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Revision #2

Repeated, multiscale, magmatic erosion and recycling in an upper-crustal pluton: implications for magma chamber dynamics and magma volume estimates

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Abstract

The Tuolumne Intrusive Complex, an upper-crustal (7-11 kilometer emplacement depths), incrementally constructed (95-85 Ma growth history) plutonic complex (~1100 km^2) preserves evidence from a number of datasets indicating the repeated, multi-scale, magmatic erosion of older units occurred and that some eroded material was recycled into younger magma batches. These include (1) map patterns of internal contacts (1000s kilometers) that show local hybrid units, truncations, and evidence of removal of older units by younger; (2) the presence of widespread xenolith and cognate inclusions (1000s), including "composite" inclusions; (3) the presence of widespread enclaves (millions), including "composite" enclaves, plus local enclave swarms that include xenoliths and cognate inclusions; (4) the presence of widespread schlieren-bound magmatic structures (>9000) showing evidence of local (meter-scale) truncations and erosion; (5) antecrystic zircons (billions) and

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other antecrystic minerals from older units now residing in younger units; (6) whole rock geochemistry including major element, REE, and isotopic data; and (7) single mineral petrographic and geochemical studies indicating mixing of distinct populations of the same mineral. Synthesis of the above suggest that some erosion and mixing occurred at greater crustal depths but that 1000s of "erosion events" at the emplacement site resulted in removal of ~35-55% of the original plutonic material from the presently exposed surface with some (~25%?) being recycled into younger magmas and the remainder was either erupted or displaced downwards. The driving mechanisms for mixing/recycling are varied but likely include buoyancy driven intrusion of younger batches into older crystal mushes, collapse and avalanching along growing and oversteepened solidification fronts within active magma chambers (1 km^2 to >500 km^2 in size), and local convection in magma chambers driven by internal gradients (e.g., buoyancy, temperature, rheology).

Key words: Magmatic recycling, erosion, mixing, Sierra Nevada, solidification front erosion, Tuolumne Intrusive Complex,

Introduction

A broad discussion in the Earth Sciences community continues about the sorts of magmatic processes that take place within middle and upper crustal magma chambers (defined as interconnected regions of crystal-melt mixtures) before they crystallize into plutons (solidified intrusive bodies). Are these magma chambers small, ephemeral, and fairly static, in which little to no mingling, mixing, or

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fractionation occur, or do large, longer-lived, dynamic magma chambers sometimes form in which extensive mixing, mingling and fractionation significantly change original magma source characteristics? A slight modification of this debate focuses on the degree to, and mechanisms by, which separate magma batches (e.g., originally separate batches of crystal-melt mixtures) physically and chemically interact as they arrive in incrementally growing magma chambers or plutons (Michaut and Jaupart, 2006). As these new batches arrive preexisting material must be moved: there has been little discussion about how space is made for these younger batches as they move into the growing magma chambers or plutons. In the context of the previous questions, one can ask how younger batches displace older intrusive materials and whether some older material is recycled into arriving batches by mixing, physical incorporation or assimilation? The latter processes require some form of significant physical interaction in which younger batches, while still in a crystal mush state, can "erode" older materials and incorporate or "recycle" them into the arriving batch. Attempting to address many aspects of these debates draws attention to what sorts of processes occur along internal contacts either between separate batches (Bergantz, 2000; Žák and Paterson, 2005; Memeti et al, 2010) and/or along solidification fronts within magma chambers (Marsh, 1996, 2006; Žák and Paterson, 2010).

Below we address the above questions by examining evidence for internal processes at scales ranging from ~100 kilometers to centimeters in the Tuolumne Intrusive Complex, or TIC (Fig. 1). This pluton has an ~10 m.y. history of growth and has been

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used as an example of both an incrementally grown pluton in which little interaction has occurred between batches at the emplacement site, and thus source magma characteristics have been preserved (Coleman et al., 2004; Gray et al. 2008; Schöpa and Annen, 2013) and as a pluton where widespread magmatic mixing, mingling and fractionation resulted in significant compositional modification of source magmas at the emplacement level (Reid et al., 1993; Burgess and Miller, 2008; Žák and Paterson, 2010; Memeti et al., 2014).

Summary of previous work on Tuolumne Intrusive Complex

The TIC is one of a number of ~1000 km^2 Cretaceous intrusive complexes that are exposed along the eastern crest of the Sierra Nevada and show a normal zonation from older, more mafic marginal phases to younger, more felsic, central phases (Fig. 1; Bateman, 1992; Kistler and Fleck, 1994). These composite intrusions were emplaced at ~7–11 kilometer paleodepths (Ague and Brimhall, 1988; Memeti et al., 2009) during the Late Cretaceous (120–85 Ma) magmatic flare-up (Ducea, 2001; Paterson and Ducea, 2015). During construction of these plutons the arc was actively deforming during dextral transpression and vertical thickening (Tikoff and Greene, 1997; Tobisch et al., 2000; Cao et al., 2015). The TIC is a metaluminous, magnetite series intrusion that has a main body and four lobes extending outwards from the central body. The former consists of three major units nested into one another that become progressively younger and more SiO_2-rich inwards and to the northeast (Memeti et al., 2010b, 2014). The outer 95–92 Ma Kuna Crest granodiorite (Fig. 1) is heterogeneous, ranging from mostly fine- to medium-grained

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granodiorites and tonalites to local gabbroic, dioritic, and granitic compositions. Most Kuna Crest domains exhibit local sheeting along the outer margins of the batholith (Memeti et al., 2014) and moderately intense magmatic and local subsolidus fabrics. The 92–88 Ma Half Dome Granodiorite (Fig. 1) is subdivided into an outer equigranular unit characterized by conspicuous euhedral hornblende (≤2 centimeter lengths), biotite books and titanite, and the inner porphyritic unit, which contains K-feldspar phenocrysts up to ~3 centimeters long and fewer mafic minerals (Bateman and Chappell, 1979). The 88–85 Ma Cathedral Peak unit (Fig. 1) is mainly composed of medium-grained granodiorite with up to 12-centimeter long K-feldspar phenocrysts and 1 centimeter quartz aggregates (Bateman and Chappell, 1979). The Cathedral Peak is a granite in places, especially in the northern Cathedral Peak lobe (Memeti, 2009).

Each of the four lobes compositionally consists of one of the main units in the central body, thus forming two southern lobes, one extending out from the porphyritic Half Dome unit and one from the Kuna Crest unit, and two northern lobes, one extending out from the equigranular Half Dome unit and the other forming the NE end of the Cathedral Peak (Fig. 1). Each lobe shows internal zoning from more mafic margins to more felsic centers and inward younging U/Pb zircon ages (e.g., Memeti et al., 2010). All of these units, both in the main body and in the four lobes, locally contain regions of late leucogranites with stock-like, sheet-like or cross-cutting dike-like shapes, the largest of which is the ca. 87.5 Ma Johnson Granite Porphyry (Bateman and Chappell, 1979; Bracciali et al., 2008).

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Geochemistry: Bateman and Chappell (1979) used major- and trace element compositions to argue for closed-system fractional crystallization to explain the roughly concentric compositional zoning in the TIC. A subsequent Nd-Sr isotope study by Kistler et al. (1986) revealed differing isotopic values for different TIC units that were better explained as mixtures of mantle-derived basaltic and crustal granitic magmas, with the latter magmas increasing in younger units. Gray (2003) and Gray et al. (2008) argued that major- and trace-element variations in the TIC are decoupled from each other and that their lack of any uniform spatial pattern can be best explained by incremental emplacement through diking, implying that any geochemical variability is due to source heterogeneity. However, Memeti et al. (2009, 2014) challenged this conclusion and instead argued that major- and trace-element variations define separate but overlapping mixing trends for the different TIC units. Coleman et al. (2004) used a geochronologic database to argue that little fractionation or mixing occurred at the emplacement site (see also, Coleman, 2005; Gray, 2003; Gray et al., 2008). Memeti and colleagues examined the peripheral 10-40 km^2 magmatic lobes of the TIC to evaluate the crystallization history, causes of normal zoning (older, mafic marginal units to younger, felsic interiors) preserved in each lobe (Memeti, 2009; Memeti et al., 2010). Geochemical histories exist between lobes with the southern HD lobe dominated by fractionation (Economos et al., 2010) and others typically preserving more complicated histories of amalgamation of batches, fractionation, and mixing (Memeti et al., 2010; 2014; Barnes et al., 2016).

