The Crystal Knob volcanic neck in the central California Coast Ranges was erupted during the Pleistocene (1.65 Ma) through the Nacimiento belt of the Franciscan complex in the coastal region of central California. It is composed of olivine-plagioclase phyric basalt containing spinel peridotites and cumulate dunite xenoliths. The peridotites sample the mantle lithosphere beneath the Nacimiento Franciscan, which was constructed as an accretionary prism during Late Cretaceous subduction of the Farallon plate. The Crystal Knob locality is of further interest because it lies in immediate proximity to crystalline nappes of the “Salinia terrane”, which are traditionally interpreted as a rooted crustal sliver, but which more recent studies show are detached and lying above the Franciscan complex.

Six spinel peridotite samples were analyzed, ranging from fertile lherzolites to clinopyroxene harzburgites with modal clinopyroxene from 2-13%. They have a depleted mantle Sr and Nd isotopic signature and correspond to abyssal peridotites, most likely sourced in a relict subducted slab. Ca-exchange geothermometry shows equilibration temperatures between 950 and 1060 ºC. Rare-earth geothermometry shows similar results, but hotter for the hottest and deepest samples. Broadly the samples can be divided into two groups based on temperature of equilibration, which also manifest in trace-element patterns, Cr content in spinels, and Ca in olivine. Ca-in-olivine barometry and phase stability limits suggest that the xenoliths were sourced along a depth gradient in the 45-75 km range. This corresponds to regional geothermal gradients ranging from 70-90 mW/m2, broadly in agreement with regional heat flow measurements. The samples show variable depletion, with residual harzburgites at deeper depths. All samples show pronounced trace element re-enrichment.

Given constraints from xenolith thermobarometry, forward modeling of thermal evolution is used to construct a range of potential thermal structures that are tested against plausible tectonic scenarios for the evolution of the mantle lithosphere beneath the region. Underplated Late Cretaceous Farallon plate floored by a deep slab window, and underplated Neogene Monterey microplate equally satisfy the thermal constraints, but a shallow Neogene slab window is excluded. Underplated Farallon plate best satisfies the crustal geological constraints. Given this, the preferred tectonic scenario for the construction of the region’s mantle lithosphere is subduction underplating of Farallon plate during the latest Cretaceous followed by cooling at depth over the Paleogene, and then superposed slab window formation at ~80 km depth during the Neogene. This scenario can explain the temperature and depth distribution of the Crystal Knob xenoliths, as well as the paired depletion and re-enrichment signatures recorded by most of the samples. This work shows that the mantle lithosphere beneath the Coast Ranges was constructed in a similar fashion as the mantle lithosphere beneath the Mojave Province; two regions that were in immediate proximity to one another prior to late Neogene-Quaternary dextral offsets along the San Andreas transform system.

# Introduction

The tectonic and petrogenetic processes by which Earth’s continental mantle lithosphere develops through time are of fundamental importance in geodynamics and Earth history. What are the relative roles of subduction accretion versus subduction erosion of mantle lithosphere fragments, and what petrologic-geochemical processes operate to ultimately stabilize mantle lithosphere beneath newly formed continental crustal tracts? Volcanic rock-hosted mantle xenolith suites offer vertical sampling columns that reveal the compositional and textural states of lithospheric sections at the time of xenolith entrainment and eruption. Petrogenetic studies of such xenolith suites offer constraints on depth intervals of entrainment, thermal gradients, and geochemical evolutionary states. The integration of such constraints with regional geophysical data, and with tectonic and petrogenetic studies of the overlying crust offer critical insights into the interplay of plate tectonic and geodynamic processes responsible for the current state of region specific lithospheric domains. 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 with regional geophysical data and modeling, and with a wealth of data on the tectonic and petrogenetic development of the overlying crust.

The Crystal Knob xenolith locality samples a highly strategic lithospheric column through the Late Cretaceous convergent margin belt of the SW North American Cordillera. This regionally extensive belt is characterized by a large-volume continental magmatic arc, generated as the Farallon oceanic plate subducted eastward beneath western North America [*Ducea et al.*, 2015]. The Franciscan complex represents the crustal level accretionary complex of this subduction zone, and is widely recognized for its tectonic inclusion of Farallon plate oceanic basement and pelagic sediment fragments, as well as upper plate magmatic arc derived siliciclastic sediments [*Blake et al.*, 1988; *Chapman et al.*, 2016a; *Cowan*, 1978; *Murchey and Jones*, 1984; *Sliter*, 1984] Recently, it has been recognized that a series of far out of place crystalline nappes that were derived from the southern California segment of the Late Cretaceous magmatic arc lie 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 and Saleeby*, 2013; *Kidder and Ducea*, 2006]. In aggregate these crystalline nappes have been called “Salinia”, or the “Salinian terrane” [*Page*, 1981]. 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 Salinia 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 the application of 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 various samples into petrographic groups [*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*, 1996, *Ducea and Saleeby* [1998a]; *Galer and O’Nions*, 1989; *Jové and Coleman*, 1998; *Lee et al.*, 2006, 2001; *Livaccari and Perry*, 1993; *Luffi et al.*, 2009; *Usui et al.*, 2003].

These studies suggest 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 erupted through the Cretaceous large volume batholith, and asthenospheric suites erupted 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. 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 posed by restoring its host crustal rocks to their position prior to San Andreas transform offset. This places them outboard of the northern reaches of the southern California batholith, the southern continuation of the Sierra Nevada batholith across the Garlock fault (Figure **¿fig:context?** and Figure **¿fig:reconstruction?**, detailed below). 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 ranges of the Mojave plateau [*Barbeau et al.*, 2005; *Chapman et al.*, 2012; *Saleeby*, 2003]. These deeply exhumed batholithic rocks all 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, 2012 *Chapman et al.* [2016b]; *Ducea et al.*, 2009; *Malin et al.*, 1995; *Yan et al.*, 2005]. The underplated metaclastic rocks are exposed in a series of 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 that was vigorously 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 effect 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 state of structural attenuation of the Salinia nappes, as well as adjacent (Salinia restored position) deeply exhumed batholithic rocks, derives from both shallow subduction megathrust displacements, and subsequent large magnitude trench-directed extensional faulting that correlates in time to the Shatsky conjugate progressing deeper into the mantle beneath the North American plate [*Chapman et al.*, 2012; *Liu et al.*, 2010; *Saleeby*, 2003].

The relatively shallow level of the tectonic underplating of the schists directly beneath deep crustal large volume batholithic rocks requires the prior tectonic erosion of the mantle lithosphere (mantle wedge) that hosted the source regime for the overlying batholith. 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*, 1996, *Ducea and Saleeby* [1998a]; *Lee et al.*, 2001, *Lee et al.* [2006]; *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 traversing towards the Garlock fault, a continuous gradient to deep levels (10 kb) is exposed [*Nadin and Saleeby*, 2008]. At these deep levels the structural base of the batholith consists of the normal sense remobilized shallow subduction megathrust, beneath which lie the underplated schists [*Chapman et al.*, 2010, *Chapman et al.* [2012], *Chapman et al.* [2016b]; *Saleeby*, 2003]. Seismic reflection data image the megathrust as effectively flat beneath the western Mojave plateau [*Yan et al.*, 2005], and dipping ~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 tectonic erosion of sub-continental mantle lithosphere (including Cretaceous mantle wedge), and the underplating of Farallon plate mantle lithosphere [*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 the neck having penetrated the Franciscan accretionary complex, clearly poses the question of the Crystal Knob suite having sampled additional underplated Farallon mantle nappes, in structural sequence with 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] pose another viable possibility 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 compositions much closer to that of the convecting mantle than the Cretaceous mantle wedge suite from the central Sierra (*Ducea and Saleeby* [1996], *Ducea and Saleeby* [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. The xenolith sites within the eastern Sierra suite were erupted within the <10 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 are correlated to the opening of the Pacific-Farallon slab window [*Atwater and Stock*, 1998; *Wilson et al.*, 2005]. Both the eruption of the Crystal Knob neck 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, in conjunction with 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]. 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 texturally they grade into single grain xenocrysts, or the apparent phenocrysts, for the host lava. The spinel peridotites are volumetrically subordinate to the dunites. As discussed below, the peridotites lack textural features suggestive of igneous cumulate origin, and in conjunction with the compositional data presented below we interpret the peridotites as mantle lithosphere in origin.

Samples were collected from the Crystal Knob lava with an emphasis on volumetrically minor 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 (for example, see *Ducea and Saleeby* [1998b]). The age of the Crystal Knob host lava was determined using the / technique on phenocrystic plagioclase. A billet of the host lava (sample CK-1) containing visible plagioclase phenocrysts 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?**], and 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 (/ = 298.56) [*Cosca et al.*, 2011].

Data for stepwise heating 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, for which the entire spectrum define a similar age, within error, of 1.71 Ma. From these age data 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 probe current of 15 kV. 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 phenocrysts of feldspar, clinopyroxene, and olivine. The sample also contains dunite and multiphase peridotite fragments ranging from aggregates of a few grains to ~5 cm diameter.

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 for different 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 Crystal Knob basalt.

The Crystal Knob basalt 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. 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 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 quickly, perhaps as the flow cooled.

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?**]. Thus, olivine grains in the host lava range from true phenocrysts, to cumulates from various intermediate stages of magma evolution and entrained xenocrysts corresponding to entrained peridotites.

## 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 boundaries 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 particularly strong graphic exsolution lamellae Figure **¿fig:microscope\_images?**‌c.

Sample CK-6, the most fertile in the sample set, shows abundant pyroxene exsolution, exhibiting both graphic and vermicular textures at small scale. 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:minerals?** 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. Variation in Mg# between samples indicates differences in melt-extraction history between the samples.

Silicate phases within CK-3 and CK-4 have Mg# distributions that suggest a residual composition, while the other samples have more fertile compositions. Samples CK-2, CK-5, and CK-7 cluster tightly in Fe-Mg space, with high Mg#s indicative of fertile compositions. Sample CK-6 has the lowest Mg#s, with values as low as 87 for clinopyroxenes. However, the sample has high abundances of iron relative to the other fertile peridotite samples, with olivine iron compositions ranging up to 8%.

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 stoichiometric balance: excess Fe is removed from the octahedral site and added to the tetrahedral until with 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:spinel\_cr?**. Sample CK-2 has additional scatter imparted by this correction but the rest seem to have homogeneous oxidized iron contents within-sample. The residual samples (CK-3 and CK-4) show high Cr# indicative of residues of melt extraction, while the fertile samples show low Cr#s. Sample CK-6 has an intermediate Cr# possibly indicative of a somewhat depleted sample and related to its high iron content Figure **¿fig:major\_elements?**‌a.

The phase compositions of the "fertile" samples are quite consistent, as are those of CK-3 and CK-4 (the more depleted samples). CK-6, with silicate-phase signatures of enrichment combined with high-iron compositions and chromian spinels, potentially demonstrates depletion during partial melting followed by assimilation of a higher-iron enriching component.

