The Pleistocene (1.65 Ma) Crystal Knob volcanic neck in the California Coast Ranges is an olivine-plagioclase phyric basalt containing dunite and spinel peridotite xenoliths. The peridotites sample the mantle lithosphere beneath the Nacimiento belt of the Franciscan complex and crystalline nappes of the "Salinia terrane," which together form the exhumed remnants of the southern California Cretaceous subducting margin. Crystal Knob adds a new continental-margin constraint on mantle-lithosphere architecture, which xenolith studies have linked to crustal structure in the Mojave and Sierra Nevada. Six samples ranging from fertile lherzolites to harzburgite residues were analyzed. With of 10.3-11.0 and / of .702, they likely represent underplated sub-oceanic mantle. Pyroxene and rare-earth exchange geothermometry show equilibration at 950-1060 ºC. Phase stability, Ca-in-olivine barometry, and 65-90 mW/m^2 regional geotherms suggest entrainment at 45-75 km depth. The samples experienced variable partial melting, and re-enrichment of the hottest suggests deep melt interaction. We test the Crystal Knob temperature-depth array against a range of forward-modeled geotherms matching plausible tectonic scenarios for the origin of the regional mantle lithosphere. A shallow Mendocino slab window source is excluded, and a young Monterey-plate "stalled slab" fits xenolith thermobarometry but is geodynamically untenable. Suboceanic mantle lithosphere underplated during the Cretaceous and reheated at deep levels by the Miocene slab window matches both xenolith constraints and the processes that shaped the overriding crust and Mojave block. This suggests that the mantle lithosphere beneath the central California coast was emplaced after Cretaceous flat-slab subduction and records a thermal signature of Neogene ridge subduction.

# Introduction

The tectonic and petrogenetic processes by which Earth’s continental mantle lithosphere develops through time are of fundamental importance to geodynamics and Earth history. Volcanic rock-hosted mantle xenolith suites offer vertical sampling columns that reveal compositions and textures of the deep lithosphere at the time of xenolith entrainment and eruption. Petrogenetic studies of such xenolith suites offer constraints on depth of entrainment, thermal gradients, and geochemical evolutionary states. The integration of such constraints with regional geophysical data, as well as tectonic and petrogenetic studies of the overlying crust, offer critical insights into the interplay of plate tectonic and geodynamic processes that assembled the deep mantle lithosphere. In this contribution we present new petrologic and geochemical data on the Pleistocene Crystal Knob volcanic neck-hosted mantle xenolith locality of coastal central California, and integrate our findings both with regional geophysical and modeling studies of the mantle lithosphere and a well-developed literature on the tectonic and petrogenetic development of the overlying crust.

The Crystal Knob xenolith locality samples a crucially important lithospheric column through the Late Cretaceous convergent margin belt of the SW North American Cordillera. This regionally extensive belt is characterized by a voluminous continental magmatic arc, generated as the Farallon oceanic plate subducted eastward beneath western North America [*Ducea et al.*, 2015], and the Franciscan complex, the crustal-level accretionary complex of this subduction zone. The Franciscan is widely recognized for its tectonic inclusion of Farallon plate oceanic basement and pelagic sediment fragments, as well as upper plate siliciclastic sediments derived from the magmatic arc [*Blake et al.*, 1988; *Chapman*, 2016; *Chapman et al.*, 2016a; *Cowan*, 1978; *Murchey and Jones*, 1984; *Sliter*, 1984]. The crystalline basement of the "Salinian terrane" (or "Salinia") [*Page*, 1981] has recently been recognized as a series of far-displaced crystalline nappes, derived from the southern California segment of the Late Cretaceous magmatic arc and lying tectonically above Franciscan-affinity accretionary complex rocks in the central California coastal area [*Barth et al.*, 2003; *Chapman et al.*, 2012; *Ducea et al.*, 2009; *Hall*, 1991; *Hall and Saleeby*, 2013; *Kidder and Ducea*, 2006]. The Crystal Knob xenolith locality lies along the western margin of Salinia, adjacent to the Nacimiento fault [Figure **¿fig:context?**], a polyphase structural zone which, in its original geometry, constituted the local structural base of the Salinian crystalline nappe sequence [*Hall and Saleeby*, 2013]. In this setting, the Crystal Knob xenolith suite samples the uppermost mantle that has been constructed beneath both the Franciscan accretionary complex and its local veneer of Salinia crystalline nappes.

## Regional tectonic setting and previous mantle xenolith studies

The SW North American Cordillera hosts many xenolith localities, at which upper mantle-lower crustal rock fragments were entrained in mainly late Cenozoic volcanic eruptions. Early studies of a number of these xenolith suites focused on the systematizing of petrographic features and classifying samples into textural groups [*Nixon*, 1987; e.g. *Wilshire et al.*, 1988]. Subsequent application of modern geochemical and petrogenetic techniques has led to the recognition of distinct regional mantle lithosphere domains consisting of pre-Phanerozoic lithosphere, Cretaceous arc mantle wedge, underplated Farallon plate nappes, and late Cenozoic shallow convective asthenosphere [*Alibert*, 1994; *Beard and Glazner*, 1995; *Ducea and Saleeby*, 1998a; e.g. *Galer and O’Nions*, 1989; *Jové and Coleman*, 1998; *Lee et al.*, 2001a; *Livaccari and Perry*, 1993; *Luffi et al.*, 2009; *Usui et al.*, 2003].

These studies suggest a clear correspondence between xenolith-sampled upper mantle domains and surface geology. Sub-continental suites have generally been erupted through cratonic and peri-cratonic crust, mantle wedge suites through the Cretaceous large-volume batholith, and asthenospheric suites through active rifts. In contrast, xenolith suites derived from underplated Farallon plate mantle nappes have thus far only been recovered from more inboard crustal domains, requiring large sub-horizontal displacements and underplating along relatively shallow subduction megathrust systems [*Helmstaedt and Doig*, 1975; *Lee et al.*, 2001b]. Crystal Knob, having erupted through the Franciscan subduction accretionary complex, presents a rare opportunity to sample mantle lithosphere directly beneath the region of long-lived subduction accretion proximal to the plate edge.

The tectonic setting of the sub-Crystal Knob mantle lithosphere is best understood by restoring its host crustal rocks to their position prior to San Andreas transform offset. This places them outboard of the southern California batholith, the southern continuation of the Sierra Nevada batholith across the Garlock fault [Figure **¿fig:context?**; Figure **¿fig:reconstruction?**]. In this restored position, the crystalline nappes that constitute Salinia correlate to deeply exhumed Cretaceous arc plutonic rocks of the southernmost Sierra Nevada batholith and the northwestern zones of the southern California batholith, the latter widely exposed in the Mojave plateau [*Barbeau et al.*, 2005; *Chapman et al.*, 2012; *Saleeby*, 2003]. These deeply exhumed batholithic rocks share a common regional upper plate position above a polyphase low-angle fault system below which Franciscan-affinity, mainly metaclastic rocks were tectonically underplated in the Late Cretaceous [*Barth et al.*, 2003; *Chapman et al.*, 2010, 2016b, 2012; *Ducea et al.*, 2009; *Malin et al.*, 1995; *Yan et al.*, 2005]. The underplated metaclastic rocks are exposed in a series of tectonic windows that are labeled as subduction channel schists in Figure **¿fig:context?**. Detritus for the schist protoliths was derived from the upper plate batholithic belt, which was rapidly uplifted above a shallow flat segment of the greater Franciscan subduction megathrust system [*Barth et al.*, 2003; *Chapman et al.*, 2016a, 2013; *Saleeby et al.*, 2007]. The underplated schists tectonically encase blocks and nappes of Farallon plate oceanic basement and sediments. The shallow flat subduction megathrust segment is attributed to the ephemeral buoyancy resulting from the subduction of the conjugate massif to the Shatsky Rise oceanic LIP [*Livaccari et al.*, 1981; *Saleeby*, 2003; *Sliter*, 1984], which is currently resolved in deep seismic tomographic images beneath the interior of North America [*Liu et al.*, 2010; *Sun et al.*, 2017]. The tectonic position and structural attenuation of the Salinia nappes, as well as adjacent (restored for Neogene dextral faulting) deeply exhumed batholithic rocks, derives from both shallow subduction megathrust displacements and subsequent large magnitude trench-directed extensional faulting, correlated in time to the progress of the Shatsky conjugate deeper into the mantle beneath the North American plate [*Chapman et al.*, 2012; *Liu et al.*, 2010; *Saleeby*, 2003].

Shallow tectonic underplating of the subduction-channel schists directly beneath deep crustal large-volume batholithic rocks requires the prior tectonic erosion of the mantle wedge lithosphere (the source for the overlying batholith) and parts of the deepest arc crust. Integrated mantle xenolith studies and deep seismic imaging document this process. Late Miocene small-volume volcanic flows and plugs from the central Sierra Nevada batholith [Figure **¿fig:context?**] carry xenolith suites that sampled the Cretaceous mantle wedge of the overlying batholith [*Ducea and Saleeby*, 1998a, 1996; *Lee et al.*, 2006, 2001a; *Saleeby*, 2003]. The central and northern regions of the batholith are currently exposed over shallow to medial crustal depths (2 to 4 kb pressure), whereas at its southern reaches, a continuous gradient to deep levels (10 kb) is exposed [*Nadin and Saleeby*, 2008]. At this depth, the shallow subduction megathrust (remobilized as an extensional structure) forms the structural base of the batholith, with the underplated schists lying directly beneath [*Chapman et al.*, 2010, 2016b, 2012; *Saleeby*, 2003]. Seismic reflection imaging shows an effectively flat megathrust beneath the western Mojave plateau [*Yan et al.*, 2005] and dips of ~30ºN beneath the southernmost Sierra Nevada region [*Malin et al.*, 1995]. The Garlock fault [Figure **¿fig:context?**] nucleated during the early Miocene along this inflection in the megathrust [*Saleeby et al.*, 2016], which constituted a lateral ramp in the subduction megathrust system [*Chapman et al.*, 2016b; *Saleeby*, 2003]. In contrast to the central Sierra xenolith suite, mantle xenoliths recovered from the eastern margin of the southern California batholith record the partial tectonic erosion of sub-continental mantle lithosphere (including Cretaceous mantle wedge), and the underplating of Farallon plate mantle lithosphere [*Armytage et al.*, 2015; *Luffi et al.*, 2009; *Shervais et al.*, 1973; *Shields and Chapman*, 2016]. More specifically, the Dish Hill suite [Figure **¿fig:context?**] samples an upper mantle duplex with imbricated nappes of Farallon plate oceanic mantle lying in structural sequence beneath a relatively thin roof of attenuated continental lithosphere peridotites. The mantle duplex is interpreted to have formed as the Farallon plate retreated following Shatsky conjugate low-angle subduction [*Luffi et al.*, 2009]. The reconstructed position of Crystal Knob, directly outboard of the Dish Hill locality [Figure **¿fig:context?**], as well as its penetration of the Franciscan accretionary complex, suggests that the Crystal Knob suite may have sampled additional underplated Farallon mantle nappes outboard of the Dish Hill mantle duplex.

Mantle xenoliths of the eastern Sierra suite [Figure **¿fig:context?**], viewed in the context of late Cenozoic plate kinematics of the southern California region [*Argus and Gordon*, 1991; *Atwater and Stock*, 1998] suggest another potential origin for the sub-Crystal Knob upper mantle. The eastern Sierra suite occurs in Pliocene-Quaternary mafic lava flows and records a significantly steeper thermal gradient and more fertile compositions than the Cretaceous mantle wedge suite from the central Sierra [*Ducea and Saleeby*, 1996, 1998b]. Seismic data from the eastern Sierra region reveal asthenospheric mantle extending upwards to the base of the crust at ~30 km depth [*Frassetto et al.*, 2011; *Jones and Phinney*, 1998; *Jones et al.*, 2014; *Zandt et al.*, 2004], consistent with the eastern Sierra xenolith suite findings. Xenolith-hosting lavas within the eastern Sierra suite were erupted within the <5 Ma Owens Valley rift system [Figure **¿fig:context?**], which is presumably driven by upper mantle convection. Convective ascent of asthenosphere to relatively shallow levels and late Cenozoic regional volcanism of the central to southern California region [*Dodge*, 1988] are correlated with the opening of the Pacific-Farallon slab window [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Both the eruption of Crystal Knob and the origin of its upper mantle underpinnings could also owe their origins to asthenosphere ascended into a slab window.

Below we investigate the petrology and geochemistry of the Crystal Knob suite and link it to regional findings on other xenolith locations, as well as geophysical data and modeling, in order to pursue the origin of the upper mantle beneath the Crystal Knob eruption site and the central California Coast Ranges more generally.

# Crystal Knob xenolith locality

The Crystal Knob volcanic neck (35.806º N, 121.174º W) is a mid-Pleistocene olivine--plagioclase phyric basalt that erupted along the margin of the Franciscan assemblage 500 m west of the Nacimiento Fault in the Santa Lucia Mountains of central California [*Seiders*, 1989]. It was briefly described in *Wilshire et al.* [1988] but not studied in detail. The basaltic plug is ~80 m in diameter at the surface and has entrained abundant dunite and sparse spinel peridotite xenoliths [Figure **¿fig:field\_photo?**]. The dunites are cumulates, and grade texturally into single grain xenocrysts, or apparent phenocrysts, within the host lava. Sparse spinel peridotites are also present which lack textural features suggestive of igneous cumulate origin. In conjunction with the compositional data presented below, we interpret the peridotites as entrained fragments of the mantle lithosphere.

Samples were collected from the Crystal Knob lava with an emphasis on these polyphase peridotite xenoliths. Xenolith samples are 5-10 cm diameter friable peridotites with medium (200 µm -- 1 mm) grains. Additionally, several samples of the host basalt and dunite cumulates were collected to establish context for the xenoliths.

## Eruptive age

The ages of host lavas for mantle xenolith suites are critical for the application of their petrogenesis to tectonic and geodynamic processes [e.g. *Ducea and Saleeby*, 1998b]. The age of the Crystal Knob host lava was determined using the / technique on phenocryst plagioclase. A billet of the host lava (sample CK-1) containing visible plagioclase lathes was provided to the USGS Geochronology Laboratory in Denver, Colorado. The sample was irradiated in the USGS TRIGA reactor, and plagioclase feldspar grains were step-heated *in situ* using an infrared laser [Figure **¿fig:step\_heating?**]. loss was measured simultaneously on a Thermo Scientific Argus VI using 4 Faraday detectors (m/e 40-37) and ion counting (m/e 36). The detectors were intercalibrated using standard gas and air pipettes. The measurements are corrected for blanks above baselines, radioactive decay, and nucleogenic interferences, and standardized against a Fish Canyon sanidine with an age of 28.20 Ma. Non-radiogenic argon is assumed to have an atmospheric composition of / = 298.56 [*Cosca et al.*, 2011].

Step-heating data are presented in Table **¿tbl:step\_heating\_table?** and shown graphically in Figure **¿fig:step\_heating?**. Our preferred age of 1.65±0.06 Ma is defined by the twelve intermediate out of fifteen heating steps. The entire spectrum defines a similar age, within error, of 1.71 Ma. From these ages we infer that the xenoliths were entrained from the upper mantle directly underlying the Crystal Knob volcanic pipe during the mid-Pleistocene, at ca. 1.65 Ma.

## Petrographic and analytical methods

Polished thin sections of 250 µm thickness were prepared for six peridotite xenolith samples (CK-2 through CK-7) and the basalt host lava (CK-1). The xenolith samples were bound with epoxy prior to sectioning. Large-format rectangular thin sections were prepared for two host basalt samples dominated by dunite cumulate fragments (CK-D1 and CK-D2). The samples were evaluated under a petrographic microscope to determine their textural and mineralogic features. Characteristic textures of the xenolith samples and basaltic host are shown in Figure **¿fig:microscope-images?**.

Major-element compositions were analyzed for each polished thin section on a five-spectrometer JEOL JXA-8200 electron-probe microanalyzer at the California Institute of Technology. Abundances were counted in wavelength-dispersive mode using a 15 kV accelerating potential, a focused 25 nA beam, and counting times of 20 seconds on-peak and 10 seconds off-peak. The instrument was calibrated using natural and synthetic standards; matrix corrections were made using the CITZAF [*Armstrong*, 1988] algorithm. 1714 measurements were performed across the six peridotite samples, concentrated in 3-4 locations of interest per sample. Areas with orthopyroxene and clinopyroxene in contact were emphasized to aid in thermometry. 403 measurements of basaltic host and entrained dunites were also taken.

Electron backscatter intensity images of each thin section were collected using a ZEISS 1550 VP field emission SEM at the California Institute of Technology. These were coregistered with optical scans and electron-microprobe analyses. Minerals were automatically classified from microprobe data using a nearest-neighbor fitting algorithm between pure endmember phases. Poor-quality measurements with low totals were automatically flagged using a scheme based on that of *Taylor* [1998], and mixed phases along grain boundaries were discarded on a case-by-case basis. The resulting classification was checked for consistency with optical and backscatter imagery.

