INTRODUCTION

Given the critical role that plutons play in the growth and development of continental crust, the assembly processes of a pluton are significantly under-constrained (Glazner et al., 2004). This study focuses on the construction of the Palms pluton, which is composed of Cretaceous peraluminous granite formed at an oceanic-continental convergent margin in the North American Cordillera approximately 81-76 Ma. The construction of this pluton is analyzed via whole rock and zircon geochemistry and geochronology. This survey is distinct from previous plutonic assembly studies (Coleman et al., 2004; Glazner et al., 2004) in that many zircons preserve information about the source of the melt and that the pluton's major oxide concentrations are relatively homogeneous throughout.

SIGNIFICANCE: CENTRAL ANDES AND THE HIMALAYAS: AN INCOMPLETE COMPARISON

Understanding petrologic and tectonic processes of the past can be very useful to comprehending processes in modern settings, which may be understood from indirect observations. Comparing different tectonic settings can also be useful. There are several structural similarities that could possibly indicate a likeness in processes active in mid-crustal parts of the Andean arc and the Himalayan convergent zone. When examining a crustal cross-section of an older oceanic-continental volcanic arc analogous to the Central Andes, there are possible key differences that may identify different structures in the middle crust.

The significance of this comparison is that it may aid in the interpretation of 'bright zones' found in the Central Andes, which are identified with geophysical anomalies but cannot be observed directly. Unlike the Himalayas, the Central Andes are lacking midcrustal rocks exposed at the surface, which makes the comparison between the two tectonic settings incomplete. If analogous rocks to the Central Andes are exposed at the surface in the North American Cordilleran, however, direct observations may be made relative to the origin and evolution of mid-crustal rocks in the modern Andes and the Himalayas.

Central Andes

The Central Andes is a modern magmatic arc where the oceanic Nazca plate is currently subducting under the continental South American plate (fig 1). At 800 km wide and elevations that exceed 6000 meters, the Central Andes are the largest active mountain belt formed from oceanic and continental plate collision (Isacks, 1988). The high Western Cordillera is an active volcanic area called the Central Volcanic Zone, where andesitic stratovolcanoes were constructed. To the east, the Altiplano-Puna is a high flat plateau that was tectonically shortened/thickened beginning about 12-10 Ma. The shortening of this crust caused it to be exceptionally thick, locally exceeding 70 km (James, 1971). The Altiplano-Puna volcanic complex (APVC), which is one of the most extensive young ignimbrite provinces in the world, covers this plateau with crystal-rich, dacitic Tertiary ignimbrites (De Silva, 1989).

The crust in this region is not only extremely thick, it is also characterized by prominent geophysical anomalies, including a high conductivity zone (HCZ) revealed by magnetotelluric and geomagnetic studies in northern Chile (Schilling et al., 2006). The upper boundary for the bright zones, which are created from these geophysical anomalies, is about 20 km from the surface while the lower boundary is not well defined and could reach the mantle (Schilling et al., 2006). Other geophysical anomalies include seismic discontinuities and broad, midcrustal low-velocity zones between 20^{o} and 26^{o} south. Anomalously high heat-flow values have also been found in the Western Cordillera and Altiplano Plateau. These geophysical anomalies have been explained by a zone of partial melting in the crust (Schilling et al., 2006), which is supported by the eruption of young ignimbrites found in the Altiplano-Puna Volcanic Complex. This bright zone has been likened to the migmatitic channel flow zone located in the Himalayas (Schilling et al., 2006).

Himalayas and the Tibetan Plateau

The Himalayas are a mountain belt that stretches 2,900 km along the zone of convergence of the Indian and Eurasian plates, with peak elevations that exceed 7000 meters, including many of the highest peaks of the world (fig 2). The Himalayas are geomorphically similar to the Andes, in that the Himalayas also have a high plateau behind the main plate collision zone known as the Tibetan plateau.

Bright zones that are currently observed under the Tibetan plateau are used as evidence for the interpretation of partially molten mid-crustal zones. The bright zones are located about 15 to 20 km beneath the surface and are identified by geophysical anomalies, such as high electrical conductivity and seismic wave attenuation (Gaillard et al., 2004). The Tibetan Plateau is about twice as thick as average crust; thickened crust and high concentration of radioactive elements may have been the cause of the initial formation of a mid-crustal partially melted zone (Nelson et al., 1996). Continued shortening and southward extrusion of this weak mid-crustal layer may be expressed in the exposure of migmatitic rocks and associated granites of the High Himalayan Crystalline Series; bright zones found underneath the Tibetan Plateau suggest that similar processes that formed the High Himalayan leucogranites in the Miocene are presently operating beneath Tibet (Gaillard et al., 2004). Leucogranites formed above the High

Himalayan migmatites share similar properties, such as comparable depths and electrical conductance, with rocks thought to be present in the bright zones imaged beneath the Tibetan Plateau (Gaillard et al., 2004). Studies have also shown that the metapelites from the High Himalayan Crystalline Series could be the source rock for Himalayan leucogranites (Patiño-Douce and Harris, 1998). Therefore, workers have interpreted the High Himalayan Crystalline Series to be an extrusive migmatitic channel flow and the leucogranites to be a collection of its partial melt (Gaillard et al., 2004).

