Bulk arc strain, crustal thickening, magma emplacement, and mass balances in the Mesozoic Sierra Nevada arc

CAO, Wenrong ^{a,b,*}, caowenrong@gmail.com, phone: +1-213-568-7185

PATERSON, Scott ^a, paterson@usc.edu

SALEEBY, Jason ^c, jason@gps.caltech.edu

ZALUNARDO, Sean ^a, seanzalunardo@gmail.com

^a Department of Earth Sciences, University of Southern California, 3651 Trousdale Parkway, Los Angeles, California 90089-0740, USA

^b Department of Earth Science, Rice University, 6100 Main St. MS-126, Houston, Texas, 77005, USA

^c Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91125, USA

Abstract

Quantifying crustal deformation is important for evaluating mass balance, material transfer, and the interplay between tectonism and magmatism in continental arcs. We present a dataset of >650 finite strain analyses compiled from published works and our own studies with associated structural, geochronologic, and geobarometric information in central and southern Sierra Nevada, California, to quantify the arc crust deformation. Our results show that Mesozoic tectonism results in 65% arc-perpendicular bulk crust shortening under a more or less plane strain condition. Mesozoic arc magmatism replaced ~80% of this actively deforming arc crust with plutons requiring significantly

greater crustal thickening. We suggest that by ~85 Ma, the arc crust thickness was ~80 km with a 30-km-thick arc root, resulting in a ~ 5km elevation. Most tectonic shortening and magma emplacement must be accommodated by downward displacements of crustal materials into growing crustal roots at the estimated downward transfer rate of 2-13 km/Myr. The downward transfer of crustal materials must occur in active magma channels, or in “escape channels” in between solidified plutons that decrease in size with time and depth resulting in an increase in the intensity of constrictional strain with depth. We argue that both tectonism and magmatism control the thickness of the crust and surface elevation with slight modification by surface erosion. The downward transported crustal materials initially fertilize the MASH zone thus enhancing to the generation of additional magmas. As the crustal root grows it may potentially pinch out and cool the mantle wedge and thus cause reduction of arc magmatism.

Keywords

Finite strain, Crustal thickening, Magma emplacement, Mass balance, Mesozoic Sierra Nevada arc

1. Introduction

Quantitative mass balances in continental margin arcs, during orogeny, mountain building, magma emplacement, growth of deep crustal roots, and lower crustal

47 delamination have been difficult to calculate for four main reasons: 1) the addition of
48 large volumes of magma either overprints or entirely removes large volumes of
49 pre-existing crust, thus removing significant geologic evidence about host rock
50 displacements; (2) lower crustal sections of continental arcs are rarely exposed; (3) the
51 strong deformation and metamorphism present in these environments masks evidence
52 of displacements, and (4) widespread markers of finite strain and/or regional
53 displacements are needed and not always available in these settings. However there
54 remains a strong motivation to collect such data in order to address a series of
55 interrelated questions: (1) Can we better quantify both strain and bulk shortening in
56 continental arcs to clarify their role in mountain building, development of crustal roots,
57 and delamination? (2) How is local space made for the emplacement of plutons? (3)
58 Can we better constrain crustal thickening through time and thus determine to what
59 degree tectonic processes are linked to the “tempos” of non-steady state magmatism
60 (e.g., Ducea, 2001; DeCelles et al, 2009; Paterson and Ducea, 2015)?

61

62 Over the past 40 years a number of studies have completed extensive dating of both
63 plutonic and metamorphic rocks and collected strain data in or near the Sierra Nevada
64 batholith of California (e.g. Tobisch 1977; Paterson et al. 1989; Saleeby, 1990;
65 Busby-Spera and Saleeby, 1990; Tobisch et al., 1995; Mundil et al., 2004; Schweickert
66 and Lahren, 2006; Saleeby et al., 2008; Memeti et al., 2010; Barth et al., 2012;
67 Chapman et al., 2012; Paterson et al., 2014; Cao et al., 2015). Over the past 10 years,
68 we have renewed these efforts by continuing modern U/Pb dating of both plutonic and

metavolcanic units, plus detrital zircon dating of metasedimentary units, measuring strains in a wide variety of rock types, and integrating our structural studies to estimate a temporal history of bulk shortening throughout the Mesozoic. Below, after introducing Sierran geology, we present an integration of over 650 finite strain analyses across the central and southern Sierra Nevada with our recent structural and geochronologic data to quantify the strain and mass balances.

2. Geological background

2.1 Sierra Nevada arc

The Sierra Nevada in California ([Figure 1A](#)) represents a long-lived (~170 million years) Mesozoic continental arc formed by subduction between Farallon and North American plates. The plutonic part of the arc is referred to as the Sierra Nevada Batholith, a collective term for mainly intermediate plutons of Mesozoic age ([Figure 1B](#)). Episodic magmatism in the Sierra Nevada arc is well documented, and three magmatic flare-ups bounded by four magmatic lulls are recognized ([Ducea, 2001](#); [DeCelles et al., 2009](#); [Paterson and Ducea, 2015](#)).

Host rock pendants representing the Mesozoic volcanic and sedimentary carapace of the Sierra Nevada arc are locally preserved. In the central Sierra Nevada (36.5° N to 38.5° N), strain data have been collected and compiled from nine stratigraphically

related but spatially disconnected pendants (Figure 1C): Piute Meadow, Saddlebag Lake, Ritter Range, Jackass Lake, Iron Mountain, Mountain Morrison, Mountain Goddard, Oak Creek and Boyden Cave. Lithology of these strata is mainly andesitic to rhyodactic, clastic volcanic rocks with local, intra-arc sedimentary rocks. (e.g. Brook, 1977; Tobisch et al., 1982; Greene and Schweickert, 1995; Tobisch et al., 2000; Sharp et al., 2000; Schweickert and Lahren, 2006). The paleo-depths at which these host rock pendants are preserved range from 2 to 4 kbar, approximately 7 to 14 km (3.64 km/kbar, assume density 2.75g/cm³) (Anderson et al., 2008; Paterson et al., 2014; Nadin and Saleeby, 2008; Chapman et al., 2012).

In the southern Sierra Nevada (36°N to ~35.8°N), deeper levels of the arc are progressively exposed southwards to ~10 kbar (Pickett and Saleeby 1993; Saleeby et al., 2003; Nadin and Saleeby, 2008; Chapman et al., 2012). Extreme transposition and migmatization are pervasive in pendant rocks at >7 kbar levels, thereby inhibiting strain measurements. Strain data have been collected from two host rock pendants, Lake Isabella and Piute Lookout (Figure 1B). Both of these pendants underwent peak metamorphism and deformational conditions during ca. 100-90 Ma magmatism, Lake Isabella at ~4 kbar conditions (Dixon, 1995; Ague and Brimhall, 1988a, Saleeby, unpubl.) and Piute Lookout at ~7 kbar conditions (Saleeby et al., 2008).

These host rock pendants in the central and southern Sierra Nevada had experienced multiple phases of deformation during the Mesozoic. Field evidence of

arc-perpendicular shortening, and vertical stretching are found in many host rock pendants. This regional strain pattern is responsible for the NW-striking, steeply dipping bedding, regional foliation, and steeply plunging lineation (e.g. Tobisch and Fiske, 1982; Sharp et al., 2000; Saleeby et al., 2003; Schweickert and Lahren, 2006) (Figure 2). The tilted bedding and metamorphic foliation and lineation are formed time-transgressively during the Mesozoic (Tobisch and Fiske, 1982; Cao et al., 2015). SW- or NW-vergent thrusts and strike-slip faults (e.g. Lahren and Schweickert, 1995; Tobisch et al., 2000; Schweickert and Lahren, 2006; Cao et al., 2015) also developed during the Mesozoic. Dextral transpressional shear zones (e.g. Sierra Crest Shear Zone) developed in the Late Cretaceous as the subduction between Farallon and North America plates become more oblique (e.g. Greene and Schweickert, 1995; Tikoff and Saint-Blanquat, 1997; Nadin and Saleeby, 2008). Host rock pendants also underwent sub-greenschist to amphibolite facies regional metamorphism and local contact metamorphism in pluton aureoles (Tobisch et al., 1977; Barton and Hanson, 1989; Hanson et al., 1993; Alberty, 2006).

2.2 Western Metamorphic Belt

The Western Metamorphic Belt (WMB, Figure 1C) is a belt of metamorphic rocks along the western edge of the Sierra Nevada. The WMB consists of terranes accreted to the western margin of the North American continent and/or formed along the continental margin and infolded with North American derived marine sediments (Clark,

1964; Edelman and Sharp, 1989; Tobisch et al., 1989; Schweickert et al., 1999; Snow and Scherer, 2006). Terrane accretion began in the Permo-Triassic and ceased by Middle or Late Jurassic (e.g. Snow and Scherer, 2006; Dickinson, 2008; Saleeby, 2011). The accretion resulted in a major boundary, Permo-Triassic Foothills suture, separating displaced continental passive margin fragments from accreted terranes and several intra-arc thrust faults, the Late Jurassic Melones fault and Bear Mountain fault separating different tectonic regimes (Figure 1C, e.g. Clark, 1964; Schweickert et al., 1977; Miller and Paterson, 1991; Saleeby and Dunne, 2015). In this paper, the WMB in the central Sierra is divided into the Foothills and Eastern Belts (Figure 1C).

3. Description of datasets

3.1 Sample locations

Strain data are mainly collected from host rocks across the central Sierra Nevada with a small subset collected in the southern Sierra Nevada. Boxes in Figure 1B and Figure 1C show the area where strain data were collected. Compiled strain data and references of data source are provided in the Supporting Materials.