Isotope and trace-element geochemical techniques were applied for the harzburgite and lherzolite samples in the dataset, and are discussed throughout Section 2.4.

## The basaltic host

The Crystal Knob host rock (sample CK-1) is an alkali basalt with sparse vesicles and abundant feldspar, clinopyroxene, and olivine phenocrysts. The sample also contains dunite and multiphase peridotite fragments ranging within the section from aggregates of a few grains to ~5 cm xenoliths.

The groundmass is dominated by altered glass and microcrystalline plagioclase lathes. Though dominantly black, it is mottled with slightly greenish grey color domains at ~500 µm scale, which likely correspond to different levels of alteration. These domains are cross-cut by elongate narrow (~1 mm) flow shear bands of finer-grained material with sparse vesicles and phenocrysts.

Petrographic study of cumulate fragments within the Crystal Knob basalt reveals residues from multiple stages of melt fractionation. Samples CK-D1 and CK-D2 (dunite-containing thin sections from CK-1) are dominated by dunite and peridotite fragments up to 2 cm in diameter. The peridotite fragments show textures similar to those in samples CK-2 to CK-7. The dunites fragments are cumulate textured and finer-grained than the peridotites, with characteristic grain sizes ranging from 50--200 µm between fragments. Intergranular melt pockets are evident between olivine grains. In some samples, dunite encases peridotite fragments containing large (up to 2 mm) grains of olivine, pyroxene, and spinel. CK-D1 notably contains a large spinel grain with a pitted rim embedded in dunite. The dunite cumulates in these samples are texturally representative of the vast majority of xenoliths in the Crystal Knob basalt.

The basalt groundmass contains abundant phenocrysts of olivine and clinopyroxene, and microphenocrysts of potassium and plagioclase feldspar. A large population of clinopyroxene phenocrysts show an average Mg# of ~76 (Mg# defined as molar Mg/(Mg+Fe) 100). A separate population of high-Mg# measurements (up to Mg# 91) are hosted in xenolith fragments and relict cores of single grains. Zoned clinopyroxene grains show a stepwise crystallization history in a progressively evolving magma. The grain highlighted in Figure **¿fig:cpx\_profile?** has a large core (likely an overgrown xenocryst) with Mg# ~90, but material with Mg# < 75 occurs only in the outermost 30 µm of the grain. This suggests that the last phase of the magma's fractionation occurred relatively, likely during its ascent and cooling.

Major-element analysis of olivine "phenocrysts" in the host lava shows a well-sampled trendline from compositions comparable to fertile peridotite xenoliths (Mg# 89) through progressively lower Mg# cumulate grains, with cumulate aggregates clustered at Mg#=68 [Figure **¿fig:major\_elements?**‌a]. Thus, olivine grains in the host lava range from true phenocrysts to xenocrysts corresponding to entrained peridotites, with cumulates from various intermediate stages of magma evolution in between.

## Peridotite xenoliths

The peridotite samples (CK-2 through CK-7) are texturally classified using the scheme of *Pike and Schwarzman* [1977]. All samples display an allotriomorphic granular texture with anisotropy largely absent. Minor plastic deformation features are petrographically observed in most samples, including slight kink bands in some olivines. However, the parallel nature of the domain boundaries and minor (~1º) angular offsets of crystal axes in most cases indicates that these were not formed under significant strain. Samples CK-2 and CK-5 exhibit a weak shape-preferred alignment in elongate spinels [Figure **¿fig:textures?**].

Minor late-stage alteration products are seen in all peridotite samples. These include variably Fe-rich grain boundaries of major phases and Ti enrichment in pyroxene rims (<10 µm from the grain edge). Sample CK-4 contains an alteration channel that cuts linearly across the thin section. This channel is bounded by resorbed edges of the major phases (olivine and orthopyroxene) and hosts microcrystalline clinopyroxene, 10 µm-scale euhedral spinels, and minor amphibole. Near this melt channel, thin streamers of intergranular fill show compositions enriched in Na and Ti. These fills are present (but less extensive) in sample CK-3, and generally absent in other samples. Samples other than CK-3 and CK-4 largely do not show melt infiltration along grain boundaries, instead having major phases in contact (though many grain boundaries are fractured). Clinopyroxene grains in all samples, but most notably CK-3 and CK-4, are strongly intergrown with orthopyroxene. Sample CK-7 shows minor exsolution lamellae of orthopyroxene and clinopyroxene. Peridotite fragments in CK-D1 show abundant exsolution lamellae Figure **¿fig:microscope-images?**|c.

Sample CK-6, the most fertile in the sample set, shows abundant pyroxene exsolution manifesting as small-scale vermicular texture. Fused pyroxenes with substantially different axial directions within the same crystal indicate recrystallization and aggregation. These features are seen only in CK-6.

### Compositions of dominant phases

Major-element abundances for the peridotite xenoliths were measured by electron microprobe using methods discussed in Section 2.2. Results are summarized in Table **¿tbl:composition?**‌a and Figure **¿fig:major\_elements?**. Generally, phase compositions show tight per-sample groupings, which suggest equilibrium within each sample. The major silicate phases show Mg# > 87, consistent with fertile or residual mantle compositions. Variations in Mg# between samples indicate differences in melt-extraction and refertilization history between the samples.

Samples CK-2, CK-5, and CK-7 cluster tightly in Fe-Mg space, with relatively low Mg#s indicative of fertile compositions similar to that of the average depleted mantle [*Workman and Hart*, 2005]. Silicate phases in CK-3 and CK-4 have Mg# > 90, suggesting a residual composition. Sample CK-6 has low Mg#s for all major phases, with values as low as 87 for clinopyroxenes. All phases in this sample show higher iron abundance than expected for fertile peridotites.

Spinel compositions give slightly more information on the modes of variation between samples. For accuracy, we correct spinel Mg# from total iron to ferrous iron basis using charge balance on a 4-oxygen basis. This correction results in spinel Mg# between 75 and 81, slightly higher than the uncorrected value. Results are shown in Table **¿tbl:spinel\_correction?** and Figure **¿fig:major\_elements?**‌c. Some samples, particularly CK-2, have scatter in corrected Mg# due to unmeasured transition metals (e.g. Zn, Co, V) that are common in oxide minerals. CK-3 and CK-4 show high Cr# (molar Cr/(Cr+Al) 100) indicative of melt extraction, while the fertile samples show low Cr#s. Sample CK-6 has an intermediate spinel Cr# that suggests a signature of depletion in tandem with abundant iron that indicates refertilization [Figure **¿fig:major\_elements?**‌a].

All samples except CK-6 contain consistently fertile or somewhat depleted phases. CK-6 shows a clear anomaly, with high-iron phases that suggest fertility and intermediate spinel Cr# suggesting a history of depletion. This suggests that its highly fertile composition was gained by assimilation of a fractionated, Fe-rich melt into a partially depleted residue. In this case, the relatively chromian spinels represent a pre-refertilization vestige.

### Modal mineralogy

The peridotite samples range from lherzolites to clinopyroxene harzburgites and are dominated by olivine and orthopyroxene. All samples contain minor (<1%wt) spinel. Grains have a characteristic scale of ~200 µm for the lherzolites, and 500 µm for the harzburgites. Sample CK-3 contains 2 mm orthopyroxene porphyroblasts.

The samples are lithologically classified using recalculated whole-rock mineral modes. Mineralogy was classified on a ~5000 pixel grid atop coregistered optical scans and electron backscatter mosaics [Figure **¿fig:textures?**]. Volumetric modes were converted to %wt using representative densities for spinel-facies peridotite phases [*Nesse*, 2000]. Results are summarized in Table **¿tbl:modal\_mineralogy?** and shown relative to abyssal peridotites in Figure **¿fig:modes?**.

CK-2 has the most fertile composition, with 12.2%wt clinopyroxene. CK-3 is the least fertile sample, with 0.91%wt clinopyroxene. All samples are Type I peridotites in the *Frey and Prinz* [1978] classification system. Broadly, phase abundances track with the mineral phase chemistry for each sample. The samples with fertile phases (CK-2, CK-5, and CK-7) show a range in olivine abundance from 65 to 75%wt, suggesting different levels of depletion without adjustment in phase equilibrium compositions. CK-3 and CK-4 have <2% clinopyroxene, and their depleted phase compositions indicate almost complete removal of incompatible lithophile elements.

### Whole-rock composition

Whole-rock major-element abundances are reconstructed from averaged mineral composition and estimated modes. Representative mineral compositions are given in Table **¿tbl:composition?**‌a, and recalculated whole-rock compositions in Table **¿tbl:composition?**‌c. Whole-rock Mg# ranges from 87 to 91. Within each sample, a consistent Mg# for all silicate phases [Figure **¿fig:major\_elements?**] is indicative of Fe-Mg equilibrium. All samples contain <1%wt spinel, with Mg# variation mirroring that of the silicate phases. A range of spinel Cr# from 10 to 27 implies variation in degree of partial melting between samples, with higher-Cr spinels in residues of higher-degree melting [*Dick and Bullen*, 1984]. We explore this further in Section 2.6.1.

Samples CK-2, CK-5, and CK-7 have Mg# between 89 and 90 (both for individual silicate phases and reconstructed whole-rock measurements). CK-3 and CK-4 have a whole-rock Mg# greater than 90. CK-6 has a whole-rock Mg# < 88 and contains substantially more Cr, Al, and Fe than the other samples. Though CK-6 is generally the most enriched in incompatible elements, sample CK-2 contains somewhat more Ca and Na. Both of these samples have fertile major element compositions.

The coincident low Mg# and spinel Cr# of CK-2, CK-5, and CK-7 implies that they are relatively fertile peridotites, with low apparent melt volumes extracted (net of possible refertilization). Within these samples, variation in whole-rock composition corresponds to the modal abundance of pyroxene phases, with sample CK-2 containing substantially more pyroxene than CK-5. The high Mg# and harzburgite classification of samples CK-3 and CK-4 cements their status as residues of high-degree partial melting. For sample CK-6, the combination of high spinel Cr# (implying depletion by partial melting) and low whole-rock Mg# (a marker of major-element fertility) suggests that this sample experienced a multistage history of depletion and re-enrichment in major elements. Assimilation of a more evolved melt might explain its higher iron content and pyroxene modes, with excess pyroxene forming due to the addition of silica to an olivine-rich residual assemblage.

## Rb-Sr and Sm-Nd isotopes

Portions of each peridotite sample were crushed using a disk mill at the California Institute of Technology. Clinopyroxene grains (150--300 µm, 35-45 mg per sample and free of visible inclusions and alteration) were picked by hand under a binocular microscope. These clinopyroxene separates were analyzed for strontium and neodymium isotopes at the University of Arizona, Tuscon following the procedures described in *Otamendi et al.* [2009] and *Drew et al.* [2009]. The samples were spiked with mixed Sm-Nd tracers [*Ducea and Saleeby*, 1998a; *Wasserburg et al.*, 1981]. Rb was measured on a quadrupole ICP-MS, while Sr was measured in static multicollector mode on a VG 54 instrument. Sm was analyzed using a static routine on a VG Sector 54 multicollector thermal-ionization mass spectrometer, and Nd was measured as an oxide on a multicollector VG 354 instrument. Results are presented in Table **¿tbl:isotopes\_table?**.

Rb-Sr and Sm-Nd radiogenic isotope data for clinopyroxene separates show these samples to be derived from the depleted convecting mantle. All xenolith samples are enriched in radiogenic ( from 10.3 to 11.0) and depleted in (/ of .702). With respect to the central California coast, values here are well below those of 0.708 recorded by Salinian granites [e.g. *Kistler and Champion*, 2001], suggesting that the mantle lithosphere sampled by Crystal Knob is sourced from a different mantle reservoir than the overlying crust. More broadly, this pattern of strong depletion in large-ion-lithophile elements rules out an origin in the western North America mantle lithosphere or the Mesozoic mantle wedge beneath western North America [*Ducea and Saleeby*, 1998b; *Luffi et al.*, 2009; *Wilshire et al.*, 1988] and suggests an origin in the asthenospheric or underplated oceanic mantle [*DePaolo and Wasserburg*, 1976; *McCulloch and Wasserburg*, 1978].

## Trace Elements

Trace element concentrations were acquired for pyroxene grains in each xenolith sample, using a Cameca IMS-7f-GEO magnetic-sector secondary ion mass spectrometer (SIMS) at the California Institute of Technology. Two to three each of orthopyroxene and clinopyroxene grains were targeted per xenolith sample. Measurements were acquired with 9 kV beam flux and a 100 µm spot size. The USGS glass standard NIST 610 was used as an external standard for all elements [*Gao et al.*, 2002]. Minimal variation in measured concentration was observed at grain and sample scale, though clinopyroxene in CK-6 and orthopyroxene in CK-7 show differences outside of analytical error in Ba, La, and Ce (potentially attributable to concentrations near SIMS detection limits). Other measurements are largely concordant and results are presented as within-sample averages. Whole-rock trace element abundances are estimated using measured concentrations in clinopyroxene and orthopyroxene and mineral modes. Olivine is excluded from calculations, which is of minimal impact as rare-earth elements (REEs) are 2-3 orders of magnitude less compatible than in clinopyroxene [*Luffi et al.*, 2009; *Witt-Eickschen and O’Neill*, 2005]. Results for measured pyroxene and recalculated whole-rock trace elements are shown in Table **¿tbl:trace\_elements\_table?**. Graphical results for pyroxene REEs are shown in Figure **¿fig:trace\_elements?**, and recalculated whole-rock concentrations are shown in Figure **¿fig:ree\_model?**.

Clinopyroxene, orthopyroxene, and recalculated whole-rock rare-earth elements show several types of variation between samples, corresponding to different amounts of depletion and re-enrichment. All samples show clear evidence of rare-earth element depletion relative to depleted MORB mantle. For clinopyroxenes, the samples show a range of depletion in rare-earth elements following patterns seen in abyssal peridotites in HREE, with variably elevated LREE content presumably due to re-enrichment in the presence of more evolved melt [Figure **¿fig:trace\_elements?**].

### Modeling depletion and re-enrichment

Clinopyroxene REE patterns tell only part of the story of whole-rock depletion and re-enrichment. Since modal pyroxene abundances have changed significantly in response to depletion, recalculated whole-rock trace elements are a valuable tool to assess the overall level of depletion in the sample. Sample CK-5 in particular is much more depleted in whole-rock composition than its clinopyroxene trace-element composition would suggest, due to low clinopyroxene modes.

We construct a melt-partitioning model to determine the depletion and re-enrichment history of the Crystal Knob sample set and distinguish between potential enriching agents [Figure **¿fig:ree\_model?**‌a]. A generic model of peridotite depletion is constructed in *alphaMELTS* [*Smith and Asimow*, 2005], in which a parcel of material starting at a mantle potential temperature of 1350ºC at 3.0 GPa and a depleted MORB mantle composition [*Workman and Hart*, 2005] is tracked along an isentropic fractional melting path with a melt porosity of 1%, using the pMELTS family of algorithms. Trace-element partitioning coefficients from *Lee et al.* [2007] are used to track these components, and we confirm that garnet is removed from the evolving system at ~2.0 GPa. These starting parameters were chosen to provide the best correspondence with the overall experimental dataset. A wide range of initial conditions (up to 1500ºC at 3.0 GPa) provide similar results for the shape of depleted trace-element profiles, though these results do not map to unique temperatures.

In the first model step, the Crystal Knob xenolith samples are fit to model steps along this adiabatic path using minimization of the squared deviations of measured values from model HREE (Er--Lu) compositions. HREE concentrations can reasonably be assumed to be a proxy for depletion, since HREEs are highly compatible and have low diffusion rates, and are not easily modified by late re-enrichment. Thus, the best-fitting decompression step gives an model composition for the samples after single-stage depletion. The measured composition of sample CK-2 matches this modeled composition across the REEs, while other samples have much higher LREE concentrations than is reasonable for residues of decompression melting.

For the depleted samples (all except CK-2), a second step is taken to find the trace-element pattern of a potential enriching agent. The fitted depleted profile is subtracted from the sample composition, and excess measured LREE is interpreted as the contribution from batch addition of an enriching melt. To compare against various potential enriching agents, the model is normalized to an average HREE of 6primitive mantle, a typical HREE concentration for both normal MORB [*Sun and McDonough*, 1989] and alkali basalt [*Farmer et al.*, 1995]. The slope of the resulting normalized profile is diagnostic of the relative abundance of trace elements in the re-enriching agent. The normalization factor employed to shift the composition of enriching fluids to this value is shown in Figure **¿fig:ree\_model?**‌b and is proportional to the relative amount of material added during re-enrichment (assuming a consistent composition).