Comparison

There are several similarities between the two different tectonic settings that suggest the bright zones may have formed by similar processes. For example, a broad orogenic plateau is located inboard of the main plate collision zone. This plateau is relatively high and is underlain by crust that is much thicker than the global average. Bright zones have been imaged at similar depths (~15-20 km) beneath these plateaus.

The comparison is incomplete, however, because the hypothesized formation of the Himalayan bright zones as a migmatitic zone is more thoroughly understood because of the exposure of the High Himalayan Crystalline Series and associated leucogranites at the surface. These exposures offer insight into the processes that may be taking place beneath the Tibetan Plateau today. The Andes, on the other hand, offer nothing analogous to the exposure of High Himalayan migmatites and leucogranites. The inferred structure of the Altiplano-Puna Plateau is therefore built only on geophysical data and petrologic inferences from volcanic rocks of the APVC. Exploring an older, non-active continental-oceanic convergent margin where the interior is presently exposed at the surface may offer insight into the processes that are responsible for the evolution of the mid-crust beneath the Altiplano-Puna Plateau and the generation of APVC ignimbrites.

North American Cordillera

The Andean Cordillera is often considered a modern analog to the older western North America Cordillera (Jordan and Allmendinger, 1986). Most workers agree that a large part of the crustal thickening in the back arc of the Central Andes is due to tectonic shortening; similar shortening processes were active in the North American Cordillera during Jurassic and Cretaceous time (Beck and Zandt, 2002). Jurassic and Cretaceous plutons nearer to the continental margin are the markers of magmatism intimately associated with this Andean-type tectonism (Ernst, 1981).

The area of the study site (Joshua Tree National Park- JTNP) includes these markers of Andean-type tectonism (fig 3). This area is particularly useful to understanding the interior of an Andean-style arc and backarc because it is part of a tilted crustal section. Because of this tilting, traveling from east to west is like descending structurally deeper from the back arc and into the arc of the Mesozoic Cordillera (Barth et al., 2008a). Therefore, the interior of the Andean-style arc and backarc are exposed at the surface. This area has the potential to provide information about the mid-crust of the Andes that is not available at the presently active subduction zone and may be analogous to the Himalayan Crystalline Series and associated leucogranites that provided insight about processes currently taking place underneath the Tibetan plateau. Consequently, understanding the construction and structure of this exposed North American Cordilleran midcrustal section may further the comparison of these two different convergent settings and offer a better understanding about the processes occurring in the interior of the Central Andes.

In order to confirm the geochemical similarities between the Central Andes magmatic arc and North American Cordilleran Cretaceous arc, Andean Central Volcanic Zone rhyolites were compared to Cretaceous granites from the North American Cordillera (figs 4 and 5). Both areas show similar, although not identical, trends for major oxides. This suggests that similar processes were taking place to form these melts. Even though these rock suites may have formed from similar processes, it is not expected that their chemistry would be identical due to many factors, including different source rocks and potential geochemical differences between plutonic and volcanic systems.

OBJECTIVES

The objectives of this study are to use whole rock and zircon geochemistry and geochronology to understand the construction of just a small portion of the middle crust exposed in JTNP in fine detail. The Palms pluton was chosen because it formed in the middle crust and because of its proximity to the Mesozoic sheeted complex. Therefore, with a detailed geochemical and geochronological understanding of this pluton, inferences can be made about processes, such as the development and transport of magmas, taking place both structurally above and below this magmatic body.

STUDY SITE

Joshua Tree National Park

JTNP is in the eastern Transverse Ranges of southern California. The oldest rock type in JTNP is the Pinto Gneiss, which thus constitutes the framework for igneous rock emplacement. This consists of Proterozoic ortho- and paragneisses that are fragments of a more widespread metamorphic complex along the Pacific margin of North America (Trent, 2004). At least five different major plutons, ranging in age from middle Proterozoic to Cretaceous, have intruded the framework rocks (Trent, 2004). In addition to plutons and framework rocks, a structurally distinct Mesozoic sheeted complex is exposed along the western margin of JTNP.

The Mesozoic sheeted complex is a body of moderately east-dipping layered rock that has been described as both migmatitic and igneous (Matti et al., 1994; Barth et al., 2008b). Paleodepths of the sheeted complex are roughly 18-20 km (Palmer et al., 2006; Needy et al., 2009). Because this sheeted complex contains abundant shallow dipping bodies of intrusive rocks lying structurally beneath upper crustal plutons and formed at similar paleodepths as the bright zones currently found under the Altiplano-Puna plateau, it is possible that the sheeted complex is a paleo 'bright spot' that would have been present during active arc magmatism in Jurassic and Cretaceous time. The Palms pluton was emplaced structurally along the top of the sheeted complex. Understanding the evolution and emplacement of this magma body may allow us to relate assembly of an upper crustal pluton to the development of the sheeted complex and therefore be a key to understanding how structures that are potentially like the Mesozoic sheeted complex form.