### Modal mineralogy

The peridotite 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 shown in Table **¿tbl:modal\_mineralogy?** and Figure **¿fig:modes?**.

The samples range from lherzolites to clinopyroxene harzburgites and are dominated by olivine and orthopyroxene. All samples contain minor (<1%wt) spinel. CK-2 has the most fertile composition, with 12.2%wt clinopyroxene. CK-3 is the least fertile sample, with 0.91%wt clinopyroxene. Weight percents derived from olivine modes range from 65 to 75%. Grain size varies between samples but generally has a characteristic scale of 200 µm. Some larger crystals are evident -- the harzburgite CK-3 contains 2 mm orthopyroxene porphyroblasts. All samples are Type I peridotites in the *Frey and Prinz* [1978] classification system.

### Whole-rock composition

Whole-rock major-element abundances are reconstructed from averaged mineral composition and estimated modes. Representative mineral and recalculated whole-rock compositions are tabulated in Table **¿tbl:minerals?**.

Whole-rock Mg# (molar Mg/(Mg+Fe) 100) 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, which show variation that mirrors that of the silicate phases. A range of spinel Cr# (molar Cr/(Cr+Al) 100) from 10 to 27 implies variation in degree of partial melting between samples [*Dick and Bullen*, 1984].

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 higher Mg#, with whole-rock Mg# greater than 90. CK-6 has a whole-rock Mg# < 88 and contains substantially more Cr and Al than the other samples. Though CK-6 is generally the most enriched in volatile elements, sample CK-2 contains substantially more Ca and Na.

The coincident Mg# and low spinel Cr# of CK-2, CK-5, and CK-7 implies that they are relatively fertile peridotites, with low melt volumes extracted. The variation in whole-rock composition within these samples is due to a change in the abundance of pyroxene phases, with sample CK-2 containing substantially more pyroxene than CK-5. The higher Mg# of samples CK-3 and CK-4 corresponds to their status as residues of high-degree partial melting, which is also evident from their harzburgite classification in modal mineralogy. For sample CK-6, the combination of high spinel Cr# (implying depletion by partial melting) and low Mg# (a marker of major-element fertility) suggests that this sample was impacted by a multistage history of depletion and re-enrichment in major elements. Assimilation of a more fractionated fluid might explain the higher iron values as well as the high pyroxene modes, with pyroxene forming due to the addition of silica to a more olivine-rich, residual component.

## Rb-Sr and Sm-Nd isotopes

Portions of each peridotite sample were crushed using a shatterbox at the California Institute of Technology. Clinopyroxene grains (150--300 µm, 35-45 g 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. The samples were spiked with mixed Sm-Nd tracers [*Wasserburg et al.*, 1981]. Samarium was analyzed using a static routine on a 54 VG Sector multicollector thermal ionization mass spectrometer (TIMS), and neodymium was measured as an oxide on a 354 VG Sector 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?**. Graphical results for pyroxene REEs are shown in Figure **¿fig:spider?**, and estimated whole-rock modes are shown in Figure **¿fig:ree\_model?**.

Clinopyroxene, orthopyroxene, and recalculated whole-rock rare-earth elements show several modes of variation between samples, corresponding to different amounts of depletion and re-enrichment. All samples show clear evidence of rare-earth element depletion, although the amount of this depletion varies. For clinopyroxenes, the samples show progressive depletion in rare-earth elements following patterns seen in abyssal peridotites [Figure **¿fig:cpx\_literature\_comparison?**].

### Modeling depletion and re-enrichment

Clinopyroxene REE patterns tell only part of the story of whole-rock depletion and re-enrichment. Combining pyroxene trace-element measurements with modal mineralogy suggests that pyroxene modes have changed significantly in response to depletion, making recalculated whole-rock trace elements 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.

A depletion model is constructed in *alphaMELTS* [*Smith and Asimow*, 2005] to illuminate the probable depletion and re-enrichment paths of the Crystal Knob sample [Figure a]. A generic model of peridotite depletion is constructed, in which a parcel of material starting at a mantle potential temperature of 1300ºC at 2.0 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%. These starting parameters were chosen to provide the best correspondence with the overall experimental dataset, but a wide range of initial starting parameters (up to 1500ºC at 3.0 GPa) provide similar results.

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. Since HREEs are not likely to be easily modified by late re-enrichment due to their low diffusion rates, the best-fitting model step is used as an estimate of single-stage depletion of the samples during decompression melting.

The difference between the sample composition and this fitted depleted profile is taken as the contribution from batch addition of an enriching fluid. The net amounts of REE added during re-enrichment are then normalized to an average of 12primitive mantle for HREE, reflecting the average HREE concentrations in likely enriching fluids including normal MORB [*Sun and McDonough*, 1989] and alkali basalt [*Farmer et al.*, 1995]. The slope of the resulting normalized profile is diagnostic of the trace-element profile of the re-enriching agent. The normalization factor employed to shift the composition of enriching fluids to this value is shown in Figure b and corresponds roughly to the amount of LREE added during re-enrichment.

The results of this model show that the samples are variably depleted and all except CK-2 are re-enriched to some extent [Figure **¿fig:ree\_model?**]. CK-2 appears to be in equilibrium with mid-ocean ridge basalt while the others appear to be in equilibrium with an enriching agent similar to alkali basalt. The strong exponential increase in the presumed LREE composition of our enriching material shows that the re-enrichment may be better modeled as a fractional process. It is likely that all samples experienced a small and similar amount of interaction with the enriching fluid Figure **¿fig:ree\_model?**‌b.

Primary depletion degrees of the xenolith samples are estimated by finding the 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:spider?**] 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 that the degree of re-fertilization is similar for all samples, with less than 1% melt assimilation in all cases (assuming a somewhat enriched fluid with HREE similar to MORB or alkali basalt). However, the more residual samples with 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 pattern of trace elements suggests that the xenolith samples are residues of progressive fractional melting of primitive mantle to form abyssal peridotites [*Johnson et al.*, 1990]. The samples underwent a multistage history of wholesale REE depletion (due to higher-degree melting) followed by later LREE re-enrichment by an enriched fluid. This general pattern is similar to that gained by emplacement at a mid-ocean ridge followed by refertilization by off-axis magmatism [*Luffi et al.*, 2009]. However, it may also have arisen during melt extraction and entrainment prior to eruption. The latter seems to demand a significant residence time of the hotter xenoliths in a magma chamber at depth to allow LREE 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 Al 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 measured on the microprobe 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 the thermometer, and the relative performance of different thermometers can be used to assess the recovery of absolute temperatures.

Within-sample variation in temperatures 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 for temperatures in 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 generally yields results coincident with BKN temperatures (sensibly, 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 of measurements 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 the bulk of the temperature signature is based on the equilibrium state of the sample.

### 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 fluid 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 interaction, lacks high-temperature rim compositions, implying more sustained equilibration with 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, as well as in other geochemical data.

This temperature distribution likely corresponds to the sourcing of these 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:depths?**.

## 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 element and Y, equivalent to . These per-element equilibration temperatures are shown in Figure **¿fig:ree\_temperatures?**. The best-fitting line from the origin through each point in vs. space (using a robust regression with a Tukey biweight norm) yields the equilibrium temperature for each sample. Significant outliers from the fit are excluded from the thermometry, and may represent effects of disequilibrium processes.

### Pyroxene rare-earth disequilibrium

CK-3 shows disequilibrium 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 an equilibrium signature.

The pattern of disequilibrium in sample CK-4 suggests that 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. The shape of this disequilibrium relationship may be traceable 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 more being incorporated into orthopyroxene, showing incomplete REE diffusion throughout the sample. This is consistent with late re-enrichment in LREEs from the Crystal Knob source magma.

All samples except CK-6 and CK-7 show results off the linear trendline for Eu. This distinct disequilibrium was also found in calibration by *Sun and Liang* [2012], and is dependent on the oxygen fugacity (and / ratio of the host magma. The exact kinetics of this scenario are unclear, but it is likely that REE equilibrium was achieved in a reducing environment *(P. Asimow and J. Blundy, personal communication, 2016)*, or Eu rapidly and differentially diffused out of the Crystal Knob xenoliths just prior to eruption. These patterns could also be due to the effect of "ghost" plagioclase on creating local europium enrichments and depletions in the neighborhood of resorbed plagioclase grains. Such a pattern would suggest that the xenoliths originated at shallow mantle lithosphere levels and was 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 rare-earth elements 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. Alternatively, partition coefficients for LREE and Eu could have been modified by an unusual late-stage melt interaction favoring the extraction of LREE and Eu from clinopyroxene. 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. Despite this disequilibrium, temperature estimates anchored by HREE perform well as measured against major-element thermometry, even for samples (such as CK-4) with generalized LREE disequilibrium.

### Comparison with major-element thermometry

Rare-earth exchange thermometry shows the samples as divided into the same two groupings of temperatures as those found by major-element thermometry. 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], it seems likely that both the TA98 and REE temperatures 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 the TA98 method. Rare-earths in pyroxene diffuse several orders of magnitude slower than 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 heating event primarily affecting the deepest samples. This is accompanied by major REE disequilibrium in sample CK-4, which also contains the most intergranular melt channels.

Sample CK-4 records a significantly higher temperature for LREE than both the other samples and its own HREE equilibration temperatures. This likely shows that the LREEs were equilibrated at a much higher temperature than the 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 in the sample set. It is likely, therefore, that CK-4 was subjected to a short, transient heating event that was not fully equilibrated in HREE. Further, since this transient heating is not reflected in major-element temperatures, it is likely that the sample was subsequently equilibrated at a lower temperature for a significant period of time. Also, given low clinopyroxene modes, rare-earths added during refertilization may not be completely homogenized throughout the sample.

## Geochemical variation within the Crystal Knob sample set

All of the Crystal Knob xenoliths analyzed are isotopically depleted, with an initial of +10, and / of .7029. This corresponds to the depleted upper mantle [e.g. *Hofmann*, 1997], with a mantle upwelling source that has seen no contribution from the western North American crust or continental lithosphere more generally. The mantle lithosphere sampled by the Crystal Knob suite was created from the same convective mantle reservoir and is distinct from the overlying crustal material.

However, the two cohorts in temperature seem to sample mantle material with somewhat different major- and trace-element characteristics. In major elements, the low-temperature cohort generally includes phases with relatively fertile compositions, with modal abundances showing a trend towards depletion by decrease in the abundance of pyroxene phases. This is supplemented by relatively undepleted pyroxene trace-element patterns: In clinopyroxene trace elements, the low-temperature cohort ranges from essentially undepleted to low levels of depletion characteristic of the least-depleted abyssal peridotites [Figure **¿fig:cpx\_literature\_comparison?**].

