**CONSTRUCTION OF MANTLE LITHOSPHERE BENEATH THE COASTAL CENTRAL CALIFORNIA SUBDUCTION ACCRETION COMPLEX FROM CRYSTAL KNOB VOLCANIC NECK HOSTED XENOLITHS**

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**Abstract**

The Crystal Knob volcanic neck in the Santa Lucia Range, California, was erupted during the Pleistocene (1.65 Ma from Ar geochronology) through the Nacimiento belt of the Franciscan complex in the coastal region of central California. The neck erupted an olivine-plagioclase phyric basalt containing spinel peridotite and dunite cumulate xenoliths. The peridotites sample the mantle lithosphere beneath the Nacimiento Franciscan, which was constructed as an accretionary prism during the Late Cretaceous subduction of the Farallon plate. The xenolith suite is also of interest because it lies within the bounds of the “Salinia terrane”, commonly interpreted as a rooted crustal fragment of the SW Cordilleran continental arc slivered into the Franciscan complex by strike-slip faulting, but more recently shown to be a series of crystalline nappes that were emplaced above the Franciscan.

Six spinel peridotite samples were analyzed, ranging from fertile lherzolites to clinopyroxene ***(?)*** harzburgites with modal clinopyroxene from 2-13%. They have a depleted mantle 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 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. This variability means the samples were likely entrained at several depths. Considering broad limits imposed by phase stability and temperature, the samples are likely sourced between 40 and 80 km depth. The higher-temperature samples show general REE depletion but sharp enrichment in LREEs, potentially showing a history of melt extraction followed by off-axis refertilization.

Forward modeling of thermal evolution is used to construct a range of scenarios for the age of the underplated Farallon slab and structure of the margin. Scenarios ranging from a young "stalled slab" to the rollback imbrication of Farallon lithosphere in the Cretaceous to Paleocene imply relatively cool modern geotherms approaching steady-state heat flow. These scenarios generally predict colder conditions at depth than are recorded by the xenoliths or surface heat-flow data implying that the xenoliths are sourced from the bottom of their stability field, and that fluid migration following the plate boundary may be an important process for increasing local heat transfer rates. ***(Daven, we can have a much closer look at the abstract when you have finished and integrated all the text modules. It’s looking pretty good though)***

**1. 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 (Ducea et al., 2015) whose generation was driven by the eastward subduction of the Farallon oceanic plate beneath western North America (Seaton et al., 2012). 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 (Cowan, 1978; Murchy and Jones, 1984; Sliter, 1984; Blake et al., 1988; Chapman et al., 2016a). 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; Kidder and Ducea, 2006; Ducea et al., 2009; Chapman et al. 2012; Hall and Saleeby. 2013). 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 (Fig. 1b), a polyphase structural zone that 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.

**1.1 Regional tectonic setting and the application of mantle xenolith studies**

The SW North American Cordillera is endowed with an abundance of xenolith localities wherein 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 (Welshire 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 (Galer and O’Nions, 1989; Livaccari and Perry, 1993; Alibert, 1994; Beard and Glazner, 1995; Ducea and Saleeby, 1996, 1998a; Jove and Coleman, 1998; Lee et al., 2001, 2006; Usui et al., 2003; Luffi et al., 2009). In general there is a geologically reasonable correspondence between such upper mantle domains and surface geology, with sub-continental suites having 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 the restoration of its host crustal rocks to a pre-San Andreas transform offset position, outboard of the northern reaches of the southern California batholith (Fig. 1 and 18, detailed below). The southern California batholith is the southern continuation of the Sierra Nevada batholith across the Garlock fault. 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, widely exposed in ranges of the Mojave plateau (Saleeby, 2003; Barbeau et al., 2005; Chapman et al., 2012). 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 (Malin et al., 1995; Barth et al., 2003; Yan et al., 2005; Ducea et al., 2009; Chapman et al. 2010, 2012, 2016b). The underplated metaclastic rocks are exposed in a series of windows that are labeled as subduction channel schists on Figure 1. 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; Saleeby et al., 2007; Chapman et al., 2013, 2016a). 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; Sliter, 1984; Saleeby, 2003), which is currently resolved in deep seismic tomographic images beneath the interior of North America (Liu et al., 2010). 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 (Saleeby, 2003; Liu et al., 2010; Chapman et al., 2012).

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 (Fig. 3) carry xenolith suites that sampled the Cretaceous mantle wedge of the overlying batholith (Ducea and Saleeby, 1996, 1998a; Lee et al., 2001, 2006; Saleeby et al., 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 (Saleeby, 2003; Chapman et al., 2010, 2012, 2016b). Seismic reflection data image the megathrust as effectively flat beneath the western Mojave plateau (Yan et al., 2005), and dipping ~30°N beneath the southernmost Sierra Nevada region (Malin et al. 1995). The Garlock fault (Fig. 1a) nucleated during the early Miocene along the inflection in the megathrust (Saleeby et al., 2016), with the inflection constituting a lateral ramp in the subduction megathrust system (Saleeby, 2003; Chapman et al., 2016b). 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 (Shervais et al., 1973; Luffi et al., 2009). More specifically the Dish Hill suite (Fig. 1) 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 (Fig. 1), 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 (Fig. 3), 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, 1998b). Seismic data from the eastern Sierra region reveal asthenospheric mantle extending upwards to the base of the crust at ~30 km depth (Jones and Phinney, 1998; Zandt et al., 2004; Frassetto et al., 2011; Jones et al., 2014), consistent with the eastern Sierra xenolith suite findings. The eastern Sierra (suite) eruption sites encompass the <10 Ma old Owens Valley rift system (Fig. 1a), which is 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 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 employ petrologic and geochemical investigations on 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.

**1.2 Crystal Knob Mantle Xenolith Locality**

The Crystal Knob volcanic neck (35.806º N, 121.174º W) is a mid-Pleistocene olivine--plagioclase phyric basalt that erupted through the Nacimiento belt of the Franciscan complex, ~500 m west of the Nacimiento fault in the Santa Lucia Mountains of central California [*Seiders*, [1989](#ref-BurnettPeak1989)]. The basaltic neck is ~80 m in diameter at the surface and has entrained abundant dunite and sparse spinel peridotite xenoliths (Fig. 2). 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.

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 Ar/Ar radioisotope technique on in situ plagioclase phenocrysts. ***(Daven, confirm whether or not the analyses were in situ, or from plagioclase separates, I seem to remember in situ)*** A billet of the host lava (sample CK-1) was provided to the USGS Geochronology Laboratory in Denver, Colorado. Plagioclase phenocrysts were selected from the sample and irradiated in the USGS TRIGA reactor. The samples were step-heated using an infrared laser, and Ar 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 assumed to have Ar/Ar = 298.56 (atmospheric composition) [*Cosca et al.*, [2011](#ref-Cosca2011)]. Data for stepwise heating are presented in **Table x**, and are presented graphically in **Figure x *(Daven insert these in sequence)***. 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 Crystal Knob lavas entrained xenoliths from the immediately underlying upper mantle at ca. 1.7 Ma, or mid-Pleistocene time.