Palms Pluton

The western part of JTNP was originally mapped by Rogers (1954). According to his map, the pluton analyzed for this research was just a portion of Palms quartz monzonite (unit C), which he extended much farther east in the Queen Mountain and Queen Valley areas (units A, B, and C). Rogers also described this pluton to be homogeneous and mostly massive.

Dibblee (1967) remapped this area and regrouped the pluton to be part of the White Tank monzogranite. Brand (1985) also describes this pluton as being a part of the White Tank monzogranite. These workers describe the pluton as a homogeneous, medium-grained equigranular phase of the White Tank monzogranite.

In this paper and other recent papers (Palmer, 2006; Barth et al., 2008a) this pluton has been renamed to the Palms pluton, because it is distinct in space, age, and chemistry from the White Tank pluton and because this is its original name from Rogers (1954).

HYPOTHESIS

This study proposes that the Palms pluton is comprised of several discrete intrusions rather than a single homogenous batch of melt, in contrast to Dibblee (1967) and Brand (1985) who describe the pluton as a homogenous magmatic intrusion. This hypothesis will be tested by constructing a thermal evolutionary model using the HEAT program of Wohletz (2007), which was used to model the crystallization time to cool one body of melt the size of the pluton below its solidus (fig 6), assuming a starting temperature of 1000°C in a magmatic body 10 km across emplaced 18-21 km beneath the surface. Other model assumptions included a surface temperature of 25°C, and a 20°C per km thermal gradient. Heat sources included in the model are magma convection, latent heat, and heat from radioactive decay. Maximum values for assumptions were chosen intentionally in order to determine a maximum cooling time. According to this model, it would take approximately 450 ky for a magma body of this size to cool from 1000°C and solidify. If the age range detected from U/Pb ratios in the magmatic zircon is greater than 450 ky, then the hypothesis that the Palms pluton formed from a series of discrete intrusions can not be rejected. If the age differences in magmatic zircons is less than 450 ky and a mixing/fractionation model can be applied to explain geochemical variation in this intrusion, then the hypothesis can be rejected.

METHODS

Collection of Samples

Samples of the Palms pluton had been collected before this study (Brand, 1985; Palmer, 2005; Wooden and Barth, unpublished data) and additional samples were collected to better understand tentative geochemical zones mapped on satellite images. The satellite images were processed using four band ratios emphasizing ferric iron, ferrous iron, carbonate and clay concentrations, and an average reflectance (Jarvis et al., 2001). This technique allowed identification of more subtle chemical variations in rock than by color alone (Jarvis et al., 2001). Locations of previously collected samples were overlain on this map (fig 7). Additional samples were collected to characterize zones that had few or no samples from previous studies.

Whole Rock Analysis

Whole rock analysis was done in order to determine if there was a spatial pattern of geochemical variation in the Palms pluton. Samples collected from the field were crushed and powdered into a fine-grained rock flour using a ceramic ball mill. Samples were fused into glass disks using high dilution fusion techniques similar to methods developed by Nakada (1985) and Mashima (2002). Nine grams of Li tetraborate were mixed with one gram of rock powder ($\pm$ 0.0005 g). The sample-flux mixture was heated in platinum alloy crucibles at high temperatures (800 to 1200^{o}C) and the molten mixture poured into a platinum mold. The platinum mold was then placed on a hot plate heated to ~500^{o}F to allow the disc to cool slowly and prevent cracking. The overall composition and cooling conditions were such that the end product after cooling was a one-phase glass.

Chemical analysis was performed using a Bruker S4 PIONEER4 kW wavelength dispersive X-ray fluorescence spectrometer (XRF), housed at the Michigan State University Department of Geological Sciences. Data reduction was performed with Bruker's SPECTRAplus software using fundamental parameters.

Zircon Analysis

Zircons were extracted from five granite samples from the Palms pluton (fig 8), which represent three of the zones mapped using the satellite image. Analyzing zircons from various parts of the pluton was essential to testing the hypothesis of multi-stage development of this midcrustal pluton.

Zircons were handpicked from crushed samples and were embedded in 25.4 mm epoxy rounds. A 25.4 mm cylindrical Teflon mold was positioned so that it surrounded the grains, and a thoroughly blended mixture (25:3 by weight) of Struers EPOES Resin and EPOAR Hardener was poured over them to a depth of 1-1.5 cm. The epoxy was left to cure for 12-24 hours. The sample mount was put in an oven at 60^{o}C for about 2 hours and then cooled. The mold was removed. The epoxy plug was trimmed on a lathe to form a disc about 6 mm thick, and polished to expose the individual grains first with

1500 grit wet/dry sandpaper, followed by 6 μm, and finally a 1 μm diamond powder slurry on a Struers LabPol5 rotary polisher.

