Late Cretaceous crustal hydration in the Colorado Plateau, USA, from xenolith petrology and monazite geochronology

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ABSTRACT

Formation and subsequent modification processes of lower crust play important roles in evolution and rheological models for continental lithosphere. In the last two decades, numerous xenolith studies have documented metasomatism of continental mantle in the Rocky Mountain region of North America. However, much less attention has been paid to whether and to what extent these processes may have affected the crust. We address the nature and timing of hydrous alteration of extant deep Proterozoic crust in the Colorado Plateau through a petrological and geochronological study of a xenolith from the Red Mesa diatreme in the 30–20 Ma Navajo volcanic field. Fluid-related alteration is widespread in sample RM-21, with the main features recorded by breakdown effects in feldspar, garnet, and allanite, and the production of secondary assemblages characterized by lower Ca plagioclase (or albite) + muscovite + biotite + calcite + monazite ± zoisite at estimated conditions of 0.8–0.7 GPa (or ~27 km depth) and 480 °C. Th-Pb dating by ion probe reveals a period of monazite crystallization at 70–65 Ma, interpreted to reflect late alteration of the high-temperature metamorphic assemblage with hydrous fluid introduced by the shallowly subducting Farallon slab. Similar to previous suggestions involving mantle hydration, the growth of low-density hydrous phases at the expense of high-density, anhydrous minerals, which are abundant in unaltered Proterozoic crust, if sufficiently widespread, could have contributed to elevated topography in the Colorado Plateau and regionally across the Rocky Mountains and High Plains.

INTRODUCTION

Physical and chemical studies of xenoliths are powerful tools for reconstructing modification processes and alteration histories of crust and mantle lithosphere. In the last two decades, numerous xenolith studies have documented metasomatism of continental mantle in the Rocky Mountain region of North America. Interpretations range from coincidence with Paleoproterozoic accretionary events (Carlson and Irving, 1994; Lester and Farmer, 1998; Downes et al., 2004; Farmer et al., 2005; Facer et al., 2009) to Late Cretaceous–Paleogene hydration of the lithospheric mantle by dewatering of a shallow Farallon slab (e.g., Smith et al., 2004). The latter interpretation has a number of significant implications, including the potential for mantle de-densification to produce subsequent uplift (Hess, 1954; McGetchin and Silver, 1972; Silver and McGetchin, 1978; Humphreys et al., 2003) and the distance inboard at which subducting slabs can retain their ocean-derived fluid (e.g., Lee, 2005; Currie and Beaumont, 2011; Sommer and Gauert, 2011). However, much less attention has been paid to whether and to what extent this process may have affected the crust, in part, due to far fewer documented records of crustal hydration and because dating metasomatic processes is notoriously difficult.

In the Colorado Plateau region, ultramafic mantle and crustal xenoliths from the Navajo volcanic field (NVF) have all been long recognized as recording evidence of extensive hydration (e.g., McGetchin and Silver, 1972; Smith and Levy, 1976). Several eclogite and peridotite xenolith studies have suggested mantle hydration events ranging in age from the Late Cretaceous to the Oligocene, based on U-(Th)-Pb dating of zircon or monazite interpreted to reflect growth or recrystallization catalyzed by a hydrous fluid (Usui et al., 2003; Smith et al., 2004; Smith and Griffin, 2005; Schulze et al., 2015). Several suites of deep crustal xenoliths sourced from depths of 20 km or more reveal variable degrees of hydration that occurred at depth prior to eruption (Broadhurst, 1986; Selverstone et al., 1999), but again the timing of this hydration has remained elusive. This study provides new mineralogical and textural observations, and geochronological data from one of the Red Mesa diatreme samples studied by Selverstone et al. (1999). Results from in situ Th/Pb dating of monazite associated with secondary alteration assemblages suggest that deep crustal hydration occurred at the end of the Late Cretaceous Epoch, and they indicate that fluids with a similar source to those that permeated much of the lithospheric mantle also made it into the lower ~20 km of the crust. If the hydration reactions were sufficiently pervasive and widespread, the associated de-densification could have significantly influenced the surface uplift history of the Colorado Plateau.

GEOLOGIC AND GEOPHYSICAL SETTING AND BACKGROUND

Colorado Plateau

The Colorado Plateau covers an area of ~3.4 × 10⁵ km² in Colorado, New Mexico, Utah, and Arizona (Fig. 1). Lateral accretion of island-arc and oceanic terranes to the southern margin of the Archean Wyoming craton between 1.8 and 1.6 Ga created much of the crystalline continental crust and mantle lithosphere in the southwestern United States, including the Colorado Plateau (Bowring and Karlstrom, 1990). This period of prolonged convergence established two distinct, NE-trending crustal...
Figure 1. (A) Simplified map of Colorado Plateau region after Selverstone et al. (1999; where CO—Colorado, NM—New Mexico, UT—Utah, AZ—Arizona). Gray-filled regions are exposed Proterozoic basement rocks. Key minettes and serpentinized ultramafic breccia pipes (SUM) of the Navajo volcanic field are labeled in bold (RM—Red Mesa, ME—Mule Ear, MR—Moses Rock, CV—Cane Valley, GR—Garnet Ridge, A—Agathla Peak, CB—Chaistla Butte, BP—Buell Parks, GK—Green Knobs, TT—The Thumb, MB—Mitten Rock, S—Shiprock). The approximate locations of laccolith intrusions are shown labeled in italics and include CM—Carrizo Mountains, UM—Ute Mountain, LP—La Plata, R—Rico, HM—Henry Mountains, AB—Abajo Mountain, and LS—La Sal Mountains. The traces of two seismic lines through the area are shown (Roller, 1965; Wilson et al., 2005 for LA RISTRA) along with EarthScope USArray station locations (open circles). Proposed Yavapai-Mazatzal Province boundary traces from several studies are shown by thick dashed lines. (B) Inset showing region of western United States, major Proterozoic terrane boundaries from Whitmeyer and Karlstrom (2007), and key mantle and crustal xenolith localities (SH—Sweetgrass Hills, BM—Bearpaw Mountains, Hs—Homestead, LH—Leucite Hills, SL—Stateline, NVF—Navajo volcanic field, VT—Vulcans Throne, KH—Kilbourne Hole, DH—Dish Hill, Ci—Cima, OC—Oak Creek, BC—Big Creek, GFTZ—Great Falls Tectonic Zone).
provinces (e.g., Yavapai and Mazatzal) of largely juvenile crust, with some older intermixed crustal and/or underlying mantle components (Bennett and DePaolo, 1987; Karstrom and Bowring, 1988, 1993; Livacari and Perry, 1993; Bickford et al., 2008; Schulze et al., 2008). Widespread granitic magmatism that affected much of the region in the Mesoproterozoic was long considered to be anorogenic (Anderson, 1983). However, much of the previously presumed Paleoproterozoic aluminosilicate “triple point” metasedimentary domain in northern New Mexico is now known to have 1.49–1.46 Ga depositional ages, thus requiring ca. 1.45 Ga burial and widespread regional metamorphism (e.g., “Picuris orogeny”; Daniel et al., 2013). Extensional basins and NW-striking fault networks in the Grand Canyon Supergroup record two periods of intracratonic deformation at 1.1 and 700–800 Ma, coincident with the far-field Grenville tectonic events for the former and the breakup of Rodinia and initiation of the Cordilleran rift margin for the latter (Timmons et al., 2001, 2005).

