GSA ANNUAL FIELD TRIP #X:
TECTONIC DEVELOPMENT OF THE COLORADO PLATEAU TRANSITION ZONE, CENTRAL ARIZONA: INSIGHTS FROM LOWER LITHOSPHERE XENOLITHS AND VOLCANIC HOST ROCKS

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ABSTRACT
A growing body of evidence suggests that continental arc lower crust and underlying mantle wedge assemblages native to the Mojave Desert (i.e., the southern California batholith; SCB) were displaced eastward during Laramide shallow-angle subduction and reattached to the base of the Colorado Plateau transition zone (central Arizona) and further inboard. We highlight on this field trip two xenolith localities from the transition zone (Camp Creek and Chino Valley) that likely contain remnants of the missing Mojave lithosphere. At these localities, nodules of garnet clinopyroxenite, the dominant xenolith type at both studied localities, yield low jadeite components in clinopyroxene, chemically homogeneous "type-B" garnet, and peak conditions of equilibration from 600-900 ˚C and 12-28 kbar. These relations strongly suggest a continental arc residue ("arclogite"), rather than a lower plate subduction ("eclogite"), origin. Zircon grains extracted from these nodules yield a bimodal age distribution with peaks at ca. 75 and 150 Ma, overlapping SCB pluton ages and suggesting a consanguineous relationship. In contrast, Mesozoic and early Cenozoic igneous rocks native to SW Arizona, with age peaks at ca. 60 and 170 Ma, do not provide as close a match. In light of these results, we suggest that transition zone xenoliths: 1) began forming in Late Jurassic time as a mafic keel to continental arc magmas emplaced into the Mojave Desert and associated with eastward subduction of the Farallon plate; 2) experienced a second ca. 80-70 Ma pulse of growth associated with increased magmatism in the SCB; 3) were transported ~500 km eastward along the leading edge of the shallowly subducting Farallon plate; and 4) were reaffixed to the base of the crust at the new location, in south central Arizona. Cenozoic zircon U-Pb, garnet-whole rock Sm-Nd, and titanite U-Pb ages suggest that displaced arclogite remained at elevated temperature (>700 ˚C) for 10s of Myr following its dispersal and until late Oligocene entrainment in host latite. The lack of arclogite and abundance of spinel peridotite xenoliths in Miocene and younger mafic volcanic host rocks, such as those examined at the San Carlos xenolith locality, and the presence of seismically fast and vertically dipping features beneath the western Colorado Plateau, suggest that arclogite has been foundering into the mantle and being replaced by upwelling asthenosphere since Miocene time.

Keywords: Colorado Plateau-Basin and Range transition zone, shallow-angle subduction, lower lithosphere xenoliths, Laramide orogeny, tectonic bulldozing, lithospheric delamination.

1. INTRODUCTION

The purpose of this trip is to visit lower crust and upper mantle (i.e. lower lithosphere) xenolith-bearing volcanic centers within the Colorado Plateau transition zone (central Arizona), as these xenoliths and host lavas provide key constraints on the Mesozoic to recent petrologic and tectonic evolution of the SW North American Cordillera. Specifically, the trip will focus on: 1) petrologic records of Late Cretaceous-early Cenozoic lower lithosphere tectonic mobility associated with the Laramide shallow-angle subduction and 2) post-Miocene delamination and associated uplift of the Colorado Plateau. During the trip, we will address a number of questions pertinent to the tectonic development of the Colorado Plateau transition zone and adjacent geologic provinces and how analogous continental margins affected by modern (e.g., Alaska, Peru, Mexico, Chile) shallow-angle subduction may evolve. For example, the upper crust of the SW Cordillera developed high-angle reverse faults and basement-cored uplifts in response to shallow-angle subduction (e.g., Burchfiel et al., 1992; Blakey and Ranney, 2018), but how was the lower lithosphere affected by this event? How does the lithosphere "heal" from the tectonic trauma of shallow subduction (i.e., how is the subcontinental mantle lithosphere reconstructed after tectonic removal)? Finally, in what ways may shallow subduction precondition the affected region to later increments of deformation (e.g., uplift of the Colorado Plateau, extension in the Basin and Range, and Cenozoic delamination)?

2. GEOLOGIC BACKGROUND

The SW North American Cordillera

The Late Cretaceous-early Paleogene Laramide orogeny was a regional compressional event that deformed the SW North American craton edge (e.g. Saleeby, 2003; DeCelles, 2004; Dickinson et al., 2009). A commonly cited mechanism for the orogeny is intensified traction and tectonic erosion of the lowermost crust and upper sub-continental mantle lithosphere (LC-SCML) due to flattening of a ~500 km-wide segment of the subducting Farallon plate (Livaccari et al., 1981; Bird, 1988; Saleeby, 2003). Parts of the central Andean orogen are regarded as the best modern analogue, where shallow slab segments coincide with colliding aseismic ridges and oceanic plateaux (e.g. Allmendinger et al., 1997; Garzione et al., 2017). Plate reconstructions for the Pacific-Farallon ridge led to the interpretation that the Laramide orogeny resulted from the subduction of conjugate massifs to the Hess and Shatsky oceanic plateaux (Livaccari et al., 1981; Liu et al., 2010). Furthermore, an ~500 km-wide Laramide deformation corridor parallels the subduction trajectory of inferred Hess and Shatsky conjugates, which were embedded in the Farallon plate as they subducted in Laramide time (Saleeby, 2003; Liu et al., 2010). The plate edge damage zone consists of the southern California batholith (SCB) of the Mojave Desert and the southernmost Sierra Nevada batholith (SNB; Fig. 1, inset).

As emphasized below, the impact of oceanic plateaux is consistent with evidence for widespread ductile thrusting within the SCB, deep crustal exhumation, tectonic underplating of trench sediments, and – the focus of this research – removal of the LC-SCML (e.g. Saleeby, 2003; Luffi et al., 2009; Chapman et al., 2012; Chapman, 2017; Ducea and Chapman, 2018).

Overview of the SCB domain of the Laramide corridor

The formerly contiguous SNB-SCB-Peninsular Ranges batholithic belt was an >2,000 km-long NNW-trending granitic arc emplaced largely during three Mesozoic magmatic "flare-up" events at ca. 230-210 Ma, ca. 160-150 Ma, and ca. 100-85 Ma (e.g., Ducea, 2001; DeCelles et al., 2009; Paterson and Ducea, 2015). In contrast to the SNB to the north and the Peninsular Ranges batholith to the south (Fig. 1), much of the ~500 km-long SCB is rootless, lying tectonically above underplated subduction accretion assemblages (the Rand and related schists in the Mojave region) that were transported inboard by shallow-angle subduction (Jacobson et al., 1988; Grove et al., 2003; Porter et al., 2011; Chapman, 2017; Ducea and Chapman, 2018). These schists are exposed in the footwall of the shallowly dipping and regionally extensive Rand fault, interpreted as a remobilized subduction megathrust (e.g., Cheadle et al., 1986; Chapman, 2017), beneath deep crustal level SCB assemblages and the southern SNB (Fig. 1).

The southern end of the SNB is tilted into a southward deepening section spanning paleodepths of ~10 to 35 km (e.g., Nadin and Saleeby, 2008). Structural and petrologic relations in the SCB and southernmost SNB indicate that the base of the batholith and underlying LC-SCML was sheared off at 30-35 km depth and replaced with trench sediments and fragments of Farallon lithosphere (Grove et al., 2003; Saleeby, 2003; Chapman, 2017).

What was the fate of the sub-SCB LC-SCML? Do remnants of the displaced material exist, and if so, what is the relationship between LC-SCML remnants and underplated schist? Some answers to these questions can be found in remote imaging studies applied to the Laramide plate edge domain. Surface exposures of schist are linked directly to deeper crustal structure by seismic data and receiver function analysis showing the schist dipping north beneath the southernmost SNB and defining a regional flat fabric with NE-SW seismic anisotropy at mid-crustal depths beneath thin (~30 km) Mojave crust (Cheadle et al., 1986; Porter et al., 2011). Additional constraints from geochemical data reveal an ~N-S trending boundary at ~116˚W, west of which lacks a lithospheric isotopic fingerprint and suggests the presence of underplated schists (Sass et al., 1994; Miller et al., 1996; Chapman et al., 2013). Collectively, these studies resolve a major compositional boundary within the mid-to deep crust of the central Mojave demarcating a western schist-bearing domain, and an eastern domain lacking significant schist and containing remnants of ancient LC-SCML (Fig. 1).

