Multimethod detrital thermochronology of the Great Valley Group near New Idria, California

Pieter Vermeesch[†]
Department of Geological and Environmental Sciences, Stanford University, Braun Hall, room 320, 450 Serra Mall, Stanford, California 94305-2115, USA

Donald D. Miller[‡]
Aera Energy, LLC, P.O. Box 11164, Bakersfield, California 93389-1164, USA

Stephan A. Graham[§]
Department of Geological and Environmental Sciences, Stanford University, Braun Hall, room 320, 450 Serra Mall, Stanford, California 94305-2115, USA

Johan De Grave[#]
Geological Institute, University of Gent, Krijgslaan 281, B-9000 Gent, Belgium

Michael O. McWilliams[††]
Department of Geological and Environmental Sciences, Stanford University, Braun Hall, room 320, 450 Serra Mall, Stanford, California 94305-2115, USA

ABSTRACT

The simultaneous use of several thermochronological methods on replicate sedimentary rock samples can reveal their pre- and postdepositional history. Single grain U/Pb dating of zircon, zircon and apatite fission track dating, and vitrinite reflectance measurements were performed on Cretaceous through Miocene sedimentary rocks of the Great Valley Group and the Temblor Formation near Coalinga and New Idria, California. The data show that the Sierra Nevada was exhumed and cooled at ~0.5–1 km/m.y. or ~20 °C/m.y. during the Cretaceous. After deposition in the Great Valley forearc basin, Sierra Nevada erosional products were buried at great depth under low thermal gradients. At ca. 12–14 Ma, northward progression of the Mendocino triple junction triggered folding on the eastern flanks of the California Coast Ranges and rapid exhumation of the New Idria serpentinite diapir. This middle Miocene event caused the deposition of spectacular deposits of sedimentary serpentinite (Big Blue Formation). The rapid rise of the hot serpentinite body created a thermal pulse that may have provided the enigmatic heat source for oil fields in the shallow Vallecitos syncline, a few kilometers north of New Idria.

Keywords: detrital thermochronology, fission track, Great Valley Group, serpentinite, Sierra Nevada, exhumation.

INTRODUCTION

The topography and geology of central California are dominated by petrotectonic elements of the Mesozoic convergent margin: the Sierra Nevada magmatic arc, the Great Valley forearc basin, and the Franciscan accretionary prism. These three domains are genetically linked, inasmuch as they were jointly formed by Pacific plate subduction beginning in the Late Jurassic (Dickinson et al., 1996) until the transformation of the subduction zone into a transform margin during the Late Cenozoic (Atwater, 1970). The upper Mesozoic strata of the Great Valley Group (or Sequence: Bailey et al., 1964) filled the forearc basin and comprise one of the thickest sequences of Cretaceous sediments known (Ingersoll, 1982). These Great Valley Group strata crop out in a homocline along the western margin of California's Central Valley and are in fault contact with the Franciscan accretionary complex (Dickinson et al., 1996).

This paper focuses on samples of the Great Valley Group collected in and around Joaquin Ridge near Coalinga (Fig. 1). These sedimentary rocks contain geochronological and petrographic information about the evolution of their source region, the Mesozoic Sierra Nevada magmatic arc. They also contain clues about their postdepositional history and the tectonic evolution of the Diablo Range in which they are now exposed. Conventional petrographic studies of Great Valley Group rocks indicate that with time the lithic fraction of the sediments decreased, the percentage of total feldspar increased, and the composition of the feldspars became more potassic. These trends reflect increasing chemical and geomorphic maturity of the magmatic arc (Ingersoll, 1979, 1983). Paleocurrents are dominantly west directed, indicating that the source was the southern Sierra Nevada (Ingersoll, 1979). In addition to conventional petrographic studies, samples in the vicinity of Coalinga were studied by bulk-rock ε_{Nd} and ε_{Sr} analysis (Linn et al., 1991, 1992), confirming the petrographic results. Between Cenomanian and Maastrichtian time, ε_{Nd} systematically decreased from –0.7 to –5.0 (Linn et al., 1991), reflecting an eastward migration of the drainage divide, to where the Sierran magmas are more continental in composition (De Paolo, 1981). Some of the samples of Linn et al. (1991) were used by DeGraaff-Surpless et al. (2002) for SHRIMP single grain zircon U/Pb age measurements,

[†]E-mail: pvermees@pangea.stanford.edu.
[‡]E-mail: ddmiller@aeraenergy.com.
[§]E-mail: graham@pangea.stanford.edu.
[#]E-mail: johan.degrave@ugent.be.
[††]E-mail: mac@pangea.stanford.edu.

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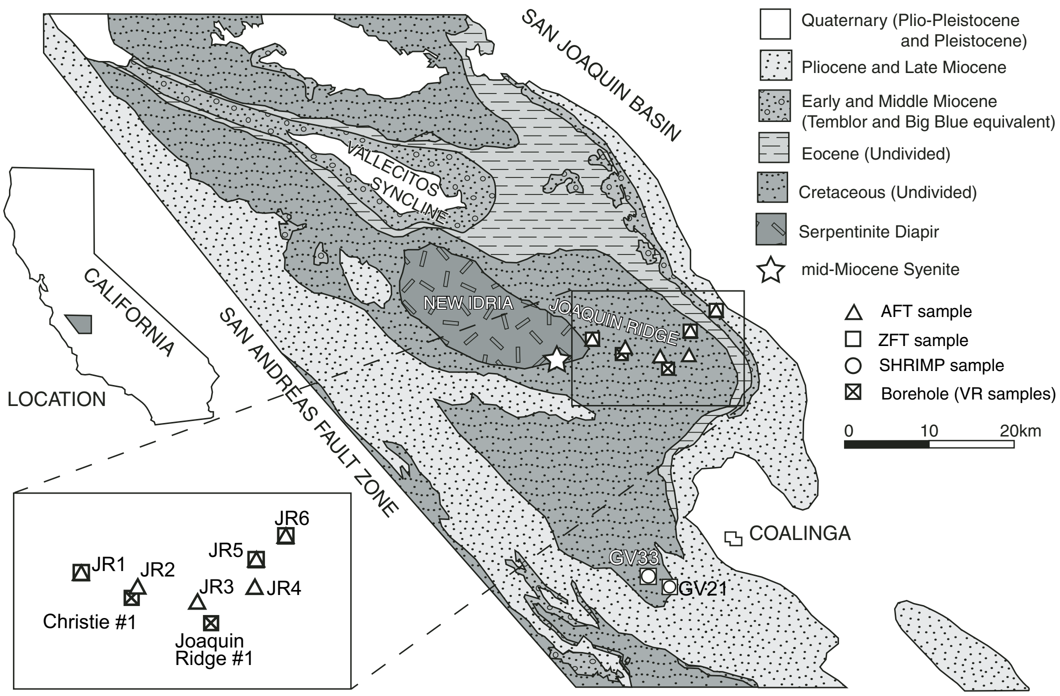