The high-temperature samples show major-element compositions with lower Mg#s, suggesting that depletion progressed differently, with all phases losing volatile elements during depletion. Within this cohort, CK-3 and CK-4 show significantly higher levels of REE depletion and extremely low clinopyroxene modes, as well as distinct enrichments in clinopyroxene LREEs, which likely correspond to assimilation of enriching fluid and are notably different from profiles found in abyssal peridotites [Figure **¿fig:cpx\_literature\_comparison?**]. They also contain chromian spinels, which are a product of increased levels of melt extraction.

Phases within sample CK-6 shows uniformly low Mg#s, suggesting depletion. However, high modal abundance of pyroxene phases suggests some amount of major-element re-enrichment after formation. This is bolstered by petrographic evidence of significant growth and aggregation of pyroxene grains, unique in the sample set. It is relatively undepleted in REEs and shows no significant temperature disequilibrium in any thermometer including REE. This major-element refertilization is likely a result of assimilation of mantle material and re-equilibration near the solidus. It implies significant mass exchange and may be due to interaction with deep upwelling fluids.

Broadly, the results of geochemical and thermometric studies of the Crystal Knob suite suggest that the mantle lithosphere beneath central California was affected by differing levels of depletion and re-enrichment. The shallower samples are less depleted, and the hottest and deepest samples are melt residues, implying more melt extraction at deeper levels of the lithospheric column. Modeling suggests that re-enrichment happened throughout the sample set, and all clinopyroxene rare-earth elements show effects of re-enrichment except for CK-2. However, disequilibrium in REE (CK-4) and unusual major element signatures (CK-6) occur in the hotter samples. This suggests that the samples were affected by alteration that was the most intense at deeper depths sometime in their post-emplacement and pre-eruptive history.

## 

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 determined 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 80 km.

Several of the techniques below produce estimates of pressure, rather than depth. We use a geothermal gradient based on integration of the crustal and mantle densities given in Table **¿tbl:model\_parameters?**. This yields a gradient of ~0.03 GPa per km across the mantle lithosphere.

### 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 **¿sec:regional\_structure?**. Another minimum depth constraint is the plagioclase--spinel peridotite facies transition, which occurs at depths of 20-30 km [*Borghini et al.*, 2009; *Green and Ringwood*, 1970].

The high-pressure boundary of spinel stability limits maximum possible entrainment depths. The spinel--garnet peridotite phase 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], a slightly deeper estimate than *O’Neill* [1981]. 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 given the unconstrained assumption of ideal garnet--spinel mixing, a spinel-weighted metastable assemblage is possible even at higher pressures.

As shown in Figure **¿fig:spinel\_cr?**, 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 chief differences from a simple CMAS model are focused at high temperatures and pressures. The rough estimate of the garnet-in pressure given by *O’Neill* [1981] is sufficiently accurate 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

Peridotite barometers are based on the decreasing Al content of orthopyroxene with depth [*Brey and Köhler*, 1990; *Nickel and Green*, 1985; *Nimis and Taylor*, 2000]. However, in the absence of garnet, the reaction is purely thermometric, with nearly vertical isopleths in P-T space [*Gasparik*, 2000; *Herzberg*, 1978].

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 cation 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 that 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 **¿sec:modeling?**]. This yields a slightly "hotter" geotherm throughout the mantle lithosphere.

The stability-field 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 age distributions broadly correspond to a range in surface heat flow from ~70 to 110 mW/m^2.

Using a database of surface heat flows for North America, *Erkan and Blackwell* [2009] estimates 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 is within the spinel stability field and falls near the center of the depth distributions extracted using Ca-in-olivine barometry. We discuss the accuracy of heat flow data further in Section 3.1

### Summary of depth constraints

The integration of depth constraints on the xenolith samples from multiple sources gives a broad set of constraints 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 45-70 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 of the subcontinental lithosphere-asthenosphere boundary occurring 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.

# Origin of the mantle lithosphere beneath Crystal Knob

Rb-Sr and Sm-Nd isotopic and trace-element data on peridotite xenoliths from this study demonstrate that the mantle lithosphere that was sampled by the Crystal Knob volcanic neck is sourced from the depleted convecting mantle [e.g. *Hofmann*, 1997] with no contribution from recycled crustal material, nor ancient sub-continental mantle lithosphere. This is consistent with the neck having penetrated through the Franciscan accretionary complex, and also with the observations that Salinia continental arc rocks of the region are unrooted nappes that lie structurally above Franciscan complex rocks.

In that the Franciscan complex of the region was assembled by long-lived subduction of the Farallon plate encompassing Cretaceous-early Tertiary time [*Chapman et al.*, 2016a; *Cowan*, 1978; *Saleeby and contributors*, 1986; *Seton et al.*, 2012], it follows that the mantle lithosphere of the region was constructed from partly subducted Farallon plate upper mantle at some point late in the Franciscan accretionary history, or by some other mechanism following the cessation of Farallon plate subduction with the establishment of the San Andreas transform system. Based on this 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.

## 

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" (CRTA) 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 flows in the Coast Ranges range from 60-90 mW/m^2. These are much higher than those observed in the adjacent Central Valley and Sierra Nevada, but are similar to those measured in the Mojave province. They are also on the relatively high end of the global range of regionally averaged continental heat flows, which range 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].

hesitant to invoke fluid flow to elevate the regionally averaged thermal gradient. 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.

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.

the thermal evolution predicted by the three tectonic scenarios discussed above. To support the evaluation of the possible tectonic scenarios for the rocks, we model the relaxation of the geotherm during subduction, underplating, and slab window opening.

## Late Cenozoic tectonic history and regional crustal structure

In Oligocene to early Miocene time the Pacific-Farallon spreading ridge obliquely impinged into 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 as well as 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 linked to slab window formation by the 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 formation, 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]. Late Cenozoic volcanism of the region, in this scenario, 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 upon closer analysis appears to require a combination of both 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 low east-dipping mafic lower crustal layer that extends beneath central California from the offshore region into proximity of the San Andreas fault, and which thickens 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 most logically is 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 III*, 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. In Figure **¿fig:reconstruction?**, 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 westwards across Nacimiento belt Franciscan an unknown distance [*Hall and Saleeby*, 2013], but have been eroded off their lower plate complex as the coastal region has risen 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 projections of the Pacific-Farallon slab window and partially subducted Monterey plate at ca. 19 Ma [*Wilson et al.*, 2005]. The slab window formed by the 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 edge of the Monterey plate. Diffuse volcanism, some clearly derived from decompression partial melting of convecting mantle, is widespread for 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. This "dangling" horizontally translated slab descends eastwards beneath the Central Valley, corresponding to the "Isabella" high-wave speed seismic anomaly [*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 that its downdip extension was translated effectively horizontally along, or sub-parallel to 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.

As shown in the Figure **¿fig:reconstruction?** reconstruction, the Crystal Knob eruption site was 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, as a result of its coupling to the Pacific plate. The likelihood of extensional attenuation of the underthrust portion of the Monterey plate during such divergent motion is non-explored, but strongly implied in the *Bohannon and Parsons* [1995] reconstruction. 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. 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 untenable notion that a structurally continuous mafic layer, representing the stalled Monterey plate, 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 the structural continuity that these studies show 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], Figure **¿fig:context?**. 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 transform offsets 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 as well [*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. As discussed above, *Luffi et al.* [2009] present findings on the Dish Hill and Cima mantle xenolith sites [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. In that the crustal structural sequence of the western Mojave region correlates closely to that of the Salinia nappes, spatially and temporally [*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 [after *Saleeby*, 2003, p.[@Luffi2009]]. This is shown to have 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, as applied to our thermal modeling presented in Section **¿sec:modeling?**, are integrated from *Kidder and Ducea* [2006];*Chapman et al.* [2010];*Chapman et al.* [2012];*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 down the subduction zone. Principal crustal responses are shown as large magnitude trench-directed extension coupled to regional extrusion of the underplated subduction channel schists, which was driven by suction forces of the retreating slab. In the Figure **¿fig:cross\_sections?**'c to d transition, accelerated rollback is driven by duplex formation from Farallon plate mantle nappes. We suspect that mantle nappe detachment was controlled primarily by the temperature control on 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, an estimated ~700-800 ºC control on this 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 the retreat of the slab as it subducted imparted a significant tensile stress component within the slab that was oriented at high angle to the subduction megathrust, which further promoted nappe detachment. The nucleation of detachment surfaces was likely controlled by hydration fronts that followed primary normal and transform faults within the upper oceanic lithosphere. The lack of high-pressure mafic schist samples in both the Crystal Knob and Dish Hill xenolith suites suggests that oceanic crust was detached during mantle nappe detachment, presumably at 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 the regional structural evolution of the central to southern California basement, and on the petrogenetic history recorded in the region's mantle xenolith suites, we consider the Figure **¿fig:cross\_sections?**‌d section to be that most likely sampled by the Crystal Knob eruption. This section is idealized for Late Cretaceous time, and below we layer on the complexity of late Cenozoic tectonics in our analysis.

## A deep slab window beneath relict lithosphere

## Kinematic

reconstructions of the impingement of the Pacific-Farallon spreading center with the SW Cordilleran subducting trench 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] have 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 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 50-80 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 Farallon Plate, Monterey Plate, and slab window 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, a forward model of the geotherm implied by each of the tectonic scenarios shown in Figure **¿fig:neogene\_sections?** is constructed. 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 modeled oceanic (or asthenospheric) geotherms and relaxed towards the present by iteratively solving the heat-flow equation using finite differences. The entire model is implemented in Python, with finite-difference modeling based on 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 model the evolution of a geotherm during subduction on a continuously subducting model. Standard values are used for oceanic and continental material properties, and are given in Table **¿tbl:model\_parameters?**. More information about model setup and integration is given in Section 7.

## 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 time of 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 geotherms too hot to correspond to the modern regional geotherm. We confirm this assessment, finding that this scenario produces extremely "hot" geotherms that are at the upper boundary of spinel lherzolite stability for much of the temperature domain of interest [Figure **¿fig:model\_comparison?**], reproducing neither the xenolith geotherm determined in this study nor the seismically-determined depth of the lithosphere-asthenosphere boundary [e.g. *Li et al.*, 2007].

### 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, reflecting the long subduction history of the Farallon plate beneath the central California coast through the Cretaceous and Paleogene. In principle, backstepping of the subduction megathrust and underplating of a slice of mantle lithosphere could have occurred at any time during this history. However, only the oldest and youngest stalled slab models correspond to geodynamic and geological evidence of a specific episode of subduction instability.

We model a series of scenarios with differently-timed underplating events, with the start of subduction ranging from 22 to 80 Ma. These subduction times, T, set the beginning of the models shown in Figure **¿fig:model\_tracers?**‌b and are shown in 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 the subduction time moves towards 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. In the oldest model with a subduction time of 80 Ma, the oceanic lithosphere at the time of subduction is 60 myr old, meaning that the oceanic crust in this model was generated beneath the Pacific--Farallon spreading ridge at 140 Ma.