Wide uplifts and basinal structures developed in the northern part of the plateau during the late Paleozoic (Ancestral Rockies; e.g., Marshak et al., 2000), whereas relatively minor faults and monoclinal folds developed in the plateau during the Late Cretaceous to Early Tertiary Laramide orogeny (e.g., Davis, 1978). Syn-Laramide magmatism (75–45 Ma) was almost exclusively limited to the Colorado Mineral Belt (Armstrong, 1969; Mutschler et al., 1987), which extends from the northern Colorado Front Range southwest to the Four Corners region. These include the ca. 70 Ma granitoid laccoliths of the Rico and La Plata Mountains on the eastern edge of the plateau and the Ute and Carrizo Mountains near the Four Corners area (Semken and McIntosh, 1997). Other small-volume granitoid intrusions on the plateau include the 31–20 Ma La Sal, Henry, and Abajo Mountain laccoliths (Nelson et al., 1992). The extent of Laramide orogenesis on the plateau stands in stark contrast to that in adjacent provinces, i.e., the Basin and Range, Southern Rocky Mountains, and Rio Grande Rift, which experienced substantial deformation and magmatism during this time. Accumulation of marine and nonmarine sediments, and the widespread occurrence of the Mancos Shale constrain the elevation of the plateau to near sea level for much of the Phanerozoic (e.g., Morgan and Swanberg, 1985). Present elevations in the plateau therefore require ~2 km of surface uplift since the Late Cretaceous.

**Xenoliths of the Navajo Volcanic Field**

The NVF encompasses a >30,000 km² region in the central Colorado Plateau that contains more than 50 volcanic centers of two principal types—minettes and serpentinized ultramafic microbreccias (Fig. 1). Minettes are thought to be products of small-percentage melting of volatile-rich upper-mantle garnet peridotite (e.g., Roden, 1981), whereas serpentinized ultramafic microbreccias are thought to represent subsolidus eruptions of upper-mantle garnet lherzolite (e.g., Roden, 1981), whereas serpentinized mantle peridotite and other ultramafic xenoliths in Late Cretaceous to Tertiary subduction complexes in the western United States (Hunter and Smith, 1981; Smith, 1995; Smith et al., 1999; Lee, 2005; Smith and Griffin, 2005; Li et al., 2008; Sommer et al., 2012).

A more limited number of studies focused on crustal xenoliths from the NVF and yielded valuable information about the age, evolution, and chemical composition of the crust. These studies revealed details of the Paleoproterozoic and Mesoproterozoic crustal growth processes and associated deep crustal metamorphism (Wendlandt et al., 1996; Silverstone et al., 1999; Crowley et al., 2006). Geochemical, isotopic, and petrologic studies showed that most crustal xenoliths from the NVF are consistent with original magmatic or depositional formation in a 1.8–1.6 Ga juvenile island-arc setting (Wendlandt et al., 1993; Mattie et al., 1997; Silverstone et al., 1999). Distinct pressure-temperature (P-T) paths recorded by crustal xenoliths from the northwest versus southeast portions of the NVF are interpreted to reflect different types of crustal formation during northwest-dipping subduction along the proposed Yavapai-Mazatzal lithospheric boundary ca. 1.7 Ga (Silverstone et al., 1999). Xenoliths from the northwest diatremes record counterclockwise P-T paths, consistent with heating, burial, and subsequent reequilibration during subduction or with magmatic heating of crust above the slab followed by accretion of the arc. Clockwise P-T paths reported for southeast crustal xenoliths, on the other hand, suggest that these rocks underwent only partial burial prior to unroofing and were therefore not part of the overriding plate. Both felsic and mafic crustal xenoliths record 1.4 Ga as the most recent period of high-grade metamorphism (Wendlandt et al., 1996; Crowley et al., 2006). U-Pb zircon dates for most felsic igneous xenoliths in the NVF record crystallization events between 1.75 and 1.69 Ga, consistent with proposed Proterozoic subduction (Condie et al., 1999; Crowley et al., 2006), whereas a subset of mafic xenoliths crystallized at 1.43 Ga (Crowley et al., 2006).

As in mantle xenoliths, evidence of late-stage hydrous alteration exists in deep crustal xenoliths in the NVF. While most studies have argued that the alteration occurred before the xenoliths were brought to the surface, based on observed high-pressure minerals like phengitic muscovite and zoisite + kyanite, the absolute age and regional extent of the alteration have been debated (McGetchin and Silver, 1972; Ehrenberg and Griffin, 1979; Padovani et al., 1982; Broadhurst, 1986; Silverstone et al., 1999). The various models for crustal hydration hinge on its timing (see Discussion), but no geochronological constraints have previously been available, other than the observation that hydration must have postdated 1.4 Ga deep crustal magmatism (Crowley et al., 2006).

**Modern Crustal Structure and Composition from Xenoliths and Geophysics**

The plethora of xenolith-based studies and both active and passive source geophysical surveys have inspired several attempts to compile crustal structure and lithologic properties for the central Colorado Plateau (Fig. 2). Crustal thickness estimates in this region range from ~40 to 48 km, and most seismic studies appear to agree on the presence of an ~10–12-km-thick lower-crustal layer with distinctly higher velocity than that of the middle and upper crust (e.g., Roller, 1965; Wolf and Cipar, 1993; Wilson et al., 2005; Bashir et al., 2011). A relatively mafic composition for this deep layer has been suggested based on high absolute velocities, high Vp/Vs values, and gravity models that are consistent with
Figure 2. Crustal structure for the Colorado Plateau near the Four Corners from geophysics and xenolith compilations. (A) Crustal model from Condie and Selverstone (1999). Note this column extends to base of the crust, whereas the other two panels (B and C) extend to 100 km depth. Bt—biotite; Grt—garnet; Px—pyroxene. (B) P-wave velocity models from xenolith studies (McGetchin and Silver, 1972; Padovani et al., 1982; this study), seismic refraction (Wolf and Cipar, 1993; reinterpretation of data from Roller, 1965), and passive source seismic studies (receiver function analysis from USArray by Bashir et al., 2011). Receiver function study of Bashir et al. (2011) identified a prominent lower-crustal layer constrained between 30 ± 0.9 km and the Moho at 42 ± 0.8 km with Vp/Vs of >2.0, suggesting mafic composition. (C) Crustal and eclogite xenolith groups (light gray—felsic and intermediate; medium gray—mafic amphibolite and granulite; black—eclogite) plotted by approximated depth at time of Navajo volcanic field eruption and versus approximate density. Pressure estimates, which are converted to depth, for xenolith groups are from Selverstone et al. (1999), Wendlandt et al. (1993), Bashir et al. (2011), Smith et al. (2004), and Schulze et al. (2015). Densities for crustal xenoliths are from approximations by Condie and Selverstone (1999). Eclogite xenolith density estimate was taken from McGetchin and Silver (1972). P-wave velocity and density for sample RM-21 (star in B and C, respectively) were calculated at 480 °C and 0.75 GPa using the Hacker and Abers (2004) physical properties spreadsheet. Ctd—chloritoid. The numbers (1-6) next to each rectangle in C are the number of xenolith samples in that group.