# Methods

Polished thin sections of 250 µm thickness were prepared for the basaltic host lava (CK-1), six peridotite xenolith samples (CK-2 through CK-7), and two dunite cumulate samples (CK-D1 and CK-D2). The samples were evaluated under a petrographic microscope to determine their textural and mineralogic variation. Electron backscatter intensity images were collected using a ZEISS 1550 VP field emission SEM. For each sample, modal mineralogy was estimated by classifying mineralogy on a ~5000 pixel grid atop coregistered optical scans and electron backscatter mosaics . Volumetric modes were converted to weight percents using representative densities for spinel-facies peridotites. Results are shown in Figure .

[[textures|figure]]

Major-element mineral compositions were analyzed for each 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](#ref-Armstrong1988)] 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. Whole-rock major-element abundances are reconstructed from mineral composition and estimated modes. Representative mineral and recalculated whole-rock compositions are tabulated in Table .

Trace element concentrations of the xenolith samples were analyzed 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](#ref-Gao2002)]. 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 ignored in calculations because rare-earth elements (REEs) are 2-3 orders of magnitude less compatible than in clinopyroxene [*Luffi et al.*, [2009](#ref-Luffi2009); *Witt-Eickschen and O’Neill*, [2005](#ref-WittEickschen2005)]. Results for measured pyroxene and recalculated whole-rock trace elements are shown in Table .

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 seperates 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](#ref-Wasserburg1981)]. 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 .

#### Host Lava (section moved to above text on Crystal Knob)

# Results

[[modes|figure]]

## Petrography

Six peridotite samples (CK-2 through CK-7) were collected and analyzed. As collected, samples were 5-10 cm diameter friable peridotites with medium (200 µm -- 1 mm) grains. Additionally, several samples consisting of the host lava (CK-1) and containing fragments of peridotite and dunite cumulates (CK-D1 and CK-D2) were prepared. ***(Daven, we should have a table that lists all samples, and gives a petrographic overview)***

Thin sections of each peridotite sample were texturally classified using the scheme of *Pike and Schwarzman* [[1977](#ref-Pike1977)]. All samples display an allotriomorphic granular texture with anisotropy or visible zoning largely absent. There is no preferred orientation in silicate grains, although several samples (notably CK-2 and CK-5) exhibit a weak shape-preferred orientation in elongate spinels***. (Daven, we should state whether or not three are petrographically observed plastic deformation features in the peridotites)***

Minor late-stage alteration products are seen in all peridotite samples. This includes 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 which 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 euhedral spinels, and minor amphibole. Near this melt channel, thin streamers of intergranular fills show compositions enriched in Na and Ti. These intergranular 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 most are fractured along these surfaces). {{ More info about grain intergrowth (CPX grains are intergrown) }} Sample CK-7 is notable for the abundance of exsolution lamellae and graphic recrystallization of orthopyroxene and clinopyroxene.

Petrographic study of the host lava and cumulate samples reveals products of multiple stages of melt generation. Sample CK-1 includes abundant phenocrysts and xenocrysts of olivine, as well as small (<1 cm) dunite and peridotite fragments. Samples CK-D1 and CK-D2 are similar but also contain larger (up to 2 cm) dunite and peridotite fragments. The peridotite fragments include both relatively pristine textures similar to those in samples CK-2 to CK-7, to poikilitic and resorbed grains with prominent reaction rims. Taken together, these samples appear to represent interactions with the melt throughout their history.

Lithologic classifications were determined using modal mineralogy. The samples range from lherzolites to clinopyroxene harzburgites and are dominated by olivine and orthopyroxene. All samples contain minor (<1%) spinel. CK-2 has the most fertile composition, with 12.2 clinopyroxene. CK-3 is the least fertile sample, with 0.91 clinopyroxene. Olivine modes range from 65 to 75. Grain size varies between samples but generally has a characteristic scale of 200 µm. The harzburgite CK-3 contains 2 mm orthopyroxene porphyroblasts. All samples are Type I peridotites in the *Frey and Prinz* [[1978](#ref-Frey1978)] classification system.

## Major elements

[[minerals|table]][[major\_elements|figure]]

Error includes both analytical and sampling errors.

Mg# (molar Mg/(Mg+Fe) 100) ranges from 87 to 91. Within each sample, a consistent Mg# for all silicate phases is indicative of Fe-Mg equilibrium. All samples contain <1 spinel. A range of spinel Cr# (molar Cr/(Cr+Al) 100) from 10 to 27 potentially implies variation in degree of partial melting between samples [*Dick and Bullen*, [1984](#ref-Dick1984)].

Sample CK-4 is the most depleted, with a distinct signature of melt extraction

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.

[[whole\_rock\_major|table]]

[[spinel\_cr|figure]]

## Trace Elements

[[spider|figure]][[enrichment\_trends|figure]] [[cpx\_literature\_comparison|figure]][[spinel\_correction|table]]

## Clinopyroxene rare-earth abundances

Clinopyroxene rare-earth element abundances are consistent within each sample, but show several distinct modes of variation between samples.

The cooler samples show progressive depletion in light rare-earth elements (LREEs) and undepleted heavy rare-earths (HREEs). The warmer samples show progressive HREE depletion and enrichment in LREEs. These trends are inverted, so CK-4, the most HREE-depleted sample, is also the most LREE-enriched.

This pattern suggests that the cooler samples were depleted of (relatively more incompatible) LREEs as residues of progressive fractional melting [*Johnson et al.*, [1990](#ref-Johnson1990)]. The warmer group of samples underwent a multistage history of wholesale REE depletion (due to higher-degree melting) followed by later LREE re-enrichment. This overprinting relationship is similar to that observed in mid-ocean ridge peridotites, with ridge crest depletion followed by off axis refertilization via infiltration of low-melt fraction magmas [*Luffi et al.*, [2009](#ref-Luffi2009)]. However, this relationship may also have arisen during melt extraction and entrainment prior ***(confusing, do you mean during eruption/entrainment here?)***to eruption. The latter seems to demand a significant residence time of the hotter xenoliths in , or proximal to, a magma chamber at depth to allow LREE refertilization.

[[trace\_elements|table]]

## Rb-Sr and Sm-Nd isotopes

[[isotopes|figure]]

Rb-Sr and Sm-Nd radiogenic isotope data for clinopyroxene separates show these samples to be derived from the depleted convecting mantle. All samples are enriched in radiogenic ( from 10.3 to 11.0) and depleted in (/ of .702). This pattern of strong depletion in large-ion-lithophile elements rules out a continental lithosphere or Mesozoic mantle wedge source and suggests an origin in the asthenospheric or underplated oceanic mantle [*DePaolo and Wasserburg*, [1976](#ref-DePaolo1976b); *McCulloch and Wasserburg*, [1978](#ref-McCulloch1978)].

