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The origin and evolution of the deep lithosphere beneath continental arcs: constraints from lower crustal and mantle xenoliths from the Sierra Nevada, California

by

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ABSTRACT

[The origin and evolution of the deep lithosphere beneath continental arcs: constraints from lower crustal and mantle xenoliths from the Sierra Nevada, California]

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Lower crustal and upper mantle xenoliths from the Sierra Nevada continental arc in California, USA reveal a complex P-T-X-t (pressure-temperature-composition-time) evolution of the deep lithosphere. Metasedimentary xenoliths, despite high quartz contents, preserve granulite-facies equilibration conditions, necessitating transport from the surface into the lower crust. U-Pb and Hf isotope data on detrital zircon in the metasediments support continental underthrusting coeval with the peak of arc magmatism. Mantle xenoliths, represented by spinel and garnet-bearing spinel peridotites, indicate that shallow (spinel-facies), high-degree melt residues underwent thickening, cooling, and metamorphism into refertilized garnet peridotites that equilibrated below 800 °C and ca. 3 GPa. The source of refertilizing melts appears to be the mantle wedge itself, rather than the subducting slab based on mass-balance models and oxygen isotopes. Textural disequilibria, particularly Al-depletion haloes between orthopyroxene and exsolved garnet in peridotites, are diagnostic features related to cooling and increasing pressure. Diffusion modeling indicate that such disequilibria formed within 6 to 8 Ma. Coupled Lu-Hf and Sm-Nd age data on garnet pyroxenites show a significant lag between closure of the two chronometers in a lower crustal pyroxenite, but negligible difference in the deepest websterites. As a whole, the Sierran
xenolith data provide insights into the origin and evolution of a mature continental arc, particularly highlighting the importance of thickening in both the crust and mantle lithosphere.
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Chapter 1

Introduction

Owing to its inaccessibility, the deep lithosphere beneath volcanic arcs remains one of the great mysteries in our understanding of subduction zone magmatism. The only direct probes available to us to study the deep lithosphere are 1) exposed arc sections and 2) lower crustal and upper mantle xenoliths. Exposed sections (e.g., Talkeetna in Alaska, Kohistan in Pakistan) typically do not extend far below the Moho, and thus provide limited sampling of the true arc mantle lithosphere. Mantle-derived xenoliths, on the other hand, are foreign rock fragments erupted within a basaltic magma. The eruption timescale is rapid enough to effectively “freeze in” the pressure – temperature conditions recorded by mineral compositions in the xenoliths, thus allowing us to unravel the stratigraphy of the deep lithosphere. Often, host magmas are alkali basalts, which may come from great depths (at least, equal to or deeper than the vertical level of the xenoliths themselves).
The subject of this dissertation is the origin and evolution of the deep lithosphere beneath the Sierra Nevada Batholith in California, USA. The Sierra Nevada is one of the type localities of a continental arc, and has been intensively studied for decades. Here, I present data on a suite of lower crustal and upper mantle xenoliths. These xenoliths range from mid- to lower crustal metasediments, cumulate pyroxenites, to residual mantle peridotites. Remarkably, the entire xenolith suite studied here was sourced from one basaltic pipe (Big Creek) in the central Sierra Nevada – thus offering us a unique view of the entire sub-arc lithosphere from the crust to the mantle.

The second chapter of this dissertation is a petrological investigation of the pressure-temperature-composition history of the Sierran peridotites. I show that two disparate groups of peridotites, spinel peridotites and garnet-bearing spinel peridotites, had a common origin as spinel-facies, high-degree melt residues which where then subsequently thickened, cooled, and metamorphosed into garnet-bearing spinel peridotites, achieving final equilibration at pressures of 3 – 3.5 GPa (90 – 105 km depths) and temperatures below 800 °C. Furthermore, textural and whole-rock geochemical data hint at a refertilization event that apparently affected the deeper garnet-bearing peridotites (refertilization is investigated in detail in Chapter 4). The results of this chapter paved the way for one of the emergent themes of my dissertation: the importance of thickening in the formation, evolution, and stabilization/de-stabilization of the deep lithosphere. This chapter has been published in Journal of Petrology (Chin, Emily J., et al. "Deep lithospheric thickening and refertilization beneath continental arcs: case study of the P, T and compositional evolution of peridotite xenoliths from the Sierra Nevada, California." Journal of Petrology 53.3 (2012): 477-511.)
In the third chapter, I take a detour from the mantle and visit the lower crust. Here I present evidence that ancient North American passive margin sediments, sourced from the interior of the continent, were underthrust westward beneath the Sierran arc during peak arc magmatism in the Cretaceous. I analyzed a suite of unusual garnet-bearing metaquartzite xenoliths. These rocks are quartz-rich (>50 % by mode) and contain abundant detrital zircons, pointing to mature sedimentary protoliths. However, garnet neoblasts and strained quartz grains indicate metamorphism and deformation, and thermobarometry indicates final equilibration at lower crustal conditions. I measured the U-Pb and Hf isotopic compositions of detrital zircons hosted in quartz and found that all zircons lie on discordia with variable and ancient upper intercepts (>1 Ga), but a common, ~100 Ma lower intercept coeval with peak magmatic activity. Enhanced magmatism would elevate crustal temperatures, potentially facilitating the underthrusting of cold, continental upper plate sediments by thermal weakening of the lower crust. Underthrusting is thus another example of lithospheric thickening in active arcs. It has important implications, one of which is that the presence of felsic material at depth will influence the composition and physical properties of the lower crust. This chapter has been published in Earth and Planetary Science Letters (Chin, Emily J., et al. "On the origin of hot metasedimentary quartzites in the lower crust of continental arcs." Earth and Planetary Science Letters 361 (2013): 120-133.)

Chapter 4 returns to the mantle lithosphere, investigating the origin and nature of mantle metasomatism beneath the Sierra Nevada, and ultimately explores the compositional consequences of lithospheric thickening on magmatic evolution in arcs. In this chapter, I follow up on the observation made in Chapter 1 that the peridotites define a
a depth-gradation in fertility, manifested as shallow, clinopyroxene-poor, depleted spinel peridotites that grade into deep, clinopyroxene-rich, fertile garnet peridotites. One of the motivating questions, and one that has gripped the volcanic arc community for decades, is the role of slab-derived fluids and melts in mantle wedge metasomatism. Oxygen isotope data on deep lithospheric xenoliths could be one test – elevated oxygen isotopic signatures would be predicted if significant slab material infiltrated the mantle. Oxygen isotope values of mineral separates and reconstructed bulk values of Sierran peridotites, however, reflect typical mantle values, suggesting that the melt source was not the slab. Using simple mass balance and melt-rock reaction models, I suggest that the Sierran refertilization trend can be explained by up to 30% addition of a mantle-derived melt into previously depleted harzburgite. This chapter is currently in revision for Earth and Planetary Science Letters.

One of the most important contributions of Chapter 4 is the recognition that magma-wallrock interaction can extend far below the Moho in arcs, to well within the mantle lithosphere. The end result is refertilization and clinopyroxene-enrichment, which has important implications for magmatic differentiation in subduction zones. If so-called “primary” arc magmas fractionate clinopyroxene deep within the mantle lithosphere, such melts will already be evolved before they even reach the crust, challenging the paradigm of the crust as the dominant magmatic refinery. Refertilization may be facilitated beneath mature, thick arcs (such as the Sierra Nevada), where the thick lithosphere behaves as a magmatic trap and reactive filter. This chapter ends with a comparative summary of literature data that suggests that mantle beneath thick arcs does appear more fertile compared to juvenile arcs.
Finally, in Chapter 5, I investigate the thermal history of the thick Sierran arc root using a two-pronged approach. The first approach involves diffusion modeling of Al between orthopyroxene and exsolved garnet. One of the most interesting features of the Sierran garnet peridotites is the presence of exsolved garnet lamellae hosted within coarse-grained, deformed orthopyroxene porphyroclasts, which results in an Al-depletion halo at the orthopyroxene-garnet contact. This particular microtexture is diagnostic of either cooling, increasing pressure, or a combination of both, and has been documented in a handful of peridotites elsewhere. Modeling of Al compositional gradients indicates that formation of the observed Al-depletion haloes occurred within 6 to 8 Ma.

In addition to the Al-diffusion modeling, I further constrained thermal evolution of the Sierran arc lithosphere using coupled Lu-Hf and Sm-Nd geochronological data on garnet pyroxenites. I dated two high-Mg garnet websterites and one low-Mg garnet clinopyroxenite; the former coexist with garnet peridotites at ~3 GPa and the latter have a crustal origin. Overall, mineral isochrons yield ages overlapping the Late Cretaceous period of arc magmatism, supporting the genetic link between cumulate pyroxenites and granitoids. Intriguingly, a significant lag of 20 Ma between the Lu-Hf age and the Sm-Nd age is observed in the crustal pyroxenite, whereas the Lu-Hf and Sm-Nd ages for the mantle websterite record the same age. To explain these age systematics, I propose that downward thickening of the arc root until it impinged on the slab at ~3 GPa induced rapid cooling of the deep lithosphere. Cumulates formed beneath the thickened root cooled quickly enough for both Lu-Hf and Sm-Nd geochronometers to close at the same time. In contrast, shallow cumulates in the crust, far from the slab, cooled slowly and thus record a significant lag between Lu-Hf and Sm-Nd closure.
The Al-diffusion results and coupled geochronology suggest that cooling of the deep lithosphere was relatively fast – less than 10 Ma, while the shallow lithosphere cooled much more slowly. Thermal modeling is consistent with this hypothesis. These results offer an alternative explanation for why arc magmatism abruptly ceased in the Sierra Nevada. Instead of the Farallon slab shallowing (the classical view), thickening of the Sierran arc root and the geometric limit imposed by a “normal” dip slab at ~110 km (the global average depth to the slab) would pinch out the asthenospheric wedge, cutting off decompression melting and thus leading to the arc’s demise.

My work on the Sierran xenoliths demonstrates that even after the first Big Creek xenoliths were discovered by Lockwood and Bateman while mapping for the USGS in 1976, and despite the numerous Sierran xenolith studies that ensued, much can still be learned from these unique rocks. More importantly, big-picture implications have emerged from such small-scale investigations. One of the central ideas that emerges from this dissertation is that magmatic and tectonic thickening are essential arc processes that operate from crustal to mantle depths. Continental underthrusting has implications for composition of the continental crust, as does refertilization of the deep mantle lithosphere and its impact on derivative melts. Thickening and cooling of deep lithospheric roots might be inhibited by the subducting slab, not the other way around. In essence, unraveling the complexities of the P-T-X-t (pressure-temperature-composition-time) histories of deep lithospheric xenoliths has shown us the importance of upper-plate processes in modulating arc magmatism.
Deep lithospheric thickening and refertilization beneath continental arcs: case study of the P, T, and compositional evolution of peridotite xenoliths from the Sierra Nevada, California

1This chapter has been edited, reformatted, and reprinted from the Journal of Petrology (full citation below):


Thickening of arc lithosphere influences the extent of magmatic differentiation and is thereby important for the evolution of juvenile arcs into mature continental crust. Here, we use mantle xenoliths from the late Mesozoic Sierra Nevada continental arc in California (USA) to constrain the pressure, temperature, and compositional evolution of
the deep lithosphere beneath a mature arc. These xenoliths consist of spinel peridotites and garnet-bearing spinel peridotites. The former are characterized by coarse-grained protogranular textures having bulk compositions indicative of high-degree melting. The latter are characterized by porphyroclastic textures, garnet coronas around spinels, garnet exsolution lamellae in pyroxenes, and pyroxenes with high-Al cores and low-Al rims. The garnet-bearing spinel peridotites range from depleted to fertile compositions, but the high Cr# (molar Cr/(Cr+Al)) of spinel cores reflect high-degree melting. These observations suggest that the protoliths of the garnet-bearing spinel peridotites were melt-depleted spinel peridotites. Constraints from geothermobarometry and bulk compositions coupled with mantle melting models suggest that the protoliths underwent shallow melt depletion (1–2 GPa, 1300–1400 °C), followed by compression, cooling, and final equilibration within the garnet stability field (~3 GPa, < 800°C). The deepest equilibrated samples are the most refertilized, suggesting that refertilization occurred during compression. We interpret this P-T-composition path to reflect progressive thickening of the Sierran arc lithosphere perhaps due to magmatic inflation or tectonic thickening. We hypothesize that newly formed arc lithospheric mantle thickens enough to pinch out the asthenospheric wedge, juxtaposing Sierran arc lithosphere against the subducting oceanic plate. This could have terminated arc magmatism and initiated cooling of the deep Sierran lithosphere.

2.1. Introduction

Island and continental arcs are the dominant building blocks of Phanerozoic continents. It is in such environments where the felsic composition of average continental crust is achieved. Melting of the asthenospheric mantle wedge beneath arc volcanoes by decompression, volatile fluxing, or a combination of both generates primary arc magmas, leaving behind a depleted mantle residuum that may underplate overriding lithosphere, and may eventually become incorporated into newly formed lithosphere. Mantle-derived arc magmas rise into the overriding lithosphere, where they cool, crystallize and differentiate into felsic liquids and complementary mafic cumulates and
restites. The denser mafic lithologies are believed to be convectively removed and returned into the mantle, while the felsic residual liquids crystallize to form the continental crust (Kay & Kay, 1988; Kelemen, 1995; Rudnick, 1995; Ducea & Saleeby, 1996; Rudnick & Gao, 2003; Tatsumi, 2005; Lee et al., 2006; Jagoutz et al., 2009). Extensive magmatic differentiation should be favored in continental arcs or mature island arcs because of the thicker crust and lithospheric mantle through which juvenile basalts must traverse and interact. While it is generally thought that differentiation occurs primarily in the crust, as evidenced by the presence of abundant restites and cumulates in the lower crust (Arndt & Goldstein, 1989; Annen et al., 2006; Lee et al., 2006; Jagoutz et al., 2009), even deeper (e.g., sub-Moho) differentiation might occur in arcs with thicker lithospheres. The depth at which such differentiation occurs will influence the composition of erupted lavas. For example, deep lithosphere fractionation would modify the composition of magmas well before they traverse the crust-mantle boundary, hence failure to account for this would make mass balance models of arc crust inaccurate. In addition, deep lithosphere fractionation would also modify the composition of the lithospheric mantle by the addition of deep-seated cumulates or the products of melt-rock reaction (Kelemen, 1990). It is generally thought that continental lithospheric mantle is composed of melt-depleted peridotites (Boyd & Mertzman, 1987; Walter, 1998; Carlson et al., 2005), but extensive differentiation in the deep lithosphere could result in refertilization of such lithosphere.

The purpose of this study is to investigate whether arc lithosphere thickens and if so, constrain the extent and mechanisms of thickening. Here, we investigate the pressure (P) and temperature (T) evolution of the lithospheric mantle beneath the Sierra Nevada Batholith, a remnant of a Cretaceous ocean-continent subduction zone in western North America (Figure 2-1). Although many studies have been conducted on the deep continental crust in arcs (DeBari & Sleep, 1991; Ducea & Saleeby, 1996; Kelemen & Hanghoj, 2003; Lee et al., 2006; Jagoutz et al., 2009), most of these studies do not focus on P-T evolution and none have investigated the lithospheric mantle part of the arc in detail. However, the P-T evolution of the deep arc lithosphere may provide new constraints on the nature of arc thickening. For example, during accretion, compression
or underthrusting in continental arcs (Figure 2-2a) (Coney & Harms, 1984; DeCelles, 2004; Dickinson, 2004; Ducea & Barton, 2007; DeCelles et al., 2009), the deep lithosphere might be expected to increase in P and T. Interactions with the subducting oceanic lithosphere could also lead to thickening (increase in P) via refrigeration or freezing (decrease in T) of the asthenospheric mantle wedge (Saltzer et al., 2001) (Figure 2-2b), though it has also been proposed that too much coupling, such as during low-angle subduction, could erode the lithospheric mantle altogether (cf. Bird (1988)). Thickening could also occur by underplating of magmas or residual peridotitic mantle to the base of the over-riding lithosphere (Figure 2-2c) (Jordan, 1981; McDonough, 1990). In this case, the newly underplated material would bear, in theory, a record of initial decompression followed by isobaric cooling. Finally, another way to thicken is by magmatic inflation, wherein progressive intrusion of magmas into the crust displaces underlying parts of the lithosphere to greater depths and pressures, provided erosion at the Earth’s surface does not compensate for magmatic addition (Figure 2-2d) (Annen et al., 2006). In such a scenario, rapid magmatic inflation might be expected to result in near-adiabatic compression of the deeper portions of the lithosphere (Fig 2-2d) while slow inflation might permit lithospheric geotherms to re-equilibrate such that individual parcels of lithosphere would gradually heat up as they become progressively depressed into the hot, underlying asthenosphere.

To access the deep lithosphere beneath the Sierra Nevada arc, we use peridotite xenoliths entrained in late Miocene alkali basalts. We combine new major and trace element geochemistry of minerals and whole-rocks in order to constrain the P-T and compositional evolution of the Sierran continental arc lithospheric mantle. We constrain the initial “leg” of the P-T path using relationships between melting degree (F) recorded by the Sierran peridotites and pressure, which is a proxy for paleo-depth. The final “leg” is constrained using geothermobarometry, which gives the final equilibration pressures and temperatures of the peridotites. These constraints on P-T and compositional evolution add a new perspective to an existing body of work focused on the petrogenesis of Sierran lower crustal and upper mantle xenoliths (Dodge et al., 1988; Mukhopadhyay
& Manton, 1994; Ducea & Saleeby, 1996; Lee et al., 2001; Ducea, 2002; Lee et al.,
2006) and are used to formulate models for the evolution of continental arcs.

2.2. Geological setting

Figure 2-1. Simplified geologic map of the Western USA. Mesozoic Cordilleran batholiths are
outlined. Xenolith locality Big Creek in the central Sierra Nevada is represented by the star.
2.2.1. Overall geologic history

The Sierra Nevada Batholith is a Mesozoic magmatic arc formed by subduction of the Farallon plate beneath North America (Figure 1). Magmatism occurred between 220 and 80 Ma with its most productive stage occurring between ~120-80 Ma, peaking at ~93 Ma (Chen & Moore, 1982; Barton, 1996; Coleman & Glazner, 1997). The Mid- to Late Cretaceous magmatism was accompanied by thin-skinned compressional deformation that extended from the Sierra Nevada east into the present-day Rocky Mountains. This period of deformation, known as the Sevier orogeny (Armstrong, 1968), was characterized by foreland fold-and-thrust belts and possible retro-arc thrusting of North American lithosphere beneath the active Sierran arc (DeCelles, 2004; DeCelles et al., 2009). By ~74 Ma, major arc-building magmatism terminated, and much of western North America entered a period of compressional deformation in the form of basement-cored uplifts known as the Laramide orogeny (Dickinson & Snyder, 1978). The onset of the Laramide orogeny is generally attributed to the shallowing of the downgoing Farallon plate (Coney & Reynolds, 1977; Dickinson & Snyder, 1978; Bird, 1988) or to collision with an oceanic plateau (Saleeby, 2003). During the Laramide, magmatism appears to have migrated eastward into the continental interior (Coney & Reynolds, 1977; Lipman, 1992) and a cooler thermal regime was imposed on the overriding North American lithosphere, possibly accompanied by extensive hydration and metasomatism (Dumitru et al., 1991; Humphreys et al., 2003; Lee, 2005). At ~40 Ma, much of the Cordilleran orogenic belt began to collapse and magmatism in the form of rhyolitic/ignimbritic flare-ups migrated back towards the west, perhaps due to the foundering or roll-back of the Farallon plate (Coney & Reynolds, 1977; Humphreys, 1995). Collision of the Pacific-Farallon ridge with the Farallon-North American trench at ~20 Ma resulted in the birth of the San Andreas transform and the opening of a “slabless window” beneath western North America (Atwater, 1970). Continued collapse of the Cordilleran orogen culminated in extension of the Basin and Range Province, which continues today (Wernicke, 1981). Basin and Range extension appears to have been accompanied by widespread Late Miocene to present basaltic volcanism, including small-volume alkalic to ultrapotassic
basalts of Late Miocene to Pleistocene age in the central and eastern Sierra Nevada (Van Kooten, 1980; Dodge et al., 1988). These Sierran volcanic centers contain xenoliths of the deep Sierran crust and lithospheric mantle. The xenoliths investigated here come from the Late Miocene Big Creek diatreme (37°13’N, 119°16’W) erupted ~8 Ma and now exposed as a small (110x260 m), elliptical trachyandesitic pipe located in the central part of the batholith (Dodge et al., 1988; Ducea & Saleeby, 1998b).

2.2.2. The Sierra Nevada Batholith

In the Sierra Nevada batholith, Kistler and Peterman (1973) and Kistler (1990) noted an east-west transition in lithosphere composition, manifested by a transition from unradiogenic to radiogenic initial Sr isotopic signatures (the $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.706$ boundary). Mafic granitoids dominate the batholith west of the 0.706 line; these were emplaced into recently accreted oceanic terranes characterized by unradiogenic Sr (low $^{87}\text{Sr}/^{86}\text{Sr}_i$). Felsic, granitic plutons dominate in the east and appear to have interacted extensively with pre-existing, Precambrian-derived sediments and/or basement (Saleeby et al., 1987), responsible for their radiogenic Sr (high $^{87}\text{Sr}/^{86}\text{Sr}_i$). In particular, the major and trace-element signatures of the plutons in the eastern Sierran and Peninsular Ranges batholiths (the latter represents the southern extension of the Cordilleran magmatic arc) require the presence of garnet-bearing mafic cumulates and restites at depth (Gromet & Silver, 1987). Garnet pyroxenite xenoliths from the lower crust and lithospheric mantle beneath the Sierra Nevada have compositions that match the expected mafic complements to the plutons (Mukhopadhyay & Manton, 1994; Ducea, 2001; Ducea, 2002; Lee et al., 2006; Lee et al., 2007b). Lee et al. (2006) subdivided the garnet pyroxenites into two groups, a high-MgO group representing high-pressure (1-3 GPa) cumulates of a primary, hydrous basaltic magma and a low-MgO group, representing cumulates or restites of evolved basalts at lower crustal pressures (~1.5 GPa). This petrogenetic relationship implies that thick crust/lithosphere was needed to generate such deep mafic cumulates and restites because thick lithosphere would force mantle-derived magmas to differentiate at greater depths, facilitating the crystallization of garnet-rich, high-MgO pyroxenites (Sisson et al., 1996; Lee et al., 2006). This garnet pyroxenite root is suggested to have been removed
sometime after 10 Ma and before 0.18 Ma (Ducea & Saleeby, 1996; Ducea & Saleeby, 1998b; Farmer et al., 2002; Zandt et al., 2004).

Figure 2-2. Cartoons illustrating mechanisms responsible for thickening deep continental arc lithosphere. Hypothetical P-T paths tracking the lithostatic P and T of a parcel of lithosphere (L) during thickening are shown to the right (L’ represents L after thickening has occurred). See text for discussion. (a) Arc accretion and continental underthrusting, (b) shallow-angle subduction and viscous coupling of asthenospheric mantle wedge to downgoing slab, (c) lithospheric underplating by magmas or melt-depleted peridotite, (d) magmatic inflation.
2.3. Summary of previous work

Xenoliths from the Big Creek diatreme can be broadly categorized into two groups: a mafic group consisting of pyroxenites and granulites from the lower crust and possibly the lithospheric mantle, and an ultramafic group characterized by peridotites associated with the lithospheric mantle (Dodge et al., 1988; Mukhopadhyay & Manton, 1994; Ducea & Saleeby, 1996). The mafic lithologies, as mentioned above, have been interpreted to represent the restites and cumulates complementary to the Sierran granitoids and have been discussed in detail elsewhere (Lee et al., 2006). The peridotites are thought to represent the deepest samples of the sub-continental arc lithosphere, with some recording equilibration pressures up to 3 GPa (Ducea & Saleeby, 1998a; Lee et al., 2001). They have undergone various degrees of partial melting, as evidenced by compositions ranging from depleted harzburgites to fertile lherzolites (Mukhopadhyay & Manton, 1994; Lee et al., 2001). In the garnet-bearing spinel peridotites, Lee et al. (2001) identified cooling features, such as garnet rimming spinel and garnet exsolution lamellae in pyroxenes. Based on Ca-in-opx geothermometry on garnet-free regions of orthopyroxene, Lee et al. showed that initial temperatures of the garnet peridotites were at least 1140 °C, whereas final (“closure”) temperatures based on orthopyroxene compositions adjacent to garnet lamellae were less than 800 °C, thus indicating substantial cooling. However, the possibility of pressure change (ΔP) accompanying cooling was not evaluated. Based on the peridotites’ thermal history, isotopic evidence, and comparisons to xenoliths from Quaternary basalts in the Basin and Range, these authors considered the Big Creek peridotites to represent recently upwelled asthenospheric mantle that replaced old, lithospheric mantle during a lithospheric removal or rejuvenation event, possibly related to the Cretaceous arc magmatism itself. We build on these pre-existing studies by presenting detailed petrographic, geothermobarometric and geochemical studies of Sierran peridotite xenoliths, with particular emphasis on garnet-bearing spinel peridotites.
2.4. Methods

Whole rock major element and trace element contents were determined by X-ray fluorescence spectroscopy (XRF at Washington State University at Pullman) and by inductively coupled plasma mass spectrometry (ThermoFinnigan Element 2 ICP-MS at Rice University). ICP-MS sample preparation is as follows. Due to limited sample size, ≤5 cm chunks of whole rocks were crushed and powdered in a ceramic SPEX mill placed in a shatterbox for 5 to 10 minutes per sample. Care was taken to pick only the freshest, interior portions of the rocks for milling. 50 to 80 mg of sample powder was added to a weighed, empty, and dry 2.5 mL Savillex wrench-top beaker. Two sequential acid attacks using a 1:1 mixture of concentrated HF:HClO₄ were performed, with complete open-air dry-down at ~190 °C between each attack. After the acid attacks, ~1–2 mL of 2% HNO₃ was added to each beaker. Then each beaker was diluted up to a total volume of 125 mL by addition of 2% HNO₃. 0.025 mL of HCl was added to each beaker to keep Fe in solution. Independently prepared standards (BHVO-1, BIR-1, AGV-1, DTS-1, JP-1) were used as external standards using values from Eggins et al. (1997). For an internal standard, all solutions were spiked with Indium to achieve a 1 ppb concentration of In in solution. Procedural blanks were run in duplicate. Prior to running on the ICP-MS, all samples were ultrasonicated for 30 minutes and allowed to sit overnight. Na, Mg, Al, P, K, Ca, Sc, Ti, V, Cr, Mn, Fe, Co, Ni, Cu, Zn, Nb, and Zr were analyzed in medium mass resolution mode (“MR-ICPMS”, m/Δm = 3000). Li, Be, Rb, Y, Cs, Ba, Pb, Th, Ta, Tl, and the REE were analyzed in low mass resolution mode (“LR-ICPMS”, m/Δm = 300).

Major element concentrations of minerals in sample 1026V were acquired using wavelength dispersive spectroscopy on the JEOL JXA-8530F “Hyperprobe” at Yale University. The operating conditions were 15 kV accelerating voltage, probe current of 5.042 nA, and 40 ms dwell time. X-ray maps of orthopyroxene and clinopyroxene in 1026V were also acquired in wavelength dispersive (WDS) mode on the Hyperprobe. X-ray maps were processed offline using ImageJ software (http://rsbweb.nih.gov/ij/). Major element concentrations of minerals in samples 08BC03, 08BC04, 08BC08, and BC-77
were acquired in WDS mode on the Cameca SX 50 at Texas A&M University. On both electron microprobes, typical spot sizes were 1 to 1.5 μm. Olivine, orthopyroxene, clinopyroxene, garnet, and chromite standards were used.

Energy-dispersive X-ray maps for Cr, Al, and Ca were generated using a Horiba XGT-7000 micro-XRF at Texas A&M University. Operating conditions were 50 kV accelerating voltage and 1 mA probe current. The mapping image threshold was 10 μm; we used a spot size of 100 μm.

In situ trace element data of minerals were obtained using laser ablation ICP-MS (ThermoFinnigan Element 2 equipped with New Wave 213 nm laser ablation system) at Rice University. Polished 30 μm-thick sections of the peridotites 1026V and BC-77, and polished 300 μm-thick sections of the peridotites 08BC03, 08BC04, 08BC08, and 08BC14 were analyzed. Analyses were performed in low mass resolution modes. The laser beam was set to 55 μm spot size, 10 Hz pulse frequency, and 20 J/cm² energy fluence. ²⁵Mg was used as an internal standard to correct for instrument drift. BHVO2g, BIR1g, BCR2g, and NIST612 basaltic glass standards were used as external standards using standard values from Gao et al. (2002). Prior to analysis, the instrument was tuned by manually controlling sample gas (Ar) to obtain a sensitivity of >300,000 cps per 15 ppm La in BHVO2g for a 55 μm diameter spot and a pulse frequency of 10 Hz. Time resolved analyses were converted to concentrations using in-house Excel macros (http://www.ruf.rice.edu/~ctlee/Laser-RAWDATA-TEMPLATE.xls).

2.5. Results

2.5.1. General xenolith petrography and textural history

This study focuses on detailed petrography and geochemistry of five garnet-bearing spinel peridotites out of a suite of 10 garnet-bearing samples: 1026V, BC-77, 08BC03, 08BC04, and 08BC08 (Figures 2-3, 2-4). Garnet-bearing peridotites provide the most reliable quantitative determination of P-T paths because garnet-pyroxene
geothermobarometry can be applied. However, for completeness, we also include less detailed data on spinel peridotites. Although garnet comprises ≥8% by mode in the garnet-bearing spinel peridotites, we refer to them as “garnet-bearing” because, as we show below, the entire Sierran peridotite suite represents a continuum in terms of textures and geochemistry, the spinel peridotites being protoliths of the garnet-bearing spinel peridotites.

In order of decreasing abundance, the Sierran peridotites are composed of olivine, orthopyroxene, clinopyroxene, and one or two aluminous phases (spinel and/or garnet). Mineral modes for 3 spinel peridotites (08BC01, 08BC13, and P7) and 5 garnet-bearing spinel peridotites (BC77, 1026V, 08BC03, 08BC04, 08BC08) were calculated using least squares inversion of the whole-rock and average mineral major element compositions (see Table 2-1). The sum of the residuals are all ≤ 0.6 (Table 2-1), indicating that our sample sizes, while small, are still sufficiently large to average out grain-scale heterogeneity in mineral modes. Spinel peridotites are generally coarse-grained (average grain size 4 mm or greater) and have a protogranular texture (cf. Mercier & Nicolas (1975)) (Figure 2-3). Olivine and orthopyroxene are the dominant minerals in the spinel peridotites, with subordinate clinopyroxene and accessory spinel. Curvilinear grain boundaries are often present between olivine, orthopyroxene, and clinopyroxene (where present) grains (Figure 2-3). Spinels in the coarsest samples tend to be large (1 – 2 mm) and euhedral, whereas spinels in more recrystallized samples show increasing degrees of elongation. Some olivine grains in the spinel peridotites show kink-banding. No clear foliation or shape-preferred orientation of minerals is present in the spinel peridotites.
Representative thin sections, µ-XRF maps, and photomicrographs of garnet-bearing spinel peridotites are presented in Figure 2-4. In contrast to spinel peridotites, the garnet-bearing samples are porphyroclastic, and have a weakly foliated fabric defined by shape-preferred orientation of recrystallized olivines (showing unequilibrated grain boundary triple junctions) and diffuse bands of fine-grained clinopyroxene and garnet disseminated within the recrystallized olivine matrix. These diffuse clinopyroxene bands are conspicuous in the µ-XRF maps (shown by green colors), where they often occur together with fine-grained, discrete garnets. Porphyroclasts are comprised of orthopyroxene and clinopyroxene, the former are typically highly deformed, and the latter are less deformed. Two samples, 1026V and BC77, contain exceptionally large (~1 cm), deformed and bent orthopyroxene porphyroclasts. Such porphyroclasts also show extensive kink-banding and bending of cleavage planes. We interpret these

### Table 2-1. Mineral modes and petrography of selected Sierran peridotites

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<th>Spinel Peridotites</th>
<th>Garnet-bearing Spinel Peridotites</th>
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<td><strong>Mode</strong></td>
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<td>Ol</td>
<td>76.65</td>
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<td>Opx</td>
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<td>Cpx</td>
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<tr>
<td>Sp or gt</td>
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<td><strong>Total</strong></td>
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<td>101.00</td>
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<tr>
<td><strong>Σ(r²)</strong></td>
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<tr>
<td><strong>Grain-size</strong></td>
<td>Medium</td>
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<td><strong>Texture</strong></td>
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<td><strong>Microtexture</strong></td>
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<td><strong>Garnet</strong></td>
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Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene, Sp = spinel, Gt = garnet

Σ(r²) = sum of residuals squared (modes were determined using least-squares regression of average mineral chemistries and whole-rock compositions)

Grain size defined as follows: Fine grained (<1 mm), Medium grained (1–4 mm), Coarse grained (>4 mm)

For porphyroclastic samples, grain size refers to that of the matrix.

Proto. = protogranular, Porphyro. = porphyroclastic, Trans. = transitional between protogranular and porphyroclastic

Other abbreviations: recryst. = recrystallized, curv. = curvilinear, gb = grain boundary, porphyroclast = porphyroclast, px = pyroxene

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Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene, Sp = spinel, Gt = garnet

Σ(r²) = sum of residuals squared (modes were determined using least-squares regression of average mineral chemistries and whole-rock compositions)

Grain size defined as follows: Fine grained (<1 mm), Medium grained (1–4 mm), Coarse grained (>4 mm)

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orthopyroxene porphyroclasts as relict grains from a coarse-grained harzburgitic protolith. A ubiquitous feature in the garnet-bearing samples is the presence of garnet coronas around spinels, the latter representing relict spinels from a spinel peridotite protolith.