Stalled slab scenarios with subduction ages as young as 30 Ma (all but the last scenario presented on Figure **¿fig:model\_tracers?**‌b) model rollback during sustained Farallon-plate subduction. The final 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 stalled 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 the 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 the 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 coincides with the entrainment constraints on the Crystal Knob xenoliths. Without consideration of potential bias towards colder measurements in the modeled geotherms, this appears to be the best model. When accounting for possible external effects [Section 4.3], it may predict a hotter geotherm than 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\_tracers?**‌c and Figure **¿fig:model\_results?**‌c. This scenario initiates in a similar fashion to the model runs for the stalled slab scenario Figure **¿fig:model\_tracers?**‌b with the oldest emplacement ages. 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 models 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 maximum 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.*, 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, and then the geotherm is perturbed by the underplating of asthenosphere by an ~80 km deep slab window. We present models with asthenosphere with an adiabatic temperature gradient held against the base of the lithosphere for periods ranging from 0 Myr (instantaneous contact followed immediately by conductive relaxation) to 6 Myr. The model for 6 Myr of sustained upwelling at the base of the lithosphere produces the "kinked" geotherm seen in panel 4 of Figure **¿fig:model\_results?**‌c at the 18 Ma time step, due to imposition of a mantle adiabat below 80 km depth. A single model without slab window heating highlighted in **¿fig:model\_comparison?** predicts much cooler geotherms that do not match the mantle geothermal constraints developed in this study.

Figure 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.

## 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]. This suggests that the high geothermal temperatures predicted for the younger stalled slab scenarios in model group **B**, including the Monterey plate scenario, may be overestimates.

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 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 [*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 **¿fig:model\_results?**‌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 **¿fig:model\_results?**‌c and Figure **¿fig:model\_comparison?**

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. Additional discussion of these factors can be found in Section 7.

## 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 (deep slab window reheated) Cretaceous mantle nappe scenarios recover the geotherm determined by xenolith thermobarometry, while not violating constraints posed by heat flow data.

Exhumation/erosion, while not accounted for in our models, pushes the modeled geotherm for the shallow slab window scenario outside of the limit of the thermobarometric constraints on xenolith entrainment, while pushing the Monterey plate endmember stalled-slab scenario towards the upper limit of this field, and aligns the reheated mantle nappe scenario with the center of the field. Unfortunately, we are prevented from incorporating the effects of exhumation/erosion due to the lack of temporally-defined geologic constraints on these processes that can be properly posed within the context of such modeling.

# Contemporary lithospheric structure and thermal state

In this section we integrate the results of our thermal modeling with our petrogenetic findings on the Crystal Knob xenoliths, regional crustal structure and evolution, and the timing and map position 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 shallow slab window emplaced asthenosphere case 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 a wide spectrum of geologic and geodynamic factors, that were 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 Monterey Plate underthrusting beneath the southern California borderland [Figure **¿fig:neogene\_sections?**], its structural integrity following its coupling to borderland transrotational rifting, and on what proportion of the partially subducted plate is now represented by 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 deformational responses to horizontal shear stresses in the lower crust. Such surface deformation patterns are not expressed for late Cenozoic time north of the Transverse Ranges. 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 of our findings on the Crystal Knob xenolith suite not being able to discriminate between the stalled Monterey plate, or underplated Farallon nappe cases, the stalled Monterey plate case creates more problems than it solves.

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 an 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]. As posited above, a thick and relatively cool lithospheric lid inhibited widespread voluminous volcanism in response to asthenospheric upwelling during the opening of the slab window. 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 Figure **¿fig:neogene\_sections?**‌c.

## Implications of petrologic complexities to thermal state

The Crystal Knob xenolith suite shows petrologic variation consistent with reheating from below. As discussed in Section 2.8.2, the LREE disequilibrium seen in several samples (especially CK-4) could be the signature of a fossil heating event not fully re-equilibrated in LREEs. This is underscored by the presence of melt-infiltration textures in sample CK-4. The apparent polyphase major-element refertilization in sample CK-6 as well as the apparently re-enriched clinopyroxenes across the sample set attest to the assimilation of varying amounts of material after the initial crystallization of these xenoliths, and re-equilibration of major-element compositions after this event.

The sourcing of the Crystal Knob pipe from relatively deep levels of the mantle lithosphere, the highly fractionated nature of the magma with zoned phenocrysts, 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 >50 km, such as that investigated by *Kelemen et al.* [2000], which includes fractal scaling of melt migration channels hosting cumulates, which are sometimes re-entrained by small-volume volcanism. Such a network is linked to progressive percolation of melt upwards 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 (e.g. the Morro Rock--Islay Hills complex [*Stanley et al.*, 2000] and the Cambria Felsite [*Ernst and Hall*, 1974]) and later deeply sourced small-volume eruptions such as Crystal Knob.

## Origin of the Crystal Knob basalt

The time lag between deep slab window opening and the ca. 1.7 Ma eruptive age of Crystal Knob presents a problem for the origin of the Crystal Knob lava. Our studies on 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 old 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], which yields an ~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 basaltic flows 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 that markedly post-dates any possible slab window opening, and 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 dominance of dunite cumulate xenoliths that appear to be related to the Crystal Knob lava, and that volumetrically far exceed the lithospheric peridotite xenoliths, attest to at least another reheating event at ca. 1.7 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.

# Conclusion

The lithosphere of southern California, to first order created by Cretaceous convergent margin tectonics, was severely structurally overprinted by two subsequent tectonic episodes, with the impact and subduction of the Shatsky Rise large igneous province conjugate during the Late Cretaceous followed by 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 constraints, 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. Geochemical 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.

# Supplementary file: modeling setup

## Model setups

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.

## Factors not incorporated into 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 **¿fig:neogene\_sections?**‌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 **¿fig:model\_comparisons?**‌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-Creteaceous 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 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, as well as its reconstruction for dextral offset on the Neogene San Andreas transform system. This reconstruction was created independently using the regional paleomagnetic framework of *Wilson et al.* [2005] with the restoration of slip along San Andreas--system faults [*Dickinson et al.*, 2005] approaches. The methods 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 Salinian block.

## 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 Cretaceous batholithic belt. 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. Exposures of Salinia nappes are shown as red patches west of the future San Andreas Fault. The view shows the disaggregated Mojave--Salinia batholith and surface outcrops of subduction channel schists in the Mojave province. Reconstruction of Salinian lithologic features 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].

## 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.

## ca\_in\_olivine

Calcium abundance in olivine for xenolith samples, showing the separability of each sample's cluster in the dataset.

## spinel\_cr

Spinel Cr# vs. Mg# showing two groups of samples with low and high Cr content, corresponding to the temperature cohorts of the dataset.

## 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.

## spider

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?** presented with best-fitting modeled compositions for depleted peridotite and enriching fluid per sample, 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.
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 data showing the two temperature cohorts of the dataset, which remain separable for all thermometers and are centered roughly 80ºC apart. The higher temperature estimates for the REE thermometer for samples CK-4 and CK-6 may reflect a fossil heating event.

## 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 ~50--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 a final depth of 40 and 75 km depth in the model domain (dashed and solid lines, respectively), bracketing the depth-domain 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 [*Wilson et al.*, 2005], in which the crust is underplated by asthenospheric mantle which convects for a period of time (several durations of active convection from 0-6 Myr are shown) beginning at 24 Ma. (b) shows a range of scenarios corresponding to oceanic lithosphere slices underplated at different times during the subduction history 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 older models shown in panel B 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. Modeled 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?**. It is similar to the older models of **B** but is based on key geologic constraints from subduction channel schists [*Kidder et al.*, 2003]. The effects of deep upwelling corresponding with the Mendocino slab window are shown for several of the models, in a manner similar to **A**.

## model\_comparison

Comparisons of "modern" (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 4.3.

## cross\_sections

Cross sections showing the evolution of southern California during subduction of a large oceanic plateau during the late Cretaceous, and underplating of Farallon-plate mantle nappes during slab rollback.

## 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 abyssal material. **B**: Translation of the Monterey plate stalled slab along the former subduction megathrust to a current position beneath the California Coast Ranges [after *Pikser et al.*, 2012] **C**: Translation of Monterey plate fragment along Hosgri fault after following slab breakoff in the Transverse Ranges region.

## monterey\_plate

Schematic representation of the Monterey plate dangling slab scenario for the origin of the sub-Salinian mantle lithosphere as envisioned by *Van Wijk et al.* [2001] and *Pikser et al.* [2012], among others.

Alibert, C. (1994), Peridotite xenoliths from western grand canyon and the thumb: A probe into the subcontinental mantle of the colorado plateau, *Journal of Geophysical Research: Solid Earth*, *99*(B11), 21605–21620, doi:[10.1029/94jb01555](https://doi.org/10.1029/94jb01555).

Argus, D. F., and R. G. Gordon (1991), Current sierra nevada-north america motion from very long baseline interferometry:Implications for the kinematics of the western united states, *Geology*, *19*(11), 1085, doi:[10.1130/0091-7613(1991)019<1085:csnnam>2.3.co;2](https://doi.org/10.1130/0091-7613(1991)019<1085:csnnam>2.3.co;2).

Armstrong, J. T. (1988), Quantitative analysis of silicate and oxide materials: Comparison of monte carlo, zaf, and phi-rho-z procedures, 239–246.

Asimow, P. (1999), A model that reconciles major-and trace-element data from abyssal peridotites, *Earth and Planetary Science Letters*, *169*, 303–319.

Atwater, T. (1970), Implications of plate tectonics for the cenozoic tectonic evolution of western north america, doi:[10.1130/0016-7606(1970)81](https://doi.org/10.1130/0016-7606(1970)81).

Atwater, T., and J. Severinghaus (1989), Tectonic maps of the northeast pacific, in *The geology of north america, vol. n, the eastern pacific ocean and hawaii*, edited by E. Winterer, D. Hussong, and R. Decker, pp. 15–20, Geological Society of America, Boulder, Colorado.

Atwater, T., and J. Stock (1998), Pacific-north america plate tectonics of the neogene southwestern united states: An update, *International Geology Review*, *40*(5), 375–402, doi:[10.1080/00206819809465216](https://doi.org/10.1080/00206819809465216).

Baker, M. B., and J. R. Beckett (1999), The origin of abyssal peridotites: A reinterpretation of constraints based on primary bulk compositions, *Earth and Planetary Science Letters*, *171*(1), 49–61, doi:[10.1016/S0012-821X(99)00130-2](https://doi.org/10.1016/S0012-821X(99)00130-2).

Barbeau, D. L., M. N. Ducea, G. E. Gehrels, S. Kidder, P. H. Wetmore, and J. B. Saleeby (2005), U-pb detrital-zircon geochronology of northern salinian basement and cover rocks, *Geological Society of America Bulletin*, *117*(3), 466, doi:[10.1130/b25496.1](https://doi.org/10.1130/b25496.1).

Barth, A. P., J. L. Wooden, M. Grove, C. E. Jacobson, and J. N. Pedrick (2003), U-pb zircon geochronology of rocks in the salinas valley region of california: A reevaluation of the crustal structure and origin of the salinian block, *Geology*, *31*(6), 517.