SAMPLE DESCRIPTION, PETROLOGY, AND GEOCHRONOLOGY

The studied sample (RM-21) is from the Red Mesa ultramafic microbreccia diatreme (Fig. 1) in the northeastern NVF (~37.01°N, 109.5°W on the map of Smith and Levy, 1976). Selverstone et al. (1999) classified the sample as belonging to a group of felsic garnet-biotite gneisses with either granitoid or sedimentary arkose protoliths. Estimated modal proportions from quantitative automated scanning electron microscope analysis (QEMSCAN) in this study are listed in Table 1 and Figure 3. The sample contains a primary mineral assemblage of Qz + Pl + Kfs + Bt + Ms + Grt (mineral abbreviations after Whitney and Evans, 2010), with easily recognizable accessory allanite, zircon, ilmenite, and rare pyrite. Static alteration is moderate but widespread and is most obviously manifest by replacement textures such as more sodic plagioclase, albite, and secondary muscovite after primary plagioclase (Fig. 3C and 4). Additional secondary minerals associated with the alteration assemblage include calcite, monazite, rutile, and zoisite (the latter reported by Selverstone et al., 1999). A weak gneissic foliation is defined by shape-preferred orientation of quartz andfeldspar aggregates and by aligned mica.

Major Mineral Textures and Compositions

Modal mineralogy (Table 1) was determined using QEMSCAN (Hoal et al., 2009) at the Colorado School of Mines. Microprobe analysis on silicate minerals was conducted with a JEOL-8600 electron microprobe
Figure 3. Images showing simplified modal mineralogy and textural characteristics of thin section from sample RM-21. (A) Full thin-section plane-polarized light photomicrograph. (B) Near full-section quantitative automated scanning electron microscope analysis (QEMSCAN) map. Area mapped in A is shown by black outline. Pixel resolution is 20 μm. Lettered regions correspond to geochronological target areas. (C) Higher-resolution QEMSCAN map of subarea shown in B. Pixel resolution is 4 μm. Mineral abbreviations are after Whitney and Evans (2010). The mineral subscripts refer to different generations of those minerals, as explained further in the text.
equipped with a W-filament at the University of Colorado–Boulder using natural and synthetic standards. A beam current of 100 nA, accelerating voltage of 15 kV, focused beam (~1 µm), and count times of 50–100 ms were used during X-ray mapping of garnet and plagioclase to identify zoning patterns. For quantitative analysis, an accelerating voltage of 15 kV, beam current 20 nA, a focused beam for garnet and a defocused beam (spot size of 5–10 µm) for mica and feldspar, to reduce volatilization of Na, Ca, and K, were used. Additional plagioclase analyses at identical beam conditions were obtained on a JEOL-8230 microprobe at the University of Colorado to include Sr analysis. Quantitative analyses of monazite were acquired with a Cameca SX-100 at the University of Massachusetts–Amherst and with a JEOL-8230 at the University of Colorado. Both instruments are equipped with a LaB6 electron gun for enhanced beam resolution, and with 4 (out of 5) wavelength-dispersive spectrometers equipped with large-area monochromators. The same correction routine was used, and results from both instruments are identical within error. Analytical conditions were 15 keV, 200 nA, and a focused beam (~0.7 µm). At these conditions, the analytical volume in monazite is roughly a 1-µm-diameter sphere. Details of the analytical setup on the Cameca SX-100 can be found in Condit et al. (2015). A similar setup was used on the JEOL-8230, with the difference that a lower current was used (50 nA) to further reduce the beam diameter (~0.4 µm), and the peak intensities were background-corrected using the mean atomic number correction for all elements (Donovan and Tingle, 1996; Donovan et al., 2016). Additional details on the analyzed elements, standards, counting times, detection limits, and interference corrections for analyses obtained on the JEOL-8230 are listed in Data Repository Table DR1.

The mineral compositions for RM-21 reported by Selverstone et al. (1999) are supplemented by results from this study (Table 2). Garnet occurs as locally resorbed anhedral porphyroblasts ranging from 300 to

### Table 2. Representative Silicate Mineral Compositions for Sample RM-21

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<th>Oxide</th>
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<th>Ms2</th>
<th>P1</th>
<th>P3</th>
<th>Ab</th>
<th>P1</th>
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**End members**

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**Note:** Mineral abbreviations are from Whitney and Evans (2010). The mineral subscripts refer to different generations of those minerals, as explained further in the text. n.a.—not analyzed. Mineral formula recalculations are based on 12 oxygens in Grt, 11 for Ms and Bt, and 8 for Pl.

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1. GSA Data Repository Item 2017180, Table DR1: Analytical setup for monazite analysis (JEOL-8230 University of Colorado–Boulder), is available at http://www.geosociety.org/datarepository/2017, or on request from editing@geosociety.org.

Figure 4. Wavelength dispersive X-ray maps showing major compositional variation and replacement features of an altered plagioclase domain in sample RM-21. (A) K X-ray map. (B) Ca X-ray map. Blue dots show locations of secondary monazite. Not all individual grains are shown by dots because several occur closely spaced. The allanite grain partially visible on the right side of the map is the same grain shown in Figures 6C, 6D, and 7A. Mineral abbreviations are after Whitney and Evans (2010).
1200 μm in their largest dimension. X-ray maps and quantitative transects indicate relatively uniform garnet interior compositions, wherein the highest Mg # (0.08), highest Ca content (X_{ca} = 0.18), and lowest Mn content (X_{mn} = 0.18) occur. Gradual decreases in Mg# toward the margins, particularly where the garnet is in contact with biotite, and increases in Mn likely reflect exchange and net transfer reactions, respectively, during retrogression. Fractures in garnet are locally lined with chlorite.

Similar to Selverstone et al. (1999), multiple generations of plagioclase and mica (biotite and white mica) are recognized (Fig. 5). In this study, two generations of plagioclase as well as albite are distinguished. Pl1 is the most anorthite-rich generation (X_{an} = 0.22–0.15) and occurs as locally preserved unzoned cores (Figs. 3C, 4, and 5A). The second generation (Pl2, X_{an} = 0.12–0.07) occurs in cracks, along twin contacts in Pl1, and as other patchy domains of intergrowth with secondary muscovite, biotite, and calcite that are contained within Pl1 grains. The albite (X_{ab} = 0.03–0.01) occurs locally as clean overgrowths, particularly where in contact with resorbed K-feldspar and early biotite. Two texturally and chemically distinct generations of each muscovite and biotite are recognized (Figs. 5B and 5C). Early Ms1 and Bt1 are coarser (up to several hundred micrometers wide), are scattered throughout the matrix, and help define a weak foliation. Secondary Ms2 occurs as fine (<10–100 μm) grains, are scattered throughout the matrix, and as a replacement phase in and around allanite, but there are two morphological subvarieties. Some grains, particularly those associated with large, moderately well-preserved allanite grains, show evidence of recrystallization, with compositionally distinct secondary allanite locally replacing the original mineral (Fig. 6). This secondary allanite (Al1) is slightly more calcic (Fig. 6D) and more depleted in light rare earth elements (LREEs) than the original igneous allanite (Alig).

