gao et al., 2004; West et al., 2004; Wilson et al., 2005). Here, the low mantle velocities associated with the RGR are the more robust signal.

The CD-ROM refraction line across the eastern JL (Fig. 2) indicates that the crust is thinner and that upper mantle velocities are lower under the JL compared to areas to the north in Colorado (Snelson et al., 2005; Levander et al., 2005). Other data collected as part of the CD-ROM experiment, including seismic reflection (Magnani et al., 2004, 2005) and passive seismic surveys (Yuan and Dueker, 2005; Zurek and Dueker, 2005), reveal complex crustal structure and low P- and S-wave velocities in the mantle beneath the JL near Las Vegas, New Mexico, along the northern edge of the Ocate volcanic field (Figs. 2, 3). Data from these profiles will be discussed further in the "Proterozoic ancestry" section.

More recent work by Schmandt and Humphreys (2010) used available body-wave data to show that, at a depth of 90 km, the entire JL and the southeastern margin of the Colorado Plateau are underlain by low-velocity mantle that extends down to at least 195 km. Lin et al. (2014) identified low S-wave velocities in the lower crust over a broad region that includes the southern Rocky Mountains, the western JL/eastern Colorado Plateau transitional boundary, and the High Plains of northeastern New Mexico. Fu and Li (2015) investigated mantle structure using radial anisotropy. Positive radial isotropy characterizes the JL west of the RGR, indicating horizontal alignment of magma in the form of sills. In contrast, negative radial anisotropy is more common in the eastern lineament, suggesting the presence of vertical dikes or zones of migrating melts. Sosa et al. (2014) used LA RISTRA and TA data to create a 3-D model of the area surrounding the RGR. Like Gao et al. (2004) and West et al. (2004), Sosa et al. (2014) found low-velocity mantle at shallow depths below the RGR and low-velocity mantle along the eastern margin and below the Colorado Plateau.

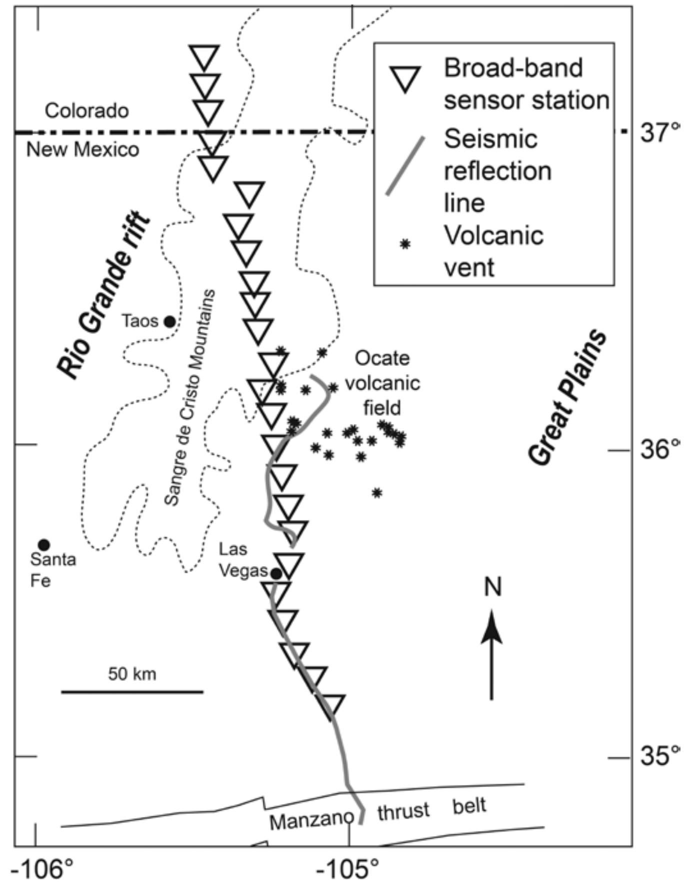

FIGURE 2. Map of the location of the CD-ROM seismic reflection line (Magnani et al., 2005) and the passive seismic array (Zurek and Dueker, 2005; Yuan and Dueker, 2005) relative to the Ocate volcanic field on the JL.

Shen and Rizwoller (2016) inverted seismic data from the TA across the entire United States; this analysis identified well-known seismic velocity anomalies in the western United States, as described above, and discovered previously unrecognized anomalies in the midwestern and eastern United States. This inversion indicates that S-wave velocities are less than 4.2 km/s at depths greater than 70 km along the southeastern margin of the Colorado Plateau, along the RGR, and along the projection of the JL onto the High Plains as far east as the western edge of the Raton–Clayton volcanic field, near Capulin and the other young volcanic centers in that field. Based on recent experimental work by Takei (2017), Schmandt et al. (2019) note that V_s<4.2 km/s at 75 km is likely indicative of small percentages of partial melt.