Figure 1. Simplified geologic map of the Great Valley Group near Coalinga with the sample locations. Modified from Bate (1985).

which confirmed that the sediment source for the Great Valley Group in the San Joaquin Valley is the southern Sierra Nevada, as the most frequently observed ages are 102–132 Ma. The spread of the Mesozoic ages and the number of peaks in the grain-age histograms increase with decreasing depositional age. This may reflect an expansion of fluvial drainage basins, resulting in a larger sediment source area. The minimum lag time between the age of zircon crystallization and deposition (determined by micropaleontology) is short (3–15 Ma). This constrains the Late Cretaceous exhumation rate of the Sierra Nevada (Surpless, 2001)—constraints that this paper will attempt to refine.

Nevertheless, considerable controversy exists about the exhumation history of the Sierra Nevada batholith. Some workers, based on the observation of uplifted and tilted Cenozoic strata, believe that the development of Sierran topography occurred in the last 10 m.y. (e.g., Unruh, 1991). Others claim that significant topography existed since at least the early Cenozoic based on old (U-Th)/He cooling ages of the Sierra Nevada granites and the fact that the spatial distribution of these ages preserves the signature of old topography (House et al., 1998, 2001). Both opinions are based upon data acquired from rocks of the modern Sierra Nevada. They seldom make use of the continuous record of sediments shed from the ancient mountain range since the Late Jurassic. These sediments are excellent “witnesses” of the evolution of the Sierra Nevada, for much of the batholith and its volcanic carapace have long since been removed.

Detritus from the Sierra Nevada was deposited in the Great Valley forearc basin. Above these strata was deposited a gradually shoaling-upward succession of Cenozoic deposits (Dibblee, 1971; Bartow, 1991). A total thickness of 6000–8000 m of the Upper Jurassic through Cretaceous Great Valley Group is exposed in a homocline on the eastern flanks of the Diablo Range (Dickinson, 2002). Superimposed on this homocline are folds formed as a result of right-lateral transpressional strike-slip deformation along the San Andreas fault (Miller, 1998). The Vallecitos syncline is one of these folds (Fig. 1). Cenozoic strata contain petroleum, although they are buried less than 1500 m deep in the syncline (Rentschler, 1985). The heat source for oil generation is enigmatic. Joaquin Ridge forms the anticline adjacent to the Vallecitos syncline. The New Idria serpentinite body is exposed in the core of this anticline. A few small intrusions of syenite crop out in the serpentinite body and are dated at ca. 12.8 Ma (no error stated) by $^{40}Ar/^{39}Ar$ dating on the amphibole barkevikite (Obradovich et al., 2000). Rb-Sr dating of benitoite yielded an age of ca. 12 Ma (no error stated; Obradovich et al., 2000). The Mendocino triple junction passed the latitude of New Idria at ca. 12–14 Ma (Johnson and O’Neil, 1984). The age of the spectacular deposits of the Big Blue Formation is also ca. 14 Ma, which consist almost solely of sedimentary serpentinite (Casey and Dickinson, 1976; Bate, 1985).

In the following sections, new fission track and vitrinite reflectance data are presented, followed by a discussion of implications of these data. To reconstruct the predepositional history of sediments, they must not be thermally reset. Therefore, the postdepositional history of the New Idria area will be discussed first. Among other conclusions, we demonstrate that until the rapid exhumation of the serpentinite dome, this area was characterized by low thermal gradients that prevented the buried sediments from being heated very much. The discussion of the predepositional history then follows. This paper illustrates the great power of simultaneously using multiple thermochronometers on detrital sediments. Each individual thermochronometer only tells part of the story, but the whole is greater than the sum of the parts. When all evidence is jointly considered, a self-consistent story emerges that traces the sediments from their crystallization in the Cretaceous Sierra Nevada until the final stages of their exhumation on Joaquin Ridge. This story not only has consequences for the regional geology of the Coalinga/New Idria area but also for the tectonic history of the Sierra Nevada and the petroleum geology of the Vallecitos syncline.

METHODS

Figure 1 shows a simplified geologic map of the field area with indication of the sample locations. Some of these samples were previously discussed by Linn et al. (1991, 1992)

TABLE 1. SUMMARY TABLE OF GEOCHRONOLOGICAL DATA

Sample name	Location		Depo age (Ma)	U/Pb age (Ma)	ZFT age (Ma)	AFT age (Ma)
	Latitude (°N)	Longitude (°W)				
GV33	36°6′	120°27′	89–95	131 ± 5	100 ± 6.5	
GV21	36°5′	120°25′	81–87	132 ± 4	98.7 ± 4.4	
JR1	36°20′	120°31.5′	90–100		98.1 ± 4.3	13.3 ± 0.9
JR2	36°19′	120°35′	90–93.5			71.9 ± 3.2
JR3	36°18.5′	120°26.5′	78–83.5			84.3 ± 3.3
JR4	36°20′	120°24′	71.3–75			80.9 ± 2.8
JR5	36°19′	120°24′	66–68		89.2 ± 4.5	81.5 ± 3.8
JR6	36°15.5′	120°23′	16.4–14.8		13.1 ± 1 and 84.3 ± 16 (*)	17.0 ± 3 and 49.7 ± 6 (*)