Prior to placing the mount in the ion microprobe, the sample was cleaned in an EDTA solution, thoroughly rinsed in de-ionized water and baked in a vacuum oven for approximately one-half hour. The sample was then covered with roughly 10 nm of gold in a Denton sputter coater in order to increase surface conductivity.

Cathodoluminescence (CL) images were prepared using a JEOL 5600LV scanning electron microscope (SEM) located at Stanford University with custom Photomultiplier tube assembly prior to chemical analysis. These images were used to reveal internal zoning related to chemical composition and also to identify inclusions.

Zircon analyses were made using the SHRIMP-RG (Sensitive High Resolution Ion Microprobe - Reverse Geometry) housed in Green Hall at Stanford University, and co-owned by the U.S. Geological Survey. For Uranium and Lead isotopic analyses, the primary oxygen beam operated at ~4-6 nA and excavated an area ~25-35 μm in diameter to a depth of ~ 1μm. For trace element analysis, the primary oxygen beam operated at ~1.5 nA with a 15-20 μm spot size. The reduced data was normalized to isotopic ratios in zircon standard R33 (Black et al., 2004); and concentrations in CZ3 (Pidgeon et al., 1994) and MAD (Mazdab and Wooden, 2006).

Zircon Geochronology

Zircon ($ZrSiO_4$) is commonly used to for isotopic dating of the formation age of igneous rocks because it is chemically inert, can survive both weathering and transport processes, and resists recrystallization during high temperature metamorphism (Hinton and Upton, 1991). The Uranium/Lead system is used for dating because the zircon structure can take up U^{+4} while excluding the larger, divalent Pb cation. We are determining ages based on a ratio between the parent and the daughter isotopes. Anything besides radioactive decay that changes this ratio will therefore pose a problem for calculating ages. One of the problems that must be addressed is the assumption of a closed system. If the zircon is heated above its closure temperature then some of the daughter lead will be lost, because it is not stably bound in the zircon structure. This problem can be addressed by looking at the $^{238}U/^{206}Pb$ and $^{235}U/^{207}Pb$ systems simultaneously. These systems have different decay constants, therefore two sets of equations can be used to solve for age (t). These solutions form a Concordia curve on a graph of $^{235}U/^{207}Pb$ vs. $^{238}U/^{206}Pb$. Data points can be plotted on the same graph and compared to the Concordia curve. Because the decay constants for the two systems are not the same but the Pb loss rate for both daughter isotopes is the same, it is possible to determine which samples have lost lead. If loss of Pb occurred at a single time, a linear trend that can be traced back to its intercept with Concordia to calculate the age prior to the loss of Pb (i.e., Pb-Pb age of discordant zircon).

There are three possible U-Pb concordia plots, Conventional (Wetherill, 1956), Tera-Wasserburg (TW; Tera and Wasserburg, 1972), and its variant (Tatsumoto et al., 1972). Zircon $^{235}U/^{207}Pb$ and $^{238}U/^{206}Pb$ data was plotted on a TW Concordia diagram ($x = {}^{238}U/^{206}Pb$ $y = {}^{207}Pb/^{206}Pb$ $z = {}^{204}Pb/^{206}Pb$). Tera-Wasserburg plots have advantages over other Concordia diagrams in visualization, because the errors in their X- and Y-values are generally much less correlated than other Concordia diagrams, so that the relative scatter of their error ellipses is more apparent (Ludwig, 2003). Furthermore, as long as the magnitude of the common-Pb correction is small, and the error in the $^{206}Pb/^{238}U$ ratio is significantly greater than for the $^{207}Pb/^{206}Pb$ ratio, error correlations for TW Concordia data can be satisfactorily approximated by zero, whereas a precise and accurate value for error correlations is always essential for the Conventional Concordia plot (Ludwig, 2003).

A TW Concordia plot was used to identify discordant zircon analyses. These analyses were excluded when calculating the ages of a sample using an error weighted average of the $^{206}Pb/^{238}U$ ages.

Magmatic vs. Premagmatic Zircons

Because zircon is such a refractory mineral, we commonly observed that zircons survived the magmatic episode(s) that formed the Palms pluton and were incorporated into the melt as solid crystals preserving their older history of initial crystallization. Therefore, some zircons in the collected samples are much older and are not related to the crystallization of the Palms pluton melt. These are described as premagmatic zircons, because they existed before this magmatic event. Zircons that were crystallized directly from the melt are named magmatic zircons. Detailed analysis of both zircon populations can provide information regarding the composition and evolution of melt, as well as the source of the melt.

In order to distinguish between premagmatic and magmatic zircons, the internal morphology of the grains was examined via CL images. Premagmatic zircons are commonly observed as a rounded core and magmatic zircons will commonly have a euhedral outer rim grown around the premagmatic core. In addition, histograms and cumulative probability plots were used to show the cumulative probability distribution of multiple analyses of a zircon age population, obtained by summing the probability distributions of a suite of data with normally-distributed errors. The number of peaks represents the number of statistically distinct ages that exist in the zircon population, and the height of the peak represents the relative likelihood that additional analyses will fall in that range.