[[isotopes\_table|table]]

## 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](#ref-Brey1990)] and TA98 [*Taylor*, [1998](#ref-Taylor1998)] are formulated based on empirical calibration of the two-pyroxene Ca exchange reaction in simple and natural systems. *Taylor* [[1998](#ref-Taylor1998)] is explicitly calibrated to account for errors arising from high Al content. The Ca-in-orthopyroxene (Ca-OPX) thermometer [*Brey and Köhler*, [1990](#ref-Brey1990)] 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 dataset.

[[thermometry|table]]

Core and rim measurements are separated to assess within-sample temperature evolution and potential late-stage 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](#ref-Taylor1998)] reports residuals of calibration of the 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. 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.

Per-sample temperature distributions are constructed by calculating a separate temperature for each individual nearest-neighbor pair of orthopyroxene and clinopyroxene. Analytical errors are propagated through the calculation. The resulting distribution of temperatures (with *n* ranging from 19 to 74 pairs per group) accounts for within-sample variations 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 corresponds well to the relationship between the two thermometers found by *Nimis and Grütter* [[2010](#ref-Nimis2010)]. This relationship can be expressed as for temperatures in ºC. The Ca-in-OPX thermometer generally yields results in between BKN and TA98, with little within-sample scatter, possibly the result of fast diffusion of small amounts of Ca in orthopyroxene.

*Nimis and Grütter* [[2010](#ref-Nimis2010)] shows that TA98 performs well against experimental results in several scenarios and advises its use over BKN. Average TA98 temperatures range from 957 to 1063ºC for cores and 955 to 1054ºC for rims . 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 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. High and variable rim compositions may also be related to the eruption event. Several samples show a wide spread rim temperatures, perhaps indicating differential exposure to interstital fluids. This internal variability is minor: for grain cores in all samples, the mean of pairwise analyses is within a few degrees of the temperature calculated by averaging all pyroxenes across the sample. This implies that the bulk of the temperature signature is based on the equilibrium state of the sample.

The samples show a clear bimodal distribution in equilibration temperature, forming distinct groups centered at 970ºC and 1060ºC (TA98) . The cooler group contains CK-2, CK-5, and CK-7, while the hotter contains CK-3, CK-4, and CK-6. This division between these two groups is robust and apparent in all thermometers. The high-temperature samples show re-enrichment in LREEs and contain chromian spinels. This temperature distribution may signify two distinct depths of origin for the studied xenoliths. 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.

[[temp\_comparisons|figure]][[temp\_summary|figure]]

### REE-in-pyroxene thermometry

The *Liang et al.* [[2013](#ref-Liang2013)] REE-in-two-pyroxene thermometer is used as an external check on the major-element equilibration temperature based on a different system. The relative immobility of REEs allows assessment of equilibrium temperatures over longer timescales than those queried with major-element thermometry.

Sample CK-4 shows major disequilibrium in the light and medium REEs, with only elements heavier than Ho retaining an equilibrium signature.

***(more descriptive text here?)***

### Depth constraints

[[depth|figure]]

Peridotite barometers are based on the decreasing Al content of orthopyroxene with depth [*Brey and Köhler*, [1990](#ref-Brey1990); *Nickel and Green*, [1985](#ref-Nickel1985); *Nimis and Taylor*, [2000](#ref-Nimis2000)]. However, in the absence of garnet, the reaction is purely thermometric, with nearly vertical isopleths in P-T space [*Gasparik*, [2000](#ref-Gasparik2000); *Herzberg*, [1978](#ref-Herzberg1978)]. With no reliable geobarometers for spinel peridotites, several less robust measures are used to evaluate the depth of the xenolith source.

Minimum entrainment depths must be greater than ~30 km, the depth of both the Moho near the eruption site [*Tréhu*, [1991](#ref-Trehu1991)], and the plagioclase-spinel peridotite facies transition [*Green and Ringwood*, [1970](#ref-Green1970a)]. The xenolith entrainment depth must exceed the boundary of spinel stability, which is composition-dependent and poorly constrained for natural systems, but thought to lie over the 50-80 km depth interval [*Gasparik*, [2000](#ref-Gasparik2000); *Kinzler*, [1997](#ref-Kinzler1997); *Klemme*, [2004](#ref-Klemme2004); *O’Neill*, [1981](#ref-ONeill1981)].

[[ree\_temperatures|figure]]

Projection of TA98 temperatures onto a geotherm using a surface heat flux [*Turcotte and Schubert*, [2002](#ref-Turcotte2002)] of 90 mW/m^2, corresponding to measured heat flows in the study region [*Erkan and Blackwell*, [2009](#ref-Erkan2009)] yields model depths ranging from 31-39 km (fig. 14). However, the surface temperatures that the heat flow data are based on may be transient, especially in a rapidly evolving region that has experience subduction, slab window opening and San Andreas transform offsets. **((**May be affected by differences in regional thermal conductivity *Lachenbruch and Sass* [[1980](#ref-Lachenbruch1980)] concluded that CRTA could not be generated by conductive fault heating alone, but was also hesitant to invoke fluid flow to regionally average thermal gradient. *Kennedy* [[1997](#ref-Kennedy1997)] shows the presence of mantle fluids in the San Andreas fault zone, giving credence to the idea deep fluid transport. These would have the effect of increasing temperatures. ***This text in the parens is confusing, and needs to be more clearly written with some specific point in mind))***

Equilibration pressure measurements are attempted for the peridotite xenoliths using the *Köhler and Brey* [[1990](#ref-Kohler1990)] Ca-in-olivine barometer. This barometer is explicitly calibrated for spinel peridotites but widely disregarded based on poor resolution, vulnerability to late-stage diffusion, and dependence on low Ca concentrations in olivine [*Medaris et al.*, [1999](#ref-Medaris1999); *O’Reilly*, [1997](#ref-OReilly1997)]. This technique yields a broad distribution in model depths, largely coincident with the spinel stability field (Fig. 14). The depth distributions are largely normal, with modes ranging from 35 to 52 km. The low and high-temperature cohorts remain separable, with high-temperature samples generally showing deeper equilibrium depths. The low-temperature samples in particular have***???***

***(((***The Cr# of the hotter samples somewhat expands the stability of spinel against garnet. Several studies have attempted to estimate the magnitude of this effect [*Klemme*, [2004](#ref-Klemme2004); *Klemme and O’Neill*, [2000](#ref-Klemme2000); *O’Neill*, [1981](#ref-ONeill1981)]. Figure shows the depth increases estimated by the simplified method given in *O’Neill* [[1981](#ref-ONeill1981)], showing that potential depths increase by up to 10 km. If these deeper depths are valid, the outlined geotherm conforms generally to temperature bounds of 700--1100ºC measured by seismic tomography at 50--100 km depth for coastal California [*Goes and Lee*, [2002](#ref-Goes2002)]. ***This paragraph is confusing. It needs more explanation, and closer ties to what is shown on Figure 14. Also the Fig. 14 caption needs more, including references for the 100, 95 and 90 geotherms)***

# Discussion

## Petrology

The peridotite samples from Crystal Knob show a range of depletion in major and trace elements. They are isotopically depleted, with an εNd of +10, and / of .7029. ***(Daven, the Nd and Sr data in Table 4 are given as measured. In the table you should also show the time corrected Nd and Sr isotopic ratios and epslon Nd value for a correction of 1.65 Ma. You should then refer to these time corrected numbers in the text and show these values on the Figure 10 Eps Nd-Sr plot. The corrections will be small, but since we have dated the host it would be best to use the time corrected values for time of entrainment.*** This corresponds to the depleted asthenosphere, or convecting upper mantle of Hoffman (2004), with a mantle upwelling source that has seen no contribution from the western North American crust or continental lithosphere, more generally.