Monazite occurs in two general textural settings, in direct association with either allanite or plagioclase. Type 1 monazite occurs exclusively as a replacement phase in and around allanite, but there are two morphological subvarieties. Some grains, particularly those associated with large, moderately well-preserved allanite grains, are equant or tabular, singular grains, most commonly less than 10 μm in average diameter. They occur in fractures and along altered allanite margins where Al1 is common (Fig. 6C), suggesting coeval growth with Alig. Thus, we refer to this population as Mnz1. In the more extensively pseudomorphed allanite grains, a second variety of type 1 monazite exhibits an irregular patchy morphology that

![Figure 5](https://example.com/figure5.png)

**Figure 5.** Mineral composition diagrams. (A) X_{An}-X_{An′}-X_{Ab} ternary diagram distinguishing Pl1, Pl2, and Ab. (B) Al-Fe-Mg ternary diagram distinguishing Bt1 and Bt2. (C) [Fe + Mg]-AlIV-AlVI ternary diagram distinguishing Ms1 and Ms2. Mineral abbreviations are after Whitney and Evans (2010).
Figure 6. (A–C) Backscatter electron images showing selected euhedral and only moderately altered allanite grains in sample RM-21. The grain in A contains readily visible oscillatory (igneous?) growth zoning. As a result of adjusting the contrast in this image to bring out the oscillatory zoning, the secondary Al domains are rendered black. The Al domains are shown in A with the red dotted lines, whereas they are visible in B and C as dark gray. Monazite occurrence in allanite from C is shown in Figure 7A. (D) Ca X-ray map of allanite grain in C showing distinctly higher Ca composition in Al, (yellow) compared to that in the igneous allanite (purple/red/orange). Mineral abbreviations are after Whitney and Evans (2010).

Figure 7. Backscatter-electron images showing selected monazite textural settings in sample RM-21. Ion probe analyses and dates are labeled. (A) Relatively well-preserved euhedral allanite grain with altered margins and numerous small type 1 monazite grains (mix of MnZ and MnZ2). (B) Type 2 monazite in altered plagioclase domain. (C) Heavily pseudomorphed allanite grain with multiple mats of polycrystalline type 2 monazite. (D) Euhedral pseudomorph of allanite with partial replacement. Mineral abbreviations are after Whitney and Evans (2010).
appears to be generated by mats of thin monazite flakes (Fig. 7C). These mats can be as large as 40 μm in their longest contiguous dimension, but they are more commonly less than 20 μm. Due to their association with the pseudomorphing assemblage, we refer to this variety as Mnz₁.

The second textural type of monazite grains (type 2) is represented by small (<5 μm) anhedral crystals that appear to be generated by mats of thin monazite flakes (Fig. 7C). These mats can be as large as 40 μm in their longest contiguous dimension, but they are more commonly less than 20 μm. Due to their association with the pseudomorphing assemblage, we refer to this variety as Mnz₁.

Due to small grain sizes and the matted nature of some monazite occurrences, obtaining high-quality chemical compositions proved challenging due to contamination from surrounding phases (recognizable by elevated elements such as Al, Si, Fe, Ca, or K). Some representative analyses of good to acceptable quality are listed in Table 3 with some notable observations. There are roughly three compositional trends recognized: rare earth element (REE)–rich, (Th,U)-rich, and (Sr,Ca,S)-rich trends (Fig. 8A). The latter is the most notable observation, with elevated SO₃ (2.5 wt%–3.5 wt%), SrO (2.7 wt%–3.5 wt%), and CaO (4.0 wt%–5.7 wt%). These include some of both type 1 (allanite association) and type 2 (plagioclase association) occurrences. Enrichments in the anhydrite-ceilstone (Ca,Sr) SO₄ component are uncommon but have been observed at or above these levels in several localities, including in monazite from kimberlites in eastern Siberia (Chakhmouradian and Mitchell, 1999) and in hydrothermal monazite from albite-muscovite schist and ore-bearing muscovite-quartz veins in the Tauern Window, Austrian Alps (Krenn et al., 2011). Sulfur is positively correlated with Ca and Sr and negatively correlated with P and REEs (Fig. 8B), suggesting that the coupled substitution REEE³⁺ + P⁵⁺ ⇔ (Ca,Sr)²⁺ + SO₄ is responsible for the incorporation of the anhydrite-celestine component into the monazite (Chakhmouradian and Mitchell, 1999; Krenn et al., 2011). Deviation from the expected 1:1 correlation in Figure 8B suggests the presence of excess Ca and Si, most likely due to the presence of silicate (allanite, plagioclase) or carbonate (calcite) in the direct vicinity of, or included in, these micron-scale monazite grains. These compositions translate to 10–24 mol% of the anhydrite (Ca,Sr) SO₄ component in the monazite solid solution (Fig. 8A). Other monazite grains, also occurring in both type 1 and type 2 settings, contain more normal concentrations of these elements (<1%). The second observation is that all type 1 monazite, with or without elevated S or Sr, has moderate to high ThO₂, typically ~3 wt% and as high as ~6 wt%, whereas most (but not all) type 2 monazite is almost pure REE-phosphate with low Sr and S contents and lower concentrations of Th, mostly near or below the microprobe detection limits, but locally up to 2 wt% ThO₂.

**Table 3. Monazite and Allanite Compositions for Sample RM-21**

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<th>Oxide</th>
<th>Mnz type 1</th>
<th>Mnz type 1 mixed</th>
<th>Mnz type 2</th>
<th>Mnz type 2 mixed</th>
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<td>0.22</td>
<td>0.32</td>
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</tr>
</tbody>
</table>

Note: n.a.—not analyzed; b.d.l.—below detection limit; Mnz—monazite.

*Analyzed at University of Massachusetts–Amherst. All others were analyzed at the University of Colorado.

1Reanalysis of same grain.
Figure 9. Pressure-temperature diagram summarizing the estimated peak and retrograde alteration conditions for crustal xenoliths from Red Mesa and Garnet Ridge diatremes. Peak and retrograde metamorphic conditions for mafic xenoliths from Selverstone et al. (1999), abbreviated as S99 in the figure, are shown with light-gray ovals. Data for felsic xenoliths (including RM-21) from Selverstone et al. (1999) include the dark shaded polygons (peak conditions) as well as the large light-gray shaded region (range of Grt-Bt temperatures from rim compositions) and the Si = 3.3 apfu for phengitic muscovite (barometer source from Massonne and Shreyer, 1987). Data shown in black are for sample RM-21 from this study and include (1) temperature estimates based on the Ti-in-biotite thermometer from Henry et al. (2005) for Bt1 (higher temperature) and Bt2 (lower temperature) and (2) Si (apfu) in phengitic muscovite from pseudosection analysis. Mineral abbreviations are after Whitney and Evans (2010).