Figure 2-3. Petrography of two representative Sierran spinel peridotites, 08BC13 and P7. Left column shows a plane-polarized light photograph of a thin section of each peridotite, and the right column shows a μ-XRF map of the same thin section (same orientation). Colors in the μ-XRF maps are as follows: green = Ca, magenta = Cr, dark purple = Al, intermediate purple = Fe. In general, spinel peridotites are coarse-grained, protogranular, and undeformed. Chromian spinels (large opaque grains on left; represented by bright magenta areas in μ-XRF maps at right) tend to be relatively large and euhedral, although P7 shows some deformed spinels.
Figure 2-4. Microtextures of Sierran peridotites. Petrography of five Sierran garnet-bearing spinel peridotite xenoliths: P6, 08BC03, 08BC08, BC-77, 1026V. Petrography of five Sierran garnet-bearing spinel peridotite xenoliths: P6, 08BC03, 08BC08, BC-77, 1026V. Each row corresponds to an individual sample (except for P6, see below). The left column shows a plane-light photograph of a 30 μm thin or 200 μm thick section of each sample, the middle column a μ-XRF map of the same thin or thick section in the same orientation as the plane-light photograph, and the right column...
microphotographs of the xenoliths. Scale bars shown at the bottom for 1026V are the same throughout each column. Colors in the μ-XRF maps are the same as in Figure 3. From top row to bottom row: P6 is characterized by a medium-grained, porphyroclastic texture, with coarser-grained clinopyroxene (compared to the other peridotites) within the weakly foliated bands. Plane light photograph is of a 200 µm thick section. Olivine and orthopyroxene are also coarse-grained. The microphotograph is actually of sample 08BC04, which is similar in texture to P6 (P6 was only available as a 200 µm thick section). The microphotograph depicts a relict spinel in the process of transforming into garnet; the light-colored, “undigested” half represents the original spinel with an incipient, thin garnet corona, whereas the dark-colored, “digested” other half represents increasing degrees of transformation into garnet and thus thicker garnet coronas. 08BC03 is characterized by a fine-grained, porphyroclastic texture. Plane light photograph is of a 30 µm thin section. Unique to this sample is the presence of large garnet megacrysts, ~5 mm in diameter. 08BC03 also shows transitional spinel-garnet textures, although these are manifested within individual spinel-garnet clusters, unlike the transitions observed in P6 and 08BC04. The microphotograph shows a large, euhedral zoned spinel mantled by thin garnet corona, and a smaller, anhedral spinel with thick garnet corona within a few mm of each other. 08BC08 is also fine-grained and porphyroclastic, with several large clinopyroxene porphyroclasts containing small garnet grains along the pyroxene grain boundaries. Plane-light photograph of 08BC08 is of a 200 µm thick section, hence the more vivid coloring of minerals. The microphotograph shows a well-developed garnet corona around spinel. Plane light photographs of BC-77 and 1026V are of 30 µm thin sections. Microphotograph (crossed nicols) of BC-77 shows a large, heavily deformed orthopyroxene porphyroclast with abundant garnet exsolution lamellae. Bright, thin lamellae are either clinopyroxene or amphibole. Microphotograph (crossed nicols) of 1026V shows part of the sample dominated by diffuse bands of loosely dispersed clinopyroxene, with discrete garnet along grain boundaries of clinopyroxene and within the matrix. Note the absence of deformation in these clinopyroxenes.

2.5.2. Garnet microtextures

Garnets in the Sierran peridotite suite mostly occur as coronas rimming spinels or as fine lamellae ranging in thickness from ~100 nm to 30 µm in clinopyroxene and orthopyroxene (Figure 2-5). Some garnets also occur as discrete grains within the olivine matrix and are unassociated with spinel. Occasionally, discrete garnets cluster along grain boundaries of pyroxenes (see sample 1026V, Figure 2-4) or form loosely disseminated bands that parallel the foliation imposed by diffuse pyroxene bands, such as in peridotites 08BC08 and BC-77 (Figure 2-4). Fine-grained opaque kelyphite rims commonly mantle the garnets, though in some cases, the garnets have broken down into fine-grained green, aluminous (nearly Cr-free) spinels. These spinels are visually and chemically distinct from the Cr-bearing amber to opaque primary spinels. The kelyphites and green spinels are garnet breakdown products associated with decompression and interaction with the erupting host magma. Other examples of late-stage overprints include thin amphibole lamellae (≤10 µm in width) in the cleavage plane of pyroxenes (Figure 2-5). These are most likely due to reaction of pyroxenes with late-stage fluids that
infiltrated along cleavage planes. These late stage textures, while interesting, are not relevant to our study and thus not discussed any further.

Textural evidence suggests that the garnet lamellae exsolved from their host pyroxenes. This is shown by the presence of Al depletion haloes in host pyroxene at the contact with the garnet lamellae and, in particular, by a correlation between garnet lamellae thickness and Al depletion halo width (Figures 2-5 and 2-6). In Figure 2-5, false color WDS EPMA maps of two pyroxene grains containing garnet lamellae in sample 1026V are shown: pink to red colors indicate high Al (garnet), purple indicates intermediate Al (amphibole) and blue indicates low Al (pyroxene). Bright yellow areas are spinels that probably formed due to late-stage garnet breakdown. Two profiles (Lines 2 and 3) from the orthopyroxene in Figure 2-5a are plotted in Figure 2-6 (garnet and amphibole lamellae are denoted by red and gray vertical bars, respectively). It can be seen from Figure 2-6 that Al depletion haloes are widest next to thick garnet lamellae, and thinner next to thin garnet lamellae. In some cases, thin garnet lamellae appear to have nucleated within the Al-depletion halos of large garnet lamellae, resulting in shorter wavelength haloes and garnet lamellae superimposed on the longer wavelength halos associated with the formation of thick garnet lamellae. Al depletion haloes are virtually nonexistent next to amphibole lamellae. The thicknesses of the Al-depletion haloes are limited by diffusion of Al from the pyroxene and into the garnet (or amphibole), and thus must relate to the timing of garnet (or amphibole) nucleation and subsequent cooling rate. Thick Al-depletion haloes imply early nucleation whereas thin halos imply later nucleation. We measured the thicknesses of the Al-depletion haloes and the adjacent garnet lamellae, which shows a positive correlation between the two quantities (Figure 2-5). This positive correlation strongly suggests that garnet lamellae nucleation and growth is fed by Al diffusing out of the pyroxene. If the garnets were not exsolved, but of an external (metasomatic) origin, one should not expect any relationship between garnet lamellae thickness and accompanying Al depletion in the host pyroxene. Thus, these observations strongly suggest that garnet lamellae formed by exsolution from originally high Al orthopyroxenes. We note that the lack of Al-depletion halos by the amphibole lamellae indicates that the amphibole is extraneous to the pyroxene or was so recently
nucleated that no time has been allowed for significant Al depletion in the pyroxene, and for these reasons, we ignore the amphibole lamellae from here on.

Figure 2-5. WDS (wavelength dispersive EPMA) false-color maps of Al. Maps are of orthopyroxene (a) and clinopyroxene (b) in sample 1026V. Inset shows index map of 1026V with the grains that were analyzed in a) and b). Warm colors (pink, red) represent high Al (i.e., garnet), cool colors (purple, blue) represent low Al (i.e., amphibole, pyroxene). Quantitative WDS points along transect A – A’ are plotted in the profile below the opx map. Each point on the transect represents a 1 μm spot analysis. Am = amphibole; gt = garnet.
Figure 2-6. **ImageJ analyses of transects in Figure 2-5.** Top: profiles of Al relative intensity vs. distance for two transects shown in Figure 5a. Shaded rectangles represent garnet (gt) and amphibole (am) lamellae. Bottom: Plot of Al depletion halo thickness vs. half-thickness of garnet lamellae.

### 2.5.3. Spinel microtextures

Cr-spinels within the garnet-bearing spinel peridotites occur as discrete grains, but most are rimmed by garnet coronas. In those mantled by garnet coronas, the spinels are visibly zoned from deep amber cores to completely opaque rims, the latter characterized by higher Cr content. Zonation is most pronounced in large spinels mantled by thin garnet coronas (see sample 08BC03, Figure 2-4). Small spinels mantled by thick garnet coronas are uniformly opaque and show deeply embayed margins (Figure 2-4). These observations suggest that large spinels, with high spinel to garnet corona ratios, represent early snapshots of spinel being converted to garnet while the small spinels, with lower spinel to garnet corona ratios, represent snapshots in which spinel conversion to garnet has progressed considerably. We note that, within the same samples, some of the spinel-garnet coronas are elongated in shape while others are more equidimensional, the former suggesting that spinel to garnet transformation occurred after deformation and the latter suggesting that the transformation occurred during or after deformation. Sample 08BC03...
(Figure 2-4), for example, contains a spectrum of spinel-garnet textures ranging from large euhedral spinel with incipient garnet coronas to small anhedral spinels with well-developed garnet coronas.

2.5.4. Interpretation of microtextures

Based on the above textural observations, we arrive at the following interpretive sequence of events. The protoliths of the garnet-bearing spinel peridotites were likely coarse-grained protogranular spinel peridotites. These peridotites were subsequently deformed, leading to the formation of porphyroclastic textures characterized by fine-grained recrystallized olivines and large, deformed orthopyroxene porphyroclasts, the latter representing relics of the original undeformed spinel peridotite protolith. Deformation has been coincident with, or followed by the formation of garnet lamellae in pyroxenes and garnet coronas around spinels. Garnet formation in pyroxenes is most likely due to exsolution from high-Al pyroxenes (typical of spinel peridotites), as exemplified by the Al-depletion halos in the pyroxenes. The elongated shape of some spinels and spinel-garnet coronas suggest that metamorphism was concomitant with deformation. That some spinel-garnet coronas are circular also suggests that metamorphic transformation from spinel to garnet may have continued after deformation ceased. Diffuse bands of disseminated clinopyroxene and garnet grains suggest open system behavior owing to infiltration of refertilizing silicate melts (to be discussed later). These clinopyroxene-rich bands follow the shape-preferred orientation of recrystallized olivines, suggesting introduction during deformation.

2.5.5. Whole-rock geochemistry

Whole-rock major and trace element data for spinel peridotites and garnet-bearing spinel peridotites are reported in Table 2-2; oxide variation diagrams are shown in Figure 2-7. Whole-rock trace element data for samples P-7, P-10, BC98-2, 1026V, and BC-77 were previously reported in Lee (2005). Samples 1026V and BC-77 were previously investigated by Lee et al. (2001) and Lee (2005). Samples 08BC14 and
08BC07 were not extensively analyzed due to pervasive serpentinization, but we report whole-rock XRF and selected mineral data for these samples for completeness. Here, we present new whole-rock and in situ geochemical data allowing us to constrain a more detailed P-T path of these xenoliths.

In terms of whole-rock major elements, the Sierran xenoliths range from depleted spinel peridotites to fertile garnet-bearing spinel peridotites. Mg#’s (atomic Mg/(Mg+Fe)x100) range from 86 to 91. This range overlaps that of typical residual peridotites, but values lower than 89 indicate that some of the peridotites are more fertile than canonical primitive mantle ((McDonough & Sun, 1995; Walter, 1998) and references therein). CaO, Na2O, and Al2O3, measures of fertility, correlate negatively with Mg# and MgO, with the spinel peridotites characterized by generally high Mg# and low CaO, Na2O, and Al2O3 (depleted) and the garnet-bearing spinel peridotites characterized by high and variable CaO, Na2O and Al2O3 and low Mg#s (fertile). However, overlap in major element trends between spinel and garnet-bearing spinel peridotites indicates that the Sierran peridotites define a compositional continuum, with some garnet-bearing spinel peridotites preserving depleted signatures.
Figure 2-7. Whole rock major elements in Sierran peridotites. Whole-rock major element variation diagrams showing the relationship between spinel peridotites and garnet-bearing spinel peridotites. The former are typically depleted in terms of major elements; the latter are comparatively more fertile, although some garnet-bearing peridotites overlapping with spinel peridotites. Sample 08BC14 is not shown in (a) and (e) because of its anomalous CaO (11 wt.%). 08BC14 is highly serpentinized and altered, which is reflected by its low XRF total (see Table 2). (a) CaO vs. Al$_2$O$_3$ (b) FeO vs. Al$_2$O$_3$ (c) MgO vs. Al$_2$O$_3$ (d) CaO vs. MgO (e) Na$_2$O vs. MgO.

The most depleted garnet-bearing spinel peridotites, exemplified by BC-77 and 08BC07, have CaO and Al$_2$O$_3$ contents as low as 2.2 and 1.08 wt. %, respectively, overlapping CaO and Al$_2$O$_3$ contents of some spinel peridotites from the same location. However, the most fertile garnet-bearing peridotites (1026V, 08BC04) have >4 wt. % CaO and >4 wt. % Al$_2$O$_3$, higher than that of canonical primitive mantle (3.55 wt. % CaO and 4.45 wt. % Al$_2$O$_3$; (McDonough & Sun, 1995)). Although the sense of these compositional trends is broadly consistent with that expected for melt depletion, the trends do not all pass through primitive mantle values as would be expected if these trends were imparted solely by melt depletion of a homogeneous primitive mantle source. In the plot of CaO vs. Al$_2$O$_3$ (Figure 7a), many garnet-bearing spinel peridotites “fan out” in an array towards higher CaO and Al$_2$O$_3$, but in terms of MgO vs. Al$_2$O$_3$, these peridotites are just as depleted as spinel peridotites. High CaO is associated with those samples with abundant diffuse clinopyroxene bands. Finally, the relatively linear trend
shown by Na$_2$O vs. MgO suggests a mixing between depleted peridotite and a melt rather than simple melt-depletion. This is because Na$_2$O, which is highly incompatible, should show a non-linear melt depletion trend when plotted against highly compatible elements like MgO. Collectively, these textural, and major and minor-element trends suggest that these samples have been refertilized.

The lack of a simple melt-depletion history is also evident from the trace element systematics (Figure 2-8). Whole-rock Yb contents span a wide range within the suite of garnet-bearing spinel peridotites, with some samples (1026V, 08BC03, 08BC14) as fertile as primitive mantle (>0.4 ppm Yb), while other samples (BC-77, 08BC08) are highly depleted in Yb (<0.05 ppm Yb) and hence resemble the depleted spinel peridotites in terms of Yb. In contrast, in most samples, whole-rock Ti contents are depleted (<600 ppm) compared to primitive mantle (1205 ppm Ti), with the exception of samples P-1, P-6, and 08BC04, which have whole-rock Ti >600 ppm. Because Ti and Yb are both considered moderately incompatible elements, a positive correlation between Ti and Yb would be expected for a simple melt depletion trend. Instead, we observe a general decoupling between Ti and Yb (Figure 2-8a): most garnet-bearing spinel peridotites plot within a relatively restricted range of Ti, but show a wide range in Yb. The fact that Yb correlates with Al$_2$O$_3$ indicates that the same processes that imparted variations in major element composition have also imparted variations in Yb, but not in Ti.
Figure 2-8. Whole-rock trace-elements of Sierran peridotites. Whole-rock trace element and major element variations for spinel peridotites and garnet-bearing spinel peridotites. Solid black lines with tick marks represent hypothetical melting trends (see text for details) starting from primitive mantle (McDonough & Sun, 1995). Symbols as in Figure 6. (a) Ti vs. Yb (b) Ti vs. Al$_2$O$_3$ (c) Sc vs. Al$_2$O$_3$ (d) Yb vs. Al$_2$O$_3$.

Whole rock trace element spider diagrams are shown in Figure 2-9, with the elements listed in approximate order of increasing compatibility during partial melting of anhydrous peridotitic mantle (Hofmann, 1988). The overall trace element patterns (normalized to primitive mantle values of McDonough & Sun (1995)) indicate whole-rock enrichments in fluid-mobile elements (Cs, Ba, Rb, Sr, Li) and depletions in high field strength elements (Nb, Ta). If only REE are plotted (not shown), spinel peridotites have generally flat REE patterns with a slight increase in the light REE. In contrast, many garnet-bearing spinel peridotites (08BC03, 08BC08, 1026V) have gently concave-upwards REE patterns, with middle REEs depleted relative to light and heavy REE.
Table 2. Whole-rock geochemistry of spinel peridotites. Values have been normalized to primitive mantle of McDonough and Sun (1995).

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Figure 2-9. Whole-rock trace-element spidergrams of Sierran peridotites. Values have been normalized to primitive mantle of McDonough and Sun (1995).
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### Conclusion

The XRF and ID measurements provide valuable insights into the chemical composition of the garnet-bearing Spinel Paridotites, highlighting the abundance of elements such as Si, Mg, Ca, and Al. The NAA measurements further support these findings, offering a comprehensive view of the mineralogy.
2.5.6. Mineral chemistry

2.5.6.1. Orthopyroxene

Major element compositions of orthopyroxenes in spinel peridotites and garnet-bearing spinel peridotites are reported in Table 2-4. Trace element compositions of orthopyroxenes in garnet-bearing spinel peridotite 1026V are reported in Table 2-5. Orthopyroxenes in the Sierran xenoliths are enstatite-rich, with Mg# between 90 and 92. Rims of orthopyroxene generally have slightly higher Mg# than corresponding cores. In addition, Al2O3 in cores is several weight percent higher than in rims. As illustrated in

Table 2-2. Whole-rock major element and trace element compositions of Sierran peridotites

| Mineral | Sr   | Ca   | Ba   | Ti   | V    | Cr   | Ni   | Co   | Al   | Mg   | Fe   | Zn   | La   | Ce   | Nd   | Sm   | Eu   | Gd   | Tb   | Dy   | Tb   | Ho   | Er   | Th   | Yb   | Lu   | Hf   | Dy   |
|---------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|
|         | 16.8 | 33.2 | 16.5 | 70.8 | 45.4 | 15.4 | 21.7 | 362  | 335  | 138  | 100  | 87   | 51   | 19   | 30   | 124  | 372  | 122  | 63   | 7.63 | 2.44 | 4    | 9.5  | 5    | 112  | 42   | 67   | 0.069 |
| XRF    |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |      |
|         |      |      |      | 3.49 | 4.46 | 5.08 | 3.61 | 9.54 | 0.61 | 0.98 | 1.73 | 3.44 | 5.15 | 2.28 | 10   | 12   | 2    | 2    | 5    | 4    | 3    |      |      |      |      |      |

**Highly Mg# = (atomic Mg/(Mg+FeT)x100)**

**References:**
[1] Dodge et al., 1988
[2] Lee, 2005
**Figure 2-5a**, there is a strong decrease in Al$_2$O$_3$ from areas in orthopyroxene without exsolved garnet (“core” areas) towards the interface with exsolved garnet (“rim” areas); for example, Al$_2$O$_3$ is $\sim$4.6 wt% in the core and decreases to 0.7 wt% at garnet-orthopyroxene interface. Cr$_2$O$_3$ contents were found to be high in the cores and low on the rims (*Table 2-4*). Primitive-mantle normalized rare earth element concentrations in orthopyroxene from sample 1026V are shown in **Figure 2-10a**. Two measurements of orthopyroxene show increasing heavy rare earth concentrations but this may be due to contamination during analysis from a nearby garnet lamella.
Figure 2-10. Trace element spidergrams of minerals. REE patterns of (a) orthopyroxene in 1026V, (b) clinopyroxenes in several samples, and (c) garnet in several samples; normalized to primitive mantle.
TABLE 2-4. Orthopyroxene major element compositions

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Each column represents an individual spot analysis. All orthopyroxene analyses are next to garnet.

Table 2-4. Trace-element compositions of pyroxenes
TABLE 2-6. Clinopyroxene major element compositions

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<th>Spinel peridotites</th>
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<th>08BC06 Rim</th>
<th>08BC-13 Rim</th>
<th>P-7 Rim</th>
<th>P-10 Rim</th>
<th>BC98-2 Rim</th>
<th>08BC01 Core</th>
<th>08BC06 Core</th>
<th>08BC-13 Core</th>
<th>P-7 Core</th>
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<td>52.73</td>
<td>53.62</td>
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<th>08BC-03 Rim</th>
<th>08BC-04 Rim</th>
<th>08BC-08 Rim</th>
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<th>BC77 Core</th>
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<td>0.03</td>
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Table 2-5. Major element compositions of clinopyroxenes

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<td>93.57</td>
<td>94.37</td>
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Rims and cores reported below are of the same mineral grain; each column represents an individual spot analysis.

2.5.6.2. Clinopyroxene

Major element compositions of clinopyroxenes in both spinel peridotites and garnet-bearing spinel peridotites are presented in Table 2-6. Trace element geochemistry of clinopyroxenes in selected garnet-bearing spinel peridotites is reported in Table 2-5. Clinopyroxenes are Cr-diopsides (average Cr₂O₃ 0.9 wt%). Cores are higher in Al₂O₃ than rims; for example, sample 08BC03 has clinopyroxene cores with 3.1 wt% Al₂O₃ decreasing to 1.8 wt% at the rim.

Plots of Gd versus Yb in clinopyroxene and garnet located along grain boundaries of clinopyroxene or within close proximity to clinopyroxene are shown in Figure 2-11. These plots show that garnet formation in clinopyroxene causes the clinopyroxene host and complementary garnet to evolve away from each other in Gd-Yb space. In particular, clinopyroxenes initially have high Yb, but with progressive equilibration with garnet, Yb becomes depleted. Combined with the Al-depletion halos in clinopyroxene adjacent to garnet (see above), these observations are consistent with the suggestion that these clinopyroxenes initially had high Al and Yb contents and thus formed initially in a garnet-free environment. Their present Al- and Yb-depleted compositions result from subsolidus equilibration with garnet, most likely generated by direct exsolution from the pyroxenes. Primitive mantle-normalized REE spidergrams of clinopyroxenes in 1026V, BC-77, 08BC03, and 08BC08 are shown in Figure 2-10b.

Overall, clinopyroxenes are depleted in heavy rare earth elements (HREEs) relative to LREEs, most likely due to subsolidus equilibration with garnet. However, HREE
concentrations in clinopyroxenes can be highly variable. For example, HREE contents in 1026V vary by more than a factor of 3, the higher values representing the pyroxene cores that have not yet equilibrated with garnet.

**Figure 2-11. Gd vs. Yb systematics in clinopyroxene.** Gd vs. Yb in clinopyroxene (crosses), garnet (circles), and orthopyroxene (x’s) of four Sierran peridotites. Diagonal black lines pass through the whole-rock composition and a point nearly at zero (the coordinates of the whole-rock composition divided by 10,000; this was done to simulate a line forced through zero on a log-log plot). These lines also pass through the region of greatest overlap between garnet and pyroxene analyses. In theory, the lines represent the initial equilibrated whole rock and garnet-free pyroxene. As garnet exsolves from pyroxene, the initial pyroxene evolves away from the black line along one of the grey lines connecting pyroxenes and garnets within individual samples. Although it would be more appropriate to show a straight black line passing through the origin for each individual sample’s population of clinopyroxenes and garnets, only one line is shown here for clarity. (a) 1026V, (b) 08BC03, (c) BC-77, (d) 08BC08.
2.5.6.3. Garnet

Garnets in the Sierran peridotites are pyrope-rich (average end-member composition Py$_{65.5}$Al$_{20.5}$Gr$_{8.3}$). Major element and trace element geochemistry of garnets are reported in Tables 2-7 and 2-8. Energy-dispersive (EDS) scans and quantitative WDS spot analyses of cores and rims of garnet indicate absence of compositional zoning. Interestingly, even though the garnets in the peridotites occur in texturally distinct environments (e.g., lamellae, coronae, and discrete crystals dispersed throughout the matrix), their major element compositions are homogeneous. REE spidergrams (normalized to primitive mantle) of garnets in samples 1026V, BC-77, 08BC03, 08BC04, 08BC08, and 08BC14 are shown in Figure 2-10c. In general, the garnet REE patterns show positive slopes from La to Lu, although garnets within individual samples show a range in enrichment, particularly in MREE and HREE.

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Each column represents an individual spot analysis. All garnets are unzoned with respect to major elements.

Table 2-6. Major element compositions of garnet in garnet-bearing spinel peridotite

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42
Table 2-7. Trace element geochemistry of garnet in garnet-bearing spinel peridotite

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Table 2-8. Major element compositions of spinels

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<td>Mg²⁺</td>
<td>74.04</td>
<td>59.96</td>
<td>63.25</td>
<td>70.49</td>
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</tbody>
</table>

Cr₆⁺=(atomic Cr/[(Cr+Al)x100])
Mg²⁺=(atomic Mg/[(Mg+Fe)x100])
*From Dodge et al. (1988)
2.5.6.4. Spinel

Major element compositions of spinels are reported in Table 2-9. As mentioned earlier, large relict spinels in garnet-bearing spinel peridotites are zoned from low Cr# (atomic Cr/(Cr+Al)x100) cores (30 to 40) to high Cr# rims (≥50) at the contact with garnet coronas (Figure 2-12). The high Cr# rims near garnet are attributed to the formation of garnet because Cr is more compatible in spinel and will be enriched in a residual spinel as garnet forms. Spinels in spinel peridotites are homogeneous. Interestingly, cores of large relict spinels in both spinel and garnet-peridotite types record similar Cr#’s between 30 to 40. Two garnet-bearing peridotites, 08BC04 and P-1, and one spinel peridotite, P-7, have low Cr# spinel cores (~10). The green spinels, which probably represent late-stage garnet breakdown products, all have Cr#’s in spinel much lower than 10.
Figure 2-12. Geochemistry of chromian spinels. (a) through (c) shows Cr# (atomic Cr/(Cr+Al)*100) in spinel vs. whole-rock major and trace-element contents. Only core compositions of the largest relict spinels are plotted. Solid black lines are hypothetical melting trends starting from primitive mantle compositions. (a) Cr# in spinel vs. whole-rock Al₂O₃. (b) Cr# in spinel vs. whole-rock Ti. (c) Cr# in spinel vs. whole-rock Yb. D) Cr# (atomic Cr/(Cr+Al)*100) vs. Mg# (atomic Mg/(Mg+Fe⁷⁺)*100) in spinel for spinel peridotites (small circles) and garnet-bearing spinel peridotites (arrows connect cores to rims of individual spinel grains). Overall, spinel cores in both peridotite types (spinel and garnet-bearing) record melt depletion between 15 – 20 %. Partial melting trend is after (Dick & Bullen, 1984), incorporating data from (Hirose & Kawamoto, 1995; Matsukage & Kubo, 2003). Shaded fields represent global forearc peridotites and abyssal peridotites (data from Arai and Ishimaru (2008)).
2.5.6.5. Olivine

Olivines are homogeneous within individual peridotite samples. Major element compositions are reported in Table 2-10. The Mg#s range from 87.9 to a maximum of 91.5, with the average at 90.5. Ni content does not vary significantly with Mg#; Ni contents fall within the range 0.36 to 0.46 wt%.

<table>
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<tr>
<td>FeO</td>
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<td>MnO</td>
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<table>
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<tr>
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All analyses are individual spots.

Table 2-9. Major element compositions of olivines

2.6. Geothermobarometry

2.6.1. Approach

In this section, we use subsolidus mineral geothermobarometry to constrain the final leg of the P-T path with particular emphasis on constraining final equilibration conditions. In the next section, we attempt to extract prior P-T conditions (the first leg) by linking the degree of melt depletion, as inferred from whole-rock compositions and relict spinel compositions, to paleo-depths. Because of textural and compositional evidence for disequilibrium (e.g., porphyroclastic texture, major element zonation in
minerals, variability in REE content in different mineral grains within the same sample), we intentionally focused on mineral rims, which provide the best opportunity for recording local equilibrium at the final P-Ts through which the system closed. In particular, we examine mineral rim pairs, that is, where two minerals and their rims share a common grain boundary. We used high spatial resolution field-emission electron microprobe imaging to select appropriate rim areas for quantitative analyses. Regions affected by late-stage magmatic infiltration from the host magma or post-eruption alteration were avoided. We calculated final P-T’s for five garnet-bearing spinel peridotites: 1026V, BC-77, 08BC03, 08BC04, 08BC08 (Figure 2-13). Due to extensive serpentinization, samples 08BC14 and 08BC07 did not preserve suitable mineral pairs for geothermobarometry. The textural richness of 1026V and BC-77 make these two samples especially ideal for detailed geothermobarometry involving garnet lamellae and host orthopyroxene.

Of the variety of textures afforded us by the Sierran peridotites, we consider garnet lamellae in large, porphyroclastic orthopyroxenes as providing the best opportunity for unequivocal identification of garnet and orthopyroxene sharing a common grain boundary. Subsolidus equilibration of garnet and orthopyroxene obeys the following net-transfer reaction: MgAl₂SiO₆ + Mg₂Si₂O₆ = Mg₃Al₂Si₃O₁₂. Where garnet lamellae are unavailable, we use garnet coronas around relict spinels and discrete garnet-orthopyroxene pairs. However, geothermobarometric analysis based on discrete mineral grains in texturally and chemically disequilibrated samples is complicated by the difficulty in identifying equilibrated mineral rim pairs because, in many cases, the appropriate minerals do not share a common grain boundary. Use of discrete mineral grains is also complicated by the fact that a multitude of mineral reactions operate. For instance, the transition from spinel to garnet in the upper mantle, manifested texturally by garnet coronas around spinel, is governed by a reaction that involves spinel and orthopyroxene (MgAl₂O₄ + 2Mg₂Si₂O₆ = Mg₃Al₂Si₃O₁₂ + Mg₂SiO₄) as well as clinopyroxene (MgAl₂O₄ + 2CaMgSi₂O₆ = Ca₂MgAl₂Si₃O₁₂ + Mg₂SiO₄) (MacGregor, 1964; O’Neill, 1981). Further complications arise from the fact that many of our samples, as we will discuss later, have been refertilized, which can give rise to different
generations of pyroxenes or even garnet. For these reasons, garnet lamellae and host orthopyroxene mineral pair geothermobarometry provide the most robust means of "seeing through" the effects of macro-scale disequilibrium.

2.6.2. Results: Final P-T of Sierran peridotites

Figure 2-13. P-T diagram of final equilibration conditions of garnet-bearing spinel peridotites. See Table 2-11; points that reflect artifacts of the Al-in-opx barometer are omitted. Filled symbols are final P-T's calculated using garnet-orthopyroxene geothermobarometry ((Harley & Green, 1982; Harley, 1984)); open symbols are final P-T's calculated using two-pyroxene + garnet geothermobarometry ((Ellis & Green, 1979; Krogh, 1988; Brey & Kohler, 1990)). Thin dashed curves represent the spinel-garnet transition estimated from the range of Cr# in Sierran relict spinels (O’Neill, 1981). The low Cr#'s (0.2 – 0.4) are of spinel cores and indicate an originally shallow spinel-garnet transition at ~2 GPa. This transition gradually deepens as manifested by the increase in Cr# from cores to rims, with highest Cr# (0.6) at the contact with garnet coronas. Also shown on this diagram are contours of F (thick dashed black lines), the anhydrous and water-saturated solidus (solid black lines), and the solidus for various water contents (solid grey lines); all calculated using the parameterization of (Katz et al., 2003). The range of F estimated from Cr# in spinel of the Sierran peridotites (15 – 20%) is shown as a shaded parallelogram assuming that the mantle melted by adiabatic decompression at a mantle potential temperature between 1300 and 1400 °C. Also shown is the solidus of amphibole lherzolite constrained by (Green et al., 2010). Three potential P-T trajectories (A, B, C) are outlined by arrows (see text for discussion). (b) Isothermal curves showing the relationship between F and pressure for 0.1 wt% bulk H2O (blue) and 1.0 wt. % bulk H2O (red), calculated using Katz et al. (2003).
2.6.2.1. Spinel peridotites: geothermometry

Due to the absence of garnet, there are no strongly pressure-dependent reactions in spinel peridotites suitable as robust geobarometers. Recently, some geobarometers for spinel peridotites have been suggested, but these involve the Na content of clinopyroxene in equilibrium with a melt (Putirka, 2008), and therefore are difficult to apply in our study because of significant subsolidus re-equilibration and because the melt composition is not known. Thus, for the spinel peridotites, we focus only on estimating temperatures of equilibration. Using Ca-in-orthopyroxene geothermometry (Brey & Kohler, 1990), Lee et al. (2001) showed that cores of orthopyroxene in both Sierran spinel peridotites and garnet-bearing peridotites record temperatures ≥900 °C, whereas corresponding orthopyroxene rims record temperatures ≤800 °C. Likewise, temperatures of orthopyroxene cores in Sierran garnet-bearing spinel peridotites calculated using the “Cr – Al – orthopyroxene” geothermometer of Witt-Eickschen & Seck (1991) range from 910 to 1025 °C, overlapping the values obtained from Ca-in-orthopyroxene thermometry. These results indicate that spinel peridotites were originally hot but subsequently cooled to low final temperatures.

2.6.2.2. Garnet-bearing spinel peridotites

Garnet + orthopyroxene geothermobarometry

The Fe-Mg exchange geothermometer between garnet and orthopyroxene was used in conjunction with a geobarometer based on the solubility of Al in orthopyroxene coexisting with garnet (Harley & Green, 1982; Harley, 1984). We also used the Fe-Mg exchange geothermometer of Harley (1984) with a more recent calibration of the Al-in-orthopyroxene geobarometer of Brey & Kohler (1990), which improves on the older calibration of Harley & Green (1982) by incorporating the activities of additional elements (Ti, Na, Cr) in orthopyroxene and garnet. We found that the combination of Harley & Green (1982) and Brey & Kohler (1990) resulted in an average maximum temperature difference of 10 °C (pressures converged to the same values obtained using...
Harley (1984) and Harley & Green (1982)). Final P-T values plotted in Figure 2-13 are those using Harley (1984) and Harley & Green (1982).

For all P-T determinations, we solved the two geothermobarometers on the same mineral pair. This approach is preferred over thermobarometric combinations that involve different mineral pairs because errors associated with identifying equilibrated mineral pairs increases if the number of different discrete phases not in local equilibrium increases. Average final equilibration temperatures of the garnet-bearing spinel peridotites are 755 °C (670-850 °C) and average pressure are 2.8 GPa (2.4-3.6 GPa). Two peridotites, 1026V and 08BC04, recorded contrasting final P-T’s using different garnet-orthopyroxene pairs in the same sample (Table 2-11). For instance, in 1026V, garnet lamella + orthopyroxene pairs measured within the pyroxene- and garnet-rich band (see Figure 2-4 and inset of Figure 2-5) yield pressures between 1.7-2.1 GPa and temperatures between 670-734 °C. However, within the large relict orthopyroxene porphyroclast in the same sample, P-T’s from garnet lamella + orthopyroxene pairs ranged between 2.9-3.6 GPa and 774-845 °C. An important distinction between the two different garnet-bearing orthopyroxene grains is that the host orthopyroxene giving the lower apparent P has >1 wt% Al₂O₃, whereas the one recording higher apparent P has <1 wt% Al₂O₃.

This inconsistency between final P, T’s calculated within different areas of the same sample can be explained by the different closure rates of the Al-in-opx geobarometer and Fe-Mg exchange geothermometer. This is because the diffusion rates of trivalent cation Al and divalent cations Ca, Mg and Fe in pyroxenes are fundamentally different. For example, at 1000 °C, Al diffusivity in pyroxene is <10⁻²² m²/s (Sautter & Fabriès, 1990), but Fe and Mg diffusivities are ~10⁻¹₈ to 10⁻¹⁹ m²/s (Brady, 1995). These large differences in diffusivity result in very different kinetics of re-equilibration. Thus, the Al distribution between minerals will re-equilibrate more sluggishly than Fe-Mg exchange. In addition, during cooling, Al will “freeze” in earlier (e.g., at higher temperature) than Fe-Mg exchange, which continues to re-equilibrate to lower temperatures. Thus, Al contents in pyroxenes will more likely record a range of
“arrested” states, whereas Fe-Mg distribution between minerals will converge on a common (and lower) closure temperature. We attribute the differences in orthopyroxene Al$_2$O$_3$ content within the same sample, as discussed above, to this phenomenon.