Pyroxene-exchange geothermometry shows that the peridotite samples form two groups in temperature with centroids separated by roughly 60ºC. These temperatures seem to correspond to the samples being sourced along a depth gradient. For reasonable slopes of the regional geotherm, this range of temperatures may sample a depth range of 5-15 km within the mantle lithosphere. These depths must be greater than 30 km, depth to Moho (appropriate reference-***right?)***, and less than 60-90 km based on the composition-dependent extent of the spinel field.

Rare-earth exchange thermometry shows the same two groupings of temperatures. Temperatures measured for the low-temperature cohort are most comparable to the TA98 results. Given that the TA98 method has been found to perform best among the pyroxene-exchange thermometers by *Nimis and Grütter* [[2010](#ref-Nimis2010)], 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](#ref-Liang2013)], 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 shows the most signatures of intergranular melt channels

The low-temperature cohort ranges from essentially undepleted (CK-7 has a flat chondrite-normalized REE profile) to low levels of depletion characteristic of the least-depleted abyssal peridotites [*Warren*, [2016](#ref-Warren2016)].

The high-temperature cohort seems to have been both depleted (based on low HREE concentrations) and re-enriched in trace elements to a much greater degree than the low-temperature cohort.

***(Daven, do you think that the high-T group attained its signature at the Farallon-Pacific spreading ridge, or from re-heating of underplated Farallon mantle by the slab window convection event?).***

### Rare-earth temperature

* Deeper samples are refertilized in LREE
  + Over and above amount of LREE in higher samples
  + Interacted with fertile fluid, but not enough to re-equilibrate major elements
* CK-4 records a significantly higher temperature for LREE
  + metasomatic processes
  + re-equilibrated at lower T?

### Trace elements

* CK-6 (hottest sample) is enriched despite trace-element history of LREE depletion and re-enrichment
* Clinopyroxene Eu anomalies: possibly a signature of ghost plagioclase
* Not a geochemically homogenous set of samples, even though they likely originated from same mantle source
* The samples are depleted differently, as shown by different amounts of major-element and Cr depletion
* Looks like MOR depletion patterns followed by re-enrichment

The variable depletion and re-enrichment in REEs is potentially a signal of mid-ocean ridge emplacement. However, this pattern does not imply a unique cause and could potentially arise from deep magma migration. The assignment of the samples to two temperature cohorts is robust and can be related to other features, particularly the observed patterns in REE abundances .

### Host lava

Extension in the lower crust? Of family with small eruptive episodes such as Coyote Creek (based on eruptive age)

Fossil heating event affecting CK-4 LREE. This is not the most recent perturbation because it's not reflected in the major element thermometers

## Origin of mantle lithosphere beneath the Crystal Knob volcanic neck

Rb-Sr and Sm-Nd isotopic 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 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 (Cowan, 1978; Saleeby et al., 1986; Seaton et al., 2012; Chapman et al., 2016a), 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. The geologic history of the region integrated with crustal structure constraints pose viable alternatives for these mechanisms.

### 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 (*Thorkelson and Taylor*, [1989](#ref-Thorkelson1989); Bohannon and Parsons, 1995; Atwater and Stock, 1998; 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 (Bohannon and Parsons, 1995; Atwater and Stock, 1998; ten 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 (Trehu and Wheeler, 1987; Brocher et al., 1999), and sharp inflections in its upper surface (Trehu, 1999) 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 18 by restoration of the San Andreas dextral transform system (Matthews, 1976; Dickinson et al., 2005; Chapman et al., 2012; Hall and Saleeby, 2013; 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 on Figure 18 along with the principal upper plate batholithic exposures. The intervening areas left white denote latest Cretaceous-Cenozoic overlap strata. On Figure 18 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 et al., 2003).

The Figure 18 reconstruction exhibits the first order crustal relations that pose three highly plausible origins for the sub-Crystal Knob mantle lithosphere: 1. shallowly ascended asthenosphere within the Pacific-Farallon slab window (Atwater and Stock, 1998); 2. subduction underplated, or stalled Monterey oceanic microplate (Bohannon and Parsons, 1995); or 3. underplated Farallon plate mantle lithosphere nappe(s) that lie in structural sequence with the upper mantle duplex resolved beneath the Dish Hill xenolith location (Luffi et al., 2009).

SLAB WINDOWS AND MONTEREY PLATE

Figure 18 shows the surface projection of the hypothetical slab window and the partially subducted Monterey plate at ca. 19 Ma (Wilson et al., 2005). The slab window is thought to have 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 an ~250 km long segment of the Pacific-Farallon ridge as an oblique rift that was rotated ~25° clockwise from the Pacific-Farallon rift axis (Atwater and Severinghaus, 1989). It’s 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 panels (Bohannon and Parsons. 1995; Stock and Atwater, 1998). The current position of the Monterey plate, relative to 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 possible additional dextral offsets in the offshore region (Fig. 3). A number of workers have suggested that translation of the Monterey plate along the San Andreas system entailed significant sub-horizontal fault segments that accommodated dextral displacements (Furlong et al., 1989; Pikser et al., 2012). As of yet, however, all remotely imaged segments of the transform system have been shown to be steeply oriented (Dietz and Ellsworth, 1990; Brocher et al., 1999; Yan et al., 2005; Titus et al., 2007; Yan and Clayton, 2007; Ozacar and Zandt, 2009).

According to the Figure 18 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 it’s 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 Bohannon and Parson’s (1995) reconstruction. Coupling of Monterey plate divergent motion across the subduction megathrust break is hypothesized to have driven dextral transrotational rifting (Bohannon and Parson, 1995). As western Transverse Range rock panels rotated into their current position from transrotational rifting, the Monterey plate continued its northward displacement along the San Gregorio-Hosgri fault system (Fig. 3). Note that on Figure 18 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 ~16° E dip to the Monterey plate offshore, with a typical abyssal crustal thickness, juxtaposed against a nearly flat thickened lower crustal layer beneath the Nacimiento Franciscan (Trehu, 1991; Nicholson et al., 1992). 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, as dismissed in the discussions above.