Figure 8. Variation in monazite composition. (A) Ternary diagram highlighting differences in monazite-xenotime (REEPO$_4$), cheralite-huttonite (Ca,Th) (Pi)O$_4$, and anhydrite-celestine (Ca,Sr)SO$_4$ components. This latter was calculated based solely on the measured S content in order to eliminate the effect of excess Ca. Regions encompassing analyses from lettered geochronological target areas and from the central plagioclase domain shown in Figure 3C are indicated. Mnz—monazite; Aln—allanite. (B) Diagram illustrating the enrichment in anhydrite-celestine component. Asterisk indicates Ca and Si are integrated in cheralite and huttonite components, and with excess Ca and Si. REE—rare earth elements; UMass—University of Massachusetts–Amherst; CU Boulder—University of Colorado–Boulder.
Thermobarometry

Peak metamorphic conditions for crustal xenoliths from the northern group of diatremes in the NVF (Red Mesa, Garnet Ridge) range from 1.15 GPa and 780 °C to 0.65 GPa and 675 °C, with the latter conditions from sample RM-21 (Fig. 9; Selverstone et al., 1999). Retrograde conditions for the hydrous alteration of these xenoliths were also estimated by Selverstone et al. (1999) as 0.7–0.9 GPa at 450–550 °C using the Fe-Mg exchange thermometer for rim compositions of garnet and biotite and Si-in-phengite barometry as calibrated by Massonne and Schreyer (1987).

We supplement these estimates further by evaluating the Ti-in-biotite thermometer (Henry et al., 2005) for both generations of biotite, and by calculating Si-in-phengite isopleths using forward petrological modeling with Perple X 6.6.8 (Connolly, 2005) and the thermodynamic database of Holland and Powell (1998; with updates through 2004). The latter is useful in light of calculations by Goncalves et al. (2012) showing the strong dependence of the phengite component in white mica on bulk composition and the buffering assemblage. The calculated Ti-in-biotite temperature for average Bt1 compositions is 654 ± 33 °C (average of 29 analyses; Fig. 9), which agrees well with peak temperature estimates from Selverstone et al. (1999). Unfortunately, most Bt2 compositions have Ti contents below the range calibrated by Henry et al. (2005), but several yielded temperatures at the lowest end of the calibration range (480–800 °C), indicating a maximum temperature of alteration of 480 °C. The Si apfu values in white mica calculated from a bulk composition estimated from the QEMSCAN modes and measured mineral compositions are also shown in Figure 9. The Si isopleths have a slightly steeper slope than those from the Massonne and Schreyer (1987) calibration but a similar location in P-T space for the interval encompassing the highest measured Si contents for Ms2 in this study and by Selverstone et al. (1999, 3.3–3.4 apfu, p. 595). In summary, the new results (0.7–0.8 GPa at 480 °C) are consistent with those determined by Selverstone et al. (1999).

Monazite Geochronology

Analytical Method

Monazite identification was facilitated by Ce X-ray mapping of full thin sections and by QEMSCAN mapping. In situ Th/Pb geochronological analysis was conducted with a Cameca IMS 1270 secondary ion mass spectrometer (SIMS) at the W.M. Keck Foundation Center for Isotope Geochemistry at the University of California–Los Angeles. Thin sections were cut with a high-precision diamond saw or drilled with a diamond core, cleaned, mounted in epoxy with a prepolished block of standards, and gold-coated with 100 Å of Au. Moderate-(554) and high-Th (Trebilcock) monazite standards were used to obtain a curve of ThO2/Th versus Pb/Th for a relative sensitivity factor before analysis and were reanalyzed after every 5–6 unknowns during data collection to evaluate changes in instrument calibration. Standard operating procedures as outlined by Harrison et al. (1995) were followed, including use of an O primary beam focused to 10–20 μm, 15 eV offset for 232Th, and a mass resolving power of 4500. High mass resolving power is necessary to distinguish between Pb and Th isotope peaks and significant molecular interferences, particularly, the PrPO4 interference at mass 204 and NdPO4 interference at mass 208. The field aperture was adjusted to within 5 μm to ensure that the instrument was sampling the dominant Th signal from the target grain, even if the pit encompassed adjacent minerals, and to minimize sampling of adjacent Pb domains (Schmitt et al., 2010; Chamberlain et al., 2010; Ault et al., 2012). Instrument precision is limited to ~2% based upon the reproducibility of the calibration curve and is most often affected by bad polish, compositional variability within a standard, or environmental conditions such as 208Pb surface contamination (Harrison et al., 1995). Therefore, the data were corrected for common Pb using a 206Pb/204Pb value of 39.34 based on San Diego wastewater (Sañudo-Wilhelmy and Flegal, 1994).

Results

Twenty-six monazite grains were analyzed with one analysis per grain (with one exception: Cm1 was analyzed twice) due to grain size limitations.
General Evidence for Retrogression

Fluid-related alteration is widespread in sample RM-21, with the main features recorded by breakdown effects in feldspar, garnet, and allanite and the production of secondary assemblages characterized by $\text{Pl}_1$ (or $\text{Ab}$) + Ms$_p$ + Bt$_i$ + Cc + Mn$_z$ + Zo. Three compositionally distinct generations of plagioclase are recognized. The preserved Ca-rich plagioclase core ($\text{Pl}_1^c$) is interpreted as the stable composition during peak metamorphism. Lower-Ca plagioclase ($\text{Pl}_1^l$ and $\text{Ab}$) reflects discontinuous breakdown of $\text{Pl}_1$ during retrogression. Primary potassium feldspar is resorbed locally along grain boundaries where Ab is particularly common. Two generations of Ms are also observed, where Ms$_r$ represents part of the peak assemblage, and Ms$_s$ is associated with the alteration assemblages in feldspar and allanite reaction zones. These observations are consistent with a family of reactions commonly observed in metasomatized granitoids (e.g., Proyer, 2003), as well as this and other felsic xenoliths from the NVF (Broadhurst, 1986; Selverstone et al., 1999), and they reflect alteration primarily via the following generalized reaction:

$$ \text{Pl}_1 + \text{Kfs} + \text{Fluid} = \text{Pl}_2 + \text{Ab} + \text{Ms}_s + \text{Cc} + \text{Zo} $$

Garnet also locally exhibits textures and replacement assemblages indicative of fluid-related alteration. A minor increase in Mn at garnet margins and fractures in garnet that are locally lined with chlorite suggest late net-transfer reactions that consumed garnet at the expense of chlorite growth. However, the lack of other obviously secondary phases that are characteristic of adjacent feldspar domains make it unclear whether the development of these alteration effects was coeval.