Laramide Imprints in Xenoliths

Volcanic hosted xenolith suites of the SW Cordillera record development of the LC-SCML prior to, during, and following the Laramide event. Proterozoic peridotitic upper mantle and granulitic lower crustal xenoliths in the Colorado Plateau and vicinity, in conjunction with Nd isotopic data on mafic volcanic rocks of the region, record local preservation of LC-SCML beneath the region through Laramide time (Livaccari and Perry, 1993).

Xenolith suites from the eastern and central Mojave region also provide evidence for underlying remnants of ancient LC-SCML. First, spinel peridotites from the Pliocene-Quaternary Cima cones yield Re-Os model ages of 1.8–3.4 Ga (Fig. 1; Lee et al., 2001), overlapping ca. 2.0 Ga Sm-Nd model ages on nearby Precambrian basement rocks (Bennett and DePaolo, 1987). Second, a subordinate group of xenoliths from the Quaternary Dish Hill cone consist of shallow-level peridotites that equilibrated at ~850ºC, and yield εNd = -6.4 to -13.0 (Fig. 1; Luffi et al., 2009). These data indicate that ancient LC-SCML was not entirely sheared off from beneath the eastern to central Mojave region by Laramide flat-slab subduction.

Remnants of pre- to syn-Laramide mantle lithosphere that constituted the mantle wedge for the SNB are present in late Miocene xenolith suites from the central SNB (Domenick et al., 1983; Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1996, 1998, Ducea, 2001; Chin et al., 2012). Xenolith constraints on central SNB "lithostratigraphy" indicate that this fossilized mantle wedge extended to ~125 km depth and cooled rapidly following the Late Cretaceous (Laramide) termination of SNB magmatism (e.g., Ducea and Saleeby, 1998; Saleeby et al., 2003; Chin et al., 2012). Peridotites and garnet websterite dominate the base of the section and grade upward into a ~45 km-thick zone of garnet clinopyroxenite followed by garnet granulite at ~40 km paleodepth. Trace element data and Sm-Nd isochron ages indicate that the garnet clinopyroxenites are partial melt residues, or deep level cumulates, linked to the overlying SNB (Ducea and Saleeby, 1998; Ducea, 2001). These garnet clinopyroxenites are commonly referred to as eclogites, but contrast with classic eclogites by having more Ca- and Mg-rich clinopyroxene, more Fe- and Ca-rich garnet ("Type-B" garnet; Coleman et al., 1965), and commonly contain accessory hornblende. These arc root cumulates are commonly referred to as "arclogites" (Anderson, 2005).

The preservation of ~125 km worth of central SNB lower lithosphere (arc root lower crust and upper mantle) through Miocene time contrasts sharply with the virtual absence of these materials beneath the SCB (Fig. 1). Geochemical proxies for crustal thickness (e.g., Sr/Y and La/Yb) strongly suggest that a deep sub-SCB root indeed existed prior to shallow-angle subduction, forming in the Late Jurassic and thickening significantly during the Late Cretaceous magmatic flare-up (Howard et al., 2016). We now focus on the fate of missing SCB lower lithosphere and the regional extent of Farallon plate mantle lithosphere underplating beneath the "unrooted" SCB.

SUB-SCB LOWER LITHOSPHERE DISPLACEMENT AND RECONSTRUCTION

It is important to reiterate here that despite LC-SCML removal from beneath the SCB, and the subsequent tectonic underplating of schist, the base of the lithosphere in the Mojave region currently exceeds ~60 km depth (Li et al., 2007; Luffi et al., 2009). This profound relationship indicates that latest Cretaceous-Cenozoic reconstruction of the mantle lithosphere beneath the schists must have taken place.

Relationships resolved in Dish Hill and Crystal Knob xenolith suites (Fig. 1) indicate that the underlying mantle lithosphere has been reconstructed by tectonic underplating of Farallon plate sub-oceanic peridotites between 80 and 30 Ma (Atwater and Stock, 1998; Luffi et al., 2009; Liu et al., 2010; Quinn et al., 2018). Considering that eclogitic fragments of the Farallon Plate (Usui et al., 2003) plus significant amounts of ancient LC-SCML both underlie the Laramide interior, the lower lithosphere beneath a significant part of the Laramide corridor must be a composite of these assemblages. As shown on the SCB section of Figure 1, the underplated schists plus underlying mantle lithosphere constitute a lithosphere-scale accretionary complex lying beneath a carapace of SCB granitoids, that was stripped of most of its underlying mantle wedge.

As to the fate of the displaced SCB LC-SCML, we identify here several latest Oligocene lower lithosphere xenolith locations from the Colorado Plateau transition zone that likely contain remnants of the missing lithosphere that was displaced eastward and reaffixed beneath the Colorado Plateau transition zone.

This hypothesis predicts that native sub-Mojave arc materials and displaced equivalents should contain similar arrays of rock types. This is indeed the case, as ca. 25 Ma latite at Camp Creek and Chino Valley localities each contain abundant nodules of, in order of decreasing abundance, garnet-pyroxene rocks, garnet granulite, peridotite, and quartzofeldspathic gneiss (Schulze and Helmstaedt, 1979; Arculus and Smith, 1979; Esperanca, 1984; Smith et al., 1994; Esperanca et al., 1988, 1997; Erdman et al., 2016). As with the sub-SNB suite, garnet clinopyroxenite xenoliths from Chino Valley and Camp Creek are arclogitic in composition, and equilibrated between 600-900 ˚C and 12-28 kbar (45-100 km depth assuming a 2800 kg/m3 overburden density; Smith et al., 1994; Esperanca et al., 1988; Erdman et al., 2016). Furthermore, Mesozoic plutons of the Mojave Desert and arclogite recovered from Chino Valley and Camp Creek all share similar isotope systematics, with $^{87}Sr/^{86}Sr$ and εNd values ranging from 0.706 to 0.711 and -2 to -10, respectively (Esperanca et al., 1988; Smith et al., 1994; Miller et al., 1996). These relations point to a thick sub-SCB residue, rather than a lower plate, origin.

This hypothesis also predicts that materials once attached to the base of the SCB should retain the high-temperature portion of the thermal history of the Mojave Desert. For example, the Mojave Desert is underlain chiefly by Middle Jurassic-Early Cretaceous (ca. 160-140 Ma) and Late Cretaceous (90-70 Ma) arc plutonic assemblages with relatively small amounts of Mesozoic to Neoproterozoic metasedimentary rocks (Wells and Hoisch, 2008; Barth et al., 2008; Needy et al., 2009; Chapman et al., 2018). Hence, if xenoliths recovered from Chino Valley and Camp Creek localities are indeed consanguineous with the SCB, the xenoliths should yield chiefly Late Cretaceous and Jurassic ages. If instead the xenoliths are native to the Colorado Plateau transition zone, which exposes mainly 1.6-1.8 Ga Yavapai-Mazatzal basement overlain by Proterozoic to Mesozoic strata, much older ages are expected. It should be noted that ca. 70-55 Ma porphyry copper deposits and a ca. 190-160 Ma magmatic arc crop out a few 10s of km SW of the studied xenolith localities (Fig. DR1; e.g., Vickre et al., 2014; Tosdal and Wooden, 2015; Chapman et al., 2018). Hence, Latest Cretaceous-early Cenozoic and Early-Middle Jurassic xenolith zircon ages may also point to a SW Arizona origin.

We highlight new data in the sections that follow, in the context of planned field trip stops, that bear directly on the hypotheses outlined above. Detailed zircon U-Pb, Hf, and trace element data plus garnet Sm-Nd geochronology are presented and discussed in two companion papers (Chapman et al., in review; Rautela et al., in review).