Note: U/Pb dating was performed on zircon and fission track dating on both zircon (ZFT) and apatite (AFT). All fission track ages are central ages (Galbraith and Green, 1993) except for JR6 (*), for which the two best fitting component ages are reported, calculated with the binomial peak fitting routine of Brandon (1996).

and DeGraaff-Surpless et al. (2002). They are labeled with the letters GV. An additional six samples were collected along a transect on Joaquin Ridge and are labeled with the letters JR. Table 1 summarizes the geochronological data. Apatites and zircons were separated from their host rock using the mineral separation techniques reported by DeGraaff-Surpless et al. (2002). All samples yielded abundant euhedral apatites and zircons. Both zircon and apatite fission track ages were measured with the external detector method (e.g., Dumitru, 2000). Figure 2 shows the apatite fission track data from Joaquin Ridge, arranged in order of decreasing stratigraphic age. Five of six samples on Joaquin Ridge came from the Great Valley Group, while the youngest sample (JR6) came from the middle Miocene Temblor Formation, beneath the middle Miocene Big Blue Formation. Figure 3 shows five zircon fission track samples labeled and arranged in decreasing order of stratigraphic age like the apatite fission track ages of Figure 2. The dark gray bands on the radial plots of samples GV21 and GV33 mark the range of crystallization ages measured by U/Pb SHRIMP dating (DeGraaff-Surpless et al., 2002). The light gray bands mark the depositional ages (Dibblee, 1971). Between $n = 20$ and $n = 41$ grains were dated per fission track sample. The probability p that either the oldest or the youngest population fraction of size f was missed by all n grains is given by:

$$p = 2(1-f)^n - (1-2f)^n.$$

For example, if $n = 30$ and $f = 0.12$, then $p = 5\%$. In other words, there is 5% chance that either the youngest or the oldest 12% of the detrital population was missed when dating 30 grains. In addition to the aforementioned outcrop samples, we also had access to material from two boreholes on Joaquin Ridge, the ARCO "Christie #1" and ARCO "Joaquin Ridge #1" wells. Vitrinite reflectance measurements were performed on 23 well cutting samples from "Joaquin Ridge #1" and three core samples from "Christie #1." Up to 100 reflected light points were measured on the vitrinite populations represented in each sample. The lowest reflectance values likely reflect contamination from organic matter in the drilling fluid, while some of the higher reflectance values may represent resedimented vitrinite. However, rather than arbitrarily rejecting some data, we have opted to contour and plot all the data (Fig. 4). The raw data for both the fission track and the vitrinite reflectance analysis are available in the Data Repository.[1]

DISCUSSION

Fission Track Data

Apatite fission tracks are immediately annealed at temperatures >~110 °C (e.g., Wagner and Van den Haute, 1992). At temperatures less than ~60 °C, apatite fission tracks are completely preserved. The temperature zone between ~60 °C and ~110 °C is named the apatite fission track partial annealing zone (e.g., Dumitru, 2000). In this zone, fission tracks are not immediately annealed but gradually shortened with time. The annealing temperature of zircon fission tracks is more controversial but generally considered to lie between 230 °C and 310 °C (see discussion by Tagami and Dumitru, 1996). In this paper, we will assume the more "conservative" value of ~230 °C.

First, we will discuss the apatite fission track data (Fig. 2). Of the five Great Valley Group samples, four have exclusively Cretaceous apatite fission track grain ages, indicating that these grains never reached temperatures greater than 110 °C since their deposition in the Great Valley Group. However, sample JR2 has the oldest depositional age of these four samples but the youngest fission track ages, with the latter being even slightly younger than the former. Therefore, JR2 has been partially reset and saw temperatures less than ~110 °C but well above ~60 °C. The fission tracks of sample JR2 are also significantly shorter than those of the other samples, an additional suggestion that JR2 must have been heated to well within the partial annealing zone. Sample JR1, located nearest to the New Idria serpentinite, has completely annealed apatite fission tracks and, therefore, was heated above ~110 °C. It dates the end of the heating event at ca. 14 Ma. Sample JR6 from the Miocene Temblor Formation contains two age components: one Cretaceous and one Miocene (Table 1). There also is a hint of bimodality in the fission track length distribution. A first group of relatively short (~9–13 µm) fission tracks formed prior to the mid-Miocene. These tracks preserve Sierran provenance ages but were partially annealed during the mid-Miocene thermal event. A second group of long fission tracks (~13–17 µm) formed after this thermal event and have not been annealed since then. Paleocurrent directions in the Temblor and Big Blue Formations are west-to-east, which is the opposite flow direction of the Great Valley Group (Casey and Dickinson, 1976; Bate, 1985; Bent, 1985). Therefore, the apatite grains of the Temblor Formation have been redeposited from the underlying Great Valley Group, some of which was thermally annealed during a mid-Miocene thermal event.