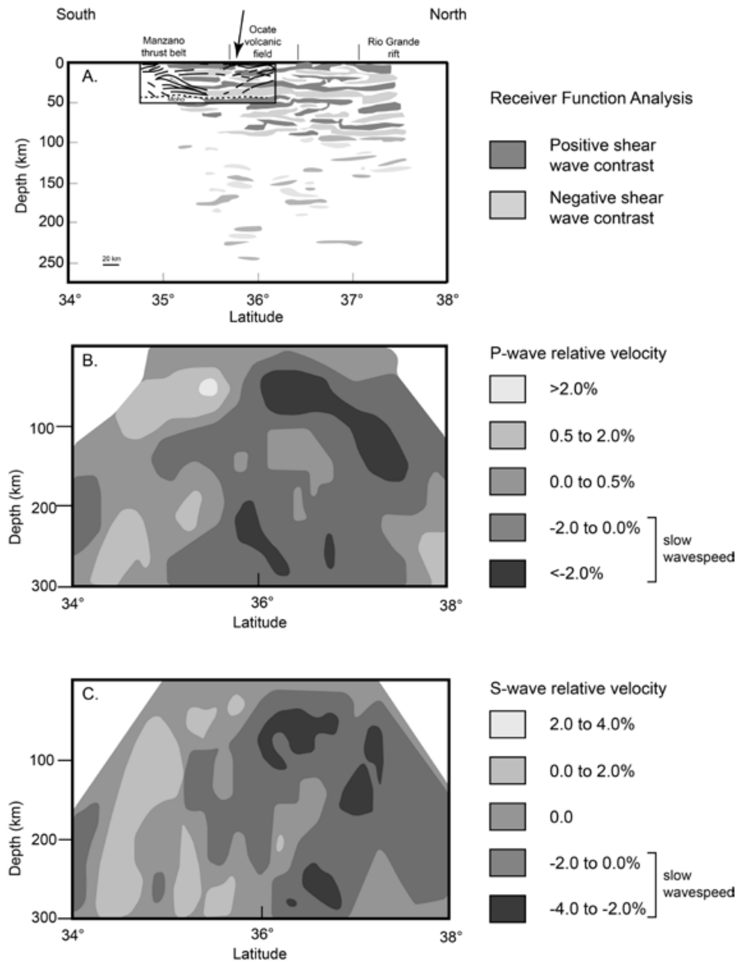

FIGURE 3. **A**) Receiver function inversion results of Yuan and Dueker (2005) and the interpretation of the seismic reflection data of Magnani et al. (2005; black lines in box near the surface between latitudes ~35° and ~36°) showing the duplex structure. The arrow points to a significant break in crustal layering, as discussed by Zurek and Dueker (2005). Note that the layering in the mantle that is imaged by the receiver function analysis is more prominent on the north end of the line. **B**) P-wave velocity inversion (Yuan and Dueker, 2005). **C**) S-wave velocity inversion (Yuan and Dueker, 2005). The darker shading in (B) and (C) highlights the relatively low P- and S-wave velocities in the mantle beneath the JL and the RGR, which are interpreted to represent mantle that is hot and/or contains partial melt. All four seismic data sets are at the same scale.

Fact 11: JL seismicity

Nakai et al. (2017) summarize seismicity in the vicinity of the JL during the timeframe when the Earthscope US Array TA and the CREST (Colorado Rockies Seismic Transect) seismic networks were deployed in Colorado and New Mexico between 2008 and 2010. The seismicity is starkly different in the western and eastern sections of the JL. Most notable is the aseismic nature of the eastern JL, which is bounded by areas of seismicity to the northwest and the southeast—a so-called "seismic halo." The seismicity to the northwest has been tied to coal-bed methane production in the Raton Basin (Rubenstein et al., 2014), but the band of seismicity to the southeast of the JL aligns with a zone of seismicity first recognized by Sanford et al. (2002). An earthquake recorded in 2010 in this area had a moment tensor indicative of strike-slip faulting. Possible hypotheses to explain the pattern of seismicity in the eastern JL will be discussed later in the "Proterozoic ancestry" section.

The alignment of seismicity along the JL west of the RGR is complex, alternating between N5E (rift structures) and N65E (Nakai et al., 2017). These authors note that the seismicity is inboard of the physiographic margin of the Colorado Plateau. Earthquakes are concentrated along the western JL, indicating complex deformation associated with the transition from northwest-southeast extension in the southeast Colorado Plateau and east-west extension within the Rio Grande rift. These authors point out the lack of seismicity in the Jemez Mountains. Sanford et al. (1991) and House and Roberts (2019) also note an absence of seismicity beneath the Valles caldera west of the RGR. Aprea et al. (2002) attribute this paucity to ductile deformation in proximity to a magma body.