3. ROAD LOG

We depart from the Phoenix Convention Center (100 N 3rd St, Phoenix, AZ 85004), and drive ~55 miles to the Sunset Point Rest Stop at Interstate 17 mile marker 252 in Black Canyon City, Arizona, where we will reset odometers to zero, and proceed north to stop 1 (Fig. 2). En route to the rest stop we will climb ~2,300 feet as we enter the Colorado Plateau Transition Zone from the Basin and Range province. As we depart the Phoenix metropolitan area to the north, we will cross the Central Arizona Project canal, which diverts water from the Colorado River to central and southern Arizona.

In the vicinity of Phoenix, note exposures of Proterozoic gneiss, schist, and granitic rocks, locally capped by late Cenozoic lava flows (e.g., roadcut beginning ~0.5 mi north of milepost 247, where pillow basalt rests above purple-brown baked basement).

In the area surrounding New River, note lightly colored and well-stratified volcaniclastic lacustrine and fluvial strata of the Upper Oligocene to Lower Miocene oreodont-bearing Chalk Canyon formation (Lander and Lindsay, 2011). The Chalk Canyon formation (informal name) and stratigraphically higher deposits (e.g., basaltic lava flows and tuffs of the Middle Miocene Hickey Formation) crop out in the Agua Fria Paleobasin, an ~N-S-trending and ~75 km-long paleotopographic low of uncertain origin (Leighty, 2007).

Paleozoic and Mesozoic strata, which blanket the Colorado Plateau to the north but are conspicuously absent here, were most likely stripped from underlying Proterozoic basement following early Mesozoic and later tectonism (e.g., McKee, 1951; Reynolds et al., 1989). Note change in landscape from desert scrub-covered mountains to grassy lava plateau ~2 mi south of Sunset Point Rest Stop.

Universal Transverse Mercator (UTM) zone 12S coordinates (WGS 84 datum) are given for each field trip stop. Road log distances are reported in miles.

Cumulative mileage	*Description*
0.0	Sunset Point Rest Stop – western edge of Black Mesa. Note views of Black Canyon and the Bradshaw Mountains, underlain by Proterozoic basement rocks, to the west. The steep cliff along the western margin of Black Mesa is the headwall scarp of a large (~7x1 km) landslide complex, destabilized by erosion of Bumble Bee Creek at the bottom of Black Canyon. Note NNE-trending low ridges of down-dropped and back-rotated (east-dipping) former mesa surface material. To the east, note Joe's Hill, the volcanic center inferred to have produced ca. 10.8 Ma basalt of the Hickey Formation which forms an ~30 m-thick cap to Black Mesa (Leighty, 1997). Formerly contiguous Black (west) and Perry (east) mesas are bisected by an ~1000' deep canyon carved by the Agua Fria River. The timing and driver(s) of deep incision are unknown. Return to vehicles and proceed north on I-17.
10.2	Merge onto I-17 N and continue for 8.9 miles. At milepost 256, I-17 descends from Black Mesa and continues on highly weathered Proterozoic basement. Take exit 262 for AZ-69 N toward Cordes Lake Rd./Prescott. Note Hickey Formation basalt atop loosely consolidated lacustrine and fluvial deposits presumably correlative to the Chalk Canyon formation.
31.7	Keep left to continue on AZ-69 N for 21.3 miles. Note excellent strike-perpendicular roadcuts of the Paleoproterozoic Mayer Group in the vicinity of Mayer (mileposts 268-277), consisting chiefly of tholeiitic, greenschist to amphibolite facies volcanic rocks including metamorphosed basaltic flows, tuff, rhyolite, iron formation, and tuffaceous sedimentary rocks (Anderson, 1989). These assemblages are thought to represent a ca. 1.80-1.74 Ga oceanic island arc formed in the ocean basin that intervened between fragments of the previously rifted Archean Wyoming craton. At Poland Junction, AZ-69 N curves northeastward and

strike-parallel roadcuts of metamorphosed andesite are visible. Turn right onto Fain Rd.

38.9 Continue north on Fain Rd. for 7.2 miles. The Black Hills of Yavapai County, the first normal fault-bounded horst block west of the Mogollon Rim on the SW edge of the Colorado Plateau, gain prominence and are visible behind low hills of locally-derived fanglomerate in the ~4.5 miles before Fain Rd veers west. Merge onto AZ-89A S.

45.2 Continue west on AZ-89A S for 6.3 miles. After 3.6 miles, the view to the south features the ca. 13.8 Ma Glassford Hill volcanic center and derivative basaltic flows and dikes of the Hickey Formation (low smooth hill; McKee and Anderson, 1971), which invade and lie above highly fractured ca. 1.4 Ga alkali granite of the Dells pluton (more jagged low peaks; Silver et al., 1984). Take exit 317 for AZ-89 toward Prescott/Chino Valley.

55.4 After 0.4 mi, keep right at the fork and merge onto AZ-89 N. Continue on AZ-89 N through three traffic circles for 9.8 mi. Enter Chino Valley and at the fourth traffic circle turn right (east) on Perkinsville Rd. Reset odometers to zero.

8.5 Continue east on Perkinsville Rd for 7.8 miles. Pavement ends 3.2 miles east of the Perkinsville Rd-AZ-89 N intersection and volcanic domes of the Sullivan Buttes latite come into view at 11 o'clock. On clear days the San Francisco Peaks, a Late Miocene to Holocene volcanic field on the northern flank of Flagstaff, are visible at 9:30. The Black Hills of Yavapai County are observed at 12 o'clock. At 6.5 and 7.5 miles, respectively, we will pass domes of pyroxene-biotite latite and low hills underlain by Devonian Martin limestone and capped by Mississippian Redwall limestone. After passing low hills of Paleozoic limestone and immediately before bridge (7.8 miles), turn left (north) on Forest Service Rd 641, labeled "Suncrest Ranch Rd" at the time of this trip. Continue for 0.7 miles, noting vertical joints in latite dome to the left and latite exposed in the ditch on the right, until the road splits. Turn the vehicles around and park. Prepare to examine exposures adjacent to dirt road and roadside xenolith "lag." We will spend at least one hour away from the vehicles. At the time of this trip, a permit is required to visit this State Trust Land and can be requested from the Arizona State Land Department.

Stop 1. Chino Valley "Lower Domes" xenolith locality (UTM 12S, 378732E, 3850027N)

Latite and trachyite in the Sullivan Buttes and correlative Wells Ranch-Boundary Tank and Lower Domes exposures (~15 km NW, 10 km north, and ~10 km east of Chino Valley, Arizona respectively; Figs. 2 and 3; Heger, 2013) were first mapped and dated by Krieger (1965) and Krieger et al. (1971), who recognized multiple flows ranging in K-Ar age from ca. 27 to 23 Ma. Each exposure contains 5-10% biotite, clinopyroxene, apatite, magnetite, and locally amphibole, all in a very fine groundmass of devitrified glass and/or pyroxene and plagioclase microlites (Krieger 1965; Tyner and Smith, 1986). This stop focuses on the Lower Domes, as ca. 23.5 Ma (Ar-Ar age; Heger, 2013) biotite trachyite at this location contains the highest proportion of lower crustal-upper mantle xenoliths (Arculus and Smith, 1979; Stefanov, 1993; Smith et al., 1994; Cunningham, 2001).

Xenoliths hosted in Lower Domes biotite trachyite are subrounded to rounded, large (generally 1-30 cm), and consist of (in order of diminishing abundance) garnet-clinopyroxene-amphibole (i.e., arclogite) assemblages, quartzofeldspathic granulite, and websterite (Smith et al., 1994; Cunningham, 2001; Heger, 2013). As observed at the Camp Creek locality (stop 2), xenoliths weather out of host trachyite and concentrate as lag deposits, most notably at topographic lows between domes, especially along Forest Service Road 641/Suncrest Ranch Road (Fig. 4). In comparison with Camp Creek xenoliths, Chino Valley nodules are less invaded by latite host lava.

Chino Valley arclogite xenoliths are moderately to strongly foliated, containing little-kelyphitized garnet (kelyphite is restricted to <100 µm thick bands along rims), clinopyroxene, and subordinate (less than 10% by mode) texturally late (overprinting clinopyroxene) pargasitic amphibole. Apatite and rutile are ubiquitous, though volumetrically minor, phases with ilmenite also present in several samples. Titanite is less common than rutile and is observed in rutile-bearing samples along rutile rims and in rutile-absent samples as primary grains. Zircon, kyanite, and allanite are present in <1% of studied arclogite nodules. Studied xenoliths generally show little textural evidence of interactions with host latite, with <10% of nodules containing pockets of glass plus quench phenocrysts of plagioclase, amphibole, pyroxene, corundum, and hercynitic spinel. Representative thin section photomicrographs are shown in Figure 5.