Several authors have examined the geochemistry of schlieren-bound magmatic

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structures within the TIC (Reid et al., 1993; Solgadi and Sawyer, 2007; Burgess and Miller, 2008; Paterson et al., 2008; Paterson, 2009; Žák et al., 2009). These studies are in general agreement and indicate that: (1) the sharp bases of schlieren show a modal increase in biotite and hornblende and an unusually high increase in accessory minerals including zircon, sphene, apatite, rare allanite, magnetite and other oxides and thus are cumulates; (2) that the composition of the schlieren are dramatically different than host magma compositions, do not follow the trends seen in the main units, have high REE enrichments reaching 800-1000 times chondritic values for LREE's, and are compatible with depletion in Al_2O_3, Na_2O and enrichment in Rb, MgO, CaO, K_2O, TiO_2, P_2O_5, Zr, Y, La, and Ce relative to nearby plutonic material; and (3) that the origin of schlieren cannot be due to a simple accumulation of crystals during fractional crystallization but require mixing/mingling and/or erosion and recycling of material from older batches (Ardill et al., 2015). Solgadi and Sawyer (2007) particularly noted that microprobe analyses of hornblendes from these schlieren define three distinct populations that compositionally match those seen in the three main units (KC, HD, CP) of the TIC. They concluded that mechanical erosion, mixing of crystals, and redeposition played a dominant role in schlieren formation.

Wallace and Bergantz (2002), Krause et al. (2009, 2010), Memeti et al. (2009, 2013, 2014), and Barnes et al. (2016) have recently focused on mineral-scale geochemical studies and concluded that the geochemical zonation pattern of K-feldspars, plagioclase, hornblende, biotite, and titanite show varied internal patterns that often

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require mixing of disparate crystal populations. Thus a large, published geochemical dataset, including both whole rock and mineral-scale major element, REE, and isotopic data support the conclusion that crystal-crystal and crystal-melt mixing, from hand sample to > 10 kilometer scales, is widespread in the TIC.

Geochronology of the TIC: Early studies of the TIC using U/Pb, Rb-Sr, K-Ar, and $^{40}Ar/^{39}Ar$ techniques recognized that its assembly took millions of years (Stern et al., 1981; Fleck and Kistler, 1994; Kistler and Fleck, 1994). A more recent study carried out by Coleman et al. (2004) determined an ~9 m.y. construction history for the TIC and concluded that the TIC in general, and the Half Dome pluton in particular, must have been formed from multiple small batches of magma, perhaps as a series of sheets or dikes. Furthermore, they argued that convecting, fractionating and mixing magma chambers were rare or absent and as a consequence, that the geochemical trends observed in the TIC must be inherited from the magma source characteristics rather than modified by in situ processes. However, the Coleman et al. (2004) study used multi-crystal analyses of zircon populations that have been shown to be particularly prone to unrecognized Pb loss and age-averaging effects of analyzing mixed autocrysts, antecrysts and xenocrysts (Mundil et al., 2001; Miller et al. 2007; Memeti et al., 2014).

It is now well established in a number of plutonic and volcanic systems, including the TIC, that zircon populations often contain mixed zircon populations (Fig. 2a) including xenocrysts (inherited from older host rocks or melt source regions),

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antecrysts (recycled from earlier batches of magma), and autocrysts (grown in final melts)(e.g., Hildreth, 2004; Bacon and Lowenstern, 2005; Miller et al., 2007). In such mixed populations the youngest zircon ages (Fig. 2a) probably best represent the time when zircon crystallization ended if Pb loss can be entirely excluded (Irmis et al., 2011). Because zircon crystallizes over a fairly narrow temperature range in calc-alkaline magmas as a function of Zr saturation, zircon ages cannot be equated to an emplacement or arrival age of magma batches but instead are a better estimate of late crystallization. Memeti et al. (2010) presented new U-Pb, CA-ID-TIMS, single-crystal analyses on 224 individual zircons or zircon fragments from 26 samples across the TIC. Figure 1 displays these new TIC ages determined from the youngest autocrystic zircons (see Ickert and Mundil, 2015 for calculation methods). The new data show a general inwards and northerly younging pattern in the main pluton and local inward younging patterns of different ages in the 4 lobes extending outward from the main pluton (Memeti et al., 2014).

Matzel et al. (2006b) and Memeti et al. (2014) also discuss new $^{40}Ar/^{39}Ar$ ages from the TIC and adjacent older units. Samples collected along two SW-NE corridors across the TIC, in the lobes, and in local host units were used to separate two different biotite and hornblende grain-size fractions (800–900 μm and 150– 180 μm). $^{40}Ar/^{39}Ar$ step heating for large single crystals (hornblende and biotite) as well as total fusion experiments (small crystals) were performed on these populations from each sample (Fig. 2b). Besides establishing a general subsolidus cooling pattern in the TIC and nearby host rocks that roughly mimics the zircon age pattern,

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these biotite ages show that cooling from the margin to central parts of the TIC was non-linear and that large parts of the TIC show similar cooling ages (Fig. 2b).

In regard to the theme of the present paper, there are several important aspects of the new geochronologic data: (1) zircon antecrysts are rare in the outer/older Kuna Crest units and increase in abundance in the inner units (Fig. 2a). Memeti et al. (2014) concluded that this result occurred from mixing or recycling of zircons from older units into younger units; (2) a comparison of local gradients in zircon ages across mapped internal contacts indicate that monotonous inward younging (argued for by Coleman et al., 2004) is not the general pattern and instead several "kinks" with gentle slopes at ~94 Ma, 90 Ma, and particularly at ~88-87 Ma, are spatially associated with mapped internal contacts (Memeti et al., 2014). These observations suggest that there are both age jumps across internal contacts and that at certain times simultaneous zircon crystallization occurred over broad areas in the TIC, indicating the presence of larger magma chambers; (3) zircons with ages of these broader regions occur as antecrysts in younger units suggesting that the original extent of older units (e.g., Kuna Crest or Half Dome granodiorites) were significantly larger and subsequently displaced and partly assimilated during intrusion of the younger units; (4) autocrystic zircon ages typically show an age dispersion (greater than errors) along concordia that extend from a few hundred thousand to over 1 million years (Fig. 2a), raising the possibility that melt from which these zircons grew also existed in the TIC for at least these durations; (5) the biotite cooling ages also show broad regions with similar cooling ages (Fig. 2b),

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suggesting that either larger magma batches arrived in the growing pluton or if these regions grew incrementally, they did so rapidly and formed large magma chamber(s) that shared similar autocrystic zircon growth and biotite cooling histories.

We suggest that the geochronology supports the following hypotheses: (1) batch arrival times remain poorly constrained but must be slightly before the ages in Figure 1; (2) batch sizes are variable and remain poorly constrained; (3) zircon and biotite ages indicate that at least three times during construction of the TIC fairly large magma chambers existed in the main batholith and smaller ones in the lobes. (4) Durations of these magma chambers could have easily ranged from 0.5 (lobes) to 1.5 m.y. (main pluton) given the spread of autocrystic zircon ages plus any additional melt residence time before zircon growth. This conclusion is supported by thermal models of the TIC (Paterson et al. 2011), by the transfer of latent heat of crystallization to the melt (e.g., Michaut and Jaupart, 2006) and by the recognition that magmas are "thermally insulating" as thermal diffusivity and conductivity are drastically decreased in magma mushes at higher temperatures (Whittington et al., 2009; Nabelek et al., 2012).