The analysis presented above argues against models of the Monterey plate having been translated horizontally as a “dangling slab” through the upper mantle of the coastal region of southern to central California by its coupling to Pacific plate motions (Van Wijk et al., 2001; Pikser et al., 2012; Wang et al., 2013). This can be argued against on the basis of seismological, geodynamic and surface geological relations. All such models rely on the untenable notion that a structurally continuous mafic layer, representing stalled Monterey plate, constitutes the lower crust beneath the entire coastal central California and offshore region (see above). 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 (Fig. 1), commonly called the “Isabella anomaly”. However, seismological and geodynamic studies much more rigorously show that this anomaly represents the convectively mobilized mantle wedge, or mantle lithosphere, derived from beneath the eastern and southern Sierra Nevada batholith (Zandt et al., 2004; Frassetto et al., 2011; Gilbert et al., 2012; Saleeby et al., 2012; Jones et al., 2014; Levandowski and Jones, 2015). 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 Great Valley and Sierra Nevada (Fig. 3), these studies show that the volume of the Isabella anomaly far exceeds a reasonable volume estimate 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 ignored in the dangling slab hypothesis. Surface geological effects of such melting and topographic transients are closely correlated to the convective mobilization of the sub-Sierran mantle lithosphere (Ducea and Saleeby, 1998b; Farmer et al., 2002; Saleeby et al., 2013; Cecil et al., 2014; Levandowski and Jones, 2015). 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 (Fig. 3; Bohannon and Parsons, 1995; Wilson et al., 2005). This is in line with the Monterey slab’s limited down-dip extent as bounded by the Monterey-Farallon slab window segment shown on Figure 18. 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 (Saleeby et al., 2013; Cecil et al., 2014). 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 (Fig. 3), and also have a firm geodynamic basis as such (Burkett an Billen, 2009).

According to the Wilson et al. (1995) reconstruction of the Pacific-Farallon slab window and adjacent Monterey plate (Fig. 18), the Crystal Knob eruption site was located above a slab window in the early Neogene, proximal to 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 (Hurst, 1982; Sharma et al., 1991; Cole and Basu, 1995; 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, which we return to below.

UNDERPLATED FARALLON PLATE MANTLE NAPPES

The reconstruction of the Crystal Knob eruption site to its pre-San Andreas position (Fig. 18) poses a highly viable alternative for the development of the site’s underlying mantle lithosphere. This is particularly viable in light of the neck having erupted through the Nacimiento belt of the Franciscan complex, immediately adjacent to the current outer limit of Salinia crystalline nappes (Fig. 1). Accretion of the Nacimiento belt occurred in the Late Cretaceous beneath the outer reaches of the Salinia nappe sequence (Hall and Saleeby, 2013; Chapman et al., 2016a). In their core area the Salinia nappes rode westwards on slightly older, higher metamorphic grade, Franciscan rocks that are shown on Figures 1 and 18 as windows into subduction channel schists (Barth et al., 2003; Kidder and Ducea, 2006; Ducea et al., 2009). As discussed earlier, 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 (Saleeby, 2003; Chapman et al., 2010, 2012, 2016b). 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 (Fig. 1) that suggest the presence of a mantle lithosphere duplex with multiple Farallon plate upper mantle nappes in structural sequence beneath a 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 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 17 we present a model for Farallon plate mantle lithosphere having tectonically underplated the Mojave-Salinia-Nacimiento segment of the SW Cordilleran convergent margin in the Late Cretaceous (after Saleeby, 2003 and Luffi et al., 2009). This is shown to have occurred in conjunction with shallow flat subduction of the Shatsky Rise conjugate LIP (after Saleeby, 2003; Liu et al., 2010). The approximate age of Farallon plate entering the trench is shown on each frame [after *Seton et al.*, [2012](#ref-Seton2012)]. Crustal deformation, timing and thermal conditions, as applied to our thermal modeling presented below, are integrated from Kidder and Ducea (2007) and Chapman et al. (2010, 2012, 2016a). Figure 17a 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](#ref-Kidder2006)].

In Figures 17c and d we adopt the focused slab rollback and mantle lithosphere underplating models of *Saleeby* [[2003](#ref-Saleeby2003)] and *Luffi et al.* [[2009](#ref-Luffi2009)], for the dynamic response of normal thickness oceanic lithosphere following the crustal 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 17c to d transition, accelerated rollback is accomplished 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 (Figs. 17c and d), an estimated ~700-800 °C control on this transition (Warren and Hirth, 2006; Burgmann and Dresen, 2008; Mei et al., 2010) 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 (Trehu, 1991; Brocher et al., 1999). 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 17d 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 faulting in our analysis. We focus now to thermal history in order to more thoroughly pursue the possible origins of the sub-Crystal Knob mantle lithosphere.

#### Thermal considerations

Regardless of the generalized lithospheric structure depicted in Figure 17, 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). 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, its volume has not been consistent with other episodes of shallow asthenospheric upwelling in the Cordillera, particularly in the forearc region of coastal central California [*Humphreys*, [1995](#ref-Humphreys1995) ***not familiar with this ref, please double check for its applicability***]. This 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 (Fig. 14), 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.

## Geothermal implications

The Farallon Plate, Monterey Plate, and slab window scenarios for mantle lithosphere emplacement all imply a peridotite composition with a depleted (convecting-mantle) isotopic and trace-element signature. Though petrographic and geochemical variations in our peridotite xenolith samples can provide information on depletion, enrichment and possibly melting history, they cannot discriminate between these potential depleted convecting mantle sources. However, these emplacement scenarios present potentially distinct thermal structures due to the large differences in timescales of cooling.

The Farallon-- and Monterey--plate scenarios are qualitatively similar, with initial emplacement beneath a mid-ocean ridge and cooling on the seafloor. 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 entrainment and eruption. However, the timescales of cooling are significantly different. In the Farallon plate scenario, the maximum age of underplating is 70 Ma, based on the youngest ages of the most pertinent (Sierra de Salinas and correlative San Emigdio-Rand) schist bodies [Barth et al., 2003; Grove et al., 2003; Saleeby et al., 2007; *Chapman et al.*, [2010](#ref-Chapman2010)]. Seafloor being subducted at that time was 40 Myr old [Seton et al., 2012; *Liu et al.*, [2010](#ref-Liu2010)]. This 110 Myr cooling history implies a relatively cold modern geotherm. The potential Monterey Plate mantle lithosphere would have been emplaced under the ridge at 27 Ma (corresponding to the chron 9 magnetic anomaly) and subducted shortly thereafter [*Atwater and Stock*, [1998](#ref-Atwater1998); Wilson et al., 2005], leaving a much shorter period for relaxation of the geotherm. The emplacement of slab-window asthenosphere directly under the coastal central California crust entails the truncation of a low-temperature forearc geotherm at 19 Ma [*Atwater and Stock*, [1998](#ref-Atwater1998)] and the substitution of an asthenospheric adiabat below this level. This scenario would provide the hottest modern geotherm, which, according to *Erkan and Blackwell* [[2008](#ref-Erkan2008)], is too hot to correspond to the modern regional geotherm. ***(Daven, are they referring to the region under CRTA, or the cooler Great Valley and Sierra Nevada)***