The results of this model confirm that all samples except CK-2 are variably depleted and re-enriched to some extent, while CK-2 shows neither depletion nor re-enrichment [Figure **¿fig:ree\_model?**]. The re-enriched samples have assimilated varying amounts of an enriching agent with a trace-element pattern similar to alkali basalt. The pronounced LREE enrichment of samples CK-3 and CK-4 is due to the lack of REE and low pyroxene modes in the sample, so assimilation of small masses of enriching melt can greatly increase normalized abundances. It is likely that all samples experienced a small (<1% assimilation) interaction with a fractionated basaltic melt [Figure **¿fig:ree\_model?**‌b]. However, if HREEs have been substantially enriched (as seems to be the case for sample CK-6) excess LREE is a minimum constraint on re-enrichment.

Primary depletion degrees of the xenolith samples are estimated by finding the pMELTS model compositions that best fit the whole-rock HREE, MgO, and composition of each sample. Results are summarized in Table **¿tbl:depletion\_degrees?** and show trends superficially similar to those visible in modal abundance [Figure **¿fig:modes?**] and trace element [Figure **¿fig:trace\_elements?**] data. The degree of depletion generally increases with the modeled temperature of the sample, with the notable exception of sample CK-6, which is by far the least-depleted sample by major-element proxies and only moderately depleted in heavy rare-earth elements, although it has the hottest modeled temperature.

### Discussion of trace elements

The trace element dataset developed here suggests that the samples were variably depleted of REEs following progressive fractional melting [*Johnson et al.*, 1990]. All of the samples underwent wholesale REE depletion (due to higher-degree melting) followed by later LREE re-enrichment. Our modeling suggests a similar degree of LREE re-enrichment for all samples, with less than 1% melt assimilation in all cases (assuming a somewhat HREE-enriched melt similar to MORB or alkali basalt). However, the more depleted samples having the lowest clinopyroxene modes show pronounced re-enrichment in clinopyroxene REEs, which likely arises from rare-earth inputs being primarily focused in clinopyroxene. Still the patterns of re-enrichment found in these samples are not found in abyssal peridotites [*Warren*, 2016], suggesting that this pattern arose from secondary re-enrichment by LREE-rich material at depth.

The overall trace element pattern suggests that the xenolith samples are residues of progressive fractional melting of primitive mantle at the mid-ocean ridge, a single-stage process similar to that recorded in abyssal peridotites [*Johnson et al.*, 1990; *Warren*, 2016]. The samples underwent a multistage history of REE depletion followed by later LREE re-enrichment by an enriched melt. Trace element patterns suggest that the Crystal Knob samples record a range of fertility, with sample CK-2 essentially undepleted. The alignment of HREE fertility with patterns expressed in phase composition and modal abundance suggests that HREE depletion is primary, without significant re-enrichment. All samples show evidence of a distinct but volumetrically minor assimilation of LREE-enriched melt. Sample CK-6 is an exception to this alignment, with high LREEs in excess of those formed at any stage of fractional melting. This, along with anomalous phase compositions, suggests that this sample assimilated a significant amount of more-evolved melt, and its trace-element profile (including HREE composition) is formed by refertilization.

## Major-element thermometry

Electron-microprobe major-element data is used as the basis for pyroxene Ca--exchange geothermometry. Several formulations of this reaction are tested: BKN [*Brey and Köhler*, 1990] and TA98 [*Taylor*, 1998] are two slightly different formulations based on empirical calibration of the two-pyroxene Ca exchange reaction in simple and natural systems. *Taylor* [1998] is explicitly calibrated to account for errors arising from high Na content. The Ca-in-orthopyroxene (Ca-OPX) thermometer [*Brey and Köhler*, 1990] is formulated for use in the absence of clinopyroxene. Together, these thermometers can query the full range of major-element compositions seen in the Crystal Knob xenolith samples. Results are shown in Table **¿tbl:thermometry?** and Figure **¿fig:temp\_comparisons?**.

### Error in thermometer calibrations

Core and rim compositions are separated to assess within-sample temperature disequilibrium and late-stage (e.g. eruptive) heating. Analytical errors (caused by uncertainty in microprobe data) are small, on the order of 5ºC (1). Other sources of error include the calibration of the thermometer and potential bias from within-sample disequilibrium. *Taylor* [1998] reports residuals of calibration of the thermometer to experimental data which yield total errors of 50-60ºC (1). Unreported calibration errors for the BKN and Ca-OPX thermometers are likely similar in scale [*Taylor*, 1998]. In practice, error distributions based on calibration with heterogeneous experimental samples likely form an upper bound on relative errors. Within-sample scatter in measured temperatures can be used to estimate the relative error of recovered temperatures, and the relative performance of different thermometers can be used to infer absolute temperatures.

Within-sample temperature variation can be useful in assessing the potential errors in calculated temperatures. The dataset of pyroxene composition measurements is grouped by location for thermometry, with a separate temperature calculated for each individual nearest-neighbor pair of orthopyroxene and clinopyroxene. Analytical errors are propagated through the calculation. The resulting distribution of temperatures for grain cores and rims for each sample (with *n* ranging from 19 to 74 pairs per group) accounts for within-sample variation and provides an approximation of measurement precision.

TA98 and BKN temperatures have a strong linear relationship, with BKN temperature estimates higher by up to 50ºC. The disparity decreases towards higher temperatures and conforms to the relationship between the two thermometers found by *Nimis and Grütter* [2010]. This relationship can be expressed as [ºC]. *Nimis and Grütter* [2010] shows that TA98 performs well against experimental results in several scenarios and advises its use over BKN. The Ca-in-OPX thermometer sensibly yields results coincident with BKN temperatures (as they were calibrated from the same dataset) [*Brey and Köhler*, 1990]. The low within-sample scatter of the Ca-in-OPX thermometer possibly results from the fast diffusion and complete re-equilibration of small amounts of Ca in orthopyroxene, or of the stability of relatively refractory orthopyroxene against late magmatic modification.

### Core temperatures

Average TA98 temperatures range from 957 to 1063ºC for cores and 955 to 1054ºC for rims [Table **¿tbl:thermometry?**]. CK-2 core temperatures indicate more complete equilibration, with a standard deviation of only 2.3ºC (compared with 8.2-12.4ºC for all other samples). Temperatures are distributed roughly normally for most samples, but outlying clusters in CK-4 and CK-6 may indicate two-pyroxene major element disequilibrium at millimeter scale. In CK-4, a few grain cores with TA98 temperatures of 1100ºC are likely related to late-stage diffusion during entrainment and eruption. Within-sample temperature variability is a relatively minor feature: for grain cores in all samples, the mean of pairwise analyses is within a few degrees of the temperature calculated for average compositions of pyroxene phases across the sample. This implies that measured temperatures are representative of equilibrium.

### Rim temperatures

Rim temperatures (measured ~10 µm from grain edges) are generally higher than core temperatures, although the level of disparity varies widely between samples. CK-2 shows only modestly elevated rim temperatures, while CK-3 and CK-6 show significant scatter to temperatures ~180 ºC higher than grain cores (CK-5 and CK-7 contain a few measurements of this type as well). High and variable rim compositions may be related to melt infiltration during entrainment and eruption of the xenoliths, but with significant mobilization of cations limited to grain rims. CK-4, which shows high core temperatures and the most significant petrographic evidence of melt infiltration, lacks high-temperature rim compositions, suggesting equilibration at high temperature, possibly in the Crystal Knob melt.

### Two temperature cohorts

The samples can be divided into two clear cohorts based on equilibration temperatures. A cooler group of samples, with a distribution of grain core temperatures centered at ~970ºC (TA98), contains CK-2, CK-5, and CK-7. A hotter group, with a mean temperature of ~1050ºC (TA98) contains samples CK-3, CK-4, and CK-6. This division between these two groups is robust and apparent in all thermometers, and aligns with geochemical differences within the sample set, with the hotter samples having higher spinel Cr# and greater HREE depletion. The range of temperatures likely records the sourcing of two sets of xenoliths from different depths within a magmatic ascent system. Throughout this paper, the samples are color-coded, with blue-green corresponding to the low-temperature array, and red-yellow representing the high-temperature samples [Figure **¿fig:major\_elements?**‌-‌Figure **¿fig:depth?**].

## REE-in-pyroxene thermometry

We use the *Liang et al.* [2013] REE-in-two-pyroxene thermometer to estimate the equilibration temperature of the samples using an independent system. The relative immobility of REEs allows assessment of equilibrium temperatures over longer timescales than those queried with two-pyroxene cation exchange thermometry.

Rare-earth abundances are compiled for SIMS measurements of pyroxene phases in contact (2--3 pairs) for each xenolith sample. A two-pyroxene equilibrium is calculated for each REE and Y, and is shown in Figure **¿fig:ree\_temperatures?**. A robust regression with a Tukey biweight norm is used to find the equilibrium temperature for each sample. Significant outliers from the fit are excluded from thermometry, and may represent effects of disequilibrium processes.

### Pyroxene rare-earth disequilibrium

CK-3 shows misfit in La only, while CK-5 and CK-7 have disequilibrium in several of the LREEs. Sample CK-4 shows major disequilibrium in the light and medium REEs, with only elements heavier than Ho retaining a linear relationship.

The pattern of disequilibrium in sample CK-4 suggests that the partition coefficient for the LREEs is larger than anticipated relative to that for HREEs. This is perhaps due to low clinopyroxene modes in the harzburgite CK-4, forcing assimilated REEs into orthopyroxene. The shape of this disequilibrium may be due to the parabolic nature of mineral-melt partition curves for both pyroxene phases, which are incompletely modeled by a linear relationship when offset [*Blundy and Wood*, 2003; *Sun and Liang*, 2012]. Alternatively, low clinopyroxene modes in the harzburgite CK-4 could result in the incorporation of more REEs into orthopyroxene than is predicted by the thermometer (Figure **¿fig:trace\_elements?** shows pronounced LREE enrichment in orthopyroxenes for CK-3 and CK-4). Regardless, since measured LREEs are consistently enriched, with low errors for both orthopyroxene and clinopyroxene, we conclude that anomalies in LREE temperature do not signal disequilibrium compositions.

All samples except CK-6 and CK-7 show results off the linear trendline for Eu. This distinct disequilibrium was discussed *Sun and Liang* [2012], and is dependent on the oxygen fugacity (and / ratio) of the host magma. In normal magmas, this patterns could also arise from the effect of "ghost" plagioclase, which could create local europium enrichments and depletions in the neighborhood of resorbed plagioclase grains. This would suggest that the xenoliths originated at shallow mantle lithosphere levels and were transported deeper, causing plagioclase breakdown. Discerning between these scenarios is difficult due to the low Eu counts measured using ion-microprobe techniques.

General LREE disequilibrium between pyroxenes can be explained by a fossil heating event that was retained only in REEs due to their slow diffusion rates. This must have happened prior to subsolidus re-equilibration in major elements, which implies that it is not related to the eruptive episode. Overall, disequilibrium patterns in REE between pyroxenes allude to possible focused heating of sample CK-4, poorly understood equilibrium partition coefficients (for instance, due to reducing mantle conditions), and incomplete linearizing assumptions in the *Liang et al.* [2013] thermometer. Untangling these effects is beyond the scope of this work but presents several opportunities for further study. Even for samples with pronounced LREE disequilibrium, temperature estimates anchored by HREE perform well against major-element thermometry.

### Comparison with major-element thermometry

Rare-earth exchange thermometry shows the samples as divided into the same two temperature groupings as those found by major-element thermometry. REE temperatures measured for the low-temperature cohort are most comparable to the TA98 results [Figure **¿fig:ree\_temperatures?**]. Given that the TA98 method has been found to perform best among the pyroxene-exchange thermometers by *Nimis and Grütter* [2010], both the TA98 and REE temperatures likely show long-term equilibrium with no significant thermal perturbations. For the high-temperature cohort, particularly samples CK-4 and CK-6, the REE method shows significantly higher equilibration temperatures than TA98. Trivalent rare-earths in pyroxene diffuse several orders of magnitude slower than bivalent major elements [*Liang et al.*, 2013], so early thermal events can leave an imprint on the distribution of rare-earths for much longer than with major elements. It is likely that these higher temperatures are a signature of a fossil high-temperature event primarily affecting the deepest samples.

Sample CK-4 records a significantly higher temperature for individual LREEs than both the other samples and its own HREE equilibration temperatures. This likely shows that LREEs were equilibrated at a much higher temperature than HREEs. This pronounced within-sample disequilibrium could be the result of metasomatic processes, which is bolstered by the fact that CK-4 shows the only significant melt-infiltration textures (intergranular melt channels) in the sample set [Figure **¿fig:textures?**]. CK-4 was likely subjected to a transient heating event that was not fully equilibrated in HREEs. Sample CK-6 has a high-temperature signature across the entire range of rare-earth elements, while CK-3 records a much lower temperature in line with major-element thermometry. Excess heating recorded by REE thermometry for CK-4 and CK-6 is not reflected in major-element temperatures, which show that the sample reached a lower-temperature equilibrium after this heating event.

## Depth constraints

Pyroxene-exchange geothermometry shows that the peridotite samples form two groups in temperature with centroids separated by roughly 60ºC. This temperature range likely corresponds to an array of sample sources along a depth gradient. The depth of the xenolith samples in the mantle lithosphere, coupled with equilibration temperatures, provides a fully-defined constraint on the geotherm beneath Crystal Knob at the time of eruption. For spinel peridotites, equilibration depths can only be analytically estimated within broad boundaries. With no reliable geobarometers for spinel peridotites, several less robust metrics are used to evaluate the depth of the xenolith source. We present several lines of reasoning suggesting that the xenoliths were sourced along a depth gradient relatively deep within the spinel stability field, between roughly 45 and 75 km.

Several of the techniques below produce estimates of pressure, rather than depth. To discuss these data in depth--temperature space, we correct them using a hydrostatic gradient of ~0.03 GPa per km across the mantle lithosphere, based on integration of the crustal and mantle densities given in Table **¿tbl:model\_parameters?**.

### Limits of spinel stability

Entrainment depths of all peridotite xenoliths must be greater than ~30 km, the depth of the Moho near the Crystal Knob eruption site [*Tréhu*, 1991], which will be discussed in more detail in Section 3.2. Another minimum depth constraint is the plagioclase--spinel peridotite facies transition, which occurs at depths of 20-30 km but is highly composition-dependent [*Borghini et al.*, 2009; *Green and Ringwood*, 1970], with high-Cr harzburgites stable to the surface.

The high-pressure boundary of spinel stability limits maximum possible entrainment depths. The spinel--garnet peridotite facies transition is composition-dependent and poorly constrained for natural systems, but thought to lie over the 50-80 km depth interval [*Gasparik*, 2000; *Kinzler*, 1997; *Klemme*, 2004; *O’Neill*, 1981]. The breakdown depth of spinel is strongly dependent on temperature and composition, particularly the amount of refractory Cr hosted by spinel. Several experimental and thermodynamic studies have attempted to estimate the magnitude of this effect. *O’Neill* [1981] presented experiments both with and without Cr and described a simple empirical relationship of spinel-out depth with Cr content and temperature. *Robinson and Wood* [1998] suggests that, given fertile "pyrolite" compositions with little Cr, garnet is unstable at depths less than 80 km at the peridotite solidus (~1470ºC at this depth). Subsolidus experimental results show that the maximum depth of the spinel stability field in the absence of Cr ranges from 1.8-2.0 GPa (55-60 km) at 1000-1200ºC [*Klemme and O’Neill*, 2000]. Chromian spinels can be stable to much greater depth: thermodynamic modeling by *Klemme* [2004] suggests a broad garnet-spinel co-stability field (up to a spinel-out reaction at 5 GPa for Cr# of ~30), but a spinel-weighted metastable assemblage is possible even at higher pressures.

As shown in Figure **¿fig:major\_elements?**‌c, samples in the high-temperature cohort (CK-3, CK-4, and CK-6) have higher spinel Cr# than the low-temperature samples. This enrichment in refractory Cr arises from the increased depletion of these samples and expands the stability field of spinel against garnet to deeper depths.

Though *Robinson and Wood* [1998], *Klemme and O’Neill* [2000], and *Klemme* [2004] show a high-pressure phase transition with a complex compositional dependence, the rough estimate of the garnet-in pressure given by *O’Neill* [1981] performs sufficiently well at T < 1200 ºC. This empirical relationship is used in Figure **¿fig:depth?** to graphically illustrate the phase-transition depths given the Cr# of each sample (with error bars of 0.15 GPa).

This simple treatment provides a high-pressure constraint on the Crystal Knob xenolith source. The maximum possible entrainment depths of the high-Cr samples increase by up to 15 km relative to Cr-free compositions, from ~65 km for the low-temperature samples to maximum depths of ~80 km for the high-temperature cohort.