An unusual northeast-trending swarm of 49 earthquakes (M_L 0.8–2.0) occurred over the course of a year (starting in November of 2008; most were within the first 60 days) in Zuni Canyon, which is on the north flank of the Zuni Mountain near the terminus of one of the Zuni-Bandera lava flows (Nakai et al. 2017). The activity migrated toward the southwest during the swarm. In the Mount Taylor region, the small earthquakes measured between 2008 and 2010 (M_L 1.0–2.5) form a diffuse, northeast-trending array on Mesa Chivato (Fig. 4). One swarm lies at the northeast tip of the mesa on a N25E-trending structure that parallels the faults of the Cuates graben of Goff et al. (2019).

Fact 12: JL electrical resistivity

Feucht et al. (2019) measured electrical resistivity of the crust and upper mantle using magnetotelluric methods along a profile line at latitude 36.25° that crosses the JL in the vicinity of the Ocate volcanic field and the very western end of the Raton–Clayton field. Both broadband and long period instruments were used. The phase-tensor azimuths are bimodal, striking northwest (N25W, paralleling the tipper strike) and northeast (N75E), indicative of anisotropy. The upper crust is resistive to depths of 15–25 km, and the middle to lower crust is conductive under the rift. The conductive zone is present both east (150 km) and west (100 km) of the rift. Measurements of the electrical fields are sensitive to north-south and east-west orientations of features along two-dimensional profiles, and the data collected during this study record strong anisotropy parallel to and perpendicular to the rift (Feucht et al., 2019). In detail, the electrical resistivity structure sensed by east-west electrical fields shows pockets of conductive material less than 50 ohm-m in the vicinity of the Ocate (40 km) and Raton–Clayton (25 km) volcanic fields. The short segment of the mantle lithosphere imaged beneath the High Plains north of the primary trend of the JL is resistive (100–1000 ohm-m).