The zircon fission track ages for four of the five samples are older than the age of Great Valley Group deposition (Fig. 3). Sample JR1, which had completely annealed apatite fission tracks, also has unreset zircon fission track ages. Therefore, sample JR1 was heated to more than ~110 °C but less than ~230 °C after its deposition. The lag between crystallization, exhumation, and deposition times were short, which means that the source area of these sediments exhumed rapidly. The most surprising observation is that the middle Miocene sample JR6, which had a bimodal apatite fission track age distribution, also has a bimodal zircon fission

[1]GSA Data Repository item 2006017, vitrinite reflectance data, apatite fission track ages, apatite fission track lengths, zircon fission track ages, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

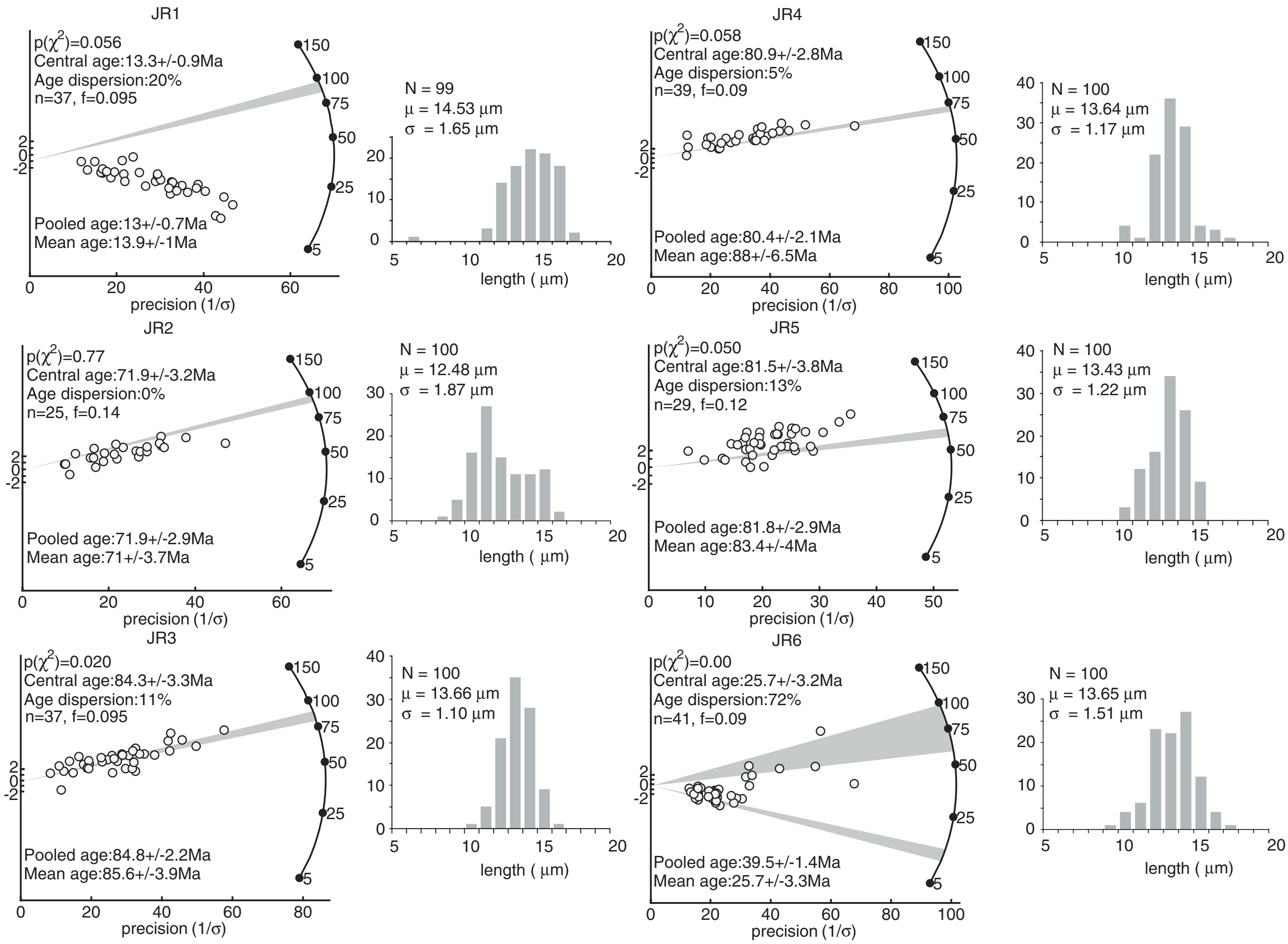

Figure 2. Apatite fission track radial plots (Galbraith, 1990) of Joaquin Ridge. The gray bands represent depositional ages. The histograms show the fission track length distributions. n—number of grains, f—largest population fraction of older/younger grains that are p = 5% likely to have been missed, N—number of confined tracks.

track distribution. The oldest mode of Mesozoic ages is compatible with the unreset fission track ages of the Joaquin Ridge samples located away from the serpentinite body (Table 1). The youngest age peak is concordant with the ca. 14 Ma apatite fission track age of sample JR1 and with the youngest mode of the apatite fission track age distribution of JR6. Because not all the apatite grains in JR6 were reset at ca. 14 Ma, we know that this sample was not heated to more than ~110 °C. In fact, there is ample evidence that the Temblor formation did not see temperatures higher than ~56 °C east of Joaquin Ridge (see below). Therefore, the ca. 14 Ma old zircons must have been annealed prior to deposition in the Temblor and Big Blue formations. Recalling the eastward paleocurrents of these deposits, this indicates that at least part of the provenance area for the Temblor Formation, which is Joaquin Ridge, reached temperatures as high as ~230 °C as recently as ca. 14 Ma.

Vitrinite Reflectance Data

Additional evidence for a heating event comes from vitrinite reflectance data from two wells on Joaquin Ridge (see Fig. 1 for the location of the wells and Fig. 4 for the vitrinite reflectance data). The "Joaquin Ridge #1" well is 4390 m deep and located in the vicinity of sample JR3. The "Christie #1" well is 1380 m deep and located next to sample JR2. Micropaleontological ages have been obtained for both wells (M.B. Lagoe, 1984, personal commun.). "Joaquin Ridge #1" sediments are of Coniacian (at 910 m) to Cenomanian age (at TD), whereas the "Christie #1" samples at TD are Cenomanian (Fig. 4). Twenty-three samples from the "Joaquin Ridge #1" well were analyzed, from depths of 300–4390 m. They show Ro values of 0.6%–1.5%, explaining why apatite fission track samples JR3, JR4, and JR5 have not been thermally reset. These samples are all located upsection from the shallowest "Joaquin Ridge #1" vitrinite reflectance samples and should, therefore, correspond to Ro values less than 0.6%, or maximum paleotemperatures less than ~85 °C (Sweeney and Burnham, 1990). The "Joaquin Ridge #1" vitrinite reflectance data imply a thermal gradient of ~14 °C /km, which was normal in the Great Valley forearc basin (Dumitru, 1988). There exists substantial evidence that the geothermal gradient in the Great Valley Group was very low during the Cretaceous and the beginning of the Tertiary. This is postulated to have been caused by the refrigerating effect of the subducting Farallon