### Ca-in-olivine barometer

Equilibration pressure measurements are attempted for the peridotite xenoliths using the *Köhler and Brey* [1990] Ca-in-olivine barometer, which is based on the decreasing abundance of Ca in olivine with pressure. This barometer is explicitly calibrated for spinel peridotites but should be treated with caution based on poor resolution, high temperature dependence,vulnerability to late-stage diffusion, and dependence on low Ca concentrations in olivine near analytical thresholds for electron microprobe analysis [*Medaris et al.*, 1999; *O’Reilly*, 1997].

To model the variability of model pressures due to analytical uncertainties, barometry is applied separately for nearest-neighbor pyroxene and olivine measurements. Analytical errors are propagated through the calculation. To correct for the mild pressure dependence of the two-pyroxene thermometer, and the temperature dependence of the olivine barometer, we jointly solve temperature and pressure by iteratively optimizing to a common solution for each set of microprobe measurements. This yields a set of separate internally consistent depth and temperature measurements for each sample corresponding to individual pairs of microprobe measurements. In Figure **¿fig:depth?**, we show the full pressure--temperature error space for each sample by applying a Monte Carlo random sampling () to the analytical errors on each pressure estimate.

The Ca-in-olivine barometer yields a broad distribution in model depths, largely coincident with the spinel stability field [Figure **¿fig:depth?**]. The depth distributions are largely normal, with modes ranging from 40 to 53 km. Within the Crystal Knob sample set, the low and high-temperature cohorts remain separable, with high-temperature samples showing deeper equilibrium depths. The scale of the errors within a single sample reflects the barometer's strong covariance with major-element thermometers, as well as its sensitivity to small variations in Ca concentrations. The bulk of the spread in the data reflects the poor calibration of the barometer itself. The low-temperature samples in particular have significant scatter towards depths above the spinel-in isograd. The small Ca cation diffuses rapidly during transient heating [*Köhler and Brey*, 1990], which produces a shallowing bias on the depth distribution. This may explain why CK-4, the most altered sample, has a depth mode ~10 km shallower than the other samples (CK-3 and CK-6) with similar equilibration temperatures. The temperatures derived from the independent REE system are higher than the major-element temperatures for several samples in the high-temperature cohort (CK-4 and CK-6), which may point to these samples being derived from a greater depth within the distribution of Ca-in-olivine depths.

Despite the imprecision of the method, Ca-in-olivine barometry suggests that the samples were sourced from relatively deep within the spinel stability field, at depths of ~40 km or greater. This preference is amplified by comparisons with depth estimates of the thermal state of the mantle lithosphere derived from regional heat flow datasets.

### Comparisons with steady-state heat flow

The depth constraints derived from xenolith thermobarometry above can be compared to surface heat-flow and seismic constraints on the regional geotherm.

Given our high-confidence temperature measurements for the Crystal Knob xenolith suite, we can generate model entrainment depths by pinning the samples to a conductive geotherm constrained by surface heat flux. These can be useful for comparisons with our intrinsic depth constraints from thermobarometry. In Figure **¿fig:depth?**, we show potential steady-state geotherms for surface heat fluxes ranging from 60 to 120 mW/m^2, all of which intersect the potential depth distributions from spinel stability and Ca-in-olivine barometry.

These geothermal gradients are calculated using thermal conductivity and diffusivity given in Table **¿tbl:model\_parameters?** for the crust to a depth of 30 km, and mantle lithosphere below this level. No fixed amount of radiogenic heat production is assumed, but the average empirical factor of 0.6 proposed by *Pollack and Chapman* [1977] is used to reduce surface heat flux to a presumed mantle contribution, with the remainder being taken up by radiogenic heat production near the surface. We use a radiogenic contribution that exponentially decreases with depth with a characteristic length scale of 10 km. The amount of heat emanating from the mantle and the presumed thermal conductivity across the lithosphere are the main controls on the slope of the modeled geothermal gradient. This methodology is developed in *Turcotte and Schubert* [2002] and is identical to that used by *Luffi et al.* [2009], except that crustal thermal conductivity is reduced to match our conditions for dynamic thermal modeling [Section 4]. This yields a slightly "hotter" geotherm throughout the mantle lithosphere.

Phase stability constraints on the Crystal Knob xenoliths correspond to a broad range of plausible lithospheric conductive geotherms. The hottest potential geotherm keeping the sample set within the spinel stability field is > 120 mW/m^2 at the surface. Accounting for the Cr-dependent depth of the spinel--garnet transition, the 65 mW/m^2 conductive geotherm is the coolest that places all samples within the spinel stability field. The centroids of the Ca-in-olivine model depth distributions broadly correspond to surface heat flows ranging from ~70 to 110 mW/m^2.

Using a database of surface heat flows for North America, *Erkan and Blackwell* [2009] estimate regionally-averaged heat flows of 80-90 mW/m^2 for the central California coast, including the vicinity of Crystal Knob. Projection of the TA98 temperature distribution onto our calculated steady-state geotherms yields model depths of ~45-55 km for the Crystal Knob sample set. This depth range falls within the spinel stability field and near the center of the depth distributions extracted using Ca-in-olivine barometry. However, depths derived from surface heat flows may be underestimates, for reasons discussed in Section 3.1.

### Summary of depth constraints

The integration of depth constraints on the xenolith samples from multiple techniques gives a broad boundaries on the depth of origin of the Crystal Knob xenoliths within the mantle lithosphere at the time of their eruption. For reasonable slopes of the sub-Salinian geotherm, the range of temperatures in the sample set indicates sourcing over a depth range of 5-10 km within the mantle lithosphere. These depths must be greater than 30 km, the depth to the Moho, and less than 60-90 km based on the composition-dependent lower limit of the spinel stability field. Ca-in-olivine barometry suggests a tighter set of constraints near the center of the spinel stability field. Model depths of 45-55 km prepared from steady-state geotherms agree with this assessment, but might be underestimates. Given the bias in both Ca-in-olivine and heat-flow measurements towards shallower depths, we take the depth range of 45-75 km as a likely entrainment depth for the Crystal Knob xenoliths.

This assessment of relatively deep entrainment of the Crystal Knob xenoliths along a fairly "cool" geotherm conforms to constraints from independent studies. Estimates of the thermal state of the deep lithosphere derived from seismic tomography show temperatures of 700--1100ºC occurring at depths of 50--100 km for coastal California [*Goes and Lee*, 2002], corresponding to cooler geotherms than predicted by surface heat flow. More recent estimates put the depth of the lithosphere-asthenosphere boundary at roughly 70 km in the southern Coast Ranges [*Li et al.*, 2007]. Given estimates that the subcontinental lithosphere-asthenosphere boundary occurs at 1200-1300ºC [*Fischer et al.*, 2010; e.g. *O’Reilly and Griffin*, 2010], this corresponds to steady-state geotherms of 70-80 mW/m^2.

We next turn to the accuracy of this extrapolation from heat flow values, and its implication for understanding the thermal structure of the lithosphere.

## Geochemical variation with depth and temperature

The Crystal Knob peridotite xenoliths studied here are uniformly isotopically depleted, with an of +10.3-11.0, and / of .7023-.7024. These are closely related to values found in the depleted upper mantle [e.g. *Hofmann*, 1997] and require a mantle source that has seen no contribution from the western North American subcontinental mantle or continental lithosphere more generally. The mantle lithosphere sampled by the Crystal Knob suite matches slightly depleted convective mantle material and is unrelated to the overlying continental crust. However, the xenoliths can be divided both along chemical lines and by their equilibration temperature. The two temperature cohorts sample mantle material with somewhat different major- and trace-element characteristics. Integrating depth constraints derived from geothermal data and Ca-in-olivine geobarometry, we can note the pattern of xenolith chemistry with depth.

The low-temperature and shallower cohort includes phases with relatively fertile compositions in both major and trace elements. The samples have phase compositions similar to depleted MORB mantle [*Workman and Hart*, 2005], and clinopyroxene trace elements range from essentially undepleted to low levels of depletion characteristic of the least-depleted abyssal peridotites [Figure **¿fig:trace\_elements?**].

The high-temperature, deeper samples show two distinct sets of chemical characteristics. Samples CK-3 and CK-4 are characterized by phases with high Mg#s, suggesting that all phases lost incompatible elements during high-degree depletion. These features, along with high levels of REE depletion, extremely low clinopyroxene modes, and chromian spinels, indicate significant melt extraction. Distinct enrichments in LREE likely correspond to assimilation of a fractionated melt, though this is modeled to be volumetrically insignificant.

Sample CK-6, the hottest and deepest sample, has a highly fertile major-element composition. However, high modal abundance of pyroxenes and spinel, and simultaneously elevated Cr, Al, and LREE suggest both significant depletion and re-enrichment. This is bolstered by petrographic evidence of substantial intergrowth and aggregation of pyroxene grains, unique in the sample set. Elevated REE-in-two-pyroxene temperatures for samples CK-4 and CK-6 suggest a fossil heating event not shown in major-element thermometry. This event may have been more completely felt by sample CK-6, which has a fully equilibrated REE temperature measurement [Figure **¿fig:ree\_temperatures?**] and appears to have assimilated a sizable amount of enriched material.

Altogether, the pattern of increasing depletion with depth suggests increased melt extraction deeper in the lithospheric column, over a <10 km depth range (given reasonable geothermal gradients). However, variation in depletion over a small depth range (anchored by similar equilibration temperatures) suggest some heterogeneity to melting. This pattern is at odds with the expected signature of decompression melting, which implies monotonically increasing melting degrees at shallower depths. *Luffi et al.* [2009] found a similar inverted pattern within mantle lithosphere packages beneath Dish Hill, ascribing fertile lherzolites at the top of the package to refertilization of suboceanic mantle lithosphere.

Wholesale refertilization seems unlikely for the Crystal Knob sample set due to the consistency of phase abundances with major and trace-element chemistry within all samples except CK-6. The low-temperature samples are petrologically similar to abyssal peridotites [Figure **¿fig:modes?** and Figure **¿fig:trace\_elements?**], suggesting they could represent un-refertilized suboceanic mantle lithosphere. The depleted harzburgites CK-3 and CK-4 could be tectonically juxtaposed to deeper levels by duplexing [as in *Luffi et al.*, 2009] or depleted after emplacement. Along these lines, the obviously refertilized CK-6 at the base of the column clearly shows melt-rock interactions that were most intense at deep levels. This spatial pattern is commonly associated with flux melting in subarc settings [e.g. *Jean et al.*, 2010] but could also arise from intense heating from below [*Thorkelson and Breitsprecher*, 2005]. We now turn to the tectonic constraints on the structure of the mantle lithosphere beneath coastal southern California, which provide a framework in which to consider this deep lithospheric geochemical and thermal structure.

# Origin of the mantle lithosphere beneath Crystal Knob

Rb-Sr and Sm-Nd isotopic and trace-element compositions of peridotite xenoliths demonstrate that the mantle lithosphere that was sampled by the Crystal Knob volcanic neck is sourced from the depleted mantle with no contribution from either recycled crustal material or ancient sub-continental mantle lithosphere The lack of radiogenically enriched arc residues confirms the unrooted nature of Salinia continental arc rocks [e.g. *Hall and Saleeby*, 2013] that lie slivered above the Franciscan accretionary complex that dominates the regional crustal column.

The Franciscan complex of the region was assembled by sustained subduction of the Farallon plate during the Cretaceous and early Tertiary [*Chapman et al.*, 2016a; *Cowan*, 1978; *Saleeby et al.*, 1986; *Seton et al.*, 2012], and the mantle lithosphere of the region may have been constructed from partly subducted Farallon plate upper mantle late in this Franciscan accretionary history. Alternatively, another mechanism could have operated following the cessation of Farallon plate subduction and during the establishment of the San Andreas transform system. Based on the regional geologic history and constraints from crustal structure, we identify several viable tectonic scenarios for the formation of the mantle lithosphere beneath coastal central California, which we summarize in Section 3.3‌-‌Section 3.5.

## Thermal constraints on lithospheric structure and history

In this section, we summarize previous efforts to use surface heat flow to query the deep lithospheric structure of coastal California. Studies relying on surface heat flow form the basis of attempts to estimate the thermal structure and evolutionary history of the mantle lithosphere beneath the central California coast ranges. *Erkan and Blackwell* [2008] and *Erkan and Blackwell* [2009] studied heat flow data across the western U.S. based on borehole measurements of heat flux in wells > 100 m deep. These workers found a "Coast Range Thermal Anomaly" of high surface heat flux measurements in a broad swath across the entire Coast Range belt, both inboard and outboard of the San Andreas fault zone.

Heat flow in the Coast Ranges range from 60 to 90 mW/m^2, much higher than in the adjacent Central Valley and Sierra Nevada, but similar to that measured in the Mojave province. It is also on the relatively high end of the global distribution of regionally averaged continental heat flows, which ranges from lows of 20 mW/m^2 in cratonic cores to 120 mW/m^2 in focused areas of active mantle upwelling (e.g. the southern Salton Trough) [*Erkan and Blackwell*, 2009; *Pollack and Chapman*, 1977].

Factors such as shear heating, fluid circulation along faults, and rapid surface uplift have been proposed to account for high heat flows in the Coast Ranges [*Erkan and Blackwell*, 2009]. However, the extent to which these surface heat flows are representative of the true integrated geotherm over the entire lithospheric column is unclear. Evaluating the nature of these constraints on the actual thermal structure of the lithosphere is critical to understanding the tectonic history of the Coast Range belt.

Heat flow measurements in the Coast Ranges are likely to be maximum estimates of the regionally-averaged heat flow and geothermal gradient due to several factors including localized uplift, highly radiogenic crust near the surface (e.g. Salinian granodiorites and tonalites), and regional heat diffusion by fluid circulation within the San Andreas transform system. Recent increases in rock uplift and erosion such as those documented since 2 Ma in the Santa Lucia Range [*Ducea et al.*, 2003] could significantly raise the geothermal gradient in the Coast Ranges. Under any of these scenarios, it seems likely that the gradients given in *Erkan and Blackwell* [2009] are a maximum limit on the geothermal gradient in the mantle lithosphere.

The geothermal gradient may be affected by regional differences in thermal conductivity, which can be enhanced by hydrothermal circulation in the presence of fractures. *Lachenbruch and Sass* [1980] concluded that high heat flows in the Coast Ranges could not be generated by conductive fault heating alone, but were hesitant to invoke elevation of the regionally averaged thermal gradient by fluid flow. However, areas underlain by an intact batholithic crustal column (e.g. the Sierra Nevada the Peninsular Ranges) show relatively low heat flows. This suggests that fluid transport or localized upwelling along the San Andreas transform system may contribute significantly to anomalously high surface heat flow in the Coast Ranges. *Kennedy* [1997] shows the presence of mantle fluids in the San Andreas fault zone, suggesting that deep fluid transport may occur in the region. The eruption of Crystal Knob itself demonstrates that a fully conductive geotherm is unreasonable, as at least *some* heat was advected to the surface by this and other volcanic edifices, such as Coyote Lake [Figure **¿fig:context?**], since the Pleistocene.

In sum, estimates of the geothermal gradient derived from surface heat flow may overestimate the geothermal gradient at depth, especially in the presence of fluid flow and radiogenic rocks near the surface. Thus, extrapolations of the lithosphere-averaged geotherm from surface heat flow estimates likely produce minimum estimates of sample depth. Next, we outline several tectonic features that could affect the thermal structure of the California coastal region.

## Late Cenozoic tectonic history and regional crustal structure

In the Oligocene to early Miocene, the Pacific-Farallon spreading ridge obliquely intersected the SW Cordillera subduction zone, leading to the development of the San Andreas transform system [*Atwater*, 1970]. Ridge impingement was kinematically complex due to large-offset ridge-ridge transforms, resulting in the opening of a geometrically complex slab window and the production of the Monterey microplate, which nucleated as an oblique intra-oceanic rift along an ~250 km long segment of the Pacific-Farallon ridge [*Atwater and Stock*, 1998; *Bohannon and Parsons*, 1995; *Thorkelson and Taylor*, 1989; *Wilson et al.*, 2005]. Late Cenozoic volcanism of the coastal region of central California has been attributed to partial melting of asthenosphere as it ascended into the slab window [*Wilson et al.*, 2005].

Alternatively, it has been suggested that microplate formation along the impinging Pacific-Farallon ridge was more dominant than slab window upwelling, and that these microplates stalled beneath coastal central California as the Farallon plate continued to subduct deeper into the mantle [*Bohannon and Parsons*, 1995; *Brocher et al.*, 1999; *Van Wijk et al.*, 2001]. In this scenario, late Cenozoic volcanism in the region is linked to the youthfulness of the subducted microplate(s), implying an “upside down” partial melting mechanism within and immediately adjacent to the lithospheric lid. Both the slab window (or gap) and stalled microplate hypotheses are based on plate kinematic relationships which require a combination of slab window and stalled oceanic microplate segments [*Atwater and Stock*, 1998; *Bohannon and Parsons*, 1995; *Brink et al.*, 1999; *Wilson et al.*, 2005].