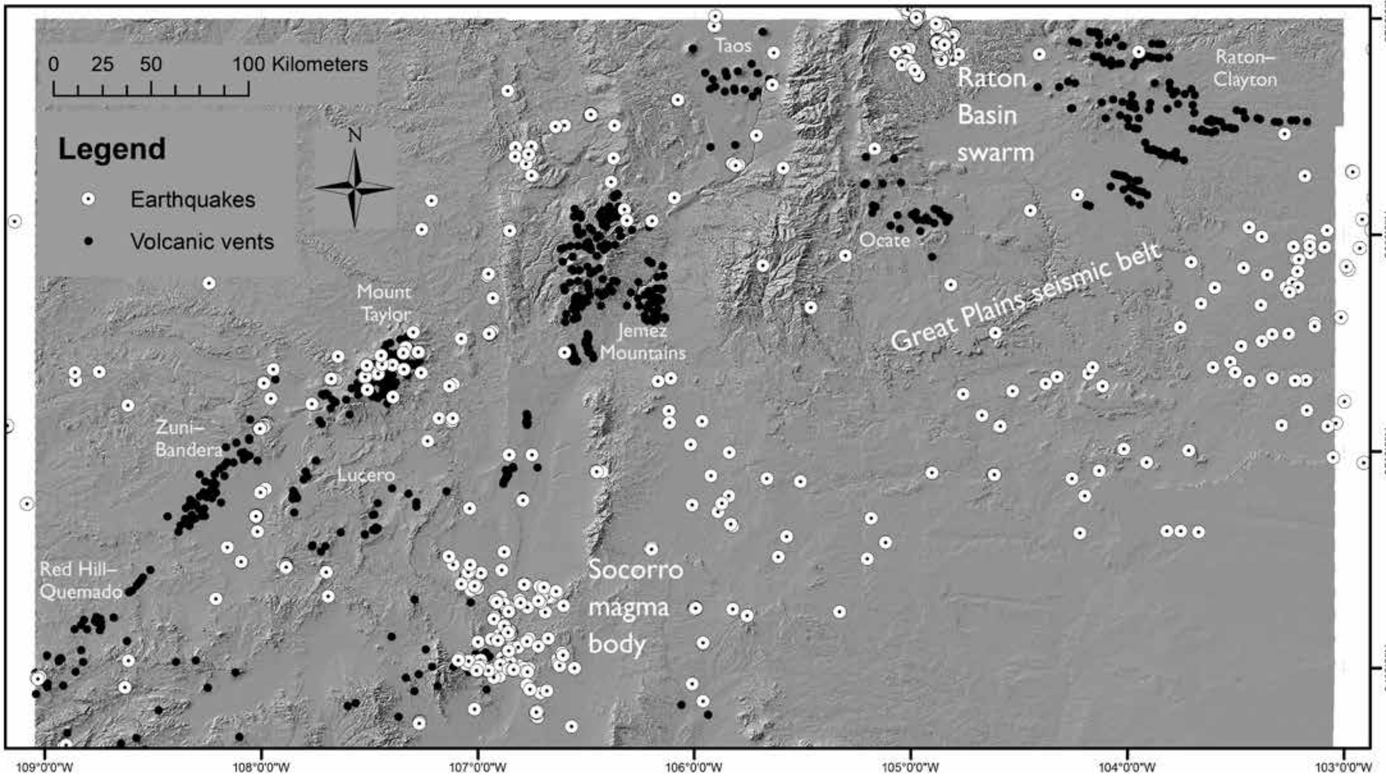

FIGURE 4. Earthquakes detected between 2008 and 2010 by Nakai et al. (2017) across northern New Mexico and volcanic vents in the JL. Note the clustering of earthquakes in the Raton Basin, which likely have an anthropogenic origin (Rubinstein et al., 2014), and the diffuse northeast-striking band of earthquakes south of the Ocate and Raton–Clayton volcanic fields. Also note the difference in the alignment of vents east and west of the RGR.

Fact 13: JL heat flow

Heat flow distribution and maps for the High Plains of northern New Mexico and southern Colorado were presented in a series of papers by Reiter et al. (1975), Edwards et al. (1978), and Swanberg (1979). A broad thermal anomaly that is characterized by an average heat flow value of 100 mW/m^2 (60–138 mW/m^2) has been observed at the boundary of the southern Rocky Mountains and the High Plains in the Raton Basin near Ocate and Raton–Clayton volcanic fields (Reiter et al., 1975; Edwards et al., 1978; Swanberg, 1979; Kelley, 2015). For comparison, heat flow on the Taos Plateau in the RGR is 115–130 mW/m^2 (Reiter et al., 1975).

Less is known about heat flow west of the RGR in the vicinity of the JL. Heat flow values in the general vicinity of Mount Taylor are 69–107 mW/m^2 (Eggleston and Reiter, 1984). Heat flow values in the Zuni-Bandera field are 75–191 mW/m^2 (Minier, 1987), and values in the Red Hill to Zuni Salt Lake region are highly variable at 43–170 mW/m^2. The 170 mW/m^2 value is associated with 13.4–9.9 ka Zuni Salt Lake maar (Onken and Forman, 2017). Recent work by Kelley et al. (2016) in the general vicinity of the 8.3 to 0.192 Ma Lucero volcanic field (Baldridge et al., 1987) suggests that two distinct bands of water wells located near the Hickman and Red Lake fault zones on the margins of the Acoma basin tap the San Andres-Glorieta aquifer and have elevated discharge temperatures of 34 to 52.8°C (depths 615 to 884 m).

In contrast, deep conductive heat flow in the central Jemez Mountains within the western Valles caldera is 200–400 mW/m^2 and can exceed 450 mW/m^2 (Sass and Morgan, 1988), which can only be caused by magma underlying the caldera at 7±1 km depth (Aprea et al., 2002). This is 2.7 times greater than maximum heat flow along the RGR and 4 to 5 times greater than heat flow along the rest of the JL.

Heat flow contour maps drawn by Edwards et al. (1978) and Blackwell et al. (2011) show elevated heat flow that roughly follows the JL. The elevated heat flow zone also wraps around the southeastern and southern edge of the Colorado Plateau. Elevated heat flow greater than 100 mW/m^2 associated with the RGR is also highlighted on these maps.

Fact 14: Geothermal systems along the JL

Considerable research since the middle 1970s shows that the only high-temperature geothermal system along the JL, and indeed in all of New Mexico, circulates within the southwest segment of the Valles caldera (Goff and Gardner, 1994; Goff and Janik, 2002). The Valles system consists of a near-boiling to boiling, gas-rich acid cap overlying a 200 to 300°C, chloride-rich brine at depths of 0.4 to 2 km. Although capable of producing modest amounts of electric power (20-30 MWe), the geothermal resource has been retired and is now part of the Valles Caldera National Preserve (Goff and Goff, 2017). Hot springs that discharge along the Jemez fault zone in Cañon de

San Diego are part of a hydrothermal outflow plume originating in the caldera and do not have high-temperature geothermal applications.

Several other hot springs and shallow thermal aquifers occur within the JL and its broad dogleg across the north-central RGR (Fig. 1; e.g., Goff and Goff, 2015; see Summers, 1976, for a comprehensive citation list). All such sites have had geothermal evaluations conducted at some time during the last 40+ years (Montezuma, Ojo Caliente, Taos area, San Ysidro/ Jemez Pueblo, Lucero, Acoma Basin, Upper Frisco, Springerville). None of them have electrical potential, one is a successful commercial spa, one has possible green-house potential and most are currently "wild" (e.g., Stone, 1979; Vuataz et al., 1984; Albrecht et al., 2011; Goff and Goff, 2017; Blomgren et al., 2016).