Magmatic Ages

After the distinction between magmatic and premagmatic ages was established, the magmatic age of the sample was determined using a suite of SHRIMP-RG spot analyses from each sample. The magmatic age of each sample was calculated using an error weighted average of the spot analyses.

Zircon Geochemistry

The chemistry of a magma system can change over time. Although this can be apparent in many chemical components, in analysis of zircons it is the easiest to identify in rare earth elements (REE). This is because REE and other trace elements are far more sensitive to igneous fractionation processes than major elements, and middle and heavy REE are enriched in zircon relative to silicate melts by factors of 3 to 1000 (Sano et al., 2002). REE patterns were normalized to chondrite REE values (Anders and Grevesse, 1989) to identify how this magma compares to, and evolved from a primitive mantle source.

Major elements in potential included silicate and oxide mineral phases were analyzed to identify inclusions. Some analyses were excluded based on these major element data.

Zircon Temperatures

Ferry and Watson (2007) established the following relationship among temperature, melt composition, and the solubility of Ti in zircon:

$$\log(\text{ppm Ti-in-zircon}) + \log\alpha_{SiO_2} - \log\alpha_{TiO_2} = A_2 + B_2/T$$

where A and B are constants, α values are the activity coefficients of SiO_2 and TiO_2, and T is the crystallization temperature (^{o}K) of zircon. This thermometer is particularly useful because it can be applied to individual grains or compositional zones within grains and it provides estimates of crystallization T for both magmatic zircons and premagmatic zircons which have been removed from their original petrologic context. The activities of SiO_2 and TiO_2 were estimated using both the calculated zircon saturation temperature (Watson and Harrison, 1983) and the estimated solidus temperature (Piwinskii, 1968).

RESULTS

Whole Rock Geochemistry

Whole rock geochemistry displays consistent systematic variations between previously collected data and data collected for this study. Most major elements in the Palms pluton decrease systematically with increasing SiO_2 (fig 9). However, this chemical variation can also be observed spatially. Figure 10 reveals trends of change in both trace and major element concentrations from east to west of the pluton. These changes do not follow the previously observed zones that were drawn from satellite imagery (fig 7).

Zircon Ages

Some age data for sample JW341 was previously reported by Barth et al. (2004) and other age data for samples in western Joshua Tree National Park, including the Palms pluton, are reported in Needy et al. (2009). However, in this study these data are reinterpreted as a group of related rocks, with inclusion of additional age data as well as additional trace element compositional data.

Even though this pluton formed in Cretaceous time, older zircons were preserved in the samples. Age distribution is dominated by a peak at 77 Ma but also contain significant populations at 81 Ma, ~100 Ma, ~ 250 Ma, and ~1700 Ma (fig 12). The premagmatic zircons indicate a hybridized source from the Mesozoic and Proterozoic times (fig 11). The Mesozoic premagmatic population is much closer in age to the magmatic zircons, but is still older than the magmatic age. The magmatic zircons are the most prevalent and are from Cretaceous time.

By examining zircon populations in more detail, premagmatic zircons can be broken down into three different sources (fig 12). There is one Proterozoic population that peaks at roughly 1650 Ma, and two Mesozoic populations that are from the early Cretaceous and the early Triassic periods.

Magmatic zircons represent the youngest coherent age group in each sample, and show two statistically significant peaks at approximately 77 and 81 Ma. Samples JW221 and JW222, however, contain abundant magmatic zircons in the younger of these two age groups, there are also a few zircons with ~81 Ma ages. These grains are considered to be premagmatic for these samples because the probability that all of these grains are from one population is 0.000 (fig 13) and their geochemistry is distinct from the younger zircon crystals, which is discussed further in the next section.

Magmatic Zircon Ages

Two calculated ages are presented for samples JW221 and JW222 (figs 13 and 14). One includes older grains while the other excludes these grains. We excluded the very old grains (81 Ma) because, as mentioned previously, there is a very low probability that these zircon crystals are from one population and they are therefore likely to be premagmatic for JW221. Geochemistry is also distinct. These older grains from sample JW221 share (fig 15) a similar Eu/Eu* value as the older JW224 sample but are distinct from the younger grains in JW221. It is hypothesized that single the older grain in JW222 is from a different population, but cannot be excluded based on age, and zircon geochemistry is not available for comparison for these spot analyses. Therefore, the older grain is included for calculating the age of JW222.

The calculated ages of all the Palms pluton samples (figs 16-19) indicate a difference in age between samples of at least 3 million years, considering the uncertainty limits of the oldest (JW224) and youngest (JW341) samples. There is some doubt, however, that JW222 is part of the Palms pluton. JW222 was collected from a stream boulder that does not correspond in composition to any nearby known outcrops.

Zircon Geochemistry

Inclusions in the zircon grains were identified by comparing major oxides (fig 20). The outliers are interpreted as inclusions and are excluded in geochemical analyses but their values are displayed in Appendix A.