Allanite and monazite typically have a close association in metamorphic rocks. In prograde metamorphism of metapelites, monazite is common at greenschist-facies conditions but reacts to allanite at uppermost greenschist- to amphibolite-facies conditions before reappearing again at higher grades (e.g., Wing et al., 2003; Krenn and Finger, 2007; Janots et al., 2008; Spear, 2010). Sample RM-21 has a granitic rather than metapelitic bulk composition, but nevertheless the monazite-allanite relations are consistent with these general trends. The distinct euhedral to subhedral morphology and concentric zoning patterns of large well-preserved allanite grains suggest that they are an original magmatic accessory phase but more than half (12) are between 107 ± 4 Ma and 57.7 ± 2.4 Ma. The remaining five analyses are from type 2 monazite grains (Mnz$_2$), which are texturally associated with altered $\text{Pl}_1$. Four of these are between 71 ± 8 Ma and 66 ± 1 Ma (Cm1-a, Cm1-b, Dm1, Dm2), and the fifth is 373 ± 35 Ma (Fm1). The majority of type 2 monazite grains were not viable for analysis because of either small size or low Th and Pb contents.
and replaced with metamorphic $\text{Al}_2 + \text{Mnz}_2$. This may have developed during peak metamorphism, which is estimated to have occurred at $-650$-$675$ °C (this study; Selverstone et al., 1999). Second, all allanite grains and both $\text{Al}_2$ and $\text{Al}_1$ are resorbed and replaced to varying degrees by the secondary assemblage of $\text{Ab} + \text{Ms} + \text{Rt} + \text{Cc} + \text{Qz} + \text{Mnz}_2$. The similarity of this assemblage to that of adjacent feldspar domains suggests retrograde alteration in a similar and possibly coeval fluid-rich environment. Similar retrograde monazite after allanite relationships have been observed by others (e.g., Krenn and Finger, 2007; Janots et al., 2008; Krenn et al., 2011). From previous work and additional thermobarometric constraints presented here, the pressure and temperature conditions of alteration of this xenolith are estimated to between 400 and 480 °C and between 0.6 and 0.9 GPa, with alteration pressures (at similar temperatures) for mafic granulite xenoliths in the NVF suite as high as 1.1 GPa (Selverstone et al., 1999).

The monazite chemistry in RM-21 is also notable for its retrograde signature and appears to at least partly have been controlled by proximity to preexisting phases in the immediate surroundings. Some monazite in RM-21 has elevated sulfur and strontium. Krenn et al. (2011) suggested that S-bearing monazite could be a sensitive monitor for S-rich fluids, although this seems unlikely given that not all secondary monazite in the sample is S-rich. An alternative local source could be preexisting sulfides, and altered pyrite was noted in proximity to several of these monazite grains. Similarly, plagioclase could have provided a local Sr source, as has been suggested by Krenn and Finger (2004) and Krenn et al. (2011) for other Sr-bearing monazite occurrences, although no distinct trend in Sr content was noted among plagioclase generations or albite in this sample (all have $-0.20$ to $-0.25$ wt% SrO). Microprobe investigations of other potentially local Sr sources like allanite or secondary calcite do not reveal Sr above trace-element levels. All type 1 monazite contains significant Th concentrations (up to $6 \text{ wt}\% \text{ ThO}_2$), whereas only relatively rare occurrences of type 2 monazite contain Th above electron microprobe detection limits. Since the allanite has significant Th, its recrystallization during $\text{Mnz}_2$ growth and consumption during $\text{Mnz}_2$ growth likely provided a local Th source for incorporation into the monazite. This type of local control on monazite chemistry has been noted in several previous studies (e.g., Mahan et al., 2006; Williams and Jercinovic, 2012).

**Age of Alteration from Monazite Geochronology**

Monazite is a particularly reactive radiogenic phosphate, which makes it a powerful chronometer for crustal events, including diagenesis, prograde metamorphism, magma crystallization, and fluid flow (e.g., Parrish, 1990; Williams et al., 2007; Janots et al., 2012; Villa and Williams, 2013). However, this same susceptibility to reactions warrants special care in studies where fluid-rock interactions have taken place because of the potential for earlier monazite to be partially or completely recrystallized or to incorporate common Pb (e.g., Harlov et al., 2011; Williams et al., 2011; Seydoux-Guillaume et al., 2012; Grand’Homme et al., 2016). In this study, all monazite occurrence is secondary, either in association with altered allanite (all $\text{Mnz}_2$, and some $\text{Mnz}_1$) or altered feldspar ($\text{Mnz}_2$). The association of $\text{Mnz}_2$ with the secondary hydrated assemblage in both textural settings suggests a genetic link between monazite crystallization and wholesale fluid alteration of the rock, which allows for geochronological constraints to be made on the timing of deep crustal hydration of the xenoliths. Type 1 monazite after allanite is the most viable for dating by in situ Th/Pb SIMS due to relatively high Th content. With four exceptions, all investigated type 2 monazite had insufficient Th signal to be dateable.

All five monazite grains with dates between 1871 and 1480 Ma are type 1 grains, three of which are $\text{Mnz}_1$ from the largest and most well-preserved allanite grain (e.g., Am4, Am6, Am7 in Fig. 7A). These grains are interpreted to reflect early recrystallization of igneous allanite to metamorphic $\text{Al}_2 + \text{Mnz}_2$, during metamorphism associated with the Yavapai and Mazatzal orogenies (Bowring and Karlstrom, 1990) and possibly subsequent ca. 1.4 Ga magmatic (Crowley et al., 2006) or Picuris orogenic (Daniel et al., 2013) activity. All of these events have previously been recognized in Sm-Nd (whole rock, garnet, and clinopyroxene) and U-Pb zircon studies of crustal xenoliths from the NVF, including xenoliths from Red Mesa (Wendlandt et al., 1993, 1996; Mattie et al., 1997; Condie et al., 1999; Crowley et al., 2006). However, the small number of analyses of this oldest suite of monazite coupled with the likelihood that it would have been the most susceptible to subsequent alteration-related disturbance, if any occurred, preclude further interpretations at this time.

A second small group of five dates scatters broadly between 850 and 174 Ma, whereas a large group of 16 dates spans fairly evenly between 107 and 58 Ma. Whereas the former range of dates does not readily correlate with known tectonic, magmatic, or fluid-fluxing events in the region, the latter range is largely coincident with the timing of shallow subduction beneath western North America (e.g., Conen and Reynolds, 1977; Lipman, 1980) and the Laramide orogeny. The younger range of dates also overlaps with U-Pb zircon dates from 85 to 33 Ma reported for hydrated eclogite (Usui et al., 2003; Smith et al., 2004) and mantle (Smith and Griffin, 2005) xenoliths from the NVF (Fig. 10). However, several sources of potential error need to be evaluated before further interpretation can be made.

Given the small grain sizes relative to the ion beam diameter, some of the scattered dates could represent artifacts from sampling non-monazite matrix. However, this source of error was minimized by reducing the field aperture to avoid sampling of ions from other host phases (Chamberlain et al., 2010; Schmitt et al., 2010) and by monitoring for systematic variations in PrPO$_2$ counts. Low PrPO$_2$ counts may reflect contact with a non-P- or non-Pr-bearing phase. Because no systematic trends were observed, it appears unlikely that sampling of non-monazite domains contributed significantly to the observed wide span of dates. Furthermore, all of this is a real source of error for some analyses, the implications are unlikely to significantly affect the dates, based on the assumption that the major source of common Pb is modern because $^{208}$Pb/$^{204}$Pb changes relatively slowly with time. For example, using an alternative correction employing a 2 Ga model (Stacey and Kramers, 1975) common Pb composition (e.g., if sampled from surrounding feldspar or allanite) changes the calculated dates by less than the stated analytical uncertainties even for analyses with intermediate (65%) radiogenic Pb contents (A. Schmitt, 2016, personal commun.). Instead, we suspect that the observed wide range of dates largely reflects sampling of mixed monazite domains of Paleoproterozoic or Mesoproterozoic and Late Cretaceous to Paleocene ages. This is easily envisioned when considering the commonly complex, matted appearance of $\text{Mnz}_2$ in type 1 settings (good examples are Bm3 and Bm5 in Fig. 7C).