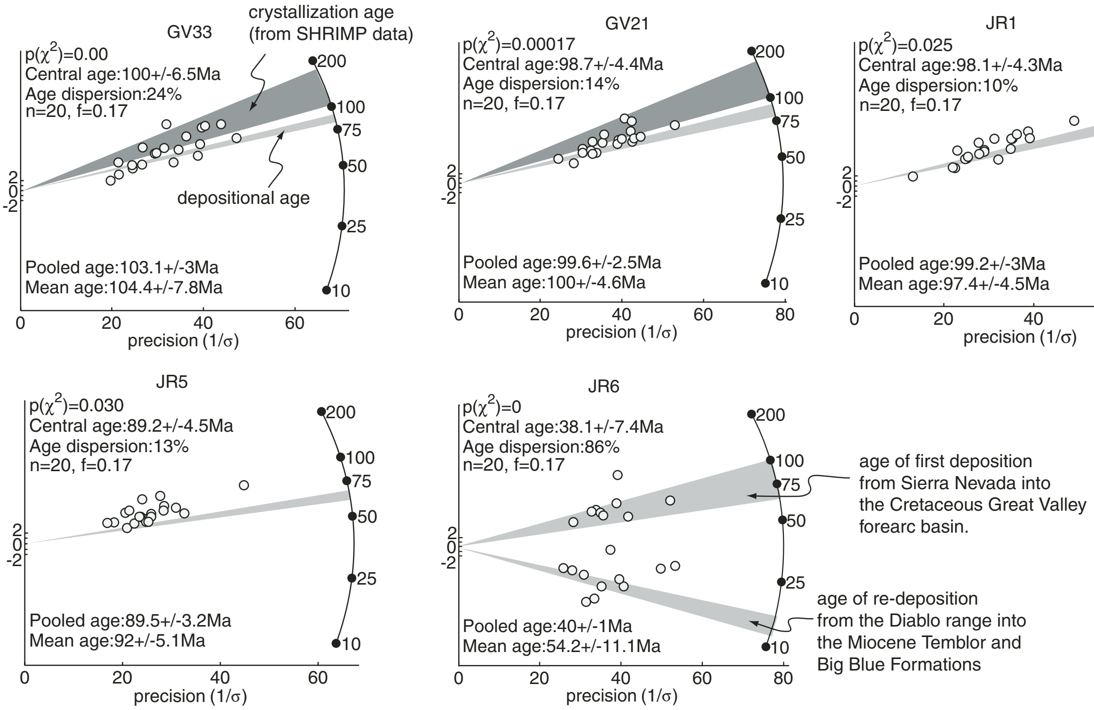

Figure 3. Zircon fission track radial plots. As in Figure 2, the light gray bands represent depositional ages. The dark gray bands mark the crystallization ages, as measured by DeGraaff-Surpless et al. (2002) using the U/Pb method on zircon. Other abbreviations as in Figure 2.

plate (Dumitru, 1988). The "Christie #1" well is located near sample JR2. Three cores taken in "Christie #1" at 1250–1380 m are characterized by vitrinite reflectance values of ~1.9%–2.0% or maximum paleotemperatures of ~180 °C, very hot for the Great Valley Group. However, apatite fission track sample JR2, located ~1500–2000 m upsection from the "Christie #1" vitrinite reflectance samples, has not been reset. Assuming a thermal gradient similar to that inferred from the "Joaquin Ridge #1" well, this would lower the predicted vitrinite reflectance value for sample JR2 to about Ro = 1.3% (maximum paleotemperature ~150 °C). This rough estimate, if correct, conflicts with the observation that apatite fission track sample JR2 has not been completely annealed. This would imply that the high Ro values of "Christie #1" are due to a thermal anomaly, that the thermal gradient in this well was not equal to that of "Joaquin Ridge #1," and/or that this gradient was not linear. Although not reset, sample JR2 has been at paleotemperatures above those of samples JR3, JR4, and JR5. The high paleotemperatures of the "Christie #1" samples are unlikely to be the result of simple burial but are instead interpreted to be the result of a middle Miocene heating spike, associated with the upward protrusion and tectonic denudation of the New Idria serpentinite body.

IMPLICATIONS FOR THE POSTDEPOSITIONAL HISTORY OF THE GREAT VALLEY GROUP

Partial annealing of apatite fission track sample JR2 and complete annealing of JR1 at ca. 14 Ma by burial-induced heating alone is improbable for at least two reasons.

1. Maximum temperatures of the Great Valley Group on Joaquin Ridge did not coincide with the timing of maximum burial. In the Panoche Hills area, north of Joaquin Ridge, subaerial exposure and erosion occurred intermittently in the Paleocene and Eocene and almost continuously since the Oligocene (Moxon, 1990; Bartow, 1991). Furthermore, deposits as old as the Eocene Kreyenhagen Formation of Oil Canyon, 10 km north of Coalinga, contain biogenic silica in the form of opal-A (Milam, 1985), indicating that they were never heated above 28 °C–56 °C (Murata and Larson, 1975). The total thickness of Tertiary deposits presently outcropping at the nose of Joaquin Ridge is only ~1.5 km (Dibblee, 1971).

2. The age of the annealing (as dated by sample JR1) exactly coincides with the exhumation of the nearby New Idria serpentinite body (see below). It is unlikely that this coincidence is mere chance.