Chemical distinctions between magmatic and premagmatic zircons allow us to identify four distinct groups. Grains that were identified as magmatic have a relatively small Eu anomaly and low Yb/Gd. There is a general pattern of deeper europium anomalies with increasing Yb/Gd values for magmatic grains (fig 21).

The grains that were identified by age as premagmatic generally have a deeper europium anomaly and greater variability in Yb/Gd than magmatic grains (fig 21). Based on age and the variations in REE abundances, premagmatic grains have three (or possibly four) distinct populations. One population, which is observed only in JW221, follows a trend that is similar to the magmatic zircon trend. These premagmatic zircons formed in Triassic and Cretaceous time.

The other premagmatic populations are from Proterozoic time, with two younger Mesozoic grains. One population (I) has a high Yb/Gd value and a low Eu/Eu* value. A second (II) has a low Eu/Eu* value and a low Yb/Gd value. This population is comprised almost entirely of Proterozoic grains. One zircon crystal from JW224, however, is Mesozoic. Populations I and II are present in all samples. An additional population is observed in JW341 (fig 22), which also falls into the magmatic trend. While one grain is Mesozoic, the other is Proterozoic. The distinction between magmatic and Mesozoic premagmatic is somewhat arbitrary in this section. While there is a clear distinction in the detailed SHRIMP-RG isotopic U/Pb analysis, trace element geochemical analyses run only a brief scan on U/Pb ages. Therefore, U/Pb ages are imprecise and there may be some overlap between magmatic grains and Mesozoic premagmatic grains. This issue is further addressed in the discussion section.

Barth et al. (2009) analyzed Proterozoic zircons from basement rocks in the San Bernardino Mountains and the eastern Transverse Ranges, and observed that metamorphic zircons were distinguishable from other populations due to their high U and relatively low Th concentrations, yielding a Th/U ratio of approximately 0.01-0.1. The samples from population I (fig 21) yield a similar Th/U ratio. The oldest sample (JW224), however, is an exception to this and has more variability than the other samples from population I.

Zircon Temperatures

Magmatic zircons show a clear trend of decreasing calculated temperature with increasing Hf content (fig 22). This trend discontinues, however, at the Hf value of approximately 1400 ppm. The Hf concentrations of premagmatic zircons are generally shifted to the right of the magmatic values in this plot. In JW224, the Mesozoic premagmatic zircon crystal follows the magmatic zircon temperature trend. In JW341 and in JW221 there is no clear distinction between Proterozoic premagmatic and Mesozoic premagmatic grains.

DISCUSSION

Magmatic zircon ages support the hypothesis that the Palms pluton was formed from multiple discrete intrusions rather than a single homogenous batch of melt according to the thermal evolutionary model developed from the HEAT program of Wohletz (2007). Based on this model, it would take approximately 450 ky for one body of melt the size of the Palms pluton to crystallize (fig 6). However, U/Pb ages of magmatic zircons indicate at least a 3 my span of crystallization, regardless of pooling data from all samples (fig 12) or observing individual sample ages (fig 19).

In addition to multiple magmatic intrusions, the Palms pluton was formed from melting multiple sources. One method of identifying sources is by observing the premagmatic zircon ages. The dominant component of the melt source is Proterozoic but there is also a younger Mesozoic component. Barth and Wooden (2006) identified the melt sources for Triassic arc plutons in the same region as the Palms pluton. These Triassic plutons have at least two components that fall into Proterozoic time, peaking at roughly 1400 Ma and 1700 Ma. This 1700 Ma peak is present in the Palms pluton but the 1400 Ma peak is absent. An interesting question that could be pursued is what caused the change in magma source between Triassic and Cretaceous time? It could be due to tectonic shuffling, a depletion of the once fertile 1400 Ma source rocks, or a significant change in the geothermal gradient.

A second, and more detailed, method for distinguishing sources is by examining the trace element chemistry of premagmatic zircons. This can also be done in order to determine if there were unique magma sources for individual intrusions. By using geochemistry, two distinct populations of premagmatic zircon crystals were identified in all samples, indicating a genetic relationship of source rocks between the distinct pulses of magma. As indicated by the distinct ages and chemistry of other premagmatic zircons, there were also sources that are possibly unique to individual samples, such as the older Mesozoic source component in JW221 and the higher Th/U zircons in JW224. These different sources found in few samples also support the conclusion that the Palms pluton is a result of multiple distinct intrusions.

A potential problem for many of these trace element analyses, however, is that many of the zircon crystals lacked both a detailed trace element and detailed age analysis. Each SHRIMP-RG trace element analysis had a brief, but imprecise, U/Pb isotope analysis to calculate a rough age with up to 25% error. Although these techniques should cause no problems in distinguishing between Proterozoic premagmatic grains and magmatic grains because their difference in age is two orders of magnitude, it may have caused some overlap between premagmatic Mesozoic grains and magmatic grains because they are much closer in age. The brief U/Pb age analysis may not be precise enough to distinguish between these two groups. This problem could be addressed by having both SHRIMP-RG trace element and isotopic scans on the same zone of the same zircon crystal. This data, however, should still be used with caution because each analysis will use unique material. It is therefore possible that analyses could unintentionally sputter material from a different zircon domain.