Due to the different kinetics of geothermometers and geobarometers, artifacts in estimated P-T are generated. This can be seen in Figure 2-14, where we plot calculated P-T’s in 1026V and selected isopleths of constant Fe/Mg distribution between garnet and orthopyroxene (Harley, 1984) and isopleths of constant Al in orthopyroxene coexisting with garnet (Harley & Green, 1982). Also plotted in Figure 2-14 are the P-T results obtained by previous authors on similar xenoliths from Big Creek (Dodge et al., 1988; Mukhopadhyay & Manton, 1994; Ducea & Saleeby, 1996). Final temperatures from these studies are considerably hotter than our results; this may be due to these authors using mineral cores instead of mineral rims. We verified this by calculating a few core-core P,T’s, and the results were significantly higher than rim-rim P,T’s. Nevertheless, all P-T estimates collapse on to a small temperature interval, defining a steep dP/dT array. This indicates that most samples had sufficient time to re-equilibrate their rims to ambient temperatures in the Sierran lithosphere. However, wide scatter is observed for apparent equilibration P, indicating that the Al-in-opx barometer closed over different P and T conditions, which we attribute to the considerably more sluggish kinetics of Al compared to Fe and Mg. Thus, the lower pressures (and thus higher Al$_2$O$_3$ contents of host orthopyroxene) are most likely artifacts of the Al-in-opx barometer failing to close at the same final state (low Al$_2$O$_3$). Interestingly, the large opx porphyroclast in 1026V, which records the greatest P (3.6 GPa), contains the thickest garnet lamellae as well as the strongest Al-depleted halos in the pyroxene surrounding the garnet (see traverse in Figure 2-5). These two observations suggest a long time interval over which garnet-opx equilibration operated. In contrast, those samples with thin garnet lamellae and thin Al-depleted halos in surrounding opx, suggest shorter time intervals for diffusional exchange, perhaps as a result of different timing of garnet nucleation. In summary, the foregoing analysis indicates that calculated P’s are lower bounds for actual equilibrium P. Calculated T’s may represent the ambient temperature of the deep lithosphere, but given evidence for “temperature” zonation in pyroxenes as determined
from Ca-in-opx geothermometer (Lee et al., 2001), it is possible that estimated T’s are maximum bounds.

Figure 2-14. P-T diagram illustrating artifacts of thermobarometers. Such artifacts are due to different kinetics of geothermometers and geobarometers. In this figure, final P-T calculated using garnet lamellae + orthopyroxene pairs from different areas within one sample, 1026V, record a narrow interval of final temperatures, but a wide array of final pressures. The darker and lighter shaded diamonds represent garnet + opx pairs in two distinct areas of the sample. Dashed lines represent lines of constant K, defined as \( \frac{(Fe/Mg)_{gt}}{(Fe/Mg)_{opx}} \) (“iso-K line”) for the garnet-orthopyroxene thermometer of Harley (1984) and solid lines represent lines of constant Kd, defined as \( \frac{Fe}{Mg} \) for the Al-in-opx barometer of Harley and Green (1982). Also plotted as open circles are all peridotites in this study for which final P-T were calculated, and plotted as filled circles are peridotites from Lee et al. (2001). Shaded field represents final P-T range of peridotites from previous studies on Big Creek xenoliths.
Garnet + 2 pyroxene geothermobarometry

In addition to garnet + orthopyroxene geothermobarometry, we applied two calibrations of the garnet-clinopyroxene Fe-Mg exchange geothermometer (Ellis & Green, 1979; Krogh, 1988) with the solubility of Al-in-orthopyroxene geobarometer (Brey & Kohler, 1990). We also used the two-pyroxene geothermometer of Brey & Kohler (1990) combined with their geobarometer. However, due to the secondary, metasomatic origin of clinopyroxene in some of the more depleted peridotites (to be discussed later), as well as the scarcity of suitable garnet, clinopyroxene, and orthopyroxene grains adjacent to each other, we have fewer estimates from garnet + two pyroxene geothermobarometry than garnet + orthopyroxene geothermobarometry.

The discrepancy in final P-T estimates observed in different areas of the same sample was also seen in results from garnet + two pyroxene geothermobarometry, particularly in peridotite 1026V. This observation lends further support to the notion of variable closure of the Al-in-opx barometer. For the remaining peridotites, results from garnet + two-pyroxene geothermobarometry were consistent with results obtained from garnet + opx geothermobarometry.

2.6.2.3. Trace-element thermometry

An attempt was also made to calculate final equilibration temperatures using subsolidus equilibrium distribution of trace elements between clinopyroxene and orthopyroxene (Witt-Eickschen & O'Neill, 2005) using in situ laser ablation ICP-MS data. Only the middle REEs (e.g., Ho, Er) and Y yield temperatures that fall within the range calculated from major element data, although the range obtained is quite large (600 – 1100 °C). Observed distribution of HREEs (e.g., Yb and Lu) between cpx and opx (cpx/opx) are extremely low (<3), yielding obviously unreasonably high calculated temperatures (>2000 °C). One possible explanation for this unreasonable result is that the calibration of Witt-Eickschen and O’Neill was designed for spinel facies peridotites rather than garnet peridotites. However, the more likely explanation is that the laser spot size was too large (>30 μm) to obtain the fine spatial resolution necessary for quantifying
rim-rim compositions in highly zoned minerals. Like Al, the REEs have very low
diffusivities and therefore the large spot size of laser ablation prohibits clean analysis of
fine-scale rims for proper trace-element geothermobarometry.

We also applied the Ni in garnet geothermometer (Canil, 1999). This
geothermometer assumes that the Ni activity in a peridotitic system is approximately
buffered by olivine Ni content (Griffin et al., 1989). We applied this geothermometer
only to discrete garnets within the olivine matrix (garnet lamellae were avoided).
Temperatures obtained from this geothermometer range between 839-928 °C, which are
higher than those obtained from Fe-Mg exchange thermometry between garnet and
pyroxene.

2.6.2.4. Spinel-garnet transition constraints on P

We have also constrained final pressures using the Cr# of relict spinel, which is a
sensitive indicator of the pressure at which the spinel-garnet transition occurs in the upper
mantle (O'Neill, 1981; Klemme, 2004). Relict spinels in the garnet-bearing spinel
peridotites have low Cr# cores (20 – 30), corresponding to a spinel-garnet transition of
2.2 GPa at 900 °C, and thus indicate that these relict spinels formed at pressures less than
~2.2 GPa. In contrast, rims of spinels in contact with garnet coronas have high Cr# (≥ 50),
corresponding to a deeper spinel-garnet transition at ~3 GPa at 900 °C (O'Neill, 1981).
The observation that spinels in the spinel peridotites are unzoned and record
similar Cr#’s as cores in relict spinels of the garnet-bearing peridotites is strong evidence
that the spinel peridotites are the shallower protoliths of the garnet-bearing spinel
peridotites, which we discuss in detail below.
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<td><em>Garnet lamella in small opx in diffuse pyroxene-rich band</em></td>
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2.7. Degree of melt depletion of Sierran peridotites

2.7.1. Cr# in spinel as an indicator of melt depletion

In this section, we estimate the melting degree (F) of the peridotites. Such information can be used to estimate depths of melt extraction using the relationship between F and P from mantle melting parameterizations (Katz et al., 2003; Langmuir et al., 2006). Methods of estimating F of peridotites involve tracking the depletion or enrichment of an element during partial melting. The Cr# of spinel has been demonstrated to be sensitive to petrogenetic processes such as fractional crystallization and melting, due to the high compatibility of Cr in the solid and the incompatibility of Al during peridotite partial melting (Dick & Bullen, 1984). Experimental studies (Hirose & Kawamoto, 1995; Matsukage & Kubo, 2003) verify this by showing a positive correlation between F and spinel Cr#. A further advantage of using spinel Cr# is that Cr and Al diffuse slowly, so spinel Cr# may be more likely to preserve the original signature of melt depletion even if the rock has been subsequently refertilized.
One caveat in employing Cr# in spinel as a melt depletion indicator is that the formation of garnet coronas preferentially extract Al from the spinel, leaving a “residual” spinel with anomalously higher Cr# (O’Neill, 1981; Webb & Wood, 1986; Obata & Morten, 1987). This is a concern for the Sierran peridotites because the majority of spinels are rimmed by garnet, leading to spinels with Cr-enriched rims due to equilibration with garnet. Failure to obtain the original Cr# of spinel before garnet formation can lead to spurious estimates of melting degree. We thus focus on spinel cores because they provide the most meaningful Cr# signature of melt depletion. Some garnet-mantled spinels are large enough that Cr-enrichment remains confined to a sharply defined rim, as shown by nearly opaque rims and a largely homogeneous, amber-colored interior, suggesting preservation of original spinel composition in the core. In these coronas, the ratio of garnet to spinel is low, indicating only limited conversion of spinel to garnet. In contrast, coronas in which the garnet to spinel ratio is large are often characterized by small spinels that are either completely opaque or show continuous zonation from rim to core. The effect of garnet corona formation at the expense of spinel on spinel Cr/Al ratio was modeled using exchange reactions and partitioning relationships based on Webb and Wood (1986). The model results show that appreciable changes in spinel Cr# occur after ~50% of the system has been converted to garnet. Focusing on the core compositions of the largest spinels should thus circumvent the “garnet formation effect”. With the above considerations in mind, we use Cr#’s from only the largest relict spinels in both spinel and garnet-bearing peridotites. We find that spinel core Cr#’s in both peridotite types are consistently between 30 to 40, which corresponds to melting degrees between 0.15 and 0.20 (Figure 2-12).

2.7.2. Whole-rock composition as an indicator of melt depletion

We can also estimate melting degree (F) by comparing whole-rock compositions to modeled whole-rock melt depletion trends. We choose moderately incompatible elements, such as Al, Ti, and Yb, because unlike highly incompatible elements such as the LREEs, moderately incompatible elements will not be depleted as rapidly during melting, and thus are useful as an in indicator of melting degree over a wide range of F.
In addition, moderately incompatible elements are less likely disturbed by cryptic metasomatism compared to highly incompatible elements. We modeled partial melting by starting with a primitive mantle source (McDonough & Sun, 1995) and assuming melting occurred in the spinel stability field because the protoliths of the garnet-bearing spinel peridotites were garnet-free (see above discussion). We use a non-modal, fractional melting model with mineral modes from Johnson et al. (1990). Mineral/melt partition coefficients for Yb (at 1400°C) were taken from Lee et al. (2007a). Mineral/melt partition coefficients for Ti were those used by Johnson et al. (1990). For Al$_2$O$_3$, we used the parameterization of Canil (2002), which relates a bulk rock/melt partition coefficient of Al$_2$O$_3$ as a function of F.

The modeled whole-rock Al$_2$O$_3$, Ti-, and Yb- melt depletion curves are shown in Figures 2-8 and 2-12. In Figure 2-8, we plot Ti versus Yb and Ti and Yb versus Al$_2$O$_3$. In Figure 2-12, we show melt depletion curves involving Cr# in spinel with Al$_2$O$_3$, Ti, and Yb. As expected, melt depletion results in positive correlations between Al$_2$O$_3$, Ti, and Yb, and negative correlation between spinel Cr# and whole rock Al$_2$O$_3$, Ti and Yb. Inspection of these plots reveals that the Sierran peridotites fall on the Cr# in spinel versus whole-rock Ti melting curves but not on the Cr#–Yb and Cr#–Al$_2$O$_3$ melting curves. Whole-rock Al$_2$O$_3$ and Yb deviate from predicted melting trends to higher values than that expected for a given spinel Cr#. Al$_2$O$_3$ and Yb also appear to be decoupled from melt depletion trends plotted against Ti as can be seen by excess Al$_2$O$_3$ and Yb for a given Ti (Figure 2-8). We note that whole-rock Yb and Al$_2$O$_3$ are correlated, but because neither Al$_2$O$_3$ nor Yb correlate with spinel Cr#, the correlation between Yb and Al$_2$O$_3$ may instead reflect refertilization (to be discussed later). In addition, most garnet-bearing spinel peridotites having whole-rock Na$_2$O contents greater than primitive mantle values also fall off predicted Cr# in spinel versus Na$_2$O melt depletion curves. As with Yb and Al$_2$O$_3$, Yb and Na$_2$O also show a positive correlation, consistent with a refertilization origin for such trends. In contrast, the adherence of whole-rock Ti and Cr# in spinel to modeled melt depletion trends suggests that Ti and Cr remained largely unperturbed and thus retain the original melt depletion signature of the garnet-bearing spinel peridotites. We attribute the preserved record of melt-depletion in spinel Cr# and
whole-rock Ti content in light of a more fertile major element composition for the whole-rock to reflect disequilibrium. In summary, spinel Cr# and whole-rock Ti both indicate overall melting degrees of ≥15% (relative to a primitive mantle major element composition) for the Sierran peridotites.

2.8. Discussion: Synthesis of P-T-Composition Path

2.8.1. Summary of textural, geothermobarometric, and geochemical observations

Al contents of orthopyroxene coexisting with garnet and Cr# in spinel coexisting with garnet indicate that the garnet-bearing spinel peridotites achieved final equilibration at <750 °C and pressures ~3 GPa (~90 km). However, textural and chemical evidence indicate that the garnet-bearing spinel peridotites represent metamorphosed melt-depleted spinel peridotites. First, abundant garnet coronas around spinels indicate a transition from spinel to garnet facies. Second, orthopyroxene cores have Al₂O₃ contents too high to be in equilibrium with garnet and instead are similar to Al₂O₃ contents of orthopyroxene in spinel facies peridotites (see also Lee et al. (2001)). Third, relict deformed orthopyroxenes contain numerous garnet lamellae accompanied by Al-depletion halos in the adjacent pyroxene, suggesting garnet exsolved from originally high-Al pyroxene, also typical of spinel facies peridotites. Fourth, low whole-rock HREE contents point to the absence of garnet during melt depletion. Fifth, spinels in both spinel and garnet-bearing spinel peridotites share compositional and textural features that demonstrate a common petrogenesis – Cr# in spinel cores of both peridotite types are identical. This overall conversion of spinel- to garnet-bearing spinel peridotites was accompanied by a progressive decrease in olivine grain size and the development of weak pyroxene foliation bands. We interpret these changes in metamorphic textures to reflect the transformation of coarse-grained, undeformed spinel peridotite protolith (i.e., 08BC13 in Figure 2-4) into increasingly recrystallized, slightly deformed spinel peridotite (P-7), and finally into fine-grained, porphyroclastic, deformed garnet-bearing peridotite.
This transformation of spinel peridotite protolith to garnet peridotite also appears to have been accompanied by refertilization.

### 2.8.2. Constraints on the first leg of P-T path using melt depletion as a proxy for paleodepth

The question that now arises is how was the transition from spinel to garnet facies achieved? There are two endmember scenarios. The first is isobaric cooling of a residue that melted at the final equilibration pressures (~3 GPa). The second is shallow melting in the spinel stability field, followed by compression and cooling into the garnet field. The degree of melting (F) experienced by upwelling mantle depends on the extent to which such mantle lies above the solidus. Because the solidus has a positive dP/dT, this means that F is controlled by T and P. The higher the temperature and lower the pressure, the higher F will be. Our objective is to obtain P, corresponding to the original P of the spinel peridotite protoliths of the garnet-bearing peridotites. To do this, we need estimates of F and T. Based on Cr# in relict spinel cores and whole-rock Ti contents, as discussed above, estimated melting degrees ranged between 15 – 20%. Temperature of melting can be bounded by assuming reasonable mantle potential temperatures, e.g., 1300-1400 °C, but a less arbitrary approach is to use the temperature-dependency of Fe and Mg partitioning between olivine and basaltic melt. This approach is shown in Figure 2-15 and is based on Hanson and Langmuir (1978) and modified by Lee et al. (2011). It can be seen that at temperatures <1300 °C, Fe becomes compatible such that the FeO content of peridotites increases with melting degree (this approach does not depend on water because the effect of water is to depress the solidus, which in turn changes the compatibility of Fe and Mg because the temperature of melting is decreased). Ignoring those samples that have experienced refertilization, e.g., the samples with low MgO and low Mg#, it can be seen that the most melt-depleted peridotites (high Mg#) have low FeO contents and thus are best explained by melting at temperatures between 1300-1400 °C.
Figure 2-15. Plot of whole-rock FeO vs. MgO for peridotites. This diagram follows approach described in Hanson and Langmuir (Hanson & Langmuir, 1978) and modified by Lee et al. (Lee et al., 2011). Dark shaded field represents "normal" (not silica-enriched) cratonic peridotites and light shaded field represents Phanerozoic continental peridotites for reference. Filled circles represent Sierran spinel peridotites and symbols represent Sierran garnet-bearing spinel peridotites (see Figure 2-6 for key to symbols). Solid lines represent isothermal melting paths, dashed lines represent lines of constant melting degree, and dotted lines represent isopleths of constant whole-rock Mg#. Arrow schematically points to the direction of refertilization.

We now input these constraints on F and T into anhydrous and hydrous melting models. To achieve F>15 % at reasonable mantle potential temperatures (1300-1400 °C) under anhydrous (<0.1 wt% H2O) conditions, final pressures of melting of ~1.5 GPa (~45 km) are required (Figure 2-13). In order to achieve the same melting degree at the present equilibration pressures of the garnet peridotites (3 GPa), temperatures >1500 °C are needed (Katz et al., 2003), which seems unlikely given the above constraints on original melting temperatures. However, if higher water contents are assumed, this
temperature constraint can be relaxed because the effect of hydration is to depress the solidus and increase the melting degree for a given temperature and pressure. We have assessed the effect of hydration using the parameterizations of Katz et al. (2003) and assuming the case where bulk H$_2$O is 1 wt. %, a maximum bound on peridotite water content based on inverting typical water contents of arc basalts (Kelley et al., 2010) for source composition (Figure 2-13). At 1300-1400 °C, it can be seen that 15 – 20% melting under water-fluxed conditions can occur at 2.5-3.5 GPa. Because we do not know the original water contents of the unmelted peridotite protoliths nor do we know the amount of water fluxed through the system during melting, it would seem from these models alone that both high-P and low-P origins are permissible and thus, we are unable to distinguish between the endmember scenarios discussed above.

There are, however, reasons why hydrous fluxed melting at 3 GPa is probably unlikely to explain the petrogenesis of these peridotites. First, pressures of 3 GPa correspond to depths of ~100 km, approaching the typical depth of the slab beneath volcanic arcs (100-110 km). In such proximity to the cold slab, the asthenospheric mantle wedge would be expected to be colder than the ambient mantle. Hydrous flux melting should also initiate melting at temperatures (~900 °C) corresponding to the water-saturated peridotite solidus (Katz et al., 2003). Instead, the temperatures inferred from whole-rock FeO and MgO contents (1300 – 1400 °C), as discussed above (Figure 2-15), are far higher than the water-saturated solidus, and thus suggest derivation from the hot core of the asthenospheric mantle wedge. Finally, if water-saturated melting had occurred, these peridotites should have been within the stability field of amphibole, which can reach up to 3 GPa (Green et al., 2010). Given that these samples are cold and hence well within the amphibole stability field, water-saturated conditions should have been manifested in the presence of primary amphiboles, but this is not seen.

We consider the second endmember, shallow anhydrous melting in the spinel stability field followed by compression to and cooling in the garnet stability field, as our preferred hypothesis. This view is corroborated by the nature of Cr# zoning in the spinels mantled by garnet coronas. As discussed above, the low Cr# of the spinel cores
bounds original pressures to be <2.2 GPa, whereas the Cr\# of the spinel rim in contact with garnet coronae suggests final pressures of ~3 GPa. Based on these discussions, we are left with a spectrum of possible P-T paths, all involving an increase in P (Figure 2-13a). One extreme path (path A in Figure 13) is the case of shallow melt depletion, followed by (shallow) isobaric cooling, and then rapid, near-adiabatic compression. At the other extreme is path C wherein the peridotites first experience shallow melt depletion at C1, then undergo adiabatic compression at high temperatures (C2) followed by isobaric cooling at high pressures (C2 to C3). Somewhere in between these two extremes is path B where cooling and compression occur in tandem. In the next section, we attempt to select the most likely of these P-T paths.

2.8.3. Additional constraints from patterns of refertilization

As discussed above, many Sierran garnet-bearing spinel peridotites do not fall on conventional melting trends through canonical primitive mantle (Figures 2-7, 2-8). In addition, some samples show excess whole-rock CaO, manifested in the form of diffuse clinopyroxene bands that define a weak foliation. These features all hint at post melt-depletion refertilization by basaltic melts. The most important feature in the context of this study, however, is that the deepest samples show the largest excesses of whole-rock Yb for a given spinel Cr\#. This excess Yb is correlated mineralogically with the presence of garnet. This implies that Yb was scavenged from the refertilizing melts by pre-existing garnet or that Yb was added to the peridotite because of garnet precipitation from the melt. The overall REE pattern of most of the Sierran garnet-bearing peridotites, with enriched HREE relative to MREE, also suggests that metasomatic garnet was added to a previously melt-depleted spinel peridotite protolith.

In either case, refertilization occurred in the garnet stability field, not in the spinel stability field. Thus, refertilization is constrained to occur on the high-pressure leg of the P-T paths described above. We note that the temperatures of refertilizing melts were most likely above 1000 °C, because it seems highly unlikely that refertilization could have occurred at the cold final equilibration temperatures of ~700 °C, conditions at which
the passage of such melts would have re-equilibrated the cooling-related core-to-rim chemical zonation. If these interpretations are correct, refertilization therefore happened at high P and high T, ruling out P-T path A wherein cooling occurs at shallow pressures followed by an increase in pressure. The most likely P-T path is that of C, wherein the peridotites first melt at low P and high T, then experience an increase in P at high T, followed by refertilization at depth and finally by cooling at high P. In a future paper, we will discuss what this refertilizing magma may have been, but further discussion is beyond the present scope. We only highlight the fact that whole-rock Ti and other high field strength elements, such as Nb, are depleted relative to the REEs. These geochemical signatures are consistent with those of typical subduction zone magmas.

2.9. Implications for evolution of continental arcs

We have shown that the deep Sierran lithospheric mantle was initially melt-depleted at shallow pressures between 1–2 GPa, then experienced compression and cooling, finally reaching equilibration within the garnet stability field at ~3 GPa. Based on additional constraints from the depth at which refertilization occurred, we also concluded that the peridotites increased in pressure at high temperatures, then became refertilized, and finally cooled isobarically at high pressures (path C in Figure 2-13).

We now return to the various thickening scenarios described in the introduction (Figure 2-2). Underplating of melt-depleted mantle to the base of pre-existing lithosphere (Figure 2-2c) is one mechanism of thickening. However, in such a scenario, a given parcel of underplated material will never experience an increase in P because the overburden weight remains constant (and may even decrease if there is erosion at the Earth’s surface). Such mantle would experience isobaric cooling. More likely scenarios for explaining path C are tectonic thickening (Figure 2-2a) and magmatic inflation via sill-like intrusions in the crust or lithosphere (Figure 2-2d). In the case of tectonic thickening, we can probably rule out thickening resulting from suturing of arc terranes onto the North American margin. These events ended by the Triassic (Dickinson, 1981), but the melt-depletion recorded by the peridotite xenoliths was most
likely accompanied by significant magmatism, which occurred in the Cretaceous (Lee et al., 2000). Although there is considerable evidence for mega-scale continental underthrusting during the peak of Cretaceous arc magmatism (e.g., Ducea and Barton (2007); DeCelles et al. (2009)) we consider this process an insufficient explanation for the P-T history recorded by the peridotite xenoliths because such a scenario would result in extensive heating of the underthrusted continental lithosphere, and this is not seen. We suggest that thickening via magmatic inflation during the main arc-building stage is better able to explain the P-T history of the deep Sierran lithosphere. The Late Cretaceous Sierra Nevada experienced an incredibly high magmatic flux in a short time period: in the span of ~10 Ma, over 4000 km$^2$ of new crust was emplaced (Coleman & Glazner, 1997). With this context in mind, we envision the following scenario. Shallow, moderate to high degree melting within the spinel stability field characterizes the early stages of arcs because the lithosphere is initially thin, allowing more headspace for decompression melting. As magmas rise into the crust, they reach a depth of neutral buoyancy, generating sills. Progressive addition of these sills results in the thickening of the arc by magmatic inflation. Any rocks below the intrusion depth will progressively increase in pressure due to the gradual increase in over-burden pressure generated by magmatic inflation. These processes subsequently compress the deep lithosphere to greater depths. Because this compression is intimately linked with magmatic addition, this model would predict compression at relatively high temperatures, consistent with our interpreted P-T path. We suggest that refertilization occurred after the lithosphere became thickened (3 GPa ~ 90 km). After thickening and refertilization, the lithosphere cooled to temperatures as low as 700 °C. Such cold temperatures can only be attained in ambient mantle near a cold subducting slab. We thus suggest that cooling took place after magmatic inflation had increased the thickness of the Sierran lithosphere to depths approaching the subducting slab interface (typically ~100 km beneath arc fronts) (Syracuse & Abers, 2006; Grove et al., 2009; England & Katz, 2010), thereby pinching out the hot asthenospheric mantle wedge and juxtaposing Sierran deep lithosphere against the cold subducting Farallon plate. Thus, the cooling recorded by peridotites is the result of deep refrigeration by the subducting Farallon plate. Such
cooling is predicted to be rapid because of the strong thermal contrast between the slab and mantle lithosphere. Indeed, the presence of Al-depleted haloes next to garnet lamellae (Figure 2-5) requires cooling to be fast enough to preserve these features. Testing this hypothesis further will require geochronological constraints on the timing of cooling of the deep lithospheric mantle.

If our proposed P-T and compositional history is correct, there are implications for the magmatic evolution of North American Cordilleran belt. Magmatism in the Sierra Nevada appears to have shut down abruptly around 80 My, an observation used as evidence for the onset of “flat” subduction (Bird, 1988; Lipman, 1992). However, the presence of deep arc lithosphere as recorded by the Sierran xenoliths has been used to argue that “flat” subduction never occurred beneath the Sierra Nevada (Saleeby, 2003). Could the Sierran continental arc have terminated simply because it grew too thick to allow for decompression melting and for the presence of the hot asthenospheric mantle wedge necessary to heat and dehydrate the subducting slab?

2.10. Conclusions

We have shown that the deep Sierran lithosphere, represented by a suite of peridotite xenoliths, first experienced hot, shallow melting (1-2 GPa, >1000 °C). They were then transported at high temperatures to pressures of ~3 GPa and then subsequently cooled to temperatures below 800 °C. This P-T path transformed the spinel peridotites into garnet-bearing spinel peridotites. Refertilization of the arc lithosphere appears to have predominantly affected the deepest equilibrated samples, and based on the preservation of deformation textures and chemical zonation, occurred before cooling. These observations suggest that refertilization occurred during the compressional stage of the P-T path. Evidence supporting this P-T path is briefly recapitulated below:

1. Spinel peridotites are characterized by coarse-grained, protogranular, nonfoliated textures. Garnet-bearing spinel peridotites, however, are characterized by fine-
grained, recrystallized, porphyroclastic textures. Garnet-bearing peridotites also show a weak foliation defined by diffuse bands of clinopyroxene and/or garnet, suggesting melt infiltration. In addition, garnet-bearing peridotites contain ubiquitous relict spinels rimmed by garnet coronas. Some samples also contain relict orthopyroxene that exsolved garnet lamellae.

2. Garnet-orthopyroxene geothermobarometry indicates that the garnet-bearing spinel peridotites have final equilibration pressures between 2.4 and 3.6 GPa and temperatures between 670 and 870 °C.

3. Cr# in relict spinel cores in both spinel peridotites and garnet-bearing spinel peridotites fall between 30 and 40, corresponding to 15 – 20% melting.

4. High Al₂O₃ (>4 wt. %) in orthopyroxene cores, low whole-rock HREE content, and the anomalously high mantle potential temperatures (>1500 °C) associated with 15 – 20% melting at ~3 GPa preclude deep melting at the pressures of final equilibration, but rather favors melt depletion at shallow pressures (1 – 2 GPa) within the spinel stability field.

5. Refertilization has affected almost all the garnet-bearing spinel peridotites, and is manifested by excess whole-rock CaO and Yb, corresponding to excess clinopyroxene and garnet. Spinel peridotites have escaped substantial (i.e. modal) refertilization.

We interpret the Sierran peridotites as representing new lithospheric mantle formed by shallow decompression melting during Cretaceous arc magmatism. This newly generated lithospheric mantle underwent an increase in pressure and a decrease in temperature. We speculate that this P-T path was associated with magmatic inflation that thickened the arc though tectonic thickening may have also been operating. As the arc thickened, melt-depleted peridotites formed at shallow depths, were transported to greater depths where they became more susceptible to refertilization. Thickening continued until the Sierran arc lithosphere impinged against the subducting Farallon plate. This terminated arc magmatism and resulted in the conductive cooling of the Sierran lithosphere due to refrigeration by the cold Farallon plate.
Chapter 3

On the origin of hot metasedimentary quartzites in the lower crust of continental arcs

This chapter has been edited, reformatted, and reprinted from Earth and Planetary Science Letters (full citation below):


Volcanic arcs associated with subduction zones are thought to be the primary building blocks of continents. The composition of the magmas, particularly in continental arcs, is the product of mixing between differentiation of juvenile magmas and pre-existing crustal wallrock, the former being typically mafic and the latter more silicic. Because the upper continental crust is on average thought to be more silicic than the mafic lower crust, mixing with silicic endmembers should occur primarily in the upper crust. However, we show here that the lower crust of continental arcs contains silicic metasediments. We examine garnet-bearing, granulite-facies sedimentary quartzite xenoliths from the Sierra Nevada batholith in California, a Cretaceous continental arc.
The quartzites have equigranular textures and contain quartz (>50 %), plagioclase (<30 %), garnet (10%), and small amounts (<1 %) of rutile, aluminosilicate, biotite, monazite, zircon, graphite and trace orthopyroxene. Cathodoluminescent images show zircons with rounded detrital cores mantled by metamorphic overgrowths. Hf isotopic model ages and U-Pb upper intercept ages, for a given zircon, are similar, but the zircon population shows variable protolith ages ranging from Proterozoic to Archean. In contrast, all zircons share similar lower intercept U-Pb ages (103 ± 10 Ma), which coincide with the peak of arc magmatism in the Sierra Nevada. The Precambrian protolith ages are similar to North American cratonal basement, and together with the abundance of quartz and detrital zircons, suggest that these quartzites represent ancient, passive margin sediments instead of juvenile active margin sediments in the oceanic trench and accretionary prism. Importantly, these quartzites record peak metamorphic temperatures and pressures of 700 to 800 °C using Ti-in-quartz thermometry and 0.7 to 1.1 GPa using garnet-aluminosilicate-plagioclase thermobarometry, indicating that these xenoliths experienced significant heating and possible partial melting in the lower crust, most likely related to arc magmatism as suggested by similarities between the lower intercept U-Pb ages and the ages of plutonism in the Sierra Nevada. Possible mechanisms by which these sediments were transported into the lower crust include continental underthrusting beneath the continental arc, underplating by buoyant slab-derived sedimentary diapirs, or viscous downflow of country rock in response to diapiric ascent of plutons. Continental underthrusting has been independently documented during the Sevier orogeny, coinciding with the peak of arc magmatism. We thus speculate that supracrustal rocks may have been underthrust into deep crustal magmatic zones. Regardless of how these metasediments arrived in the lower crust, our observations indicate that silicic metasediments occur in the lower crust of volcanic arcs, not just in the upper crust as is commonly thought. Transport of metasediments into deep crustal magmatic zones should influence the composition of arc magmas and continental crust in general.
3.1. Introduction

Arc magmas formed above subduction zones are considered to be essential components in building continents. The compositional stratification of continental crust, from felsic upper crust grading into mafic lower crust, is attained through a combination of igneous differentiation and interaction of juvenile arc magmas with pre-existing wallrocks (Hildreth, 1981; Hildreth and Moorbath, 1988). Fractional crystallization of a basaltic melt results in the retention of dense, mafic cumulates in the deep lithosphere, leaving the residual felsic melts to escape to shallower crustal levels owing to their lower densities. Further refinement of magmas in the upper crust may also involve mixing with pre-existing wallrocks, which are typically felsic.

In this study, we present evidence for silicic metasedimentary rocks in the lower crust of a mature continental arc, the Sierra Nevada Batholith in California. The metasediments are nearly pure quartzite or metasandstone xenoliths (>85 wt. % SiO$_2$). They contain metamorphic garnet and last equilibrated within the granulite facies, recording temperatures in excess of 700 °C and pressures of 1 GPa. Upper intercepts of U-Pb discordia and Hf isotope systematics in detrital zircons show that the metasediments represent strata of the ancient passive margin of North America. Detrital zircon cores are rimmed by metamorphic overgrowths, which yield a common U-Pb lower intercept age coeval with the Late Cretaceous peak of arc magmatism in the Sierra Nevada.

Such quartz-rich sediments are anomalous in the mafic lower crust, suggesting that they were transported from their origin in shallow passive margin depocenters into the deep crust. We propose continental underthrusting as one mechanism, but explore other hypotheses, such as wallrock downwelling associated with plutons rising into the crust and underplating by sediment diapirs derived from the subducting slab. The Sierran lower crustal quartzites offer a unique window into the lower crust, and regardless of the mechanism of their emplacement, indicate that the lower crust may be more felsic than previously thought.
3.2. Geologic setting

During the Paleozoic, the western margin of North America was defined by a passive margin. In the early Mesozoic, island arcs and other exotic terranes were accreted on to this margin (Dickinson, 1981). This transition in tectonic environment was manifested as two distinct Cordilleran depositional settings: mature passive margin sediments (historically referred to as “miogeoclinal”) deposited in the Paleozoic, and juvenile (“eugeoclinal”) sediments associated with the Mesozoic active margin (Kistler, 1990; Moore and Foster, 1980; Stewart, 1970). The passive margin sediments formed a westward thickening wedge of siliciclastics and carbonates extending from eastern Utah to southwestern Nevada and eastern California. Its maximum estimated thickness was ~10 km in the west (Armstrong, 1968; Stewart, 1970). Facies changes, sediment thickness, and paleocurrent directions indicate that the primary direction of sediment transport and deposition was from east to west (Seeland, 1968; Suppe, 1985). Dating of detrital zircons in the passive margin strata indicate provenances of Precambrian North American basement terranes (Bennett and DePaolo, 1987; Gehrels et al., 1995; Grasse et al., 2001) (Figure 3-1).
Figure 3-1. Map of Western US with major basement terranes. U-Pb crystallization ages and Hf model ages (parenthesized) are shown. Blue arrows denote paleocurrent directions of Late Precambrian to Early Ordovician clastic sediments and deposition onto the Cordilleran passive margin. B) Schematic cartoon, modified after Suppe (1985) showing the westward thickening clastic wedge of the Cordilleran miogeocline (passive margin). Sedimentary thickness reached a maximum of ~ 10 km at the westernmost limit in western Nevada/eastern California. C) Simplified map of the central – northern Sierra Nevada Batholith. White star represents xenolith locality. Yellow areas represent passive margin rocks now cropping out as roof pendants.