INTERNAL CONTACTS

Internal contacts provide information either about magma chamber or pluton construction during incremental growth or about internal magma chamber processes that may be somewhat independent of pluton growth (e.g., Pitcher and

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Berger, 1972; Hardee, 1982; Hutton, 1982, 1992; Lagarde et al., 1990; Paterson and Vernon, 1995; McNulty et al., 1996; Vigneresse and Bouchez, 1997; Paterson and Miller, 1998; Wiebe and Collins, 1998; Johnson et al., 1999; Paterson et al., 2011). Internal contacts in granitoids have variable characteristics such as: (1) sharp or gradational; (2) defined by distinct compositions and/or distinct microstructures; (3) along-strike lengths that range from centimeters to kilometers, and (4) the presence or absence of magmatic structures (e.g., host rock rafts, troughs, enclaves, and schlieren) or objects (e.g., enclaves, stoped blocks, rafts) concentrated along them. Contacts between the TIC units vary from knife sharp to gradational over hundreds of meters (Bateman and Chappell, 1979; Bateman, 1992; Žák and Paterson, 2005). The different types of internal contacts may form by (1) the addition of new magma batches into a magma chamber, (2) localized flow of magma with different viscosities within an existing magma chamber, and (3) processes during crystallization, such as crystal-liquid fractionation. The latter two processes imply that internal contacts may not necessarily imply the arrival of multiple magma batches during growth of a pluton.

Figure 1 displays the position and general nature of the main internal contacts in the TIC. Several patterns have emerged. First, these contacts can be followed for distances from 1 to 100 kilometers. Second, a single contact will often vary in character from sharp to gradational along strike sometimes over scales of 100s of meters. Even so, certain contacts tend to typically be sharp or gradational no matter which older units are juxtaposed across the contact. For example, the outer

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Cathedral Peak contact is commonly quite sharp against either pre-TIC host rocks or the porphyritic Half Dome unit. Some late leucogranites also have sharp contacts against older units. However, the contact between the Kuna Crest-equigranular Half Dome units is more often gradational or has preserved hybrid units along it (Fig. 1). Although the previously discussed zircon geochronology was not collected with the target of evaluating age variations across internal contacts, both the ages reported by Coleman et al. (2004) and by Memeti et al. (2010) are suggestive of a relationship between contact sharpness and an increasing magnitude of age jumps across the contacts.

A number of maps of internal contacts, constructed at 1:10,000 scale, have been completed in the TIC (Fig. 1), and four examples (Fig. 3, 4, 5, 6) indicate the variable nature of these contacts, including the presence of "hybrid" magmatic zones locally formed between units and the presence of a number of truncations of older units by younger.

Glen Aulin: The ~102 Ma El Capitan granite forms a prominent eastward protrusion that TIC units wrap around near the western margin of the TIC. Here the outermost units of the TIC tend to broadly follow the border of this protrusion (Fig. 3). In detail some important map patterns exist. First small pendants and rafts of metamorphic host rock, all considered to be pieces of the former Snow Lake block consisting of Paleozoic passive margin assemblages (e.g., Memeti et al., 2010a), occur near or at the external contact between the El Capitan and TIC. The most

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remarkable case is a 5 centimeter to meter wide strip of metamorphic rock immediately at the contact between the two plutons north of Glen Aulin that, in spite of its thin width, is still preserved. Other host rock occurs as blocks or rafts within the two plutons. Here the outermost unit of the TIC, excluding local mafic cumulates, is called the Glen Aulin tonalite, which is considered to be part of the Kuna Crest unit. The next major unit is the equigranular Half Dome granodiorite with its prominent euhedral hornblendes and sphene. Locally a hybrid unit with shared Kuna and Half Dome characteristics occurs between the two units (Fig. 3). Along strike the contact is sharp and no hybrid unit exists. The equigranular Half Dome granodiorite changes gradually inwards into the porphyritic Half Dome granodiorite, which is also discontinuous along strike (Fig. 3). The most extensive, innermost and youngest unit mapped in this area is the Cathedral Peak granodiorite with its large K-feldspar megacrysts and more abundant biotite than hornblende. Close inspection shows that the Cathedral Peak contact is typically sharp and in places cuts across both Half Dome units and directly intrudes the Kuna Crest. Except for the Cathedral Peak granodiorite, all other units are surprisingly thin in this area compared to their exposed thicknesses elsewhere in the TIC.

K*una Crest units in Upper Lyell Canyon*: Two NW-striking, steeply dipping tonalite to granodiorite Kuna Crest units have been mapped along the SW margin of this lobe by Memeti et al. (2010). These units are sharply truncated (near Ireland Lake on Fig. 4) by an ~east-west striking, steeply dipping, younger Kuna Crest granodiorite unit and an east-west striking fairly homogenized, Kuna Crest-Half Dome hybrid unit

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sharing field and geochemical characteristics of both of these units (Fig. 4). The detailed characteristics and geochemistry of intrusive units in this lobe are presented by Memeti et al. (2010, 2014).

Fletcher Peak area: Žák and Paterson (2005) examined the geology in the Fletcher Peak area near the southern end of the TIC (Fig. 1, 5). Here the ~97 Ma Ireland Lake granite intrudes older metavolcanic units and together these two units make up the host rock that the TIC intruded. The Kuna Crest granodiorite intrudes these host rocks, contains host rock blocks and rafts in it, and forms an east-west striking, steeply dipping intrusive unit that strikes towards the east-west Kuna Crest unit in Figure 4. The western end of this east-west Kuna Crest unit is truncated by a complex mingling zone consisting of hybrid units, local sheeting, and a variety of magmatic structures. The porphyritic Half Dome unit intrudes this mingling zone along a sharp contact that truncates structures within the mingling zone. No equigranular Half Dome unit is preserved here except in local areas within the mingling zone or within magmatic structures in this zone. Finally, the Cathedral Peak unit intrudes the inner part of the porphyritic Half Dome unit along another sharp contact.

Sawmill Canyon area: Paterson et al. (2008; 2014) discuss complex cross-cutting intrusive relationships in the Sawmill Canyon area along the eastern margin of the TIC (Fig. 1, 6). These authors established that the northward striking and steeply dipping outer and older (~94-95 Ma) Kuna Crest unit and adjacent inner and

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younger (~92-88 Ma) equigranular Half Dome granodiorite are sharply truncated, at the centimeter to meter scale, by an east-west striking, complicated sheeted and mingled zone dominated by a variety of magmatic structures. This sheeted zone swings to the NNW along the NNW striking eastern margin of the TIC (Fig. 6) and then is eventually truncated by the younger (~88 Ma) Cathedral Peak granodiorite. North of the Sawmill Canyon area, the Kuna Crest unit is completely absent. Only two blocks, one large (100s of meters) and one small (several meters) of Half Dome granodiorite have been found north of the Sawmill Canyon area, and are now largely surrounded by Cathedral Peak granodiorite (Fig. 6a). Field mapping and geochemical studies from within the “sheeted zone” (Paterson et al., 2008) indicate that much of the material is made up of porphyritic Half Dome magma, but that a complete mingling of Kuna Crest, Half Dome, and Cathedral Peak phases (both solid pieces and crystal mush mingled phases) make up this zone. Smaller versions of these sheeted zones crop out immediately south of the Sawmill Canyon area along what are interpreted to be east-west striking cracks that were filled by pulses of magma (Paterson et al., 2008). The main Sawmill Canyon sheeted zone is interpreted to have formed along one of these cracks across which the northern Kuna Crest and Half Dome phases were eventually removed.

The map patterns presented above provide examples of two important concepts. The first is that hybrid magmatic zones bordered by gradational contacts sometimes form in between the main TIC units. These hybrid units range from fairly homogenized zones with mineralogical characteristics of the two adjacent units, to

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irregular, heterogeneous (less homogenized) zones with different parts sharing characteristics of one of the two adjacent rock types. This suggests that some degree of mixing or mingling between adjacent units occurred and required at least sufficient interconnected melt in both units. Sometimes these "hybrid" zones can be very complex (e.g., Fletcher Peak and Sawmill Canyon) with sheeting and a variety of compositionally defined magmatic structures suggesting that they were active zones of magma pulsing, mixing, mingling, and local convection.