The distinctions of premagmatic and magmatic zircon are observed from not only age and geochemistry, but also in calculated temperatures. Ferry and Watson (2007) established that there is a relationship between the temperature of zircon crystallization and the composition and solubility of Ti in zircon. Other factors that are necessary to consider in this calculation are the activity coefficients of TiO_2 and SiO_2. In addition to this calculation, zircon saturation temperatures were calculated using the relationship between zircon solubility, temperature, and major element composition of the melt established by Watson and Harrison (1983).

The zircon temperatures shown in figure 23 may be unrealistic because many zircon grains yield temperatures above the zircon saturation temperature. One important component of this calculation is the concentration of Zr in the melt. Whole rock Zr concentrations, however, are unrealistic estimates of melt Zr because many zircon grains are premagmatic, which means that they were never a component of the melt. This means that whole rock Zr concentration is higher than melt Zr concentration. A rough calculation was used to determine the relative portions of premagmatic and magmatic zircon in order to address this problem. This was done by treating each zircon crystal as a sphere and calculating the volume of a premagmatic zircon core and the volume of a magmatic zircon rim. Comparing these two volumes, premagmatic zircons account for approximately 20-25% of all zircons. However, correcting for this estimate only exaggerates the problem. For example, lowering the Zr concentration in the melt by 20% would lower the saturation temperature by approximately 25^{o}C. In addition, it is likely that zircon crystals are not the only premagmatic minerals. If there are other premagmatic minerals, then this would concentrate Zr in the melt and therefore raise zircon saturation temperature, unless this effect greatly changes alkalinity of the melt. The net effect of these considerations therefore appears to be minimal on changing the calculated saturation temperature in these samples.

An unrealistic zircon saturation temperature is not the only possible explanation for the temperature discrepancies. The activity coefficients used in the Ti in zircon thermometer (Ferry and Watson, 2007) may be incorrect for at least some of the zircon grains. Although it is likely that the silica activity is near 1 for a silica-rich melt, if some premagmatic zircons were from a more mafic source, the SiO_2 activity would be too high for these grains. Lowering the SiO_2 activity coefficient would lower the calculated temperature.

Another possibility is that all of the calculations are correct, but the Palms pluton was frozen in a state of disequilibrium. In other words, the intrusive melt was hot enough to begin dissolving premagmatic zircons, but cooled too quickly to completely dissolve the premagmatic grains. Watson (1996) addressed this issue by calculating dissolution rates of zircons in melts of granitic composition. Watson considered zircon behavior over three different hypothetical thermal histories for crustal magmatic events, some peaking at 900^{o}C. In all three models for the time-temperature paths, zircon dissolution was slow enough that larger zircon grains survived without complete dissolution. Based on Watson's models (1996), relatively rapid melting and magma ascent may preserve the cores of premagmatic zircon crystals even if melt temperatures were above zircon saturation temperatures.

Although observations support the hypothesis that this pluton did not form from one batch of melt, it is not clear how many batches were involved in pluton assembly. According to magmatic zircon ages, there were at least two batches of melt. These batches do not follow the zonal patterns drawn from the satellite images (fig 7). Zircon age and trace element composition indicate a more complicated zonal pattern than the zones previously drawn. This may be due to remote sensing imaging recording weathering patterns of the rock rather than the internal chemistry of the granite. Pulses of magma may also occur more frequently than can be discerned with the precision of existing dating methods. If there are enough magmatic zircons lacking a complicated core, Mattinson's (2005) chemical abrasion-thermal ionization mass spectrometry (CA-TIMS) technique would provide more precise ages. Combining this with Ti-in-zircon thermometer (Ferry and Watson, 2007), we might be able to 'see' pulses of magma crystallize and also detect generations of new pulses.

An alternative to the conclusion that the Palms pluton formed from at least two melt batches would be that JW224, which is the only sample that is significantly older, is not actually a part of the pluton. It may be a 'pocket' of preserved granite from a previous intrusion, such as Squaw Tank. Although Squaw Tank is shallower (fig 24) than the Palms pluton, it could be a relic from the plumbing system. This alternative conclusion could be tested by a detailed comparison of Squaw Tank and JW224 magmatic zircons.

The Palms pluton is composed of multiple melt batches, and therefore it may be considered an extension of the Mesozoic sheeted complex. Even though the sheets are discrete, this extension is not initially apparent because the contrast between the compositionally similar sheets in the Palms pluton is not as striking as the alternating felsic and mafic sheets in the sheeted complex. A working model of the formation of the middle crust was developed (figs 25A and 25B). Future work to test this idea could include a detailed study of the Mesozoic sheeted complex to obtain a more complete understanding of the relationship between its component and Palms granites, as well as other surrounding plutons.