An alternative explanation for the vitrinite reflectance and fission track data is that the Great Valley Group was heated by the serpentinite diapir. Three lines of evidence suggest that the serpentinite body was hot when it breached the surface. Most importantly, mineral assemblages of Franciscan inclusions in the serpentinite body indicate that it rose from depths of as much as ~20 km, which even under the lowest thermal gradients would make them relatively hot (>200 °C; Coleman, 1996a, 1996b). Second, the serpentinite diapir rose very rapidly. Evidence for the massive size and sudden nature of this event is contained in the middle Miocene Big Blue Formation, which crops out ~15 km east of the serpentinite dome. The Big Blue Formation consists almost entirely of serpentinite clasts, some of which are house sized (Anderson and Pack, 1915). Paleocurrents indicate flow toward the east, in the opposite direction of the "normal" Great Valley Group paleocurrents (Casey and Dickinson, 1976; Bate, 1985). The facies gradient from sheared protrusive serpentinite through braided stream deposits to marine tidal flat facies evinces an eastward-facing paleoslope (Bate, 1985). The fluvial deposits preserving paleocurrents were shed from the gradually spreading flank of a New Idria serpentinite protrusion that breached the surface to form a domelike mass that spread laterally as additional serpentinite was supplied to the surface by upward diapiric flowage from within the crust. If the serpentinite body rose to the surface extremely rapidly, it is likely to have remained hot all that time. Finally, serpentinization reactions are exothermic:

Antigorite: $34Mg_2SiO_4 + 51H_2O \rightarrow Mg_{48}Si_{34}O_{85}(OH)_{62} + 20Mg(OH)_2$

Chrysotile: $2Mg_2SiO_4 + 3H_2O \rightarrow Mg_3Si_2O_5(OH)_4 + Mg(OH)_2$

Both of these reactions require a lot of water. This could be the reason the serpentinization did not happen before the middle Miocene. At that time, the Mendocino triple junction passed the latitude of Coalinga. The faulting and folding caused by the San Andreas fault might have introduced a pathway for fluids in the ophiolitic crust that underlies the Great Valley Group. Both the reaction enthalpy ΔH of the serpentinization reactions and the heat capacity C_p of the serpentinite minerals vary with temperature (Holland and Powell, 1998). At the conditions relevant to the New Idria serpentinite dome, the pressure effect is negligible. Making the simplifying assumption that serpentinization of the entire body occurred at the same time, we can calculate a first-order approximation of the maximum temperature increase that could be caused by the serpentinization:

$$\Delta T = \Delta H / C_p.$$

The evolution of this reaction temperature as a function of ambient temperature is shown in Figure 5. The lower the ambient temperature, the more exothermic the serpentinization reactions are, but above a few hundred degrees, they can even become endothermic. The buffering thermodynamics of the serpentinization reactions are such that the serpentinite body must have been hot when it formed, either because the ambient temperature was high or because of its own reaction heat. A rough estimate of the thermal effect that a hot, spherical body the size of the New Idria diapir would have on the adjacent country rock can be calculated assuming simple conductive cooling:

$$T(r,t) = \frac{T_i}{2}\left(erf\left(\frac{R-r}{2\sqrt{\kappa t}}\right) + erf\left(\frac{R+r}{2\sqrt{\kappa t}}\right)\right) + T_i \frac{\sqrt{\kappa t}}{r\sqrt{\pi}}\left(exp\left(-\frac{(r+R)^2}{4\kappa t}\right) - exp\left(-\frac{(r-R)^2}{4\kappa t}\right)\right)$$

with T_i being the initial temperature difference between the serpentinite and the country rock, R the radius of the sphere (~10 km), r the distance from the center of the sphere, κ the diffusivity (~10^{-6} m^2/s), and t, time (modified from Carslaw and Jaeger, 1959). Figure 6 shows the result of this calculation. It indicates that during ~10^6 yr after the intrusion of a hot body the size of the New Idria diapir, the rocks within a few km of the contact would experience a transient heating spike. Added to the preexisting background geothermal gradient, this spike could explain both the vitrinite reflectance data and the apatite fission track annealing behavior. Thermal halos around protrusive serpentinite bodies of west-central California have been described by Murata et al. (1979), who traced the Marca Shale Member of the Upper Cretaceous and Paleocene

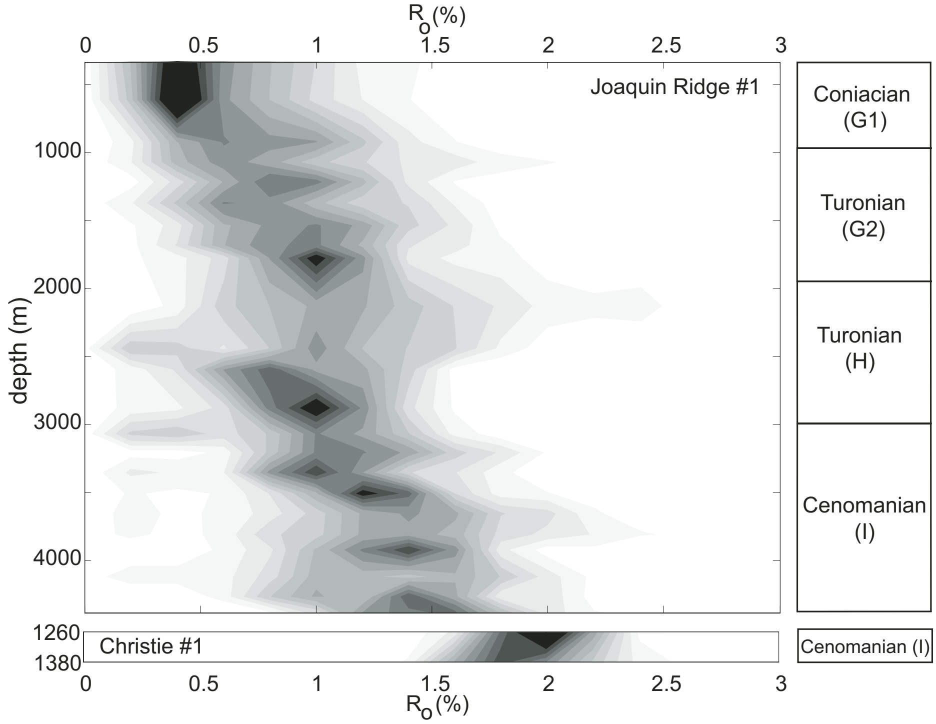

Figure 4. Contoured vitrinite reflectance results for two wells on Joaquin Ridge (see Fig. 1 for their locations).