Constraints on the pressure and temperature history of Chino Valley and Camp Creek arclogite xenoliths are imprecise as no robust geobarometers exist for bimodal garnet-clinopyroxene assemblages. Thermobarometry of websterite (i.e., mosaic orthopyroxene-bearing assemblages) and plagioclase-bearing arclogite bracket paleopressures of plagioclase- and orthopyroxene-absent arclogite (which presumably resided between websterite and plagioclase-bearing arclogite) to 9-28 kbar (33-100 km depth assuming a 2800 kg/m^3 overburden density; Smith et al., 1994; Esperanca et al., 1988; Erdman et al., 2016). Garnet-clinopyroxene, Ti-in-zircon, and Zr-in-rutile thermometry yield equilibration temperatures of 600-900 ˚C, calculated at 9-28 kbar. Construction of Chino Valley arclogite-specific pressure-temperature pseudosections do not constrain the conditions of equilibration further, as the field garnet-clinopyroxene-rutile-ilmenite extends over the range previously determined via thermobarometry. However, garnet zonation patterns, notably decreasing Mg# (Mg/(Mg+Fe)) from core to rim, are only possible via near-isobaric cooling of ~300 ˚C.

Geochemically, arclogite whole rock powders yield elevated Ni and Cr contents, lack significant Eu anomalies, and are not significantly enriched in HREE (Esperanca, 1984; Smith et al., 1994). Respectively, these relations suggest ultimate derivation from basaltic melts, that plagioclase was not an important fractionating phase, and that garnet growth in these rocks occurred below the solidus. Further evidence for metamorphic garnet growth comes from textural relations and zircon trace element geochemistry, which shows a steadily decreasing HREE slope with zircon U-Pb age for both Chino Valley and Camp Creek xenoliths (Fig. 6).

Garnet-whole rock Sm-Nd, titanite U-Pb, and zircon U-Pb data provide geo-/thermochronologic constraints on the evolution of Chino Valley xenoliths. Gneissose lower crustal xenoliths, characterized by abundant primary plagioclase and clinopyroxene (largely altered to amphibole), with minor amounts of apatite and ilmenite yield zircon U-Pb ages of ca. 1.7 Ga and time-corrected zircon εHf values of +9.6 (Fig. 7), interpreted as in situ depleted mantle-derived cratonic lower crust. In contrast, zircon grains extracted from relatively clinopyroxene-rich, garnet-poor, and amphibole-absent arclogite yield a unimodal spread of concordant Late Cretaceous to early Cenozoic ages ranging from ca. 100 to 50 Ma and a peak centered at ca. 75 Ma (Fig. 8; Chapman et al., in review). Schulze et al. (2017) report a similar Late Cretaceous-early Cenozoic distribution of U-Pb zircon ages from amphibole-bearing Chino Valley arclogite, with an additional Late Jurassic-Early Cretaceous age peak also observed at Camp Creek (see stop 2). These ages are interpreted to record zircon growth, rather than cooling, as closure temperatures calculated for sub-orogenic plateau (i.e., very slowly cooled) 25-100 µm grains (e.g., Dodson, 1973; Cherniak and Watson, 2000) exceed equilibration temperatures of the studied xenoliths. Zircon U-Pb data from Chino Valley arclogite are limited relative to Camp Creek, where two groups of samples (high-MgO and clinopyroxene-rich nodules with exclusively Late Cretaceous-early Cenozoic ages versus low-MgO and amphibole-rich nodules with Late Jurassic-Early Cretaceous and Late Cretaceous-early Cenozoic age populations) are readily distinguishable and discussed further at stop 2.

Whole rock Nd isotopic values vary widely depending on lithology. A single sample containing garnet, clinopyroxene, and mosaic orthopyroxene yields an εNd(0) value of +8, identical to values determined from websterite by Smith et al. (1994). Websterite whole rock Nd values mated with garnet values yield a ca. 155 Ma two-point isochron (Fig. 9; Rautela et al., in review). In light of paleopressures in the 17 to 23 kbar range calculated from websterite (Smith et al., 1994), aluminous orthopyroxene (Al2O3 ranging from 6-7 wt.%; Arculus and Smith, 1979), enriched Nd isotopic values, and a garnet-whole rock isochron age, websterite nodules from Chino Valley likely represent Late Jurassic mantle melts quenched ~75 km below the surface.

In contrast, garnet-clinopyroxenite nodules lacking orthopyroxene yield far less radiogenic whole rock εNd(0) values, ranging from -6 to -9 (Smith et al., 1994; Rautela et al., in review). Amphibole-rich and amphibole-poor garnet-clinopyroxene assemblages appear to span similar ranges in whole rock eNd values (Smith et al., 1994). Garnet from amphibole-rich assemblages is Such low negative εNd(0) values preclude derivation from oceanic crust or young depleted mantle, instead requiring extraction from Precambrian lower lithosphere (i.e., lower crust and/or mantle material). In comparison with websterite xenoliths, arclogite yields much younger garnet-whole rock isochron ages, ranging from 33 to 48 Ma (Fig. 9; Rautela et al., in review). Titanite derived from multiple samples of Chino Valley garnet clinopyroxenite yields an age of ca. 57 Ma that overlaps garnet-whole rock ages at the 1σ level (Erdman et al., 2016). Considering that arclogitic nodules equilibrated at temperatures overlapping garnet Sm-Nd and titanite U-Pb closure (Cherniak 1993; Ganguly et al., 1998; Frost et al. 2000), the observed Cenozoic spread in garnet and titanite ages point to long-term (10s of Myr) residence within the lower lithosphere and/or incomplete reequilibration with ca. 25 Ma host latite.

Interpretations of the above field, petrographic, geochronologic, and geochemical data are presented in the context of additional observations presented at stops 2 and 3 in “Discussion.” Return to vehicles and retrace route to Sunset Point Rest Stop for lunch.

After lunch, return to vehicles, reset odometers to zero, and proceed south to stop 2.

Cumulative mileage	*Description*
0.0	Merge onto I-17 south and continue for 27 miles to exit 223 for Carefree, reset odometers to zero.
12.1	Turn left (east) onto E. Carefree Highway and continue 12.1 miles.
13.1	Turn left (north) on N. Tom Darlington Dr. and continue 1 mile.
14.1	Turn right (east) onto E. Stage Coach Pass and continue 1 mile.
14.3	Turn left (north) onto Mule Train Rd and continue 0.2 miles.
19.5	Turn right (east) onto E Cave Creek Rd and continue 5.2 miles.
2.8	Turn slight right (east) onto Bartlett Dam Rd/Srv Rd 205 and zero odometers. Latite exposures at Blue Mountain come into view 1.2 miles after turning onto Bartlett Dam Rd/Srv Rd 205. Disaggregated Proterozoic granite and grus in road cuts as we descend into the valley floored by Camp Creek Wash. After 2.8 miles, turn right (south) and take Tonto Recreation Alliance – Great Western Trail road 413 into Camp Creek Wash.
3.4	Turn right onto road 413 noting exposures of basin fill conglomerate, continue 0.6 miles, turn vehicles around and park. Prepare to examine exposures on the east side of Camp Creek Wash. We will spend at least 30 minutes away from the vehicles. Anticipate temperatures around 100 ˚F and make sure to drink plenty of water.