As stated already, the largest group of dates (59% of the data set) spans from 107 to 58 Ma. These $\text{Mnz}_2$ dates are from grains in both types of textural settings, but they include four of the five analyses from type 2 grains (71.0 ± 9.4 to 66.2 ± 1.3 Ma), which are directly associated with the most abundant evidence for late-stage fluid-assisted alteration of the xenolith assemblage (Fig. 4). The range of dates could indicate that $\text{Mnz}_2$ growth, and thus hydrous alteration, occurred over a prolonged period from the Late Cretaceous to Paleocene rather than during a single punctuated event. However, there are several reasons that lead us to interpret that the main episode of $\text{Mnz}_2$ crystallization, and thus alteration, occurred at 70–65 Ma. First, the weighted mean of all 15 dates (not including replicate Cml-b) is 71.1 ± 6.0 Ma (2σ error). The mean square of weighted deviates (MSWD) for this calculation is 16.3, and while its relatively large magnitude supports the protracted monazite growth hypothesis, it may also simply mean that additional uncertainties are not adequately accounted for. As stated
earlier, the most substantial uncertainty that remains unaccounted for is sampling mixed domains of different ages, and the six oldest dates in this range are all from grains that have the matted texture indicative of multiple intergrown flakes. Exclusion of analyses from grains with this texture yields a weighted mean of 65.7 ± 3.4 Ma and MSWD = 2.51 (n = 8). Further exclusion of all analyses with <50% radiogenic 208Pb, which are exclusively restricted to the youngest end of the spectrum of dates, yields a weighted mean of 67.4 ± 2.1 Ma and MSWD = 0.76 (n = 5). Thus, the most reliable monazite dates point to a relatively punctuated event of hydrous alteration near the end of the Late Cretaceous.

A single monazite date of 22 ± 3.5 Ma is the youngest and is from a type 1 grain in allanite (Bm2 in Fig. 7C). This date coincides with the 33–19 Ma age range of diatreme emplacement, which is based on a variety of isotopic constraints from volcanic host rocks and xenoliths (Laughlin et al., 1986; McDowell et al., 1986; Wendlandt et al., 1996; Usui et al., 2003; Smith et al., 2004). This grain is also inside one of a subset of partial allanite pseudomorphs that additionally contain mats of clay. Thus, this date may reflect very late near-surface alteration (Mnz, ?).

Potential Fluid Sources for Late Cretaceous Hydration Beneath the Colorado Plateau

Potential sources of fluid that could have induced the observed crustal alteration include those related to local granitoid laccoliths or mid-Tertiary host diatreme emplacement (Ehrenberg and Griffin, 1979; Padovani et al., 1982; Broadhurst, 1986), Proterozoic tectonic or magmatic processes (e.g., Selverstone et al., 1999), and Late Cretaceous to mid-Paleogene subduction-related fluids from the Farallon slab (e.g., Broadhurst, 1986; Humphreys et al., 2003; Smith et al., 2004; Lee, 2005). Ehrenberg and Griffin (1979) observed secondary alteration of xenoliths hosted by the serpentinitized ultramafic microbreccias but not in those hosted by the minettes, and they suggested that the hydration effects therefore likely reflect local interaction with the host gas-fluid mixture just prior to eruption (see also Padovani et al. [1982] based on seismic velocity properties). Broadhurst (1986) similarly argued for what was interpreted as the second of two episodes of hydration recorded by crustal xenoliths as a reaction with the fluid component of the host microbreccia, but still before incorporation into the upper part of the diatremes, because conodonts from limestone xenoliths do not display evidence for maturation to temperature levels of even the lower-temperature “second” hydration (≤330 °C). The timing of alteration by this type of source fluid should obviously coincide with diatreme emplacement, which is constrained from a variety of studies to have occurred in the 33–19 Ma interval. While this study includes one 22 Ma date, and several other studies reported emplacement-related secondary zircon and monazite in eclogite xenoliths with dates between 35 and 28 Ma (Usui et al., 2003; Smith et al., 2004; Schulze et al., 2015), all other dates from this study (96%) point to hydration well before emplacement of the host volcanics and regional lachlactic intrusions like those in the Abajo, Henry, and La Sal Mountains (30–20 Ma; Nelson et al., 1992). Hydrous alteration features have now also been recognized in the microbreccia- and minette-hosted xenoliths (Selverstone et al., 1999), removing that basis of argument for very localized fluids.

The Yavapai-Mazatzal Province boundary is thought by many to pass through the Colorado Plateau, and Selverstone et al. (1999) suggested that hydration of the xenoliths might be sourced from Paleoproterozoic subduction, based upon systematic differences between the degree of alteration observed in xenoliths sourced from diatremes northwest and southeast of their proposed trace of this tectonic boundary (Fig. 1). The recognition of Paleoproterozoic Mnz in the Red Mesa xenolith from this study, notably associated with recrystallization of igneous allanite, might support this argument. However, several mafic xenoliths that contain the characteristic high-pressure alteration features, including three of five studied by Crowley et al. (2006) from the Red Mesa diatreme, have 1.43 Ga zircon U-Pb crystallization ages, indicating that significant alteration must have occurred at or after 1.4 Ga. This is consistent with Mnz, in RM-21 being a relatively minor constituent of the overall age population and therefore its restriction to relatively early textural settings, albeit secondary with respect to the igneous allanite.

We interpret monazite data from sample RM-21 to indicate that the bulk of the observed hydrous alteration occurred at 70–65 Ma. This age approximately coincides with emplacement of the Carrizo Mountains diorite laccolith (74–70 Ma; Semken and McIntosh, 1997), in northwestern Arizona (Fig. 1), which is the southwesternmost intrusive complex in the Colorado Mineral Belt trend, and which could conceivably have been the source of fluids inducing the alteration. However, this seems unlikely given the small volume and shallow level of emplacement of the intrusion (into Permian to Cretaceous sedimentary rocks) relative to the mid-crustal depths of xenolith alteration, and also considering that mineralization associated with the Late Cretaceous components of the Colorado Mineral Belt was apparently extremely limited (Cunningham et al., 1994). A more likely fluid source emerges when considering that the RM-21 xenolith is one of a large suite of crustal and mantle xenoliths in the NVP for which similar alteration has been described, and that the age of alteration is also coincident with the timing of shallow subduction beneath western North America (e.g., Coney and Reynolds, 1977; Lipman, 1980).