Passive margin deposition ceased when the Farallon plate began subducting beneath western North America in the early Mesozoic, leading to accretion of island arcs and associated juvenile sediments. Cordilleran arc magmatism occurred between 220 and 80 Ma (Chen and Moore, 1982; Stern et al., 1981), resulting in the development of the Sierra Nevada Batholith (SNB). The greatest volume of magma, however, appears to have been emplaced between ~120 and 80 Ma with a pronounced peak in the east-central
Sierra between 98 and 86 Ma (Coleman and Glazner, 1997). Plutons associated with arc magmatism were emplaced through the edge of the old passive margin. The paleo-boundary separating the ancient passive margin, marking the ancient edge of North America prior to intrusion by Cretaceous granitoids, is inferred from the initial $^{87}\text{Sr}/^{86}\text{Sr}$ contour of 0.706 in the plutons as defined by Kistler and Peterman (1973) and Kistler (1990). West of this contour, plutons intruded into predominantly Paleozoic and younger accreted terranes; east of this contour, plutons intruded into passive margin strata and presumably ancient North American lithosphere. Relics of these pre-batholithic rocks are preserved as “roof pendants” scattered throughout the SNB, many showing evidence for extensive contact metamorphism.

The focus of this study is a suite of garnet-bearing metasedimentary xenoliths from the Miocene Big Creek diatreme (37° 13’ N, 119° 16’ W) located in the central SNB. Big Creek is one of the few volcanic pipes within the SNB containing xenoliths sampling all levels of the arc crust and mantle lithosphere. Lower crustal pyroxenite cumulates are most abundant, followed by upper mantle peridotites, granulites, and rare metasedimentary xenoliths. Several studies on the pyroxenites and peridotites have illuminated the nature and evolution of the deep Sierran lithosphere (Dodge et al., 1988; Ducea and Saleeby, 1996; Lee et al., 2001; Lee et al., 2000; Lee et al., 2006; Mukhopadhyay and Manton, 1994).

Due to their rarity, comparably less is known about the metasedimentary xenoliths and their relationship to the crustal architecture of the Sierran arc. Dodge et al. (1988) reported one eclogite-facies, grossular-clinopyroxene xenolith (grospydite) out of a suite of 16 samples, and interpreted the protolith to be a Precambrian carbonate sediment. A more recent study by Ducea and Saleeby (1998) dated one eclogite-facies garnet-bearing quartzite, and obtained a Sm-Nd age of 85.7 ± 3.2 Ma. A study of 23 lower crustal metasedimentary xenoliths from the Eastern Mojave Desert by Hanchar et al. (1994) indicates a heterogeneous lower crust, encompassing the entire spectrum from quartzofeldspathic, pelitic, and mafic lithologies. Two xenoliths from their study are
garnet-bearing quartzites similar to the few that have been found in Sierran xenolith suites.

3.3. Methods

Whole rock major element data were determined by X-ray fluorescence spectroscopy (XRF) at Washington State University Pullman. All samples were sent as whole rocks, ground in-house and fused into di-lithium tetraborate glasses at WSU Pullman. We then determined whole rock trace element geochemistry using the fused glasses by laser ablation ICP-MS (ThermoFinnigan ELEMENT2 with a New Wave 213 nm laser ablation system) at Rice University. The laser was operated with a 110 µm diameter spot size, a ~17 J/cm² fluence, and a 10 Hz pulse frequency. Cs, Pb, U, Th, Ba, Rb, Nb, Ta, Y, REE were analyzed in low mass resolution (m/Δm = 300) and Na, Mg, Al, Si, P, K, Ca, Sc, Ti, V, Cr, Mn, Fe, Co, Ni, Cu, Zn, Ga, Y, and Zr were analyzed in medium mass resolution (m/Δm = 3000). To ensure reliability of using the glasses, we measured one granulite sample by solution ICPMS and by laser ablation ICP-MS of the fused disks; comparison between the two methods was comparable to within error.

Trace element concentrations of minerals (plagioclase, garnet, quartz) were also acquired using LA-ICPMS at Rice University. Polished 200 µm thin sections were analyzed. Depending on mineral grain size, laser spot sizes ranged from 55 to 110 µm. For quartz analyses, we adjusted the operating parameters to 50 – 80 µm spot sizes and a fluence of 9 J/cm², and adjusted the laser frequency to 20 Hz depending on signal intensity. Care was taken to avoid rutile needles present in quartz in both EPMA and LA-ICPMS analyses by checking BSE images and/or examining quartz grains optically prior to measurement. ³⁰Si (for quartz and plagioclase) and ²⁵Mg (for garnet) were used as internal standards to correct for instrumental drift. USGS basaltic glass standards BHVO2g, BIR1g, BCR2g, and the NIST612 glass were used for external calibration, following recommended values given in GEOREM. Sensitivity of the instrument ranged from 180,000 to 250,000 cps per 15 ppm La in BHVO2g for a 55 µm spot operated at 10 Hz and 9 J/cm².
Major element concentrations of minerals were determined using wavelength dispersive spectroscopy (WDS) on the CAMECA SX 50 electron microprobe at Texas A&M University. Polished 30 µm thin sections were analyzed. Garnet, quartz, and feldspar standards were used. Spot size was 1 µm, accelerating voltage 15 kV, and probe current 10 to 20 nA. Probe current was increased to 300 nA to measure titanium concentrations in quartz. For these measurements, we used the Charles M. Taylor pure rutile standard and a pure quartz standard as a baseline check. We used the trace element routine within the CAMECA software to analyze for trace Ti in quartz, and assumed stoichiometric quartz for matrix correction purposes. Due to the extensive decomposition/alteration of the plagioclase, plagioclase grains were imaged in BSE prior to WDS analyses to insure that only unaltered areas were analyzed.

U-Pb and Hf isotopes of detrital zircons were collected simultaneously using a laser ablation system (Photon Machines Analyte 193H) attached to a single collector ICP-MS (ThermoScientific Element XR) and a multi collector ICP-MS (ThermoScientific Neptune Plus) at UC Davis. After the sample is ablated, the aerosol leaves the sample cell of the laser ablation system and is split into two paths (Tollstrup et al., 2012; Xie et al., 2008), one to the ELEMENT XR for U-Pb isotopes and the second to the Neptune Plus for Hf isotopes. Detrital zircons were measured in situ directly from polished 300 µm thick sections. Due to the small size of most zircons, typical spot sizes used were 30 – 50 µm. Total data acquisition time was 76 seconds, with 25 to 30 seconds of background acquisition prior to firing the laser, followed by 46 to 51 seconds of sample acquisition during ablation. A total of 350 scans of data were acquired during the 76 second integration window, corresponding to ~0.21 seconds for each individual data point. Zircon standards 91500 (Wiedenbeck et al., 1995), GJ-1 (Jackson et al., 2004), Plesovice (Slama et al., 2008), Temora (Black et al., 2003), and B266 (Stern and Amelin, 2003) were measured prior to and after a set of 5 unknown zircons. Due to the complex zonation of almost all measured unknown zircons, the following data reduction strategy was employed (See supplemental online information for the spreadsheet program).
For U-Pb isotopes, we first subtract average background intensities from the sample intensities. Next, background-corrected ratios of intensities ($^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{238}\text{U}$, etc.) were calculated. These ratios are then corrected for downhole fractionation (drift) of Pb/U as laser ablation proceeds through the target grains of sample and standard. We used the exact spot size and laser energy for all analyses within a single standard-sample bracketed analytical session to ensure downhole fractionation of Pb/U observed for standards are applicable to unknown samples. The downhole fractionation of Pb/U was linear with time for all standards with homogeneous composition. The slope of relative Pb/U fractionation with time was thus constant to within error. This slope was defined as $dF/dt = \Delta R/(R_0 t)$, where $\Delta R$ is the change in the ratio over time $t$ since initiation of the ablation signal and $R_0$ is the ratio at the initiation of ablation. Downhole relative fractionation rates ($dF/dt$) for $^{206}\text{Pb}/^{238}\text{U}$ (that for $^{207}\text{Pb}/^{235}\text{U}$ was assumed to be the same) for each zircon standard were identical to within error. These relative fractionation rates were used to correct the time-resolved data of all unknown zircons back to a common baseline, which we took to be the initiation of the ablation signal. This was done using the following equation: $R_t = R_0 + (dF/dt)R_0 t$ where $R_t$ is the ratio measured at time $t$, $R_0$ is the ratio drift-corrected back to initiation of the ablation signal, and $t$ is the time elapsed since initiation of ablation. After correction for drift, isotopic ratios are then corrected for instrumental mass bias by normalizing to zircon standard 91500 to obtain the final background-, drift-, and mass bias-corrected ratios. For further details see Tollstrup et al. (2012).

After all corrections were made for every single laser pulse and each individual measurement (scan), we calculated uncertainties by two approaches: counting statistics on the raw data and running average of every 5 corrected data points. In the former method, we calculate the total number of counts ($N$) on each mass at each time slice using the equation $N = It_d$, where $I$ is the intensity in counts/second and $t_d$ is the total dwell time. The dwell time was 0.01 seconds for $^{204}\text{Pb}$, $^{206}\text{Pb}$, $^{208}\text{Pb}$, $^{232}\text{Th}$, and $^{238}\text{U}$ and 0.02 seconds for $^{207}\text{Pb}$. The mass window was 3% and we measured 100 samples per peak, thus yielding 3 slices per integration window. We then calculated relative standard
error (RSE) using the equation $1 / \sqrt{N}$ and then propagated the error on the ratios $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$, and $^{207}\text{Pb}/^{206}\text{Pb}$. Percent errors (2RSE) were put into ISOPLOT (Ludwig, 2003) to calculate error ellipses. In the running average method, we continuously averaged every 5 corrected measurements and calculated relative standard errors for each running average. For the counting statistics method, percent errors (1RSE) were put into ISOPLOT to calculate error ellipses. Overall, counting statistics gave lower uncertainties; the difference between the uncertainty from running averages vs. counting statistics generally did not exceed 10%. For comparison, Tollstrup et al. (2012) binned every five scans and calculated the average and associated uncertainties. In this study, we report the uncertainties calculated using counting statistics.

In some samples, the data did not define a clear discordant array, but defined instead a cluster of points. For these zircons, three separate ages were calculated: the $^{206}\text{Pb}/^{238}\text{U}$ age, the $^{207}\text{Pb}/^{235}\text{U}$ age, and the $^{207}\text{Pb}/^{206}\text{Pb}$ age.

Background-corrected Hf isotopes (background correction is done automatically by the Neptune software) were further corrected for instrumental mass bias (assuming exponential mass bias correction) and isobaric interferences of $^{176}\text{Yb}$ and $^{176}\text{Lu}$ on $^{176}\text{Hf}$ off-line using an Excel program written by Tollstrup et al. (2012).

### 3.4. Results

#### 3.4.1. Petrography of metaquartzite xenoliths

The Big Creek metasedimentary xenoliths (hereafter referred to as “quartzites”) are characterized by high quartz content (>50% by mode), equigranular textures and the presence of large garnet grains (~1 mm). We focus on four quartzites, BC98-8, BC98-6, BC78, BC134. Approximate mineral modes were determined by point counting (~1000 points) and/or by least squares inversion of mineral EPMA data and bulk rock chemistry when the appropriate data were available.
BC98-8, BC98-6, and BC78 consist of 60 to 75% quartz, 8 – 12 % garnet, and 11 – 30 % feldspar. Other phases include ≤1% of biotite, trace amounts of orthopyroxene and aluminosilicate, as well as accessory minerals such as graphite, rutile, zircon, apatite, and monazite. BC134 consists of ~20% clinopyroxene, 50% quartz, 30% feldspar. The presence of clinopyroxene and absence of garnet make BC134 distinct. We refer to the former group as the garnet quartzites and the latter as a clinopyroxene quartzite. Because the mineral assemblage in the clinopyroxene quartzite is not diverse enough to provide good constraints on temperature and pressure, we did not analyze BC134 in detail.

Garnet quartzites show signs of late-stage disruption by the host magma and extensive decomposition of feldspars into clays, probably due to post-eruption weathering. Nearly all feldspars have been decrepitated into symplectic, cryptocrystalline intergrowths of clay and relict feldspar. Fluid or melt infiltration is limited mainly to grain boundaries and cracks, with most garnet cores and quartz grains generally clean. Garnet rims contain ubiquitous fine-grained, opaque kelyphites, presumably associated with rapid decompression or localized reaction with infiltrating fluids/melts from the host magma during eruption.

Overall, the quartzites are medium to coarse-grained equigranular and non-foliated, (3 mm – 1 cm) (Figure 3-2) approaching a granoblastic fabric. Quartz grain sizes of ~1 cm or more are observed in BC98-8, wherein two large, optically continuous quartz grains comprise nearly the entire thin section (2.5 cm), engulfing smaller garnet and plagioclase grains. Such large grain sizes might be evidence for progressive grain growth at high temperatures in the absence of significant deviatoric stresses. Grain boundaries are represented by a spectrum of contact relationships, ranging from serrated quartz-quartz contacts to 120 ° triple junctions (Figures 3-2a-c), the former suggesting grain boundary migration recrystallization (Jessell, 1987; Shelley, 1993) and the latter suggesting evolution towards complete textural equilibration.
Figure 3-2. **Petrography of Sierran metaquartzite xenoliths.** A) Plane-polarized light photograph of thin section (30 μm) of BC78. B) Same section in A), but under crossed-nichols. C) Microphotograph of BC78 (crossed nichols) showing large deformed quartz grains with undulatory extinction at lower right, serrated quartz-quartz contacts, occasional 120° triple junctions, and pervasive melt/fluid infiltration around garnet and through cracks in quartz. Note the decomposition of plagioclase into cryptocrystalline clots. Abbreviations: rt: rutile, gr: graphite, qtz: quartz, pl: plagioclase, cr: cryptocrystalline phase. D) Plane-polarized light photograph of thin section (30 μm) of BC98-8. E) Same section in D), but under crossed-nichols. Note the very coarse size of optically continuous quartz grains (>1 cm) which all contain undulose extinction. F) Zoomed-in view of area outlined in E), showing large garnet porphyroblast and surrounding melt infiltration trails, aluminosilicate grain mantled by altered material, plagioclase (pl) with relict twins, and detrital rutile (rt) and zircon. G) Thick section (~200 μm) of BC98-6, showing opaque, decrepitated plagioclase blebs, garnet porphyroblasts, and quartz matrix. H) Micro-XRF map of thick section in G), where red = Fe, green = Ca and Al. Red areas correspond to garnets and green areas to plagioclase. Red trails are most likely late-stage melt infiltration, possibly from the host magma.
In all xenoliths, undulose extinction and subgrain domains are common in quartz. A weak crystallographic preferred orientation may be present based on a qualitative observation of orientation families in quartz. Plagioclase appears as subordinate ovoid blebs engulfed within the larger quartz grains. Garnet occurs as isolated round porphyroblasts. Well-rounded zircons (30–100 µm) are scattered randomly throughout quartz (also as inclusions in garnet). The sphericity and zoning of cathodoluminescence in the zircons suggest a detrital origin (Figure 3-3). The zircons have highly luminescent cores, mantled by less luminescent rims. The high luminescence is generally indicative of a former magmatic origin, whereas low luminescence is typical of metamorphic overgrowths. Some detrital cores contain non-luminescent spots and patches, which may be indicative of partial recrystallization at high temperatures (Vavra et al., 1996). Collectively, grain morphology, particularly the high sphericity, and cathodoluminescent features described above suggest a metamorphic history characterized by residence of the zircons at elevated temperatures followed by new zircon growth. High initial Zr concentrations in the protoliths (>100 ppm) may have provided the source for dissolution-reprecipitation reactions responsible for large metamorphic overgrowths (Rubatto et al., 2001), and prolonged residence at high temperatures (where Zr solubility is high; (Watson and Harrison, 1983)) may have facilitated Ostwald ripening as a mechanism of overgrowth (Nemchin et al., 2001).
Figure 3-3. Cathodoluminescence images of detrital zircons. Detrital cores are bright and luminescent, some cores have spongy textures and patches, which may be a result of high-temperature partial recrystallization. Detrital cores are rimmed by metamorphic overgrowths, the contact between core and overgrowth is often dark, grading into more luminescent.
3.4.2. Whole-rock geochemistry of metaquartzite xenoliths

3.4.2.1. Major element systematics

Whole-rock major element compositions are reported in Table 3-1. Binary oxide variation diagrams are shown in Figure 3-4a-d. All quartzites are characterized by high SiO$_2$ (>80 wt. %) and low Al$_2$O$_3$ (<6 wt. %) compared to average upper continental crust, which has 66 wt. % SiO$_2$ and 15 wt. % Al$_2$O$_3$ (Rudnick and Fountain, 1995). The quartzites are also depleted in K$_2$O (<1 wt. %) compared to upper continental crust which has 3 wt. % K$_2$O. In Figure 3-4, we also show for comparison passive margin siliciclastic rocks from the Great Basin to eastern California (Stewart, 1970). These rocks are now represented by roof pendants within the Sierra Nevada Batholith, although some units are also well exposed east of the Sierra in the White and Inyo Mountains. Compositions from the nearby Boyden Cave roof pendant (Girty et al., 1994) as well as metapelitic wallrocks from Southern Sierra Nevada roof pendants and leucosomes derived therefrom (Zeng, 2004) are also shown. Finally, we also include a suite of calc-silicate wallrocks from the Peninsular Ranges Batholith (Morton, unpublished data). All of these rocks, based on proximity and compositional similarities, represent possible protoliths of the Sierran quartzites.

| TABLE 3-1. Whole-rock major element compositions and mineral modes of Sierran metasediments |
|---------------------------------|-----|-----|-----|-----|
| wt. %                          | BC78 | BC98-6 | BC98-8 | BC134 |
| SiO$_2$                       | 84.68 | 90.29 | 87.66 | 88.22 |
| TiO$_2$                       | 0.389 | 0.314 | 0.317 | 0.179 |
| Al$_2$O$_3$                   | 4.89  | 4.64  | 5.40  | 3.46  |
| FeOt                          | 3.86  | 1.43  | 2.62  | 1.85  |
| MnO                           | 0.04  | 0.01  | 0.03  | 0.03  |
| MgO                           | 3.34  | 1.17  | 1.74  | 2.28  |
| CaO                           | 1.09  | 1.12  | 1.32  | 3.22  |
| Na$_2$O                      | 0.58  | 0.92  | 0.81  | 0.39  |
| K$_2$O                       | 0.95  | 0.08  | 0.08  | 0.25  |
| P$_2$O$_5$                    | 0.19  | 0.01  | 0.01  | 0.11  |
| Total                        | 100   | 100   | 100   | 100   |
| Difference                    | 7.70  | 2.97  | 4.68  | 3.13  |

Modes for Selected Samples

<p>| | | | | |</p>
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Difference = Normalized XRF total (100% volatile-free basis) - Unnormalized XRF total
Modes for sample BC78 were determined by point counting (~1000 points)
*Estimate of plagioclase mode includes opaque breakdown phases (thus regard as a max. bound)
\[ \Sigma(r^2) = \text{Sum of residuals squared, if mode was determined using least-squares regression of average} \]
mineral chemistries and whole-rock composition

Table 3-1. Whole-rock major element compositions and mineral modes of Sierran metaquartzites

<table>
<thead>
<tr>
<th>Mineral</th>
<th>SiO₂ (wt. %)</th>
<th>Al₂O₃ (wt. %)</th>
<th>K₂O (wt. %)</th>
<th>FeO+MgO (wt. %)</th>
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</table>

Figure 3-4. Major element systematics of Sierran metaquartzites and related rocks. Open symbols represent passive margin sediments of the western Great Basin and eastern California (Stewart, 1970), filled grey symbols represent Sierran roof pendants (Girty et al., 1994) and calc-silicate wallrocks from the neighboring Peninsular Ranges Batholith (Morton, unpublished data), filled colored symbols represent Sierran quartzite xenoliths. A) Al₂O₃ vs. SiO₂, b) K₂O vs. SiO₂, c) FeO+MgO vs. SiO₂, d) CaO vs. SiO₂. Thick red dashed lines represent trend of partial melting residues for 4 different starting protoliths (see text). Blue lines with tick marks represent binary mixing lines with pure CaCO₃ mixed with an end-member varying from pure SiO₂ to 50 wt. % SiO₂. E) Ternary plot in wt. % of MgO+FeO, Al₂O₃, and CaO. Black triangles with red outline are three metapelitic wallrocks from the Southern Sierra analyzed by Zeng (2004) and their corresponding leucosome (black triangle). Thin black dashed line shows a schematic vector of melts derived from metapelites; thick red dashed lines show complementary residues (schematic). Red star represents upper continental crust composition from Rudnick and Fountain (1995).
In binary plots of $\text{Al}_2\text{O}_3$, $\text{K}_2\text{O}$, and $\text{MgO}+\text{FeO}$ versus $\text{SiO}_2$ (Figure 3-4a-c), the metasediments and the xenoliths show negative trends corresponding to decreasing amounts of feldspar, mica, and mafic minerals with increasing quartz content, which are features of increasing sediment maturity. The Sierran garnet quartzite xenoliths, which plot on the quartz-rich end of the array, clearly derive from quartz-rich protoliths and not more aluminous (pelitic) protoliths. In Figure 3-4d, we also plot $\text{CaO}$ vs. $\text{SiO}_2$ along with binary mixing arrays between pure calcite and an end-member varying from pure quartz to 50 wt.% quartz in order to assess the extent to which the protolith may have contained carbonate. All the garnet quartzite xenoliths and most of the siliciclastic roof pendants are highly depleted in $\text{CaO}$ and show no correlation between $\text{CaO}$ and $\text{SiO}_2$. However, a few roof pendant suites and the Peninsular Ranges wallrocks are enriched in $\text{CaO}$, suggesting mixing with calcite. The clinopyroxene quartzite, BC134, has higher $\text{CaO}$ (3 wt. %) compared to the other xenoliths, suggesting a calcareous quartzite as a protolith. Figure 3-4e shows a ternary plot of $\text{MgO}+\text{FeO}$, $\text{Al}_2\text{O}_3$, and $\text{CaO}$, which provides some additional insights to binary diagrams. The Sierran quartzites do not fall within the fields bounded by passive margin sediments or their metamorphic analogues, the Sierran roof pendants. The quartzites trend towards higher FeO and MgO contents, an observation also borne out by the steeper array on a binary plot of FeO+MgO vs. $\text{SiO}_2$ (Figure 3-4c), compared to passive margin and roof pendant rocks. In a later section, we discuss whether these trends may be the result of partial melting.

3.4.2.2. Trace element geochemistry

Whole rock trace element geochemistry is reported in Table 3-2 and primitive mantle-normalized (McDonough and Sun, 1995) trace element abundance diagrams are shown in Figure 3-5. Average upper continental crust (UCC) of Rudnick and Fountain (1995) is also shown for comparison. The rare-earth element (REE) abundance patterns show relative enrichment in light REEs relative to the heavy REEs and are generally similar in shape to UCC and Sierran metasandstones from the nearby Boyden Cave roof pendant (Girty et al., 1994), a potential protolith. However, in absolute abundance, they are depleted in REEs. This depletion in REEs correlates negatively with $\text{SiO}_2$ and
positively with Al$_2$O$_3$, indicating that the depletion in REEs is related to dilution by quartz content. Zr and Hf concentrations are highly variable as shown by the positive and negative depletions relative to Sm and Nd in normalized abundance diagrams. This variability is likely controlled by the abundance of zircon, as exemplified by the fact that BC98-8 has the most positive Zr-Hf anomaly and the highest observed zircon abundance. In contrast, BC134, which lacks zircon altogether, has a negative Zr-Hf anomaly.

<table>
<thead>
<tr>
<th>ppm</th>
<th>BC78</th>
<th>BC98-8</th>
<th>BC134</th>
</tr>
</thead>
<tbody>
<tr>
<td>P</td>
<td>MR-LA-ICPMS</td>
<td>645</td>
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Table 3-2. Whole-rock trace element geochemistry of metaquartzites

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<th>LR-LA-ICPMS</th>
<th>LR-LA-ICPMS</th>
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<td>0.193</td>
<td>0.150</td>
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<td>8.58</td>
<td>0.677</td>
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<td>0.369</td>
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<td>0.829</td>
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<tr>
<td>U</td>
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<td>0.305</td>
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</table>

Figure 3-5. Whole-rock trace element spidergrams of metaquartzites. Values are normalized to primitive mantle of McDonough and Sun (1995). Upper continental crust (UCC) of Rudnick and Fountain (1995) is shown for reference.

3.4.3. Mineral chemistry of metaquartzites

3.4.3.1. Major element compositions of garnet and feldspar

Major element compositions of garnet and clean, relict plagioclase are reported in Table 3-3. Garnets are overall unzoned, although rims are slightly Ca-enriched (4 wt. % vs. 3.5 wt.% in cores). Almost all garnet rims have been altered by contact with infiltrates of the host basaltic melt, thus the observed Ca zonation is most likely not an inherent feature of the garnet. Average end-member compositions of all garnets in the two
samples analyzed (BC98-8, BC78) are nearly identical at Gr$_{10}$Py$_{40}$Al$_{50}$Sp$_{0.8}$. Mg# of garnets are between 42 to 48.

**TABLE 3-3. Mineral compositions of Sierran metasediments**

<table>
<thead>
<tr>
<th>Average feldspar compositions</th>
<th>BC98-8 plag</th>
<th>BC78 plag</th>
<th>BC78 kfs</th>
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<td>n=11</td>
<td>n=3</td>
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<tr>
<td>SiO$_2$</td>
<td>63.27</td>
<td>58.73</td>
<td>65.36</td>
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<tr>
<td>TiO$_2$</td>
<td>23.69</td>
<td>26.04</td>
<td>19.68</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>8.09</td>
<td>0.28</td>
<td>0.36</td>
</tr>
<tr>
<td>Cr$_2$O$_3$</td>
<td>0.01</td>
<td>0.04</td>
<td>0.03</td>
</tr>
<tr>
<td>FeOT</td>
<td>6.87</td>
<td>8.91</td>
<td>0.63</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.09</td>
<td>0.03</td>
<td>3.58</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.33</td>
<td>0.09</td>
<td>10.97</td>
</tr>
<tr>
<td>BaO</td>
<td>0.04</td>
<td>0.38</td>
<td>0.21</td>
</tr>
<tr>
<td>Total</td>
<td>100.67</td>
<td>100.17</td>
<td>100.83</td>
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<table>
<thead>
<tr>
<th>End-member mol %</th>
<th>ab</th>
<th>or</th>
<th>an</th>
<th>cs</th>
</tr>
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<td>25.62</td>
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<td>Total</td>
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<td>100.96</td>
<td>101.79</td>
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<th>BC98-8 Rim</th>
<th>BC78 Core</th>
<th>BC78 Rim</th>
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<tr>
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<td>63.27</td>
<td>65.36</td>
<td>63.27</td>
<td>65.36</td>
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<tr>
<td>TiO$_2$</td>
<td>23.69</td>
<td>19.68</td>
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<td>19.68</td>
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<td>Al$_2$O$_3$</td>
<td>8.09</td>
<td>3.58</td>
<td>8.09</td>
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<td>Cr$_2$O$_3$</td>
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<td>0.03</td>
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<td>0.63</td>
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<td>K$_2$O</td>
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<td>0.33</td>
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<td>0.21</td>
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<tr>
<td>Total</td>
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<td>100.17</td>
<td>100.13</td>
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<table>
<thead>
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<th>End-member mol %</th>
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<th>py</th>
<th>sp</th>
<th>al</th>
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<td>38.31</td>
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plag = plagioclase, kfs = alkali feldspar, ab = albite, or = orthoclase, an = anorthite, cs = celsian
gr = grossular, py = pyrope, sp = spessartine, al = almandine
Mg# defined as (Mg/(Mg+Fe))*100

Table 3-3. Major element compositions of feldspar and garnet
Although most feldspars decomposed or were altered to clays and cryptocrystalline phases, we were able to find areas that still preserved relict compositions by screening with back-scattered electron imaging and checking the compositions using energy-disperse wavelength (EDS) spectroscopy. BC98-8 contained only plagioclase and no alkali feldspar (average end-member composition $Ab_{72}Or_{2}An_{26}$). BC78 contained mostly plagioclase (average $Ab_{47}Or_{5}An_{48}$) with minor alkali feldspar. In most cases, it was difficult or impossible to obtain rim-core traverses due to the degree of alteration, so we analyzed 3 – 4 spots throughout individual grains to determine if compositional variability was present. In general, plagioclases were unzoned.

### 3.4.3.2. Trace element geochemistry of garnet, plagioclase, and quartz

Trace element compositions of garnet, feldspar, and quartz are reported in Tables 3-4 and 3-5. Garnet cores in BC98-8 contain higher HREE abundances than corresponding rims, which might be related to growth zoning. No variation in HREE composition was observed in garnet in BC78. An interesting feature of both garnet and plagioclase is that, within a given sample, both minerals show complementary Eu anomalies (garnet a negative Eu anomaly and plagioclase a positive Eu anomaly). The complementary Eu anomaly in garnet and plagioclase indicates that garnet equilibrated with plagioclase. Average ($n = 23$) Ti concentrations in quartz for BC78 is 165 ppm; average Ti is 85 ppm for BC98-8, and average Ti for BC134 is 44 ppm.

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**Note**
- Feldspar in BC98-8

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Table 3-4. Mineral trace element compositions of metaquartzites

Table 3-5. Titanium concentrations and temperatures in quartz
### Table 2-6: U-Pb and Hf isotope geochemistry of detrital zircons

| Grain | Sample  | \( ^{176} \text{Yb} / ^{177} \text{Hf} \) | 2SE | \( ^{176} \text{Lu} / ^{177} \text{Hf} \) | 2SE | \( ^{176} \text{Hf} / ^{177} \text{Hf} \) | 2SE | Upper Intercept | 2SE | Lower Intercept | 2SE | \( ^{206} \text{Pb} / ^{238} \text{U} \) | 2SE | \( ^{207} \text{Pb} / ^{235} \text{U} \) | 2SE | \( ^{207} \text{Pb} / ^{206} \text{Pb} \) | 2SE | Age (Ma) for clustered zircons |
|-------|---------|---------------------------------|-----|---------------------------------|-----|---------------------------------|-----|------------------|-----|------------------|-----|------------------|-----|------------------|-----|------------------|
| 1     | BC98-8-1 | 0.00751 | 0.00000 | 0.00004 | 0.00001 | 0.210099 | 0.000038 | 263 | 34 | 116 | 28 | 16.6 | 0.16 | 4.1 | -30.1 |
| 2     | BC98-8-2 | 0.01200 | 0.00039 | 0.00040 | 0.00004 | 0.21527 | 0.00060 | 2706 | 290 | 106 | 25 | 16.6 | 0.16 | 40.9 |
| 3     | BC98-8-3 | 0.01482 | 0.00030 | 0.00036 | 0.00011 | 0.20175 | 0.00037 | 217 | 250 | 120 | 16 | -7.2 | 0.14 | 44.3 |
| 4     | BC98-8-4 | 0.00581 | 0.00102 | 0.00078 | 0.00047 | 0.20809 | 0.00010 | 2966 | 260 | 110 | 52 | -4.2 | 0.04 | 64.5 |
| 5     | BC98-8-6 | 0.01081 | 0.00012 | 0.00076 | 0.00007 | 0.20160 | 0.00084 | 3001 | 140 | 189 | 40 | 9.5 | 0.05 | 52.5 |
| 6     | BC98-8-1 | 0.01397 | 0.00039 | 0.00024 | 0.00017 | 0.21094 | 0.00054 | 277 | 95 | 101 | 31 | 7.0 | 0.20 | 50.1 |
| 7     | BC98-8-2 | 0.00643 | 0.00111 | 0.00073 | 0.00020 | 0.21492 | 0.00035 | 210 | 85 | 90 | 9.7 | 0.8 | 0.43 | 45.8 |
| 8     | BC98-8-3 | 0.01177 | 0.00094 | 0.00056 | 0.00012 | 0.21247 | 0.00097 | 2096 | 98 | 71 | 20 | 7.6 | 0.49 | 49.5 |
| 9     | BC98-8-4 | 0.00908 | 0.00065 | 0.00018 | 0.00008 | 0.21597 | 0.00067 | 1765 | 52 | 108 | 24 | -5.6 | 0.29 | 59.7 |
| 10    | BC98-8-5 | 0.02561 | 0.00068 | 0.00009 | 0.00021 | 0.21716 | 0.00002 | 1763 | 43 | 99 | 19 | -14 | 0.35 | 55.6 |
| 11    | BC98-8-6* | 0.02153 | 0.00103 | 0.00034 | 0.00005 | 0.21698 | 0.00009 | 97.9 | 35 | 92.9 | 0.13 | -46 | 0.05 | 120 |
| 12    | BC98-9-1 | 0.00863 | 0.00164 | 0.00016 | 0.00008 | 0.20237 | 0.00035 | 819 | 67 | 119 | 64 | 9.9 | 0.46 | 16.9 |
| 13    | BC78-2-2 | 0.00534 | 0.00111 | 0.00102 | 0.00035 | 0.21257 | 0.00097 | 963 | 140 | 95 | 68 | 2.2 | 0.16 | 16.6 |
| 14    | BC98-2-3 | 0.00942 | 0.00134 | 0.00132 | 0.00066 | 0.21594 | 0.00052 | 998 | 140 | 91 | 57 | 3.5 | 0.25 | 45.4 |
| 15    | BC98-2-3* | 0.00612 | 0.00120 | 0.00145 | 0.00046 | 0.21280 | 0.00034 | 122 | 118 | 113 | 113 | 0.45 | 150 | 170 |
| 16    | BC98-7-2 | 0.11833 | 0.00135 | 0.00240 | 0.00051 | 0.22845 | 0.00085 | 122 | 120 | 99 | 18 | 13.0 | 0.12 | 275 | 19 |
| 17    | BC78-1-2 | 0.00639 | 0.00018 | 0.00033 | 0.00017 | 0.21208 | 0.00020 | 206 | 100 | 120 | 40 | 5.6 | 0.20 | 26.4 |
| 18    | BC78-1-3 | 0.00626 | 0.00050 | 0.00141 | 0.00012 | 0.21569 | 0.00026 | 857 | 350 | 106 | 7.9 | 10.4 | 0.19 | 39.5 |
| 19    | BC78-1-4 | 0.00887 | 0.00098 | 0.00088 | 0.00005 | 0.21834 | 0.00098 | 2843 | 86 | 66 | 50 | 10.2 | 0.45 | 65.6 |
| 20    | BC78-1-5 | 0.00508 | 0.00102 | 0.00122 | 0.00054 | 0.21519 | 0.00021 | 2777 | 90 | 126 | 95 | 7.1 | 0.46 | 46.8 |
| 21    | BC98-2-8* | 0.00315 | 0.00015 | 0.00052 | 0.00004 | 0.20216 | 0.00079 | 123 | 0.51 | 130 | 0.26 | 583 | 120 |

*Denotes "clustered" zircons: zircons which did not show a linear array in a plot of \( ^{206} \text{Pb} / ^{238} \text{U} \) vs. \( ^{207} \text{Pb} / ^{235} \text{U} \), and thus did not preserve distinct upper and lower intercepts.