Stop 2. Camp Creek xenolith locality (UTM 12S, 427959E, 3744672N)

The Camp Creek xenolith locality is mapped on Blue Mountain in the Wildcat Hill quadrangle, ~5 km east of the fringe of the Phoenix metropolitan area (Figs. 2 and 10; Skotnicki et al., 1997). This locality was first described by Esperanca (1984), who noted distinct gray and pink high-K latite volcanic necks of limited lateral extent (<2 km) – each erupting through Proterozoic granite and early Cenozoic fanglomerate, likely related to Colorado Plateau “rim gravels,” – that flowed southward (Fig. 11). Both latites contain phlogopite, diopsidic clinopyroxene, and apatite as phenocryst phases (5-10% by volume) in a fine groundmass of flow-aligned sanidine, clinopyroxene, biotite, and magnetite (Esperanca and Holloway, 1986), though the pink latite has relatively elevated Fe_2O_3 (likely contributing to its color), is generally coarser grained, contains more xenoliths, and weathers more readily. No precise ages exist for latite at Blue Mountain, though lithologic and isotopic similarities permit tentative correlation with 26.5 Ma (Ar-Ar age) latite in the vicinity of New River, AZ, ~30 km WNW of Blue Mountain (Leighty, 1997).

Both latites contain subrounded to rounded xenoliths ranging from 1-30 cm in diameter of (in order of diminishing abundance) garnet-amphibole-clinopyroxene assemblages (arclogite), granulite, and websterite (Esperanca, 1984; Esperanca and Holloway, 1986; Esperanca et al., 1988; Erdman et al., 2016). Xenoliths are common as lag deposits in dry creeks and slopes flanking Blue Mountain (Fig. 11), especially adjacent to exposures of pink latite.

Two groups of arclogite xenoliths are observed at Camp Creek (Esperanca et al., 1988; Erdman et al., 2016). The first are moderately to strongly foliated garnet-clinopyroxene-amphibole rock with poikilitic texture and relatively low bulk-rock MgO (<13 wt. %) and SiO_2 (<48% wt. %). Garnet is more abundant than clinopyroxene in this group and is partially (along rims) to entirely replaced by kelyphite. Diopsidic clinopyroxene exhibits a cloudy/turbid texture, particularly along rims, and is partially to completely replaced and pseudomorphed by pargasitic amphibole, which makes up at least 10% by mode. Amphibole commonly exhibits cores with higher order birefringence than rims. Apatite and rutile are ubiquitous, though volumetrically minor, phases in this group, with ilmenite and titanite also present in several samples. Injected latite melt, characterized by feldspar- and biotite-rich pathways and quench phenocrysts identical to those observed at stop 1, make up a significant portion (up to ~30%) of the xenolith volume for this group. In contrast, the second arclogite group has high bulk-rock MgO (>13 wt. %) and SiO_2 (>48% wt. %) and exhibits an equigranular texture composed of relatively fresh clinopyroxene, subordinate garnet (kelyphite is restricted to <100 µm thick bands along rims), minor apatite and rutile, and less than 10% pargasitic amphibole as a clinopyroxene replacement product. Latite veins are less prevalent than in group 1 arclogite and make up <10% of the xenolith volume. Zircon is observed in ~10% of arclogite nodules at the Camp Creek locality. Representative thin section photomicrographs are shown in Figure 5.

The degree to which clinopyroxene is replaced by amphibole in arclogite is higher at Camp Creek compared to Chino Valley. Thermodynamic modeling requires that Camp Creek arclogite absorbed 0.2 to 0.5 wt% H_2O with respect to solid phases present in order reproduce observed amphibole modal abundance (generally 5-30%) and composition (hornblende and pargasite). We infer that Camp Creek, and to a lesser extent Chino Valley, arclogite experienced a petrogenetically late hydration event. Numerous studies suggest that flattening of the Farallon slab introduced water into the lower lithosphere of North America in Late Cretaceous-early Cenozoic time (e.g., Humphreys et al., 2003; Smith et al., 2004; Lee, 2005), a mechanism that may also explain the extent and timing of hydration of studied xenoliths.

Zircon grains extracted from Group 1 nodules yield a bimodal age distribution consisting chiefly of Late Jurassic (Kernel Density Estimate [KDE] peak at ca. 150 Ma) ages with a lower proportion (~25%) of Late Cretaceous-early Cenozoic (KDE peak at ca. 70 Ma) grains (Fig. 8). This age distribution is similar to that reported from amphibole-rich arclogite at Chino Valley (Schulze et al., 2017). Group 2 xenoliths yield a unimodal spread of concordant Late Cretaceous to early Cenozoic zircon U-Pb ages ranging from ca. 100 to 50 Ma and a peak centered at ca. 75 Ma (Fig. 8).

Whole rock Nd isotopic values from Camp Creek garnet-amphibole-clinopyroxene assemblages overlap those observed from Chino Valley, ranging from -3 to -9 (Esperanca et al., 1988; Rautela et al., in review). Group 1 and group 2 xenoliths tend toward the bottom and top of this isotopic range, respectively. Similarly, zircon extracted from Camp Creek garnet-clinopyroxenite xenoliths yields low negative εHf values (-3 to -13), time-corrected to corresponding U-Pb ages, with lower (-6 to -13) and higher (-7 to -3) values observed for group 1 and group 2 samples, respectively (Fig. 12).

A significant Precambrian lower lithosphere component is required to explain low negative eNd and εHf values from these nodules. Intriguingly, group 2 (i.e., amphibole-poor and high MgO) arclogite appear to require a lower proportion of ancient lithosphere and a higher proportion of more radiogenic component(s) such as asthenosphere and/or Farallon lithosphere (Fig. 12). Websterite nodules yield lower values than those at Chino Valley, but are still significantly more radiogenic (+1 to +2) than orthopyroxene-absent assemblages (Esperanca et al., 1988) and probably also represent deep-seated fossilized mantle melts.

Camp Creek arclogite samples yield a spread of garnet-whole rock isochron ages that overlap and extend to older ages compared to Chino Valley, ranging from 30 to 60 Ma (Esperanca et al., 1988; Rautela et al., in review). These data support the inference drawn from Chino Valley that arclogite xenoliths resided within the lower lithosphere and/or partially reequilibrated with ca. 25 Ma host latite.

Return to vehicles. Group will divide into those needing to return to Phoenix by late afternoon-early evening and those wanting to visit stop 3, who don't mind returning to Phoenix by mid-evening. For those continuing to stop 3, reset odometers to zero.

Cumulative mileage	*Description*
7.9	Return to E. Cave Creek Road- Bartlett Dam Rd/Srv Rd 205 and continue west on E. Cave Creek Road to intersection with N. Pima Road (7.9 miles). Turn left (south) on N. Pima Road.
20.0	Continue south on N. Pima Road for 12.1 miles
37.8	Take the ramp onto AZ-101 Loop S, merge onto AZ-101 Loop S, and continue for 17.5 miles.
113.7	Take exit 55A-55B to merge onto US-60 E toward Globe, stay on US-60 E for 79.9 miles. At mileage 56.2 (between mile markers 194 and 195), we will cross the Central Arizona Project canal for the second time on the trip. Note views of a resurgent dome complex of the western Superstition Mountains at 11 o'clock. The Superstition Mountains and adjacent Goldfield Mountains are underlain almost entirely by basaltic to rhyolitic ash-flow tuffs, lavas, and breccias of the Superior volcanic center, generated in at least five caldera-forming events from 30 to 14 Ma (Nealey and Sheridan,1989). These volcanic products are extensive, blanketing >5000 km^2 of the transition zone, and are observed two more times along US-60 E: at mile 224 (Picketpost Mountain, an eroded volcanic vent) and again over an ~ 8 mile stretch from Superior to Top-of-the-World. Five miles east of Florence Junction, at the Tonto National Forest boundary (odometer mileage 78.9 and east of mile marker 217), exposures of chiefly greenschist facies argillite, pelite, and tuff protoliths belonging to the Paleoproterozoic Pinal schist appear in road cuts (Anderson, 1989). The Pinal schist is thought to represent deep-water turbidites that accumulated in the forearc basin and/or accretionary complex of the Paleoproterozoic Mazatzal arc (Meijer, 2014). East of Superior (odometer mileage 88.6 and at mile marker 227), US-60 E crosses a N-S striking and W-side-down normal fault (the Concentrator fault) that bounds