3.2 Methods of finite strain measurement

The strain measurements compiled in this paper have been obtained by a number of

different people using several different methods (e.g., Tobisch et al, 1977, 2000; Longiaru, 1987; Paterson et al., 1989; Albertz, 2006; Horseman et al., 2008). A great majority of the strains were determined from clastic objects that can be approximated by ellipsoidal shapes (e.g., pebbles in conglomerates, lithic fragments and pumice lapilli in volcanics, sand grains in sandstones). These were analyzed using one of the following two approaches:

1. When logistically feasible, oriented samples with clastic objects were collected and brought back to the lab. Three perpendicular surfaces either parallel to principal planes (Tobisch et al, 1977) or arbitrary (Paterson et al., 1989) were cut in each sample and then labeled with sample coordinate axes. In each 2D surface the orientation of the long axis of clasts relative to the sample coordinates axes and the long axis/short axis ratio of least ~25-50 objects per surface were measured (by hand in older studies and digitally from images of the surface in recent studies). R_f/ϕ techniques (e.g., Shimamoto and Ikeda 1976; Lisle 1979, 1985) were used to determine the shape and orientation of 2D ellipses for each surface and then combined into a single 3D ellipsoid (Shimamoto and Ikeda, 1976; Miller and Oertel, 1979). The orientations of 3D ellipsoids were reoriented to true geographic orientations using the sample orientation data. Given sufficient sample sizes, precision errors are in the range of 5-10% (see Miller and Oertel, 1979 for error estimates).

2. When not feasible to collect oriented samples, surfaces axial ratios and orientations

of clastic objects were obtained on principal planes. Since each two principal planes share one axis of strain, a 3D ellipsoid ($X \geq Y \geq Z$) can be determined by setting $Z = 1$ and determining the ratios of X/Z and Y/Z from the 2D averages (Ramsay, 1967; Lisle, 1977; Paterson, 1983). Strain ellipsoid orientations for these field determinations are assumed to be parallel to the cleavage (X-Y plane) and mineral or clast lineation (X-axis). Precision errors using this approach were in the range of 5-20% when fitting 2D surfaces to 3D results.

The above approaches assume constant volume strain at the scale of individual clasts. In most cases we see little evidence of large volume changes (e.g., fibers in pressure shadows or dissolution of clasts), thus supporting the above assumption. These approaches also ignore the existence of any primary fabrics in samples, which may be present if clasts were even weakly aligned during deposition and burial. Lisle (1979), Paterson et al. (1989) and Paterson and Yu (1993) discuss approaches for correcting final strains for the presence of primary fabrics. Note that many primary fabric ellipsoids have small axial ratios (Paterson et al., 1995). Since these corrections are only available for some of the data discussed in this paper, we only use and compare uncorrected data. Finally R_f/ϕ techniques assume that there is no difference in the viscosity of clasts and matrix and thus clast strain equals bulk strain of the rock. Any effect of viscosity differences decrease as the percent of clasts in samples increase and as the sample becomes clast supported (i.e., at around >50% clasts). The existence of viscosity contrasts can be evaluated by examining whether the cleavage is deflected

around clasts or passes undeflected through them. Most samples in this study show little evidence of significant viscosity contrasts.

Strains in slate and phyllite samples from the WMB were determined by Paterson et al. (1989) using the March method (March, 1932; Oertel, 1983). Precision errors are in the range of 5-10%. These samples record different strains than nearby (at cm to m scale) samples measured by R_f/ϕ techniques and are interpreted to include a component of flattening strain formed during burial and dewatering (Oertel, 1983; Paterson et al., 1989; Paterson et al., 1995).

A very few of the strains presented in this paper were determined using the Fry method (Crespi 1986; Erslev, 1988). In this method 3 perpendicular cuts are made in oriented samples and for each face the distance and direction between centers of objects are measured and used to construct an “all-objects-separation” plot. The inner cut-off of plotted points is used to determine the ratio and orientation of 2D ellipses. These ellipses are combined into a 3D ellipsoid using both Shimamoto and Ikeda (1976) and Miller and Oertel (1979) techniques. Precision errors for Fry data are in the range of 5-20% depending on the number of objects used (see Crespi, 1986 for error estimates).

3.3 Geochronologic data

Numerous studies have determined absolute ages of plutons and host rocks in the

central Sierra Nevada (e.g. Bateman and Chappell, 1979; Busby-Spera and Saleeby, 1990; Schweickert and Lahren, 1993; Tobisch et al., 1995; Schweickert and Lahren, 2006; Saleeby et al., 2008; Memeti et al., 2010; Chapman et al., 2012; Barth et al., 2012; Paterson et al., 2014; Cao et al., 2015). We compiled these ages and integrated them with the strain data when available. We focused on U/Pb zircon ages of plutons and volcanics or interpreted ages of deposition for sedimentary units based on youngest age peaks of detrital zircons. For host rocks without dated absolute ages, we assign approximate ages based on stratigraphic correlations. Mesozoic rock ages are binned into age groups (see Section 4.4). While the majority of rock ages in our dataset are Mesozoic, a few Paleozoic ages were included from continental margin tectonic slices now residing in the Sierra Nevada.

3.4 Geobarometric data

The Aluminum-in-hornblende barometer (Johnson and Rutherford, 1989; Schmidt, 1992; Holland and Blundy, 1994; Anderson and Smith, 1995;) was used to determine emplacement pressure/depth of plutons in the central Sierra Nevada (e.g. Ague and Brimhall; 1988a, 1988b; Ague and Brandon, 1992; Pickett and Saleeby, 1993; Ague, 1997; Saleeby et al., 2003; Nadin et al., 2008; Saleeby et al., 2008; Anderson et al., 2008; Chapman et al., 2012; Paterson et al., 2014). We compiled the geobarometric data and integrated them with the strain data when available and used the emplacement pressure of nearby intruding plutons as the maximum paleo-depth of host rocks.

Paleo-depths are then binned into different groups (see Section 4.5). The rocks of shallower paleo-depth (1-3 kbars) are from the central Sierra Nevada whereas rocks of deeper paleo-depth (4-7 kbars) are from the southern Sierra Nevada.

4. Characteristics of finite strain in Mesozoic Sierra Nevada

4.1 Orientations of Metamorphic foliation and lineation in central Sierra Nevada

Metamorphic foliation and lineation are well developed in the WMB and host rock pendants in the central Sierra Nevada. Foliation type varies from slaty cleavage to schistosity to mylonitic in ductile shear zones. Mineral lineations are defined by aligned stretched clasts, pebbles, and mineral grains. [Figure 3](#) shows the metamorphic foliation and lineation from the WMB and several host rock pendants in the central Sierra Nevada. The measurements are from both Paleozoic and Mesozoic rocks but the great majority are from Mesozoic rocks. The cylindrical best fits are calculated ([Allmendinger et al., 2012](#); [Cardozo and Allmendinger, 2013](#)). Regional foliations in most of the areas dip steeply and strike NW to NNW parallel to the trend of the arc. The foliation best fits only vary within 17° between most areas except in the Iron Mountain and Jackass Lake areas, where the foliation best fits strike slightly more northerly. Lineations in all areas are steeply plunging (plunge > 70°). We interpret the foliations and lineations to approximate the X-Y planes and X-axes of finite strain ellipsoids (X, Y, Z are the longest, intermediate, and shortest axes of a strain ellipsoid, respectively).

Comparisons of calculated X-Y planes and X-axes of strain from samples analyzed in the lab support this assumption (e.g., Paterson et al., 1989). The consistent orientations of regional foliation and lineation across the central Sierra Nevada suggest a more or less co-axial arc-perpendicular ductile shortening during Late Paleozoic and Mesozoic, or approximate the shear plane orientations of arc-parallel shear zones and thus the X-Y plane of high shear related strains, which is consistent with observations from many other researchers (e.g. Tobisch and Fiske, 1982; Greene and Schweickert, 1995; Sharp et al., 2000; Tobisch et al., 2000; Albertz, 2006; Horsman et al., 2008; Cao et al., 2015).

4.2 General patterns of finite strain in Mesozoic Sierra Nevada

We plotted 612 strain measurements with age constraints on a Hsu diagram (Hus, 1966; Hossack, 1967) (Figure 4). Most measurements plot at a strain magnitude less than 1.8 with a few, typically in shear zones, ranging up to 3.4. Lode's ratios (Lode, 1926) are quite variable with most between -0.75 and 1. Triassic and Jurassic rocks show slightly higher magnitudes with very few (4 points) >3.0. All Cretaceous rocks show magnitudes less than ~2.1. The data are slightly concentrated in the flattening region ($\gamma > 0$) but show an overall symmetrical distribution.

We also plotted strain measurements in the space of X- and Z-extension percent (Figure 5). The central bold red curve ($Y=0$) represents the relationship between X- and

Z-extension under plane strain. The thin curves on either side of the central bold curve show the relation between X- and Z-extension if volume loss and/or Y-extension occurred during deformation. [Figure 5](#) shows that the majority of data clustering about the red curves with X-direction extension ranging from <5% to about 500%. Most data cluster in the space defined by 25%-150% X-direction extension, 20%-70% Z-direction shortening, and between Y=0% and Y=20% extension curves. Average Z-direction shortening is 49.4% with a standard deviation of 16.7%. Since the volume change has not been determined by strain measurements in the present study, the deviation of points plotted off the central “plane strain” curve in [Figure 5](#) is interpreted as the results of length change along the Y-direction. In reality, both length change along the Y-direction and volume change can cause the deviation from the plane strain. Most of the measurements from slate fall in the flattening field on the diagram, which is interpreted as recording primary compaction during burial and dewatering ([Pfiffner and Ramsay, 1982; Paterson et al., 1995](#)). This is consistent with the observation in slates that cleavage is commonly parallel to bedding. If we correct for this early compaction, the resulting tectonic strains shift upwards and approach the central “plane strain” curve ([Paterson et al., 1989](#)).

In summary, [Figure 5](#) suggests that although Sierran strain measurements have a wide range of X- and Z- direction length change, most points are within the $Y=0 \pm 20\%$ range suggesting roughly plane strain at the arc scale. Since the Y-Z plane parallels regional NW-striking foliation and X-direction parallels lineation ([Figure 3](#)), [Figure 5](#)

suggests the finite ductile strain in these pendant rocks is 50% arc-parallel shortening,
~100% vertical thickening and limited arc-parallel length and/or volume change.