An underground aquifer reported to be at least 40°C circulates at depths of ±1500 m in the large uranium mines flanking the northwest and east sides of the Mount Taylor volcanic field (Goff et al., 2019). This aquifer (or aquifers) has never been evaluated for geothermal potential, but the reported temperature and depth would match measured geothermal gradients in this region (25 to 35°C/km) and calculated heat flow of about 105 mW/m^2 (Nathenson et al., 1982). The age and volume of volcanic rocks in the Mount Taylor region are too old and too small (≥1.3 Ma; 120 km^3) to indicate much, if any, high-temperature geothermal reservoir potential (T≥200°C; Duffield and Sass, 2003).

Fact 15: CO_2 gas fields along the JL

Two carbon-dioxide (CO_2) gas fields, influenced by relatively young magmatism and high thermal gradients in the region, are located within the JL. The 3600 km^2 Bravo Dome CO_2 gas field is found in the southeastern portion of the Raton–Clayton volcanic field (Fig. 1; Broadhead, 1990, 2019; Sathaye et al., 2016; Brennan, 2017). The "discovery well" was drilled in 1916 to a depth of 763 m. Although this was an oil test well, 25 million cubic feet gas per day (MCFGPD) was produced from the Permian Tubb Sandstone. By the 1930s, a few wells from the field supplied gas to a processing plant that sold dry ice and bottled CO_2 to markets in Colorado and surrounding states. This enterprise went out of business long ago; now, the Bravo Dome field is unitized and supplies CO_2 gas by pipeline south to the Permian Basin oil fields of west Texas and southeastern New Mexico for enhanced oil recovery. About 47 trillion cubic feet of gas was produced between 2004 and 2018.

The gas in this field is nearly pure, greater than 98 mol-% CO_2 and is primarily produced from the Tubb Sandstone. The overlying Cimarron Anhydrite seals the reservoir. However, carbon and helium isotope studies show that the gases are mantle derived and rise in the north part of the field adjacent to the Raton–Clayton volcanics (Gilfillan et al., 2008; Brennan, 2017). The estimated pre-production volume of trapped gas at Bravo Dome is 1.3 Gt (Sathaye et al., 2016).

Although isotopic data indicates a mantle origin for the gases, the vast quantities of CO_2 in the Bravo Dome field have elicited controversy (Broadhead, 1990; Sathaye et al., 2014). Could all that CO_2 really come from the mantle or melts within it? Assume for the moment that the gas originated solely from degassed mafic magma along the JL in the Raton–Clayton volcanic field. The maximum solubility of CO_2 in tholeiitic to basanitic magma at 1100°C and 15 kbar (60 km) is roughly 4500 ppm (0.45 wt-%, Holloway and Blank, 1994, fig. 14A). The original volume of gas in the field (1.3 Gt) requires 290 x 109 metric tons of mafic magma to degas. The density of mafic magma is roughly 3 g/cm3 or 3 x 109 metric tons per km3. Thus, the original volume of unproduced gas could come from 96.6 km3 of mafic magma. Coincidentally, this volume is about the same as the volume of erupted products at the Raton–Clayton volcanic field (100 km3±20%, Table 1). Thus, deep magmatism along the JL is a plausible source for the CO_2 at Bravo Dome.

The 1800 km^2 St. Johns CO_2 gas field is located in east-central Arizona in the northwest part of the Springerville volcanics. Although the first gas well was drilled in 1939, this field is less studied than Bravo Dome (Rauzi, 1999; Gilfillan et al., 2008; Eastman and Muir, 2012). The main CO_2 reservoirs are in sandstones of the Permian Supai Formation (200 to 700 m deep) that directly overlie Proterozoic basement at depths of 800 to 1300 m. Intercalated anhydrite beds in the Supai provide localized seals or cap rocks to the CO_2. Since the mid-1990s, producers have intended to ship gas by pipeline to the Permian Basin for enhanced oil recovery, but financing for this endeavor was never successful. A secondary intent, also unsuccessful, was to separate and sell helium gas. More recently, a DOE-financed project intends to develop an enhanced geothermal (or "hot dry rock") project in Proterozoic basement using compressed CO_2 as the heat exchange fluid rather than water (Eastman and Muir, 2012). Gas compositions at the St. Johns field are variable, about 83 to 98 mol-% CO_2, yet the carbon and helium isotope values indicate a mantle origin for these constituents (Gilfillan et al., 2008). Presumably, the source for the gases is the Springerville magmatic system. Estimated CO_2 reserves are 445 billion m^3. This volume of gas is considerably smaller than the original volume of gas trapped at the Bravo Dome field (1.3 Gt); thus, degassing of deep mafic magma in the Springerville area of the JL is a reasonable source for the CO_2 trapped in the St Johns field.