The second concept is that the map patterns show examples of repeated truncations, and thus implied removal or erosion of older by younger intrusive units along these contacts. These truncations, sometimes regional in extent, can be viewed at the meter to centimeter scale and in cases shown to truncate local compositionally defined contacts, or compositionally defined magmatic structures, and in rare cases even large phenocrysts. The truncation of these objects suggests that complex brittle-ductile behavior occurred during removal or "erosion" of older materials (typically crystal-rich mushes) during intrusion of younger magma batches.

Outcrop-scale examples of erosion/recycling

At 1 to 10 meter scales three features support repeated erosion/recycling: (1) "composite" xenoliths of host rocks or cognate inclusions of older parts of the TIC now present as blocks in younger plutonic phases; (2) widespread and fairly evenly distributed mafic enclaves and local enclave swarms with xenoliths; and (3) a variety of compositionally defined magmatic structures that are sharply truncated

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by younger magma batches implying erosion of older crystal mushes and removal or recycling of these mushes. We present examples of these below.

Xenoliths (solid objects of material foreign to the TIC) and cognate inclusions (fairly coherent blocks of solid material or crystal mushes from an earlier part of the TIC) occur in all units within the TIC (Fig. 7, 8, 9, 10). The most exciting examples include "composite" cases where an older object is surrounded by an older plutonic phase and in turn both are surrounded by younger plutonic phase. Fig. 7 shows a number of examples from the Kuna Crest unit. Figure 7a shows Paleozoic metasedimentary stoped blocks, 7b older plutonic, volcanic and mafic enclaves in a large mafic enclave now in a Kuna Crest matrix, 7c, 7d, and 7e depict "composite" blocks of Jurassic volcanic blocks (7c, 7e) and calc-silicate metasediments (7d), respectively, surrounded by mafic Kuna Crest tonalite in more felsic Kuna Crest granodiorite. Figure 7f displays numerous cognate inclusions and mafic enclaves and rare xenoliths in a hybrid Kuna Crest granodiorite. Note that in all of the above examples, the compositional heterogeneity over short distances and variable structural orientations between nearby blocks support that these blocks have moved significantly from their site of origin accumulated in the present location.

Figure 8 shows similar examples residing in a matrix of Half Dome granodiorite. Figures 8a and 8b show metavolcanic and metasedimentary xenoliths in the Half Dome granodiorite. Figures 8c and 8d show composite cognate inclusions of former enclave swarms with local xenoliths that are attached to a matrix of Kuna Crest

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granodiorite and now reside together in equigranular Half Dome granodiorite. Note the sharp boundaries defining the rectangular shape of the blocks that cut and truncate individual enclaves indicating that a former more elongate enclave swarm (e.g., Tobisch et al., 1997) was broken apart into rectangular pieces, presumably during movement of Kuna Crest magma, and subsequently was incorporated into Half Dome magma. Figure 8e displays a composite block of volcanic pieces locally attached to and intruded by Kuna Crest magma now surrounded by a Half Dome granodiorite matrix. Figure 8f shows a collection of xenoliths, a large cognate inclusion of layered plutonic rock and numerous enclaves near the Kuna Crest – Half Dome contact.

Figure 9 shows inclusions residing in the Cathedral Peak granodiorite. Figure 9a shows a folded metavolcanic xenolith, 9b a cognate inclusion of Half Dome granodiorite, 9c a layered cognate inclusion of porphyritic Half Dome magma and 9d a composite inclusion of layered plutonic rock rimmed by Half Dome plutonic material in the Cathedral Peak granodiorite. Figure 9e shows a schlieren-bound tube sharply truncated by Cathedral Peak magma and 9f shows a composite inclusion with a small piece of layered plutonic rock (white arrow) in a matrix of layered Half Dome granodiorite (HD), now entirely surrounded by Cathedral Peak granodiorite.

Figure 10 shows various pieces of the Cathedral Peak granodiorite in the Johnson granite. Figure 10a shows Cathedral Peak K-feldspar megacrystic antecrysts, 10b an angular piece of Cathedral Peak granodiorite, and Figure 10c and 10d rounded

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(presumably by partial melting) pieces of Cathedral Peak, each with one large K-feldspar megacryst in a medium-grained Cathedral Peak granodiorite matrix, in the fine grained Johnson granite.

Figures 7-10 document that pieces of host rock or older parts of the TIC have been broken, moved, and are now in younger parts of the intrusive complex and thus provide direct evidence of "recycling" of material into new magma batches. The composite examples are particularly exciting since they require at least two "recycling events" to form. There is no evidence that any of these inclusions are traditional rafts that haven't moved from their site of origin. Instead all are now located at some distance from the nearest likely sources, show along-strike changes in rock type with nearby blocks, and typically show internal structures that are rotated from both regional structural patterns and nearby blocks. If one argues that the blocks haven't moved very far, the presence of older magmatic phases of the TIC in regions where only a younger phase is now preserved indicates that the older phases were originally more extensive and have since been removed and/or displaced out of the present plane of exposure.

Another feature in the TIC that deserves careful consideration in regards to mixing/recycling of magmas are the characteristics of enclaves and enclave swarms (Fig. 11). A great deal of attention was paid to enclaves in past Sierran studies (e.g. Pabst, 1928; Frost and Mahood, 1987; Barbarin, 1990, 2005; Tobisch et al., 1997) but little work has been published on them recently. Barbarin (2005) noted that

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both mafic dikes and mafic enclaves in central Sierran plutons exhibit a bulk compositional diversity but with broadly overlapping major, trace element and isotopic compositions and argued that both were produced by the mixing of the same two components in different proportions. Based on enclave microstructures and geochemistry, Barbarin (2005) concluded that a few enclaves formed from early cumulates at depth, whereas most formed by disruption and mingling/mixing of mafic dikes (e.g., Frost and Mahood, 1987) and some by disruption of schlieren at the level of emplacement. Their initial formation was followed by the sequential mixing and mingling during widespread dispersion, followed locally by collection processes to form enclave swarms (see also Tobisch et al., 1997). Barbarin (2005) also noted less common "composite enclaves" that consist of either host rock xenoliths enclosed in a mafic enclave, or one style of mafic enclave enclosed in another, both now residing in a younger more felsic granitic matrix.

Mafic microgranitoid enclaves are widespread in the TIC and occur in all of the three main units, although they are rare in the late leucogranites (Fig. 11). We also find less common synmagmatic mafic dikes (Fig. 11a) in various degrees of disaggregation (Fig. 11b) and enclave formation (Memeti and Paterson, 2007). Most TIC enclaves form roughly elliptical shapes and are remarkably evenly spaced (Fig. 11c). Local enclave swarms are fairly common in the TIC (Fig. 11d). These enclave swarms often include not only enclaves but cognate inclusions and host rock xenoliths supporting the interpretation that they are physical accumulations of formerly dispersed objects in the magma chamber. Composite enclaves locally occur

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(Fig. 11e) and some rectangular cognate blocks consist of former enclave swarms (Fig. 8c, 8d, 11f). These observations thus imply that many enclaves formed from disaggregated dikes, followed by dispersion in a magma chamber (by convection), followed by some form of local flow sorting resulting in accumulations of not only enclaves but previously formed xenoliths and cognate inclusions. Finally sometimes the enclave swarms are reintruded and broken into new cognate blocks.

Reid et al. (1993), Burgess and Miller (2008), Solgadi and Sawyer (2008), Paterson (2009), and Hodge et al. (2012) described a variety of compositionally defined magmatic structures in the TIC. For purposes of this paper, we focus on schlieren-defined tubes and troughs that repeatedly show sharp truncations of older parts of these structures and sometimes show evidence of reintrusion and recycling of the broken pieces into younger magmatic phases. Fig. 12 shows examples of these structures, which occur in all units of the TIC although they are most common in the Cathedral Peak and Half Dome granodiorites. Figures 12a and 12b show mafic schlieren defining two sets of magmatic troughs showing examples of sharp truncations of older schlieren layers. These truncations are visible at the centimeter scale and sharply cut across both mafic and felsic layers as well as subparallel magmatic mineral foliations with no deflection of the older layers. A new magmatic foliation is subparallel to the new, cross-cutting schlieren layer. Figure 12c shows a package of mafic-felsic schlieren layers of Half Dome granodiorite, defined by grading in proportions of mafic minerals, that were sharply truncated by younger batches of magma. None of the layers show any deflection or gradation into the

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younger plutonic material. Figure 12d shows schlieren in porphyritic Half Dome granodiorite that are sharply truncated by younger schlieren layers. Figure 12e shows an example of a "bifurcated" schlieren-bound migrating tube(s), with repeated truncations of older tube margins by younger, now surrounded and locally reintruded by Cathedral Peak granodiorite. These structures have magmatic fabrics parallel to the bases of the mafic layers that are overprinted by magmatic fabrics that are continuous from the surrounding Cathedral Peak granodiorite into the magmatic structures (e.g. Paterson, 2009). Figure 12f is an example in Cathedral Peak granodiorite of planar schlieren and subparallel magmatic foliation truncated by and presumably removed during the formation of a schlieren bound trough in which a new schlieren-parallel foliation is formed.