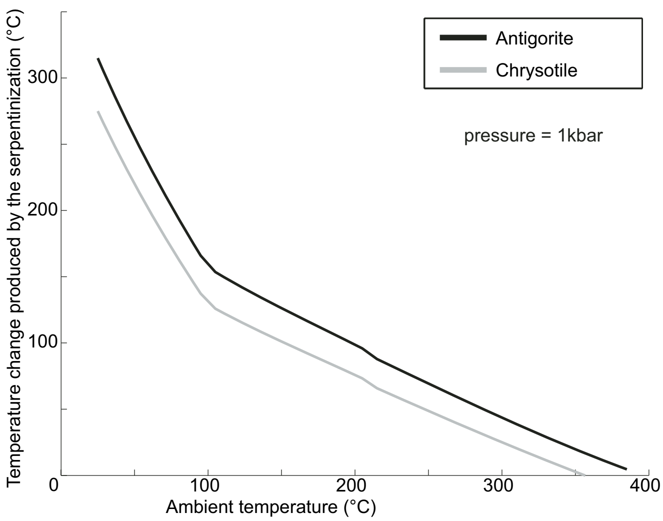

Figure 5. Serpentinization reactions are less exothermic as the ambient temperature increases. The New Idria serpentinite body must have been relatively hot for one of two reasons: (1) it formed under high ambient temperatures, or (2) it generated the heat itself.

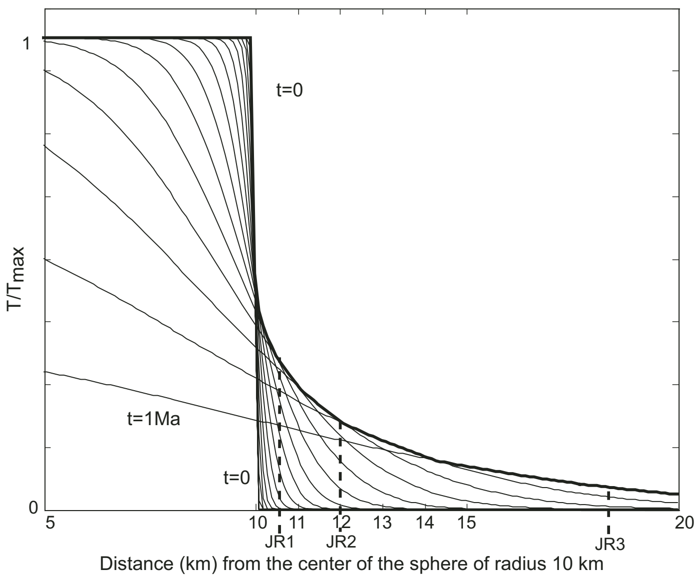

Figure 6. Conductive cooling of a hot sphere surrounded by a cooler material creates a transient heating signal in the latter. The sphere (radius = 10 km, left side of the figure) represents the New Idria serpentinite body, rapidly rising in a Great Valley Group country rock (right side of the figure) with thermal diffusivity $\kappa = 10^{-6}$ m^2/s. The thin black lines show the evolution with time (from 0 to 1 Ma) of the thermal contact. The thick black line connects the maximum temperatures reached at different distances. The location of three of the fission track samples is also marked on the figure.

petroliferous Moreno Shale over a distance of 120 km and found that biogenic silica in this unit was cristobalitic everywhere, except for the northern flank of Joaquin Ridge, where it comes within less than 1 km of the New Idria serpentinite body. This is the only place where quartz-phase silica exists, indicating maximum temperatures less than ~80 °C.

We have not attempted to model rigorously petroleum generation and trapping in the Vallecitos syncline, but current studies by the United States Geological Survey may better constrain the petroleum history (Peters et al., 2005). Nevertheless, various relations provide general constraints on petroleum generation and accumulation and thus provide a point of comparison for our interpretation. Biomarkers recently collected from the Vallecitos oil field by the United States Geological Survey show biomarker and isotope compositions indicative of upper Eocene Kreyenhagen source rocks (K.E. Peters, 2004, personal commun.). However, the small pools in the syncline occur mainly in fault traps located under the Kreyenhagen Formation (California Division of Oil and Gas, 1982). Therefore, the Maastrichtian-Danian Moreno Formation might be a more plausible source of these hydrocarbons. Although not understood in detail, the fault traps likely developed when the syncline folded in the late middle Miocene (Rentschler, 1985). Thus, maturation, migration, and entrapment likely occurred no earlier than the late Miocene, consistent with our data and interpretation.

IMPLICATIONS FOR THE CRETACEOUS HISTORY OF THE SIERRA NEVADA

Most geochronological methods have a closure temperature (Dodson, 1973). However, in the apatite fission track method, for example, there is not one distinct temperature but a rather diffuse zone over which the geochronological system "closes." Nevertheless, for our purposes, the rather crude concept of a closure temperature is still useful. For the U/Pb system in zircon, the closure temperature is as high as ~900 °C (Dahl, 1997; Miller et al., 2003). The closure temperature of the zircon fission track method is somewhere between ~230 °C and 310 °C (Wagner and Van den Haute, 1992; Tagami and Dumitru, 1996). For the apatite fission track method, a closure temperature of ~100 °C can be used, although this value varies with apatite chemistry (e.g., Gleadow and Duddy, 1981).