Fact 16: Mineral/ore deposits associated with volcanism along JL

Ignoring sand, gravel, road metal, and cinder quarries, there are several locations along the JL where volcanism has formed extractable mineral deposits. Of course, pre-Puebloan and later Native American cultures mined obsidian from specific rhyolite domes and tuffs in the Jemez Mountains, Taos, and Mount Taylor volcanic fields (Shackley, 1998, 2005; Glascock et al., 1999; Shackley and Goff, 2016). Pumice has been mined extensively from both the basal pyroclastic fall deposits of the Otowi Member, Bandelier Tuff (Guaje Pumice Beds), and from the El Cajete Pyroclastic Beds in the Jemez Mountains volcanic field (McLemore and Austin, 2017). The pumice is used for construction, decorative stone, abrasives, and stonewashed jeans. Small amounts of pumice were sporadically mined from the east end of Grants Ridge Rhyolite center in the Mount Taylor volcanic field during 1941 to 1967 (McLemore and Austin, 2017).

New Mexico is the leading producer of perlite in the United States. Perlite is extracted from several large mines in the No Agua rhyolite domes and flows of the Taos volcanic field. Perlite was mined from 1953 to 1991 at the Grants Ridge Rhyolite center in the Mount Taylor volcanic field. Perlite also occurs in older rhyolites and tuffs in the southern Jemez Mountains volcanic field but has not been mined (McLemore and Austin, 2017).

Epithermal gold-silver quartz veins and related base metals were explored and mined in the Cochiti mining district in the southeast Jemez Mountains volcanic field from the late 1880s until about 1916. Sporadic production continued into the 1940s (Hoard, 2007), and several mining companies reexamined the district into the 2000s. The ore is high-grade but low tonnage; gold values are highest near the tops of the veins. Mineralization is associated with 7 to 6 Ma Bearhead Rhyolite dikes, small intrusions, domes and flows (WoldeGabriel and Goff, 1989). An estimated $1.4 million dollars in gold, silver, copper, and lead was produced from 1894 to 1963, but the value of Jemez Mountains pumice (estimated $31 million) exceeds the value of precious metals (McLemore and Lueth, 2017). The forest fires of 2011 in the southeast Jemez Mountains have now made this mining district nearly inaccessible. Additional major gold-silver mining districts, such as the Old and New Placers, Cerrillos, and Elizabethtown-Baldy deposits (McLemore and Lueth, 2017), lie along the JL, but the ages of source volcanic and intrusive rocks pre-date the age of volcanism along the JL.

Until the 1980s, the Grants district was the largest uranium producing area in the United States and was possibly fourth worldwide. Two large mines lie on opposite sides of Mount Taylor; San Mateo is to the northwest and Jackpile is to the southeast. Several hundred bore holes were drilled through the volcanic rocks to trace uranium-bearing Jurassic strata, mostly the Westwater Canyon Member of the Morrison Formation, beneath the western and northern parts of the volcanic field (McLemore and Chenoweth, 2017; Goff et al., 2019). Renewed interest starting in 2007 sparked re-staking of claims in the Mount Taylor area, but the 2011 Fukushima nuclear accident caused a precipitous drop in uranium prices and demand. For the most part, uranium deposits have a Jurassic age of roughly 130 Ma, but there are a few redistributed uranium deposits that date as young as 3 to 12 Ma (McLemore and Chenoweth, 2017, table 7). It is presently not known if some redistributed uranium in the Grants area results directly from magmatic activity associated with the Mount Taylor volcanic field.

HYPOTHESES ABOUT THE FORMATION OF THE JEMEZ LINEAMENT

Proterozoic Ancestry

The JL has long been considered to be a boundary between Proterozoic crustal provinces (e.g., Cordell, 1978; Lipman and Mehnert, 1979; Aldrich et al., 1981). The aligned volcanic fields of the JL overlie a complex, NE-trending transition zone between two Proterozoic terranes that were accreted against Laurentia. The Yavapai province lies to the northwest and is comprised of several 1.80–1.70 Ga oceanic arcs that came together along the southern margin of the continent during the Yavapai orogeny at 1.71–1.68 Ga (Whitmeyer and Karlstrom, 2007). The Mazatzal province located to the southeast of the lineament is composed of 1.68–1.60 Ga rocks that formed in continental-margin and back-arc settings; these rocks were deformed during the 1.65–1.60 Ga Mazatzal orogeny (Whitmeyer and Karlstrom, 2007). The Mazatzal orogeny, which is related to subduction of the southern margin of the Yavapai block beneath the Mazatzal block, created a broad zone of mid-crustal folding and faulting that is preserved in modern mountain ranges in northern New Mexico and southern Colorado (Shaw and Karlstrom, 1999). The JL is considered to be a fundamental crustal boundary, because it marks the southern extent of 1.7 Ga rocks at the surface (Karlstrom and Humphreys, 1998), and lead isotope data in Arizona indicate different crustal compositions on either side (Wooden and DeWitt, 1991). The Paleoproterozoic crust was subsequently affected by widespread 1.45 to 1.35 Ga felsic volcanism.