Beard, B. L., and A. F. Glazner (1995), Trace element and sr and nd isotopic composition of mantle xenoliths from the big pine volcanic field, california, *Journal of Geophysical Research: Solid Earth*, *100*(B3), 4169–4179, doi:[10.1029/94jb02883](https://doi.org/10.1029/94jb02883).

Blake, M. C. J., A. S. Jayko, R. J. McLaughlin, and M. B. Underwood (1988), Metamorphic and tectonic evolution of the franciscan complex, northern california, in *Metamorphism and crustal evolution of the western united states*, vol. 8, pp. 1035–1060, Prentice-Hall Englewood Cliffs, New Jersey.

Blundy, J., and B. Wood (2003), Partitioning of trace elements between crystals and melts, *Earth and Planetary Science Letters*, *210*(3-4), 383–397, doi:[10.1016/S0012-821X(03)00129-8](https://doi.org/10.1016/S0012-821X(03)00129-8).

Bohannon, R. G., and T. Parsons (1995), Tectonic implications of post-30 ma pacific and north american relative plate motions, *Geological Society of America Bulletin*, *107*(8), 937–959, doi:[10.1130/0016-7606(1995)107<0937:TIOPMP>2.3.CO;2](https://doi.org/10.1130/0016-7606(1995)107<0937:TIOPMP>2.3.CO;2).

Borghini, G., P. Fumagalli, and E. Rampone (2009), The stability of plagioclase in the upper mantle: Subsolidus experiments on fertile and depleted lherzolite, *Journal of Petrology*, *51*(1-2), 229–254, doi:[10.1093/petrology/egp079](https://doi.org/10.1093/petrology/egp079).

Brady, R. J., M. N. Ducea, S. B. Kidder, and J. B. Saleeby (2006), The distribution of radiogenic heat production as a function of depth in the sierra nevada batholith, california, *Lithos*, *86*(3-4), 229–244, doi:[10.1016/j.lithos.2005.06.003](https://doi.org/10.1016/j.lithos.2005.06.003).

Brey, G., and T. Köhler (1990), Geothermobarometry in four-phase lherzolites ii. new thermobarometers, and practical assessment of existing thermobarometers, *Journal of Petrology*, *31*(c), 1353–1378.

Brink, U. ten, N. Shimizu, and P. Molzer (1999), Plate deformation at depth under northern california: Slab gap or stretched slab?, *Tectonics*, *18*(6), 1084–1098.

Brocher, T. M., U. S. .. T. B. Brink, and T. Abramovitz (1999), Synthesis of crustal seismic structure and implications for the concept of a slab gap beneath coastal california, *International Geology Review*, *41*(3), 263–274, doi:[10.1080/00206819909465142](https://doi.org/10.1080/00206819909465142).

Burkett, E. R., and M. I. Billen (2009), Dynamics and implications of slab detachment due to ridge-trench collision, *Journal of Geophysical Research: Solid Earth*, *114*(12), 1–16, doi:[10.1029/2009JB006402](https://doi.org/10.1029/2009JB006402).

Bürgmann, R., and G. Dresen (2008), Rheology of the lower crust and upper mantle: Evidence from rock mechanics, geodesy, and field observations, *Annual Review of Earth and Planetary Sciences*, *36*(1), 531–567, doi:[10.1146/annurev.earth.36.031207.124326](https://doi.org/10.1146/annurev.earth.36.031207.124326).

Cecil, M. R., Z. Saleeby, J. Saleeby, and K. A. Farley (2014), Pliocene-quaternary subsidence and exhumation of the southeastern san joaquin basin, california, in response to mantle lithosphere removal, *Geosphere*, *10*(1), 129–147, doi:[10.1130/GES00882.1](https://doi.org/10.1130/GES00882.1).

Chapman, A. D. (2016), The pelona – orocopia – rand and related schists of southern california : A review of the best-known archive of shallow subduction on the planet, *International Geology Review*.

Chapman, A. D., C. E. Jacobson, W. Ernst, M. Grove, T. Dumitru, J. Hourigan, and M. N. Ducea (2016a), Assembling the world’s type shallow subduction complex: Detrital zircon geochronologic constraints on the origin of the nacimiento block, central california coast ranges, *Geosphere*, *12*(2), GES01257.1, doi:[10.1130/GES01257.1](https://doi.org/10.1130/GES01257.1).

Chapman, A. D., S. Kidder, J. B. Saleeby, and M. N. Ducea (2010), Role of extrusion of the rand and sierra de salinas schists in late cretaceous extension and rotation of the southern sierra nevada and vicinity, *Tectonics*, *29*(5), 1–21, doi:[10.1029/2009TC002597](https://doi.org/10.1029/2009TC002597).

Chapman, A. D., J. B. Saleeby, and J. Eiler (2013), Slab flattening trigger for isotopic disturbance and magmatic flare-up in the southernmost sierra nevada batholith, california, *Geology*, *41*(9), 1007–1010, doi:[10.1130/g34445.1](https://doi.org/10.1130/g34445.1).

Chapman, A. D., D. Wood, J. B. Saleeby, and Z. Saleeby (2016b), Late cretaceous to early neogene tectonic development of the southern sierra nevada region, in *Geological society of america fieldtrip guide*, vol. 7, Ontario, California.

Chapman, A., J. Saleeby, D. Wood, A. Piasecki, S. Kidder, M. Ducea, and K. Farley (2012), Late cretaceous gravitational collapse of the southern sierra nevada batholith, california, *Geosphere*, *8*(2), 314–341, doi:[10.1130/GES00740.1](https://doi.org/10.1130/GES00740.1).

Cole, R. B., and A. R. Basu (1995), Nd-sr isotopic geochemistry and tectonics of ridge subduction and middle cenozoic volcanism in western california, *Geological Society of America Bulletin*, *107*(2), 167–179, doi:[10.1130/0016-7606(1995)107<0167:NSIGAT>2.3.CO;2](https://doi.org/10.1130/0016-7606(1995)107<0167:NSIGAT>2.3.CO;2).

Cosca, M., H. Stunitz, A.-L. Bourgeix, and J. P. Lee (2011), 40Ar∗ loss in experimentally deformed muscovite and biotite with implications for 40Ar/39Ar geochronology of naturally deformed rocks, *Geochimica et Cosmochimica Acta*, *75*(24), 7759–7778, doi:[10.1016/j.gca.2011.10.012](https://doi.org/10.1016/j.gca.2011.10.012).

Cowan, D. S. (1978), Origin of blueschist-bearing chaotic rocks in the franciscan complex, san simeon, california, *Geological Society of America Bulletin*, *89*(9), 1415, doi:[10.1130/0016-7606(1978)89<1415:oobcri>2.0.co;2](https://doi.org/10.1130/0016-7606(1978)89<1415:oobcri>2.0.co;2).

Crank, J., and P. Nicolson (1947), A practical method for numerical evaluation of solutions of partial differential equations of the heat-conduction type, *Mathematical Proceedings of the Cambridge Philosophical Society*, *43*(01), 50–67.

DePaolo, D., and G. Wasserburg (1976), Inferences about magma sources and mantle structure from variations of 143Nd/144Nd, *Geophysical Research Letters*, *3*(12).

Dick, H., and T. Bullen (1984), Chromian spinel as a petrogenetic indicator in abyssal and alpine-type peridotites and spatially associated lavas, *Contributions to Mineralogy and Petrology*, *86*, 54–76.

Dickinson, W., M. N. Ducea, L. I. Rosenberg, G. H. Greene, S. A. Graham, J. C. Clark, G. E. Weber, S. Kidder, W. G. Ernst, and E. E. Brabb (2005), Net dextral slip, neogene san gregorio-hosgri fault zone, coastal california: Geologic evidence and tectonic implications, *Geological Society of America Special Papers*, *391*, 43, doi:[10.1130/2005.2391.](https://doi.org/10.1130/2005.2391.)

Dietz, L. D., and W. L. Ellsworth (1990), The october 17, 1989, loma prieta, california, earthquake and its aftershocks: Geometry of the sequence from high-resolution locations, *Geophysical Research Letters*, *17*(9), 1417–1420, doi:[10.1029/gl017i009p01417](https://doi.org/10.1029/gl017i009p01417).

Doin, M. P., and L. Fleitout (1996), Thermal evolution of the oceanic lithosphere: An alternative view, *Earth and Planetary Science Letters*, *142*(1-2), 121–136, doi:[10.1016/0012-821x(96)00082-9](https://doi.org/10.1016/0012-821x(96)00082-9).

Ducea, M. N. (2003), Arc composition at mid-crustal depths: Insights from the coast ridge belt, santa lucia mountains, california, *Geophysical Research Letters*, *30*(13), 0–3.

Ducea, M. N., and J. B. Saleeby (1998a), The age and origin of a thick mafic-ultramafic keel from beneath the sierra nevada batholith, *Contributions in Mineralogy and Petrology*, *133*, 169–185.

Ducea, M. N., S. Kidder, J. T. Chesley, and J. B. Saleeby (2009), Tectonic underplating of trench sediments beneath magmatic arcs: The central california example, *International Geology Review*, *51*(1), 1–26, doi:[10.1080/00206810802602767](https://doi.org/10.1080/00206810802602767).

Ducea, M. N., J. B. Saleeby, and G. Bergantz (2015), The architecture, chemistry, and evolution of continental magmatic arcs, *Annual Review of Earth and Planetary Sciences*, *43*(1), 299–333, doi:[10.1146/annurev-earth-060614-105049](https://doi.org/10.1146/annurev-earth-060614-105049).

Ducea, M., and J. Saleeby (1996), Buoyancy sources for a large, unrooted mountain range, the sierra nevada, california: Evidence from xenolith thermobarometry, *Journal of Geophysical Research*, *101*(B4), 8229–8244.

Ducea, M., and J. Saleeby (1998b), A case for delamination of the deep batholithic crust beneath the sierra nevada, california, *International Geology Review*, *40*(1), 78–93, doi:[10.1080/00206819809465199](https://doi.org/10.1080/00206819809465199).

Ducea, M., M. A. House, and S. Kidder (2003), Late cenozoic denudation and uplift rates in the santa lucia mountains, california, *Geology*, *31*(2), 139, doi:[10.1130/0091-7613(2003)031<0139:lcdaur>2.0.co;2](https://doi.org/10.1130/0091-7613(2003)031<0139:lcdaur>2.0.co;2).

England, P., and P. Molnar (1990), Surface uplift, uplift of rocks, and exhumation of rocks, *18*, 1173–1177, doi:[10.1130/0091-7613(1990)018<1173:SUUORA>2.3.CO](https://doi.org/10.1130/0091-7613(1990)018<1173:SUUORA>2.3.CO).

Erkan, K., and D. Blackwell (2009), Transient thermal regimes in the sierra nevada and baja california extinct outer arcs following the cessation of farallon subduction, *Journal of Geophysical Research*, *114*(B2), B02107, doi:[10.1029/2007JB005498](https://doi.org/10.1029/2007JB005498).