Seismic data cited in support of the stalled slab hypothesis consist of an 8-15 km thick, shallowly east-dipping mafic lower crustal layer that extends beneath central California from the offshore region into proximity of the San Andreas fault, thickening eastwards over Moho depths of ~12-30 km [*Brocher et al.*, 1999]. Strong internal reflectivity within this layer [*Brocher et al.*, 1999; *Trehu and Wheeler*, 1987], and sharp inflections in its upper surface [*Tréhu*, 1991], indicate that this mafic layer is internally deformed and imbricated, which accounts for its thickness exceeding typical oceanic mafic crust by a factor of two to three.

Such imbrication and underplating require a basal detachment, which is most logically located at the underlying Moho. In this context, the regions’s lower crustal mafic layer is more plausibly interpreted as a regional underplated duplex of Farallon plate oceanic crustal nappes that accreted during Franciscan subduction. The underlying mantle lithosphere could be underplated Farallon plate mantle, and/or Monterey microplate mantle with its crustal section left imbricated along the toe of the mafic duplex in the offshore region. The Crystal Knob xenolith suite is the only known direct sampling of this underplated mantle.

The pre-Neogene tectonic setting of the Crystal Knob eruption site is shown in Figure **¿fig:reconstruction?** by restoration of the San Andreas dextral transform system [*Chapman et al.*, 2012; *Dickinson et al.*, 2005; *Hall and Saleeby*, 2013; *Matthews*, 1976; *Sharman et al.*, 2013]. The Crystal Knob eruption site restores to a position outboard of the southern California batholith. The principal windows into shallowly underplated subduction channel schists are shown in Figure **¿fig:reconstruction?** along with the principal upper plate batholithic exposures. The current western extent of the Salinia crystalline nappes is shown as the Nacimiento fault and the offshore Farallon escarpment. Crystalline rocks of the Salinia nappes extended an unknown distance westwards across the Nacimiento belt Franciscan [*Hall and Saleeby*, 2013], but were eroded off the lower plate complex as the coastal region rose in the Pliocene [*Ducea*, 2003].

The cross-sections presented in Figure **¿fig:neogene\_sections?** show the first-order crustal relations that are implied by three potential origins for the sub-Crystal Knob mantle lithosphere: **A**. shallowly ascended asthenosphere within the Pacific-Farallon slab window [*Atwater and Stock*, 1998]; **B**. subduction of an underplated, or stalled, Monterey oceanic microplate [*Bohannon and Parsons*, 1995]; or **C**. underplated Farallon plate mantle lithosphere nappe(s) that lie in structural sequence with the upper mantle duplex resolved beneath the Dish Hill xenolith location in the Mojave region [*Luffi et al.*, 2009]. These scenarios have been posed by a number of workers as detailed below.

## The Neogene slab window

Figure **¿fig:reconstruction?** shows the hypothetical surface projection of the Pacific-Farallon slab window and partially subducted Monterey plate at ca. 19 Ma [*Wilson et al.*, 2005]. The slab window formed by subduction of the trailing edge of the Farallon plate, unsupported by sea floor spreading along the former spreading axis with the Pacific plate. The Monterey plate nucleated along a ~250 km segment of the Pacific-Farallon ridge as an oblique rift that was rotated ~25º clockwise from the Pacific-Farallon rift axis [*Atwater and Severinghaus*, 1989]. Its generation was synchronous with the early stages of Pacific-Farallon plate convergence into the Cordilleran subduction zone along the southern California coastal region, and coincided with transrotational rifting of the continental borderland region and displacement of the western Transverse Ranges bedrock [*Atwater and Stock*, 1998; *Bohannon and Parsons*, 1995].

According to the *Wilson et al.* [2005] reconstruction of the Pacific-Farallon slab window and adjacent Monterey plate [Figure **¿fig:reconstruction?**], the Crystal Knob eruption site was located above a slab window in the early Neogene, ~50-100 km northeast of the northeastern boundary transform of the Monterey plate. Diffuse volcanism, some clearly derived from decompression partial melting of convecting mantle, is widespread at this time period across the region of the reconstructed slab window [*Cole and Basu*, 1995; *Hurst*, 1982; *Sharma et al.*, 1991; *Wilson et al.*, 2005]. However, this phase of slab window opening and related volcanism cannot account for the eruption of the ca. 1.7 Ma Crystal Knob volcanic neck itself, which we return to in Section 5.2.

## The Monterey plate

A stalled partially subducted Monterey microplate has been invoked as the source of the lithospheric mantle beneath the central Coast Ranges [*Erkan and Blackwell*, 2008]. When the East Pacific Rise first reached the North American plate at 28.5 Ma, the Monterey microplate broke from the Farallon slab and subducted independently until 19.5 Ma, while rotating clockwise with respect to the Pacific plate [*Wilson et al.*, 2005], Figure **¿fig:reconstruction?**. The remnant microplate has been integrated into the Pacific plate and still forms part of the abyssal seafloor in the proximal offshore region [Figure **¿fig:context?**]. A recent permutation of the stalled slab hypothesis is that the Monterey plate extends eastward of the San Andreas fault as a horizontally translated "dangling" slab [*Brink et al.*, 1999; *Pikser et al.*, 2012; *Wang et al.*, 2013], Figure **¿fig:context?**.

The current position of the Monterey plate offshore of the Crystal Knob eruption site is a result of dextral displacements linked to borderland transrotational rifting, subsequent ~155 km-scale dextral offsets along the San Gregorio-Hosgri fault system, and ~100 km of additional dextral offsets in the offshore region [Figure **¿fig:context?**] as modeled both by geologic reconstruction of fault offsets [*Dickinson et al.*, 2005] and plate kinematic reconstructions [*Wilson et al.*, 2005]. Continuation of the Monterey plate east of the San Gregorio-Hosgri fault requires the translation of its downdip extension laterally along the former subduction interface beneath the Coast Ranges. Its hypothetical extension east of the San Andreas fault as a "dangling slab" requires that this subduction interface likewise extended eastwards beyond the San Andreas fault. Seismological, geodynamic, and surface geological evidence presented here argues against models invoking horizontal translation of the Monterey plate, both beneath the Coast Ranges and as a “dangling slab” beneath the Central Valley.

In the Figure **¿fig:reconstruction?** reconstruction, the future Crystal Knob eruption site sits above the Pacific-Farallon slab window ~50-100 km north of the northeast margin of the partially subducted Monterey plate. The narrow slab window segment shown along the eastern edge of the partially subducted plate marks the plate’s separation locus with the Farallon plate, which subsequently opened wider beneath the southern California region as the Farallon plate descended deeper into the mantle [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Over the time interval of ca. 22-10 Ma, the Monterey plate’s dextral motion relative to the subducting trench had a nontrivial divergence component, due to its coupling to the Pacific plate. Potential extensional attenuation of the underthrust portion of the Monterey plate during such divergent motion is strongly implied in the *Bohannon and Parsons* [1995] reconstruction, though not explicitly explored. Coupling of Monterey plate divergent motion across the subduction megathrust break is hypothesized to have driven dextral transrotational rifting [*Bohannon and Parsons*, 1995].

As western Transverse Range crustal panels rotated into their current position during transrotational rifting, the Monterey plate continued its northward displacement along the San Gregorio-Hosgri fault system [Figure **¿fig:context?**]. Note that in Figure **¿fig:neogene\_sections?** the outer edge of the Farallon-Monterey slab window is on trend with the San Gregorio-Hosgri fault system [*Dickinson et al.*, 2005]. Distinct steps and inflections in lower crustal velocity structure across this fault system [*Brocher et al.*, 1999] indicates that it cuts the entire crust. This poses the likely possibility that the San Gregorio-Hosgri fault system bounds the eastern margin of underplated Monterey plate in the coastal central California region. This is in line with seismic observations showing an ~16º eastward-dipping Monterey plate offshore, with a typical abyssal crustal thickness, juxtaposed against a nearly flat thickened lower crustal layer beneath the Nacimiento Franciscan [*Nicholson et al.*, 1992; *Tréhu*, 1991]. These observations are in direct conflict with the notion that a structurally continuous mafic layer constitutes the lower crust beneath the central coastal California and adjacent offshore region.

Studies proposing a deep Monterey plate "dangling slab" [e.g. *Furlong et al.*, 1989; *Pikser et al.*, 2012] have suggested that translation of the Monterey plate along the San Andreas system entailed significant sub-horizontal fault segments that accommodated dextral displacements [Figure **¿fig:neogene\_sections?**‌c]. As of yet, however, all seismically imaged segments of the transform system have been shown to be steeply oriented [*Brocher et al.*, 1999; *Dietz and Ellsworth*, 1990; *Ozacar and Zandt*, 2009; *Titus et al.*, 2007; *Yan and Clayton*, 2007; *Yan et al.*, 2005].

This assertion is generally paired with the discussion of a structurally continuous mafic layer, which represents the stalled Monterey plate and constitutes the lower crust beneath the entire coastal central California and offshore region. These studies further suggest that the Monterey "dangling slab" currently corresponds to the high-wave speed anomaly of the southern Sierra Nevada-Great Valley region [Figure **¿fig:context?**], commonly called the “Isabella anomaly”.

Seismological and geodynamic studies show that the Isabella anomaly is derived primarily from the convectively mobilized mantle wedge, or mantle lithosphere of the southern Sierra Nevada batholith [*Frassetto et al.*, 2011; *Gilbert et al.*, 2012; *Jones et al.*, 2014; *Levandowski and Jones*, 2015; *Saleeby et al.*, 2012; *Zandt et al.*, 2004]. In addition to showing structural continuity between the seismic anomaly and the residual mantle lithosphere that is still in place beneath the Central Valley and Sierra Nevada [Figure **¿fig:context?**], these studies show that the volume of the Isabella anomaly far exceeds reasonable volume estimates for the attenuated terminus of a hypothetical translated Monterey slab. These studies also provide mechanisms for lower crustal plastic deformation, observable surface faulting, upper mantle--lower crustal partial melting and dynamic topographic effects that are all ignored in the dangling slab hypothesis.

First-order geological effects such as volcanism and topographic transients are closely correlated to the convective mobilization of the sub-Sierran mantle lithosphere, and its current expression as the Isabella anomaly [*Cecil et al.*, 2014; *Ducea and Saleeby*, 1998b; *Farmer et al.*, 2002; *Levandowski and Jones*, 2015; *Saleeby et al.*, 2013]. The surface effects of Monterey plate partial subduction followed by transtensional coupling to Pacific plate motions are closely correlated to transrotational rifting in the southern California Borderland and the linked clockwise rotation of western Transverse Ranges bedrock panels [*Bohannon and Parsons*, 1995; *Wilson et al.*, 2005]. This is in line with the Monterey slab’s limited down-dip extent as bounded by the Monterey-Farallon slab window segment shown in Figure **¿fig:reconstruction?**. If the hypothetical Monterey "dangling slab" were of proper proportion to form the Isabella anomaly, then why were its effects on surface geology restricted to the Borderland and Transverse Ranges? Epeirogenic transients that correlate to the convective mobilization of the sub-Sierran mantle lithosphere as the Isabella anomaly are highly out of phase with the predicted translation pattern for a “dangling” Monterey slab [*Cecil et al.*, 2014; *Saleeby et al.*, 2013]. Possible remnants of necked off partially subducted Monterey plate are more plausibly correlated to the Transverse Ranges high-wave speed anomaly in terms of position and volume [Figure **¿fig:context?**], and also have a firm geodynamic basis as such [*Burkett and Billen*, 2009].

Much of the attractiveness of the stalled slab hypothesis lies in its utility to explain sparse Neogene volcanism in the Coast Ranges as well as modern heat flow values that are too low to be explained by shallow underplated asthenosphere emplaced within a slab window. *Erkan and Blackwell* [2008] examined whether the migrating slab window caused wholesale replacement of underlying mantle lithosphere against a possible "stalled slab", and concluded that the stalled slab was more likely due to the lower thermal gradients and heat flows predicted by this mechanism. Though their model techniques were sound, they failed to consider another distinct tectonic mechanism that could result in similarly lower thermal gradients than a shallow slab window. We summarize this mechanism and the supporting geologic data below.

## Underplated Farallon Plate mantle nappes

The reconstruction of the Crystal Knob eruption site to its pre-San Andreas position [Figure **¿fig:reconstruction?**] suggests the underplating of Farallon-plate mantle nappes prior to Neogene transform faulting as a highly viable alternative for the development of the site’s underlying mantle lithosphere.

The Crystal Knob neck erupted through the Nacimiento belt of the Franciscan complex, immediately adjacent to the current western limit of Salinia crystalline nappes [Figure **¿fig:context?**]. Accretion of the Nacimiento belt occurred in the Late Cretaceous beneath the outer reaches of the Salinia nappe sequence [*Chapman et al.*, 2016a; *Hall and Saleeby*, 2013]. In their core area, the Salinia nappes rode westwards on slightly older, higher metamorphic grade, Franciscan rocks that are shown in Figure **¿fig:context?** and Figure **¿fig:reconstruction?** as windows into subduction channel schists [*Barth et al.*, 2003; *Ducea et al.*, 2009; *Kidder and Ducea*, 2006]. The southernmost Sierra Nevada-western Mojave "autochthon" for the Salinia nappes is likewise detached from its original mantle wedge underpinnings, and shingled into crystalline nappes that lie on underplated high-grade subduction channel schists [*Chapman et al.*, 2010, 2016b, 2012; *Saleeby*, 2003]. Tectonic erosion of the mantle wedge followed by shallow subduction underplating of Franciscan rocks requires subsequent reconstruction of the current mantle lithosphere. *Luffi et al.* [2009] and *Armytage et al.* [2015] present petrologic studies on the Dish Hill and Cima mantle xenolith suites [Figure **¿fig:context?**] that suggest the presence of a mantle lithosphere duplex with multiple Farallon plate upper mantle nappes in structural sequence beneath an eastern residual roof of continental mantle lithosphere. Since the crustal structural sequence of the western Mojave region is closely spatially and temporally correlated to that of the Salinia nappes [*Chapman*, 2016; *Chapman et al.*, 2010, 2012], it stands to reason that upper mantle duplex accretion progressed westwards from the Mojave region to beneath the Salinia nappes as well as the Nacimiento belt of the Franciscan.

In Figure **¿fig:cross\_sections?**, we present a model for the tectonic underplating of the Farallon plate mantle lithosphere beneath the Mojave-Salinia-Nacimiento segment of the Late Cretaceous convergent margin [*Luffi et al.*, 2009; after *Saleeby*, 2003]. This occurred in conjunction with shallow flat subduction of the Shatsky Rise conjugate Large Igneous Province [*Liu et al.*, 2010; *Saleeby*, 2003; *Sun et al.*, 2017]. The approximate age of Farallon plate entering the trench is shown on each frame [after *Seton et al.*, 2012]. Crustal deformation, timing, and thermal conditions are integrated from *Kidder and Ducea* [2006], *Chapman et al.* [2010], *Chapman et al.* [2012], and *Chapman et al.* [2016a]. Figure **¿fig:cross\_sections?**‌a and b show the arrival of the oceanic plateau into the subducting trench, and plateau buoyancy driven shallowing of the subduction megathrust, which drove tectonic erosion of the mantle wedge. Temperature conditions along the flat subduction megathrust initiated at ~900ºC, ambient conditions within the deep levels of the then-active arc, and retrogressed to ~715ºC, peak temperatures recorded in shallowly subducted metaclastic rocks of the Sierra de Salinas schist, exposed in the principal Salinia window into the subduction channel schists [*Kidder and Ducea*, 2006].

In Figure **¿fig:cross\_sections?**‌c and d we adopt the focused slab rollback and mantle lithosphere underplating models of *Saleeby* [2003] and *Luffi et al.* [2009], for the dynamic response of normal-thickness oceanic lithosphere following the thickened oceanic plateau into the subduction zone. Suction forces on the retreating slab drove crustal responses including large-magnitude, trench-directed extension coupled to regional extrusion of the underplated subduction channel schists. In the Figure **¿fig:cross\_sections?**‌c to d transition, accelerated rollback is driven by the formation of Farallon-plate mantle duplexes. We suspect that mantle nappe detachment was controlled primarily by the temperature of the brittle-plastic transition in olivine. For ca. 40-50 m.y. old oceanic lithosphere entering the subduction zone Figure **¿fig:cross\_sections?**‌c and d, this ~700-800 ºC transition [*Bürgmann and Dresen*, 2008; *Mei et al.*, 2010; *Warren and Hirth*, 2006] occurs at ~25-40 km depth in the slab [*Doin and Fleitout*, 1996]. We also suspect that retreating subduction imparted a significant tensile stress within the slab that further promoted nappe detachment. Observational data and laboratory experiments show that profound co-seismic dilation transients can develop along subduction megathrusts [*Gabuchian et al.*, 2017], underscoring the potential importance of tensile stresses within subducting slabs.