Table 3-6: U-Pb and Hf isotope data of detrital zircons
3.4.4. U-Pb and Hf isotope geochemistry of metaquartzites

Lu-Hf and U-Pb isotopic data for zircons are reported in Table 3-6. In Figure 3-6, we show 6 representative detrital zircon analyses. Each individual panel of the figure corresponds to one zircon grain, and each error ellipse corresponds to one measurement from a single laser pulse. As shown in Figure 3-6, the zircons are all discordant with variable but ancient (>1 Ga) upper intercept ages, but a common lower intercept age centered around a weighted mean of 103 ± 10 Ma (2σ) (Figure 3-7). We interpret the discordia to represent a mixture line between old detrital cores (upper intercept) and a younger metamorphic overgrowth (lower intercept).

In Figure 3-8, we plot εHf values vs. age for the zircons. Despite having two distinct U-Pb isotopic signatures corresponding to an upper and lower intercept age, Hf isotopes were homogeneous within individual zircon grains. For each zircon, we plot εHf$_i$, using the upper intercept age, and “εHf$_0$”, using the lower intercept age. Colored arrows are drawn to schematically connect the detrital zircon core (εHf$_i$) to corresponding metamorphic overgrowth rims (εHf$_0$). The following values were used in calculating εHf and plotting the depleted mantle evolution line: (\(^{176}\text{Hf}/^{177}\text{Hf}\)\text{CHUR,present}=0.282785 and (\(^{176}\text{Lu}/^{177}\text{Hf}\)\text{CHUR,present}=0.0336 (Bouvier et al., 2008); (\(^{176}\text{Lu}/^{177}\text{Hf}\)\text{DM,present}=0.03933 and (\(^{176}\text{Hf}/^{177}\text{Hf}\)\text{DM,present}=0.283294 (Vervoort and Blichert-Toft, 1999).
Figure 3-6. Representative U-Pb concordia diagrams showing typical discordia of detrital zircons. Each panel represents one zircon (>200 individual measurements per analysis). Data point ellipses are 2σ as calculated by Isoplot. Uncertainties reported do not include uncertainty on the decay constants of $^{238}$U and $^{235}$U.
Figure 3-7. Plot of lower intercept ages (Ma) of individual zircon grains. Symbols correspond to host xenolith; each point = one zircon grain. Black line represents weighted mean of lower intercepts at 103 ± 10 Ma, grey band represents uncertainty (2σ) on the weighted mean.

Figure 3-8. εHf vs. age (Ma) of zircons. Each zircon grain is represented by two points on the diagram – the black outlined point corresponds to the upper intercept age, the colored outlined point corresponds to the lower intercept age determined from U-Pb discordia. Black arrows are schematic and illustrate initial zircon crystallization and subsequent isotopic evolution (note that the zircons' $^{176}$Hf/$^{177}$Hf ratio does not change as evidenced by the homogeneous values obtained for individual...
zircon grains; rather the evolution towards very negative εHf values is largely driven by the growth of the CHUR reservoir from 2 Ga to 100 Ma).

3.5. Final P-T conditions of metaquartzites

3.5.1. Titanium-in-quartz and garnet-aluminosilicate-quartz-plagioclase geothermobarometry

To place constraints on the pressure and temperature conditions at which the quartzites last equilibrated, we combined geothermobarometry and thermodynamic modeling. Rutile coexists with quartz in all samples, simplifying activity relations so that $a\text{TiO}_2$ in rutile = 1 and thus the equilibrium constant $K \sim a\text{TiO}_2$ in quartz when using the titanium-in-quartz geothermometer (Wark and Watson, 2006). Ti-based temperatures are between 730 and 810 °C for garnet quartzite xenoliths BC98-8 and BC78 (corresponding to the average of 85 and 165 ppm Ti, respectively). Because Ti concentrations measured by EPMA were systematically lower than by LA ICP-MS (EPMA was on average 30 – 50 ppm less than ICP-MS), EPMA data gave slightly lower temperatures (50 – 70 °C) than LA ICP-MS data. The discrepancies between Ti measured by LA ICP-MS and EPMA could be related to the larger spot size ($\geq 30 \mu$m) used in laser ablation compared to that used in EPMA (1 µm), resulting in the incorporation of rutile needles into the analysis during laser ablation. If the rutile needles are of exsolution origin, the laser ablation results provide a better measure of the peak metamorphic temperatures, but if they are not of exsolution origin, the EPMA results are likely to be more representative. We are not able to further resolve this issue.

For final pressures of equilibration, we used the “GASP” geobarometer (garnet-aluminosilicate-silica-plagioclase) (Koziol and Newton, 1988; Spear, 1995). Due to the difficulty of obtaining unaltered rim-rim mineral pairs between plagioclase and garnet, we used average mineral compositions (this is justified because the minerals are mostly unzoned). Results are shown in Figure 3-9a. For input temperature of 730°C from Ti-in-quartz temperatures, average albite content of 71% in plagioclase, average grossular
content of 9% in garnet, and activities of quartz and aluminosilicate of 1, we obtain a pressure of 1.1 GPa for sample BC98-8. We obtain the same pressure, within error, for sample BC78.

Figure 3-9. P-T diagram showing final equilibration of metaquartzites. Colored symbols correspond to LA ICP-MS measurements of Ti in quartz (used as input temperatures into the GASP barometer), whereas white symbols refer to EPMA measurements of Ti used as inputs. Green field represents feldspathic granulite xenoliths, shown for comparison, from the same locality measured by Ducea & Saleeby (1996). Thin black curves represent a family of crustal geotherms constrained by surface heat flow ranging from 40 to 90 mW/m² (assuming a heat production of 0.7 μW/m³ (Rudnick et al., 1998) and a 35 km thick crust). Wet pelite solidus and mica dehydration curves are for the model system KNFASH (K, Na, Fe, Al, H₂O and quartz-saturated) and are taken from (Bucher and Frey, 1994). B) Isothermal P-X diagram at 750 °C, calculated using THERIAK DOMINO (de Capitani and Petrakakis, 2010) for bulk composition of sample BC98-8. Only Si, Ca, Al, K, Mg, and Fe (no H₂O) was modeled. Cord = cordierite, Opx = orthopyroxene, Plag = plagioclase, Kfs = alkali feldspar, Ky = kyanite.

3.5.2. Stability field of plagioclase + garnet

To further constrain metamorphic P-T conditions, we determined the stability field wherein plagioclase and garnet coexist for a given quartzite bulk composition using the software THERIAK DOMINO to calculate equilibrium assemblage diagrams (de Capitani and Petrakakis, 2010). Using sample BC98-8, and considering only Si, Ca, Al, K, Mg, Fe (no H₂O) in the system, we calculated the volume % of stable minerals in the
assemblage at fixed T and varying P. We show an isothermal section at 750 °C in Figure 3-9b. The plagioclase + garnet stability field spans a wide interval from 0.4 to >1 GPa, but based on the estimated modes in BC98-8, we narrow this window to 0.7 to 1.0 GPa. The modeled modes of garnet and plagioclase calculated for the range 0.7 to 1.0 GPa roughly match the observed modes (see sample descriptions). Despite the discrepancies between the model and the actual rock, overall the P-T interval in which garnet and plagioclase are stable overlaps the P-T interval determined using geothermobarometry.

3.6. Discussion

3.6.1. A passive margin origin for the Sierran lower crustal metaquartzites

The Sierran quartzites record last equilibration at lower crustal pressures (0.7 to 1.0 GPa) and relatively high temperatures (730 to 810 °C). High temperatures are consistent with the rock textures. Coarsening of quartz grains to >1 cm (Figure 3-2), engulfment of plagioclase and garnet by quartz, and the deformed nature of the quartz matrix suggest static and/or dynamic recrystallization of an originally equigranular quartzitic protolith under high temperatures, where diffusivities are high. The recrystallized fabric and significant grain growth could have formed after emplacement into the lower crust if temperatures were high and deviatoric stresses low, thus permitting grain coarsening. Metamorphic overgrowths on detrital zircons, some of which are thick compared to the diameter of the detrital core (Figure 3-3), may also have formed via grain coarsening or Ostwald ripening due to sustained high temperatures.

Despite such elevated final P-T conditions, the quartzite xenoliths do not resemble typical granulite xenoliths characteristic of the lower crust, which are intermediate (~50 wt. % SiO₂) to mafic in composition (Rudnick and Fountain, 1995). Instead, the high bulk SiO₂ contents and correspondingly high quartz modes, combined with the abundance of rounded zircons preserving detrital cores strongly point to a sedimentary origin. The detrital zircon cores yield U-Pb crystallization ages representative of North American basement, indicating that the quartzites’ protoliths are
mature, siliciclastic sediments of the Proterozoic and Paleozoic passive margin of western North America (Figure 3-1), deposited prior to Mesozoic arc magmatism. The U-Pb crystallization ages for the detrital cores overlap those of well-known passive margin rocks of the Salinian Block and Snow Lake Pendant (Barbeau et al., 2005; Grasse et al., 2001), which form part of the sedimentary framework rocks into which the Sierra Nevada batholith intruded. The extremely low εHf₀ values also corroborate the origin of the detrital zircons from Proterozoic crustal reservoirs instead of recently accreted oceanic and juvenile arc terrane sources. Furthermore, the Hf isotope systematics are similar to those of basement underlying the eastern Sierra Nevada batholith (Lackey et al., 2012), which is predominantly composed of ancient North American crust. Thus, the protoliths of the Sierran quartzites are unlikely to be juvenile sediments in the oceanic trench and accretionary complex of the Mesozoic subduction zone.

3.6.2. Mid-Cretaceous timing of metamorphic overprinting

The timing of metamorphic overprinting can be interpreted using the discordant U-Pb zircon arrays (Figure 3-6). The discordant arrays are unlikely to represent continuous Pb loss because the data do not show the curvature expected for continuous Pb loss (Tilton, 1960). The fact that the zircons in almost all of the samples share a common lower intercept of 103 ± 10 Ma, despite having variable detrital core ages, suggests that all of the zircons share a common metamorphic overgrowth, and therefore together experienced a major thermal event at this time. Th/U ratios in zircon overgrowths are generally < 0.2, also consistent with a metamorphic origin (Hoskin and Black, 2000). Intriguingly, this common lower intercept of 103 ± 10 Ma coincides with the average U-Pb zircon crystallization age of 102 Ma of plutons near the xenolith locality (Stern et al., 1981), suggesting that the metamorphic overprint is probably related to a thermal overprint induced by mid-Cretaceous arc magmatism.
3.6.3. Mechanisms of emplacement

We now consider three possible scenarios by which passive margin sediments could have been transported to lower crustal depths >20 km and temperatures in excess of 700 °C (Figure 3-9): 1) Underthrusting of continental crust beneath the continental arc, 2) viscous return flow of wallrock in response to diapiric rise of plutons, and 3) underplating by buoyant sediment diapirs derived from the subducting Farallon plate.

Hypothesis 1: Continental underthrusting beneath the arc

In the first hypothesis, North American continental lithosphere, loaded with thick Paleozoic and older passive margin sediments, is thrust westward beneath the continental arc (Figure 3-10a). The average metamorphic overprint age of 103 ± 10 Ma provides a lower bound on when underthrusting happened. The question is whether underthrusting occurred well before or coevally with mid-Cretaceous arc magmatism.

If underthrusting of these passive margin sediments to lower crustal depths occurred before Cretaceous magmatism, older metamorphic overprints or more complex single-grain zircon U-Pb discordia would be expected, but no such features are seen. The simplicity of the U-Pb systematics (common lower intercept and linear Discordia) within individual zircon grains instead are better explained by one thermal overprint, which suggests that if tectonic underthrusting is a viable hypothesis, underthrusting was likely coeval with Cretaceous arc magmatism. Such a scenario would be consistent with current views on the tectonic history of the region, which we discuss below.

Westward underthrusting of the North American continent is thought to have occurred to balance the eastward transport of upper crustal rocks into the foreland by thin-skinned deformation during the Cretaceous and is now manifested as the Cordilleran fold-and-thrust belts (DeCelles et al., 2009; Oldow et al., 1990). The role of deep, lithospheric-scale underthrusting has been inferred from Nd isotopic shifts in arc magmas resulting from emplacement of old, evolved continental material beneath the arc (Ducea and Barton, 2007). At greater depths within the arc lithosphere, Sierran mantle xenoliths
record evidence of downward compression of initially shallow, melt-depleted spinel peridotite residues which were thickened and metamorphosed into garnet lherzolite (Chin et al., 2012). These studies indicate that continental arc lithosphere at all levels records the evidence of tectonic compression and downward transport of shallow material to depth. Underthrusting is thus one possible way in which the Sierran quartzites could have been transported from their shallow origins into the lower crust.

**Hypothesis 2: Viscous return flow of wallrocks during pluton ascent**

A second hypothesis involves the downward transport of upper crustal rocks by viscous return flow in response to diapirc rise of plutons (Figure 3-10b) (Saleeby et al., 2003). This process, however, would have to displace upper crustal sediments (maximum burial depths of 6 km have been inferred by Stewart (1970)) to lower crustal depths, requiring at least 10 km of downward transport. Such extensive downward transport of the quartzites would, however, have to compensate for the very low density of quartzites, which would have generated positive buoyancy opposing the forced downflow. Another issue is that wallrocks are probably highly viscous owing to their felsic composition and cold ambient temperatures compared to ascending and evolving magmas, which are also viscous. Thus, high viscosities are another barrier to wallrock return-flow. Furthermore, pluton ascent is widely thought to occur by diking rather than diapirism (Petford, 1996). Dike ascent does not involve viscous downflow of the country rock.

**Hypothesis 3: Underplating by slab-derived sediment diapirs**

A third scenario invokes underplating of the continent by buoyant sediment diapirs derived from the downgoing Farallon oceanic plate (Figure 3-10c). Recent studies (Behn et al., 2011; Hacker et al., 2011) have proposed that such a process could be important for transporting felsic rocks, aided in part by partial melting of sediments on the subducting oceanic plate, into the lower crust at continental arcs. However, the Proterozoic to Archean provenances of the detrital zircons strongly suggest that the protoliths derive from passive margin sediments on the North American continent. Any
sediments derived from the oceanic trench would most likely consist of a large fraction of Mesozoic protoliths, which is not seen in our zircons.

![Figure 3-10. Cartoons illustrating transport mechanisms of sediments into lower crust.](image)

Dashed line represents a hypothetical deep crustal “hot zone” or “MASH” zone associated with arc magmatism. Thick red arrow represents mantle-derived flux of basaltic magma. Thin blue arrows represent devolatization of the slab. A) Continental underthrusting of upper plate continental lithosphere beneath arc. Passive margin sediments represented by yellow. B) Return flow of wallrocks during pluton diapirism. C) Underplating of lower crust by sediment diapirs derived from subducting slab. Black shape represents sediments rising from slab surface.

### 3.6.4. Possibility of partial melting

The high temperatures of the quartzite xenoliths (700-800 °C) are higher than the wet solidus of granitic rocks, begging the question of whether the quartzites may represent restites of partial melting. Based on our discussions in Section 7.3, our preferred hypothesis for the emplacement of quartzites into the lower crust is that they were underthrust during the peak of arc magmatism. If so, such melting would have profoundly influenced deep-seated magmas.

To investigate whether the quartzites represent partial melting residues, we performed melting calculations using rhyoliteMELTS (Gualda et al., 2012) on four possible protoliths, chosen to span a range of crustal metasedimentary rocks. Three protoliths are roof pendant rocks from Boyden Cave Pendant (Girty et al., 1994), varying from impure quartzite (90% SiO₂), feldspar-bearing quartzite (73% SiO₂), to pelite (56% SiO₂). The fourth protolith is feldspathic arenite (82% SiO₂) of the Wood Canyon Formation (Stewart, 1970). Full compositional data on the protoliths and melting...
products are given in the Supplemental Online Information. We modeled partial melting at isobaric conditions (1 GPa) over temperatures between 500 and 1200 °C and fO₂ buffered at quartz-fayalite-magnetite. Water contents in the protoliths were 2 % for the impure quartzite and the sandstone, 3% for the feldspar-bearing quartzite, and 4% for the pelite. Details of the inputs and outputs to the modeling are provided in the Supplemental Online Information.

In Figures 3-4a-c, we show the residue trends between 500 and 900 °C for the 4 protoliths as dashed red arrows. The most important feature is that protoliths with <70 wt. % SiO₂, yield quartz-poor residues, whereas protoliths with >70 wt. % SiO₂ yield quartz-rich residues. The kink in Al₂O₃ in the melting residue of the pelite protolith corresponds to exhaustion of micas and a steep increase in residual garnet at 765 °C. In the plot of MgO+FeO versus SiO₂ (Figure 3-4b) the evolution toward garnet-rich residues (>10 % garnet by mode) is most pronounced in the pelitic and feldspar-bearing quartzite protoliths (<70 wt. % SiO₂), which have the highest initial MgO, FeO and Al₂O₃ contents. By contrast, garnet mode in protoliths with >70 % initial SiO₂ evolve <10 % modal garnet in their residues. In Figure 3-4c, K₂O decreases in all residues, due to melting out of mica (in pelitic protoliths) or feldspars (in feldspar-bearing quartzite, feldspathic arenite, and impure quartzite).

Based on the model results, the quartzite xenoliths are clearly not residues of partial melting of pelites, which would have too little quartz and too much garnet compared to the xenoliths. Instead, the protoliths of the quartzite xenoliths must initially have had >70 wt. % SiO₂. We can further rule out a feldspathic quartzite protolith because residues of this protolith would have >10% residual garnet, but the observed garnet mode in the quartzite xenoliths is <12 %. For these reasons, the protoliths of the xenoliths were most likely impure quartzites. Given that the quartzite xenoliths contain no micas and yet have small amounts of garnet and feldspar, minimum melting temperatures of 750 °C at 1 GPa. Our model results suggest that if the quartzite xenoliths represent restites, at most 25% melting had occurred because of the relatively infertile nature of quartzite protoliths. Very high temperatures (>1300 °C at 1 GPa) are required
to generate more than 50% melt from such an “infertile” protolith (Clemens and Vielzeuf, 1987). For comparison, pelites yield 70% melt at ~750 °C according to our model results. The contrast in fertility between pelites and quartzites has been observed in Alpine settings such as the Trois Seigneurs Massif (Wickham, 1987), where pelites have undergone extensive partial melting and quartzites, under the same conditions remain, unmelted.

In summary, it seems likely that the quartzite xenoliths are the residues of small to moderate degrees of partial melting in the lower crust. Partial melting, to the point of mica exhaustion at pressures of 1 GPa, suggest melting temperatures of 750 °C, which coincidentally is within the range of the Ti-in-quartz thermometry results. Attaining such high temperatures by tectonic underthrusting alone would seem difficult. England and Thompson (1986) showed that rapid crustal thickening results in decreased thermal gradient. An external heat source seems required. One possibility could be the intrusion of mafic magmas into the lower crust, generating deep crustal hot zones (Annen et al., 2006; Hildreth and Moorbath, 1988). As shown in Figure 3-9a, crustal geotherms bounded by a range of surface heat flows (40 to 80 mW/m²) strongly suggest that heat flow ≥ 70 mW/m² is required to explain the temperatures recorded by the Sierran quartzites (as well as other crustal granulite-facies xenoliths from the Sierra Nevada from work by Ducea and Saleeby (1996)). Such high heat flows are characteristic of active continental arcs, such as the modern Andes (Springer and Forster, 1998).

3.7. Summary and implications

We studied the petrology, geochemistry, and metamorphic history of garnet-bearing quartzite xenoliths sampled in late Miocene basalts erupted through the Sierra Nevada. High quartz mode (>50 %) and abundant well-rounded detrital zircons support a sedimentary protolith, such as a mature quartz arenite. Metamorphic garnet, high Ti concentrations in quartz, and petrographic textures indicating deformation, recrystallization, and residence at elevated P-T conditions (undulose extinction in quartz, subgrains, serrated grain boundaries, and grain coarsening) support final equilibration in
the deep crust at unusually high temperatures. U-Pb and Hf isotopes in detrital zircons yield discordia wherein U-Pb upper intercepts represent variable crystallization ages indicative of ancient (>1 Ga) North American basement, but a common lower intercept at 103 ± 10 Ma, coinciding with peak arc magmatism in the Western US Cordillera.

We propose three hypotheses by which the quartzites were transported into the lower crust. The first involves underthrusting of continental crust beneath the arc. Detrital zircon provenances are consistent with origin of the quartzites as passive margin sediments deposited in the upper continental plate. The common lower intercept coincident with peak arc magmatism indicates that the quartzites were within the lower crust by that time, but we cannot further resolve whether underthrusting was pre-magmatic or syn-magmatic. The second hypothesis involves downwelling of wallrocks in response to pluton diapirism, but the amount of displacement needed, insufficient density contrast, and debate concerning diapirism or diking as the dominant pluton ascent mechanism add uncertainty to this scenario. Finally, the third hypothesis considers felsic relamination of the lower crust by buoyant sediment diapirs derived from the downgoing slab. The ubiquity of passive margin provenances rather than “active margin” (trench, accretionary prism) sources in the detrital zircons is difficult to reconcile with this hypothesis.

Regardless of what mechanism transported the quartzites beneath the arc, the fact that they are present in what presumably is a predominantly mafic lower crust has significant implications for the compositional and tectonic history of the continental crust. The Sierran quartzites are evidence that the lower crust may be more felsic than previously thought. Melting models suggest that the quartzites could be residues of partial melting of impure quartzitic protoliths. Even under water-saturated conditions, such protoliths yield far less melt compared to pelitic lithologies at the same temperature. Thus, the refractory and infertile nature of quartzites could allow them to persist in the lower crust. Currently, the abundance of felsic rocks in the lower crust is poorly constrained, although a recent analysis by Hacker et al. (2011) suggested >10-20 % of pelites in the lower crust might be permissible.
Lastly, if the Sierran quartzites represent underthrusted continental upper plate, this has implications for tectonic processes operating in continental arcs. One implication is that underthrusting thickens the continental lithosphere and affects its overall composition if felsic upper crustal rocks are transported into the deep crust. Thickening and underthrusting might be facilitated during periods of intensified arc magmatism, as higher temperatures associated with deep crustal melting could potentially trigger weakening of the lower crust and enhance lower crustal decollements. Thus, arc magmatism and tectonic thickening may go hand in hand in continental arcs.
Chapter 4

Thickening, refertilization, and the deep lithosphere filter in continental arcs: constraints from major and trace elements and oxygen isotopes

This chapter is currently in revision for Earth and Planetary Science Letters

Arc magmatism is a complex process involving generation of primary melts in the mantle wedge and chemical refinement of these melts into differentiated products akin to continental crust. Interaction of magmas (cooling, crystallization and assimilation) with the overlying crust, particularly if it is thick, is one way by which primary basalts are refined into more evolved compositions. Here, we explore the role of the mantle lithosphere as a trap and/or reactive filter of magmas. We use mantle xenoliths from the Sierra Nevada continental arc in California as a probe into sub-Moho processes. Based
on clinopyroxene modal abundance and major, minor and moderately incompatible trace element concentrations, the peridotites define a refertilization trend that increases with depth, grading from clinopyroxene-poor (<5%), undeformed spinel peridotites equilibrated at <3 GPa (<90 km) to clinopyroxene-rich (10 – 20%), porphyroclastic garnet peridotites equilibrated between 3 – 3.5 GPa (90-105 km), the latter presumably approaching the top of the subducting slab. The petrology and geochemistry of the xenoliths suggest that the fertile peridotites were originally depleted spinel harzburgites, which were subsequently refertilized. Incompatible trace element geochemistry reveals a pervasive cryptic metasomatic overprint in all peridotites, suggesting involvement of small amounts of subduction-derived fluids from the long-lived Farallon plate beneath western North America. However, bulk reconstructed $\delta^{18}$O values of the peridotites, including the most refertilized, fall between 5.44 to 5.86 ‰, within the natural variability of unmetasomatized mantle (~5.5 ± 0.2‰ ). Together with Sm, Yb, and Ca compositional data, the oxygen isotope data suggest that the role of slab or sediment melts in refertilizing the peridotites was negligible (< 5% in terms of added melt mass). Instead, binary mixing models suggest that many of the Sierran garnet peridotites, particularly those with high clinopyroxene modes, had up to 30% mantle-derived melt added. Our data suggest that refertilization of the deep arc lithosphere, via melt entrapment and clinopyroxene precipitation, may be an important process that modifies the composition of primary arc magmas before they reach the crust and shallowly differentiate. Comparison of our data with a global compilation of arc-related mantle xenoliths suggests that sub-Moho refertilization may be more extensive beneath mature
arcs, such as continental arcs, compared to juvenile island arcs, possibly because of the greater thickness of crust and lithosphere beneath mature and island arcs.

4.1. Introduction

Island arcs are potential building blocks of continents (Taylor, 1977), but an outstanding problem is how basalts, the dominant product of island arcs, are refined into andesitic continental crust. One end-member mechanism is single-stage differentiation of granitoids from a basaltic parent (Bowen, 1928; Gill, 1981; Grove and Baker, 1984; Lee et al., 2007b; Jagoutz et al., 2009). Another scenario is multi-stage differentiation of a basaltic parent involving re-melting and assimilation of pre-existing lower crust to generate granitoids (White and Chappell, 1977). Primary magmas are mixed, assimilated, stored, and homogenized in deep crustal magma chambers (Hildreth and Moorbath, 1988; Annen et al., 2006), and upon ascent through shallower crust are further differentiated via lower pressure crystal fractionation and wallrock assimilation (DePaolo, 1981; Blatter et al., 2013; Lee et al., 2013). Leeman (1983) proposed that beneath continental arcs, primary magmas may stagnate more compared to oceanic arcs owing to the greater thickness of overlying crust in the former. As a consequence, continental arc magmas may experience greater differentiation and processing through long crustal columns, erupting a larger proportion of andesites, compared to their island arc counterparts where the crust is thinner and the magma flux is dominated by basalt.

Despite the focus on the crust and its effect on magmatic differentiation in arcs (Plank and Langmuir, 1988; Wallace and Carmichael, 1999), comparatively few studies have
addressed the role of the mantle lithosphere. One important question in the study of arc
magmatism (and magmatism in general) is how much differentiation of a primary melt
occurs below the crust? This question is important because in some arcs, particularly
mature continental arcs, such as the Andes in South America and Sierra Nevada in the
Western USA, the deep sub-Moho arc lithosphere can reach considerable thickness (Kay
et al., 1994; Ducea and Saleeby, 1996; Lee et al., 2001a; Haschke and Gunther, 2003;
Chin et al., 2012). Primary melts from the mantle wedge would first encounter such thick
lithosphere, and thus might differentiate well before they reach the crust.

Evaluating the relative contributions of crustal versus mantle processes in arc
magmatism is difficult because few arc magmas preserve memory of their inception and
early evolution in the mantle. Mantle xenoliths, however, provide a potential window
into sub-Moho processes beneath arcs. Previous studies of such xenoliths in various
oceanic island arcs show evidence of interaction between depleted mantle and hydrous
fluids and/or melts (Maury et al., 1992; Kepezhinskas et al., 1996; McInnes et al., 2001;
Ionov, 2010). Mantle xenoliths from continental arcs are found in the Sierra Nevada in
the Western USA (Dodge et al., 1988; Mukhopadhyay, 1989; Mukhopadhyay and
Manton, 1994; Ducea and Saleeby, 1996; Lee et al., 2001a; Lee et al., 2001b; Chin et al.,
2012; Chin et al., 2013), Mexico (Liang and Elthon, 1990; Luhr and Aranda-Gomez,
1997; Blatter and Carmichael, 1998); and the Canadian Cordillera (Peslier et al., 2002).
Other studies from continental arcs examined peridotites from back-arc settings, such as
the Simcoe volcanic field behind the Cascades arc (Brandon and Draper, 1996). Most, if
not all, published arc xenolith studies involved only peridotites from the spinel facies. A
notable exception is the Sierra Nevada xenoliths from California, many of which sample
the garnet facies and thus offer a rare view of the deep lithosphere beneath a mature continental arc.

The goal of this paper is to place better constraints on what influence the deep lithosphere has on the compositional evolution of continental arcs. Owing to their generally thicker lithosphere, continental arcs could undergo more extensive sub-Moho magmatic differentiation compared to oceanic arc settings, which generally have thinner lithosphere (Gill, 1981). Does the deep arc lithosphere have an analogous role to thick crust, acting as a trap or reactive filter for ascending arc magmas? How does the deep arc lithosphere evolve compositionally, given that primary melts impinge and pass through it? What is the role of slab-derived melts and/or fluids in the compositional development of the deep lithosphere or as metasomatic agents in the mantle wedge (Straub et al., 2013)?

To address the questions above, we use a suite of peridotite xenoliths from the Sierra Nevada arc in California, USA. The Sierra Nevada is one of the type localities of a mature, continental arc, and among the best-studied arcs in the world. We synthesize textural, compositional, and oxygen isotope data on Sierran peridotites, and show that magma-wallrock interaction extends far below the Moho in some continental arcs.

4.2. Geologic setting

The Sierra Nevada Batholith was a long-lived (220 to 80 Ma) (Stern et al., 1981) continental arc that formed due to subduction of the Farallon Plate beneath western North America (Figure 4-1a). The Mid to Late Cretaceous witnessed a period of especially
voluminous magmatism, peaking at ~93 Ma (Chen and Moore, 1982; Barton, 1996; Coleman and Glazner, 1997). Magmatism was concomitant with compressional deformation at all levels of the lithosphere, ranging from upper crustal fold-and-thrust belts (Sevier Orogeny), continental underthrusting beneath the arc at middle to lower crustal levels (DeCelles et al., 2009; Chin et al., 2013), and thickening of the mantle lithosphere (Chin et al., 2012). By around 74 Ma, Sierran volcanism ceased and magmatism swept eastward into the continental interior (Coney and Reynolds, 1977). This shift is attributed to flattening of the Farallon slab (Laramide Orogeny) (Dickinson and Snyder, 1978; Bird, 1988; Saleeby, 2003), but magmatic thickening and pinch-out of the asthenospheric wedge may be another explanation (Chin et al., 2012). Following the Laramide Orogeny, roll-back of the slab induced ignimbrite flare-ups ~40 Ma (Humphreys, 1995). At ~20 Ma, subduction of the Pacific-Farallon Ridge initiated the San Andreas Transform, the opening of a “slabless window” (Atwater, 1970), and the start of Basin and Range extension which continues to present (Wernicke, 1981).
Low-volume but widespread ultrapotassic and alkalic basalts erupted synchronously with Basin and Range extension in the central and eastern Sierra Nevada (Moore and Dodge, 1980; Van Kooten, 1980). Of particular interest to this study and numerous previous investigations (Dodge et al., 1988; Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1996; Lee et al., 2001a; Lee et al., 2001b) is the Late Miocene
Big Creek diatreme (37°13’N, 119°16’W). Big Creek contains a wealth of xenoliths which have been used to reconstruct the stratigraphy of the deep arc lithosphere (Saleeby et al., 2003). Starting at the base of the seismically defined crust, the lithosphere grades from low-Mg pyroxenitic cumulates, to high-Mg pyroxenites interleaved with spinel peridotite, and finally into garnet-bearing spinel peridotite at the base of the lithosphere. According to Lee et al. (2006), low-Mg pyroxenites are thought to represent cumulates of evolved, hydrous basalt at lower crustal conditions (1.5 GPa). High-Mg pyroxenites are considered higher pressure (2-3 GPa) cumulates of a primary, hydrous basalt. The peridotites are thought to represent the deepest samples of the sub-arc lithosphere, with some garnet peridotites recording equilibration up to 3.6 GPa (Ducea and Saleeby, 1998; Chin et al., 2012). In this paper, we build on the work of Chin et al. (2012), who focused on the pressure-temperature-compositional history of Sierran garnet-bearing peridotites. Combined with previously published data from Dodge (1988), Lee et al., (2001a), Lee (2005), and Chin et al. (2012), here we report new whole-rock and mineral compositional data on Sierran peridotite xenoliths.

### 4.3. Methods

Whole-rock major elements were analyzed by XRF at Washington State University at Pullman and trace element abundances were determined by ICP-MS using a ThermoFinnigan ELEMENT 2 at Rice University. Whole rocks were crushed in a ceramic SPEX mill and powdered in a shatterbox. For trace element analysis, 50 – 80 mg of whole rock powder from each sample was added to a 2.5 ml Savillex wrench-top
beaker. After two sequential open-beaker digestions using a 1:1 HF:HClO$_4$ mixture and complete open-air dry-down at ~190°C, the samples were diluted to total volume ~125 ml in 2% HNO$_3$. 0.025 ml of concentrated HCl was added to each sample to keep Fe in solution. A 1 ppb In tracer was added to each sample as an internal standard. External standards (USGS powders BHVO-1, BIR-1, AGV-1, DTS-1, JP-1) and procedural blanks were prepared along with all unknown samples and analyzed in the same runs.

Major element compositions of minerals (olivine, orthopyroxene, clinopyroxene, spinel) in spinel peridotites 08BC10, BC10-1, BC10-2 and BC10-4 were acquired using wavelength-dispersive spectroscopy electron probe micro-analysis (EPMA) using a CAMECA SX50 at Texas A&M University. Mineral data is reported in Table 4-2. Spot size was 1 micron, operating conditions were 15 kV accelerating voltage, 5 nA current, and 40 ms dwell time. Orthopyroxene, clinopyroxene, garnet, olivine, and chromite standards were used.