Surface geologic mapping of Proterozoic rocks has yet to reveal the exact location of the boundary between provinces. Grambling et al. (2015) found that 1.4 Ga granites at the south end of the Sierra Nacimiento were derived from pre-1.7 Ga crust using hafnium isotope data, which indicates that Yavapai rocks extend further south than previously thought; thus the Proterozoic boundary is somewhere south of the Sierra Nacimiento. Similarly, Proterozoic rocks dated at 1.63 and 1.43 Ga (Strickland et al., 2003) are exposed in the core of the Zuni Mountains, which lie between the Zuni-Bandera volcanic field and Mount Taylor within the JL. Although a few scattered outcrops of ultramafic rocks that are chemically consistent with origin in ocean crust are preserved in the range, no clear evidence of the Mazatzal orogeny is preserved in the Zuni Mountains. Instead, the deformation that is preserved is 1.43 Ga (Strickland et al., 2003). The boundary must be north of the Zuni Mountains.

A seismic reflection line that was part of the CD-ROM experiment (Figs. 2, 3) images the Proterozoic boundary in the eastern JL in unprecedented detail, showing oppositely dipping reflection bands that converge near the volcanic lineament (Magnani et al., 2004; 2005). South of the latitude of Las Vegas, New Mexico, crustal reflectors at depths greater than 15 km dip north, whereas deep crustal reflectors north of Las Vegas dip south, forming what is interpreted to be a “doubly vergent crustal duplex.” At shallower crustal levels, the north-vergent Manzano thrust zone near the south end of the line and north-vergent Pecos thrust and nappe near the north end of the line form prominent reflectors. The crustal boundary here is broad (~100 km) and diffuse because of the collision of a Mazatzal island arc with the Yavapai margin. As collision continued, the margin was subducted along a south-dipping zone, creating the bivergent convergence geometry. Notable sub-horizontal, fairly continuous, strong reflectors on the seismic line at depths of 10–15 km are interpreted as mafic sills of possible Mesoproterozoic (Amarante et al., 2005) or late Cenozoic age. Given the fact that many of the lavas in the Raton–Clayton volcanic field show evidence of crustal contamination (e.g., Ramos et al., 2019), the latter interpretation is more likely.

Zurek and Dueker (2005) used receiver functions, which image sharp vertical changes in velocity, to map layering in the crust and the mantle across the JL (Figs. 2, 3). Many of the layers and truncations of layers in the receiver function image align nicely with features identified on the seismic reflection line (Fig. 3a), but one south-dipping truncation in the receiver function profile, highlighted by the arrow in Figure 3a, is not obvious in the reflection line because a data gap (the reflection line could not be run through the town of Las Vegas). Thus, the two methods complement one another and the receiver function image supports the "doubly vergent crustal duplex" interpretation. Note that this feature does not penetrate all the way through the crust; it probably soles into the brittle-ductile transition in the crust. The Moho is relatively flat at a depth of 40 km across this profile, with one small step up at the north end of the reflection line. Model resolution is poor north of latitude 37°.

The tomographic body-wave images of Yuan and Dueker (2005; Figs. 3b, 3c) show that both P- and S-velocities are reduced beneath and north of the Ocate volcanic field, in the same region of pronounced layering in the 100-km-thick mantle lithosphere. Zurek and Dueker (2005) and Yuan and Dueker (2005) attribute these observations to a juxtaposition of lithospheric materials of different chemical composition. Hydration of the mantle during Proterozoic subduction has resulted in the preservation of "sub-solidus material" in lithosphere that can more readily form partial melt (Yuan and Dueker, 2005).

Similarly, Wolff et al. (2005) identified four potential mantle sources for volcanic rocks in the Jemez Mountains volcanic field but concluded that only two sources are likely important: 1) Proterozoic oceanic lithosphere enriched by basaltic melt and 2) convecting asthenosphere. Modification of the mantle by subduction of Proterozoic oceanic crust associated with complex, piecemeal accretion of island arcs against the North American continent could be the source of the array of volcanic rock types preserved along the JL (Dueker et al., 2001; Wolff et al., 2005). In a later study, Rowe et al. (2015) examined spatial and temporal variations in water, chlorine, fluorine, and sulfur in basaltic magmas in the RGR and on either end of the JL and found an east to west decrease in volatile enrichment that is likely explained by a combination of varying mantle sources and early removal of metasomatized lithospheric mantle.