4.3 Strain vs. rock types

To investigate the relationship between strain and rock type in the central Sierra Nevada, we plot strain data from 9 major rock types (pebbly mudstone, sandstone, conglomerate, slate/phyllite, andesite, rhyolite/dacite, tuff/lapilli stone, volcanic breccia, and unclassified volcanics) on Hsu diagrams (Figure 6). The strain plot of pebbly mudstone (Figure 6A) shows strain magnitude below 1.8 and variable strain shape. Sandstone (Figure 6B) shows a concentrated distribution in strain magnitude less than 1.2 and variable strain shape. Conglomerate (Figure 6C) shows two groups: one with strain magnitudes less than 1.8 and with slightly more flattening shape, the second with higher strain magnitudes higher than 2.1 and constrictional strain shape. The strain measurements in the first and second groups are from conglomerates in the central and southern Sierra Nevada, respectively. Slate and phyllite (Figure 6D) cluster in the flattening regime, probably due to compaction during sedimentation and lithification or volume loss during plane strain deformation (Ramsay and Wood, 1973; Pfiffner and Ramsay, 1982). Andesite (Figure 6E) shows strain magnitudes ≤ 1.5 based on 12 data points. Rhyolite and dacite (Figure 6F), and tuff-lapilli stone (Figure 6G) show slightly concentrated distribution along the Lode's ratio = 0 line suggesting that these rocks are deformed close to "plane strain". Volcanic breccia (Figure 6H) shows

higher strain magnitudes with many points above 1.8, plotting slightly in the flattening regime. [Figure 6I](#) shows volcanic rocks that are not classified into composition groups. Strain magnitudes of these volcanic rocks are mostly below 2.1 and are slightly more distributed in the flattening regimes.

In summary, [Figure 6](#) shows that rock type affects strain magnitude and strain shape but not significantly. Strain magnitude is typically less than 1.8. Strain shape is generally evenly distributed in constrictional and flattening regions but for some rock types (slate and phyllite, volcanics unclassified), flattening, presumably due to compaction during lithification, is more common.

4.4 Strain vs. rock ages

To investigate the relationship between strain and rock age in the central Sierra Nevada, we plot the strain magnitude and Lode's ratio against rock ages ([Figure 7](#)). Notice the age of a rock is only an upper limit of the age of deformation but can still be used to bin data into different lengths of time.

Strain magnitudes ([Figure 7A](#)) of Triassic, Jurassic, and Cretaceous rocks are fairly scattered but the overall trend is a slight decrease in strain magnitude with younging. The trend of decreasing strain magnitude with younging can be seen both from the distribution of individual data and the trend of averaged magnitudes. Triassic rocks of

220-230 Ma ages have strain magnitude average between 1.0 and 1.5 while Jurassic rocks fluctuate around 1.0. Early Cretaceous rocks (145-100 Ma) show strain magnitude average no more than 1.0. Average strain magnitude of Late Cretaceous rocks (100-80 Ma) show a wider spread ranging from 0.6 to 1.7. The trend confirms that older rocks are on average more deformed although local exceptions occur for Late Cretaceous rocks. [Figure 7B](#) also shows a scattered pattern but the averages of Lode's ratio slightly increases from Triassic to Jurassic. Since most slate and phyllite are Jurassic, the Jurassic Lode's ratios are likely to be affected by the flattening during compaction and lithification.

4.5 Strain vs. paleo-depths

To investigate the relationship between strain and rock ages in the central Sierra Nevada, we plot the strain magnitude and Lode's ratio against the paleo-depths of rocks ([Figure 8](#)). The paleo-depths of host rocks are represented by the paleo-depth of plutons intruding the host rocks. Thus the paleo-depths of host rocks do not give full information about the change in a rock's vertical position through time: A rock may be displaced vertically prior or subsequent to the locking in at the emplacement depth. We report paleo-depths as pressure in units of kilobar (kbar). A conversion constant of 3.64 km/kbar can be used to transform pressure to depth.

Figure 8A shows that the average strain magnitude generally increases from 0.7 at 1 kbar to 1.2 at 3 kbars except the lower magnitude ($\epsilon \approx 1.0$) of rocks at 2.5 kbars. Strain magnitude average increases to 2.3 and 2.4 of the rocks at 4 kbars and 7 kbars, respectively. Figure 8B shows that the average Lode's ratio has the highest value at 1 kbar ($\gamma \approx 0.38$) and drops to $\gamma \approx 0.2$ for rocks at 1.5-3 kbars. The average Lode's ratio then sharply decreases to -0.3 and -0.65 at 4 kbars and 7 kbars, respectively. The overall pattern of Lode's ratio suggests rocks at 1-3 kbars have more plane to flattened shapes while the rocks at deeper paleo-depths typically become more constrictional. These data sets (Figure 8A and 8B) also suggest that there is a general negative correlation between strain magnitude and Lode's ratio as the paleo-depth increases. At shallower depth, strain magnitude averages are small ($\epsilon < 1.5$) and Lode's ratio averages are positive ($\gamma > 0$). Strain magnitude average increases to greater values and Lode's ratio average decreases to negative values at greater depth indicating that rocks at deeper paleo-depths are associated with higher strain magnitude and more constrictional strain shape.

4.6 Strain near pluton margins

One challenge in studying finite strain in continental arcs is to differentiate strain caused by regional tectonism from strain caused by pluton emplacement. A pluton potentially generates a localized structural aureole in which strain in the host rocks is higher than strain outside the aureole (e.g., Paterson and Vernon, 1995). In order to

constrain strain related to pluton emplacement, we compiled Lode's ratios, strain magnitudes, and Z-shortening in host rock against the distance from the nearest contact of the Cretaceous Tuolumne Intrusive Complex in the central Sierra Nevada (Figure 9). Locations of transects are shown in Figure 1C.

As shown in the plot of Lode's ratio vs. distance from pluton contact (Figure 9A), there is no clear correlation between strain shape and distance to the contact: Strain can be flattening ($\gamma > 0$), or constrictional ($\gamma < 0$), or close to plane strain ($\gamma \approx 0$) at any distance. Maximum strain magnitudes (Figure 9B) and Z-shortening (Figure 9C) suggest that higher strain in host rocks only locally occur within <200 m distance from the Tuolumne Intrusive Complex. Furthermore, pluton aureole strain can be evaluated by examining the deflection of host rock markers (e.g., bedding, contacts, and pre-emplacement dikes, faults and foliations) as pluton margins are approached. We have used this approach while mapping a number of pluton aureoles throughout the central Sierra and find that zones of increased strain near the margin are rare and only locally developed and large sections of the pluton margin are quite discordant to host rock bedding and foliations indicating little to no strain increase towards the pluton margin. Such observations suggests that even for a pluton of a fairly large size as the Tuolumne Intrusive Complex, higher host rock strain may be absent or quite localized within a few hundred meters from the contact.

4.7 Rate estimation of downward transport of host rocks

Downward transfer of host rocks is proposed by some researchers to solve the “space problem” during pluton emplacement (e.g. Saleeby, 1990; Tobisch et al., 2000; Paterson and Farris, 2008). The evidence supporting downward transport of host rocks includes the subvertical host rock stretching lineation and magmatic lineation in plutons (Saleeby et al., 2003; Paterson and Farris, 2008; Cao et al., 2015; this study), stoped blocks in plutons, and pluton rim and roof synclines and monoclines in adjacent host rocks (e.g. Paterson and Vernon, 1995; Paterson and Farris, 2008). Plutons intruding older volcanic rocks or sedimentary rocks (Figure 2D, Figure 2H) suggest that these supracrustal materials must be transported downwards to emplacement levels, which range from 2-3 kbars in the central Sierra Nevada to 9-10 kbars in the southern Sierra Nevada (Saleeby et al., 1990; Chapman et al, 2012).

With the geobarometry and geochronology data sets, we are able to estimate the minimum rate of downward transport of host rocks (mm/yr or km/Myr). The transported distance can be calculated using the emplacement pressure of pluton and the time interval is the age difference between host rocks and intruding pluton. Figure 10 shows the estimated minimum rates in different host rock pendants: most of them show rates of 2-13 mm/yr with faster rates found in the Cinko Lake, Jackass Lake and Eagle Peak areas, where upper crust plutons (~3 kbars) intrude slightly older (< 1 Ma) volcanic rocks suggesting a absolute rate of >10 mm/yr. The calculated absolute rates

are the minimum values since supracrustal materials can stay at surface until a pluton intruded later. This might be the case for some apparent slower rates found in Ritter Range, Green Lake and Saddlebag Lake areas.

5. Discussion

5.1 Possible first-order mechanisms contributing to finite bulk arc strain

Before we establish a comprehensive model of first-order mechanisms of strain, we will first examine two end member models and their resulting strain implications (Figure 11). Model A (Figure 11-A1) shows a crust only strained by homogenous, constant volume tectonic shortening and thickening with no magmatism. Above the brittle-ductile transition, crustal shortening occurs by brittle thrust faults, tilting of bedding, and duplex structures. Below the brittle-ductile transition, crustal shortening occurs by regional ductile strain and localized ductile shears, which also rotate bedding. Since the deformation is constant volume, plane strain, shortening will be balanced by crustal thickening. Development of LS-fabrics are predicted below the brittle-ductile transition under plane strain condition. (Sullivan, 2013). If some Y-direction length change or volume loss occurs, strain shape will deviate away from plane strain and become slightly flattening (Figure 11-A2). In a homogenous crust, strain shape and magnitude should be the same throughout the crustal column and plot as a horizontal

line in the strain shape vs. depth (Figure 11-A2) and strain magnitude vs. depth plots (Figure 11-A3).

Model B (Figure 11-B1) shows a crustal column being intruded by pulses of ascending magma but with no tectonic deformation. During magma ascent and trapping (e.g., emplacement) structural aureoles remain small or non-existent. Little to no evidence of regional lateral displacement of host rock units exists. This leaves the likelihood that complex processes within laterally growing plutonic bodies and narrow aureoles drive vertical displacements. Regional stress fields generated during emplacement of magmas may also drive host rock tilting and vertical displacements. As the arc matures the volume of plutonic rocks increases and a greater amount of magma are preferentially trapped at increasing depths (Saleeby, 1990; Paterson and Ducea, 2015).

The magmas begin life as low viscosity bodies (with dike, sheet, diapir, or irregular shapes) but over the short timescales of crystallization become high strength objects that are mechanically part of the host rock. Several processes deserve consideration. First there is a clear mass balance issue: as plutons emplace, an equal volume of host rock must be displaced to make the room. Structural and strain patterns require vertical motion either in or immediately next to the magma columns or in volumetrically decreasing “host rock channels” between solidified plutons (Figure 11-B1). In the Sierran arc greater than 80% of original host rock is now displaced requiring almost doubling of the crustal thickness. Second, strain within host rocks will increase in intensity and will become more constructional with depth as host rock is displaced

downwards through increasingly narrow channels (Figure 11-B2, B3). Host rocks are forced to “escape” vertically, through these “escape channels” either within magma columns or bounded by solidified plutonic bodies. The downward escape of host rock is potentially complicated by dragging forces, up or downwards within narrow, discontinuous aureoles. Similar shear forces along the magma-host rock contact had been proposed to contribute to host rock bedding rotation (Tobisch et al., 2000) and constrictional strain of host rocks in pluton aureoles (Paterson and Farris, 2008).