	the Pinal Mountains. The Concentrator fault is considered inactive as it is overlapped by Miocene volcanic rocks (Richard and Spencer, 1998). Pre-Miocene footwall strata are tilted to the east resulting from motion along the Concentrator fault. Superior owes its >140 year history of silver and copper mining to mineralization along pre-Cenozoic faults ~3,000 feet below the highway. Currently, the proposed Resolution Copper project aims to begin extracting an estimated ~1.6 billion metric tons of porphyry copper deposits from a depth of over 7,000 feet within the next decade. From here, US-60 E ascends quickly parallel to Queen Creek, the major drainage of the catchment area. The angular unconformity between late Cenozoic volcanic rocks atop Paleozoic passive margin strata (Cambrian Bolsa Quartzite, Devonian Martin Limestone, Mississippian Escabrosa Limestone, and Pennsylvanian Naco Limestone), is on spectacular display at the Queen Creek Bridge, ~1 mile east of Superior (Fig. 13). Exposures of the ca. 64 Ma Schultze granitic stock (e.g., Spencer et al., 2003) begin at Top-of-the-World (odometer mileage 96.5 and east of mile marker 235) and are recognized by pervasive fracturing in roadcuts and spheroidal weathering in natural exposures. The Schultze stock is one of numerous porphyritic to equigranular granitic bodies exposed in Basin and Range and Colorado Plateau transition zone provinces of Laramide age. Most economic copper deposits in Arizona are associated with this unit, including the multi-pit Miami mine visible over the next 15 miles, and are thus named "porphyry copper deposits."
114.5	Enter Globe, AZ, turn right (SE) on US-70 E/Ash St. and continue 0.8 miles to Circle K gas station (2011 E. Ash St.). Fill up on gas, use restrooms, and pick up San Carlos Apache Tribe recreation permits.

Return to vehicles and reset odometers to zero.

Cumulative mileage	*Description*
15.5	Globe, Arizona Circle K gas station (2011 E. Ash St.). Proceed east on US-70 E to stop 3. In road cuts note exposures of locally-derived and poorly sorted coarse sand and fanglomerate of the Pliocene-early Pleistocene Gila conglomerate (Peterson, 1962). This unit accumulated in the Safford basin, a discontinuous >300 km-long (extending from Globe to northern Mexico) graben produced during low-angle normal faulting associated with metamorphic core complex formation and later reinvigorated by high-angle Basin and Range extension (Gootee, 2012). In early Pleistocene time, as deposition of the Gila conglomerate waned, the stretch from Globe to San Carlos was a part of a broad intermontane basin occupied by a shallow internally drained lake. Note lacustrine deposits of sandstone, clay, and marl beginning a few miles east of Globe. After 6.5 miles, keep straight to stay on US-70 E. At 15.5 miles, turn left off US-70 E and onto an unmarked dirt road. Note sign reading "Non-members permit required beyond this point."

16.4 Proceed on road to right (not road that parallels US-70 E. At 16.1 miles, keep right at branch in road.

16.9 At mileage 16.4, we enter the vent from which basalt at Peridot Mesa erupted from the south. After traversing the vent, we ascend along its north rim (mileage 16.8), noting exposures of pyroclastic surge deposits. Upon leaving the vent and emerging atop Peridot Mesa, continue left on better maintained road (mileage 16.9).

18.3 Road forks at 17.0, stay right. Road forks at 17.3, stay left. Keep straight (no sharp left) at 17.9. Turn left at 18.0 and park at 18.3 next to tailings. We will spend at least 30 minutes inspecting xenoliths in the excavated drainage that bisects Peridot Mesa (peridotite xenoliths are most abundant ~7-10 m below the mesa surface. Again, plan for temperatures around 100 ˚F and make sure to drink plenty of water.

Stop 3 (optional). Peridot Mesa volcanic center and San Carlos mantle xenoliths (UTM 12S, 548632E, 3688525N)

At Peridot Mesa we will examine volcanic products and xenoliths that differ significantly from those visited at stops 1 and 2. The profound difference from stops 1 and 2 to stop 3 invites discussion of changes in style and composition of volcanism and xenolith suites that occurred between 15 and 10 Ma in central Arizona.

Peridot Mesa is located on San Carlos Apache Tribe land and a permit is required to visit the mesa (Fig. 14). Native Americans have long used the area to mine for gem-quality olivine. Approximately 90% of worldwide gem-quality olivine crystals (referred to in the gemstone world as "peridot" or "peridot stone") come from here.

We will see a basanitic diatreme that erupted some 500,000 years ago in several stages. Magmatism is broadly related to Basin and Range extension in this region. All but the most explosive, initial stage of the eruptive cycle, contain mantle-derived xenoliths. This is an indication that magma extracted from the mantle traveled through the crust at high speeds without stopping in intermediate magma chambers. The energy of the initial eruption was probably due to the high amount of volatiles (water and carbon dioxide) that originate in the mantle source region.

Stop 3, like stops 1 and 2, resides within the Colorado Plateau transition zone (Fig. 2). Basement rocks in the area include Proterozoic granitoids, miogeoclinal (western North American passive margin) rocks of Paleozoic age (note examples of each after descending from Peridot Mesa and driving west on AZ-170 N), and Laramide-age subduction-related granitic bodies (e.g., the Schultze stock). The higher stratigraphic units in the area are alluvial (e.g., the Gila Conglomerate) and lacustrine deposits of Pliocene-Quaternary age. The Peridot Mesa volcanic sequence lies directly over these lake beds; volcanism took place after the San Carlos lake dried out. Basaltic layers are found interbedded with the extensive lake deposits and attest to a longer history of volcanism in the area.

The field trip area comprises several mafic volcanic products that are part of the San Carlos Volcanic field (7-0.5 Ma). The lava flow that makes Peridot Mesa was dated at 0.58 Ma (±0.21, Bernatowicz, 1981), consistent with the Quaternary age of underlying sediments. The entire volcanic suite that vented from Peridot Hill erupted within a few weeks or less. Peridot Hill is a classic diatreme that started as a phreatomagmatic eruption and was later followed by lava flows. The Peridot Mesa eruption sequence consists of a lower pyroclastic surge unit overlain by welded scoria with cinder, bombs and spatter abundant near the vent, and at least two mappable massive NE-trending lava flows that were later incised by canyons (Fig. 14). The surface of the upper lava flow makes up Peridot Mesa. These flows contain abundant ultramafic xenoliths that are concentrated at the base of the flows (Fig. 15).

Soda Springs vent is a neighboring vent of mugearittic composition (alkali basaltic, relatively high in silica) that yields a K-Ar age of 5.3 ± 0.13 Ma (Holloway and Cross, 1978). Several other undated (but clearly late Pliocene to Quaternary) volcanic centers are found nearby. Xenocrysts of anorthoclase and kaersutite have been reported from other vents though it is unclear if other vents contain mantle xenoliths.

Ultramafic xenoliths from San Carlos are predominantly spinel lherzolite, but clinopyroxenite, orthopyroxenite, websterite, wehrlite, dunite, and gabbro are also found. Most samples are coarse grained and commonly display cross-cutting relationships that record melt percolation events through the mantle. The abundance, size (up to 80 cm!), and freshness of xenoliths have made this location a favorite for materials used in a variety of geological and geophysical studies. Peridotite nodules from here are the standards for many geochemical techniques, and starting materials for experimental petrology or rock physics. Ironically, the geology of these xenoliths remains somewhat enigmatic, despite some classic studies that were carried out here and that paved the way for modern petrologic studies of xenoliths. Of these efforts, the cornerstone is Frey and Prinz (1978), which defined Cr-diopside (Group I) and Al-augite (Group II) mantle xenoliths. Cr-diopside-bearing rocks are the standard upper mantle assemblages, whereas Al-augite-bearing rocks are frozen melts or cumulate of melts residing in the mantle. It is widely accepted that Group I xenoliths make up the bulk of the solid uppermost mantle, although statistically they do not predominate in xenoliths here or elsewhere.

Further complicating the petrography of the continental mantle are phlogopite and/or kaersutite veins that document the presence of aqueous fluids (melts?). Check for rare veins with phlogopite found in some peridotite xenoliths. Kaersutite (an amphibole stable in the mantle) is typically found as xenocrysts, rather than in xenoliths; because of that, some have suggested that they may be cognate and not accidental.

San Carlos ultramafic rocks are almost without exception undeformed, in contrast to other localities in SW North America where sheared xenoliths are also present (e.g., Basu et al., 1977). This may be due to the high equilibration temperature that leads to annealing of any older, deformed fabrics. Large, up to 5 cm "porphyroblasts" of olivine (the gem quality crystals the Native Americans are interested in) are particularly puzzling as they are not typical for the overall grain size of the samples here or elsewhere in continental regions. They may be secondary crystals that grew during partial melting events.