Volcano Spacing and Ancient Plumes

If the subduction suture theory can be entertained, then possibly the spacing between JL volcanic centers may reflect inherited conduits, plumes, or vertical zones of weakness caused by upward magma migration in a chain of a previous arc system. Several researchers have commented on "regular spacing" in subduction zone (arc) volcanoes (i.e., Marsh, 1979; Tatsumi and Kogiso, 2003). Marsh (1979) proposed that arc centers result from gravitational instabilities (i.e., Rayleigh-Taylor instabilities) in developing magma(s) at the approximate boundary of a downward plunging slab and overlying asthenosphere. Shimozuru and Kubo (1983) published an average volcano spacing of 58±24 km (1σ) for currently active subduction zones. These authors also noted that single chain arcs generally indicate high-angle dip and faster subduction rate of the subducted slab, whereas broad or multi-chain arcs, perhaps reflected in the broad width of the JL, signified lower dip angle and slower rates of convergence. On the other hand, de Bremond d'Ars et al. (1995) measured volcano distributions at 16 arcs containing 479 volcanic systems and concluded that volcanoes are randomly distributed at convergent margins. In spite of these contradictions, we measured the spacing between the approximate centers of the 10 volcanic fields along the JL and the average value is 94±24 km (1σ). Whether or not JL volcanic centers reveal a "regular" or random spacing reflective of past subduction in the Proterozoic, current intraplate volcanism (the last 20 my) is not compositionally similar to "andesitic" volcanism typical of convergent margins.

Thermal and Structural Connections to the Colorado Plateau

The western JL is closely aligned with the southeastern margin of the Colorado Plateau. Several researchers (e.g., Chamberlin, 2007; Crow et al., 2011) have noted a robust pattern of inboard sweep of volcanism through time attributed to thermal erosion along the western and southern margins of the Plateau, but that pattern is not as clear along the southeastern margin. Aldrich and Laughlin (1984) found that extension directions in the Colorado Plateau are southwest-northeast, whereas extension directions are east-west in the RGR, and that the difference in orientation is accommodated in the JL on the west side of the RGR. The accommodating structures, which are oriented north-northeast, disrupted the thermal erosion pattern and focused volcanism into voluminous centers like Mount Taylor and the Jemez Mountains volcanic fields.

East versus West

Chapin et al. (1978) defined the JL to include the aligned volcanic centers toward the northeast across the RGR onto the High Plains with a dogleg, but Aldrich et al. (1981) restricted the extent of the JL to the section west of the RGR. Both the eastern and western segments share many similarities. The compositional ranges and general timing of volcanism (mostly less than 15 Ma) along each belt are similar. The entire JL is underlain by upper mantle with low P- and S-wave speeds (e.g., V_s<4.2; Shen and Ritzwoller, 2016) and is characterized by elevated heat flow. However, there are significant differences, particularly in the stress orientations and distribution of modern seismicity. Aldrich et al. (1981) determined that the least principal stresses, derived from the orientations of faults and dikes that are less than 5 Ma, are northwest-southeast in the western JL and northeast-southwest in the eastern JL. Vent alignments west of the RGR trend northeast and have more east-southeast to easterly striking trends to the east, especially in the Raton–Clayton field (Fig. 4). Seismicity levels are concentrated perpendicular to sigma 3 in the western JL. In contrast, few earthquakes occur in the eastern JL and the diffuse events generally trend northeast (Fig. 4).

Nakai et al. (2017) offer two explanations for the essentially aseismic nature of the eastern JL. First, these authors note that the presence of solidified mafic sills of Cenozoic or Mesoproterozoic age imaged on the reflection line of Magnani et al. (2004, 2005) may have strengthened the crust. Nakai et al. (2017) use the analogy of the seismic parabola around the Snake River Plain in Idaho and Wyoming as a hypothesis to explain the observations in northeastern New Mexico, although the driver for the northeastern New Mexico system is in the lithospheric mantle. Alternatively, the low-velocity zone in the mid-crust found by Lin et al. (2014) and Fu and Li (2015) may indicate elevated temperature in the crust, which could cause aseismic ductile deformation in the eastern JL that is surrounded by brittle failure induced by differential thermal stresses.

We reiterate that the center of the Jemez Mountains occupied by the Valles caldera and the underlying magma chamber is also aseismic, presumably because molten rock attenuates seismic waves. This seismic characteristic was first recognized by Suhr (1981; see also Goff et al., 1989, p. 395-396) but has been discussed most recently by Sanford et al. (1991) and House and Roberts (2019).

CONCLUSIONS

Surface and subsurface characteristics of volcanic fields along the JL have been extensively studied since the 1970s, resulting in many factual observations. Geophysical and geochemical measurements and data support the idea that the Paleoproterozoic ancestry of this feature created fertile mantle lithosphere that resides within the North American plate (Spence and Gross, 1990). Exactly what triggers the seemingly random temporal and spatial pattern of volcanic eruptions along this zone is a bit of a mystery, but subsequent tectonic events, including the hydration of the mantle by subduction of the Farallon Plate during Laramide deformation and younger regional-scale uplift of the Rocky Mountain region (e.g., Nereson et al., 2013), has created rising magmas that erupt at any time along this zone. Recent $^{40}Ar/^{39}Ar$ dating in the Raton–Clayton field (Zimmerer, 2019) suggests eastward migration of volcanism paralleling plate motion. The plate motion signal might constantly be present on timescales <1 Ma, but we do not yet have the resolution in our current sampling strategies to see that pattern in older volcanic rocks of the JL.

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