Erkan, K., and D. D. Blackwell (2008), A thermal test of the post-subduction tectonic evolution along the california transform margin, *Geophysical Research Letters*, *35*(7), n/a–n/a, doi:[10.1029/2008GL033479](https://doi.org/10.1029/2008GL033479).

Ernst, W. G., and C. A. Hall (1974), Geology and petrology of the cambria felsite, a new oligocene formation, west-central california coast ranges, *Bulletin of the Geological Society of America*, *85*(4), 523–532, doi:[10.1130/0016-7606(1974)85<523:GAPOTC>2.0.CO;2](https://doi.org/10.1130/0016-7606(1974)85<523:GAPOTC>2.0.CO;2).

Farmer, G. L., A. F. Glazner, and C. R. Manley (2002), Did lithospheric delamination trigger late cenozoic potassic volcanism in the southern sierra nevada, california?, *Geological Society of America Bulletin*, *114*(6), 754–768, doi:[10.1130/0016-7606(2002)114<0754:dldtlc>2.0.co;2](https://doi.org/10.1130/0016-7606(2002)114<0754:dldtlc>2.0.co;2).

Farmer, G. L., a. F. Glazner, H. G. Wilshire, J. L. Wooden, W. J. Pickthorn, and M. Katz (1995), Origin of late cenozoic basalts at the cima volcanic field, mojave desert, california, *Journal of Geophysical Research*, *100*(B5), 8399, doi:[10.1029/95JB00070](https://doi.org/10.1029/95JB00070).

Fischer, K. M., H. A. Ford, D. L. Abt, and C. A. Rychert (2010), The lithosphere-asthenosphere boundary, *Annual Review of Earth and Planetary Sciences*, *38*(1), 551–575, doi:[10.1146/annurev-earth-040809-152438](https://doi.org/10.1146/annurev-earth-040809-152438).

Fowler, C. (2005), *The solid earth: An introduction to global geophysics*, 1st ed., Cambridge University Press, Cambridge.

Frassetto, A. M., G. Zandt, H. Gilbert, T. J. Owens, and C. H. Jones (2011), Structure of the sierra nevada from receiver functions and implications for lithospheric foundering, *Geosphere*, *7*(4), 898–921, doi:[10.1130/ges00570.1](https://doi.org/10.1130/ges00570.1).

Frey, F. A., and M. Prinz (1978), Ultramafic inclusions from san carlos, arizona - petrologic and geochemical data bearing on their petrogenesis, *Earth and Planetary Science Letters*, *38*, 129–176, doi:[10.1016/0012-821X(78)90130-9](https://doi.org/10.1016/0012-821X(78)90130-9).

Furlong, K. P., W. D. Hugo, and G. Zandt (1989), Geometry and evolution of the san andreas fault zone in northern california, *J. Geophys. Res.*, *94*(B3), 3100–3110, doi:[10.1029/jb094ib03p03100](https://doi.org/10.1029/jb094ib03p03100).

Galer, S. J. G., and R. K. O’Nions (1989), Chemical and isotopic studies of ultramafic inclusions from the san carlos volcanic field, arizona: A bearing on their petrogenesis, *Journal of Petrology*, *30*(4), 1033–1064, doi:[10.1093/petrology/30.4.1033](https://doi.org/10.1093/petrology/30.4.1033).

Gao, S., X. Liu, H. Yuan, B. Hattendorf, D. Günther, L. Chen, and S. Hu (2002), Determination of forty two major and trace elements in usgs and nist srm glasses by laser ablation-inductively coupled plasma-mass spectrometry, *Geostandards and Geoanalytical Research*, *26*(2), 181–196, doi:[10.1111/j.1751-908X.2002.tb00886.x](https://doi.org/10.1111/j.1751-908X.2002.tb00886.x).

Gasparik, T. (2000), An internally consistent thermodynamic model for the system cao‐MgO‐Al2O3‐SiO2 derived primarily from phase equilibrium data, *The Journal of Geology*, *108*, 103–119.

Gilbert, H., Y. Yang, D. W. Forsyth, C. H. Jones, T. J. Owens, G. Zandt, and J. C. Stachnik (2012), Imaging lithospheric foundering in the structure of the sierra nevada, *Geosphere*, *8*(6), 1310–1330, doi:[10.1130/GES00790.1](https://doi.org/10.1130/GES00790.1).

Goes, S., and S. V. D. Lee (2002), Thermal structure of the north american uppermost mantle inferred from seismic tomography, *Journal of Geophysical Research*, *107*.

Green, D., and A. Ringwood (1970), Mineralogy of peridotitic compositions under upper mantle conditions, *Physics of the Earth and Planetary Interiors*, *3*, 359–371, doi:[10.1016/0031-9201(70)90076-2](https://doi.org/10.1016/0031-9201(70)90076-2).

Grove, M., C. E. Jacobson, A. P. Barth, and A. Vucic (2003), Temporal and spatial trends of late cretaceous-early tertiary underplating ofPelona and related schist beneath southern california and southwestern arizona, *Tectonic evolution of northwestern Mexico and the southwestern USA*, *374*, 381.

Guyer, J. E., D. Wheeler, and J. A. Warren (2009), FiPy: Partial differential equations with python, *Computing in Science and Engineering*, *11*, 6–15, doi:[10.1109/MCSE.2009.52](https://doi.org/10.1109/MCSE.2009.52).

Hall, C. A., and J. B. Saleeby (2013), Salinia revisited: A crystalline nappe sequence lying above the nacimiento fault and dispersed along the san andreas fault system, central california, *International Geology Review*, *55*(13), 1575–1615, doi:[10.1080/00206814.2013.825141](https://doi.org/10.1080/00206814.2013.825141).

Hardebeck, J. L. (2010), Seismotectonics and fault structure of the california central coast, *Bulletin of the Seismological Society of America*, *100*(3), 1031–1050, doi:[10.1785/0120090307](https://doi.org/10.1785/0120090307).

Herzberg, C. T. (1978), Pyroxene geothermometry and geobarometry: Experimental and thermodynamic evaluation of some subsolidus phase relations involving pyroxenes in the system cao-mgo-al2o3-sio2, *Geochimica et Cosmochimica Acta*, *42*, 945–957.

Hofmann, A. W. (1997), Mantle geochemistry: The message from oceanic volcanism, *Nature*, *385*, 219–229, doi:[10.1038/385219a0](https://doi.org/10.1038/385219a0).

Humphreys, E. D. (1995), Post-laramide removal of the farallon slab,western united states, *Geology*, *23*, 987–990, doi:[10.1130/0091-7613(1995)023<0987](https://doi.org/10.1130/0091-7613(1995)023<0987).

Hurst, R. W. (1982), Petrogenesis of the conejo volcanic suite, southern california: Evidence for mid-ocean ridge–continental margin interactions, *Geology*, *10*(5), 267, doi:[10.1130/0091-7613(1982)10<267:potcvs>2.0.co;2](https://doi.org/10.1130/0091-7613(1982)10<267:potcvs>2.0.co;2).

Johnson, K., H. Dick, and N. Shimizu (1990), Melting in the oceanic upper mantle: An ion microprobe study of diopsides in abyssal peridotites, *Journal of Geophysical Research*, *95*(89), 2661–2678.

Jones, C. H., and R. A. Phinney (1998), Seismic structure of the lithosphere from teleseismic converted arrivals observed at small arrays in the southern sierra nevada and vicinity, california, *Journal of Geophysical Research: Solid Earth*, *103*(B5), 10065–10090, doi:[10.1029/97jb03540](https://doi.org/10.1029/97jb03540).

Jones, C. H., H. Reeg, G. Zandt, H. Gilbert, T. J. Owens, and J. Stachnik (2014), P-wave tomography of potential convective downwellings and their source regions, sierra nevada, california, *Geosphere*, *10*(3), 505–533, doi:[10.1130/ges00961.1](https://doi.org/10.1130/ges00961.1).

Jové, C. F., and R. G. Coleman (1998), Extension and mantle upwelling within the san andreas fault zone, san francisco bay area, california, *Tectonics*, *17*(6), 883–890, doi:[10.1029/1998tc900012](https://doi.org/10.1029/1998tc900012).

Kelemen, P. B., M. Braun, and G. Hirth (2000), Spatial distribution of melt conduits in the mantle beneath oceanic spreading ridges: Observations from the ingalls and oman ophiolites, *Geochemistry Geophysics Geosystems*, *1*(1), doi:[10.1029/1999GC000012](https://doi.org/10.1029/1999GC000012).

Kennedy, B. M. (1997), Mantle fluids in the san andreas fault system, california, *Science*, *278*(June), 1278–1281, doi:[10.1126/science.278.5341.1278](https://doi.org/10.1126/science.278.5341.1278).

Kidder, S. B., F. Herman, J. Saleeby, J.-P. Avouac, M. N. Ducea, and a. Chapman (2013), Shear heating not a cause of inverted metamorphism, *Geology*, *41*(8), 899–902, doi:[10.1130/G34289.1](https://doi.org/10.1130/G34289.1).

Kidder, S., and M. N. Ducea (2006), High temperatures and inverted metamorphism in the schist of sierra de salinas, california, *Earth and Planetary Science Letters*, *241*(3-4), 422–437.

Kidder, S., M. Ducea, G. Gehrels, P. J. Patchett, and J. Vervoort (2003), Tectonic and magmatic development of the salinian coast ridge belt, california, *Tectonics*, *22*(5), 1058, doi:[10.1029/2002TC001409](https://doi.org/10.1029/2002TC001409).

Kinzler, R. J. (1997), Melting of mantle peridotite at pressures approaching the spinel to garnet transition: Application to mid-ocean ridge basalt petrogenesis, *Journal of Geophysical Research*, *102*(B1), 853, doi:[10.1029/96JB00988](https://doi.org/10.1029/96JB00988).

Kistler, R. W., and D. E. Champion (2001), Rb-sr whole-rock and mineral ages, k-ar , ar/ar, and u-pb mineral ages, and strontium, lead, neodymium, and oxygen isotopic compositions for granitic rocks from the salinian composite terrane, california, *USGS Open File Report*, *01-453*, 1–80.

Klemme, S. (2004), The influence of cr on the garnet–spinel transition in the earth’s mantle: Experiments in the system mgo–Cr2O3–SiO2 and thermodynamic modelling, *Lithos*, *77*(1-4), 639–646, doi:[10.1016/j.lithos.2004.03.017](https://doi.org/10.1016/j.lithos.2004.03.017).

Klemme, S., and H. S. C. O’Neill (2000), The near-solidus transition from garnet iherzolite to spinel iherzolite, *Contributions to Mineralogy and Petrology*, *138*(3), 237–248, doi:[10.1007/s004100050560](https://doi.org/10.1007/s004100050560).