A frequent practice in igneous and metamorphic geochronology is the simultaneous use of several dating techniques on the same sample. A graph of apparent age versus closure temperature is then used to estimate the cooling history of such a sample (e.g., Harrison and McDougall, 1980). A similar approach can be used for detrital samples. Conservatively assuming that the tops of the plutons in the southern Sierra Nevada were emplaced at 2–3 km depth (Ague and Brimhall, 1988), Surpless (2001) argued that the relatively short minimum lag times between the U/Pb zircon ages and the depositional ages of 3–15 Ma indicated rapid exhumation of the Cretaceous Sierra Nevada at rates of ~0.6–1 mm/yr. We can extend this method to the lower temperature thermochronometers of Table 1.

If we had access to double-dated grains, as in Rahl et al. (2003), estimating the probability distribution of provenance cooling rates would be a trivial exercise. However, because of the uncontroversial provenance of our samples and the fact that the southern Sierra Nevada can be considered a structurally more or less homogeneous fault block, we might be able to proceed without such data. Thus, we can get a first-order estimate of the cooling rates by looking at the time lag between the mean (or central) ages of two thermochronological grain-age populations (e.g., apatite and zircon fission tracks) or at the time lag between the mean (or central) age of a grain-age population and the depositional age of the sample.

Doing this for the thermally unreset data of Table 1 yields seven apparent cooling rates that do not show any systematic variation with depositional age. The weighted mean of these estimates yields an apparent cooling rate of ~21 °C/m.y. Depending on the thermal gradient (22–40 °C/km; Rothstein and Manning, 2003), this then corresponds to exhumation rates of ~0.5–1 mm/yr, which agrees with the estimates of Surpless (2001) and Ague and Brimhall (1988). These are rather high rates, but this does not come as a surprise when we consider the amount of sediment deposited in the Great Valley Group at the time.

SUMMARY AND CONCLUSIONS

Figure 7 summarizes the history of the Great Valley Group near Coalinga and New Idria.

1. The U/Pb and fission track ages of unreset apatite and zircon show that the southern Sierra Nevada underwent rapid and steady exhumation during the Late Cretaceous. As the exhuming mountain range was eroded, it shed sediments in the Great Valley forearc basin. These sediments are presently exposed in a homocline on the eastern flank of the Diablo Range.

2. The Great Valley Group turbidites were buried to great depths but under refrigerated thermal gradients produced by Franciscan subduction (Dumitru, 1988).

3. During the middle to late Miocene (ca. 12–14 Ma), the Mendocino triple junction migrated from south to north to the west of Coalinga (Johnson and O'Neil, 1984). Its passage was associated with minor igneous activity, dated at ca. 12.8 Ma (Obradovich et al., 2000).

4. The faulting and folding caused by the San Andreas fault introduced fluids in the ophiolitic crust that underlies the Great Valley Group. These fluids reacted with the peridotitic rocks and formed serpentine minerals. Because of its low density and the relative ease with which it deformed, the serpentinite body rose rapidly. The combined effect of the folding and the rise of the serpentinite diapir caused substantial uplift of Joaquin Ridge and secondary folding and faulting in the adjacent Vallecitos Syncline (Rentschler, 1985). Paleocurrent directions reversed, and resedimentation of Great Valley Group detritus formed the middle Miocene Temblor and younger formations. The Big Blue came to the surface as a rising hot protrusion, heating the Great Valley Group on the way up, and spreading at once over the countryside.

5. Because it formed at great depth or because the serpentinization reactions are exothermic, the New Idria serpentinite body was hot when it approached the surface, forming a thermal halo. The heating was sufficient to completely anneal apatites in the surrounding country rock and explain the relatively high vitrinite reflectance values that are observed near the contact between the Great Valley Group and the serpentinite body.

6. The heat released by the serpentinite produced diagenetic alteration of cristobalitic to quartzose silica in the Moreno Shale, which underlies the Vallecitos syncline (Murata et al., 1979), pushing the source rocks of this basin into the thermal oil window.

7. Alternatively, it is possible that syenitic intrusions in the serpentinite body provided the "missing" heat source. However, we think that the serpentinite body itself is a more realistic alternative because it is size rather than temperature that determines the amount of time it takes for a hot body to cool and the size of the resulting thermal halo (time ≈ size^2/κ; e.g., Turcotte and Schubert, 1982). A small igneous intrusion may have been hotter than the serpentinite protrusion, but it would not be felt as long and as far into the country rock.

8. Deformation and folding continue in the Joaquin Ridge area today, witnessed by the 1983 Coalinga earthquake. However, gravity data suggesting that the root of the New Idria body is only a few kilometers deep imply that the exhumation of the serpentinite diapir has almost stopped (Byerly, 1966; Casey and Dickinson, 1976).

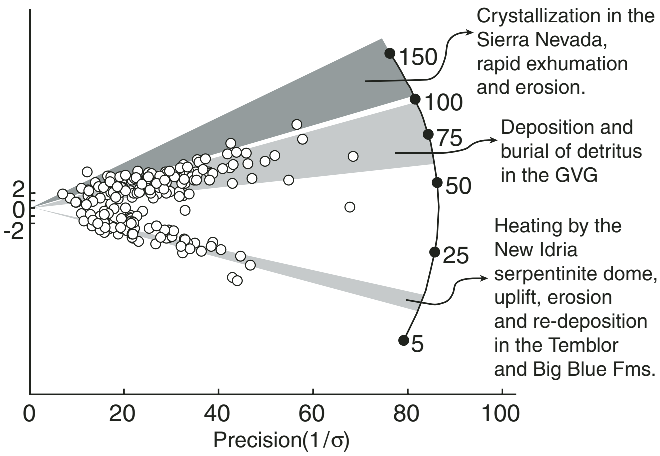

Figure 7. The history of Joaquin Ridge sediments summarized on a single fission track radial plot, compiled from all the apatite fission track data. GVG—Great Valley Group.

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