Hydration at mantle depths beneath the Colorado Plateau during the Late Cretaceous to Eocene is well documented from previous xenolith studies, but to our knowledge, the data presented here are the first to constrain crustal hydration to the same time interval. The earliest studies invoking an important role for Laramide-aged fluids beneath the Colorado Plateau focused on eclogite xenoliths, where Helmstaedt and Doig (1975) argued for them as direct samples from the shallowly subducted Farallon slab (see also Usui et al., 2003). Several geochemical and petrological xenolith studies also documented significant hydration of mantle lithologies from the Colorado Plateau, many of which either inferred or constrained a Laramide age for these processes (e.g., Smith, 1995; Smith and Griffin, 2005; Lee, 2005; Li et al., 2008).

While subsequent chemical and isotopic studies have shown that most of the eclogite xenoliths more likely originated from Proterozoic subduction processes (Wendlandt et al., 1993; Smith et al., 2004), most recent studies have concluded that significant recrystallization in the presence of a hydrous fluid occurred during the Laramide. Concordant U-Pb zircon dates, interpreted to record fluid-induced recrystallization, are reported from eclogite xenoliths in the ranges of 70–35 Ma (Smith et al., 2004) and 81–47 Ma (Usui et al., 2003), and from an ultramafic mantle garnetite xenolith at 85–60 Ma (Smith and Griffin, 2005). The temporal overlap between zircon recrystallization in the mantle and monazite dates presented here (Fig. 10) suggests that hydration extended from mantle depths into the lower crust during the period of low-angle subduction of the Farallon slab, as also hypothesized by Broadhurst (1986) and Smith et al. (2004).

Potential Implications of a De-Densified Lower Crust

Hydrous fluids interacting with deep continental lithosphere can have profound influences on rheological and other physical properties, and magmatic patterns and compositions, as well as the dynamic evolution and ultimate survival of continental lithosphere. Those effects receiving the most attention with regard to Laramide-aged hydration of lithospheric mantle in the western United States have centered on the potential for increased buoyancy to drive Cenozoic surface uplift and on how the process may...
have fertilized the mantle for subsequent large-volume ignimbrite eruptions once the shallow slab was removed (e.g., Humphreys et al., 2003).

The full impact of these effects relies on the spatial extent of the hydration process. Whereas this study focuses on only one xenolith locality, and some have suggested that fluid alteration in the mantle beneath the Colorado Plateau was confined to serpentinized fracture zones (Smith et al., 2004), other considerations suggest that deep alteration was both vertically and laterally extensive across much of the broader region. First, this and other studies of xenoliths from the NVF suggest that at least the lower 20 km of the crust and mantle to 140 km experienced hydrous alteration. Sample RM-21 is one of the shallower crustal xenoliths (alteration depth at ~27 km from 0.75 GPa pressure calculation using the density profile of Christensen and Mooney, 1995), and others record pressures as high as 1.2 GPa (Selverstone et al., 1999), corresponding to ~43 km depth, which is near current estimates of crustal thickness (~45 km; Fig. 2B; Bashir et al., 2011; Gilbert, 2012; Shen et al., 2013). Xenolith petrology also requires that hydration of the mantle lithosphere occurred to depths as shallow as 70 km, and the chemistry of the serpentinized ultramafic breccias requires a hydrated eruptive source, which means that hydration occurred to depths greater than the ~140 km recorded by the deepest xenoliths (e.g., Smith, 1995; Smith et al., 2004; Smith and Griffin, 2005; Li et al., 2008).

Second, xenolith studies along transects across the western United States show predicted changes in metasomatizing fluids that are consistent with progressive west-to-east dehydration of a slab with increasing distance from the trench, and they link mantle metasomatism in the Sierran arc and Colorado Plateau to the same Farallon subduction system (Lee, 2005; Li et al., 2008). Similarly, mantle xenoliths in the northern Rocky Mountains of Montana show evidence for extensive metasomatism, some of which is thought to have occurred during the Laramide orogeny (Carlson et al., 2004; Downes et al., 2004; Facer et al., 2009). These data, supporting widespread mantle hydration, are also consistent with numerical modeling studies that predict hydrous minerals in a subducting slab to persist to well over 1000 km inboard from the trench (e.g., Li et al., 2008; Currie and Beaumont, 2011).

The conversion of olivine to serpentine during hydration reduces the density of peridotite (e.g., Hess, 1954), and this process has been proposed to have provided a regional source of buoyancy to drive surface uplift of large tracts of the Rocky Mountain region during the Laramide orogeny (Humphreys et al., 2003; see also Schulze et al., 2015). Jones et al. (2015) made a similar argument for crustal hydration as a potential buoyancy source, where hydration reactions consume denser crustal minerals like garnet to produce lower-density secondary phases such as mica and amphibole. That study emphasized that the lower-crustal seismic velocity structure in the Rocky Mountains, from both active and passive sources, and shear wave velocity gradients in the uppermost mantle are also consistent with regional and progressive northeastward hydration of crust. From variations in densities and density–velocity relationships in Montana and Wyoming crustal xenoliths, this mechanism was shown to be capable of producing more than the observed elevation differences between northern Montana and southern Wyoming. Similarly, three-dimensional (3-D) density models derived from regional seismic, heat-flow, and gravity data are consistent with up to 900 m of surface uplift relative to the eastern Great Plains due to variations in crustal density (Levandowski et al., 2014; W. Levandowski, 2016, personal commun.).

The xenolith sample studied here records these types of hydration reactions as having occurred during the Laramide orogeny. While this particular xenolith is one of the least altered of the suite, and its calculated density is not dramatically different before and after alteration (less than 0.01 g/cm³; Fig. 2C), many of the deeper and more mafic xenoliths record more extensive hydration reactions, with textures indicating the loss of garnet and pyroxene as well as plagioclase (Broadhurst, 1986; Selverstone et al., 1999; Crowley et al., 2006; Butcher, 2013). For example, preliminary estimates from another sample from the Selverstone et al. (1999) suite, an extensively altered garnet amphibolite from Garnet Ridge, suggests that pre- and postalteration density differences may be as high as 0.07 g/cm³. Depending on the thickness of the altered layer, these values for density change would predict isostatic surface uplift on the order of several hundred meters. A more comprehensive evaluation of a broader suite of crustal xenoliths from the Colorado Plateau awaits a future study. The densification mechanism for surface uplift is particularly attractive because it provides an explanation for uplift of relatively undeformed regions of the greater Rocky Mountains such as the Colorado Plateau and High Plains.

CONCLUSIONS

Xenoliths from the Navajo volcanic field in the central Colorado Plateau commonly exhibit distinct textures and secondary mineralogical assemblages reflecting high-pressure, low-temperature hydration of the deep crust and lithospheric mantle sometime prior to the 30–20 Ma eruption of the host volcanics. In line with several past studies of mantle xenoliths from the field, a crustal xenolith from the Red Mesa diatreme reveals a period of monazite crystallization at 70–65 Ma. We interpret this Late Cretaceous monazite growth to reflect late alteration of the high-temperature Proterozoic metamorphic assemblage with hydrous fluid introduced by the shallowly subducting Farallon slab. Similar to previous suggestions involving mantle hydration, the growth of low-density hydrous phases at the expense of high-density, anhydrous minerals, which are abundant in unaltered Proterozoic crust, could have contributed to elevated topography in the Colorado Plateau. If crustal hydration was sufficiently widespread, this crustal densification mechanism could have contributed to regional uplift across the Rocky Mountain region and High Plains.

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