Köhler, T., and G. Brey (1990), Calcium exchange between olivine and clinopyroxene calibrated as a geothermobarometer for natural peridotites from 2 to 60 kb with applications, *Geochimica et Cosmochimica Acta*, *54*, 2375–2388.

Lachenbruch, A., and J. Sass (1980), Heat flow and energetics of the san andreas fault zone, *Journal of Geophysical Research*, *85*, 6185–6222.

Lee, C.-T. A., X. Cheng, and U. Horodyskyj (2006), The development and refinement of continental arcs by primary basaltic magmatism, garnet pyroxenite accumulation, basaltic recharge and delamination: Insights from the sierra nevada, california, *Contributions to Mineralogy and Petrology*, *151*(2), 222–242, doi:[10.1007/s00410-005-0056-1](https://doi.org/10.1007/s00410-005-0056-1).

Lee, C.-T., R. L. Rudnick, and G. H. Brimhall (2001), Deep lithospheric dynamics beneath the sierra nevada during the mesozoic and cenozoic as inferred from xenolith petrology, *Geochemistry, Geophysics, Geosystems*, *2*(12), n/a—–n/a, doi:[10.1029/2001gc000152](https://doi.org/10.1029/2001gc000152).

Levandowski, W., and C. H. Jones (2015), Linking sierra nevada, california, uplift to subsidence of the tulare basin using a seismically derived density model, *Tectonics*, *34*(11), 2349–2358, doi:[10.1002/2015tc003824](https://doi.org/10.1002/2015tc003824).

Li, X., X. Yuan, and R. Kind (2007), The lithosphere-asthenosphere boundary beneath the western united states, *Geophysical Journal International*, *170*(2), 700–710, doi:[10.1111/j.1365-246x.2007.03428.x](https://doi.org/10.1111/j.1365-246x.2007.03428.x).

Liang, Y., C. Sun, and L. Yao (2013), A ree-in-two-pyroxene thermometer for mafic and ultramafic rocks, *Geochimica et Cosmochimica Acta*, *102*, 246–260, doi:[10.1016/j.gca.2012.10.035](https://doi.org/10.1016/j.gca.2012.10.035).

Liu, L., M. Gurnis, M. Seton, J. Saleeby, R. D. Müller, and J. M. Jackson (2010), The role of oceanic plateau subduction in the laramide orogeny, *Nature Geoscience*, *3*(5), 353–357, doi:[10.1038/ngeo829](https://doi.org/10.1038/ngeo829).

Livaccari, R. F., and F. V. Perry (1993), Isotopic evidence for preservation of cordilleran lithospheric mantle during the sevier-laramide orogeny, western united states, *Geology*, *21*(8), 719, doi:[10.1130/0091-7613(1993)021<0719:iefpoc>2.3.co;2](https://doi.org/10.1130/0091-7613(1993)021<0719:iefpoc>2.3.co;2).

Livaccari, R. F., K. Burke, and a. M. C. Şengör (1981), Was the laramide orogeny related to subduction of an oceanic plateau?, *289*, 276–278, doi:[10.1038/289276a0](https://doi.org/10.1038/289276a0).

Luffi, P., J. B. Saleeby, C.-T. a. Lee, and M. N. Ducea (2009), Lithospheric mantle duplex beneath the central mojave desert revealed by xenoliths from dish hill, california, *Journal of Geophysical Research*, *114*(B3), B03202, doi:[10.1029/2008JB005906](https://doi.org/10.1029/2008JB005906).

Malin, P. E., E. D. Goodman, T. L. Henyey, Y. G. Li, D. A. Okaya, and J. B. Saleeby (1995), Significance of seismic reflections beneath a tilted exposure of deep continental crust, tehachapi mountains, california, *Journal of Geophysical Research: Solid Earth*, *100*(B2), 2069–2087, doi:[10.1029/94jb02127](https://doi.org/10.1029/94jb02127).

Mancktelow, N. S., and B. Grasemann (1997), Time-dependent effects of heat advection and topography on cooling histories during erosion, *Tectonophysics*, *270*(3-4), 167–195, doi:[10.1016/S0040-1951(96)00279-X](https://doi.org/10.1016/S0040-1951(96)00279-X).

Matthews III, V. (1976), Correlation of pinnacles and neenach volcanic formations and their bearing on san andreas fault problem, *AAPG Bulletin*, *60*, doi:[10.1306/c1ea3a82-16c9-11d7-8645000102c1865d](https://doi.org/10.1306/c1ea3a82-16c9-11d7-8645000102c1865d).

McCulloch, M., and G. Wasserburg (1978), Sm-nd and rb-sr chronology of continental crust formation, *Science*, *200*(4345), 1003–1011.

Medaris, G., H. Wang, and J. Fournelle (1999), A cautionary tale of spinel peridotite thermobarometry: An example from xenoliths of kozákov volcano, czech republic, *Geolines*, *9*, 92.

Mei, S., A. M. Suzuki, D. L. Kohlstedt, N. A. Dixon, and W. B. Durham (2010), Experimental constraints on the strength of the lithospheric mantle, *Journal of Geophysical Research*, *115*(B8), doi:[10.1029/2009jb006873](https://doi.org/10.1029/2009jb006873).

Murchey, B. L., and D. L. Jones (1984), Age and significance of chert in the franciscan complex in the san francisco bay region, *Franciscan Geology of Northern California*.

Nadin, E. S., and J. B. Saleeby (2008), *Disruption of regional primary structure of the sierra nevada batholith by the kern canyon fault system, california elisabeth*, 15.

Nesse, W. D. (2000), *Introduction to mineralogy*, Oxford University Press, Oxford.

Nicholson, C., C. C. Sorlien, and B. P. Luyendyk (1992), Deep crustal structure and tectonics in the offshore southern santa-maria basin, california, *Geology*, *20*(3), 239 –&, doi:[10.1130/0091-7613(1992)020<0239](https://doi.org/10.1130/0091-7613(1992)020<0239).

Nickel, K., and D. Green (1985), Empirical geothermobarometry for garnet peridotites and implications for the nature of the lithosphere, kimberlites and diamonds, *Earth and Planetary Science Letters*, *73*(1), 158–170, doi:[10.1016/0012-821X(85)90043-3](https://doi.org/10.1016/0012-821X(85)90043-3).

Nimis, P., and H. Grütter (2010), Internally consistent geothermometers for garnet peridotites and pyroxenites, *Contributions to Mineralogy and Petrology*, *159*(3), 411–427, doi:[10.1007/s00410-009-0455-9](https://doi.org/10.1007/s00410-009-0455-9).

Nimis, P., and W. R. Taylor (2000), Single clinopyroxene thermobarometry for garnet peridotites. part i. calibration and testing of a cr-in-cpx barometer and an enstatite-in-cpx thermometer, *Contributions to Mineralogy and Petrology*, *139*(5), 541–554, doi:[10.1007/s004100000156](https://doi.org/10.1007/s004100000156).

Ozacar, A. A., and G. Zandt (2009), Crustal structure and seismic anisotropy near the san andreas fault at parkfield, california, *Geophysical Journal International*, *178*(2), 1098–1104, doi:[10.1111/j.1365-246x.2009.04198.x](https://doi.org/10.1111/j.1365-246x.2009.04198.x).

O’Neill, H. (1981), The transition between spinel lherzolite and garnet lherzolite, and its use as a geobarometer, *Contributions to Mineralogy and Petrology*, *77*, 185–194.

O’Reilly, S. Y. (1997), Minor elements in olivine from spinel lherzolite xenoliths: Implications for thermobarometry, *Mineralogical Magazine*, *61*(405), 257–269, doi:[10.1180/minmag.1997.061.405.09](https://doi.org/10.1180/minmag.1997.061.405.09).

O’Reilly, S. Y., and W. L. Griffin (2010), The continental lithosphere-asthenosphere boundary: Can we sample it?, *Lithos*, *120*(1-2), 1–13, doi:[10.1016/j.lithos.2010.03.016](https://doi.org/10.1016/j.lithos.2010.03.016).

Page, B. M. (1981), The southern coast ranges, edited by W. Ernst, *The geotectonic development of California; Rubey Volume I*, *1*, 329–417.

Pike, J. E. N., and E. C. Schwarzman (1977), Classification of textures in ultramafic xenoliths, *The Journal of Geology*, *85*(1), 49–61, doi:[10.2307/30068676](https://doi.org/10.2307/30068676).

Pikser, J. E., D. W. Forsyth, and G. Hirth (2012), Along-strike translation of a fossil slab, *Earth and Planetary Science Letters*, *331-332*, 315–321, doi:[10.1016/j.epsl.2012.03.027](https://doi.org/10.1016/j.epsl.2012.03.027).

Platt, J. P., and W. M. Behr (2011), Deep structure of lithospheric fault zones, *Geophysical Research Letters*, *38*(24), doi:[10.1029/2011GL049719](https://doi.org/10.1029/2011GL049719).

Pollack, H. N., and D. S. Chapman (1977), On the regional variation of heat flow, geotherms, and lithospheric thickness, *Tectonophysics*, *38*(3-4), 279–296, doi:[10.1016/0040-1951(77)90215-3](https://doi.org/10.1016/0040-1951(77)90215-3).

Robinson, J. A. C., and B. J. Wood (1998), The depth of the spinel to garnet transition at the peridotite solidus, *Earth and Planetary Science Letters*, *164*(1-2), 277–284, doi:[10.1016/S0012-821X(98)00213-1](https://doi.org/10.1016/S0012-821X(98)00213-1).

Royden, L. (1993), The steady state thermal structure of eroding orogenic belts and accretionary prisms, *Journal of Geophysical Research: Solid Earth*, *98*, 4487–4507.

Saleeby, J. (2003), Production and loss of high-density batholithic root, southern sierra nevada, california, *Tectonics*, *22*(6), 1064, doi:[10.1029/2002TC001374](https://doi.org/10.1029/2002TC001374).

Saleeby, J. B., Z. Saleeby, J. Robbins, and J. Gillespie (2016), Sediment provenance and dispersal of neogene-quaternary strata of the southeastern san joaquin basin and its transition into the southern sierra nevada, california, *Geosphere*.

Saleeby, J., and 1. contributors (1986), Continent-ocean transect, corridor c2, monterey bay offshore to the colorado plateau, in *Geologic society of america map and chart series tra c2, 2 sheets, scale 1:500,000*, p. 87.

Saleeby, J., K. A. Farley, R. W. Kistler, and R. J. Fleck (2007), Thermal evolution and exhumation of deep-level batholithic exposures, southernmost sierra nevada, california, in *Special paper 419: Convergent margin terranes and associated regions: A tribute to w.G. ernst*, pp. 39–66, Geological Society of America.

Saleeby, J., L. Le Pourhiet, Z. Saleeby, and M. Gurnis (2012), Epeirogenic transients related to mantle lithosphere removal in the southern sierra nevada region, california, part i: Implications of thermomechanical modeling, *Geosphere*, *8*(6), 1286–1309, doi:[10.1130/GES00746.1](https://doi.org/10.1130/GES00746.1).