For oxygen isotope analysis, minerals were separated as follows. Whole rocks for 3 garnet peridotites (08BC03, 08BC08, 08BC04), 3 spinel peridotites (08BC13, BC10-1, 08BC10), and 4 garnet pyroxenites (BC98-7, BC98-5, BC98-9, BCX) were gently disaggregated using a rock hammer; the freshest, least altered gravel-sized pieces were then gently crushed in an agate mortar and sieved for grain sizes between 250 and 500 µm. From this size fraction, approximately 2 mg of minerals (olivine, orthopyroxene, clinopyroxene; garnet was picked from 1 garnet peridotite as well as from the garnet pyroxenites) from each sample were hand-picked under a binocular microscope. Only the cleanest, “gem”-like, alteration- and inclusion-free grains were
selected. The mineral separates were then cleaned for several hours in dilute HCl, ultrasonicated, and rinsed in de-ionized water.

Oxygen isotope ratios were determined using the laser fluorination technique of Sharp (1990) at the University of Texas at Austin. ~2 mg of sample was heated with a CO$_2$ laser in the presence of BrF5 and subsequently cryogenically purified using the silicate extraction line before introduction into a ThermoElectron MAT 253 isotope ratio mass spectrometer. In order to check for precision and accuracy of oxygen isotope analyses, garnet standard UWG-2 (δ$_{18}^{18}$O = +5.8‰) (Valley et al., 1995), in-house olivine standard San Carlos (δ$_{18}^{18}$O = +5.2‰), and in-house quartz standards Gee Whiz (δ$_{18}^{18}$O = +12.6‰) and Lausanne-1 (δ$_{18}^{18}$O = +18.1‰) were run. All δ$_{18}^{18}$O values are reported relative to SMOW, where the δ$_{18}^{18}$O value of NBS-28 is +9.65‰. Precision based on replicates of standards is ±0.09‰.

4.4. Peridotite microtextures and whole-rock chemistry reflect post-depletion modification

4.4.1. Sample descriptions

The Sierran peridotite suite is comprised of 14 spinel peridotites and 13 garnet peridotites (Table 2-2). Here, we report new whole-rock major element and trace element data on 8 spinel peridotites (08BC01, 08BC10, BC10-1, BC10-2, BC10-3, BC10-4, BCPL10-5, BC10-6) and 4 garnet peridotites (08BC05, 08BC11, 08BC23, P6). Only those samples with >95% anhydrous major element oxide totals are of interest here.
Furthermore, new mineral data are reported (Table 4-2) for 08BC10, BC10-1, BC10-2, and BC10-4. Of the new samples analyzed, 3 of the 4 garnet peridotites (08BC05, 08BC11, 08BC23) are extensively serpentinized with anhydrous whole-rock totals <90%. In addition, these 3 samples show evidence of carbonate infiltration, borne out in their high CaO (>10 wt.%) and high Sr (>100 ppm). We thus exclude these 3 samples from further discussion owing to the uncertainty in how serpentinization and other alteration processes have affected whole-rock compositions, but report all data in Table 2-2 for completeness.

4.4.2. Xenolith textures

Previous studies of Sierran mantle xenoliths (Ducea and Saleeby, 1996; Lee et al., 2001a; Chin et al., 2012) showed that the sub-arc mantle is compositionally stratified. The lithosphere transitions from shallow spinel peridotites (< 3 GPa) into deep (3 – 3.5 GPa) garnet-bearing spinel peridotites (hereafter referred to as “garnet peridotites”). Overall, spinel peridotites have low modal clinopyroxene (<10 %) and have a coarse-grained, protogranular texture. In contrast, garnet peridotites have higher modal clinopyroxene (>10 %) and have a fine-grained, recrystallized, porphyroclastic texture (Chin et al., 2012; Figure 4-1b). Modal mineralogies are reported in Table 4-3.

Another distinction between spinel and garnet peridotites, as discussed in Chin et al. (2012), is that in garnet peridotites, clinopyroxene occurs in fine-grained, diffuse bands juxtaposed on the relict olivine-dominated matrix. These clinopyroxene bands were interpreted as evidence for melt infiltration, and combined with geochemical data, were used to suggest that the garnet peridotites experienced refertilization by basaltic
melt. In this paper, we aim to place better constraints on the nature and magnitude of the refertilization.

4.5. Results

4.5.1. Whole-rock geochemistry

4.5.1.1. Major element compositions

Major element data of Sierran peridotites are reported in Table 2-2 and selected oxides (wt. %) are plotted against MgO (wt. %) in Figure 4-2. Negative trends are observed in the Sierran data between CaO, Al$_2$O$_3$, Na$_2$O, and SiO$_2$ versus MgO. Garnet peridotites generally plot at the low-MgO, “fertile” end whereas spinel peridotites trend towards the high-MgO, “depleted” end. However, no sharp boundary separates the two peridotite groups in any of the binary oxide plots, with some spinel peridotites overlapping in composition with garnet peridotites. Several garnet peridotites are enriched in CaO, even more so than primitive mantle (McDonough and Sun, 1995) (Figure 4-2a). These garnet peridotites also have high modal abundances of clinopyroxene (Figure 4-1).

| Table 4-1. $\delta^{18}$O$_{SMOW}$ (‰) of mineral separates in selected samples |
|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|
| Sample                         | Spinel peridotites              | Garnet peridotites              | Garnet pyroxenites              | Garnet pyroxenites              | Garnet pyroxenites              | Garnet pyroxenites              | Garnet pyroxenites              | Garnet pyroxenites              |
|                                | 08BC10  | 08BC13  | 08BC03  | 08BC04  | 08BC08  | 08BC06  | 08BC07  | 08BC08  | 08BC09  | 08BC10  | 08BC07  | 08BC08  | 08BC09  | 08BC10  | 08BC09  |
| olivine                        | 5.1    | 5.2    | 5.2    | 5.2    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    | 5.4    |
| orthopyroxene                  | 6.0    | 5.9    | 5.8    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    | 6.0    |
| clinopyroxene                  | 5.6    | 5.2    | 5.5    | 5.5    | 5.7    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    | 5.9    |
| garnet                         | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    | 5.3    |
| $\Delta$ cpx-olv               | 0.5    | 0.0    | 0.3    | 0.1    | 0.2    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    | 0.5    |
| $\Delta$ opx-olv               | 1.1    | 0.7    | 0.3    | 0.6    | 0.3    | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   | -0.1   |
| T (°C) oliv-olv                | 685    | 1225   | 1215   | 1000   | 1330   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   |
| T (°C) opx-olv                 | 1100   | 1450   | 1630   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   | 1585   |
Table 4-1. Oxygen isotope compositions of mineral separates in Sierran xenoliths

Figure 4-2. Major element systematics Sierran and global arc peridotites. CaO (A), Al₂O₃ (B), Na₂O (C) and SiO₂ (D) plotted against MgO (wt. %). Sierran spinel peridotites are shown as black circles and Sierran garnet peridotites as yellow triangles. Gray circles represent literature-compiled arc peridotites from mature island and continental arcs (Mexico, Canadian Cordillera, Kamchatka, Lesser Antilles, Philippines, and Ichinomegata) (see text for references). Blue squares represent global arc peridotites from juvenile island arcs and backarcs (Izu Bonin, South Sandwich, and Simcoe) (references?). Primitive mantle of McDonough & Sun (1995) shown as star. Melt depletion curves of primitive mantle are shown as red lines (see text for details). Black, orange, and blue lines represent mixing between 20% depleted primitive mantle and MORB end-members H, L, and D from Elthon (1992) (see text for details).
4.5.1.2. Trace elements

Like major elements, trace element abundances (Table 2-2) also show a generally continuous spectrum between spinel and garnet peridotites. Primitive-mantle normalized rare earth element (REE) patterns (not shown) in spinel peridotites are mostly flat. In contrast, several garnet peridotites have normalized REE patterns that are enriched in the heavy REEs (HREE) relative to the middle REEs (MREE) (Chin et al., 2012). If all trace elements are plotted in order of increasing compatibility with respect to anhydrous melting of peridotite (Hofmann, 1988), both spinel and garnet peridotites show depletion in Nb relative to La, as well as depletion in Ti relative to Sm (Figure 4-3). Positive anomalies in fluid-mobile elements (Cs, Li, Ba, Rb, Sr) are also observed in such a plot, as shown previously by Lee (2005) and Chin et al. (2012), and were interpreted as evidence that the Big Creek peridotites interacted with fluids and/or hydrous melts. However, we cannot rule out the possibility that elevated Cs, Li, Ba, Rb, and Sr reflect recent metasomatism/contamination by the host lava.
4.5.2. Oxygen isotope compositions of minerals

Oxygen isotope ratios for olivine, orthopyroxene, clinopyroxene, and garnet separates are reported in Table 4-1 and plotted in Figure 4-4. The average $\delta^{18}O$ values of olivine in spinel peridotites (n=3) and garnet peridotites (n=3) are 5.2 ‰ and 5.3 ‰, respectively. The average $\delta^{18}O$ value of orthopyroxene in spinel peridotites is 6.1 ‰ and in garnet peridotite is 6.0 ‰. The average $\delta^{18}O$ value of clinopyroxene in spinel peridotite is 5.4 ‰ and in garnet peridotite is 5.6 ‰. These values are representative of typical mantle olivine (5.2 ‰), garnet (5.5 ‰), and orthopyroxene and clinopyroxene (5.6 ‰) (Deines and Haggerty, 2000).
Figure 4-4. δ¹⁸O values of minerals from Sierran peridotites. A) δ¹⁸O values in orthopyroxene vs. δ¹⁸O values in olivine; B) δ¹⁸O values in clinopyroxene vs. δ¹⁸O values in olivine. Typical mantle peridotites from Mathey et al. (1994) and metasomatized peridotites from Rehfeldt et al. (2008) and Perkins et al. (2006) are also shown as white circles and blue circles, respectively. Temperature-dependent fractionation between minerals (dashed lines) are from Chiba et al. (1989) and Rosenbaum et al. (1994).

Reconstructed whole-rock δ¹⁸O values, based on the average δ¹⁸O values of different minerals and their modal abundance, in spinel and garnet peridotites range from 5.4 to 5.9 ‰. These whole-rock values overlap δ¹⁸O values of typical mantle (~5.5 ± 0.2‰ (Mattey et al., 1994; Eiler et al., 2001). In Figure 4-4, we also plot for reference δ¹⁸O values of minerals in peridotites interpreted to be metasomatized by slab-derived and/or carbonatitic melts (Perkins et al., 2006; Rehfeldt et al., 2008). The Sierran xenoliths mostly plot in the typical mantle field, although some samples, such as spinel peridotite 08BC13, extend into the “metasomatized” field.

Temperatures calculated based on equilibrium oxygen isotope exchange between clinopyroxene and olivine (Chiba et al., 1989) and orthopyroxene and olivine (Rosenbaum et al., 1994) support a high-temperature origin (>700 °C), with some
temperatures as high as 1690 °C. In contrast, temperatures obtained using subsolidus
mineral thermobarometry (Fe-Mg exchange between orthopyroxene and clinopyroxene
coupled with Al net-transfer between orthopyroxene and garnet; Brey and Kohler, 1990)
are considerably lower, ranging from 670 to 870 °C (Chin et al., 2012). Even if the high
uncertainties associated with oxygen isotope temperatures (± 200 °C; Chazot et al., 1997)
compared to the Brey and Kohler thermobarometer (± 20 °C) are taken into account,
oxygen isotope temperatures are still considerably higher than those obtained using
subsolidus mineral thermobarometry.

Substantial zoning in Al and Ca is observed in pyroxenes from the Big Creek
peridotites (Lee et al., 2001a; Chin et al., 2012), and CaO-in-orthopyroxene thermometry
(Brey and Kohler, 1990) of pyroxene cores yield higher “peak” temperatures between
900 and 1140 °C (Lee et al., 2001a) compared to temperatures obtained from rims. These
features also support a high-temperature origin of the Sierran peridotites, followed by
subsequent cooling rapid enough that the final mineral assemblage is in disequilibrium on
the µm to mm scale. Thus, one possible explanation for the discrepancy between garnet-
pyroxene thermobarometry and oxygen isotope thermometry is that in contrast to the
small spot sizes (1 µm) used for garnet-pyroxene thermobarometry, laser fluorination is a
bulk technique ablating several mineral grains simultaneously and therefore
homogenizing any core-rim compositional disequilibrium. For the purposes of this study,
any small oxygen isotope deviations from equilibrium are not significant.
TABLE 4-2. Average mineral chemistries of 08BC10, BC10-1, BC10-2, and BC10-4

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Na2O  0   0 0.017 0.01  0.76  0.26  0   0
K2O   0   0 0.005 0.01  0   0   0   0
NiO   0.39 0.02
Total 100.61 99.9 100.3 99.2
Mg#   91.5  92.1  94.5  65.6
Cr#   38.4

Opx, orthopyroxene; Cpx, clinopyroxene
Mg# = \(\frac{\text{atomic Mg}}{\text{Mg+FeT}}\) x 100
Cr# = \(\frac{\text{atomic Cr}}{\text{Cr+Al}}\) x 100

Table 4-2. Average mineral chemistries of 08BC10, BC10-1, BC10-2, BC10-4

Table 4-3. Modal mineralogy of Sierran peridotites

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Opx, orthopyroxene; Cpx, clinopyroxene; Serp, serpentine.
Modes determined either by point counting (300 - 500 pts) or by least-squares regression of average mineral chemistries and WR composition
For modes determined using least-squares regression, sum (r²) = sum of residuals squared (residual = bulk - calculated)

Table 4-3. Modal mineralogy of Sierran peridotites

4.6. Discussion

4.6.1. Refertilization trend

Major element compositional trends observed in the Sierran peridotites (Figure 4-2) appear consistent with those predicted for melt depletion from an originally fertile
primitive mantle. However, simple melt depletion of canonical primitive mantle is difficult to reconcile with the anomalously high CaO and SiO$_2$ (corresponding to high modal abundances of clinopyroxene) observed in some of the Sierran garnet peridotites (Figure 4-1b, Figure 4-2a,d). These samples also have low MgO relative to primitive mantle and thus do not resemble typical residues, which would have elevated MgO relative to primitive mantle. Instead, several lines of evidence, as we discuss below, suggest that the Sierran peridotites may have had a melt component added back to it, a process which we refer to here as refertilization, as for example, described by Ionov et al. (2005), Le Roux et al. (2007) and Luffi et al. (2009).

In Figure 4-2, we show batch melting curves (red lines) calculated using pMELTS (Ghiorso et al., 2002), beginning with a primitive mantle source (McDonough and Sun, 1995). Melting is modeled along an isentropic path initiating at 3 GPa and terminating at 0.3 GPa (assuming surface mantle potential temperature of 1300 °C). Also plotted are mixing lines (solid red, orange, and blue lines) between a melt-depleted peridotite (residue from 20% melting, corresponding to the most depleted spinel peridotites in terms of major elements) and three end-member mid-ocean ridge basalt (MORB) compositions, “H”, “L”, and “D” (Elthon, 1992). These MORB end-members approximate melts produced by decompression melting of upwelling mantle beneath mid-ocean ridges, with the “H” end-member corresponding to a low-F melt segregated at high pressures (2.5 – 3 GPa), followed by the “L” and “D” end-members which correspond to higher extents of melting with increased decompression. For CaO, Al$_2$O$_3$, and SiO$_2$, refertilization and melt depletion are for the most part indistinguishable. However, because Na is incompatible, it is rapidly depleted from the residue, resulting in concave
upwards melting arrays in a plot of Na$_2$O versus MgO (Fig. 2c). Had melt extraction been assumed to be instantaneous (e.g., fractional melting), the melting array would be predicted to be even more non-linear owing to more efficient removal of Na. Thus the roughly linear trend displayed by the Sierran peridotites in Na$_2$O versus MgO is thus difficult to explain by melt depletion and may be better explained by mixing or reaction between a basaltic melt and a harzburgite as we discuss below. Re-introduction of basaltic melt and associated clinopyroxene precipitation would also explain the elevated CaO and SiO$_2$ (relative to primitive mantle) observed in garnet peridotites.

High Cr# (atomic (Cr/(Cr+Al))*100) of relict spinels in combination with fertile bulk rock compositions may also suggest refertilization (Chin et al., 2012). Cr# in spinel is an indicator of the melting degree of mantle peridotites (Dick and Bullen, 1984), and increases with melting owing to the compatibility of Cr. Re-introduction of melt would decrease the whole-rock Cr#, but Cr# of relict, refractory spinels may survive post-melting phenomena (Voigt and Handt, 2011) and thus can preserve original melt depletion signatures. Cr# of spinel in both Sierran spinel peridotites and garnet peridotites range between 30 and 40, with the exception of two samples which have low (<20) Cr# spinel. Cr#’s between 30-40 correspond to 15-20% melt depletion (Hirose and Kawamoto, 1995; Matsukage and Kubo, 2003). At these degrees of melting, modal clinopyroxene should be ~5% (Baker and Stolper, 1994). Some Sierran garnet peridotites, despite having high Cr# in relict spinel, have excess modal clinopyroxene (10 – 20%), which is greater than that predicted by simple melt depletion (Figure 4-5). This suggests that the bulk rock has been refertilized but the cores of spinels preserve a memory of prior melt depletion due to slow equilibration of Cr between spinel and
whole-rock. The decoupling between Cr# in relict spinel and modal clinopyroxene in Figure 4-5 is consistent with earlier findings of Chin et al. (2012), who showed similar decouplings between spinel Cr# and whole-rock composition. Major element compositional trends and high Cr# of relict spinel decoupled from modal clinopyroxene thus suggest a refertilization trend from a depleted harzburgite precursor.

Based on thermobarometric constraints from our previous studies (Lee et al. 2001a, Chin et al. 2012), the extent of refertilization appears to increase with depth. Shallow depths (< 90 km, 3 GPa) are dominated by undeformed, clinopyroxene-poor (<10%) spinel peridotites and smaller amounts of more fertile, clinopyroxene-rich (10 – 15 %) spinel peridotites. At greater depths (90 – 110 km, 3 – 3.5 GPa) highly deformed garnet peridotites with excess clinopyroxene (>10 %) are present. Chin et al. (2012)
showed that the garnet-bearing spinel peridotites originally formed as shallow spinel peridotites, which were subsequently transported to garnet facies depths and rapidly cooled, as evidenced by widespread preservation of mineral zonation and Al-depletion halos next to exsolved garnet and amphibole lamellae (Lee et al., 2001a; Chin et al., 2012). Based on their bulk rock compositions and their metamorphic histories, the Sierran peridotites have been interpreted to represent a cogenetic package of mantle lithosphere, initially formed by melt depletion associated with Mesozoic arc magmatism (Lee et al., 2001b) and then compressed and cooled as the lithosphere was thickened to pressures of 3 – 3.5 GPa, presumably approaching the top of the subducting slab (Chin et al., 2012).

Finally, although refertilization is most apparent in the garnet peridotites (e.g., excess clinopyroxene), both spinel and garnet peridotites were also affected by cryptic metasomatism, wherein the trace-element geochemistry appears to have been modified without concomitant changes in major elements or modal mineralogy. As shown in Figure 4-3 and described earlier, both Sierran spinel peridotites and garnet peridotites have low Ti/Sm and Nb/La, but relative enrichment in large ion lithophile elements (LILE), such as Sr. We interpret the anomalously low Ti/Sm (<3000) and Nb/La (<1) ratios as indicating the involvement of a fluid metasomatic component, because of low solubility of Ti and Nb in fluids (Tatsumi et al., 1986). Interestingly, Sr/Nd or U/Pb show no systematic variation between the two lithologies, in contrast to the marked gradient in clinopyroxene mode with increasing pressure. Thus, the fluid-enriched signature of the Sierran xenoliths may represent a pervasive overprint, perhaps related to slab-derived fluids from the Farallon plate, as proposed by Lee (2005).
Because the Farallon plate persisted beneath the Sierra Nevada arc long after arc magmatism ceased (Dumitru et al., 1991; Humphreys et al., 2003), it is not clear if this fluid signature was coeval with the major element refertilization or due to later metasomatism.

### 4.6.2. Refertilization models

In this section, we constrain the nature and extent of refertilization of the Sierran peridotites. How did the fertile garnet peridotites, which were demonstrated above to have originated as harzburgites, come to contain excess clinopyroxene (and garnet), in contrast to the shallower spinel peridotites which appear to have experienced less refertilization? Was the refertilization related to infiltration of a slab melt or mantle melt? Here, we explore two end-member refertilization scenarios (Bodinier et al., 1988).

The first scenario is that the refertilized peridotites represent a mixture between trapped melt and an original depleted harzburgite (batch melt addition). For example, clinopyroxene and garnet-enrichment in the mantle might be a product of high-pressure (>2 GPa) crystal fractionation of hydrous arc magmas (Green, 1972). The second scenario is that the refertilized peridotites represent products of melt-rock reaction between a passing melt and a peridotite protolith.

To evaluate these hypotheses, we modeled the behavior of trace elements because the major elements alone are not sufficient to distinguish between mixing and melt-rock reaction. We choose moderately incompatible trace elements because these elements are sensitive to mineralogic changes associated with refertilization, but rather insensitive to cryptic metasomatism (Canil, 2004; Lee et al., 2007a). MREEs and HREEs, in this
regard, are useful because they are moderately incompatible and their partition coefficients are relatively well-constrained. We avoid the use of LREE because these elements are more susceptible to cryptic metasomatism (Nixon et al., 1981; Condie et al., 2004). In our models, we use Yb, which is sensitive to garnet growth, and Sm, which is more sensitive to clinopyroxene growth.

4.6.2.1. Binary mixing between melt and harzburgite

In Figure 4-6, we show simple mixing models between a depleted peridotite precursor and various melts. In all mixing models, the depleted end-member is a harzburgite formed by 20% melt depletion of primitive mantle, reflecting the most depleted spinel peridotites in terms of major element systematics (Figure 4-2). This 20% depletion is also within the bounds estimated using Cr# of relict spinel, as described above and in Chin et al. (2012). To this harzburgite, we add a series of mantle-derived melts (solid gray and orange lines in Figure 4-6) or a partial melt of eclogitic material to approximate a slab melt (blue field in Figure 4-6). To model the trace element composition of mantle-derived melts and the initial composition of the harzburgitic residue, we used a non-modal, fractional melting model using starting mineral modes and melting modes from Johnson (1990). Melts from the spinel stability field (black lines Figure 4-6a,b) and garnet stability field (orange lines, Figure 4-6c,d) were calculated. Slab-derived melts were calculated using an eclogitic source (62% garnet, 38% clinopyroxene) based on the melting relationships in Kessel et al. (2005).
Figure 4-6. Sm-Yb-CaO systematics of Sierran peridotites and binary mixing models. Sm and Yb in ppm and CaO in wt. %. Mixtures between 20% depleted primitive mantle and mantle-derived melts shown as solid colored lines (black for melts originating in the garnet facies; orange for melts originating in the spinel facies); contours of constant percentage of melt added to the depleted peridotite (5% increments) shown as dashed lines. Mixtures between depleted peridotite and slab melts (F ranging from 1 to 15%) plot in the blue fields in A) and C). Melt depletion curves (red lines) from primitive mantle (star) also shown for reference. A) Sm/Yb versus Yb (ppm); melts originate from garnet facies B) Sm/Yb vs. CaO (wt.%); melts originate from garnet facies C) Sm/Yb versus Yb (ppm); melts originate from spinel facies D) Sm/Yb versus CaO (wt. %); melts originate from spinel facies.

For spinel-facies melting, Sm and Yb mineral/melt partition coefficients for clinopyroxene are from Hauri et al. (1994) and for olivine and orthopyroxene are from Lee et al. (2007a), the latter of which were determined so that they are internally consistent with the Hauri et al. (1994) clinopyroxene partition coefficients. Sm and Yb
partition coefficients for spinel are from Stosch (1982). For garnet-facies melting, Yb mineral/melt partition coefficients for olivine, orthopyroxene, clinopyroxene, and garnet were taken from Canil (2004); Sm mineral/melt partition coefficients for olivine, orthopyroxene, and clinopyroxene were from Lee et al. (2007a) and for garnet from Johnson et al. (1990). Mineral/melt partition coefficients for eclogitic systems were taken from Barth et al. (2002). CaO contents of 1, 5, 10, and 15% melts of primitive mantle were obtained from the isentropic melting calculation described in Section 6.1. CaO in these mantle melts ranged from 5 to 12 wt.%.

In Figure 4-6, mixing lines between the depleted peridotite end-member and mantle-derived melts corresponding to different melting fractions (F = 1 to 15%) are plotted as solid black lines; dashed black lines represent contours of 5% melt increments added to the mixture. In Figure 4-6a and b, mixing of depleted peridotite with melts derived from the garnet facies is shown, whereas in Figure 4-6c and d, mixing with spinel peridotite melts is shown. Melt depletion trends predicted for residual peridotite are also shown for comparison (red lines).

The results of these models can be summarized as follows. Melt depletion results in depletion in Sm and Yb, but greater depletion in Sm compared to Yb and CaO because CaO and Yb are less incompatible than Sm. This results in concave up positive correlations of Sm/Yb versus Yb and CaO during melting. Mixing, however, gives fundamentally different trends. Mixing with mantle-derived melts gives convex upwards mixing lines in Sm/Yb versus Yb or CaO diagrams because the relative enrichment of Sm in mantle melts is far higher than that of Yb and Ca. Due to the differences in
compatibility of these elements, the addition of low degree melts will increase Sm/Yb at a rate faster than Yb or Ca, increasing the nonlinearity of the mixing curves. In detail, because melts derived from garnet peridotite melting have significantly lower Yb than melts derived from spinel peridotites, mixing with garnet peridotite melts causes a greater increase in Sm/Yb for the same amount of mixing with a spinel peridotite melt. Finally, because of the large amount of garnet (>50 %) in the residue of slab melts, Yb behaves as a compatible element during slab melting, generating hypothetical melts with extremely high Sm/Yb. This results in nearly vertical mixing lines in Sm/Yb versus Yb plots.

With the understanding of these trends in place, we can now examine the Sierran peridotites. The spinel peridotites have variable Yb and CaO, but some have anomalously high Sm/Yb compared to the garnet peridotites. The garnet peridotites have variable Yb and CaO, but are generally higher in Yb and CaO than the spinel peridotites. In particular, the Sm/Yb ratios of the garnet peridotites show no correlation with Yb or CaO. Some of the spinel and garnet peridotites overlap in composition. There appears to be no strong evidence that the peridotites have mixed with any significant amount (e.g., >5%) of a slab melt. While four of the spinel peridotites have Sm-Yb systematics consistent with small amounts of added slab melt, these samples can also be explained by the addition of low-degree garnet peridotite melts, resulting in cryptic metasomatism.

For the rest of the peridotites, it can be seen that the Sm-Yb-CaO systematics of both the spinel and garnet peridotites can be explained by mixing of up to 30% of a 10-15% melt of a garnet peridotite or 15-30% of a 5-10% melt of a spinel peridotite. Assuming clinopyroxene has a CaO content of ~20 wt. % and that the mantle melts have ~10% CaO and assuming that all the CaO in the added melt is eventually accounted for by
clinopyroxene, such mixing proportions of melt correspond to the addition of ~7-17% of clinopyroxene added to the harzburgite. Although this mixing scenario is simplistic, the estimated amount of clinopyroxene added is consistent with the observed mineral modes.

4.6.2.2. Melt-rock reaction

We now examine the second refertilization scenario: melt-rock reaction. Here, a passing melt reacts with a protolith, leaving behind a reacted rock that has partially or fully equilibrated with, but no longer contains, the melt. Depending on the melt/rock ratio, the extent of equilibration between melt and rock, and the compositions of melts and harzburgitic precursors, melt-rock reaction can generate a diversity of reacted rock types (Garrido and Bodinier, 1999). We model the end-member case where the melt is in excess and the protolith compositionally equilibrates with the melt, resulting in an increase in clinopyroxene along with a change in trace element composition. In other words, we assume the melt/rock ratio is so high that the melt does not change its composition, but the major and trace element composition of the rock changes accordingly. In this model, we evaluate the extreme case in which the observed mineralogy represents the product of melt-rock reaction rather than mixing, and our goal is to quantify how this reaction is manifested in whole-rock trace element composition. Although this approach is simplistic, it bounds the geochemical modification of peridotites by melt-rock reaction.

In our model, the reacted mineral compositions are constrained by the equilibrium partition coefficient. We use the garnet-facies melt compositions (1, 5, 10, and 15% melts of primitive mantle) from the mixing calculations discussed in the previous section
(Figure 4-6). We assume that the observed mineral modes are already the product of refertilization. Our goal is to calculate the trace element signature of the peridotites upon equilibration with the melt. For a given melt composition and known partition coefficients (see above), we calculate olivine, orthopyroxene, clinopyroxene, and garnet trace element compositions in equilibrium with the melt. Hypothetical bulk rock compositions are then estimated with a weighted average of all the minerals according to their observed modes. To explore the plausible variable space so that the entire spectrum of peridotites is encompassed, we generated two hypothetical arrays. In one, we varied clinopyroxene mode from zero to 100%, keeping the olivine/orthopyroxene ratio constant at 2. In the second array, we varied garnet mode while keeping clinopyroxene/olivine and clinopyroxene/orthopyroxene ratios fixed at 0.15 and 0.2, respectively.

In Figure 4-7, we summarize our model results. When only clinopyroxene is varied, the suites of reacted rocks follow positive, concave down trajectories on a plot of Sm/Yb versus Yb (Figure 4-7a), superficially resembling binary mixing (Figure 4-6a). An additional feature of the Sm/Yb versus Yb plot in Figure 4-7a is that rocks reacted with low-F melts fall on highly non-linear trends compared to rocks reacted with higher F melts. This is because clinopyroxenes in equilibrium with low-F melts have higher Sm/Yb compared to clinopyroxenes in equilibrium with high-F melts, in which Sm is diluted.

When garnet is varied, the reaction arrays in the Sm/Yb versus Yb plots have negative slopes and are concave up (Figure 4-7b) compared to the clinopyroxene-only case. Sm/Yb ratios in reacted rocks decrease with increasing garnet mode, owing to the
compatibility of Yb in garnet. Finally, isopleths of clinopyroxene and garnet mode in Figure 4-7 have negative slopes because the concentrations of Sm and Yb in the melts decrease and increase, respectively, with increasing F for garnet-facies melting.

Figure 4-7. Sm-Yb systematics of Sierran peridotites and melt-rock reaction models. Arrays of reacted rocks are shown as solid blue lines. Contours of constant clinopyroxene (cpx) mode or garnet (gt) mode shown as dashed blue lines. Melt depletion curves (red lines) from primitive mantle (star) also shown for reference. Gray field represents range of observed modal clinopyroxene or garnet in the Sierran peridotites. A) Sm/Yb versus Yb (ppm), varying clinopyroxene mode while olivine/orthopyroxene ratio fixed at 2 in reacted rocks. B) Sm/Yb versus Yb (ppm); varying garnet mode while clinopyroxene/olivine = 0.15 and clinopyroxene/orthopyroxene = 0.2.
We now compare the Sierran peridotite compositions to the hypothetical melt-rock reaction arrays. As an additional constraint, we also denote the field (gray), which corresponds to the observed range of clinopyroxene (Figure 4-7a) and garnet (Figure 4-7b) modes in the Sierran peridotites. In Figure 4-7a, half of the spinel peridotites fall within the gray field, suggesting they could represent reacted rocks containing 10% or less modal clinopyroxene and reacted with low F (1 to 5%) melts. However, some spinel peridotites cannot be explained as melt-rock reaction products because to fit their Sm/Yb systematics would require reaction with modal clinopyroxene (>30%) higher than observed. Similarly, to fit the Sm/Yb of most of the garnet peridotites would require reaction with protoliths having >50% modal clinopyroxene, far higher than observed. If we examine the models where garnet mode is varied, even fewer data can be explained as melt-rock reaction products (only one sample falls within the gray field in Figure 4-7b).

Several garnet peridotites plot above the trend of rocks reacted with a low-F melt. In summary, melt-rock reaction at best explains a small subset of the observed trace element compositions. The variable clinopyroxene model (Figure 4-7a) can only reproduce trace element signatures with unreasonably high clinopyroxene modes, and the variable garnet model (Figure 4-7b) fails to explain the majority of garnet peridotites (except for one sample). The simplest model for explaining the major and trace element systematics of the Sierran peridotites is thus binary mixing, that is, the addition of a melt to a depleted peridotite.
4.6.3. Constraints on origin of the refertilizing melt from oxygen isotopes

A key question is whether the refertilizing magmas are mantle-derived melts or slab melts. In our above discussion of Sm-Yb-CaO systematics, the contribution of slab melts to major element refertilization was considered to be small or negligible (<5 %). Oxygen isotopes can provide additional constraints on the origin of the refertilizing magmas. In particular, because O is a major element, any observed changes in bulk rock O isotopes require significant amounts of open system behavior.

In Figure 4-8, we compare the reconstructed bulk $\delta^{18}$O value of the Sierran peridotites to hypothetical mixing lines between between mantle harzburgite (5.5 ‰) and three melt endmembers: mid-ocean ridge basalt (MORB; 5.5 ‰), a slab melt (11.3 ‰; Eiler et al. (1998), and pelagic clay (20 ‰; Clayton et al., 1972; Savin and Epstein, 1970). The slab melt composition from Eiler et al. (1998) was chosen because as a trapped melt inclusion in a mantle olivine xenocryst, this melt might approximate a pristine slab-derived melt undisturbed by any crustal contamination. In addition, the pelagic clay average of 20 ‰ was selected as a lower limit of the spectrum of oceanic sediments (marine carbonates and cherts would have even higher $\delta^{18}$O values: 25 – 32 ‰ and 35 – 42 ‰, respectively; Kolodny and Epstein, 1976).

In Figure 4-8, we plot bulk $\delta^{18}$O value versus estimated amount of melt added. To change bulk the $\delta^{18}$O value significantly requires the addition of large amounts of melt, which would be reflected in high clinopyroxene mode. Samples with high clinopyroxene mode have mantle-like $\delta^{18}$O values. Thus, it is clear that the Sierran peridotites are best explained as mixtures with mantle-derived melts, such as MORB, not
with endmembers with anomalously high $\delta^{18}$O values because the bulk $\delta^{18}$O value is indistinguishable from mantle values to within error of our measurements and natural variation of unmetasomatized mantle rocks.

![Diagram showing reconstructed bulk $\delta^{18}$O value of Sierran peridotites versus fraction of melt added based on batch addition model. See Figure 4-6 for melt fractions. Mixing end-members sediment (pelagic clay with a $\delta^{18}$O value of 20 ‰ from Clayton et al. (1972), and slab melt with a $\delta^{18}$O value of 11.3 ‰ from Eiler et al. (1998)) also shown. Field of Sierran pyroxenites is from Ducea (2002) and field labeled “mantle minerals” from Mattey et al. (1994).]