Studies of the compositions of these structures (e.g., Reid et al., 1993; Solgadi and Sawyer, 2007; Burgess and Miller, 2008; Paterson et al., 2008; Žák et al., 2009; Paterson, 2009) are in general agreement and indicate that: (1) the sharp bases of schlieren are marked by an abruptly higher concentration of biotite and hornblende and an unusually high concentration of accessory minerals including zircon, sphene, apatite, rare allanite, magnetite and other oxides, and thus are cumulates; (2) the compositions of the schlieren are dramatically different than host magma compositions, do not follow the trends seen in the main units, have high REE abundances reaching 800-1000 times chondrites for LREE's, lower abundances in Al_2O_3, Na_2O_3 and higher abundances in Rb, MgO, CaO, K_2O, TiO_2, P_2O_5, Zr, Y, La, and Ce; and (3) that the origin of the schlieren cannot be simple accumulations of the

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mafic and accessory minerals from a single magma, but instead require two or more sources for these minerals. Solgadi and Sawyer (2007) particularly noted that compositions of hornblendes from schlieren in the Sawmill Canyon sheeted zone define three distinct populations that match those seen in the three main units of the TIC (KC, HD, CP). They concluded that mechanical erosion, mixing of crystals, and magmatic redeposition played a dominant role in schlieren formation.

We suggest that these structures have several implications: (1) The reintrusion of magmatic structures by the surrounding magmas and overprinting of the structures by magmatic fabrics support the hypothesis that melt was present in the structures and the surrounding plutonic material when the structures formed; (2) magmatic structures preserve evidence for crystal sorting during local movement of magmas and thus that at least local non-static magma chambers existed; (3) the sharp truncations of compositionally-defined layers and subparallel mineral fabrics argue for erosion of crystal mushes and removal from the exposed plane of the eroded material. We find it very unlikely that strictly chemical processes formed the unusual model mineral proportions, the magmatic fabrics, and the sharp truncations; (4) the truncations, reintrusion and disruption of structures show that at the meter scale some materials were recycled into the surrounding magma mushes; and (5) the composition of the structures is best explained by crystal sorting of pre-mixed magmas and/or during erosion/recycling events.

Mineral-scale examples of erosion/recycling

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The following evidence for mixing and recycling in the TIC occurs at the mineral scale. First, as noted above, a close inspection of single zircon ages indicates that the number of antecrystic zircons statistically increases from the older marginal units to the younger interior units (Memeti et al., 2014), and that the ages of the antecrystic zircons in the interior units always match the ages of the older, outer units of the TIC (e.g., Fig. 2). Rare xenocrystic zircons occur, but these are statistically minor and occur in samples from both outer and inner units. A very likely scenario to explain this zircon pattern is that as inner magma batches arrive in or ascend through older magma mush units, they incorporate zircons from these units.

A second observation is that some minerals that likely formed in older units are found in younger units and thus are antecrystic or cognate minerals (Fig. 13). For example we find 1-3 cm, euhedral hornblendes in the Cathedral Peak unit (Fig. 13a), which typically lacks or is poor in hornblende. Memeti et al. (2014) determined that these hornblendes are surrounded by a Cathedral Peak phase characterized by biotite and, even though they maintain their hornblende shape, are largely replaced by biotite. We thus interpret these to be hornblendes that formed in the Half Dome granodiorite and were recycled into Cathedral Peak magma, where they were unstable and pseudomorphed by biotite.

A potentially similar example is the presence of K-feldspar megacrysts in the late leucogranites, such as in the Johnson porphyry granite (Fig. 10). Megacrysts are not usually found in these leucogranites and the presence of resorbed and rounded

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medium grained, biotite-bearing granodiorite around some megacrysts indicates that the megacrysts grew in Cathedral Peak magma before being recycled into the fine grained, biotite-poor leucogranites.

We have also suggested that some of the particularly large K-feldspar megacrysts in the outer parts of the Cathedral Peak (e.g., Memeti and Paterson, 2008) initially crystallized in the porphyritic Half Dome granodiorite and have continued their growth after being mixed into the Cathedral Peak magmas (Fig. 13c, d). This suggestion has yet to be fully tested.

Krause et al. (2009, 2010), Memeti et al. (2009, 2013, 2014), and Barnes et al. (2016) recently focused on the geochemistry of individual crystals, especially the geochemical zoning patterns of K-feldspars, but also plagioclase, hornblende, biotite, and titanite. The main results derived from electron microprobe x-ray element intensity mapping, and core-to-rim transects of electron microprobe spot and laser ablation-ICP-MS analyses are the following: (1) multiple populations with distinct crystal sizes and morphologies exist for some minerals (e.g., Solgadi and Sawyer, 2008; Barnes et al., 2016); (2) all minerals preserve zones if slow diffusing elements are examined, and the number and character of growth zones differ both in different mineral populations and spatially throughout the TIC; (3) growth zones in some minerals such as K-feldspar preserve many repeated compositional patterns suggesting that either multiple magma batches were added to the magma chamber, or that crystals were repeatedly moved into different parts of the magma chamber

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during growth (Memeti et al., 2014); (4) that multiple populations of each crystal, each with different trace element zoning patterns, are now mixed together at hand sample scale (Barnes et al., 2016), and (5) that simpler zoning and less heterogeneity (e.g., fewer different crystal populations) is preserved in crystals from the simpler and shorter duration lobes compared to rocks in the main TIC (Krause et al., 2009; Memeti et al., 2009, 2014). These authors proposed a model of crystals initially forming in different melt compositions and/or PT environments and then being mixed together into a new hybrid magma composition. Continued growth of minerals in the hybrid magma resulted in the similar REE patterns in rims of crystals that have different morphologies and mineral core chemistries.

Discussion

Summary erosion/recycling: We have summarized 7 datasets, listed below, all of which support the truncation and erosion of older magma mushes accompanied by recycling of some of the eroded materials into younger magma batches: (1) map patterns and characteristics of internal contacts that show local hybrid units, truncations, and evidence of removal of older units by younger; (2) widespread xenoliths and cognate inclusions, including "composite" examples; (3) widespread magmatic enclaves, including composite examples, plus local enclave swarms that also include xenoliths and cognate inclusions; (4) widespread schlieren-bounded magmatic structures showing evidence of truncations and erosion; (5) antecrystic zircons and other antecrystic minerals from older units now present in younger units; (6) whole rock compositions including elemental, REE, and isotopic data all

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supportive of extensive mixing; and (7) mineral petrographic and geochemical studies indicating mixing of distinct populations based on crystal morphology, elemental chemical mapping, and REE core-to-rim transects. Even if interpretations for one or two of these examples might be questioned, the combination of all seven provides compelling evidence that the erosion of older magmatic units and recycling of eroded materials were widespread in the TIC.

This erosion/recycling occurred during repeated events over a range of spatial scales. For example, (1) a comparison of the Fletcher Peak and Kuna lobe maps shows that Kuna Crest units are truncated by a younger Kuna Crest unit, which in turn is truncated by a complex sheeted unit in the Fletcher Peak map, which in turn is truncated by a unit consisting of Half Dome granodiorite. These map patterns require at least 5 erosion/recycling events at the kilometer scale; (2) "composite" cognate inclusions require at least two erosion/recycling events; (3) repeated truncations of schlieren and subparallel magmatic fabrics in troughs support repeated erosion/recycling events at the 10 meter scale; (4) gradual increase in antecrystic zircons in younger units are most easily explained if repeated erosion/recycling events occurred since it is unlikely that all units were present and in a magmatic state at the same time.