Chemically, Group I xenoliths from San Carlos range from depleted (in REEs, isotopic ratios, and clinopyroxene) to enriched - spanning a range larger than all oceanic basalts worldwide. Undoubtedly, these rocks record multiple local (tens of cm scale) melt depletion and metasomatic enrichment episodes, though the timing of these events is ambiguous.

There is no question these xenoliths represent the shallowest mantle beneath the area – they were "sampled" between 30 to 60 km below the surface. The bulk of the upper mantle is assumed to be composed of spinel lherzolites, whereas the other assemblages are veins, sills, and cumulate layers within peridotitic layers. There are three possible types of shallow mantle beneath the area: (1) lithospheric North America, (2) oceanic, Farallon slab mantle, and (3) deeper asthenospheric mantle that upwelled with the onset of Basin and Range extension. Equilibration temperatures determined from San Carlos fall within a narrow range of 1020-1060 ˚C (e.g., Galer and O'Nions, 1989). These values, which are typical for most Basin and Range xenoliths, are too high for a normal continental mantle lid, but too low to be part of the convective mantle. Paleopressures from San Carlos nodules are inferred to be in the 30-70 km range but no modern barometric studies are available .

The range of isotopic ratios of Sr, Nd and Pb (Galer and O'Nions, 1989) exceeds the entire spectrum of isotopic compositions displayed by all analyzed (many hundreds) Basin and Range basalts. These values do not reflect a single melt depletion event and instead point to multiple episodes of metasomatism (melt enrichment). San Carlos contains some of the most depleted peridotites on the planet, but also some of the most enriched ones.

We have no unambiguous results that can rule out any of the three origins listed above, based on the xenolith composition, recorded temperatures, range of isotopic composition, and lack of chronologic techniques for these rocks. The relatively low $^{3}He/^{4}He$ ratios in San Carlos xenoliths (~8x atmospheric ratios, Porcelli et al., 1992) are similar to MORB values and those data can (at least) rule out a fourth, plume-related origin. This is a great example that illustrates the compositional and thermal complexity of the continental mantle and our inability to decipher its geologic evolution. We will discuss these issues in the field in more detail.

The modern Moho beneath the San Carlos-Globe area is located some 31 km below the surface and the average crustal Vp/Vs ratio in the area is relatively low, ~1.72 km/s, consistent with a felsic crust throughout

(Frassetto et al., 2006). There are no relevant data constraining the temperatures below the Moho, or any other compositional details. The xenoliths on display here are part of the immediately sub-Moho mantle, and may represent portions of the lithosphere, asthenosphere, or perhaps "astenospherized" lithosphere, a reflection of small-scale convection processes resulting from multiple phases of Cenozoic extension in the Colorado Plateau transition zone?

Return to vehicles and reset odometers to zero.

Cumulative mileage	*Description*
1.5	Drive northeast along the most prominent road atop Peridot Mesa toward San Carlos. After 1.3 miles, descend off Peridot Mesa into San Carlos, noting the contact between dark basalt and light tan lacustrine deposits over the next 0.2 miles.
2.9	Turn left on Aravaipa Rd (returning to pavement). After 1.4 miles, turn left on AZ-170 N/Cutter Rd.
13.6	Continue on AZ-170 N for 10.7 miles, noting angular unconformity between dipping Paleozoic strata and flat-lying fanglomerate at odometer mileage=7.8 miles, to US-70 W.
	Return (96 miles) via Globe to Phoenix Convention Center. End of field trip.

SUMMARY AND DISCUSSION

New results highlighted on this fieldtrip indicate that studied arclogitic xenoliths at Camp Creek and Chino Valley, yielding chiefly Late Jurassic and Late Cretaceous zircon U-Pb ages and early Cenozoic cooling ages (titanite U-Pb and garnet-whole rock Sm-Nd) that overlap the thermal history of the SCB. With the exception of a small number (<10% of analyzed grains) of Latest Cretaceous-early Cenozoic grains, studied arclogite xenoliths are older than Arizona porphyry copper deposits (Vickre et al., 2014; Chapman et al., 2018). These relations lead us to assert that arclogitic xenoliths are not native to central Arizona and instead represent LC-SCML fragments displaced eastward from beneath the SCB and reaffixed beneath the transition zone (Axen et al., 2018; Chapman et al., in review). A profound shift from bimodal and pyroxenite-including volcanism toward more mafic and peridotite-including volcanism, highlighted on this trip at the San Carlos locality, occurred in the transition zone in Miocene time (Nealey and Sheridan, 1989). We suggest the following model for the petrologic and tectonic evolution of Colorado Plateau transition zone lower lithosphere (Figs. 16 and 17).

In Late Jurassic time, low-MgO amphibole-rich arclogite xenoliths began forming as a mafic keel to continental arc magmas emplaced into the central and eastern Mojave Desert and associated with eastward subduction of the Farallon plate (Barth et al., 2008; Needy et al., 2009; Erdman et al., 2016; Chapman et al., 2018). Following an Early Cretaceous lull in arc activity, the sub-SCB root experienced a ca. 80-70 Ma pulse of growth associated with increased magmatism in the SCB (Wells and Hoisch, 2008; Needy et al., 2009; Chapman et al., 2018), as recorded by Late Cretaceous zircon ages in low-MgO amphibole-rich and high-MgO amphibole-poor arclogite. These Late Cretaceous additions to existing SCB root material and concomitant plutonism at higher levels of the crust likely resulted from delamination and foundering of LC-SCML, destabilized by slab shallowing-related lateral stresses, and ensuing upwelling of hot asthenosphere (Leventhal et al., 1995; Wells and Hoisch, 2008).

The onset of shallow-angle subduction at the plate margin is constrained by forward and inverse plate motion modeling and marked by the underplating of schist to the ca. 90-80 Ma time interval (e.g., Grove et al., 2003; Liu et al., 2010; Chapman, 2017). The leading edge of the shallowly subducting segment likely reached the central Mojave Desert, then ~500 km inboard from the margin, by 75 Ma assuming an orthogonal converge rate of 100 km/Myr (Engebretson et al., 1985). At this point, arclogite dislodged from the SCB above and was tectonically bulldozed by the tip of the shallow slab segment from its origin at least ~500 km further inboard, perhaps as far as SW Colorado (Axen et al., 2018), over the ensuing ~5 Myr to the Colorado Plateau transition zone. In the process, the foreland crust thickened significantly, forming the "Nevadaplano" of DeCelles (2004), and magmatism swept inboard, forming ca. 75-55 Ma granitic stocks and associated copper mineralization in the transition zone (e.g., Vickre et al., 2014; Chapman et al., 2018). We suggest that the combination of magmatism and crustal thickening-related radiogenic heating precipitated Latest Cretaceous-early Cenozoic zircon growth in dislodged arclogite.

Cenozoic Sm-Nd garnet and titanite U-Pb ages (Esperanca et al., 1988; Erdman et al., 2016) indicate that displaced arclogite remained hot (>700 ˚C) for 10s of Myr following its dispersal. This observation is readily explained by early Cenozoic residence beneath thickened orogenic crust plus the well-documented Eocene to Miocene westward sweep of magmatism, including the ca. 25 Ma arclogite xenolith-hosting latite, that accompanied rollback and tearing of the Farallon slab (e.g., Coney and Reynolds, 1977; Humphreys, 1995). This time also marked a shift from convergent to extensional tectonism in the SW U.S., including the initiation of the southern belt (i.e., California, Arizona, and Sonora) of Cordilleran metamorphic core complexes (Dickinson et al., 2009; Whitney et al., 2013).