## Thermal modeling

***(somewhere near the beginning of this section we need a very explicit statement of what the heat flow data are, what the anomalies are, what is CRTA, and how these data have been interpreted-remember, very few of our readers will be familiar with these data, and potentially our most critical reviewers will be very familiar with these data, and their current interpretation. Here is a comment on Erkan and Blackwell that we should incorporate into our qualitative analysis, I leave it to you to fit it into this section in your logical progression, and wording:***

They corner themselves into needing special circumstances to explain CRTA like rapid recent uplift (which could be a contributing factor-Ducea et al., 2003-in added refs), shear heating or thermal relaxation of stalled slab nonsense, just because they insist on the low heat flow from the Great Valley-Sierra Nevada to be caused by a stalled slab-classic circular reasoning. We need to go on the offensive here. First thermal relaxation of a young stalled slab would not be much different than a deep slab window. Low heat flow values under the entire GV-SN are irrespective of where plate reconstructions show slab windows or (Monterey) microplates, but map out very close to the distribution of where Mesozoic mantle wedge/continental mantle lithosphere is still intact beneath the crust. Compare E & B’s figure 1 map with the structure contours that I added to Figure 3 for the Isabella anomaly as well as the symbolism for the adjacent SE Sierra region that has lost its mantle lithosphere. The structure contours show just the deeper part of the anomaly, leaving lots of additional cool Mesozoic mantle lithosphere beneath the Great valley to axial Sierra Nevada making the correlation with E & B’s figure 1 pattern tighter. Then we can now layer on what we know about lithosphere–scale heat production in the Sierra Nevada from Brady et al. (2006-emailed to you separately and in added ref list). I also email to you our “Epi II” paper on delamination which has a helpful section on SN heat flow, right column, pg. 404). Bottom line is that GV-SN heat flow, except where mantle lithosphere has delaminated, reflects a relict Late Cretaceous geotherm developed during extended flat slab subduction at the time or arc cessation. Crystal Knob and Coyote Lake, in the Coast Ranges clearly show that there IS recent upwelling under the Coast Ranges, not to mention the Clear Lake volcanic field. E and B have about every aspect of their interpretation wrong, but we should be careful and constructive in our criticism of their work, for their data is very useful.

* Model setups corresponding to tectonic scenarios
* Model the relaxation of the geotherm during subduction and underplating

### Setups

* Global Depth and Heat model for oceanic crust
* Forearc geotherm model [*Royden*, [1993b](#ref-Royden1993)]
* Standard values for oceanic and continental material properties
* We ignore the effects of a downgoing slab below ***(Not quite sure in what context you are referring to here)***
  + may be significant over long timescales
  + will generally serve only to cool geotherms

Given the range of potential geothermal scenarios, models for the emplacement of depleted mantle lithosphere under the central coastal California region can be tested by comparison of their implied geothermal structures with xenolith geothermometry. However, this analysis is a crude approximation due to a lack of well-constrained geobarometers for spinel peridotites.

[[model\_setups|table]]

## Model setup

Evaluation of potential thermal scenarios for the mantle lithosphere sampled by Crystal Knob require constraints on the implied present geotherm. Forward-modeling of the initial thermal conditions will help assess the level of separation between the scenarios after 25 Myr or more of re-equilibration beneath the continental margin. To distinguish between potential emplacement mechanisms for this mantle lithosphere, a forward model of the geotherm implied by each case is constructed. A model based on the one-dimensional heat-flow equation is used to track a vertical profile through the lithosphere. This framework is used to follow the thermal state of the xenolith source region from its most recent thermal peak, regardless of tectonic setting, to its final emplacement beneath the Crystal Knob eruption site.

### Oceanic geotherm

For the stalled-slab and Late Cretaceous rollback scenarios ***(why not call them “stalled Monterey plate” and “Farallon mantle nappe” scenarios)***, the Global Depth and Heat (GDH) model [*Stein and Stein*, [1992](#ref-Stein1992)] is used to trace the thermal evolution of the oceanic lithosphere from its emplacement at the spreading ridge until subduction. This model is an 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 ***??what??***(e.g., *Fowler* [[2005](#ref-Fowler2005)]) and tends to yield higher geotherms for old geothermal lithosphere. All 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](#ref-Stein1995); *Stein and Stein*, [1992](#ref-Stein1992)]. This may result in overestimates of geothermal gradients for the youngest subduction scenarios.

The oceanic mantle is defined with a thermal conductivity of , a specific heat of , a density of , and radioactive heat generation of . For simplicity, oceanic crust is not considered separately within the model framework.

Increasing the thermal conductivity of the model domain substantially flattens the modeled geotherms.

[[reconstruction|figure]]

### Supra-subduction geotherm

The geotherm of the forearc wedge during subduction is calculated using the *Royden* [[1993a](#ref-Royden1993a)] analytical solution for the steady-state thermal structure of continuously-subducting systems. Shear heating on the subduction thrust is taken to be . Rates of surface erosion in the forearc and subduction accretion are taken to be 0.

The coastal California accretionary crust is represented homogenously as a material with a thermal conductivity of ***?***, specific heat capacity of ***?***, density of ***?***and a radiogenic heat flux of ***?***, values that are close to average for the continental crust [*Fowler*, [2005](#ref-Fowler2005)] and those used by *Kidder et al.* [[2013](#ref-Kidder2013)] to model the thermal conditions along the Late Cretaceous shallow subduction megathrust segment.

[[model\_results|figure]]

### Underplating

To simulate subduction and underplating, the forearc geotherm is stacked atop the modeled oceanic geotherm 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](#ref-Guyer2009)]. Explicit and implicit finite difference approaches are combined using a two-sweep Crank-Nicholson technique [*Crank and Nicolson*, [1947](#ref-Crank1947)] to ensure a stable result.

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* [[1993a](#ref-Royden1993a)] thermal model using the parameters outlined above. For all cases, the final depth of the underplated subduction interface is taken to be 30 km (corresponding to a "flat-slab" regime), 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. Increasing the slab dip angle will have the effect of speeding the equilibration of the geotherm, but interface temperatures will be lower. The overall effect on the evolution of the thermal scenarios appears to be minimal.

### Limitations of the model

This model framework has several simplifications. Surface erosion after underplating is taken to be zero. 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 and Saleeby*, [2012](#ref-Chapman2012)]. Also, the underplating scenario is modeled incompletely -- when the trench interface jumps to with the emplacement of a nappe of oceanic mantle 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 from the final trench interface, the scope for further rollback after emplacement of the nappe is of interest is limited.