Spatial and temporal scales of these processes: When only a few representative examples are presented, it is difficult to assess the total number and spatial distribution of different features in a pluton and thus determine the magnitude of

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the inferred processes. We attempt to address this by estimating the approximate number, size and spatial distribution of the compositional and structural examples discussed above by extrapolating local examples to a pluton wide scale. Present topography allows us to examine vertical thicknesses of a few hundred to 3000 meters. Since true vertical thicknesses are uncertain, volumes are difficult to determine: we therefore use the total number of objects, surface areas and lengths of internal contacts in the largely 2D exposure of the TIC as "reference frames" in this discussion. Our goal is a first order estimate of the numbers and spatial scales of the various examples of erosion and recycling discussed above.

The TIC is ~60-70 kilometers long and anywhere from 10 to 35 kilometers wide with a total surface area ~1,100 km^2. Preserved surface areas of TIC units indicate that the Kuna Crest makes up 10%, the Half Dome 27%, the Cathedral Peak 58% and all leucogranites combined about 4% of the total area. If our interpretation that inner parts of the Kuna Crest and Half Dome units have been partially removed is correct, it is interesting to estimate the potential size of removed material. We have done so by using present exposures, evidence of truncated margins, location of cognate inclusions, zircon age distributions and guesses of the likely minimum and maximum extents of these units prior to intrusion of subsequent units (Fig. 14). We assume that the external margin of the Cathedral Peak and late leucogranites have not been changed because large, younger magma batches are not present. Throughout the entire growth history of the TIC, the minimum estimate of total removed area is ~595 km^2, which is about 35% of the likely minimum original size

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of the TIC. The maximum estimate is ~1295 km^2, which is about 54% of original material removed from the TIC. If we assume a minimum 5 kilometer thickness over which removal of plutonic material occurred, then a minimum of 2955 km^3 and maximum of 11,975 km^3 of material was moved out of this 5 km section of the crust and/or recycled into younger units during the 10 m.y. lifespan of TIC. Internal (i.e., within one magma unit), repeated erosion/recycling events might increase these numbers or simply reflect local recycling. The above estimates can be evaluated if we imagine that over 10 m.y. magmatic material was either removed by one or more volcanic eruptions, recycled into younger batches or by downward movement to deeper crustal levels in the magmatic system as younger batches ascended (Fig. 16). Volumes of volcanic eruptions in continental arcs range from 0.1 km^3 to more than 1000 km^3 over average durations of ~50 days (e.g, Simkin and Siebert, 1994). Thus three to 10 large eruptions, or a greater number of smaller volume eruptions could accommodate all of the removed materials. Total eruptive volumes in volcanic fields of similar dimensions and durations to the TIC range from 100s to 1000s of km^3 of volcanic materials (Lipman 2008, Grunder et al., 2008). If some eroded material was recycled into younger magma batches, this would reduce the amounts of “removed” magma or plutonic material: below we estimate that potential magnitudes of recycled materials may range up to 25%. The magnitude of downward movement of magmatic/plutonic material is the least constrained.

Figure 1 displays the length of the main contacts in the TIC that are sharp at the meter scale, gradational over 10-100 meter scales, or are associated with hybrid

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zones. The contact between Half Dome and Kuna Crest units is ~95 kilometers of which ~17% is sharp, 58% is gradational, and ~24% is associated with mappable hybrid units. The Cathedral Peak contact against older units is ~115 kilometers long, 50 kilometers of which forms discordant contacts with metamorphic host rocks. Of the remaining 65 kilometers (mostly against Half Dome units) 58% is sharp, 30% gradational and 12% is associated with hybrid units. The late leucogranites, although volumetrically small, have >>100 kilometers of contacts because of their thin elongate shapes. Greater than 70% of these contacts are sharp and the remaining are gradational. Clearly the Cathedral Peak contact is typically much more discordant than the Kuna Crest-Half Dome contacts. Thus field observations and geochronology indicate that the Cathedral Peak contact is typically an erosional contact with an age gap. The Kuna Crest-Half Dome contact, although mostly gradational or associated with hybrid zones, still preserves good evidence for truncation and erosion. But presumably the eroding magma batches and crystal mushes being eroded had enough melt to continue to interact both physically and chemically. Bergantz et al. (2015) recently modeled such behavior showing that crystal-rich systems can simultaneously behave as a "soft" viscoplastic material while failing in a Mohr-Coulomb fashion, thus forming linear (sharp) and diffusive (gradational) contacts during the same intrusive event.

Schlieren-bound troughs provide one example of local internal contacts associated with erosion in the TIC. We have measured >1000 troughs in local domains where detailed mapping was completed. If a conservative estimate of the spatial

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distribution of troughs (number/km^2) is extrapolated to entire units, we estimate a minimum number of >9000 troughs for the entire TIC. Contacts in these troughs range from 10s of centimeters to 100s meters in length: if we use a conservative average of 5 meter length/trough, then there is a minimum of 45 kilometers of internal trough contacts. Troughs tend to cluster (Memeti et al., 2014; Stanback et al., 2016). If we assume that one "event" causes 9 clustered troughs, then we need a minimum of 1000 "local events" to form the troughs in the TIC. We have also mapped 1-10 km-scale zones of fairly planar schlieren (Fig. 1) with local schlieren truncations that may be a form of broad trough or "magma mush avalanche" surfaces. Our estimates are that schlieren in these planar schlieren zones define around 17,000 kilometers of internal contacts and would require many 1000's of events to form. In spite of these impressive lengths, if we assume that schlieren bands are around 5 centimeter thick on average, then all of the planar schlieren only make up about 0.85 km^2 in area or less than 1% of the total area of the TIC. Thus they represent a spectacular record of repeated erosion/recycling events in an active magma chamber but only have a minor effect on magma compositions.

It is equally intriguing to estimate the number of enclaves (Fig. 11), cognate inclusions (Fig. 7-11), and mixed minerals in the TIC and their scales of distribution. Using a conservative estimate established from field measurements of ~0.1 enclave/m^2, there are greater than 70 million exposed enclaves. Even if we drastically reduce this estimate, there are clearly millions of enclaves widely distributed throughout all three units. An estimate for the average area of a single

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enclave is ~ 50 cm^2. The calculated total area of enclaves for the exposed TIC is about 0.0365 km^2 much less than 1% of the total area: thus the distribution of enclaves requires the formation and dispersal of a huge number of objects, although their total volume is small.

To date we have found evidence of mineral-scale mixing in almost every sample analyzed for ages or geochemistry. So far the only potential exceptions occur in the southern Half Dome lobe and in some late leucogranites. To exemplify the numbers of crystals involved we can make a preliminary estimate using zircons. We typically get >1000 zircons in separates from ~1000 cm^3 samples. Assuming that ~10 zircons occurred in any 100 cm^2 surface cut in these samples, this number scaled to the entire TIC surface area, would imply around a trillion zircons. Memeti et al. (2014) noted that the number of antecrystic zircons increase in younger units and statistically make up around 25% of the total zircons in the inner Half Dome and Cathedral Peak units (Fig. 2a). This would imply that a minimum of 250 billion antecrystic zircons occur in the present surface exposure of the Half Dome and Cathedral Peak units. Even if this estimate is off by 2 orders of magnitude, there must be billions of antecrystic zircons mixed into the Half Dome and Cathedral Peak units.

All of the above measurements draw attention to the scale of the processes involved in erosion/recycling in the TIC. These processes likely involved erosion along 100s to 1000s of kilometers of contacts, during 1000s of “events” and involving 1000s of

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cognate blocks, millions of enclaves and billions of antecrystic crystals, resulting in ~35-55% of material eroded from the present surface level, some (~25%? based on zircon estimates) recycled into younger batches and the majority displaced either up or down relative to the present surface.