The lack of high-pressure mafic schists in both the Crystal Knob and Dish Hill xenolith suites further suggests that slivers of oceanic crust were detached along with mantle lithosphere nappes (presumably from oceanic Moho depths) to be underplated as the seismically imaged thickened mafic lower crust of the region [*Brocher et al.*, 1999; *Tréhu*, 1991]. On the basis of both the regional structural evolution of the central to southern California basement and the petrogenetic history recorded in the region's mantle xenolith suites, we suspsect the Figure **¿fig:cross\_sections?**‌d section most accurately represents the construction of the mantle domain sampled by Crystal Knob. This section is idealized for Late Cretaceous time, and we now layer the complexity of late Cenozoic tectonics onto this framework.

## A deep slab window beneath relict lithosphere

Kinematic reconstructions of the impingement of the Pacific-Farallon spreading center on the SW Cordilleran subduction zone require a slab window beneath the Crystal Knob eruption site in the early Neogene [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Previous modeling of thermal effects of the slab window [*Erkan and Blackwell*, 2008] has only investigated the resulting emplacement of asthenosphere at immediate subcrustal levels. However, the depth of asthenospheric underplating related to slab window opening is poorly constrained, and likely to vary geographically as a function of thickness and thermal variations in the pre-existing lithospheric lid, as well its state of stress and structural coherency. Though volcanism in the central California Coast Ranges has been tied to slab window opening [e.g *Ernst and Hall*, 1974], it has been volumetrically insignificant when compared to that generated by other coeval examples of shallow asthenospheric upwelling in the Cordillera such as the high-flux volcanism generated in the Basin and Range province in the Eocene--Miocene [e.g. *Humphreys*, 1995].

The lack of a signature of shallow asthenospheric mantle in the Miocene is readily explained if the slab window opened beneath pre-existing mantle lithosphere originating in the Cretaceous. This would consist of a tiered duplex of underplated Farallon mantle nappes, roofed by a duplex of underplated Farallon oceanic crust (lower crustal mafic layer), in turn roofed by the Nacimiento Franciscan and Salinia nappes. Our estimate of a 45-75 km depth interval over which the Crystal Knob lavas sampled the underlying mantle lithosphere [Figure **¿fig:depth?**], coupled with a general lack of significant late Cenozoic extensional faulting in the immediate region, implies a strong thermo-mechanical lid that likely suppressed the ascent of voluminous asthenosphere derived magmas that were hypothetically sourced from a deep underlying slab window.

We now model the thermal evolution of the coastal California mantle corresponding to the scenarios outlined above and test the resulting predictions against the sub-Crystal Knob geothermal state of the mantle lithosphere.

# Thermal modeling of tectonic scenarios

The slab window, Monterey Plate, and Cretaceous mantle duplexing scenarios for the source of the Crystal Knob xenoliths all imply a peridotite composition with a depleted (convecting-mantle) isotopic and trace-element signature. Though petrographic and geochemical variations provide information on the depletion history, they cannot discriminate between these potential depleted convecting mantle sources. However, these emplacement scenarios present potentially distinct thermal structures due to large differences in timescales of cooling. Tectonic models for the emplacement of depleted mantle lithosphere under the central coastal California region can be tested by comparison of their implied geothermal structure with xenolith geothermometry.

## Model setup

To distinguish between potential emplacement mechanisms for the mantle lithosphere sampled by Crystal Knob, we construct a forward model of the geotherm implied by each of the tectonic scenarios shown in Figure **¿fig:neogene\_sections?**. A model based on the one-dimensional heat-flow equation is used to track the evolution of the lithospheric geotherm predicted by the three tectonic scenarios presented above.

To simulate both subduction and slab-window driven mantle underplating, the forearc geotherm is stacked atop a modeled sub-oceanic or asthenospheric geotherm and relaxed towards the present by iteratively solving the heat-flow equation. Finite-difference modeling is implemented in Python using the FiPy software package [*Guyer et al.*, 2009]. Explicit and implicit finite difference approaches are combined using a two-sweep technique [*Crank and Nicolson*, 1947] to ensure a stable result. The model is run to a depth of 500 km to remove the effects of unknown mantle heat flux.

Several auxiliary analytical models are used to constrain portions of our modeled scenarios. We use the Global Depth and Heat (GDH) model for oceanic crust [*Stein and Stein*, 1992], and the *Royden* [1993] forearc geotherm model to track the evolution of a geotherm during continuous subduction. Standard values used for oceanic and continental material properties are given in [Table **¿tbl:model\_parameters?**]. More information about model setup and integration is given in Supplementary Materials.

## Model results

Model results are presented as geotherms corresponding to specific model steps in Figure **¿fig:model\_results?** and as temperature--time tracers in Figure **¿fig:model\_tracers?**.

### Shallow slab window

The geologic context of the shallow slab window scenario is shown in Figure **¿fig:neogene\_sections?**‌a, and our thermal modeling for this scenario (model group **A**) is displayed in Figure **¿fig:model\_results?**‌a and Figure **¿fig:model\_tracers?**‌a. The model begins at 24 Ma, corresponding to the opening of the Mendocino slab window under southern California [*Wilson et al.*, 2005]. A steady-state profile through the crust is truncated by a mantle adiabat to simulate direct contact with the ascended asthenosphere (for 0-6 Myr), after which the domain relaxes conductively to the conclusion of the model. Previous modeling by *Erkan and Blackwell* [2008] suggests that this scenario yields temperatures too hot to correspond to the modern regional geotherm. We confirm this assessment, finding that this scenario produces extremely steep geotherms at the upper boundary of spinel lherzolite stability for much of the temperature domain of interest [Figure **¿fig:model\_comparison?**], reproducing neither the xenolith pressure--temperature array developed in this study nor the seismically-inferred depth of the lithosphere-asthenosphere boundary [e.g. *Li et al.*, 2007].

### Neogene stalled slab

The geologic context of the stalled slab scenario is shown in Figure **¿fig:neogene\_sections?**‌b, and our thermal modeling of this scenario (model group **B**) is displayed in Figure **¿fig:model\_results?**‌b and Figure **¿fig:model\_tracers?**‌b. This scenario tracks the potential thermal structure of oceanic plates stalled under the forearc at a range of times. Each run begins at a specified time with the subduction of oceanic lithosphere assigned an initial thermal structure corresponding to the Global Depth and Heat model [*Stein and Stein*, 1992] for oceanic lithosphere of a given age of oceanic crust.

We model cooling scenarios for a wide range of underplating times, with the start of subduction ranging from 22 to 80 Ma. This reflects the long subduction history of the Farallon plate beneath the central California coast through the Cretaceous and Paleogene. These subduction times, T, set the initial conditions shown in Figure **¿fig:model\_tracers?**‌b and the first panel of Figure **¿fig:model\_results?**‌b. Each model operates on oceanic crust of the appropriate age for the time of subduction, given the geometry of Farallon plate subduction over the Cretaceous and Paleogene [*Liu et al.*, 2010; *Seton et al.*, 2012]. As T approaches the present, the age of subducted oceanic crust generally decreases, reflecting the approach of the Pacific--Farallon spreading ridge to the western margin of North America.

Stalled slab scenarios with subduction ages older than 30 Ma simulate rollback during sustained Farallon-plate subduction. While backstepping of the subduction megathrust and underplating of a slice of mantle lithosphere could, in principle, occur at any time during subduction, these older stalled-slab models do not correspond to geodynamic and geological evidence of a specific episode of subduction instability. Though improbable, these models are included to fully explore the model space between model groups **B** and **C**, and are represented with a reduced opacity on Figure **¿fig:model\_tracers?**‌b. In the oldest model with a subduction time of 70 Ma, the oceanic lithosphere is 50 Myr old at the time of subduction. At this time, the Shatsky conjugate had already subducted to beneath the Cordilleran interior [*Liu et al.*, 2010] and the Nacimiento belt of the Franciscan was in its later stages of subduction accretion [*Chapman et al.*, 2016a]. This is the earliest time a stalled slab could have developed outside of the specific scenario treated in model group **C**.

The youngest model run in Figure **¿fig:model\_tracers?**‌b corresponds to the "Monterey plate" hypothesis [*Pikser et al.*, 2012; *Van Wijk et al.*, 2001], which entails hypothetical northward lateral translation on a shallowly-dipping arrested subduction megathrust. The potential thermal effects of the required anhydrous shearing of the underplated oceanic lithosphere along a ~300 km flat displacement trajectory [Figure **¿fig:context?**] are not accounted for in model **B**. Instead, this scenario is modeled simply as a young endmember stalled-slab scenario, with generation of mantle lithosphere beneath the oceanic spreading ridge at 27 Ma (corresponding to the chron 7 magnetic anomaly) and subduction shortly thereafter [*Atwater and Stock*, 1998; *Wilson et al.*, 2005].

Overall, the stalled-slab underplating scenarios represented in **B** result in cooler geotherms than shallow slab window underplating, matching the broad thermobarometric constraints placing Crystal Knob xenolith entrainment relatively deep within the spinel stability field [Figure **¿fig:model\_comparison?**]. The Monterey plate subduction scenario likewise predicts a modern geotherm that matches the entrainment constraints on the Crystal Knob xenoliths. Without consideration of potential bias towards colder measurements in the modeled geotherms, this appears to match our xenolith data. When accounting for possible external effects Section 3.1 and Supplementary Materials, it suggests a hotter geotherm than that derived from the thermobarometric constraints.

### Late-Cretaceous mantle nappe underplating

The geologic context of the Late Cretaceous mantle nappe underplating scenario is shown in Figure **¿fig:neogene\_sections?**‌c, and our thermal modeling of this scenario (model group **C**) is displayed in Figure **¿fig:model\_results?**‌c and Figure **¿fig:model\_tracers?**‌c. The initiation of this scenario is similar to the older stalled slab scenarios [Figure **¿fig:model\_tracers?**‌b]. In both cases, the oceanic mantle forms under the Pacific--Farallon spreading ridge during the Late Cretaceous, thermally matures to form a mantle lithosphere lid during oceanic plate transport, and subducts beneath the southwest Cordilleran margin later in the Cretaceous. Thus, the initial conditions and thermal evolution of scenario **C** are qualitatively similar to the older runs of **B**, except that this scenario incorporates more crustal geological constraints that pertain to its subduction history. In model **C**, the *Royden* [1993] forearc geotherm is tied to temperature constraint of 715ºC at 25 km depth based on garnet-biotite thermometry of Salinia granites that lie tectonically above the subduction complex, and ~575ºC at 30 km depth based on garnet-biotite thermobarometry on the proximally underplated schist of Sierra de Salinas schist [*Ducea*, 2003; *Kidder and Ducea*, 2006]. The subduction conditions and mantle lithosphere structure implied by this scenario are shown in Figure **¿fig:cross\_sections?**.

In model **C**, the age of subduction and underplating is taken as ~70 Ma, based on the youngest ages of the Sierra de Salinas/Nacimiento Franciscan subduction complex [*Barth et al.*, 2003; *Chapman*, 2016; *Chapman et al.*, 2016a, 2010; *Grove et al.*, 2003; *Saleeby et al.*, 2007]. Seafloor being subducted at that time was 40 Myr old [*Liu et al.*, 2010; *Seton et al.*, 2012]. In this tectonic scenario Figure **¿fig:neogene\_sections?**‌c and Figure **¿fig:cross\_sections?**, Farallon oceanic lithosphere continued to subduct after mantle nappe detachment until the Pacific--Farallon spreading ridge encountered the trench in the Neogene. In the thermal model [Figure **¿fig:model\_results?**], the underplated mantle nappe(s) cool beneath the forearc for 50 Myr, after which the geotherm is perturbed by the underplating of asthenosphere at ~80 km depth, corresponding to a deep slab window.

In several model runs, asthenosphere is held against the base of the lithosphere for periods ranging from 0 Myr to 6 Myr. An adiabatic temperature gradient with a mantle potential temperature of 1450ºC is held against the base of the lithosphere for the duration of contact. The model for 0 Myr entails instantaneous contact followed immediately by conductive relaxation, while 6 Myr of sustained upwelling produces the "kinked" geotherm seen in panel 4 of Figure **¿fig:model\_results?**‌c at the 18 Ma time step, due to continuing imposition of a mantle adiabat below 80 km depth. A single model without slab window heating [highlighted in Figure **¿fig:model\_comparison?**] predicts much cooler geotherms that do not match the mantle geothermal constraints developed in this study.

Figure **¿fig:model\_results?**‌c, panel 2 shows the thermobarometric constraints and inverted metamorphic gradient recorded by subduction-channel schists for this episode of subduction [*Kidder et al.*, 2013; *Kidder and Ducea*, 2006] and used to tune the *Royden* [1993] forearc geotherm model. These high subduction temperatures constrained by crustal geothemometry make little difference to the final thermal structure of the mantle lithosphere [Figure **¿fig:model\_tracers?**‌c]. When not reheated by a deep slab window, the Cretaceous underplating scenario has a similar final thermal structure to the longest-running stalled slab scenarios in **B** [Figure **¿fig:model\_comparison?**]. This reflects the model's basic correspondence with a generalized Farallon plate mantle lithosphere underplating event of similar age. High subduction-channel temperatures experienced during late-Cretaceous flat slab subduction and schist metamorphism did not have a long-lasting impact on the thermal structure of the margin. Thus, heating by a Miocene deep slab window is required for Cretaceous mantle nappe underplating scenarios to produce warm mantle lithosphere.

## Summary of model results

Our thermal modeling predicts much higher temperatures within the mantle lithosphere, and much higher geothermal gradients, for the shallow slab window than for the stalled-slab or underplated mantle nappe models. The geothermal gradients implied for the shallow slab window scenario are much higher than those suggested by heat flow data in the Coast Ranges, leading *Erkan and Blackwell* [2008] to favor a stalled slab tectonic scenario. Our modeling predicts that both the stalled slab and Cretaceous mantle nappes reheated by a deep slab window recover the geotherm determined by xenolith thermobarometry, while not violating constraints posed by heat flow data. An assessment of model sensitivity presented in Supplementary Materials includes biases that may influence the model results. Most pertinent is the underestimation of modeled geothermal gradients due to lack of accounting for exhumation during lithospheric thermal equilibration. This would: 1. for the slab window scenario, further push the model results out of the field of acceptable mantle lithosphere geotherms derived from xenolith constraints; 2. for the Monterey plate, push it to the upper margin of the field; and 3. for the reheated mantle nappe scenario, push it towards the centroid of the field.

# Contemporary lithospheric structure and thermal state

In this section we integrate the results of thermal modeling with petrogenetic findings on the Crystal Knob xenoliths, regional crustal structure and evolution, and the timing and location of xenolith entrainment. Of the three plausible scenarios depicted for the evolution of the sub-Crystal Knob mantle lithosphere in Figure **¿fig:neogene\_sections?**, we reject the scenario of asthenosphere emplaced in a shallow slab window, based on our thermal modeling presented above. The Monterey plate stalled slab and underplated Farallon plate mantle nappe cases are equally plausible based on our thermal modeling. Based on the wide spectrum of geologic and geodynamic factors discussed above, we dismiss the notion suggested by *Pikser et al.* [2012] of a regionally extensive Monterey Plate "dangling slab" extending far to the east of the San Andreas fault.

Depending on the original scale of the subducted portion of the Monterey Plate beneath the southern California borderland [Figure **¿fig:neogene\_sections?**], its structural integrity following its coupling to borderland transrotational rifting, and the potential correspondence of the partially subducted plate to the detached portion that forms the Transverse Ranges high-wave speed anomaly [Figure **¿fig:context?**], an argument could be made that an underthrust portion of the Monterey Plate has been translated northwards horizontally beneath the Crystal Knob eruption site [Figure **¿fig:neogene\_sections?**‌b]. Although kinematically plausible, in theory, this case seems unlikely based on dynamic factors. As with the "dangling slab" version of this case, horizontal translation of such a large mantle mass along the base of the crust should manifest at the surface by transients in dynamic topography, as well as brittle crustal deformation in response to horizontal shear stresses in the lower crust. Such surface deformation patterns are not expressed north of the Transverse Ranges in late Cenozoic time. Furthermore, horizontal translation of a previously underthrust slab provides no melting mechanism for the early Neogene or Pleistocene volcanic centers of the region. We thus conclude that, despite the inability the Crystal Knob xenolith suite to fully rule it out, an eastward-extending stalled Monterey plate is geologically and geodynamically untenable.

In Figure **¿fig:neogene\_sections?**‌c we show the partially subducted terminus of the Monterey plate bounded to the east by the San Gregorio-Hosgri fault, based on our above discussion of the Figure **¿fig:reconstruction?** reconstruction. East of the fault lies the Nacimiento Franciscan complex and its tectonic veneer of Salinia nappes (not differentiated on the figure), and its lower crustal oceanic crustal duplex lying tectonically above underplated Farallon-plate mantle nappes. The structural profile shown on Figure **¿fig:neogene\_sections?**‌c between the San Andreas and San Gregorio-Hosgri faults was constructed at southern California latitudes in continuity with that of the southernmost Sierra Nevada and adjacent Mojave plateau region [Figure **¿fig:reconstruction?**; Figure **¿fig:cross\_sections?**]. Partial subduction, or stalling, of the Monterey plate occurred along the outer edge of Franciscan complex, further south than rocks of the Nacimiento belt [Figure **¿fig:reconstruction?**].