We propose that together Model A and B depict many but not all aspects of the deformation and observed strain patterns in the Sierra Nevada. The LS-fabrics (Figure 3) in host rock pendant in central Sierra Nevada as well as strain plots (Figure 4 and 5) suggest that plane strain dominates the deformation in shallower arc crust. Minor Y-extension and/or volume loss may shift the strain toward to a slightly more flattening strain regime. Such strain patterns can be explained by Model A. The increase of strain and more constrictional strain with depth can only be explained by the increasing number of plutons and magma with depth as predicted in Model B. Combining Model A and B, Model C (Figure 11-C1) shows a crust undergoing simultaneous tectonism and magmatism. In this model, combined tectonic and magmatic stresses result in complex horizontal shortening and vertical thickening that must accommodate both tectonic and magma addition mass balance requirements. Magmatic bodies may be internally strained and have their shapes influenced by tectonism. Solidified plutons will act as a “plutonic framework”. As the crust is tectonically shortened, the distances

between plutons may decrease resulting a more constrictional strain in host rocks with depth. We propose Model C could depict the observed strain patterns in both shallower and deeper parts of crust (Figure 11-C2, C3).

Figure 12 shows crustal column of Sierra Nevada arc with increasing strain magnitude and constrictional shapes with depth based on Model C. The limited increase in Mesozoic surface elevations and upper crustal exhumation (discussed in next section) requires major downwards transfer of host rocks to balance the crustal shortening and emplacement of plutons. The downward transported host rocks may spread laterally during growth of large orogenic roots resulting in the vertical constructional strains evolving to plane to flattening strains in the lower crust (Figure 12). And as descending host rocks enter the MASH zone (Hildreth and Moorbath, 1988), they may be disaggregated, partly assimilated and recycled into new ascending magma batches (e.g. Zellmer and Annen, 2008).

The goal of our models (Figure 11, 12) is to give a first-order approximation of processes that need evaluation as crust is strained during magmatism and tectonism. We recognize that more complicated processes occur during arc deformation including structures that broke through the Mesozoic Sierran arc, such as regional strike-slip faults, ductile shear zones, and low-angle normal faults (e.g. Lahren and Schweickert, 1989; Kistler, 1990; Saleeby and Busby, 1993; Nadin and Saleeby, 2007; Chapman et al., 2015; Saleeby and Dunne, 2015). Since most strike slip faults trend ~parallel to the

arc, they may shuffle units parallel to the arc but not significantly affect the overall shortening and thickening. Dip slip faults/shears are less common but at times may play a greater role in arc deformation and strain that deserves greater attention in future studies. These structures will also aid in the overall arc shortening/thickening and tilt of units discussed in this paper and may add localized concentrated increments of strain to the arc crust. When possible such strain increments are included in the finite bulk arc strain. The model shown in [Figure 12](#) of course is not a perfect explanation for arc strain. We do not fully understand some observed strain patterns. For example, the spread of the average strain magnitude and Lode's ratio in Late Cretaceous rocks (mainly volcanic rocks of 100-80 Ma in age) ([Figure 7A and 7B](#)). We suspect such temporally restricted strain variations might be a consequence of volumetrically large addition of magma in the Late Cretaceous leading to a rheologically very heterogeneous crust.

5.2 The finite bulk arc strain, mass balance and crustal thickening

Above we concluded that the ductile strain in the WMB and Sierran host rock pendants resulted in about 50% arc-perpendicular shortening in a more or less plane strain style. In the field, at the paleo-depths of 5-10 km, different packages of ductilely deformed rocks are commonly separated by Mesozoic NE- or SW-vergent thrust faults ([Tobisch et al., 2000](#); [Schweickert and Lahren, 2006](#); [Cao et al., 2015](#)), which have been subsequently rotated with bedding to a sub-vertical orientation. To estimate the bulk

deformation in host rocks, other deformation mechanisms need to be evaluated as well.

[Tobisch et al. \(2000\)](#) proposed that thrust-related duplex structures can rotate the bedding to 45° at maximum, which correspond to about 30% shortening along horizontal direction ([Cao et al., 2015](#)). Since both ductile and brittle strain suggest the same NE-SW shortening kinematics, we combine 50% ductile shortening with 30% shortening caused by thrust-related duplex, for a finite shortening strain is about 65% shortening ($1-[1-0.5[1-0.3]=0.65$).

To estimate the finite bulk arc strain and crustal thickening, in addition to the host rocks, one has to take the plutons into account. Sierran plutons record multiple strain events caused by both internal magma flow and external tectonic shortening ([Paterson et al., 1998](#); [Žák et al., 2007](#)) The tectonic strain in plutons are represented by the ~NW-striking magmatic-subsolidus foliation and subvertical lineation (e.g. [Žák et al., 2007](#); [Cao et al., 2015](#)). Strain in plutons is difficult to retrieve quantitatively due to the lack of strain markers. In a few cases, strain can be estimated using the folded veins/dikes within the plutons. Our estimates of averaged tectonic strains in plutons range from 0~30% shortening with a common average around 10% shortening (e.g., [Paterson et al., 2014](#)). Others have suggested that pluton shapes may reflect tectonic strains (e.g. [Weinberg et al., 2004](#)). Some plutons in central Sierra Nevada have more or less circular shapes whereas others have length/width ratios of up to 3/1. However neither show internal tectonic fabrics that are more intense than discussed above. Nor do they show evidence at their NW or SE “ends” of expansion in either the plutons or

host rocks. Thus we suspect that plutons haven't been significantly strained by regional shortening. Even so for our first-order calculations below, we assume the plutons have been totally shortened ~20% of which the fabrics record about half.

In an active continental arc, the thickening of crust is caused by (1) tectonic thickening compensating the tectonic shortening and (2) magmatic intrusion (e.g. Allmendinger et al., 1997; Karlstrom et al., 2014; Lee et al., 2015). To calculate the final crustal thickness, a Matlab script (see Supporting Materials) is written to explore the simultaneous incremental addition of magmatism and incremental regional strain to a crust. For the Sierra Nevada, we assume an initial crust thickness of ~25 km when the arc started (~250 Ma) and that the pluton fraction in the crust is about 0.8 by ~85 Ma. The calculation also assumes 20% shortening in plutons and 20% volume loss/Y-direction extension of host rocks into account. Figure 13 shows the final crustal thickness in a parameter space of host rock strain (ϵ_H) and pluton fraction (β) for a given initial crustal thickness of 25 km. When $\epsilon_H = -0.65$ and $\beta = 0.8$, the final crustal thickness is ~97 km. Next, we consider how erosion and exhumation modifies the crustal thickness.

In the central Sierra Nevada, Al-in-Hornblende barometer on Mesozoic plutons establishes that Triassic plutons in the eastern Sierra emplaced at ~3.2 kbars are juxtaposed by Late Cretaceous (~86 Ma) plutons at ~2 kbars, suggesting a Mesozoic exhumation < 5 km (Cao et al., 2015). In the WMB, Triassic volcanics and Jurassic

marine sediments are exposed at the surface and show lower to upper greenschist grades of metamorphism. In addition, Late Jurassic plutons in the WMB (e.g., the Guadalupe Igneous Complex) intruded these sedimentary rocks at shallow levels corresponding to 1-2 kbars (Haeussler and Paterson, 1993; Putirka et al., 2014) and are intruded by Late Cretaceous plutons with slightly higher Al-in-hornblende pressures of ~3 kbars (Ague and Brimhall, 1988a; Haeussler and Paterson, 1993), which suggests that large exhumation did not occur in the WMB early in the Mesozoic and if anything depths increased in the Cretaceous, the latter consistent with Cretaceous crustal thickening.

In the southern Sierra, Early Triassic plutons emplaced at 10-15 km depth are intruded by Middle Jurassic and Late Cretaceous plutons emplaced at 5-6 km depth (Nadin and Saleeby, 2008; Saleeby and Dunne, 2015), suggesting that mid-crustal exhumation from Early Triassic to Late Cretaceous in southern Sierra is ~10 km. Exhumation in the southern Sierra along the Kaweah River based on barometry in plutons indicates an average 0.5 mm/yr exhumation between ~100 Ma and 85 Ma although the western and eastern parts of the southern Sierras have different exhumation rates (Saleeby and Sousa, personal communication).

In summary, by ~85 Ma, the exhumation of southern Sierra is larger (~10 km) than the exhumation in the central Sierra (< 5 km). However, the exhumation based on barometry only gives a minimum value of rocks eroded (Figure 14). The correction of

exhumation with respect of crustal thickening should be made. In this paper, for a first-order approximation, we use a more generous range of 10-20 km for the amount of removed crust during Mesozoic thickening of the Sierra Nevada. Since we earlier estimated that crustal thickness is ~97 km at 85 Ma, surface erosion will reduce the Late Cretaceous crustal thickness to ~77-87 km.

5.3 Implications on downward transfer of host rocks, paleo-elevation and arc behaviors

As we discussed in the previous sections, downward transfer of host rocks satisfy the strain patterns observed and helps both the tectonically driven thickening and root development and the “space problem” during magma ascent and emplacement (e.g. Saleeby et al., 1990; Paterson and Farris, 2008; Anderson, et al., 2010; Chapman et al., 2014). The downward transfer of host rocks is required to make the mass balance in the arc crust: about ~50-km-thick materials have to be transported downwards to accommodate the thickened crust if the initial crust thickness is 25 km. The minimum rates of such processes are shown in Figure 10. In a 10 Myr interval, 20-130 km of downward transport of host rock material is feasible given these rates. Although this is a rough estimate, it shows that materials can be displaced downward by tens-of-kilometers during the duration of a magmatic flare event, making enough space for emplacement of voluminous magma.