Location of erosion/recycling: Where did this magmatic erosion and mixing/recycling occur? It is certainly possible that some occurred either at the site of magma generation in the lower crust/upper mantle or during magma ascent, although both of these would require that the source rocks and/or the units bordering the ascent path shared identical age and similar geochemical characteristics with the older TIC units exposed at the present crustal level. It is also possible that some recycling occurred at higher crustal levels and subsequently sank to the present levels. This would require a vertically connected magma plumbing system in a rheological state that would permit such sinking.

However, several observations suggest that a fair amount of erosion and recycling occurred at the presently exposed emplacement site. First, at the scale of 10s of kilometers, the presence of hybrid zones along the main internal contacts and truncations of units along these contacts seem to require, at least in part, in situ formation. The large scale, the consistent position between the two units with which the hybrid unit shares characteristics, and gradational contacts with these units would make it remarkably fortuitous if separate batches of magma were sequentially emplaced in the correct spatial order, or if the hybrid units formed

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elsewhere and were transported to their present location as a package. We think it is more likely that the hybrid units represent in situ mixing between the adjacent units.

Second, Memeti et al. (2010) pointed out that the older, more isotopically primitive, shorter-lived southern lobes typically show significantly less evidence for mixing than similar units in the main plutonic body. If all recycling/mixing occurred during magma generation and ascent, then there is no reason that mixing should not be equally prevalent in these lobes.

Another intriguing, although less conclusive observation, concerns the composition of rocks making up local sheeted zones. For example, in both the Fletcher Peak (Fig. 5) and Sawmill Canyon areas (Fig. 6), the zones of sheeting and mingling were partly formed from porphyritic Half Dome magmas, which is what would normally be the next innermost unit. But in both cases the porphyritic Half Dome (Fletcher Peak) or entire Half Dome (Sawmill Canyon) is entirely missing and instead younger units are now present. Thus one likely implication is that when these sheeted zones were forming, Half Dome magmas were present along the interior edges of these zones and were removed during intrusion of younger magma. The presence of Half Dome blocks and Half Dome aged zircons in these younger magmas supports this interpretation and, if correct, the implication that large volumes of plutonic material were eroded and removed.

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At the meter to 10 meter scale, the preservation of numerous schlieren-bound magmatic structures, often with similar structural patterns (e.g., orientations, younging directions, migration directions) in the same spatial domains, also argue for in situ formation.

It is less certain where the dispersal of enclaves, xenoliths and cognate inclusions occurred, although the concentrations of these objects into "enclave swarms" likely occurred in situ. Many enclave swarms are one-half to a few meters wide and 10s of meters long, and like the magmatic structures, it seems highly unlikely that they moved any significant distance without being disrupted. We suspect that a great percentage of the millions of enclaves were formed and dispersed at significantly deeper crustal levels where conditions more favorable for widespread dispersal of mafic magmas exist, Preservation of disrupted and mingled mafic dikes occur in the TIC (Memeti and Paterson, 2007), but examples are widely distributed at the emplacement site (Fig. 11). Arguments about the origins of the cognate inclusions are the same as presented below for zircons in that it is uncertain where they formed but they either formed near the present emplacement level or their presence has implications for the types of materials being recycled from other locations. Finally, it is intriguing that the densities of many of these xenoliths and cognate inclusions are greater than the densities of the surrounding plutonic rocks, which were even lower densities when magma, implying that enclaves, most cognate inclusions, and mafic structures had negative buoyancies. This implies that

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the combination of host magma viscosities and ascent rates must have been sufficient to bring the enclaves and xenoliths to their present sites and trap them.

The presence of antecrystic zircons in inner units implies either (1) zircons with identical ages occur in the magma source region, (2) zircons with identical ages occur in units making up the "host rock walls" that the magma ascended through, or (3) zircon was recycled at the emplacement site. The latter implies that more than one recycling event is needed to transport Kuna Crest zircons into the central parts of the Cathedral Peak. In spite of extensive TIMS and LA-ICP-MS dating of zircon cores and rims, statistically older zircon core ages have not been found (Memeti et al., 2010). Zircon saturation temperatures range from $\sim$780 to $\sim$710 ^{o}C (e.g., Miller et al., 2007), and these temperatures plus preserved antecrystic zircons suggest that zircon growth did not start early in the magma history. The present vertical thickness of the TIC is unknown: but the above observations suggest to us that most zircon mixing did not begin until magma ascent in the middle crust and during upper crustal emplacement.

Finally, although much more extensive single mineral geochemistry is needed, the presence of antecrystic zircons and preliminary observations show direct evidence that mixing of crystal populations is widespread (Krause et al., 2010; Memeti et al., 2014; Barnes et al., 2016). However where this mixing occurred is uncertain. The convergence of REE values in the rims of K-feldspar crystals suggests that at least the final mixing of these crystal populations occurred in hybrid melts that may be

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preserved in the plutonic material at the present emplacement site. However, this hypothesis needs further testing (e.g., Barnes et al., 2016).

The above discussion suggests to us that some mixing of enclaves and crystals probably occurred by erosion/recycling/mixing during ascent but that a significant amount occurred at the present emplacement site. The preserved truncations, hybrid zones, local magmatic structures, cognate blocks, enclave and crystal mixing, and local enclave swarms and schlieren layers at the emplacement site indicates that these erosion and recycling/mixing processes occurred over an impressive range of scales during pluton growth. The likely result was removal of at least 35% of previous plutonic material from the present level of exposure and recycling of at least a component of this removed material into younger magmas.

Size and number of TIC magma chambers: Given the well established 10 m.y. growth history of the TIC, its irregular shape, and clear evidence of incremental growth, it is expected that at least several, and potentially many, ephemeral magma chambers existed (Fig. 14). All researchers would likely agree that individual batches formed at least short-lived magma chambers upon arrival. But there remains a great deal of debate about whether larger and longer lived magma chambers formed either because arriving batches were larger or that the rate of arrival of smaller batches was fast enough to allow coalescence into larger magma chambers (e.g., Michaut and Jaupart, 2006; Paterson et al., 2011). Our conclusion about the extent of

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erosion/recycling supports the presence of large and active magma chambers and here we explore evidence for the number and size of potential magma chambers.

Memeti et al. (2010) used a combination of map patterns, TIMS U/Pb zircon ages, thermal modeling, magmatic fabric patterns, and bulk-rock compositional data to argue that each of the four TIC lobes at one time consisted of magma chambers roughly the size of the inner portions of the lobes (Fig. 14) and with hypersolidus durations of ~0.4 to 2.3 m.y. By comparing locations of steep gradients in a histogram plot of all TIMS single zircon ages from the TIC to the size of map regions with similar zircon ages, Memeti et al. (2014) also argued that large magma chambers grew in the main plutonic body at ~94 Ma (Kuna Crest magmas), 90.5 Ma (Half Dome magmas), and 88.5 Ma (Cathedral Peak magmas). If we link map patterns, contoured age distributions (e.g. Fig 1, 2 and Memeti et al., 2014), and the age spread of autocrystic zircons in dated samples, these data suggest that these latter magma chambers at one point were at least 100s km^2 in size with local hypersolidus durations of ½ to 1.5 m.y. and total hypersolidus histories of 1.5 to 2.5 m.y (Fig. 14). The Kuna Crest and Half Dome magma chambers would have been centered in the SW to S part of the TIC (Fig. 14) whereas the Cathedral Peak magma chamber initially formed in the southern center of the TIC and then migrated northwards over several million years. Paterson et al. (2011) presented thermal models of the incremental growth of the TIC, which suggested that new batches arrived (assuming zircon crystallization ages give an approximate proxy for relative magma batch arrival times) in the locations where temperatures remained the

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highest and presumably where the final crystallization of the previous magma chambers occurred. Both the zircon age patterns and the broad areas of similar biotite cooling ages of 91-90 Ma in the southern TIC and 88-86 Ma everywhere around the CP support this hypothesis (Fig. 1, 2).