The potential for disequilibrium processes such as slab window heating to disrupt the geotherm at depth during the long residence time beneath the crust are not considered in the model. ***(This is confusing, I thought that was one of our scenarios. This needs to be clarified)***

There are no reliable estimates of the mantle heat flux that cover the model domain, and the thermal environment for underplated mantle lithosphere is complicated by the presence of a subducting slab below the model domain at some depth. The model is run to great depth to avoid any influence on the surface geotherm. However, Farallon and forearc scenarios can be treated as maximum temperatures because of the influence of the overriding slab. ***(not sure what you are talking about here. What is “forearc scenarios”)***

### Model results

[[model\_tracers|figure]]

Model results are presented in Figs. 19 and 20? . For the Farallon mantle nappe scenarios ***(why plural?)***, the *Royden* [[1993a](#ref-Royden1993a)] forearc model predicts low temperatures (~235-245) at the subduction interface. This is quite low relative to the temperatures predicted for the Pelona schist (~700 C) by *Kidder et al.* [[2013](#ref-Kidder2013)], or 700-800 C derived as an emplacement constraint for the Sierra de Salinas schist [***correct ref here is:*** Kidder and Ducea, 2006 ***,right?****Barth et al.*, [2003](#ref-Barth2003), *Grove et al.* [[2003](#ref-Grove2003)], *Ducea et al.* [[2009](#ref-Ducea2009)]]. In the stalled Monterey plate scenario, temperature is predicted to be 980 ***(where, subduction interface, base of lithosphere???)***. Despite this uncertainty in subduction conditions, the model is much more sensitive to the thermal history of the oceanic plate than to the forearc geotherm. For all longer run-time models (both the forearc ***???***and Farallon scenarios), this may prove to be a major factor.

The baseline scenario with Late Cretaceous rollback and temperatures pinned to the Sierra de Salinas ***(not Pelona, right*** Pelona ***)*** schist has a very similar final thermal structure to the older stalled slab scenarios ***(do we show these in the model?)***, showing that the high forearc temperatures experienced during Late Cretaceous flat slab subduction and schist metamorphism do not have a lasting impact on the thermal structure of the subducted plate ***(right? Rather than:***Salinian basement***)***. Thus, downward heat conduction ***(right? =***these***)*** cannot be explain the elevated temperatures recorded in the Crystal Knob peridotites.

The model predicts much higher temperatures ***(do you mean to say a much higher thermal gradient, or surface heat flow, or both, or if not much higher temperatures where, and relative to what?)***for the slab window than for the Farallon mantle nappe or stalled Monterey plate models, which corresponds to the findings of previous studies (*Erkan and Blackwell* [[2008](#ref-Erkan2008)]. However, these studies concluded that the effect must be due to a stalled slab, where in fact underplated Farallon mantle nappes satisfy the surface heat flow data equally well. (***Right, instead of:***it seems to be possible that a Farallon-plate lithosphere is to blame, as this produces the same temperatures as found for the stalled slab.***)***

It is possible that the slab could be older. But, Thermal relaxation of an old slab would not contribute any heat to explain elevated heat flow values in Coast Ranges (CRTA). ***What is CRTA? Maybe you should email me Erkan and Blackwell. Are these, or is this, several points, or a single point of high heat flow measurements within the regional field of low heat flow measurements. If so where are these point(s). They could be important in terms of deep heat advection, in terms of why a 1.7 Ma lava would erupt in the region)***

*Erkan and Blackwell* [[2008](#ref-Erkan2008)] pointed out that heat flow for the stalled slab scenario was too low to fully explain anomalous heat flows in the Coast ranges. Model-predicted heat flows lower than measured values could be explained by added heat flux from shear heating [*Thatcher and England*, [1998](#ref-Thatcher1998)] or erosion [**???**; *Mancktelow and Grasemann*, [1997](#ref-Mancktelow1997)] ***or fault controlled heat advection form a very deep slab window remnant?***

[[model\_comparison|figure]]

Estimating erosion is beyond the scope of this study, but pulses of recent erosion in the Coast Ranges are ***see Ducea et al. (2003) for rapid late Cenozoic uplift of the Santa Lucia’s***

**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 on Figure 16 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 of a regionally extensive Monterey Plate “dangling slab” extending far to the east of the San Andreas fault as suggested by Pikser et al. (2012). Depending on the original scale of Monterey Plate underthrusting beneath the southern California borderland region (Fig. 18), 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 (Fig. 3), an argument could be made that an underthrust portion of the Monterey Plate has been translated northwards horizontally beneath the Crystal Knob eruption site (Fig. 16b). 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 surface levels by transients in dynamic topography, as well 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 though. Furthermore, such horizontal translation of a previously underthrust slab offers little in terms of melting mechanisms neither for Neogene nor 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.

On Figure 16c 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 18 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 tectonically above underplated Farallon plate mantle nappes. The structural profile shown on Figure 16c 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 (Figs. 17 and 18). Partial subduction, or stalling, of the Monterey plate occurred along the outer edge of Franciscan complex further south than rocks of the Nacimiento belt (Fig. 18).

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, the thick and relatively cool lithospheric lid that the slab window of this region opened beneath inhibited widespread voluminous volcanism in response to asthenospheric upwelling. The time lag between deep slab window opening and the ca. 1.7 Ma eruption 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 50-80 km, which is 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), based on thermal decay relations (Doin and Fleitout, 1996), would have its lithosphere-asthenosphere boundary at ~35 km. The offshore Monterey plate is at the edge of Li et al’s. (2007) resolution, 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 (Trehu, 1991), bringing the thermal maturation theoretical depth for the transition much closer to the Li et al’s. (2007) observed depth. Crystal Knob is located ~15 km east of the Hosgri fault, with it’s 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, 2012) 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, 2001; Titus et al., 2007) indicate that at lower crustal-upper mantle levels Hosgri fault shear could be distributed across 10’s of kilometers normal to the fault surface (Fig. 16c). Eruption of small volume basaltic flows of Plio-Pleistocene age, some with lower crust and upper mantle xenoliths also occurred ~150 km north of Crystal Knob along the San Andreas-Calaveras fault bifurcation zone (Jove 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 (Fig. 16c) 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 (Jove 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 (Erkin and Blackwell, 2008, Fig. 1), further suggesting recent mobilization of asthenospheric mantle that was initially emplaced into the early Neogene slab window.

# Conclusion *(Daven, once you have finished all the text above write a very succinct Conclusion section and I will look it over for possible edits during my read of the completed manuscritp draft.).*

* Slab window is really hot!
* Farallon plate and older forearc are really cold
  + Converging on “steady-state” value
* Farallon heated from below seems to work well
  + Provides a thermal buffer for low-heatflow coast range and can potentially help explain the Coast Range Thermal Anomaly

An initial petrologic study of the Crystal Knob peridotite xenoliths has been conducted, and thermobarometry has been applied to examine the thermal state of their mantle source. These xenoliths sample the lithosphere at moderate to deep (45-80 km) depths and are divided into two groups which record subtly different temperatures. Sm-Nd and Rb-Sr isotopes show that the samples are derived from the depleted convecting mantle. Given this, the samples are derived from some generation of the Farallon downgoing slab. However, the thermal models approach equilibrium at long timescales, and depth constraints are imperfect at best, so distinguishing between the plateau-emplacement and older underplating scenarios is difficult at best. No compelling features to distinguish these two scenarios are as yet identifiable.

Continuing refinement of thermometry and depth constraints along with forward-modeling of temperature scenarios will potentially improve the ability to resolve the structure of the geotherm, but the uncertainties inherent in spinel peridotite barometry may prove crippling. Regardless, the completed work and continuing petrologic analysis of the host basalt and cumulates will lead to an improved understanding of the coastal California mantle lithosphere.