Saleeby, J., Z. Saleeby, and L. Le Pourhiet (2013), Epeirogenic transients related to mantle lithosphere removal in the southern sierra nevada region, california: Part ii. implications of rock uplift and basin subsidence relations, *Geosphere*, *9*(3), 394–425, doi:[10.1130/GES00816.1](https://doi.org/10.1130/GES00816.1).

Schott, R. C., and C. M. Johnson (1998), Sedimentary record of the late cretaceous thrusting and collapse of the salinia-mojave magmatic arc, *Geology*, *26*(4), 327, doi:[10.1130/0091-7613(1998)026<0327:SROTLC>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<0327:SROTLC>2.3.CO;2).

Schott, R. C., and C. M. Johnson (2001), Garnet-bearing trondhjemite and other conglomerate clasts from the gualala basin, california: Sedimentary record of the missing western portion of the salinian magmatic arc?, *Geological Society of America Bulletin*, *113*(7), 870–880.

Seiders, V. (1989), Geologic map of the burnett peak quadrangle, monterey and san luis obispo counties, california, *U.S. Geological Survey Geologic Maps*.

Seton, M. et al. (2012), Global continental and ocean basin reconstructions since 200Ma, *Earth-Science Reviews*, *113*(3-4), 212–270, doi:[10.1016/j.earscirev.2012.03.002](https://doi.org/10.1016/j.earscirev.2012.03.002).

Sharma, M., A. R. Basu, R. B. Cole, and P. G. DeCelles (1991), Basalt-rhyolite volcanism by {MORB}-continental crust interaction: Nd, sr-isotopic and geochemical evidence from southern san joaquin basin, california, *Contributions to Mineralogy and Petrology*, *109*(2), 159–172, doi:[10.1007/bf00306476](https://doi.org/10.1007/bf00306476).

Sharman, G. R., S. A. Graham, M. Grove, and J. K. Hourigan (2013), A reappraisal of the early slip history of the san andreas fault, central california, {USA}, *Geology*, *41*(7), 727–730, doi:[10.1130/g34214.1](https://doi.org/10.1130/g34214.1).

Shervais, J. W., H. G. Wilshire, and E. C. Schwarzman (1973), Garnet clinopyroxenite xenolith from dish hill, california, *Earth and Planetary Science Letters*, *19*(2), 120–130, doi:[10.1016/0012-821x(73)90106-4](https://doi.org/10.1016/0012-821x(73)90106-4).

Shields, J. E., and A. D. Chapman (2016), Late cretaceous tectonic displacement of sub-continental mantle lithosphere beneath the sw u.S. cordillera; mantle xenolith constraints from the colorado plateau transition zone (central arizona)., *Abstracts with Programs - Geological Society of America*, *48*(7), @Abstractno.267–12.

Sliter, W. V. (1984), Foraminifers from cretaceous limestone of the franciscan complex, northern california, in *Society of economic paleontologists and mineralogists, pacific section*, edited by M. Blake, pp. 149–162, Los Angeles, California.

Smith, P. M., and P. D. Asimow (2005), Adiabat\_1ph: A new public front-end to the melts, pMELTS, and pHMELTS models, *Geochemistry, Geophysics, Geosystems*, *6*(2), n/a–n/a, doi:[10.1029/2004GC000816](https://doi.org/10.1029/2004GC000816).

Stanley, B. R. G., D. S. Wilson, and P. A. McCrory (2000), Locations and ages of middle tertiary volcanic centers in coastal california, *USGS Open File Reports*, *00*(154), 0–27.

Stein, C. A. (1995), Heat flow of the earth, *AGU Reference Shelf*, 144–158, doi:[10.1112/S0024609301008396](https://doi.org/10.1112/S0024609301008396).

Stein, C. A., and S. Stein (1992), A model for the global variation in oceanic depth and heat flow with lithospheric age, *Nature*, *356*(6391), 133–135, doi:[10.1038/359123a0](https://doi.org/10.1038/359123a0).

Sun, C., and Y. Liang (2012), Distribution of ree between clinopyroxene and basaltic melt along a mantle adiabat: Effects of major element composition, water, and temperature, *Contributions to Mineralogy and Petrology*, *163*(5), 807–823, doi:[10.1007/s00410-011-0700-x](https://doi.org/10.1007/s00410-011-0700-x).

Sun, D., M. Gurnis, J. Saleeby, and D. Helmberger (2017), A dipping, thick segment of the farallon slab beneath central us, *Journal of Geophysical Research: Solid Earth*, doi:[10.1002/2016JB013915](https://doi.org/10.1002/2016JB013915).

Sun, S.-s., and W. F. McDonough (1989), Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes, *Geological Society, London, Special Publications*, *42*(1), 313–345, doi:[10.1144/GSL.SP.1989.042.01.19](https://doi.org/10.1144/GSL.SP.1989.042.01.19).

Taylor, W. (1998), An experimental test of some geothermometer and geobarometer formulations for upper mantle peridotites with application to the thermobarometry of fertile lherzolite and garnet websterite, *Neues Jahrbuch für Mineralogie Abhandlungen*, *172*(2), 381–408.

Thorkelson, D. J., and R. P. Taylor (1989), Cordilleran slab windows, *Geology*, *17*(9), 833–836, doi:[10.1130/0091-7613(1989)017<0833:CSW>2.3.CO;2](https://doi.org/10.1130/0091-7613(1989)017<0833:CSW>2.3.CO;2).

Titus, S. J., L. G. Medaris, H. F. Wang, and B. Tikoff (2007), Continuation of the san andreas fault system into the upper mantle: Evidence from spinel peridotite xenoliths in the coyote lake basalt, central california, *Tectonophysics*, *429*(1-2), 1–20, doi:[10.1016/j.tecto.2006.07.004](https://doi.org/10.1016/j.tecto.2006.07.004).

Trehu, A. M., and W. H. Wheeler (1987), Possible evidence for subducted sedimentary materials beneath central california., *Geology*, *15*(3), 254–258, doi:[10.1130/0091-7613(1987)15<254:PEFSSM>2.0.CO;2](https://doi.org/10.1130/0091-7613(1987)15<254:PEFSSM>2.0.CO;2).

Tréhu, A. (1991), Tracing the subducted the central california continental margin: Results from ocean bottom seismometers deployed during the 1986 pacific gas and electric edge experiment, *Journal of Geophysical Research*, *96*(90), 6493–6506.

Turcotte, D., and G. Schubert (2002), *Geodynamics*, 2nd ed., Cambridge University Press, Cambridge.

Usui, T., E. Nakamura, K. Kobayashi, S. Maruyama, and H. Helmstaedt (2003), Fate of the subducted farallon plate inferred from eclogite xenoliths in the colorado plateau, *Geology*, *31*(7), 589, doi:[10.1130/0091-7613(2003)031<0589:fotsfp>2.0.co;2](https://doi.org/10.1130/0091-7613(2003)031<0589:fotsfp>2.0.co;2).

Van Wijk, J., R. Govers, and K. Furlong (2001), Three-dimensional thermal modeling of the california upper mantle: A slab window vs. stalled slab, *Earth and Planetary Science Letters*, *186*, 175–186.

Vilà, M., M. Fernández, and I. Jiménez-Munt (2010), Radiogenic heat production variability of some common lithological groups and its significance to lithospheric thermal modeling, *Tectonophysics*, *490*(3-4), 152–164, doi:[10.1016/j.tecto.2010.05.003](https://doi.org/10.1016/j.tecto.2010.05.003).

Wang, Y., D. W. Forsyth, C. J. Rau, N. Carriero, B. Schmandt, J. B. Gaherty, and B. Savage (2013), Fossil slabs attached to unsubducted fragments of the farallon plate., *Proceedings of the National Academy of Sciences*, 5342–6, doi:[10.1073/pnas.1214880110](https://doi.org/10.1073/pnas.1214880110).

Warren, J. (2016), Global variations in abyssal peridotite compositions, *Lithos*, *248-251*, 193–219, doi:[10.1016/j.lithos.2015.12.023](https://doi.org/10.1016/j.lithos.2015.12.023).

Warren, J. M., and G. Hirth (2006), Grain size sensitive deformation mechanisms in naturally deformed peridotites, *Earth and Planetary Science Letters*, *248*(1-2), 438–450, doi:[10.1016/j.epsl.2006.06.006](https://doi.org/10.1016/j.epsl.2006.06.006).

Wasserburg, G., S. Jacousen, D. DePaolo, M. McCulloch, and T. Wen (1981), Precise determination of ratios, sm and nd isotopic abundances in standard solutions, *Geochimica et Cosmochimica Acta*, *45*(Table 1), 2311–2323, doi:[10.1016/0016-7037(81)90085-5](https://doi.org/10.1016/0016-7037(81)90085-5).

Wilshire, H., C. E. Meyer, J. K. Nakata, and L. C. Calk (1988), Mafic and ultramafic xenoliths from volcanic rocks of the western united states, *USGS Open File Report*, *1443*, 179.

Wilson, D. S., P. a. McCrory, and R. G. Stanley (2005), Implications of volcanism in coastal california for the neogene deformation history of western north america, *Tectonics*, *24*, 1–22, doi:[10.1029/2003TC001621](https://doi.org/10.1029/2003TC001621).

Witt-Eickschen, G., and H. S. O’Neill (2005), The effect of temperature on the equilibrium distribution of trace elements between clinopyroxene, orthopyroxene, olivine and spinel in upper mantle peridotite, *Chemical Geology*, *221*(1-2), 65–101, doi:[10.1016/j.chemgeo.2005.04.005](https://doi.org/10.1016/j.chemgeo.2005.04.005).

Workman, R. K., and S. R. Hart (2005), Major and trace element composition of the depleted morb mantle (dmm), *Earth and Planetary Science Letters*, *231*(1-2), 53–72, doi:[10.1016/j.epsl.2004.12.005](https://doi.org/10.1016/j.epsl.2004.12.005).

Yan, Z., and R. W. Clayton (2007), A notch structure on the moho beneath the eastern san gabriel mountains, *Earth and Planetary Science Letters*, *260*(3-4), 570–581, doi:[10.1016/j.epsl.2007.06.017](https://doi.org/10.1016/j.epsl.2007.06.017).

Yan, Z., R. W. Clayton, and J. Saleeby (2005), Seismic refraction evidence for steep faults cutting highly attenuated continental basement in the central transverse ranges, california, *Geophysical Journal International*, *160*(2), 651–666, doi:[10.1111/j.1365-246x.2005.02506.x](https://doi.org/10.1111/j.1365-246x.2005.02506.x).

Zandt, G., H. Gilbert, T. J. Owens, M. Ducea, J. Saleeby, and C. H. Jones (2004), Active foundering of a continental arc root beneath the southern sierra nevada in california, *Nature*, *431*(7004), 41–46, doi:[10.1038/nature02847](https://doi.org/10.1038/nature02847).