The profound shift toward more mafic volcanism plus the lack of arclogite and abundance of spinel peridotite xenoliths between 15 and 10 Ma (e.g., the San Carlos locality highlighted on this trip; Galer and O'Nions, 1989; Nealey and Sheridan, 1989), in conjunction with vertically-dipping seismically fast features and a "double Moho" beneath the western edge of the Colorado Plateau and transition zone (Levander et al., 2011; Erdman et al., 2016), suggest that arclogite has been foundering into the mantle and being replaced by upwelling asthenosphere since at least 12 Ma and perhaps as early as 25 Ma. We suggest that post-Laramide rollback of the Farallon slab and associated influx of asthenosphere destabilized the foreign mafic keel of the Colorado Plateau, leading to its removal and heating of overlying material. Ensuing flow in the weakened lower crust, which remained feebly coupled to the upper crust, facilitated the localized surface extension that ultimately resulted in the southern belt of Cordilleran metamorphic core complexes.

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FIGURE CAPTIONS

Figure 1. Fence diagram showing idealized lithospheric structure for beginning of Cenozoic time for sections across Sierra Nevada batholith (SNB)-Great Valley forearc, southern California batholith (SCB)- Colorado Plateau transition zone, and linking section across the southern SNB. Locations of sections shown on inset. Note the position of sub-continental mantle lithosphere (SCML) and arclogite at the eastern limit of the SCB section. Abbreviations: Cz, Cenozoic; K, Cretaceous; Mz, Mesozoic; pC, Precambrian.

Figure 2. Simplified geologic and tectonic map of Arizona, highlighting Colorado Plateau, Basin and Range, and Colorado Plateau Transition Zone (CPTZ) provinces and with planned field trip stops overlain. Exposures of Proterozoic crust and overlying Paleozoic plus Mesozoic cover rocks shown in orange and blue, respectively. Exposures of Jurassic arc-related assemblages shown in black; Laramide plutons and stocks associated with porphyry copper deposits outlined in black with diagonal line pattern fill.

Figure 3. Simplified geologic map of Chino Valley xenolith locality (Modified after Stefanov, 1993 and Cunningham, 2001). Base map: U.S. Geological Survey 7.5 minute topographic map of the King Canyon Quadrangle (2018). Contour interval is 20 feet.

Figure 4. Field photographs of structural and petrologic features at Chino Valley locality. (A) View to SW showing subvertical joints in Sullivan Buttes latite. Photo is 200 m across. (B) Xenolith "lag" along Forest Service Road 641 at Chino Valley locality. Note numerous subrounded reddish-brown xenoliths weathering from gray latite. Hammer is 34 cm long. (C) Rounded garnet-clinopypyroxene xenolith set in Sullivan Buttes latite. Note "breadcrust" texture on the outside of the xenolith, most likely due to thermal expansion associated with inclusion in latite. (D) Subrounded amphibole-rich garnet-clinopypyroxene xenolith in flow-banded Sullivan Buttes latite. Pen is 14 cm long.

Figure 5. Photomicrographs of petrologic features in Colorado Plateau Transition zone xenoliths. All panels show plane-polarized light on the left and cross-polarized light on the right, respectively (A) Sample 12CC2A9—Garnet clinopyroxenite with relict clinopyroxene in cores of amphibole, which has replaced most of the primary clinopyroxene. Note plagioclase-rich injected melt pockets, kelyphitized garnet rim domains, and moderately strong foliation defined by elongate amphibole running from left to right of frame. (B) Sample 12CC2A3—Garnet clinopyroxenite with a higher proportion of primary clinopyroxene to secondary amphibole than sample 12CC2A9. (C) Sample 16CC2J— Fresh garnet-clinopyroxenite with clinopyroxene as an intercumulate phase. (D) Amphibolitized felsic granulite gneiss. Note biotite-clinopyroxene latite showing flow foliation from top to bottom of photo. Mineral abbreviations: Amph— pargasitc amphibole; Cpx—clinopyroxene; Grt—garnet; k—kelyphite; Pl—plagioclase; Rt—rutile.

Figure 6. zircon TE geochem

Figure 7. (A) Concordia diagram for studied Chino Valley granulite xenolith from LA-MC-ICP-MS zircon U-Pb data (Chapman et al., in review). Concordia intercept age after Vermeesch (2018). Errors are 2σ. MSWD—Mean square of weighted deviates. Hf evolution diagram (B) for cv1n.

Figure 8. Concordia plot (A) and (B) Non-normalized kernel density estimates with 10 Myr bandwidth comparing U-Pb zircon ages from Colorado Plateau transition zone (CPTZ, Chino Valley and Camp Creek localities) xenoliths with pluton ages from the Mojave Desert (Barth et al., 2008; Wells and Hoisch, 2008; Needy et al., 2009; Chapman et al., 2018) and Colorado Plateau transition zone (Vickre et al., 2014; Tosdal and Wooden, 2015; Chapman et al., 2018).

Figure 9. Garnet-whole-rock Sm-Nd isochron for Chino Valley and Camp Creek xenoliths from Rautela et al. (in review). Error bars and calculated two-point isochron ages are 2σ.

Figure 10. Simplified geologic map of Camp Creek xenolith locality (Modified after Skotnicki et al., 1997). Base map: U.S. Geological Survey 7.5 minute topographic map of the WIldcat Mountain Quadrangle (2018). Contour interval is 20 feet.

Figure 11. Field photographs of structural and petrologic features at Camp Creek locality. Hammer in photos B and C is 34 cm long. (A) View to ESE showing gray and pink latite stratigraphically above conglomerate (unit Tc on Figure DRX). Field of view 500 m long. (B) Detailed view of pink latite-conglomerate contact, on which hammer is placed. (C) Pink latite hosting numerous subrounded, fist-sized and smaller xenoliths. (D) Garnet-amphibole-clinopyroxenite xenolith recovered from xenolith "lag" below host latite. Average-sized human hand for scale. Note gray, foliation-parallel injections of host latite into the xenolith. Also note subrounded and smooth morphology of xenolith, the most common xenolith morphology at both Camp Creek and Chino Valley localities, interpreted to form by some combination of physical abrasion and chemical interaction between xenolith and hotter host magma (e.g., Heger, 2013).

Figure 12. Zircon Hf evolution diagram for Chino Valley and Camp Creek arclogite xenoliths.

Figure 13. Panoramic view to the SE showing angular unconformity between moderately east-dipping Paleozoic passive margin sediments below subhorizontal late Oligocene-early Miocene tuff of the Superior volcanic center. Photo is ~1 km wide.

Figure 14. Simplified geologic map of San Carlos xenolith locality (Modified after Wohletz, 1978).

Figure 15. Field photographs of structural and petrologic features at San Carlos locality. Hammer in photos A-C is 34 cm long. (A) View of quarry wall showing a concentration of peridotite xenoliths (weathered orangish-brown) at the base of basalt flow. (B) Detailed view of xenolith-rich interval, most of which are relatively fresh peridotite. (C) View on quarry floor showing diverse array of xenoliths in including peridotite (green), dark gray to black phlogopite- and/or kaersutite-rich nodules, and peridotite invaded by phlogopite and/or kaersutite veins. (D) Detailed view showing large "porphyroblasts" of gem quality peridot (light green) and clinopyroxene (dark green) in lherzolite xenoliths. Note small gabbroic nodule in upper right hand corner. Average-sized human hand for scale.

Figure 16. SW U.S. plate tectonic reconstruction for A) middle Jurassic, B) Early Cretaceous, C) Late Cretaceous, D) Latest Cretaceous, E) late Cenozoic, and F) Recent time (modified after DeCelles, 2004; Saleeby and Dunne, 2015). Farallon/Pacific plate trajectories from Engebretsen (1985). Surface outline of LC-SCML foundering from Levander et al. (2011) and Erdman et al. (2016). Core complex and schist kinematics from Dickinson (2009) and Chapman (2017), respectively. See text for details. Abbreviations: CC, Camp Creek; CV, Chino Valley; MTJ, Mendocino triple junction; POR, Pelona-Orocopia-Rand schist; RTJ, Rivera triple junction; SC, San Carlos.

Figure 17. Cross sections corresponding to x-x' locations in Figure 16. Vertical equals horizontal scale. Colors correspond to those in Figure 3. See text for details. Abbreviations: CC, Camp Creek; CO, Colorado; CV, Chino Valley; Ig., ignimbrite; LC, lower crust; MCC, metamorphic core complexes; MSL, mean sea level; SAf, San Andreas fault, SC, San Carlos; SCML, sub-continental mantle lithosphere; SOML, sub-oceanic mantle lithosphere.