636

637 Downward transport of host rocks can potentially play a role in regulating Sierran
638 magmatism and arc tempos: It transports supracrustal materials downward, fertilizing
639 the MASH in the lower crust and upper mantle in addition to crustal inputs from the
640 fore-arc, retro-arc and downgoing plate (e.g. [Kay et al., 2005](#); [Ducea, 2001](#); [DeCelles et](#)
641 [al., 2009](#); [Hacker et al., 2011](#); [Chapman et al., 2013](#); [Chapman et al., 2014](#)). $\delta^{18}\text{O}$ and
642 ϵNd results suggest that crustal sources have contributed up to 50% of Sierran
643 magmatic sources ([Ducea and Barton, 2007](#)). The isotopic data cannot constrain the
644 exact source of supracrustal materials, but leaves open the possibility that intra-arc
645 downward transfer of host rocks is one likely source. Crustal thickening not only
646 introduces greater crustal components to the arc magma source regime, but it also
647 places greater vertical intervals of the source regime in both the plagioclase out and
648 garnet in stability fields favoring the accumulation of an eclogitic residue along the
649 base of the crust ([Saleeby et al., 2003](#)). The foundering of such high-density arc roots
650 into asthenospheric mantle may further introduce supracrustal materials (e.g. crustal
651 isotopic signatures) into the mantle and contribute to the geochemical heterogeneity of
652 the mantle.

653

654 The estimated 77 to 87 km-thick crust at ~85 Ma will build a high elevation in the Sierra
655 Nevada as a result of isostasy. Elevation (h) is a function of crustal thickness (H) and
656 arc root thickness (R) if crust is isostatically balanced (Eq. 3, modified from [Lee et al.,](#)
657 [2015](#)):

658

659
$$h = h_0 + \left(1 - \frac{\rho_c}{\rho_m}\right) \cdot (H - H_0) + \left(1 - \frac{\rho_r}{\rho_m}\right) \cdot R \quad (3)$$

660

661 h_0 is the initial elevation and H_0 is the initial crustal thickness. ρ_c , ρ_m , and ρ_r are

662 the densities of crust, asthenospheric mantle and arc root, respectively. Note that

663 $\left(1 - \frac{\rho_c}{\rho_m}\right) > 0$ while $\left(1 - \frac{\rho_r}{\rho_m}\right) < 0$, which means that thickened crust will positively

664 contribute to the elevation while the arc root will reduce the elevation. [Figure 15](#) shows

665 the how the elevation changes with crustal thicknesss (H) and arc root thickness (R)

666 assuming $h_0 = 0$ and $H_0 = 25$ km. If the arc root is 30 km thick, the elevation of Sierra

667 Nevada at ~85 Ma will be 4.5~5.7 km, close to the Puna-Altiplano plateau of central

668 Andes (4-5 km, [Allmendinger et al., 1997](#)).

669

670 The thickened crust also influences the asthenospheric mantle wedge and magma

671 production. [Figure 16](#) shows a subduction zone before and after magmatism and

672 deformation. Before magmatism and deformation started ([Figure 16A](#)), the crust is 25

673 km thick and mantle flow is not impeded. After magmatism and deformation started

674 ([Figure 16B](#)), the crust is thickened to ~80 km with a 30-km-thick arc root. Mantle flow

675 will be impeded and as a result, arc magmatism will potentially migrate inboard

676 ([Karlstrom et al., 2014](#)). [Chin et al., \(2012, 2015\)](#) also suggested that such thickened

677 crust also cools down the mantle wedge causing the reduction of the production of new

678 magmas. The impeded mantle flow, lithospheric mantle cooling effect, and the slab

flattening that started ~90-85 Ma ([Saleeby, 2003](#)) in Sierra Nevada may eventually cool and thus reduce magma generation in the Sierra Nevada area and cause the eastward migration of magmatism.

6. Conclusions

A compiled dataset of >650 finite strain analyses from central and southern Sierra Nevada, California are used to help quantify the arc crust deformation in Sierra Nevada. Our results show that the overall Mesozoic tectonism results in 65% arc-perpendicular bulk crust shortening under a more or less plane strain condition with ductile strain recording about 50% shortening. The tectonism leads to at least >100% crustal thickening. Since Mesozoic magmatism replaced ~80% of the crust, this mass balance exchange drove additional thickening. We estimated that the arc crust has been significantly thickened to ~80 km by ~85 Ma due to the combined effects of tectonism and magmatism. A 30-km-thick arc root was generated and surface elevation of ~5 km at this same time. Using constraints on exhumation, most crustal materials must be transported downwards to accommodate the crustal thickening at estimated rates of 2-13 km/Myr. The downward transfer of crustal materials must occur in locations of active magma channels, or in “escape channels” in between solidified plutons that decrease in size with time and depth resulting in an increase in the intensity of constrictional strain with depth. The crustal thickness and surface elevation are thus controlled by both tectonism and magmatism but are slightly modify the surface

erosion. On the one hand, the downward transported crustal materials may fertilize the MASH zone thus enhancing the generation of additional magmas. However, as the crustal root grows it may potentially pinch out and cool the mantle wedge and thus cause reduction of arc magmatism.

Acknowledgements

Paterson acknowledges support from NSF grants EAR-1019636, EAR-0537892 and EAR-0073943 and U.S. Geological Survey EDMAP grants G13AC00120 (supporting Sean Hartman) and 03HQA0038 (supporting Vali Memeti) for work in and around the TIC. Wenrong Cao acknowledges support from EDMAP Grant G12AC20178 and Geological Society of America Graduate Student Research Grants. We thank USC undergraduate student Emily Van Guilder for the initial strain data compilation. Snir Attia, Sean Hartman, Katie Ardill, and Babsi Ratschbacher, are gratefully thanked for their discussions and comments on the manuscript. Cin-Ty Lee, Juling Zhang and Xun Yu are thanked for the discussion on isostasy and crustal thickening. Alan Chapman and an anonymous reviewer are thanked for their critical comments, which greatly help to improve the quality of the manuscript. We also thank Editor Toru Takeshita for handling the submission.

References

Ague, J.J., 1997. Thermodynamic calculation of emplacement pressures for batholithic rocks, California: Implications for the aluminum-in-hornblende barometer. *Geology* 25, 563–566.

723 Ague, J. J., and Brandon, M. T., 1992. Tilt and northward offset of Cordilleran
 724 batholiths resolved using igneous barometry. *Nature* 360, 146–149.

725 Ague, J. J., and Brimhall, G. H., 1988a. Magmatic arc asymmetry and distribution of
 726 anomalous plutonic belts in the batholiths of California: Effects of assimilation,
 727 crustal thickness, and depth of crystallization. *Geological Society of*
 728 *America Bulletin* 100, 912–927.

729 Ague, J. J., and Brimhall, G. H., 1988b. Regional variations in bulk chemistry,
 730 mineralogy, and the compositions of mafic and accessory minerals in the batholiths
 731 of California. *Geological Society of America Bulletin* 100, 891–911.

732 Albertz, M., 2006. Relationships between melt-induced rheological transitions and
 733 finite strain: Observations from host rock pendants of the Tuolumne Intrusive Suite,
 734 Sierra Nevada, California. *Journal of Structural Geology* 28, 1422–1444.

735 Allmendinger, R.W., Cardozo, N.C., and Fisher, D., 2012. *Structural Geology*
 736 *Algorithms: Vectors & Tensors*. Cambridge University Press, Cambridge, UK.

737 Allmendinger, R. W., Jordan, T. E., Kay, S. M., & Isacks, B. L. 1997. The evolution of
 738 the Altiplano-Puna plateau of the Central Andes. *Annual Review of Earth and*
 739 *Planetary Sciences* 25, 139–174.

740 Anderson, J.L., Barth, A.P., Wooden, J.L., and Mazdab, F., 2008. Thermometers and
 741 thermobarometers in Granitic Systems. *Reviews of Mineralogy and Geochemistry*
 742 69, 121–142.

743 Anderson, J.L., Paterson, S. R., Memeti, V., Zhang, Tao, Economos, R., Barth, A.,
 744 Pignotta, G., Mundil, R., Foley, B., Schmidt, K., 2010. Episodic downward crustal
 745 flow during Triassic to Cretaceous magma surges in the central Sierra arc.
 746 *Geological Society of America Abstracts with Programs* 42, 51.

747 Anderson, J.L., and Smith, D.R., 1995. The effects of temperature and f_{O_2} on the
 748 Al-in-hornblende barometer. *American Mineralogist* 80, 549–559.

749 Barth, A.P., Feilen, A.D.G., Yager, S.L., Douglas, S.R., Wooden, J.L., Riggs, N.R. and
 750 Walker, J.D., 2012. Petrogenetic connections between ash-flow tuffs and a
 751 grandioritic to granitic intrusive suite in the Sierra Nevada arc, California.
 752 *Geosphere* 8, 250–264.

753 Barton, M. D., and Hanson, R. B., 1989. Magmatism and the development of
 754 low-pressure metamorphic belts: Implications from the western United States and
 755 thermal modeling. *Geological Society of America Bulletin* 101, 1051–1065.

756 Bateman, P.C., and Chappell, B.W., 1979. Crystallization, formation, and solidification
 757 of the Tuolumne intrusive series, Yosemite National Park, California. Geological
 758 Society of America Bulletin 99, 465-482.

759 Brook, C.A., 1977. Stratigraphy and structure of the Saddlebag Lake roof pendant,
 760 Sierra Nevada, California. Geological Society of America Bulletin 88, 321-334.

761 Busby-Spera, C.J, and Saleeby, J.B., 1990. Intra-arc strike-slip fault exposed at
 762 batholithic levels in southern Sierra Nevada, California. *Geology* 18, 255-259.

763 Cao, W., Paterson, S.R., Memeti, V., Mundil, R., Anderson, L. and Schmidt, K. 2015.
 764 Tracking paleo-strain field in a continental arc: a study of incremental and finite
 765 strain in Mesozoic plutons and host rocks, respectively in central Sierra Nevada,
 766 California. *Lithosphere* 7, 296-320.

767 Cardozo, N., and Allmendinger, R.W., 2013. Spherical projections with
 768 OSXStereonet. *Computers & Geosciences* 51, 193-205.

769 Chapman, A. D., Ernst, W. G., Gottlieb, E., Powerman, V., and Metzger, E. P., 2015.
 770 Detrital zircon geochronology of Neoproterozoic-Lower Cambrian passive-margin
 771 strata of the White-Inyo Range, east-central California: Implications for the
 772 Mojave-Snow Lake fault hypothesis. *Geological Society of America Bulletin*,
 773 B31142-1, doi: 10.1130/B31142.1.