Slab window opening beneath the Crystal Knob eruption site is reconstructed to have occurred between ca. 28-23 Ma [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Based on the colocation of distinct Cretaceous structural features with the footprint of the Neogene slab window, along with our constraints from the Crystal Knob xenoliths, we conclude that a thick and relatively cool lithospheric lid inhibited widespread voluminous volcanism in response to asthenospheric upwelling during the opening of the slab window.

## Implications of petrologic complexities to thermal state

Petrologic markers within the Crystal Knob xenolith suite and the late timing of the Crystal Knob eruption can be related to interaction of the lithospheric lid from which the xenoliths were sourced with an underlying deep slab window. The Crystal Knob xenolith suite shows petrologic variation consistent with reheating from below. As discussed in Section 2.10, the array of increasing depletion with depth in the Crystal Knob sample set does not match simple decompression melting, and sample CK-6 experienced polyphase major-element refertilization by fractionated melt presumably generated below its pre-eruptive depth. At the same time, low levels of re-enrichment by alkali-basalt-like melt are evident across the sample set. Deep partial melting and alkalic melt generation are common features of slab window volcanism [*Hole et al.*, 1991] and may contribute to the signature of deep depletion and re-enrichment seen in the Crystal Knob sample set.

The sourcing of the Crystal Knob pipe from relatively deep levels of the mantle lithosphere, the highly fractionated, alkalic magma, and the presence of both dunite cumulates and peridotite xenoliths suggest that the magma was sourced from a complex multi-tiered melt percolation network within the mantle lithosphere at depths >45 km, such as the fractal scaling of melt migration channels investigated by *Kelemen et al.* [2000]. These host cumulates which are sometimes re-entrained by small-volume volcanism, during a process of progressive upwards percolation of melt from a deep mantle source.

A long-duration reservoir of slab window material locally rising through a thick lid of relict mantle lithosphere may explain both mid-Miocene hypabyssal intrusives that are directly related to the slab window episode, such as the Morro Rock--Islay Hills complex [*Stanley et al.*, 2000] and Cambria Felsite [*Ernst and Hall*, 1974], and later deeply-sourced, small-volume eruptions such as Crystal Knob.

## The Crystal Knob eruptive event

The time lag between deep slab window opening and the ca. 1.7 Ma eruptive age of Crystal Knob underscores the unknown genesis of the Crystal Knob lava. Our studies of the Crystal Knob xenoliths indicate an underlying lithosphere-asthenosphere boundary at a depth of 70-90 km, consistent with regional seismic studies placing it at ~70 km beneath the central California Coast Ranges [*Li et al.*, 2007]. In contrast, 28-20 Ma oceanic lithosphere of the adjacent Monterey plate [*Wilson et al.*, 2005], would have its lithosphere-asthenosphere boundary at ~35 km, based on thermal decay relations [*Doin and Fleitout*, 1996]. The offshore Monterey plate is at the edge of the resolution of *Li et al.* [2007], who calculate a ~50 km depth for the lithosphere-asthenosphere boundary. Proximal to the modern shoreline, the Monterey plate is thrust beneath ~12 km of sedimentary accretionary prism [*Tréhu*, 1991], bringing the theoretical thermal maturation depth for the transition much closer to the depth observed by *Li et al.* [2007]. Crystal Knob is located ~15 km east of the Hosgri fault, with its host Franciscan complex pervasively cut by faults and shear zones [*Cowan*, 1978; *Seiders*, 1989]. The geologic slip and seismicity history of the Hosgri fault [*Dickinson et al.*, 2005; *Hardebeck*, 2010] indicate that it was likely active during the eruption of the Crystal Knob neck. Integration of theoretical and observational data on intra-continental transform faults [*Platt and Behr*, 2011; *Titus et al.*, 2007] indicate that, at lower crustal--upper mantle levels, Hosgri fault shear could be distributed across 10s of kilometers normal to the fault surface [Figure **¿fig:neogene\_sections?**]. Eruption of small-volume basalts of Plio-Pleistocene age occurred elsewhere in the Coast Range belt. This includes the Coyote Lake pipe [Figure **¿fig:context?**], which occurred ~150 km north of Crystal Knob along the San Andreas-Calaveras fault bifurcation zone and entrained lower crust and upper mantle xenoliths [*Jové and Coleman*, 1998; *Titus et al.*, 2007]. Xenoliths recovered from these flows record asthenosphere ascent and partial melting markedly post-dating any possible slab window opening. Thus, the Crystal Knob small volume eruption is not an exceptional event.

Distributed shearing and strike-slip juxtaposition of the shallow sub-Monterey plate asthenosphere against underplated Farallon plate lithosphere and its deep slab window asthenosphere along the Hosgri fault [Figure **¿fig:neogene\_sections?**‌c] is a plausible mechanism for Crystal Knob basalt melt generation. This was perhaps accentuated by possible extensional transients along the fault surface as documented for the Coyote Lake basalts [*Jové and Coleman*, 1998; *Titus et al.*, 2007]. Thermal modeling presented above indicates that underplated Farallon mantle was already re-heated by the deeper Neogene slab window. The abundance of dunite cumulate xenoliths and xenocrysts in the Crystal Knob basalt suggests the generation of significant melt in the mantle lithosphere beneath the Coast Ranges, and the entrainment of peridotite xenoliths underscores its rapid ascent at ~1.65 Ma. Two of the principal thermal maxima in the Coast Range thermal anomaly occur in the areas of the Crystal Knob and Coyote Lake Plio-Pleistocene basaltic eruptions [*Erkan and Blackwell*, 2008, Figure 1], further suggesting recent mobilization of asthenospheric mantle that was initially emplaced into the early Neogene slab window, due to extension in the mantle lithosphere.

# Conclusion

The lithospheric structure of southern California, to first order created by Cretaceous convergent margin tectonics, was severely overprinted by two subsequent tectonic episodes, with the impact and subduction of the Shatsky Rise large igneous province conjugate during the Late Cretaceous and the progressive evolution of a transform boundary in the Neogene. These episodes are recorded throughout the crustal geologic record of southern California and the central California Coast Ranges outboard of the San Andreas Fault. Using constraints from the Crystal Knob xenolith suite along with thermal modeling of tectonic scenarios, we show that the mantle lithosphere beneath the central California coast was profoundly affected by both of these episodes of deformation. The Crystal Knob suite is sourced along a depth gradient from ~45-70 km depth, and isotopic constraints show that it originates from the convecting mantle, which is typical of mid-ocean ridges or shallowly-ascended asthenosphere. Samples are variably depleted, and trace-element re-enrichment (and a single example of likely major-element assimilation) suggests interaction with low-volume melts after the formation and initial thermal equilibration of this mantle lithosphere material.

Major element, trace element, and radiogenic isotope data for the Crystal Knob xenolith suite equally satisfy the first-order geochemical requisites of the shallow slab window, stalled slab, and Late Cretaceous mantle nappe tectonic scenarios. Xenolith pressure-temperature constraints, thermal modeling, and geochemical subtleties of depletion and re-enrichment together add some discriminating factors between these scenarios. A shallowly underplated slab window predicts extremely hot geotherms that are untenable for the xenolith constraints of this study. The stalled slab and mantle nappe scenarios appear equally plausible in terms of the thermal modeling, equally satisfying constraints developed from xenolith thermobarometry. When the effects of potential exhumation/erosion are qualitatively considered, the Monterey plate stalled slab endmember scenario corresponds less well to constraints on the upper-mantle geotherm. This, along with a number of crustal geologic considerations, leads us to favor the Late Cretaceous mantle nappe underplating scenario, with reheating by a deep slab window in the Neogene.

This preferred scenario for the origin of the mantle lithosphere sampled by Crystal Knob xenoliths matches a host of geologic constraints demonstrating slab rollback and regional crustal extension during the Late Cretaceous, as the Shatsky Rise conjugate subducted deeper into the mantle following its initial collision and shallow subduction beneath the southern California convergent margin. This episode built the mantle lithosphere beneath the Mojave province by mantle duplexing during the retreat of the Farallon Plate subducting slab [*Luffi et al.*, 2009], and appears to have subsequently built the outboard mantle lithosphere beneath the Crystal Knob eruption site. The outer toe of this lithosphere-scale accretionary belt was subsequently displaced along the San Andreas transform system to its current location beneath the central California Coast Ranges. This displaced package of underplated mantle lithosphere records basal reheating by the Mendocino slab window, which opened in conjunction with the San Andreas plate juncture. Peridotite major- and trace-element re-enrichment and abundant dunite cumulate xenoliths and xenocrysts within the Crystal Knob basalt record the percolation of fluids and melts through the lithosphere. This percolation, the highly fractionated Crystal Knob basaltic pipe, and the modeled Neogene thermal pulse that reheated the lithosphere can be attributed to a deep slab window segment that opened as the Pacific--Farallon ridge encountered the California convergent margin. This adds to a growing body of evidence that much of the structural complexity in the California Coast Ranges is inherited from the Late Cretaceous regime of subduction accretion.

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# Supplementary Information

## Thermal modeling setup

Standardized parameters used in modeling are justified in the text below. Standard values for thermal conductivity from *Fowler* [2005] yield good results. Increasing the thermal conductivity of the model domain substantially depresses the modeled geotherms (lowering predicted temperatures at a given depth), but does not affect the relative temperatures predicted by the geotherms. Radiogenic heat flow for the continental marginal crust is estimated conservatively, and changes result in only minor changes to modeled geotherms across the board.

### Slab window crustal replacement

In series **A**, we model shallow slab-window upwelling. The emplacement of slab-window asthenosphere directly under the coastal central California crust entails the truncation of a low-temperature forearc geotherm at the base of the crust and the substitution of an asthenospheric adiabat below this level. The model begins at 24 Ma, corresponding to the time of opening of the Mendocino slab window under southern California [*Wilson et al.*, 2005]. The geotherm begins as a steady-state profile to 600 ºC at 30 km, truncated by a mantle adiabat. The mantle is held at asthenospheric conditions for a set period which is varied between model runs (from 0 to 6 Myr) to simulate a period of active convection, after which it relaxes conductively to the conclusion of the model.

### Subduction and underplating

Thermal conditions during subduction are tracked using the *Royden* [1993] steady-state forearc model. The samples then relax to the present. After subduction and underplating, the cooled oceanic lithosphere re-equilibrates with an overlying 30 km of forearc crust until the present, or for our xenolith samples until the time of ca. 1.7 Ma entrainment and eruption.

Progressive subduction of the downgoing slab beneath the forearc wedge is modeled as stepwise advection beneath a linearly thickening forearc wedge conforming to the *Royden* [1993] thermal model using the parameters outlined above. For all cases, the final depth of the underplated subduction interface is taken to be 30 km, and the distance landward of the subduction zone is taken to be 100 km. No effort is made to differentiate 'flat-slab' and baseline subduction geometries. Though increasing the slab dip angle will result in a cooler subduction interface at a given depth, the overall effect on the evolution of the thermal scenarios appears to be minimal.

### Oceanic geotherm

For the Neogene stalled Monterey plate and Late Cretaceous Farallon mantle nappe scenarios, the Global Depth and Heat (GDH) model [*Stein and Stein*, 1992] is used to trace the thermal evolution of the oceanic lithosphere from its emplacement at the spreading ridge until subduction. This model is a Taylor-polynomial fit of cooling parameters to global heat-flow and depth datasets. This fit yields higher geotherms than half-space cooling models that are directly based on Equation 1 (e.g., *Fowler* [2005]), and tends to produce higher geotherms for old oceanic lithosphere.

With the GDH model in conjunction with the *Royden* [1993] subduction model, we predict low temperatures (~235-245 ºC) at the subduction interface for the oldest stalled slabs modeled. For the Monterey Plate scenario (with young oceanic crust) the temperature at the subduction interface is predicted to be 980 ºC.

All oceanic-cooling models, including GDH and half-space cooling models, significantly overestimate heat flux from young oceanic plates, a fact that is likely attributable to vigorous hydrothermal circulation in young submarine lithosphere [*Stein*, 1995; *Stein and Stein*, 1992]. This may result in overestimates of geothermal gradients for the scenarios with the youngest subducted oceanic crust, such as the Monterey Plate scenario at the left of Figure **¿fig:model\_tracers?**.

### Supra-subduction geotherm

The geotherm of the forearc wedge during subduction is calculated using the *Royden* [1993] analytical solution for the steady-state thermal structure of continuously-subducting systems. Shear heating on the subduction thrust is ignored, as recent studies suggest that it is not an important factor [*Kidder et al.*, 2013]. Forearc rock uplift and erosion, as well as accretion and erosion on the subduction megathrust are ignored. In reality, megathrust accretion rates of 0.2-3.6 km/Myr are favored by *Kidder et al.* [2013] based on the Pelona schist, and some rock uplift is evident for the Coast Ranges.

The coastal California accretionary crust is represented homogenously as a material with a thermal conductivity of 2.71 W/m/K, specific heat capacity of 1000 J/kg/K, density of 2800 kg/m^3 and a radiogenic heat flux of 2 uW/m^3, values that are close to average for the continental crust [*Fowler*, 2005] and those used by *Kidder et al.* [2013] to model the thermal conditions along the Late Cretaceous shallow subduction megathrust segment. A radiogenic heat production in the crust of 2 uW/m^3 is actually a relatively conservative estimate given the fluxes shown for Sierra Nevada batholithic material by *Brady et al.* [2006], and the fact that much of the Franciscan material within the subduction channel is pelitic sediment rich in radiogenic elements [*Vilà et al.*, 2010]. Still, lower radiogenic heat production in the crust yields only a slight decrease in modeled geotherms across the board, not impacting conclusions.

## Thermal model sensitivity and bias

Generally, changes in model parameters such as radiogenic heat flux, thermal conductivity, and heat capacity do not impact the relative results for modeled scenarios, due to the consistent lithologic structure of the model domains.

Due to widely varying timescale of equilibration for modeled scenarios in groups **B** and **C**, the model is sensitive to assumptions about steady-state cooling of the oceanic mantle lithosphere. The choice of the "GDH" model to track the evolution of the suboceanic thermal structure is an important control on the scale of temperature variation in Figure **¿fig:model\_tracers?**‌b. Though GDH is well-calibrated, oceanic cooling models tend to overestimate the heat flow from young oceanic plates [*Stein*, 1995]. Thus, the modeled geothermal gradients for the younger stalled slab model runs may be too high.

Another potential confounding factor affecting the older scenarios of **B** and **C** is the thermal effects of continued subduction beneath the underplated mantle nappes. After rollback and underplating of the modeled section of oceanic mantle lithosphere, a downgoing slab at depth could, depending on its age, cool the forearc lithosphere from below. However, this effect is considered minimal and diminishes over time due to the progressive subduction of younger, hotter oceanic lithosphere. Reconstruction of the Pacific--Farallon spreading ridge history show that, between ca. 70 and 30 Ma, oceanic lithosphere entering the southwest Cordilleran subduction zone got younger at a rate of ~1 Myr/Ma [*Atwater and Stock*, 1998; *Liu et al.*, 2010; *Seton et al.*, 2012] corresponding to the approach of the ridge to the subduction zone. This factor coupled with slab window emplacement starting at ca. 24 Ma leads to the interpretation that cooling from below by continued subduction was of second-order significance.

Surface erosion is not modeled, but may bias the results. Any erosion will yield higher apparent heat flows and increased geotherm convexity, as heat is advected from the top of the model domain by material removal [*England and Molnar*, 1990; *Mancktelow and Grasemann*, 1997]. Geologic constraints suggest that 15-20 km of exhumation is likely to have occurred in a major pulse of unroofing coincident with flat-slab underplating and rollback in the Cretaceous [*Chapman et al.*, 2012; *Saleeby*, 2003], and is thus likely to disproportionately affect the older models. The lack of erosion in the model framework biases towards predicting lower geothermal gradient overall. For the slab window and underplated Monterey plate scenarios (model groups **A** and **B**) this effect would push the final geotherm to or beyond the limit of xenolith thermobarometry [Figure a and b]. In the underplated mantle nappe scenario (model **C**) this effect would push the final modeled geotherm towards the centroid of the xenolith thermobarometric array [Figure c and ].

The uncertainties inherent in this model bias the results towards predicting lower-temperature, less-convex geotherms over the model domain. These potential biases affect comparisons comparisons with measured values of heat flux and xenolith thermobarometry, which are not subject to these biases [Figure **¿fig:model\_comparison?**]. Thus, geotherms predicted by this model might be underestimates for potential mantle temperature at a given depth, especially for the older tectonic scenarios modeled.