# Figures *(note that in my text edits and in the figures that I edited I used the numbering system that you used in the G-cubed formatted version that you gave me at GSA)*

***Missing: Fig X Ar/Ar step heating plot, maybe also photomicrographs showing cumulate dunite and non-cumulate peridotite textures, only if very clearly distinguished!***

***Perhaps the new Figure 3 map that I worked over should be figure 1, depending on what is referred to first in text. We should talk about the way the rock units are designated on this figurer, it’s not so good and should be altered. Also, figure 18 should probably be before figures 16 and 17. Actually, I will stop here! Looking below non of the captions agree in content with the figured manuscript that you have emailed me. I will leave it to you to number the figures in sequence with text subjects.***

***Whatever figure the old Figure 3a, now Figure 3 is in my files, here are critical references:***

Isabella anomaly and adjacent upper mantle structure: Jones and Phinney (1998), Zandt et al. (2004), Frassetto et al. (2011), Gilbert et al. (2012), Jones et al. (2014), Levandowski and Jones (2015)

Transverse Ranges anomaly: Schmandt and Humphreys (2010)

Coyote Creek xenolith location Jove and Coleman (1998), Titus et al. (2007). Let me know if you need help with the other xeno loc refs, you have them in yours+my additions ref lists all ready.

**Figure 1:** Mineral classification images of each sample (1" round thin-section) created atop coregistered electron backscatter optical imagery and showing textural variation within Crystal Knob suite.

**Figure 2:** Major element composition (oxide %, normalized to 100%) of xenolith samples recalculated from modal mineralogy.

**Figure 3:** Per-element equilibrium temperature for REE thermometry of xenolith samples. Equivalent to for in REE elements. Thin horizontal lines represent regression of against to find best-fitting temperature across all REE. Data points not corresponding to a dominantly horizontal line signify disequilibrium between Opx and Cpx.

**Figure 4:** Summary of temperature data showing the two temperature cohorts of the dataset.

**Figure 5:** Tracers for each modeled scenario following the temperature-time evolution of particles at 40 and 80 km depth in the model domain (dashed and solid lines, respectively).

**Figure 6:** 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](#ref-Wijk2001)] and *Pikser et al.* [[2012](#ref-Pikser2012)], among others.

**Figure 7:** Cross sections showing the evolution of southern California during subduction of a large oceanic plateau during the late Cretaceous.

**Figure 8:** Radiogenic isotope measurements

**Figure 9:** Clinopyroxene and orthopyroxene rare-earth element abundances.

**Figure 10:** Comparisons of different model scenarios, with most likely of each grouping highlighted.

**Figure 11:** Cr# vs. Mg# for spinels, showing two groups of samples with low and high Cr content.

**Figure 12:** Results of two-pyroxene thermometers

**Figure 13:** Comparison of results from pyroxene major-element thermometers.

**Figure 14:** Material properties used in modeling

**Figure 15:** Average composition of xenolith mineral components

**Figure 16:** Spinel ferric iron content

**Figure 17:** Tectonic reconstruction of the California margin at 19 Ma showing tectonic provinces of Salinia restored for offset along the San Andreas system. The view shows the disaggregated Mojave--Salinia batholith and surface outcrops of subduction channel schists in the Mojave province. Reconstruction of Salinian lithologic features is after *Schott and Johnson* [[1998](#ref-Schott1998)],*Schott and Johnson* [[2001](#ref-Schott2001)],*Chapman and Saleeby* [[2012](#ref-Chapman2012)],*Dickinson et al.* [[2005](#ref-Dickinson2005)]. Reconstruction of the evolving slab window and microplate detachment after *Wilson et al.* [[2005](#ref-Wilson2005)].

**Figure 18:** Paired Sm-Nd and Rb-Sr isotopes.

**Figure 19:** FeO vs. MgO for electron microprobe measurements of grain cores, showing range in major-element depletion between samples. Dotted lines show Mg# levels.

**Figure 20:** Modal composition of Crystal Knob perodotites. Abyssal [*Asimow*, [1999](#ref-Asimow1999); *Baker and Beckett*, [1999](#ref-Baker1999)] and Dish Hill [*Luffi et al.*, [2009](#ref-Luffi2009)] peridotite compositions are shown for comparison.

**Figure 21:** Map of southern California showing the geologic setting of Crystal Knob and its placement relative to other tectonic features, such as Neogene sinistral faults and the stalled Monterey microplate. Xenolith sampling locations that backed up previous studies are shown: the Central and Eastern Sierran suites show a record of delamination of a batholithic root [*Ducea and Saleeby*, [1996](#ref-Ducea1996)] and Mojave sites show underplating of Farallon-plate lithospheric nappes during the Cretaceous [*Luffi et al.*, [2009](#ref-Luffi2009)]. The reconstructed position of Crystal Knob is shown in panel *A*,. This reconstruction was created independently using the regional paleomagnetic framework of *Wilson et al.* [[2005](#ref-Wilson2005)] with the restoration of slip along San Andreas--system faults [*Dickinson et al.*, [2005](#ref-Dickinson2005)] approaches. The methods agree to within 5 km on the position of the Crystal Knob source locale at 19 Ma. Crystal Knob can be restored to ~350 km SE of its current location, accounting nicely for ~310 km displacement on modern San Andreas Fault and ~40 km remainder on the Rinconada fault within the Salinian block. Panel *B* shows the location of key mid-Miocene hypabyssal intrusives (the Morro Rock--Islay Hills complex [*Stanley et al.*, [2000](#ref-Stanley2000)] and the Cambria Felsite [*Ernst and Hall*, [1974](#ref-Ernst1974)])

**Figure 22:** Trace element abundances recalculated from whole-rock major elements.

**Figure 23:** 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**: Similar to **A**, but with heating of mantle lithosphere from below without wholesale replacement. **C**: Breakoff of slab at depth (presumably the rotated Monterey microplate) and dextral strike-slip translation beneath te continental margin, replacing marginal continental mantle lithosphere .

**Figure 24:** Results of model runs for each of the three families of tectonic scenarios.

**Figure 25:** Depths from Ca-in-olivine geobarometry plotted against TA98 temperature. 90, 95, and 100 mW/m^2 conductive geotherms are plotted in the top right. Lines show per-sample maximum emplacement depths calculated using the expanded stability of high-chromian spinel [*O’Neill*, [1981](#ref-ONeill1981)] with error bars of 0.15 GPa. This data suggests that the samples were sourced from ~50--80 km depth.

**Figure 26:** Source of Crystal Knob xenoliths

**Figure 27:** Trace element abundances (ppm)

**Figure 28:** Enrichment trends for trace elements.

**Figure 29:** Clinopyroxene trace elements for Crystal Knob compared to abyssal peridotite data compiled by *Warren* [[2016](#ref-Warren2016)]. The Crystal Knob samples show mild to moderate depletion in HREEs but samples CK-3, CK-4, and CK-6 show re-enrichment of LREEs.

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