774 Chapman, A.D., Saleeby, J.B., and Eiler, J., 2013. Slab flattening trigger for isotopic
 775 disturbance and magmatic flare-up in the southernmost Sierra Nevada batholith,
 776 California: *Geology* 41, 1007-1010, doi:10.1130/G34445.1.

777 Chapman, A. D., Saleeby, J. B., Wood, D. J., Piasecki, A., Kidder, S., Ducea, M. N., and
 778 Farley, K. A., 2012. Late Cretaceous gravitational collapse of the southern Sierra
 779 Nevada batholith, California. *Geosphere* 8, 314-341.

780 Chapman, A.D., Ducea, M.N., Kidder, S., and Petrescu, L., 2014. Geochemical
 781 constraints on the petrogenesis of the Salinian arc, central California: Implications
 782 for the origin of intermediate magmas. *Lithos* 200-201, 126-141,
 783 doi:10.1016/j.lithos.2014.04.011.

784 Chin, E.J., Lee, C.-T. A., Luffi, P., and Tice, M., 2012. Deep lithospheric thickening and
 785 refertilization beneath continental arcs: Case study of the P,T and compositional
 786 evolution of peridotite xenoliths from the Sierra Nevada, California. *Journal of*
 787 *Petrology* 53, 477-511, doi:10.1093/petrology/egr069.

788 Chin, E.J., Lee, C.-T. A., and Blichert-Toft, J., 2015. Growth of upper plate lithosphere
789 controls tempo of arc magmatism: Constraints from Al-diffusion kinetics and
790 coupled Lu-Hf and Sm-Nd chronology. *Geochem. Persp. Let.* 1, 20-32, doi:
791 10.7185/geochemlet.1503

792 Clark, L. D. 1964. Stratigraphy and structure of part of the western Sierra Nevada
793 metamorphic belt, California. U. S. Geological Survey Special Paper 410. U. S.
794 Govt. Print. Off.

795 Crespi, J. M., 1986. Some guidelines for the practical application of Fry's method of
796 strain analysis. *Journal of Structural Geology* 24, 799-808.

797 DeCelles, P.G., Ducea, M.N., Kapp, P., and Zandt, G., 2009. Cyclicality in Cordilleran
798 orogenic systems. *Nature Geoscience* 2, 251-157.

799 Dickinson, W. R. 2008. Accretionary Mesozoic–Cenozoic expansion of the Cordilleran
800 continental margin in California and adjacent Oregon. *Geosphere* 4, 329-353.

801 Dixon, E.T., 1995, An Evaluation of Hornblende Barometry, Isabella to Tehachapi
802 Region, Southern Sierra Nevada, California [M.S. thesis]: Ann Arbor, University
803 of Michigan, 150 p.

804 Ducea, M., 2001. The California Arc: Thick granitic batholiths, eclogitic residues,
805 lithospheric-scale thrusting, and magmatic flare-ups. *GSA Today* 11, 4-10.

806 Ducea, M.N., and Barton, M.D., 2007. Igniting flare-up events in Cordilleran arcs.
807 *Geology* 35, 1047–1050, doi: 10.1130/G23898A.1.

808 Edelman, S. H., and Sharp, W. D. 1989. Terranes, early faults, and pre-Late Jurassic
809 amalgamation of the western Sierra Nevada metamorphic belt,
810 California. *Geological Society of America Bulletin* 101, 420-1433.

811 Erslev, E. A. 1988. Normalized center-to-center strain analysis of packed
812 aggregates. *Journal of Structural Geology* 10, 201-209.

813 Greene, D.C., and Schweickert, R.A., 1995. The Gem Lake shear zone: Cretaceous
814 dextral transpression in the northern Ritter Range pendant, eastern Sierra Nevada,
815 California. *Tectonics* 14, 945–961, doi:10.1029/95TC01509.

816 Hacker, B.R., Kelemen, P.B., and Behn, M.D., 201. Differentiation of the continental
817 crust by relamination. *Earth and Planetary Science Letters* 307, 501–516,
818 doi:10.1016/j.epsl.2011.05.024.

819 Haeussler, P. J. and Paterson, S.R., 1993, Post-emplacement tilting and burial of the
820 Guadalupe Igneous Complex, Sierra Nevada, California. *Geological Society of*

821 America Bulletin 105, 1310-1320.

822 Hanson, R.B., Sorensen, S.S., Barton, M.D., and Fiske, R.S., 1993. Long-term
823 evolution of fluid–rock interactions in magmatic arcs: evidence from the Ritter
824 Range pendant, Sierra Nevada, California, and numerical modeling. *Journal of*
825 *Petrology* 34, 23-62.

826 Holland, T., and Blundy, J., 1994. Non-ideal interactions in calcic amphiboles and their
827 bearing on amphibole-plagioclase thermometry. *Contributions to Mineralogy and*
828 *Petrology* 116, 433-447.

829 Hildreth, W., and Moorbath, S. 1988. Crustal contributions to arc magmatism in the
830 Andes of central Chile. *Contributions to mineralogy and petrology* 98, 455-489.

831 Horsman, E., Tikoff, B., and Czeck, D., 2008. Rheological implications of
832 heterogeneous deformation at multiple scale in the Late Cretaceous Sierra Nevada,
833 California. *Geological Society of America Bulletin* 120, 238-255.

834 Hossack, J.R., 1967. Pebble deformation and thrusting in the Bygdin Area (southern
835 Norway). *Tectonophysics* 5, 315-339.

836 Hsu, T.C., 1966. The characteristics of coaxial and non-coaxial strain paths. *Journal of*
837 *strain analysis for engineering design* 1, 216-222.

838 Johnson, M. C., and Rutherford, M. J., 1989. Experimental calibration of an
839 Aluminum-in-hornblende geobarometer with application to Long Valley caldera
840 (California) volcanic rocks. *Geology* 17, 837-841.

841 Kay, S.M., Gody, E., and Kurz, A., 2005. Episodic arc migration, crustal thickening,
842 subduction erosion and magmatism in the south-central Andes: *Geological Society*
843 *of America Bulletin* 117, 67–88, doi:10.1130/B25431.1.

844 Karlstrom, L., Lee, C.-T.A., and Manga, M., 2014. The role of magmatically driven
845 lithospheric thickening on arc front migration: *Geochemistry Geophysics*
846 *Geosystems* 15, doi:10.1002/2014GC005355.

847 Kistler, R. W. 1990. Two different lithosphere types in the Sierra Nevada, California.
848 *Geological Society of America Memoirs* 174, 271-282.

849 Lahren, M.M., and Schweickert, R.A., 1989. Proterozoic and Lower Cambrian passive
850 margin rocks of Snow Lake pendant, Yosemite-Emigrant Wilderness, Sierra
851 Nevada, California: Evidence for major Early Cretaceous dextral translation.
852 *Geology* 17, 156–160.

853 Lahren, M.M., and Schweickert, R.A., 1995. Continuation of the Snow Lake block into

854 western Nevada and speculations on the geometry and tectonics of the
855 Mojave-Snow Lake fault. Geological Society of America Abstracts with Programs
856 27, 58.

857 Lee, C. T. A., Thurner, S., Paterson, S.R., and Cao, W. 2015. The rise and fall of
858 continental arcs: Interplays between magmatism, uplift, weathering, and climate.
859 Earth and Planetary Science Letters 425, 105-119.

860 Lisle, R. J., 1977. Estimation of the tectonic strain ratio from the mean shape of
861 deformed elliptical markers. *Geologie en Mijnbouw* 56, 140-144.

862 Lisle, R. J., 1979. Strain analysis using deformed pebbles: The influence of initial
863 pebble shape. *Tectonophysics* 60, 263-277.

864 Lisle, R. J., 1985. Geological strain analysis: a manual for the R_f/θ technique.
865 Pergamon Press, Oxford.

866 Lode, W., 1926. Versuche über den Einfluss der mittleren Hauptspannung auf das
867 Fließen des Metalle Eisen, Kupfer, und Nickel. *Zeitschrift für Phys.* 36, 913-939.

868 Longiaru, S., 1987. Tectonic evolution of Oak Creek volcanic roof pendant, eastern
869 Sierra Nevada, California. Ph.D. thesis, University of California Santa Cruz.

870 March, A., 1932. Mathematische Theorie der Regelung nach der Korngestalt. *Z.*
871 *Kristallogr* 81, 285-297.

872 Memeti, V., Gehrels, G.E., Paterson, S.R., Thompson, J.M., and Mueller, R.M., 2010.
873 Re-evaluating the Mojave-Snow Lake fault hypothesis and origins of central
874 Sierran metasedimentary pendant strata using detrital zircon provenance analyses.
875 *Lithosphere* 2, 341-360.

876 Miller, D. M., and Oertel, G., 1979. Strain determination from the measurement of
877 pebble shapes: a modification. *Tectonophysics* 55, T11-T13.

878 Miller, R.B., and Paterson, S.R., 1991. Geology and tectonic evolution of the Bear
879 Mountains fault zone, Foothills terrane, central Sierra Nevada, California.
880 *Tectonics* 10, 995-1006.

881 Mundil, R., Nomade, S., Paterson, S., and Renne, P.R., 2004. Geochronological
882 constraints ($^{40}\text{Ar}/^{39}\text{Ar}$ and U/Pb) on the thermal history of the Tuolumne Intrusive
883 Suite (Sierra Nevada, California). AGU Fall Meeting Abstract V53A-0616.

884 Nadin, E. S., and Saleeby, J. B., 2008. Disruption of regional primary structure of the
885 Sierra Nevada batholith by the Kern Canyon fault system, California. Geological
886 Society of America Special Paper 438, 1-26.

887 Oertel, G., 1983. The relationship of strain and preferred orientation of phyllosilicate
888 grains in rocks - a review. *Tectonophysics* 100, 413-447.

889 Paterson, S.R., 1983. A comparison of methods used in measuring finite strains from
890 ellipsoidal objects. *Journal of Structural Geology* 5, 611-618.

891 Paterson, S.R., 1989. A reinterpretation of conjugate folds in the central Sierra Nevada,
892 California. *Geological Society of America Bulletin* 101, 248-259.