## Factors not incorporated in the model

Several simplifications are made to create an internally consistent model framework. Subducted oceanic crust is not considered to have distinct thermal properties from the oceanic mantle. Additionally, though there are no reliable estimates of the mantle heat flux that cover the model domain, the model is run to great depth to avoid any influence of this uncertainty on the surface geotherm.

### Subduction zone rollback

The confounding factor of an active subduction zone just outboard of the scenarios for the older models is also not included within the model. When the trench interface jumps with the emplacement of an oceanic mantle nappe beneath the forearc, the new subduction interface will cool the detached nappe from below. This is not modeled because it would substantially increase model complexity (requiring a fully iterative approach to the forearc geotherm), and at this distance (~100 km) inboard of the final trench interface, there is limited scope for further episodic rollback after emplacement of the nappe(s) of presumed xenolith source [e.g. Figure c]. Further, although an active subduction interface at depth will cool the mantle lithosphere from below, the subduction of progressively younger crust until cessation at ~27 Ma will yield gradually increasing heat on the subduction interface [*Royden*, 1993]. The models for scenarios **B** and **C** [Figure b and c] are already near the coolest permitted by our xenolith constraints. As these geotherms are already quite cold, introducing this added complexity will not significantly change the model results. However, late-Cretaceous underplating and other stalled-slab scenarios can be treated as maximum temperatures because of the influence of the subducting slab.

### Change in convergence rate of rotating microplates

Potential Monterey Plate mantle lithosphere beneath Crystal Knob would have been emplaced under the ridge at 27 Ma (corresponding to the chron 7 magnetic anomaly) and subducted shortly thereafter [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Due to slower margin-normal convergence during microplate fragmentation and rotation [*Wilson et al.*, 2005], the parcel would take ~3 Myr to reach its final stalled position (~100 km behind the trench) as shown in Figure **¿fig:neogene\_sections?**‌b. This is responsible for the kink in the "Age of initial oceanic lithosphere" curve in Figure **¿fig:model\_tracers?**‌b. For model simplicity, we do not incorporate this disequilibrium shift into the starting parameters of the *Royden* [1993] subduction model.

### Erosion of the forearc

Surface erosion after underplating is taken to be zero. Any erosion will result in higher apparent heat flow values and increased geotherm convexity, as heat is advected from the top of the model domain by material removal. Geologic constraints suggest that the majority of erosion to the mid-crustal levels now at the surface in Salinia is likely to have occurred in a major pulse of unroofing coincident with flat-slab underplating and rollback [*Chapman et al.*, 2012; *Saleeby*, 2003], and is thus likely to disproportionately affect the older models. The 30 km of crust shown in the study area is based on modern estimates of the Moho depth, so recent erosion is unlikely to have biased the whole-lithosphere geotherm significantly. Still, the lack of erosion in the model framework will likely bias the results towards predicting a lower geothermal gradient overall, and lower temperatures in the mantle lithosphere, as upward advection of material by erosion increases the geothermal gradient [*England and Molnar*, 1990; *Mancktelow and Grasemann*, 1997]. Thus, these values need to be biased to higher temperatures to accurately capture the relationship between xenolith constraints on the actual temperature and temperatures derived from this modeling.

# Figure Captions

## field\_photo

Outcrop view of Crystal Knob peridotite xenoliths in place within the alkali basalt host lava.

## context

Map of southern California showing the geologic setting of Crystal Knob and its placement relative to key tectonic features, such as the dispersed Southern California batholith, Neogene dextral faults and the stalled Monterey microplate. Sampling locations for previous xenolith studies are shown: the Central and Eastern Sierran suites show a record of delamination of a batholithic root [*Ducea and Saleeby*, 1996] and Mojave sites show underplating of Farallon-plate lithospheric nappes during the Cretaceous [*Luffi et al.*, 2009]. The position of Crystal Knob is also shown, along with its reconstruction for dextral offset on the Neogene San Andreas transform system. Independent reconstructions using the regional paleomagnetic framework of *Wilson et al.* [2005] and restoration of slip along San Andreas--system faults [*Dickinson et al.*, 2005] agree to within 5 km on the position of the Crystal Knob source locale at 19 Ma see also **¿fig:reconstruction?**. Crystal Knob can be restored to ~350 km SE of its current location, accounting for ~310 km displacement on modern San Andreas Fault and ~40 km remainder on the Rinconada fault within the area of Salinian nappes.

## microscope-images

Optical petrographic images (2.5 mm wide field of view) showing characteristic textures found in the Crystal Knob sample set. (a) shows sample CK-D2, with the edge of a cumulate xenolith composed of equant olivine (ol) grains at ~200 µm characteristic scale, set against a host lava groundmass containing <100 µm phenocrysts of olivine, pyroxene, and plagioclase feldspar. (b) shows the spinel lherzolite sample CK-4 with >2 mm olivine, orthopyroxene (opx), clinopyroxene (cpx), and spinel (sp). (c) shows sample CK-D2, with a single large orthopyroxene crystal with augite exsolution lamellae and containing an olivine inclusion juxtaposed against dunite cumulate material consisting of mosaic-textured olivine grains.

## reconstruction

Tectonic reconstruction of the California margin at 19 Ma showing the early evolution of the San Andreas transform system, offshore oceanic plates, and the reconstructed location of modern exposures of the Cretaceous batholithic belt, the disaggregated Mojave--Salinia batholith, and surface outcrops of subduction channel schists in the Mojave province. Reconstruction of displaced features such as Salinia batholithic exposures from *Schott and Johnson* [1998], *Schott and Johnson* [2001], *Chapman et al.* [2012], and *Dickinson et al.* [2005] is combined with reconstruction of the evolving slab window and microplate detachment after *Wilson et al.* [2005]. The Monterey plate ridge ceased spreading at 20 Ma (Chron 6), which is labeled and corresponds to the similar time label on Figure **¿fig:neogene\_sections?**‌b.

## isotopes

Paired Sm-Nd and Rb-Sr isotope data for the Crystal Knob sample set contextualized relative to major Earth reservoirs. The position of Crystal Knob within the "depleted mantle" field suggests that the mantle lithosphere underlying coastal California was sourced directly from the mantle, either at a mid-ocean ridge or by direct underplating.

## cpx\_profile

Profile of Mg# measured across clinopyroxene phenocryst in the host lava sample CK-1. The grain has a partially cannibalized and fractured xenocryst core with Mg# , surrounded by successive layers with lower Mg# corresponding to crystallization in a progressively evolving magma, and shows the complex fractionation history of the magma.

## textures

Mineral classification images of each sample (1" round thin-section) created atop coregistered electron backscatter and optical imagery that show the textural variation within Crystal Knob suite. These classifications form the basis of the modal abundance measurements presented in Figure **¿fig:modes?**.

## step\_heating

Step-heating results for / dating of the Crystal Knob host basalt, showing a broad plateau for the accepted age of 1.65 Ma.

## major\_elements

1. FeO vs. MgO for electron microprobe measurements of grain cores, showing range in major-element depletion between samples. Dotted lines show Mg# levels.
2. Mg# (measured with a total iron basis) vs. for the silicate phases in the Crystal Knob peridotite xenoliths. This shows the range in major-element depletion between samples, including the low Mg# of CK-6, the most fertile sample.
3. Spinel Cr# vs. Mg# showing grouping of samples by Cr content, corresponding to different levels of depletion. Mg# is corrected based cation charge balance. The highest-Cr samples (CK-3 and CK-4) are harzburgite residues, while the CK-6 is an apparently refertilized lherzolite.

## ca\_in\_olivine

Calcium abundance in olivine for xenolith samples, showing the per-sample separability of microprobe analyses despite low measured abundances.

## spinel\_cr

Spinel Cr# vs. Mg# showing two groups of samples with low and high Cr content, corresponding to different levels of depletion. Mg# is corrected based cation charge balance. The highest-Cr samples (CK-3 and CK-4) are harzburgite residues, while the CK-6 is an apparently refertilized lherzolite.

## whole\_rock\_major

Major element composition (oxide %, normalized to 100%) of xenolith samples recalculated from modal mineralogy.

## modes

Modal composition of Crystal Knob perodotites. Abyssal [*Asimow*, 1999; *Baker and Beckett*, 1999] and Dish Hill [*Luffi et al.*, 2009] peridotite compositions are shown for comparison.

## trace\_elements

1. Chondrite-normalized pyroxene rare-earth element abundances showing the range in depletion and re-enrichment in the Crystal Knob sample set.
2. Element-ratio proxies for depletion and re-enrichment of clinopyroxene rare-earth elements, showing that samples have a range of depletion characteristics and a variety of re-enrichment patters.

## ree\_model

1. Recalculated whole-rock trace elements for xenolith samples Table **¿tbl:trace\_elements\_table?** presented with best-fitting modeled compositions for depleted peridotite and enriching fluid, using the model discussed in text. Normal mid-ocean ridge basalt (NMORB) rare-earth composition is from *Sun and McDonough* [1989], and the range presented for alkali basalts is compiled from a suite of Mojave-desert samples measured by *Farmer et al.* [1995]. These fields are presented for comparison with the modeled composition of the enriching fluids, which closely resemble alkali basalt for all samples. The calculated re-enriching fluid composition for sample CK-2 is not shown because this sample is not depleted relative to modeled values.
2. REE depletion and re-enrichment trends for xenolith samples derived from modeling in (a). For all samples, <1% assimilation of alkali-basalt-like melt is required to explain the observed trends in re-enrichment of rare-earth elements.

## cpx\_literature\_comparison

Clinopyroxene trace elements for Crystal Knob compared to abyssal peridotite data compiled by *Warren* [2016]. The Crystal Knob samples show mild to moderate depletion in HREEs characteristic abyssal peridotites, but samples CK-3, CK-4, and CK-6 show re-enrichment of LREEs that is not seen in the abyssal peridotite dataset, implying that these samples saw a second phase of enrichment after creation.

## ree\_temperatures

1. Per-element equilibrium temperatures for REE thermometry of xenolith samples. Horizontal lines represent a projection of the best-fitting line representing the equilibrium temperature for each sample. Data points far from the horizontal line signify disequilibrium between pyroxene phases, and those outliers plotted with open circles are excluded from the fit.
2. Best-fitting REE temperatures for each sample with Gaussian error bounds, plotted against a kernel density distribution of TA98 temperatures. Joint error distributions are created using a Monte Carlo approach for both error distributions. This approach shows significant disequilibrium in Eu and across LREE for sample CK-4. The samples can be grouped into two temperature cohorts, with all samples, especially the low-temperature group, agreeing well with the TA98 thermometer.

## temp\_comparisons

Comparison of results from pyroxene major-element thermometers. (a) Core and rim measurements (filled and open circles, respectively) for each sample using the *Taylor* [1998] thermometer. Samples CK-3, CK-4, and CK-7 show heating along grain rims, and the samples vary in the tightness of within-sample temperature scatter. (b) The strong linear relationship between the TA98 and BKN themometers is shown, with BKN consistently measuring temperatures 30-70º higher. (c) Ca-in-orthopyroxene temperatures against TA98, showing the reproduction of two clear temperature cohorts around 980 and 1080ºC by this thermometer for grain cores.

## temp\_summary

Summary of temperature and depletion-degree data for the peridotite xenoliths showing the two temperature cohorts of the dataset, which remain separable for all thermometers and are centered roughly 80ºC apart. Estimates for all thermometers track together except for the higher temperature estimates for the REE thermometer for samples CK-4 and CK-6, which may reflect a fossil heating event. Spinel Cr# and several melt-extraction proxies are used to assess the level of depletion of the samples. The lower-temperature samples are generally less depleted in all systems, with variable MgO and depletion reflecting low modal abundances of pyroxene. Sample CK-6 has moderately high spinel Cr# but low levels of whole-rock depletion, suggesting bulk assimilation of an enriching fluid.

## depth

Summary of depth constraints for the xenolith samples. Depths from Ca-in-olivine geobarometry are plotted against TA98 temperature. A series of steady-state conductive geotherms for values of surface heat flow are plotted beneath the data, and the hatched region represents the bounds of the potential xenolith source region assembled from the Lines show per-sample maximum emplacement depths calculated using the expanded stability of high-chromian spinel [*O’Neill*, 1981] with error bars of 0.15 GPa. The synthesis of this data suggests that the samples were sourced from ~45--70 km depth.

## model\_results

Temperature-depth profiles through the crust and upper mantle at key timesteps during the evolution of the three tectonic scenarios. Each plotted profile represents a different model run based on the same scenario. (a) presents a shallow slab window scenario, with underplating of upwelling asthenosphere truncating a forearc geotherm at 24 Ma. This asthenosphere is held against the base of the crust from 0--6 Myr, accounting for the spread of models in the second panel. The final panel tracks all models to the present. (b) shows stalled slabs of different ages, with panels corresponding to shared tectonic events, modeled at different times based on the timing of subduction and age of oceanic crust. Subduction is bracketed by T and T, with T = T -- 1.04 Myr for all cases. The youngest and hottest of these runs corresponds to the "Monterey plate" tectonic scenario. (c) tracks Farallon Plate mantle lithosphere emplaced beneath the central California coast by mantle duplexing during the late Cretaceous [Figure **¿fig:cross\_sections?**] and reheated by a pulse of heat from below during the Miocene slab window [Figure **¿fig:neogene\_sections?**]. The second panel, at the end of subduction, shows the geologic temperature constraints used to tune the model to subduction conditions on the late-Cretaceous megathrust [e.g. *Ducea*, 2003; *Kidder and Ducea*, 2006] In this scenario, oceanic lithosphere is 55 Myr old at the time of subduction.

## model\_tracers

Temperature-time tracers for each modeled scenario shown in Figure **¿fig:model\_results?**, following the evolution of particles at final depths of 40 and 75 km in the model domain (dashed and solid lines, respectively), bracketing the boundary conditions of the Crystal Knob xenolith suite. All models conclude at 1.65 Ma, the eruptive age of the Crystal Knob xenoliths. (a) shows a scenario corresponding to upwelling-driven mantle lithosphere replacement to the base of the crust during the Mendocino slab window episode beginning at 24 Ma [*Wilson et al.*, 2005]. Asthenospheric mantle is held at the base of the crust for a period of time between 0 and 6 Myr to represent potential durations of active convection. (b) shows a range of scenarios corresponding to oceanic lithosphere slices underplated at different times during the subduction of the Farallon plate until its cessation in the Neogene. The youngest of these scenarios corresponds to the likely thermal evolution of a Monterey Plate stalled slab. The models with subduction times >30 Ma are included for completeness, though none of these can be linked to geologic features of the margin as well as the Cretaceous underplating and Monterey plate scenarios. Model tracers begin at 10 and 45 km beneath the seafloor and are advected to depths of 40 and 75 km during subduction over the first 1.04 Myr of the model run. (c) tracks a Farallon-plate slab subducted and underplated during the late Cretaceous, as envisioned in Figure **¿fig:cross\_sections?**. While similar to the older models of **B**, it is tuned for key thermobarometric constraints on subduction channel schists [*Kidder et al.*, 2003]. The effect of Mendocino slab window upwelling at the base of this section is shown, with timing equivalent to the replacement of the entire mantle lithosphere represented in (a).

## model\_comparison

Comparisons of Crystal Knob eruptive (1.65 Ma) sub-Salinia geotherms for each of the modeled scenarios. The profiles corresponding with a young underplated slab (Monterey-plate equivalent), wholesale mantle-lithosphere replacement by the slab window, and a Cretaceous Farallon slab both with and without deep slab-window reheating are shown. The results as shown are purely conductive geotherms in the absence of erosion and thus might be biased towards lower temperatures relative to measured xenolith pressure--temperature constraints [see Supplementary Materials].

## cross\_sections

Cross sections showing the evolution of southern California during the subduction of a large oceanic plateau during the late Cretaceous, and underplating of Farallon-plate mantle nappes during slab rollback [*Chapman et al.*, 2010; *Luffi et al.*, 2009; after *Saleeby*, 2003].

## neogene\_sections

Schematic cross-sections showing potential scenarios for modification of the marginal mantle lithosphere at the end of subduction in the early Miocene. (a) Migration of the East Pacific mantle upwelling beneath the continental margin, forming a slab window and causing wholesale replacement of sub-Salinia mantle lithosphere with ascended asthenosphere. (b) Translation of the Monterey plate stalled slab along the former subduction megathrust to a current position beneath the California Coast Ranges [after *Bohannon and Parsons*, 1995] (c) Mantle lithosphere beneath the Crystal Knob eruption site composed of underplated Farallon plate mantle nappes reheated at their base by the Neogene slab window. The Monterey plate fragment is translated along the Hosgri fault from the Transverse Ranges region, to have its lithospheric column juxtaposed with previously underplated Farallon-plate mantle lithosphere.

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