If these bodies of rocks are related, then this provides insight into the behavior of middle crust in magmatic arcs. A coalescence of alternating mafic and felsic sheets that progresses to exclusively felsic sheets at shallower depths would be distinct from the structure resulting from partial melting and granitic plutonism in the Himalayan middle crust. Additional research is also necessary to more thoroughly understand the relationship between the Mesozoic sheeted complex and the Palms pluton. Understanding any connections between this complex and other surrounding plutons would be important as well. A comparison of the geochemistry of major oxides to other nearby Cretaceous granites is shown in figure 26. These other two plutons (Squaw Tank and Smoke Tree) are structurally shallower than the Palms pluton but formed at roughly the same time. Generally, the granites follow the same geochemical trends. Squaw Tank granites are overall less felsic with one exception and trends are unclear between SiO_2 and MnO, Na_2O, and K_2O. Future work could entail a more complete comparison of these plutons and, additionally, a geochemical comparison of these granites to the Mesozoic sheeted complex.

Even if the sheeted complex was once a bright zone found in the back arc region of the North American Cordillera, it remains unclear whether the sheeted complex is igneous or metamorphic. If it is metamorphic, then the bright zones formed in both convergent settings would develop similarly. In other words, they would both form from a partially molten zone of weakened crust with the collection of magma above the migmatitic zone.

If the sheeted complex is igneous, however, the development of bright zones in oceanic-continental collision zones is likely to have been different from the bright zones observed in continental-continental collision zones. This hypothesis can be addressed by comparing the geochemistry of plutons that are both structurally shallower and deeper than the sheeted complex. If these plutons are geochemically related, then it is unlikely that the sheeted complex is the source for the shallow plutons. The plutons would have a common source, and therefore the source must be deeper than the plutons below the sheeted complex. This deeper source would also be distinct from the Himalayan leucogranite source.

The Palms pluton, White Tank, and Squaw Tank are all granitic intrusions that are structurally shallower than the sheeted complex (fig 24). Whole rock geochemistry was used to compare the shallow intrusions to deeper granitic bodies located in the San Gabriel Mountains (Josephine Mountain intrusion and Waterman batholith). Comparing both major oxide (fig 26) and trace element (fig 27) concentrations, there is significant overlap. Lower crustal granitic rocks, which formed deeper than the sheeted complex, appear to be somewhat enriched in TiO_2, FeO, and Ba. Comparing additional chemical data, such as rare earth elements and isotopes would significantly help in determining the relationship between the plutons found above and below the sheeted complex. In their study, Wiegand et al. (2007) analyzed Sr and Nd isotopic compositions of these and additional plutonic rocks from lower to upper crustal intrusion depths to evaluate the evolution of those rocks. Initial Sr_i and εNd values range from 0.708-0.712 (-10 to-16) in upper crustal granites, and 0.708-0.714 (-11 to-16.5) in the sheeted plutonic complex. Lower crustal intrusive rocks are characterized by Sr_i and εNd values of 0.707-0.708 (-11 to -12) in the Waterman Mountain, and 0.709-0.710 (-10 to-11) in the Josephine Mountain intrusion. These values show a significant overlap between sheeted middle crustal intrusive rocks and upper crustal granites, indicating a genetic relationship between them.

If the sheeted complex is indeed an igneous body, then it may mark a boundary of neutral buoyancy where it capped relatively dense igneous intrusions. Thus, once the intrusions reached this boundary, they spread laterally into relatively flat sheets, forming the sheeted complex.

Although they share similar features, bright zones, and therefore, magmatism, observed in continental-oceanic convergent settings may develop through different processes than midcrustal magmatism observed in continental-continental convergent settings. Studying the emplacement and source of a pluton has shed light on understanding the development of magmatism in an arc setting, and therefore has begun to provide more details to cultivate a more articulated comparison of these two tectonic settings.

In order to complete the comparison, however, geochemical work should focus on determining the relationships between Cretaceous intrusions to the Mesozoic sheeted complex. If the sheeted complex had zircons that were similar in age and geochemistry of the magmatic zircons of the Cretaceous intrusions, this would support the idea that the Mesozoic sheeted complex is magmatic and therefore a unique structure to the Himalayan midcrustal migmatitic channel flow.

CONCLUSIONS

Zircon age and trace element data support the hypothesis that there were at least two discrete intrusions involved in the assembly of the Palms pluton. Zones that were identified with remote sensing imagery do not match zircon chemistry, which indicates that zoning may be more complicated. Remote sensing imaging may therefore reflect weathering patterns rather than chemistry of the granite. Not only can premagmatic zircons be identified by age, but they also display distinct REE patterns from magmatic zircon patterns. Mesozoic premagmatic zircons, however, share magmatic zircon patterns. Premagmatic zircon REE patterns indicate a genetic relationship between melt batches.

Future work should include conducting both trace element and isotopic scans on the same zircon zone. Also collecting data from other Cretaceous granites in the surrounding area and the Mesozoic sheeted complex would be necessary to more fully understand magma production and evolution in the middle crust of magmatic arcs.