893 Paterson, S.R., and Ducea, 2015. Arc magmatic tempos: Gathering the evidence.
894 *Elements* 11, 91-98.

895 Paterson, S. R., and Farris, D. W., 2008. Downward host rock transport and the
896 formation of rim monoclines during the emplacement of Cordilleran batholiths.
897 *Transactions of the Royal Society of Edinburgh: Earth Sciences* 97, 397-413.

898 Paterson, S. R., Fowler, T. K., Schmidt, K. L., Yoshinobu, A. S., Yuan, E. S., and Miller,
899 R. B., 1998. Interpreting magmatic fabric patterns in plutons. *Lithos* 44, 53-82.

900 Paterson, S. R., Tobisch, O. T., Bhattacharyya, T., 1989, Regional structural and strain
901 analyses of terranes in the Western Metamorphic belt, central Sierra Nevada,
902 California. *Journal of Structural Geology*, v. 11, p. 255-273.

903 Paterson, S.R., Memeti, V., Anderson, J.L., Cao, Wenrong, Lackey, J.S., Putirka, K.D.,
904 Miller, R.B., Miller, J.S., Mundil, R. 2014. Overview of arc processes and tempos.
905 In Memeti, V. et al., (Eds.), *Formation of the Sierra Nevada batholith: Magmatic
906 and tectonic processes and their tempos*. Geological Society of America Field
907 Guide 34, 87-104.

908 Paterson, S.R., Vernon, R.H., 1995. Bursting the bubble of ballooning plutons: a return
909 to nested diapirs emplaced by multiple processes. *Geological Society of America
910 Bulletin* 107, 1356–1380.

911 Paterson, S.R., Yu, H. 1993. Microstructures and primary grain-shape fabrics in
912 sandstones: implications for depositional processes and strain analyses. *Journal of
913 Structural Geology* 16, 505-517.

914 Paterson, S.R., Yu, H., and Oertel, G., 1995. Primary and tectonic fabric intensities in
915 mudrocks. *Tectonophysics* 247, 105-119.

916 Pfiffner, O.A. and Ramsay, J.G., 1982. Constraints on geological strain rates:
917 Arguments from finite strain states of naturally deformed rocks. *Journal of
918 Geophysical Research* 87, 311-321.

919 Pickett, D. A., and Saleeby, J. B. 1993. Thermobarometric constraints on the depth of

920 exposure and conditions of plutonism and metamorphism at deep levels of the
 921 Sierra Nevada Batholith, Tehachapi Mountains, California. *Journal of Geophysical*
 922 *Research: Solid Earth* (1978–2012) 98, 609-629.

923 Putirka, K., Canchola, J., Rash, J., Smith, O., Torrez, G., Paterson, S., and Ducea, M.,
 924 2014, Pluton assembly and the genesis of granitic magmas: insights from the GIC
 925 pluton in cross section, Sierra Nevada Batholith, California, *American*
 926 *Mineralogist* 99, 1284-1303.

927 Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.

928 Ramsay, J.G., and Wood, D., 1973. The geometric effects of volume changes during
 929 deformation processes. *Tectonophysics* 16, 263-277.

930 Saleeby, J. B., 1990. Progress in tectonic and petrogenetic studies in an exposed
 931 cross-section of young (~100 Ma) continental crust, southern Sierra Nevada,
 932 California. In Salisbury, M. H., and Fountain, D. M., (Eds.), *Exposed Cross-*
 933 *Sections of the Continental Crust* 137-158.

934 Saleeby, J.B., 2003. Segmentation of the Laramide Slab - Evidence from the southern
 935 Sierra Nevada region: *Geological Society of America Bulletin* 115, 655–668.

936 Saleeby, J., 2011. Geochemical mapping of the Kings-Kaweah ophiolite belt,
 937 southwestern Sierra Nevada Foothills-Evidence for progressive melange
 938 formation in a large offset transform-subduction initiation environment: in J.
 939 Wakabayashi and E. Dilek, (Eds.), *Geological Society of America Special Paper*
 940 *on the The Nature and Formation of Melanges*, Geological Society of America
 941 *Special Paper* 480, 31-73, doi:10.1130/2011.2480(02).

942 Saleeby, J.B., and Busby, C., 1993. Paleogeographic and tectonic setting of axial and
 943 western metamorphic framework rocks of the southern Sierra Nevada, California,
 944 *in* Dunne, G., and McDougall, K., (Eds.), *Mesozoic Paleogeography of the*
 945 *Western United States-II: Pacific Section*, Society of Economic Paleontologists
 946 and Mineralogists, Book 71, 197-226.

947 Saleeby, J. B., Ducea, M. N., Busby, C. J., Nadin, E. S., and Wetmore, P. H., 2008.
 948 Chronology of pluton emplacement and regional deformation in the southern
 949 Sierra Nevada batholith, California. *Geological Society of America Special Paper*
 950 438, 397-427.

951 Saleeby, J. B., Ducea, M.N., and Clemens, K.D., 2003. Production and loss of
 952 high-density batholithic root, southern Sierra Nevada, California. *Tectonics* 22,
 953 doi:10.1029/2002TC001374, 2003.

954 Saleeby, J.B, and Dunne, G., 2015. Temporal and tectonic rela- tions of early Mesozoic
 955 arc magmatism, southern Sierra Nevada, California. In Anderson, T. H., Didenko,
 956 A.N., John- son, C.L., Khanchuk, A.I., and MacDonald Jr., J.H., (Eds.), *Late*
 957 *Jurassic Margin of the Laurasia—A Record of Faulting Accommodating Plate*
 958 *Rotation*. Geological Society of America Special Paper 513,
 959 doi:10.1130/2015.2513(05).

960 Schmidt, M.W., 1992. Amphibole composition in tonalite as a function of pressure: an
 961 experimental calibration of the Al-in-hornblende barometer. *Contributions to*
 962 *Mineralogy and Petrology* 110, 304-310.

963 Schweickert, R. A., Hanson, R. E., and Girty, G. H., 1999. Accretionary tectonics of the
 964 western Sierra Nevada metamorphic belt. In Wagner, D. L., and Graham, S. A.,
 965 (Eds.). *Geologic field trips in northern California: Sacramento, CA, California*
 966 *Department of Conservation, Division of Mines and Geology, Centennial Meeting*
 967 *of the Cordilleran Section of the Geological Society of America, Special*
 968 *Publication* 119, 33–79.

969 Schweickert, R.A., and Lahren, M.M., 1993. Tectonics of the east-central Sierra
 970 Nevada-Saddlebag Lake and northern Ritter Range pendants. In Lahren, M.M.,
 971 Trexler, J.H., and Spinoza, C., (Eds.), *Crustal evolution of the Great Basin and the*
 972 *Sierra Nevada: Geological Society of America May 1993 Field Trip Guidebook*,
 973 *Department of Geological Sciences, University of Nevada, Reno*, 313-352.

974 Schweickert, R.A., and Lahren, M.M., 2006. Geologic evolution of Saddlebag Lake
 975 pendant, eastern Sierra Nevada, California: Tectonic implications. In Girty, G., and
 976 Cooper, J.D., (Eds.), *Using stratigraphy, sedimentology, and geochemistry to*
 977 *unravel the geologic history of the southwestern Cordillera: Pacific Section, SEPM*
 978 *(Society for Sedimentary Geology), Fullerton*, 27–56.

979 Schweickert, R.A., Saleeby, J.B., Tobisch. O.T., 1977. Paleotectonic and
 980 paleogeographic significance of the Calaveras complex, western Sierra Nevada,
 981 California. In Stewart, J.H., and Stevens, C.H., (Eds.), *Paleozoic Paleogeography*
 982 *of the Western United States: Pacific Section, Society of Economic Paleontologists*
 983 *and Mineralogists* 381-394.

984 Sharp, W.D., Tobisch, O.T., and Renne, P.R., 2000. Development of Cretaceous
985 transpressional cleavage synchronous with batholith emplacement, central Sierra
986 Nevada, California. *Geological Society of America Bulletin* 112, 1059–1066,
987 doi:10.1130/0016-7606(2000)112<1059:DOCTCS>2.0.CO;2.

988 Shimamoto, T., and Ikeda, Y., 1976. A simple algebraic method for strain estimation
989 from deformed ellipsoidal objects. 1, Basic theory. *Tectonophysics* 36, 315–337.

990 Snow, C. A., and Scherer, H., 2006. Terranes of the Western Sierra Nevada Foothills
991 Metamorphic Belt, California: A Critical Review. *International Geology Review* 48,
992 46–62, doi:10.2747/0020-6814.48.1.46.

993 Sullivan, W. A. 2013. L tectonites. *Journal of Structural Geology*, 50, 161–175.

994 Tikoff, B., and Saint-Blanquat, M., 1997. Transpressional shearing and strike-slip
995 partitioning in the Late Cretaceous Sierra Nevada magmatic arc, California.
996 *Tectonics* 16, 442–459, doi:10.1029/97TC00720.

997 Tobisch, O. T., Fiske, R. S., Sacks, S., and Taniguchi, D., 1977. Strain in
998 metamorphosed volcanoclastic rocks and its bearing on the evolution of orogenic
999 belts. *Geological Society of America Bulletin* 88, 23–40.

1000 Tobisch, O.T., and Fiske, R.S., 1982. Repeated parallel deformation in part of the
1001 eastern Sierra Nevada, California and its implications for dating structural events.
1002 *Journal of Structural Geology* 4, 177–195.

1003 Tobisch, O.T., Fiske, R.S., Saleeby, J.B., Holt, E., and Sorensen, S.S., 2000. Steep
1004 tilting of metavolcanic rocks by multiple mechanisms, central Sierra Nevada,
1005 California. *Geological Society of America Bulletin* 112, 1043–1058.

1006 Tobisch, O.T., Saleeby, J.B., Renne, R.R., McNulty, B., and Tong, W., 1995. Variations
1007 in deformation fields during development of a large volume magmatic arc, central
1008 Sierra Nevada, California. *Geological Society of America Bulletin* 107, 148–166.

1009 Weinberg, R. F., Sial, A. N., and Mariano, G., 2004. Close spatial relationship between
1010 plutons and shear zones. *Geology* 32, 377–380.

1011 Žák, J., Paterson, S.R., and Memeti, V., 2007, Four magmatic fabrics in the Tuolumne
1012 batholith, central Sierra Nevada, California (USA): Implications for interpreting
1013 fabric patterns in plutons and evolution of magma chambers in the upper crust:
1014 *Geological Society of America Bulletin* 119, 184–201.