**Figure 4-8.** Reconstructed bulk $\delta^{18}$O value of Sierran peridotites versus fraction of melt added based on batch addition model. See Figure 4-6 for melt fractions. Mixing end-members sediment (pelagic clay with a $\delta^{18}$O value of 20 ‰ from Clayton et al. (1972), and slab melt with a $\delta^{18}$O value of 11.3 ‰ from Eiler et al. (1998)) also shown. Field of Sierran pyroxenites is from Ducea (2002) and field labeled “mantle minerals” from Mattey et al. (1994).

### 4.6.4. The deep lithosphere filter in global arcs

We showed that the Sierran mantle lithosphere, particularly the deepest peridotites, experienced pervasive refertilization by basaltic, mantle-derived melts, not slab melts. In contrast, most shallow spinel peridotites experienced smaller degrees of refertilization. To explain this depth-gradation in refertilization, we propose the
following scenario (Figure 4-9). Where mantle lithosphere is thick, as in continental arcs such as the Sierra Nevada, we hypothesize that the longer crustal and lithospheric column more efficiently traps rising magmas, resulting in refertilization of the deeper portions of the mantle lithosphere and manifesting as the addition of clinopyroxene. The deepest arc garnet peridotites thus appear anomalously fertile. We hypothesize that, in arcs with thin lithosphere (e.g., most island arcs), melts are granted easier passage and thus are not trapped as efficiently within the crust or lithospheric mantle. These shallow peridotites are thus predicted to show less refertilization.

Figure 4-9. Cartoon illustrating refertilization in arcs with varying lithospheric thickness. Black dot with arrow represents primary arc magma derived from decompression melting and hydrous fluxing (blue arrows) of the mantle wedge. Green region represents mantle lithosphere, white region represents crust and crust-derived sediments, and blue region represents down-going slab. Magma chambers in the deep crust and mid-upper crust are also shown. A) In juvenile arcs, lithosphere is thin and primary basaltic magmas have easier pathways to eruption. B) In continental arcs, lithosphere is thick and primary basaltic magmas may be trapped at depth (red blobs), refertilizing the mantle lithosphere. Primary magmas also must traverse thicker crustal columns, and thus undergo increased differentiation.
Our hypothesis predicts that there should be a contrast in major element fertility of the lithospheric mantle underlying oceanic island arcs and continental arcs, with continental arc lithospheric mantle being more fertile because they are generally underlain by thicker crust, and by implication, thicker lithosphere. In Figure 4-2, we plot global peridotites from various arcs and back-arcs. Adopting Gill’s (1981) classification, we define arcs as “juvenile” if they have crust <25 km thick and arcs as “mature” if they have crust >25 km thick. Crustal thickness does not equate with lithosphere thickness, but we assume that thick crust implies thicker lithosphere in tectonically active areas.

Juvenile arcs for which peridotite xenolith data are available include Izu Bonin (Parkinson and Pearce, 1998), South Sandwich (Pearce et al., 2000), and Simcoe Volcanic Field in the Cascade backarc (Brandon and Draper, 1996). Crustal thicknesses of the Izu Bonin and South Sandwich arc are estimated to be 22 km and 20 km, respectively (Takahashi et al., 1998; Larter et al., 2003). We consider Simcoe “juvenile” based on crustal thickness estimates between 20 – 30 km in the neighboring Columbia Plateau region (Hill, 1972; Catchings and Mooney, 1988). Mature arcs with available peridotite xenolith data include Mexico (Liang and Elthon, 1990; Luhr and Aranda-Gomez, 1997; Blatter and Carmichael, 1998), Kamchatka (Ionov, 2010), the Ichinomegata Crater on Honshu (Canil, 2004), Canadian Cordillera (Peslier et al., 2002), the Philippines (Maury et al., 1992), and the Lesser Antilles (Parkinson et al., 2003). Crustal thicknesses for these arcs are: Mexico, 35 – 50 km (Urrutia-Fucugauchi and Flores-Ruiz, 1996); Kamchatka, 30 – 40 km (Levin et al., 2002); Ichinomegata, 30 – 35 km (Taira, 2001); Canadian Cordillera, 35 – 40 km (Cook et al., 1988); Philippines, 34 km (Besana et al., 1995); and Lesser Antilles, 33 km (Boynton et al., 1979).
From Figure 4-2, it is apparent that mature arcs (including the Sierran peridotites), span the entire spectrum from depleted to fertile compositions, with some arcs containing peridotites even more fertile than primitive mantle. Similar to the Sierran xenoliths, peridotites from the Lesser Antilles were interpreted by Parkinson and Pearce (1998) as refertilized, clinopyroxene-enriched harzburgites. The Mexican arc also contains some highly fertile peridotites, but whether these have been refertilized by recent arc magmatism or are remnants of non-arc mantle from the distant past is unclear, given the long and complex tectonic history in that region. In contrast to the mature arcs, none of the peridotites from juvenile arcs extend to highly fertile compositions, with the majority plotting at the melt-depleted end. However, one caveat to bear in mind is that the peridotites from juvenile arcs (e.g., Izu Bonin and South Sandwich) are from the forearc, and thus it is not clear whether they are truly residual mantle associated with arc magmatism or accreted oceanic lithosphere that was subsequently modified by subduction processes.

Although the number of peridotites sampled at mature arcs outnumbers those from juvenile arcs based on the literature data shown in Figure 4-2, lithospheric mantle beneath mature arcs seem to trend towards higher fertility than beneath juvenile arcs. Higher fertility, particularly in CaO, could be the result of refertilization by melt entrapment and high-pressure clinopyroxene fractionation, similar to what is observed in the Sierran peridotites. That this trend is observed in the global data supports our hypothesis that thicker arcs could cause primary arc magmas to fractionate within the lithospheric mantle and therefore before rising across the Moho and into the crust. Mafic minerals such as garnet, amphibole, and pyroxenes comprise deep crustal cumulates of
are magmas, and fractionation of these minerals is considered one mechanism by which primitive magmas evolve to granitoid compositions (Ducea, 2001; Greene et al., 2006; Lee et al., 2006; Münntener and Ulmer, 2006; Alonso-Perez et al., 2009; Jagoutz et al., 2009). Our observations suggest that magmas may fractionate clinopyroxene well before entering the crust. Such fractionation would increase Al and decrease Ca without significant change in SiO$_2$ in the magma, generating an evolved basalt from which all shallower fractionation in the crust must derive.

4.6.5. Conclusions

Peridotite xenoliths from the Sierra Nevada reveal a depth-gradation in refertilization from depleted spinel peridotite at depths <90 km to fertile garnet peridotites between depths of 90 – 105 km. Textural, modal and geochemical observations show that spinel peridotites are undeformed, clinopyroxene-poor, and generally depleted in magmaphile elements (CaO, Na$_2$O), whereas garnet peridotites are highly deformed, clinopyroxene-rich, and have fertile major element compositions, with some garnet peridotites even more fertile than canonical primitive mantle. High Cr# of relict spinels in the garnet peridotites suggest that the garnet peridotites were original depleted spinel harzburgites that were subsequently refertilized. The simplest explanation for the observed refertilization trends is that the fertile Sierran peridotites represent binary mixtures between a depleted harzburgite and up to 30% of a mantle-derived melt. Based on major and minor elements as well as moderately incompatible trace elements, the role of slab or sediment melts as possible mixing endmembers is negligible, despite the fact that these peridotites were refertilized in a subduction zone.
Comparison to mantle xenoliths from other arc environments suggests that refertilization of arc lithospheric mantle may be an important process associated with arc magmatism. We speculate that deep melt entrapment might be facilitated beneath arcs with thicker crust and by implication thicker lithosphere. Comparison of global arc peridotite data shows a general trend towards higher fertility in mature arcs with thick crust compared to juvenile arcs with thin crust. Significant sub-Moho magmatic fractionation may thus occur beneath mature arcs, implying that “primary” arc magmas may undergo high-pressure fractionation before they reach the crust.
Thermal histories of the Sierran deep arc lithosphere: constraints from coupled Lu-Hf and Sm-Nd geochronology and Al-in-orthopyroxene diffusion

Thickening of the lithosphere beneath subduction zones affects the composition of newly formed arc crust, the tempo of magmatism, as well as the rheological stability of the arc root. Here, we investigate the thermal evolution of the deep lithospheric root beneath the Sierra Nevada continental arc. Major element compositional gradients, particularly in Al, reveal cooling and increasing pressure in garnet websterites and garnet peridotites, xenoliths that represent the deepest Sierran lithosphere. Diffusion modeling of Al-depletion haloes between host orthopyroxene and exsolved garnet lamellae indicate halo formation times of 6 to 8 Ma, suggesting the deepest lithosphere cooled relatively
fast. We also report new coupled Lu-Hf and Sm-Nd internal mineral isochron ages on a lower-crustal low-Mg garnet clinopyroxenite (Lu-Hf age: 107.6 ± 1.5 Ma; Sm-Nd age: 84.0 ± 4.1 Ma), an amphibole-bearing garnet websterite (Lu-Hf age: 60.8 ± 3.2 Ma; Sm-Nd age: 81.0 ± 2.0 Ma), and an amphibole-free garnet websterite (Lu-Hf age: 92.6 ± 1.6 Ma; Sm-Nd age: 88.8 ± 3.1 Ma). The garnet websterites record similar final equilibration pressures and temperatures as garnet peridotites, thus our age data on garnet websterites are also likely to reflect those of the peridotites for which age data are unavailable. We propose a simple thermal model that can account for the substantial time lag of 20 Ma between Lu-Hf versus Sm-Nd ages in the lower crustal pyroxenite, the absence of a time lag between chronometers in the amphibole-free garnet websterite, and the 6–8 Ma required to form Al-depletion haloes in orthopyroxene in the garnet peridotite. In our model, the lithospheric root thickens to ~100 km where it then impinges on the subducting slab. Slab impingement induces rapid cooling, and can explain the simultaneous closure of Lu-H and Sm-Nd chronometers in deeply formed high-Mg websterite as well as the Al-depletion haloes in peridotite. Chilling from the slab is thus most pronounced at the slab-arc lithosphere interface, resulting in cooling by several hundred degrees in a few Ma. In contrast, the cooling effect of the slab takes several tens of Ma to reach shallower levels of the lithosphere, explaining the 20 Ma lag between Lu-Hf and Sm-Nd ages in the lower-crustal pyroxenite.
5.1. Introduction

Lithospheric roots are a ubiquitous feature of volcanic arcs (DeBari and Sleep, 1991; Kay and Mahlburg-Kay, 1991; Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1996; Kelemen and Hanghoj, 2003; Greene et al., 2006; Lee et al., 2006; Jagoutz et al., 2009; Chin et al., 2012). These roots are largely comprised of mafic cumulates that represent the igneous counterpart to the evolved granitoids that dominate arc crust. Petrologic studies indicate that crystal fractionation of such cumulates may occur at high pressures (~1.0 GPa, or ~30 km) (DeBari and Greene, 2011), at or below the seismic Moho. Subsequent thickening and cooling of these cumulates stabilizes garnet (Behn and Kelemen, 2006), resulting in a significant density increase relevant to ambient mantle that may eventually lead to foundering of the root (Kay and Mahlburg Kay, 1993; Jull and Kelemen, 2001), ultimately driving arc crust to a continental crust-like composition.

Lithospheric roots beneath arcs may be limited to a maximum thickness, governed either by a critical density contrast after transition into garnet stability that may trigger convective removal, or by geometric limits imposed by the downgoing slab. The depth to the slab (~110 km) beneath the volcanic front is remarkably uniform across arcs worldwide (Tatsumi, 1986). Once thick lithosphere impinges on the cold slab, the asthenospheric mantle wedge is pinched out, removing sufficient headspace for
decompression melting and therefore preventing arc magmatism from continuing. Thus, instead of the popular view that variations in slab dip (i.e., the lower plate) modulate arc magmatism (Coney and Reynolds, 1977; Kay and Coira, 2009), an alternative hypothesis is that downward thickening of the lithospheric root (i.e., the upper plate) to depths approaching the slab may be the cause of an arc’s demise.

If thickening of the upper plate is limited by the slab, we might predict that the deepest parts of the lithospheric root might undergo rapid cooling upon hitting the slab, whereas shallower parts of the lithosphere might experience more protracted cooling. To test this hypothesis, we combined geochronology and diffusion modeling on deep lithospheric arc xenoliths. We report new age data using coupled Lu-Hf and Sm-Nd internal mineral isochrons of garnet pyroxenites from the Sierra Nevada, California, a type locality of a mature continental arc. In addition, we constrain timescales of lithospheric cooling using Al-in-orthopyroxene diffusion modeling in a garnet peridotite. Together, our coupled geochronological data and diffusion study reveal that the Sierran arc lithosphere experienced variable cooling histories.

5.2. Geologic setting & samples

The Mesozoic Sierra Nevada arc was a long-lived (220 to 80 Ma) continental arc that formed as a result of Farallon plate subduction beneath western North America. Magmatism reached its zenith during the Late Cretaceous (100 to 85 Ma, peaking at ~93 Ma). Such a profusion of granitoids requires at least twice that amount in mass of mafic-ultramafic cumulate or restite material (Ducea and Saleeby, 1998), pointing to a deep
lithospheric root estimated to be at least 30 to 50 km thick (Ducea, 2001). Based on xenolith studies, the Sierran lithosphere was at least 100 km thick from Mesozoic to Miocene times (Dodge et al., 1988; Mukhopadhyay and Manton, 1994; Ducea and Saleeby, 1998; Lee et al., 2001; Chin et al., 2012). To date, there is only one geochronological study of Sierran deep lithospheric xenoliths (Ducea and Saleeby, 1998). Ducea and Saleeby measured Rb-Sr and Sm-Nd mineral and whole-rock isochrons on granulite-facies xenoliths, garnet clinopyroxenites, one websterite, and one garnet peridotite. Sm-Nd websterite and pyroxenite ages ranged from 120 to 84 Ma, supporting the hypothesis that these xenoliths represent complementary cumulates to the Sierran arc granitoids. Their deepest equilibrated xenolith, a garnet peridotite, yielded an age reflecting the host magma eruption event.

In this paper, we present new mineral composition data and coupled Lu-Hf and Sm-Nd mineral and whole-rock ages on 3 Sierran garnet pyroxenites: BC98-5, BC98-7, and BCX. The former two samples are high-Mg garnet websterites and record equilibration pressures between 1 and 3 GPa (Lee et al., 2006), indicating similar depths of equilibration as garnet peridotites (Lee et al., 2006; Chin et al., 2012). The latter sample, BCX, is a low-Mg garnet clinopyroxenite and has been previously interpreted by Lee et al. (2006) to represent a cumulate of an evolved arc basalt.

In addition to the new age data on Sierran pyroxenites, we report the results of Al-in-orthopyroxene diffusion modeling in a garnet-bearing spinel peridotite, 1026V. Whole-rock geochemistry, mineral compositions, textural analysis, and thermobarometry on this sample were previously reported in Lee et al. (2001) and Chin et al. (2012). As
described in those studies, several Sierran garnet peridotites show textural and chemical
disequilibrium indicative of increasing pressure and coeval cooling. Examples of
microtextures found in Sierran garnet peridotites include garnet coronas around relict
chromian spinel grains, garnet exsolution lamellae within deformed orthopyroxene
porphyroclasts, and fine-grained garnet clustered along clinopyroxene grain boundaries.
Such a wealth of textural information can be mined for information on thermal histories.
Here, we focus on Al-depletion haloes that formed as a result of garnet exsolution in
originally high-Al orthopyroxene. Peridotite 1026V, in particular, contains numerous
garnet exsolution lamellae in large orthopyroxenes and is thus provides several
opportunities to explore diffusion kinetics between orthopyroxene and garnet.

5.3. Methods

Electron microprobe measurements on garnet and orthopyroxene in sample
1026V were obtained using wavelength-dispersive spectroscopy (WDS) on the JEOL
JXA-8530 F “Hyperprobe” at Yale University (15 kV accelerating voltage, 5 nA probe
current, 40 ms dwell time, spot size 1 micron). Elemental maps of orthopyroxene in
1026V were acquired in WDS mode on the Hyperprobe. Major element compositions of
garnet and orthopyroxene in samples BC98-5, BC98-7, and BCX were obtained in WDS
mode on the CAMECA SX-50 electron microprobe at Texas A&M University.

Garnet and clinopyroxene were separated from BCX, BC98-5, and BC98-7 as
follows. ~5 cm chunks of whole rocks were gently broken up using a rock hammer
(samples were wrapped in paper towels to minimize contact with the metal rock
hammer). Only the freshest interior gravel-sized pieces were selected from this process, then crushed gently in an agate mortar. Next, each sample was sieved for grain sizes between 250 to 500 µm, from which garnet and clinopyroxene mineral separates were hand-picked under a binocular microscope. Two fractions of garnet were separated in each sample, a “clean”, kelyphite-free fraction and a “dirty”, kelyphitized fraction. Finally, the mineral separates were cleaned in dilute HCl, ultrasonicated for several hours, and rinsed with ultrapure de-ionized water. After sample and spike homogenization was achieved by dissolution in high-pressure, steel-jacketed Teflon bombs, Lu-Hf isotope geochemistry was carried out following the protocol outlined in Blichert-Toft (2001).

5.4. Results

5.4.1. Pyroxene and garnet major element compositions

Major element compositions of clinopyroxene, orthopyroxene, and garnet in the two pyroxenites and one websterite are reported in Tables 5-1, 5-2, and 5-3, respectively. In Figure 5-1, we plot new compositional data on orthopyroxene core and rim pairs (each set of symbols connected by an arrow corresponds to one grain) from garnet websterites BC98-5 and BC98-7. Also shown is previously published data on garnet peridotite 1026V from Lee et al. (2001) and Chin et al. (2012) (these studies did not examine garnet websterite mineral zonation). Orthopyroxene in garnet websterites contain high-Al and high-Ca cores but low-Al and low-Ca rims, similar to garnet peridotite 1026V. BC98-5 has orthopyroxene rim Al₂O₃ as low as 0.7 wt.%, approaching
the low values observed in the garnet peridotite. New garnet-orthopyroxene thermobarometry using the thermobarometers of Harley and Green (1984) and Harley (1984) for BC98-5 indicate final equilibration pressures of 2.8 GPa and final temperatures of 760 °C. This pressure overlaps the range of final pressures recorded by Sierran garnet peridotites (3 – 3.5 GPa), and clearly show that garnet websterites and garnet peridotites coexisted at similar depths.

Figure 5-1. Al$_2$O$_3$ (wt.%) in orthopyroxene in garnet websterites BC98-7, BC98-5 and garnet peridotite 1026V

A wavelength-dispersive (WDS) EPMA elemental map of Al content in an orthopyroxene porphyroclast containing garnet exsolution lamellae is shown in Figure 5-2a. The orthopyroxene area shown here is part of a much larger (~1 cm) deformed orthopyroxene porphyroclast containing several garnet and amphibole lamellae, found in sample 1026V. Chin et al. (2012) observed a positive correlation between the thickness
of the Al-depleted halo next to garnet lamellae and the thickness of garnet lamellae, strongly suggesting that garnet lamellae exsolved under closed system conditions from originally hi-Al orthopyroxene. Much thinner, but more numerous, amphibole lamellae are also observed in Figure 5-2a. In contrast to the strong Al-depletion seen at the orthopyroxene-garnet contacts, weak to negligible Al-depletion is observed at orthopyroxene-amphibole contacts, suggesting that amphibole is either a late-nucleating phase or formed by open-system processes (Chin et al., 2012). Our interest here lies in the Al-depletion haloes at orthopyroxene-garnet contacts, and thus we do not discuss the amphibole lamellae further.
Figure 5-2. WDS Al map and Al diffusion profile. Top: WDS map of Al intensity. Bottom: WDS points along transect A-B from map above. Diffusion profiles at various times from 0.1 to 8 Ma are juxtaposed on the real data.
In Figure 5-2b, a transect of quantitative WDS points is plotted as a function of distance from the host orthopyroxene at the contact with the garnet lamella (endpoint B) to 20 µm into the host orthopyroxene interior (endpoint A). Within the orthopyroxene interior, Al\textsubscript{2}O\textsubscript{3} reaches a maximum of 4.58 wt. %; at the point closest to the garnet lamella, Al\textsubscript{2}O\textsubscript{3} is much lower (0.91 wt.%).

5.4.2. Lu-Hf and Sm-Nd geochemistry

Lu-Hf and Sm-Nd garnet, clinopyroxene, and whole-rock isotope data are presented in Table 5-4 and plotted in Figure 5-3. The slopes of lines regressed through the garnet, clinopyroxene, and/or whole-rock data in Figure 5-3 yield ages broadly coeval with the magmatic peak in the Sierra Nevada (100 to 85 Ma): for BC98-5, Lu-Hf age is 92.6 ± 1.6 Ma and Sm-Nd age is 88.8 ± 3.1 Ma; for BC98-7, Lu-Hf age is 60.8 ± 3.2 Ma and Sm-Nd age is 81.0 ± 2.0 Ma; and for BCX, Lu-Hf age is 107.6 ± 1.5 Ma and Sm-Nd age is = 84.0 ± 4.1 Ma. These results confirm the earlier Sm-Nd data of Ducea and Saleeby (1998).
Figure 5-3. Internal isochrons of garnet pyroxenites. Green circle = clinopyroxene, red triangle = garnet, white circle = whole rock.

For two samples, BC98-5 and BCX, the Lu-Hf system records an older age compared to the Sm-Nd system. There is a much greater time difference ($\Delta t = (\text{Lu-Hf age}) - (\text{Sm-Nd age})$) between Lu-Hf and Sm-Nd age for BCX ($\Delta t = 23.6$ Ma) compared to that of BC98-5 ($\Delta t = 3.8$ Ma). For one sample, BC98-7, the Lu-Hf age is considerably younger (by 20 Ma) compared to the Sm-Nd age. BC98-7 is the only pyroxenite we measured that contained amphibole. It is possible that the amphibole is not primary (cf. late-stage amphibole in some Sierran garnet peridotites), and thus the young Lu-Hf age may be associated with a metasomatic overprint.
5.5. Discussion

5.5.1. Constraints on rate of $\Delta T$ from Al-in-orthopyroxene diffusion

To summarize, several textural and compositional features related to Al zonation in both garnet websterites and garnet peridotites support cooling and/or increasing pressure. In websterites, orthopyroxenes are zoned from high-Al and high-Ca cores to low-Al and low-Ca rims. Similar zonation patterns are observed in garnet peridotites, particularly where garnet exsolution lamellae are involved (Figure 5-2). The next step is to estimate the rate of $\Delta T$ and $\Delta P$ recorded by the Al zonation in orthopyroxene, which will then be used to better constrain the thermal history of the deep Sierran arc lithosphere.

To model Al diffusion in orthopyroxene, we solved the general diffusion equation,

$$\frac{\partial C}{\partial t} = D(T) \frac{\partial^2 C}{\partial x^2}$$

using an explicit finite difference program using VisualBasic Macros in Excel developed by C-T Lee.

The temperature-dependent diffusion coefficient $D$ follows an Arrhenius relation of the form,

$$D = D_o \exp(-E/(RT))$$
where E is the activation energy (kJ/mol), R the gas constant, T the temperature, and $D_0$ the diffusion coefficient at infinitely high temperature. Al diffusion studies in pyroxene are limited, owing to the analytical challenges posed by the slow diffusion of Al. For our calculations, we used an activation energy $E$ between 350 and 400 kJ/mol (results corresponding to 375 kJ/mol are reported), which is based on the experiments of Sautter et al. (1988). We estimated $D_0$ using the data of Sautter et al. (1988) and Smith et al. (1991) to be $5 \times 10^{-11} \text{ m}^2/\text{s}$.

Because we are interested in modeling the exsolution of garnet lamellae in orthopyroxene and the resulting Al-depletion in the host orthopyroxene, our Al concentration boundary conditions must change with each time step as garnet is exsolved. To define the Al concentration in orthopyroxene in equilibrium with garnet at each time step, we parameterized the garnet-orthopyroxene thermobarometer of Harley and Green (1982) as a function of temperature at constant pressure (3 GPa) (Figure 5-4). We assumed constant Fe/Mg ratio in orthopyroxene and constant Ca content in garnet. Assuming a constant Fe/Mg ratio is reasonable because the Fe-Mg $K_D$ between orthopyroxene and olivine for typical mantle compositions is independent of pressure and temperature (von Seckendorff and O'Neill, 1993). In addition, CaO in low-Cr mantle garnets (typical of off-craton peridotite suites) is relatively uniform below ~5 wt.% (Pearson et al., 2003). Our parameterized boundary condition concentration function is represented by a polynomial function,

$$ y = (3E-08)T^2 - (4E-06)T - 0.0043 $$
where \( y = \text{Al concentration in units of XAl on the orthopyroxene M1 site (XAl = Al/2 in 6 oxygen unit orthopyroxene; Harley and Green (1982))} \) and \( T \) is temperature in Celsius.

**Figure 5-4.** The Harley and Green (1982) thermobarometer. Top, as a function of temperature at constant pressure. Bottom, as a function of pressure at constant temperatures.

Lastly, we used a temperature-time path that follows an exponential decay relationship,

\[
T = T_f + (T_0 + T_i) \exp(-kt)
\]
where $T$ = temperature at each time step, $T_f$ and $T_o$ were defined above, $k$ is a
time constant that represents the e-fold timescale for cooling, and $t$ is time.

Results of Al diffusion modeling for a transect between host
orthopyroxene and exsolved garnet lamella in peridotite 1026V is shown in Figure 5-2b.
In this particular case, we ran the model using an initial temperature $T_0$ of 1275 °C,
which at 3 GPa, corresponds to the maximum XAI of 0.0393 observed at the interior of
the host orthopyroxene. The final temperature was set to 750 °C, corresponding to the
average final temperatures of equilibration estimated using garnet-orthopyroxene
thermobarometry from Chin et al. (2012). Based on Figure 5-2b, the observed Al-
depletion halo formed within 6 to 8 Ma.

5.5.2. Differential closure of Lu-Hf and Sm-Nd chronometers and
implications for thermal models of the Sierran lithosphere

In our geospeedometry models above, we find that Al-depletion haloes between
orthopyroxene and exsolved garnet in the deepest-equilibrated Sierran peridotite formed
in less than 10 Ma, between 6 to 8 Ma. Pinning exactly when the Al-depletion haloes
formed, in an absolute timeframe, would constrain when the Sierran arc root cooled and
thickened. Did cooling and thickening occur during the main phase of arc magmatism
during the Late Cretaceous (100 – 85), or does it record a later event, perhaps associated
with the Laramide Orogeny that started after 85 Ma? Obviously, obtaining coupled Lu-
Hf and Sm-Nd ages on Sierran garnet peridotites might easily address this question.
Obtaining ages on the Sierran garnet peridotites, however, is difficult (if not impossible),
owing to the extremely low Hf (<0.05 ppm) concentrations in peridotite garnets, as well
as the textural complexity of garnet occurring as exsolved lamellae and coronas around spinels (making it extremely difficult to physically separate enough garnet from the bulk rock). However, coupled Lu-Hf and Sm-Nd age data are relatively much easier to obtain on garnet-rich pyroxenites. In addition, the high-Mg garnet websterites can be used as a “proxy” for garnet peridotite, because both websterites and peridotites record similar final equilibration P-T’s.

Older Lu-Hf ages are consistent with the Lu-Hf system having an equal or higher closure temperature compared to Sm-Nd (Scherer et al., 2000). Calculation of a Dodson-style closure temperature (Dodson, 1973), however, is often ambiguous owing to uncertainties in cooling rates, diffusivities of the elements of interest, mineral composition, and grain size. Although diffusion data exist for Sm-Nd in garnet, no such data is currently available for Lu-Hf in garnet, thus it is unclear how much closure temperatures between the two systems may vary. However, many studies on high-grade metamorphic rocks using coupled Lu-Hf and Sm-Nd garnet-clinopyroxene geochronology report Lu-Hf ages older than Sm-Nd ages (Lapen et al., 2003; Lapen et al., 2005; Kylander-Clark et al., 2007; Cheng et al., 2008), qualitatively supporting a higher temperature of Lu-Hf closure.

Two of our Sierran pyroxenite samples (BC98-5 and BCX) yield Lu-Hf ages that are older (by 3.8 and 23.6 Ma, respectively) compared to Sm-Nd ages. The other pyroxenite, BC98-7, shows the opposite trend, having a Lu-Hf age that is 20 Ma younger than the Sm-Nd age. Owing to the presence of (possibly secondary) amphibole in this sample, it is possible that the Lu-Hf may have been reset by a metasomatic event in
BC98-7. For the purposes of this study, we will not discuss BC98-7 further (I also have to think more about how this younger Lu-Hf age could have formed…). Below, we focus on BC98-5 and BCX.

The two pyroxenites BCX and BC98-5 represent shallow and deep stratigraphic levels of the Sierran arc root, respectively. BCX is undoubtedly of crustal origin: its petrology suggests it is a cumulate of a more differentiated melt compared to BC98-7 (Lee et al., 2006), and garnet in BCX have elevated, crust-like δ¹⁸O of 10.6 ‰ (Chin et al., in review). In comparison, garnets in websterites such as BC98-5 have mantle-like δ¹⁸O (5.5 ‰) (Chin et al., in review). Furthermore, BC98-5 records final P-T conditions analogous to Sierran garnet peridotites (3 GPa, ca. 750 C), indicating that the websterites formed within the mantle lithosphere.

The short time lag of only a few Ma between the Lu-Hf and Sm-Nd age in BC98-5 suggests that both chronometers closed at nearly the same time in this sample. In contrast, the large difference between the two chronometers in BCX suggests that a long period of time elapsed between Lu-Hf closure and Sm-Nd closure. One possibility is that BC98-5 experienced rapid cooling after it formed, thus locking in both radiometric ages almost simultaneously. BCX, on the other hand, experienced slower cooling such that 20 Ma elapsed between Lu-Hf closure at 107.6 Ma and Sm-Nd closure at 81 Ma. One scenario that can explain the age discrepancies is that as the lithospheric root thickened, it impinged upon the slab once it reached thicknesses ~3 GPa (e.g., typical depth to the slab). Slab impingement resulted in rapid cooling of the deepest part of the arc root, represented by the garnet websterites and the garnet peridotites, and thus explains why
BC98-5 has identical (to within error) Lu-Hf and Sm-Nd ages. In contrast, a cumulate formed further up in the lithosphere, such as BCX, did not experience this chilling effect from the slab and thus cooled slowly.

Figure 5-5. Thermal model results. The top panel shows various temperature-depth profiles through a 100-km thick lithosphere at various times from 0 to 70 Ma. At 0 Ma, the lithosphere impinges on the slab and is cooled immediately to 750°C. The bottom panel shows temperature-time paths of various “depth slices” through the lithosphere (ranging from the lower crust at 40 km to the slab-mantle interface).

To test this hypothesis, we performed a simple 1-D thermal diffusion model using the finite-difference technique with constant boundary conditions and no thermal
diffusivity was assumed to be 1E-06 m\(^2\)/s. As an initial condition, we defined a linear temperature decrease (1350 °C over a 100 km thick lithosphere) with an instantaneous drop to 750 °C at 100 km, approximating the slab effect. The model result is shown in Figure 5-5. Figure 5-5a shows temperature profiles through the lithosphere at various times from 0 Ma to 70 Ma. In Figure 5-5b, the temperature-time paths of various lithospheric depth slices are plotted. From this diagram, we see that a lithosphere slice at 95 km cools by several hundred degrees in 10 Ma, while a slice at 60 km depth has cooled by barely 50 degrees in the same time period.

5.6. Conclusions & implications

Diffusion modeling of Al depletion haloes between orthopyroxene and exsolved garnet in garnet peridotite suggests that such haloes formed within 6 to 8 Ma, suggesting a relatively rapid cooling of the deepest mantle lithosphere beneath the Sierra Nevada. This timescale is within the range predicted by thermal models of the deep lithosphere at depths between 80 and 100 km, where garnet peridotites equilibrated. In addition, thermal models support the hypothesis that slab-impingement rapidly cooled the deepest arc lithosphere, resulting in a small difference between Lu-Hf and Sm-Nd ages in garnet websterites. In contrast, protracted cooling in the shallower lithosphere (mid- to lower crust) led to a substantial lag between Lu-Hf closure and Sm-Nd closure in shallower crustal pyroxenites.

Growth of the arc root by “downward thickening” until the root impinges on the slab at 3 GPa (Figure 5-6a,b) could be one scenario that can explain the Lu-Hf
and Sm-Nd age systematics. If this were the case, it implies that the arc root already extended to ~3 GPa at the time websterite BC98-5 was formed at ~90 Ma. This is not implausible, given that the most voluminous arc magmatism occurred at this time. Chilling by the slab could also be the trigger for garnet exsolution in orthopyroxene. In Figure 5-6c, we show schematic P-T paths of the deep Sierran arc lithosphere based on thermobarometry of Chin et al. (2012). Three possible P-T paths are outlined: 1 – isobaric cooling followed by rapid, isothermal thickening; 2 – coupled cooling and thickening; and 3 – isothermal thickening followed by isobaric cooling. Although we have not yet modeled each of these specific scenarios, we can rule out P-T path 1 because garnet could not exsolve from orthopyroxene during the (<3 GPa) isobaric cooling stage. Moreover, it would be difficult to exsolve garnet during isothermal thickening because at this point, temperatures would be too cold. Thus, P-T paths between 2 and 3 are more likely.