1015 Zellmer, G.F., and Annen, C., 2008. An introduction to magma dynamics. *Geological*
1016 *Society, London, Special Publications* 304, 1–13.

FIGURE CAPTIONS

Figure 1. (A) Map of California with Sierra Nevada Batholith highlighted in black. (B) Sierra Nevada Batholith with four Late Cretaceous Intrusive Complexes highlighted. (C) Simplified map of central Sierra Nevada. Boxes show the areas where strain data are collected. Gray is Western Metamorphic Belt. Green is Mesozoic arc-related volcanic rock. Dark blue are Paleozoic rocks. Light blue is sedimentary rock of various ages. WMB=Western Metamorphic Belt, WFB=Western Foothill Belt, CFB=Central Foothill Belt, EFB=Eastern Foothill Belt, SCT=Sullivan Creek Terrane, CAL=Calaveras Complex, SFF=Shoo Fly Formation, PM=Piute Meadow pendant, SL=Saddlebag Lake pendant, RR=Ritter Range pendant, JL=Jackass Lake pendant, IM=Iron Mountain pendant, MM=Mountain Morrison pendant, MG=Mountain Goddard pendant, BC=Boyden Cave pendant, OC=Oak Creek pendant, LIS=Lake Isabella pendant, PL=Piute Lookout pendant, S=Sonora Intrusive Complex, T=Tuolumne Intrusive Complex, MP=Mono Pass Intrusive Complex, MW=Mt. Whitney Intrusive Complex, SF=San Francisco, LA=Los Angeles. Index of shear zone/fault: 1=Sierra Crest Shear Zone; 2=potential Mojave-Snow Lake fault; 3=Permo-Triassic Foothills suture; 4=Melones fault; 5=Bear Mountain fault.

Figure 2. Field photos showing strain markers. (A) Triassic volcanic breccia from Sullivan Creek Terrane. (B) Pebbly mudstone from Calaveras Complex. (C) Late Triassic Cooney Lake conglomerate from Saddlebag Lake pendant. (D) Cooney Lake conglomerate cut by a Middle Jurassic pluton. Black lines show the intrusive contact. (E) Late Triassic volcanoclastic rocks from Saddlebag Lake pendant. (F) Jurassic conglomerate from Saddlebag Lake pendant. (G) Cretaceous volcanic breccia from Saddlebag Lake pendant. (H) Cretaceous volcanic rocks from Saddlebag Lake pendant with steep bedding cut by a Cretaceous pluton.

Figure 3. Lower hemisphere, equal area projections of metamorphic foliations and lineations from Western Metamorphic Belt and host rock pendants in central Sierra Nevada. Kamb contours with 2.0 sigma contour interval (C.I.) are added. Dots in the red shade are poles to the metamorphic foliation and dots in the blue shade are metamorphic lineation. N_F =Number of foliation, N_L = Number of lineation. Cylindrical best fits are calculated using Stereonet 9.0 and 95% confidence ellipses (Allmendinger et al., 2012; Cardozo and Allmendinger, 2013).

Figure 4. (A) Hus plot (Hus, 1966; Hossack, 1967) of all 612 data with age constraints. Yellow part is the zoomed area shown in plot (B). Hus plots are made using the program developed by Mookerjee and Nickleach (2011). Points are color-coded by rock ages. Lode's ratio is calculated based on Lode (1926).

Figure 5. Finite strain measurements plotted in the space of X- and Z-extension percent (negative value = shortening; X, Y, Z are the longest, intermediate and shortest axes of a finite strain ellipsoid). Each dot represents an individual strain measurement. The central coarse red curves represent the relation between X- and Z- extension if the deformation is plane strain. The thin red curves represent the relation between X- and Z- extension if there is volume loss or Y-extension during deformation. Measurements from slates are shown as yellow dots, whose finite strain ellipsoids are likely to be affected by primary compaction during sedimentation and lithification. The plot suggests that the host rocks in the central Sierra Nevada are strained at about 50% in average with <15-20% Y-direction length change and/or potential volume changes.

Figure 6. Hsu plots of 9 major rock types from Western Metamorphic Belt and central Sierra Nevada. Strain magnitude (ϵ) contour interval is 0.3. Lode's ratio (γ) contour interval is 0.25.

Figure 7. Strain magnitude (ϵ) and Lode's ratio (γ) plotted against rock age. Averages of strain magnitudes and Lode's ratios of rocks of the same/similar ages are plotted as filled circles. Red bars show 2 standard deviations. Individual measurements are plotted as empty squares. The arrows show the interpreted trends. Strain magnitude decreases from Triassic rocks to Early Cretaceous rocks and it spreads in Late Cretaceous rocks (100-80 Ma). Lode's ratio slightly increases from Triassic rocks to Early Cretaceous rocks and its range spreads in Late Cretaceous rocks (100-80 Ma).

Figure 8. Strain magnitude (ϵ) and Lode's ratio (γ) plotted against paleo-depth of rock (paleo-depth is expressed in the unit of kilobar, kbar). Averages of strain magnitudes and Lode's ratios of rocks of same paleo-depths are plotted as filled circles. Red bars show 2σ errors. Individual measurements are plotted as empty squares. The arrows show the interpreted trends. Strain magnitude generally increases from 1 kbar to 3 kbars then sharply increases at 4 kbars. Lode's ratio has the highest value at 1 kbar and decreases to about 0.2 at 1.5-3 kbars then sharply decrease to -0.65 at 7 kbars. Notice that the uncertainty associated with barometry data is usually about ± 1 kbar.

Figure 9. (A) Lode's ratio (γ), (B) strain magnitude (ϵ), and (C) Z-shortening plotted against the distance from intrusive contacts of Tuolumne Intrusive Complex. The locations of eastern and northern contacts are shown in the Figure 1.

Figure 10. Plot shows the minimum downward transport rates of host rocks in the southern and central Sierra. The horizontal axis is the age difference in million years between the host rocks and intruding plutons. The vertical axis is the emplacement depth in kilobar (kbar) of the hosting plutons determined by Aluminum-in-Hornblende barometry. The background color shows the absolute downward transport rates. The rate varies from 2-13 km/Myr for most of the data. Location is given near the data

points, which are color-coded by the age of plutons.

Figure 11. Three proposed models for ductile straining mechanisms. Model A is homogenous tectonic thickening only. Model B only has the magmatism. Model C combines tectonism and magmatism. A1, B1, and C1 show the cartoon illustrating the corresponding models. A2, B2, and C2 show predicted trends of Lode's ratio (γ), or strain shape, changing with depth. A3, B3, and C3 show predicted trends of strain magnitude (ϵ) changing with depth. See text for full discussion. BDT= Brittle-ductile transition.

Figure 12. A possible model for Mesozoic Sierra Nevada based on the Models C shown in Figure 11. An arc crustal column is shown with tectonic shortening, ascending magma bodies, and crystallized plutons. Above the brittle-ductile transition, thrusts, duplex structures and as well as other host rock drop-down processes (e.g. stoping, ridge bedding rotation) accommodate the tectonic shortening strain and transport supracrustal materials downwards. The majority part of the arc crust below the brittle-ductile transition is deformed in the ductile fashion. Host rocks are forced to transport downward through escape channels. Strain magnitude increases and strain become more constrictional towards depth. The downward transported host rocks may enter the MASH zone at lower crust at about 70 km, where host rocks might be desegregated and assimilated. The thickness of the crust shown here is based on the discussion in Section 5.2.

Figure 13. Crustal thickness determined by host rock strain (ϵ_H) and pluton volume fraction (β). The initial thickness of the crust is assumed to be 25 km and the calculation takes 20% shortening in plutons and 20% volume loss/length change along Y-direction of host rocks into account. If $\epsilon_H=-0.65$ and $\beta=0.8$, the resulting thickness of crust is ~97 km. The crustal thickness show in this plot does not include exhumation.

Exhumation will reduce the thickness to 77-87 km for Late Cretaceous Sierra Nevada (see text).

Figure 14. Cartoon showing that exhumation based on barometry of plutons may only give a minimum value of exhumation. (A) is the case when there is no crustal thickening. At Time 1, Pluton A was emplaced at 10 km depth and at Time 2, Pluton B was emplaced at 8 km depth. The difference between emplacement depth of Pluton A and B is 2 km and equals to the thickness of rock eroded. (B) is the case when there is crustal thickening. At Time 1, Pluton A was emplaced at 10 km depth and at Time 2, Pluton B was emplaced at 8 km depth. The difference between emplacement depth of Pluton A and B is less than the thickness of rock eroded.

Figure 15. Elevation as a function of crustal thickness and arc root thickness if crust is isostatically balanced (Eq. 3). For the estimated crustal thickness range of Sierra Nevada at 80 Ma (~80 km), the elevation will be 4.5-5.7 km if the arc root is 30 km. The area between horizontal dashed lines is the elevation range of Puna-Altiplano plateau of the central Andes (4-5 km, Allmendinger et al., 1997). This calculation assumes the crust density $\rho_c=2.9 \text{ g/cm}^3$, mantle density $\rho_m=3.3 \text{ g/cm}^3$, and arc root density $\rho_r=3.5 \text{ g/cm}^3$ based on Lee et al. (2015). This calculation also assumes that 0 km elevation when the crust is 25 km thick.

Figure 16. Cartoon shows how magmatism and deformation thicken the crust. (A) is the geometry of an arc system before magmatism and deformation. The initial thickness of crust (H_0) is 25 km and elevation (h_0) is at sea level (0 km). The mantle flow is not impeded. (B) is the geometry of the arc system after magmatism and deformation and it may apply to the Sierra Nevada arc at ~85 Ma. Crust is thickened (H') to ~80 km and elevation (h') is ~5km. The total thickness of crust and arc root extending into mantle wedge is ~ 80 km. Mantle flow is impeded. The thickened crust could also cool down the mantle wedge. The impeded mantle flows, cooling effect, and the slab flattening are likely to contribute to the migration of magmatism.

Fig. 1

Fig. 2

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Model A: Tectonism only

A1

A2

A3

Model B: Magmatism only

B1

B2

B3

Model C: Tectonism and magmatism combined

C1

C2

C3

Fig. 12

Fig. 13

Time 1

Time 2

A

Time 1

Time 2

B

Fig. 14

Fig. 15

Before magmatism and deformation

After magmatism and deformation

Fig. 16

Graphical Abstract