Figure 5-6. Schematic illustrating downward thickening and slab impingement

![Figure 5-6](image-url)
In the context of regional geology, our study challenges the common view that flattening of the Farallon slab beneath Western North America was responsible for cessation of arc magmatism. Instead, we suggest, as Saleeby (2003) did, that no such flattening of the slab occurred beneath the Sierra Nevada. Magmatic inflation and tectonic thickening, both processes associated with the upper plate, were sufficient to thicken an arc root until it reached slab depths, resulting in pinching out of the asthenospheric wedge and cooling of the deep lithosphere.

| TABLE 5-1. Clinopyroxene major element compositions in garnet pyroxenites |
|--------------------------|-----------------|-----------------|-----------------|
| ID          | BCX_03-cpx     | BCX_05-cpx     | BCX_08-cpx     |
|             | rim next to gt | rim next to gt | rim next to gt |
| x           | -6             | -5075          | -4444          |
| y           | -15851         | -6058          | -13636         |
| z           | 7              | -39            | -12            |
| SiO₂        | 54.85          | 54.80          | 54.53          |
| TiO₂        | 0.184          | 0.186          | 0.205          |
| Al₂O₃       | 9.34           | 9.42           | 9.24           |
| Cr₂O₃       | 0.042          | 0.004          | 0              |
| FeOt        | 4.14           | 4.17           | 4.19           |
| MnO         | 0.002          | 0.029          | 0.037          |
| MgO         | 10.57          | 10.57          | 10.60          |
| CaO         | 17.04          | 16.762         | 16.97          |
| Na₂O (wt%)  | 4.61           | 4.70           | 4.53           |
| Total       | 100.78         | 100.64         | 100.30         |

| TABLE 5-1. Clinopyroxene major element compositions in garnet pyroxenites |
|--------------------------|-----------------|-----------------|-----------------|
| ID          | BCX98-7         | BCX98-5         |
|             | BC98_05-cpxR    | BC98_05-cpxC    | BC98_09-cpxR    |
|             | rim             | core            | rim             |
| x           | 12686           | 12656           | 769             |
| y           | 14152           | 14420           | 8874            |
| z           | 151             | 151             | 140             |
| SiO₂        | 53.79           | 53.62           | 54.45           |
| TiO₂        | 0.146           | 0.242           | 0.088           |
| Al₂O₃       | 3.906           | 4.491           | 3.558           |
| Cr₂O₃       | 0.437           | 0.435           | 0.451           |
| FeOt        | 3.73            | 4.04            | 3.50            |
| MnO         | 0.084           | 0.056           | 0.033           |
| MgO         | 15.00           | 14.50           | 15.16           |
| CaO         | 21.81           | 21.38           | 21.83           |
| Na₂O (wt%)  | 1.61            | 1.67            | 1.65            |
| Total       | 100.51          | 100.44          | 100.72          |

<p>| TABLE 5-1. Clinopyroxene major element compositions in garnet pyroxenites |
|--------------------------|-----------------|-----------------|-----------------|
| ID          | BCX98-5         |
|             | BC98_05-cpxR    | BC98_05-cpxC    | BC98_15-cpxR    |
|             | rim             | core            | rim             |
| x           | 4075            | 5089            | 7454            |
| y           | 23847           | 24479           | 8124            |
| z           | 105             | 105             | 93              |
| SiO₂        | 54.35           | 53.591          | 53.408          | 54.057          | 52.982 |</p>
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Table 5-1. Clinopyroxene major element compositions in garnet pyroxenites

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Table 5-2. Orthopyroxene major element compositions in garnet websterites
### Table 5-3. Garnet major element compositions in garnet pyroxenites

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Table 5-4. Lu-Hf and Sm-Nd isotope data of garnet pyroxenites
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## Appendix A: Petrographic Atlas of Xenoliths

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Plate IA. Spinel peridotites

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200 μm thick section

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Top: 200 μm thick
Middle: 30 μm thin section
(plane light)
Bottom: 30 μm thin section
(cross-polarized)
Plate IB. Spinel peridotites

08BC06
Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized light)

08BC10
Plate IC. Spinel peridotites

BC10-6

Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized)
Plate ID. Spinel peridotites

BC10-1

Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized)
Plate IE. Spinel peridotites

BC10-2
Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized)

BC10-3
Plate IF. Spinel peridotites

BC10-4
Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized light)

BCPL10-5
Plate IG. Spinel peridotites

**P-7**
200 \( \mu \text{m} \) thick section

**BC98-2**
200 \( \mu \text{m} \) thick section
Plate 2A. Garnet-bearing spinel peridotites

08BC03

08BC04

Top: 200 \( \mu \)m thick section (plane light)
Middle: 30 \( \mu \)m thin section (plane light)
Bottom: 30 \( \mu \)m thin section (crossed polarized) light)
Plate 2B. Garnet-bearing spinel peridotites

08BC05  08BC07

Top: 200 μm thick section (plane light)
Middle: 30 μm thin section (plane light)
Bottom: 30 μm thin section (crossed polarized light)
Plate 2C. Garnet-bearing spinel peridotites

08BC08
Top: 200 μm thick section
Middle: 30 μm thin section (plane-polarized light)
Bottom: 30 μm thin section (cross-polarized)

08BC09
Plate 2D. Garnet-bearing spinel peridotites

1026V
30 μm thin section (plane light)

08BC14
30 μm thin section
left: plane light
right: cross-polarized
Plate 2E. Garnet-bearing spinel peridotites

**BC77**
Top: 30 μm thin section (plane light)
Bottom: 30 μm thin section (cross-polarized)

**P-6**
Top: 200 μm thick section (plane light)
Bottom: 200 μm thick section (cross-polarized)
Plate 2F. Garnet-bearing spinel peridotites

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Middle: 30 μm thin section (plane-polarized)
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Plate 3A. Garnet pyroxenites

08BC16

08BC17

Top: 200 µm thick
Middle: 30 µm thin section (plane light)
Bottom: 30 µm thin section (cross-polarized)

Scale bar same in all photos
Plate 3B. Garnet pyroxenites

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Scale bar same in all photos

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Plate 3C. Garnet pyroxenites

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Top: 200 μm thick section
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Top: 30 μm thin section (plane light)
Bottom: 30 μm thin section (cross-polarized)

Scale bar same in all photos
Plate 3D. Garnet pyroxenites

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Scale bar same in all photos
Plate 3F. Garnet pyroxenites

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Scale bar same in all photos
Plate 4A. Granulitic xenoliths

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Top: 200 μm thick section
Middle: 30 μm thin section (plane light)
Bottom: 30 μm thin section

08BC21
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Bottom: 30 μm thin section (cross-polarized)
Plate 4B. Granulitic xenoliths

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Top: 200 μm thick section
Middle: 30 μm thin section (plane light)
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08BC25A
Scale bar same in all photos
Plate 4C. Granulitic xenoliths

BC127
Top: 30 µm thin section (plane light)
Bottom: 30 µm thin section (cross-polarized)
Plate 5A. Metasedimentary xenoliths

Scale bar same in all photos

Left: 200 µm thick section

Top row: 30 µm thin section (plane light)
Bottom row: 30 µm thin section (cross-polarized)
Plate 5B. Metasedimentary xenoliths

BC98-8
Left: 200 μm thick section

Top row: 30 μm thin section (plane light)
Bottom row: 30 μm thin section (cross-polarized)

BC98-6
Left: 200 μm thick section
Plate 5C. Metasedimentary xenoliths

**BC134**
Left: 200 μm thick section

Top row: 30 μm thin section (plane light)
Bottom row: 30 μm thin section (cross-polarized)

**BC135**
Left: 200 μm thick section
Plate 6A. Garnet peridotites from Lethlakane mine (Botswana) and Venetia mine (South Africa)

LETH-1

VEN-1

LETH-3

VEN-2

All photos are 200 μm thick sections
Scale bar same in all photos
Appendix B: WDS maps of 1026V
Appendix C: Location maps of WDS quants in 1026V

These image maps of sample 1026V give the locations of quantitative WDS points reported in Table A-6, A-7, and A-8.
Area “BigOPX”
WDS quants locations
## Appendix D: Electron microprobe measurements of minerals in Sierran xenoliths

### TABLE A-1. Orthopyroxene probe measurements

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**2026V**

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**Descrip.**

- **SiO₂**: Silicon oxide content.
- **MgO**: Magnesium oxide content.
- **MnO**: Manganese oxide content.
- **Al**: Aluminum oxide content.
- **TiO₂**: Titanium oxide content.
- **FeOT**: Iron oxide content.
- **CaO**: Calcium oxide content.
- **Na₂O**: Sodium oxide content.
- **K₂O**: Potassium oxide content.
- **Total**: Total oxide content.

**Descrip.**

- **core**: Core of the sample.
- **rim**: Rim of the sample.

**Descrip.**

- **large opx with gt exsolution**: Sample with large opx and gt exsolution.
- **large, discrete grain, primary opx, no exsolution**: Sample with large, discrete grain, primary opx, no exsolution.
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**TABLE A.5. Spinel probe measurements**

- TiO₂: 0.157
- Al₂O₃: 41.35
- Cr₂O₃: 27.66
- FeO: 1.68
- MnO: 0.15
- MgO: 17.63
- V₂O₅: 0.158
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| Total | 100.87 | 100.87 | 100.87 | 100.87 |

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| Total | 100.69 | 100.53 | 100.51 | 99.96 |

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| Total | 100.69 | 100.53 | 100.51 | 99.96 |

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| Total | 100.69 | 100.53 | 100.51 | 99.96 |

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NiO 0.166 0.115 0.145 0.205 0.141 0.191
ZnO 0.122 0.159 0.208 0.303 0.173 0.083
Total 100.52 100.79 100.50 101.23 100.73 100.83

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**TABLE A-6 cont'd. Garnet and orthopyroxene in 1026V - "BigOPX" (Refer to Appendix C for position #a)**

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**TABLE A-6 cont'd. Garnet and orthopyroxene in 1026V - "BigOPX" (Refer to Appendix C for position #a)**

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<th>FeO</th>
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### Transect 1B

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**TABLE A-6 cont'd.** Garnet, orthopyroxene, & amphibole in 1026V - "BigOPX" (Refer to Appendix C for position #s)

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**TABLE A-6 cont'd.** Garnet, orthopyroxene, & amphibole in 1026V - "BigOPX" (Refer to Appendix C for position #s)

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**TABLE A-6 cont'd.** Garnet, orthopyroxene, & amphibole in 1026V - "BigOPX" (Refer to Appendix C for position #s)
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Position # 1F #1 1F #2 1G #1 1G #2
Acc V (kV) 15 15 15 15
Current (A) 1.508E-08 1.507E-08 1.516E-08 1.508E-08
X mm 32.6202 32.7262 32.7382
Y mm 26.2183 26.2563 26.2623

Position #: 1H #1 1H #2 1H #3 1H #4
Acc V (kV) 15 15 15 15
Current (A) 1.001E-08 1.001E-08 9.998E-09 1.001E-08
X mm 32.3749 32.3769 32.4299 32.4329
| SiO$_2$ | 44.05 | 44.00 | 44.35 | 44.50 |
| TiO$_2$ | 0.261 | 0.249 | 0.228 | 0.211 |
| Cr$_2$O$_3$ | 1.50 | 1.54 | 1.24 | 1.75 |
| FeOT | 3.18 | 3.57 | 3.28 | 2.91 |
| MnO | 0.08 | 0.105 | 0.015 | 0.023 |
| MgO | 18.91 | 18.54 | 18.99 | 18.97 |
| CaO | 11.34 | 11.11 | 10.71 | 10.94 |
| Na$_2$O | 3.213 | 3.146 | 3.203 | 3.289 |
| K$_2$O | 0.184 | 0.173 | 0.173 | 0.132 |
| Total | 97.58 | 96.78 | 96.67 | 97.48 |

**TABLE A-7. Garnet, orthopyroxene, & amphibole in 1026V - "CpxGL" (Refer to Appendix C for position #s)***

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**Detailed analysis of mineral contents in Table A-7:**

- SiO$_2$ content varies between 44.00 to 44.50, indicative of the presence of silica minerals.
- TiO$_2$ content ranges from 0.21 to 0.261, possibly indicating the presence of titania minerals.
- Al$_2$O$_3$ content varies between 13.90 to 14.42, suggesting aluminum oxide is present.
- Cr$_2$O$_3$ content ranges from 1.24 to 1.75, indicating chromium oxide.
- FeOT content ranges from 2.91 to 3.57, indicating ferrous oxide.
- MnO content ranges from 0.015 to 0.105, indicating manganese oxide.
- MgO content ranges from 18.97 to 18.99, indicating magnesium oxide.
- CaO content ranges from 10.71 to 10.94, indicating calcium oxide.
- Na$_2$O content ranges from 3.146 to 3.289, indicating sodium oxide.
- K$_2$O content ranges from 0.132 to 0.173, indicating potassium oxide.

Further analysis indicates the presence of orthopyroxene and amphibole phases, with specific compositions detailed in the table. The detailed analysis includes specific compositions and percentages of each mineral, providing a comprehensive overview of the mineralogy of the sample.
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### TABLE A-8. Garnet, orthopyroxene, spinel, & amphibole in 1026V - "OpsChin" - see Appendix C for position #s

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| SiO₂ | 44.67 | 46.22 | 44.36 | 44.38 | 22.64 | 22.63 | 22.55 | 22.67 | 22.71 | 22.79 | 4.96   | 1.54   | 21.49 |
| TiO₂ | 0.22  | 0.21  | 0.19  | 0.25  | 1.16  | 1.07  | 1.06  | 1.15  | 1.16  | 1.18  | 0.72   | 0.89   | 1.68  |
| Al₂O₃| 11.97 | 10.67 | 12.38 | 12.39 | 1.21  | 1.66  | 1.48  | 1.49  | 1.53  | 1.58  | 1.68   | 0.72   | 0.89  |
| Cr₂O₃| 1.96  | 1.75  | 1.69  | 1.69  | 2.82  | 3.80  | 3.35  | 3.32  | 3.32  | 3.32  | 3.32   | 3.32   | 3.32  |
| FeOT | 0.06  | 0.08  | 0.09  | 0.14  | 0.06  | 0.08  | 0.09  | 0.14  | 0.06  | 0.08  | 0.09   | 0.14   | 0.06  |
| MgO  | 19.02 | 20.79 | 19.17 | 18.74 | 19.02 | 20.79 | 19.17 | 18.74 | 19.02 | 20.79 | 19.17   | 18.74   | 18.74 |
| CaO  | 11.58 | 10.32 | 11.56 | 11.45 | 2.63  | 2.26  | 2.26  | 2.84  | 2.63  | 2.26  | 2.26   | 2.84   | 2.84  |
| Na₂O | 0.07  | 0.07  | 0.22  | 0.16  | 1.01  | 0.90  | 0.96  | 1.16  | 1.01  | 0.90  | 0.96   | 1.16   | 1.16  |
| Total | 96.01 | 97.06 | 96.23 | 96.52 | 96.01 | 97.06 | 96.23 | 96.52 | 96.01 | 97.06 | 96.23   | 96.52   | 96.52 |

### Transect IB


**Note:** This table includes data on the abundance of various minerals in a specific sample or transect. The data is presented in a tabular format with columns for position numbers, abundance percentages, and some additional details such as the position of the sample and the type of minerals present. The data is analyzed to provide insights into the composition and distribution of these minerals within the sample or transect.
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| SiO₂      | 56.80    | 56.28    | 56.26    |
| TiO₂      | 0.00     | 0.02     | 0.02     |
| Al₂O₃     | 2.32     | 2.47     | 2.61     |
| Cr₂O₃     | 0.76     | 0.71     | 0.59     |
| FeO       | 5.92     | 5.89     | 5.76     |
| MnO       | 0.15     | 0.14     | 0.17     |
| MgO       | 34.77    | 34.95    | 34.71    |
| CaO       | 0.17     | 0.15     | 0.16     |
| Na₂O      | 0.03     | 0.00     | 0.00     |
| NiO       | 0.09     | 0.07     | 0.12     |
| K₂O       | 0.010    | 0.012    | 0.001    |
| Total     | 101.02   | 100.68   | 100.37   |

### Transect I

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| SiO₂      | 42.02    | 42.11    | 42.71    | 41.50     | 41.15     | 41.25     |
| TiO₂      | 0.04     | 0.06     | 0.01     | 0.00      | 0.00      | 0.02      |
| Al₂O₃     | 22.39    | 22.49    | 23.09    | 0.00      | 0.02      | 0.00      |
| Cr₂O₃     | 1.41     | 1.58     | 1.46     | 0.00      | 0.04      | 0.07      |
| FeO       | 10.31    | 10.20    | 10.80    | 10.51     | 10.07     | 9.39      |
| MnO       | 0.56     | 0.66     | 0.47     | 0.20      | 0.22      | 0.24      |
| MgO       | 17.90    | 18.07    | 18.27    | 48.92     | 49.37     | 49.75     |
| CaO       | 5.24     | 5.34     | 4.94     | 0.02      | 0.03      | 0.01      |
| Na₂O      | 0.03     | 0.00     | 0.00     | 0.25      | 0.40      | 0.38      |
| K₂O       | 0.01     | 0.012    | 0.001    | 0.01      | 0.01      | 0.01      |
| Total     | 99.90    | 100.50   | 101.76   | 101.39    | 101.28    | 101.10    |

### Transect E

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<p>| SiO₂      | 42.20    | 0.08     | 0.12     | 49.71     | 37.85     | 35.58     | 0.10      | 0.10      |
| TiO₂      | 0.01     | 0.00     | 0.00     | 0.02      | 0.28      | 0.23      | 0.02      | 0.01      |
| Al₂O₃     | 22.83    | 57.86    | 59.07    | 7.98      | 11.38     | 17.78     | 57.58     | 62.69     |
| Cr₂O₃     | 1.34     | 6.32     | 5.24     | 1.72      | 3.81      | 5.75      | 8.51      | 3.91      |
| FeO       | 10.92    | 16.65    | 15.52    | 8.06      | 4.88      | 5.22      | 13.11     | 11.80     |
| MnO       | 0.55     | 0.51     | 0.46     | 0.23      | 0.01      | 0.16      | 0.19      | 0.21      |
| MgO       | 18.25    | 18.16    | 18.31    | 28.71     | 16.44     | 16.05     | 18.95     | 20.21     |
| CaO       | 4.76     | 0.00     | 0.01     | 0.64      | 9.25      | 8.87      | 0.01      | 0.00      |
| Na₂O      | 0.00     | 0.05     | 0.09     | 0.08      | 0.09      | 0.02      | 0.04      | 0.05      |
| K₂O       | 0.01     | 0.012    | 0.001    | 0.01      | 0.01      | 0.01      | 0.01      | 0.01      |
| Total     | 100.86   | 99.63    | 98.80    | 97.14     | 83.99     | 89.65     | 98.50     | 98.38     |</p>
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### Appendix E: LA ICP-MS measurements of minerals in Sierran xenoliths

**TABLE A-9.** LR-LA-ICPMS measurements of garnet in peridotites (values in ppm)

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**Table A-10.** LR-ICP-MS measurements of clinopyroxene in Sierran peridotites (values in ppm)

**P-7**

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TABLE A-11. LR-ICP-MS measurements of orthopyroxene in Sierran peridotites (values in ppm)

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05208BC01

04808BC01

04908BC01

08BC06
01108BC06

05508BC01
rim

01208BC06

01308bc06
rim

01408bc06

03508bc06

03608bc06

Rb

0.0183

0.0070

0.0590

0.0112

0.0406

0.0344

0.0239

0.0426

0.0457

0.0334

Sr
Y

1.09
0.388

0.31
0.542

0.53
0.442

0.84
0.722

2.31
0.212

1.29
0.185

0.06
0.184

0.19
0.245

0.15
0.285

0.20
0.219

Zr
Nb

2.83
0.071

1.61
0.054

1.52
0.057

2.07
0.066

0.480
0.114

0.333
0.098

0.374
0.092

0.321
0.074

0.323
0.140

0.320
0.060

Cs
Ba

0.002
0.578

0.001
0.061

0.043
0.184

0.001
0.078

0.019
0.878

0.012
0.492

0.001
0.095

0.009
0.236

0.002
0.055

0.003
0.159

La
Ce

0.074
0.124

0.031
0.050

0.033
0.066

0.070
0.125

0.017
0.044

0.012
0.023

0.002
0.006

0.005
0.010

0.004
0.024

0.005
0.007

Pr
Nd

0.012
0.046

0.007
0.038

0.010
0.025

0.015
0.059

0.004
0.016

0.002
0.019

0.002
0.030

0.002
0.018

0.004
0.012

0.003
0.009

Sm
Eu

0.019
0.008

0.015
0.009

0.015
0.007

0.023
0.008

0.008
0.005

0.007
0.005

0.008
0.006

0.009
0.005

0.008
0.004

0.005
0.004

Gd
Tb

0.033
0.007

0.031
0.008

0.024
0.006

0.042
0.010

0.010
0.003

0.011
0.005

0.008
0.008

0.017
0.005

0.015
0.003

0.011
0.002

Dy
Ho

0.056
0.015

0.073
0.021

0.058
0.015

0.092
0.025

0.024
0.008

0.018
0.007

0.023
0.007

0.026
0.009

0.026
0.009

0.023
0.008

Er
Tm

0.054
0.009

0.084
0.012

0.067
0.010

0.099
0.017

0.033
0.011

0.034
0.040

0.035
0.068

0.036
0.024

0.039
0.007

0.035
0.006

Yb
Lu

0.095
0.020

0.120
0.028

0.113
0.025

0.168
0.035

0.071
0.015

0.058
0.012

0.066
0.017

0.071
0.015

0.070
0.019

0.070
0.016

Hf
Ta

0.080
0.007

0.056
0.009

0.056
0.008

0.067
0.005

0.018
0.011

0.020
0.037

0.019
0.061

0.026
0.021

0.025
0.008

0.016

Pb
Th

0.102
0.109

0.036
0.036

0.024
0.057

0.123
0.042

0.063
0.005

0.059
0.006

0.058
0.003

0.023
0.002

0.030
0.002

0.055
0.004

U

0.097

0.046

0.056

0.050

0.009

0.009

0.001

0.002

0.003

0.003

TABLE A-12. LR-ICP-MS measurements of olivine in Sierran peridotites (values in ppm)
08BC01
ID
Descrip

08BC06

051-08BC01

017-08bc06

018-08bc06

Rb

0.0071

0.0243

0.0165

Sr
Y

0.0259
0.0071

0.0468
0.0067

0.0299
0.0053

Zr
Nb

0.1532
0.0075

0.0375
0.0061

0.0524
0.0054

Cs
Ba

0.0010
0.0430

0.0005
0.1979

0.0007
0.0460

La
Ce

0.0026
0.0089

0.0017
0.0037

0.0016
0.0181

Pr
Nd

0.0042
0.0105

0.0015
0.0122

0.0012
0.0124

Sm
Eu

0.0075
0.0049

0.0071
0.0026

0.0043
0.0038

Gd
Tb

0.0072
0.0019

0.0047
0.0014

0.0059
0.0021

Dy
Ho

0.0009

0.0050
0.0010

0.0041
0.0006

Er
Tm

0.0044

0.0030
0.0015

0.0029
0.0019

Yb
Lu

0.0115
0.0030

0.0100
0.0022

0.0150

Hf
Ta

0.0132
0.0059

0.0044

0.0037

Pb
Th

0.0028
0.0049

0.0180

0.0245
0.0033

U

0.0028

0.0015

0.0011


Appendix F: Compositional data on cumulate xenoliths from the Engorora Crater, Tanzania

TABLE A-13. LR-ICP-MS measurements of minerals in Engorora cumulate xenolith ENG (values in ppm)

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| AlO, | 0.03  | 0.03  | 0.01  | 0.01  | 0.02  | 0.02  | 0.02  | 0.02  | 0.03 |
| MnO, | 0.25  | 0.23  | 0.20  | 0.19  | 0.20  | 0.22  | 0.19  | 0.20  | 0.20 |
| MgO, | 40.24 | 40.23 | 45.43 | 45.74 | 45.42 | 45.41 | 45.51 | 45.60 | 45.48 |
| CaO, | 0.13  | 0.14  | 0.19  | 0.05  | 0.18  | 0.15  | 0.18  | 0.18  | 0.19 |
| NaO, | 0.00  | 0.02  | 0.02  | 0.00  | 0.02  | 0.00  | 0.01  | 0.01  | 0.01 |
| NiO, | 0.29  | 0.33  | 0.23  | 0.35  | 0.26  | 0.28  | 0.23  | 0.25  | 0.26 |
| Total | 100.37 | 100.45 | 100.00 | 100.25 | 100.24 | 100.22 | 100.25 | 100.50 | 100.84 |

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| SiO, | 39.68 | 39.43 | 39.60 | 39.59 | 38.45 | 38.38 | 38.41 | 38.57 | 40.33 |
| AlO, | 0.02  | 0.01  | 0.01  | 0.02  | 0.04  | 0.02  | 0.04  | 0.02  | 0.04 |
| MnO, | 0.18  | 0.19  | 0.19  | 0.19  | 0.25  | 0.24  | 0.24  | 0.25  | 0.16 |
| MgO, | 45.61 | 45.85 | 45.87 | 45.69 | 40.50 | 40.35 | 40.46 | 40.10 | 47.45 |
| CaO, | 0.17  | 0.06  | 0.06  | 0.07  | 0.20  | 0.13  | 0.17  | 0.13  | 0.20 |
| NaO, | 0.00  | 0.01  | 0.04  | 0.02  | 0.02  | 0.03  | 0.00  | 0.00  | 0.03 |
| NiO, | 0.21  | 0.38  | 0.38  | 0.38  | 0.28  | 0.31  | 0.30  | 0.30  | 0.31 |
| Total | 100.52 | 100.36 | 100.59 | 100.51 | 100.04 | 100.79 | 100.47 | 100.75 | 100.61 |

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<p>| SiO, | 40.14 | 40.18 | 40.26 | 39.86 | 39.80 | 39.71 |
| AlO, | 0.02  | 0.03  | 0.03  | 0.02  | 0.02  | 0.02 |
| FeOT, | 11.89 | 11.84 | 11.73 | 13.19 | 13.27 | 13.31 |
| MnO, | 0.18  | 0.18  | 0.16  | 0.20  | 0.20  | 0.18 |
| MgO, | 47.51 | 47.36 | 47.90 | 46.41 | 46.65 | 46.84 |
| CaO, | 0.31  | 0.20  | 0.20  | 0.19  | 0.19  | 0.20 |
| NaO, | 0.02  | 0.03  | 0.03  | 0.02  | 0.01  | 0.00 |
| NiO, | 0.31  | 0.32  | 0.31  | 0.24  | 0.27  | 0.26 |
| Total | 100.37 | 100.13 | 100.62 | 100.13 | 100.39 | 100.52 |</p>
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**TABLE A.15.** Electron microprobe measurements of clinopyroxenes in Engorora xenoliths
TABLE A-16. Electron microprobe measurement of melt pocket in Engorora xenolith ENG-14

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TABLE A-17. Electron microprobe measurements of spinel in Engorora xenoliths

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TABLE A-18. Electron microprobe measurements of orthopyroxene in Engorora xenolith ENG-7

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**TABLE A19. Whole rock XRF data on Engorora xenoliths**

| SiO<sub>2</sub> | 52.82 | 53.17 | 52.96 | 53.04 | 52.79 | 52.92 |
| TiO<sub>2</sub> | 0.13  | 0.11  | 0.21  | 0.17  | 0.11  | 0.21  |
| Al<sub>2</sub>O<sub>3</sub> | 2.22  | 2.32  | 2.30  | 2.23  | 2.15  | 2.37  |
| Cr<sub>2</sub>O<sub>3</sub> | 0.23  | 0.14  | 0.20  | 0.22  | 0.19  | 0.20  |
| FeO | 8.49  | 9.12  | 8.79  | 8.76  | 8.38  | 8.61  |
| Fe<sub>2</sub>O<sub>3</sub> | 4.81  | 4.15  | 4.61  | 4.81  | 4.96  | 4.81  |
| MnO | 0.22  | 0.20  | 0.29  | 0.23  | 0.25  | 0.25  |
| MgO | 29.79 | 29.95 | 29.48 | 29.68 | 29.86 | 29.61 |
| CaO | 0.90  | 0.62  | 1.20  | 1.05  | 0.84  | 1.17  |
| Na<sub>2</sub>O | 0.06  | 0.04  | 0.06  | 0.07  | 0.06  | 0.07  |
| K<sub>2</sub>O | 0.01  | 0.00  | 0.02  | 0.00  | 0.00  | 0.00  |
| Total | 99.67 | 99.81 | 100.11| 100.05| 99.59 | 100.32|

**Ni** 419  462  458  717  409  577  513  2603  798  609  1092
**Cr** 3113  2939  2933  3050  1446  2365  2554  820  2219  2979  2234
**Sc** 45  45  45  24  56  38  49  5  37  40  25
**V** 191  214  210  134  253  213  49  199  168  160
**Ba** 206  195  196  412  173  281  118  39  84  258  88
**Rb** 2  4  5  14  6  5  6  2  3  12  6
**Sr** 211  129  127  714  316  169  301  43  190  1053  108
**Zr** 20  35  35  50  22  34  31  14  20  28  25
**Y** 9  11  10  11  12  10  11  3  9  8  6
**Nb** 3.1  5.5  4.6  8.5  6.6  6.9  4.8  5.6  4.2  7.4  7.8
**Ga** 5  8  7  6  6  4  5  2  4  5  3
**Cu** 41  95  97  82  117  294  158  38  127  154  100
**Zn** 50  47  46  103  45  43  49  85  77  47  75
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Appendix G: Trace element compositions of minerals in garnet peridotites from Venetia and Lethlakane mines, South Africa

<p>| TABLE A-20. MR-ICP-MS measurements of minerals in 2 garnet peridotites from Venetia and Lethlakane |
|----------------------------------|---------------------------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
| ID  | 15-LET3   | 16-LET3   | 17-LET3 incl in gt | 21-LET3 | 22-LET3 | 23-LET3 | 25-LET3 | 27-LET3 of rim near gt | 29-LET3 | 31-LET3 | 33-LET3 |
|     | gt rim    | gt rim    | al-rich opx (%)    | gt core  |          |          |          |                       |          |          |          |
| Na  | 163       | 279       | 815               | 139      | 165      | 62       | 13      | 9            | 64      | 935        | 9572          |
| Mg  | 127936    | 131016    | 260428            | 131149   | 124645   | 312867   | 315531  | 299306        | 297620  | 108642     | 110995        |
| Al  | 111217    | 100932    | 26005             | 105357   | 98736    | 91       | 12      | 7            | 32      | 8077       | 9548          |
| Si  | 179823    | 183266    | 170382            | 184257   | 192376   | 164052   | 16133   | 177262        | 177956  | 231186     | 222359        |
| P   | 183.8     | 185.5     | 65.0              | 321.0    | 297.4    | 45.9     | 41.2    | 41.2          | 45.1    | 33.5       | 47.8          |
| K   | 26.8      | 337.2     | 272.2             | 12.2     | 24.1     | 7.2      | 1.9     | 5.7           | 17.9    | 105.7      | 27.3          |
| Ca  | 26629     | 24730     | 3045              | 2683     | 26126    | 125      | 85      | 100           | 166     | 150372     | 158413        |
| Sc  | 65.0      | 73.8      | 0.537             | 61.2     | 0.377    | 0.296    | 0.393   | 0.482         | 28.6    | 30.6       |               |
| Ti  | 579       | 563       | 107               | 554      | 524      | 74.6     | 74.4    | 72.5          | 79.6    | 138        | 1426          |
| V   | 161       | 167       | 3.51              | 159      | 152      | 2.48     | 2.39    | 2.58          | 2.81    | 388        | 366           |
| Cr  | 7582      | 7169      | 33.8              | 8834     | 8296     | 31.6     | 27.3    | 30.1          | 32.3    | 6435       | 6962          |
| Mn  | 3672      | 3342      | 771               | 3445     | 3192     | 757      | 817     | 823           | 835     | 414        | 419           |
| Fe  | 65162     | 73484     | 10579             | 61182    | 65966    | 69365    | 71274   | 63071         | 67368   | 11475      | 12290         |
| Co  | 41.2      | 47.9      | 165               | 38.7     | 43.4     | 135      | 133     | 123           | 126     | 14.9       | 15.0          |
| Ni  | 21.2      | 21.7      | 3720              | 21.6     | 22.0     | 3271     | 3187    | 2969          | 3016    | 323        | 340           |
| Cu  | 0.208     | 0.233     | 0.315             | 0.130    | 0.506    | 0.442    | 0.437   | 0.697         | 0.498   | 0.489      |               |
| Zn  | 6.61      | 6.80      | 44.4              | 6.85     | 6.73     | 39.4     | 37.9    | 46.2          | 45.5    | 6.58       | 8.11          |
| Ga  | 8.05      | 7.59      | 6.80              | 6.88     | 7.69     | 0.075    | 0.046   | 0.078         | 0.079   | 5.53       | 5.36          |
| Ge  | 2.07      | 2.21      | 8.57              | 1.66     | 1.86     | 0.771    | 0.624   | 0.324         | 0.525   | 1.55       | 1.33          |
| Y   | 2.07      | 2.87      | 0.698             | 2.10     | 2.27     | 0.456    | 0.422   | 0.645         | 0.619   | 0.973      | 1.26          |
| Zr  | 15.7      | 16.4      | 16.0              | 16.1     | 0.466    |          |        |               |         |            |               |</p>
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Note: The values in the table represent different properties or concentrations of the elements.
Appendix H: Whole-rock XRF data on Sierran garnet pyroxenites and granulitic xenoliths

Table A-22. Whole rock geochemistry of Sierran garnet pyroxenites

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| Cr     | 3310   | 275    | 133    | 153    | 123    | 928    | 149        | 62         | 56         | 68         | 3645   | 17     |
| Sc     | 30     | 53     | 68     | 28     | 64     | 42     | 49         | 59         | 57         | 66         | 49     | 33     |
| V      | 172    | 367    | 294    | 499    | 337    | 260    | 411        | 295        | 309        | 257        | 307    | 286    |
| Ba     | 136    | 302    | 153    | 453    | 146    | 951    | 209        | 107        | 575        | 81         | 53     | 286    |
| Rb     | 2      | 5      | 13     | 4      | 9      | 28     | 10         | 8          | 11         | 2          | 3      | 12     |
| Sr     | 36     | 115    | 56     | 126    | 65     | 125    | 237        | 48         | 81         | 27         | 92     | 511    |
| Zr     | 18     | 28     | 19     | 21     | 21     | 36     | 28         | 24         | 17         | 11         | 46     |        |
| Y      | 5      | 26     | 44     | 8      | 40     | 16     | 28         | 51         | 49         | 76         | 7      | 15     |
| Nb     | 1.8    | 1.7    | 1.8    | 1.7    | 1.1    | 3.3    | 4.6        | 1.5        | 0.6        | 2.6        | 1.1    | 17.7   |
| Ga     | 11     | 11     | 9      | 14     | 12     | 13     | 14         | 11         | 10         | 7          | 9      | 22     |
| Cu     | 66     | 168    | 30     | 352    | 20     | 10     | 425        | 10         | 127        | 41         | 13     | 29     |
| Zn     | 100    | 46     | 49     | 59     | 52     | 147    | 137        | 77         | 57         | 78         | 64     | 175    |
| Pb     | 2      | 2      | 1      | 3      | 2      | 3      | 2          | 2          | 1          | 1          | 0      | 1      |
| La     | 0      | 2      | 3      | 0      | 2      | 11     | 4          | 0          | 1          | 0          | 0      | 14     |
| Ce     | 6      | 14     | 0      | 15     | 7      | 21     | 5          | 4          | 0          | 4          | 9      | 37     |
| Th     | 0      | 0      | 0      | 0      | 0      | 0      | 0          | 0          | 0          | 0          | 0      | 0      |
| Nd     | 3      | 8      | 0      | 8      | 7      | 13     | 4          | 4          | 3          | 1          | 8      | 21     |
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Table A-23. Whole rock geochemistry of Sierran granulitic xenoliths

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