Memeti et al. (2014) noted that in their zircon age cumulative probability curve, the distinct peaks at older ages (90 and 91 Ma) are due to the occurrence of antecrysts in younger units (particularly the Cathedral Peak), which is consistent with former larger magma batches of these ages being partially assimilated into the Cathedral Peak since these peaks "represent an overrepresentation of crystals of this age compared to the relatively small observable magma volume of the equigranular Half Dome granodiorite unit." The evidence for erosion/recycling presented in this paper (e.g., cognate inclusions, "inner" magmatic material in sheeted zones, antecrystic zircons) supports this idea that the size of the original Kuna Crest and Half Dome magma chambers were significantly bigger than their present map patterns since their inner, crystal-poor parts are now partially removed or recycled (Fig. 14). This would not be the case for the Cathedral Peak and late leucogranites. Our estimates indicates that these magma chambers at one time reached areas > 500 km^2.

The estimates of removed/recycled materials in the TIC have implications for calculating pluton-scale "tempos" of magmatism. Paterson et al. (2011) and Memeti et al. (2014, Figure 4-4) estimated volume addition rates during the 10 m.y. growth of the TIC. We have revised these suggested temporal histories to include the

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minimum estimate of removed materials for older plutonic units (Fig. 14, 15). Our new estimate of the growth of the TIC (Fig. 15) suggests that the original volumes of the earlier Kuna Crest and Half Dome magma bodies were as large or larger than the Cathedral Peak body. Greater volume input of magma occurred every 3-4 m.y. If we assume (based on the antecrystic to autocrystic zircon ratio in younger units), that ~25% of the removed material is recycled into younger units, this would reduce the estimated additions for the Half Dome and Cathedral Peak magmas by around 5% on Figure 15.

A couple of issues remain unclear. Above we conclude that small magma chambers formed in the lobes and larger magma chambers formed in the main body of the TIC. It is not clear if the two magma chambers in the southern lobes were ever connected to those of similar age in the main body long enough to chemically and physically interact. Memeti et al. (2010) argue that the two northern lobes formed from magmas derived from one of the main magma chambers and the gradual northward younging age pattern supports this. Thus our working hypothesis is that there were two fairly disconnected magma chambers in southern lobes, at least three larger magma chambers in the main body that, before removal of their inner parts, reached sizes of >500 km^2 and that the locus of newly arriving magmas and/or of the remaining magma chambers migrated northward. The final large magma chamber occurred in the Cathedral Peak unit in the northeast corner of the TIC based on the young zircon and biotite ages and fairly evolved compositions in this region.

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It is also unclear to what degree the large magma chambers in the main body grew from rapidly arriving smaller batches with very similar compositions, or from large batches (that may or may not have formed from amalgamation of smaller batches at depth). Contacts between potential batches that make up these former large magma chambers are not preserved. Detailed mapping and studies of single mineral compositional and zoning histories will be needed to fully evaluate this issue.

Processes driving erosion/recycling: If the above conclusions are correct, then a large amount of magmatic erosion and remixing occurred at upper-crustal depths during growth and final hypersolidus evolution of the TIC. Although Glazner (2014) argued that such magmatic erosion and mixing should rarely occur since erosion requires turbulent flows with high Reynolds numbers (Re), there are abundant examples both in nature and in elegant models of such erosion and remixing in magmatic systems (Ruprecht et al., 2008; Burgisser and Bergantz, 2011; Barboni and Schoene, 2014; Bergantz et al., 2015). Three scenarios are worth mentioning: (1) In rare cases processes such as volatile increase may raise Re numbers to moderate values; (2) even at low Reynolds numbers there are processes that still operate to aid erosion and movement of materials; and (3) there are processes that can lead to erosion and mixing that do not rely on surface tractions or turbulent flow. Studies by Bachmann and Bergantz (2006), Michaut and Jaupart (2006), Ruprecht et al. (2008), Burgisser et al. (2011), Klemetti and Clynne (2014) and Cooper and Kent (2014) among others provide examples of processes driving the remobilization of magmas. Where gas bubbles provide the driving force, Re numbers can increase to

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about ~100-400. However, even in models of magmatic systems with low to moderate Re numbers (e.g., 1-10), nonlinear terms remain important (Burgisser et al., 2005; Del Gaudio and Ventura, 2008; Burgisser and Bergantz, 2011; Dufek, pers. comm., 2014). Plus the increasing melt viscosities and low Stokes number of crystals in magmas with low Re numbers imply that the particles will follow any fluid motion fairly closely as drag coefficients go up (Burgisser et al., 2005; Burgisser and Bergantz, 2011; Dufek, person comm., 2014).

Equally important are processes in magmatic systems that can trigger erosion and remixing without the requirement of being driven by turbulent flows. For example, Bergantz et. al. (2015) and Schleicher et al. (2016) showed models of one magma batch intruding another during the temporal evolution of magma chambers that result in extensive visco-plastic collapse of older materials and mixing into younger batches. Such a process can drive convection (Martin et al., 1987) and downward flow of the older magma in part to make space for the newly arriving magma (Fig. 16). Marsh (1996, 2006, 2013, 2015) has published a series of papers that explore the growth of solidification fronts in cooling magma chambers. If cooling is faster at the roof or upper levels of magma chambers, then these solidification fronts will form over-steepened and even overhanging zones of denser crystal mushes (Fig. 16) that are unstable and prone to collapse and avalanching into the hotter centers or onto rheological boundaries or floors of magma chambers (Žák and Paterson, 2010). Removal of magma and remixing may also occur during volcanic eruptions (Bacon and Lowenstern, 2005; Barboni and Schoene, 2014) or collapse of unstable steep

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margins between magma batches (Bergantz, 2000). Furthermore, the negative ΔV during cooling of these crystal mush zones results in contraction and tearing, which can lead to erosion and movement along magmatic faults (Paterson, 2009; Humphreys and Holness, 2010). Synmagmatic tectonic straining of magma chambers may trigger faulting, erosion, and collapse of materials. Finally, the complex thermal and strain gradients within intrusive complexes can locally drive stoping and erosion that may remove and recycle older material into younger batches (e.g., Del Gaudio and Ventura, 2008; Paterson et al., 2011) and may cause movement of local magma batches resulting in a wide variety of small-scale magmatic structures (Paterson, 2009).

Figure 16 displays a cartoon of an idealized magma chamber with (1) growing solidification fronts that are locally collapsing and in turn driving additional magmatic erosion, mixing and redeposition, (2) arriving magma batches driving convective stirring, return flow and mixing, and (3) a variety of resulting truncation surfaces, displacement of crystal mushes, remixing, and the formation of local magmatic structures. Observations presented above suggest that in the TIC 1000s of such events, or a smaller number of large events resulting in 100s of subsidiary localized events, occurred in a number of active magma chambers that rapidly evolved in space and time during the growth of this batholith (Fig. 14, 15, 16).

Processes during growth of the TIC

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The upper-crustal Tuolumne Intrusive Complex grew incrementally over 10 m.y. by a complex set of processes operating over a range of temporal and spatial scales. Growth began with the emplacement of numerous, but volumetrically small magma batches forming sheets of Kuna Crest magmas (Hardee, 1982; Memeti et al. 2010, 2014) followed by three periods of much larger magma additions, or pluton-scale flare-ups, at approximately 94-93, 91-90 and 88-87 Ma either by numerous chemically similar small batches or fewer but larger batches. During these flare-ups, smaller magma chambers (up to 40 km^2) formed in the lobes and larger magma chambers (up to >500 km^2) formed in the main part of the intrusive complex. Irrespective of magma chamber size, repeated, multiscale, magmatic erosion events of older intrusive units occurred in these magma chambers driven by a complex interplay of buoyancy-driven intrusion of younger magma into older crystal mushes, collapse and avalanching along growing solidification fronts, both broader and local convection driven by internal thermal, compositional, and rheological gradients (Martin et al., 1987) and by tectonic strains imposed on the magma chambers. Certainly some of the implied erosion and mixing occurred at greater depths resulting in the arriving magmas already being pre-mixed. The 1000s of "erosion and recycling events" at the emplacement site resulted in removal of ~35-55% of the original plutonic material in the presently exposed surface with a small portion being recycled into younger magma batches.

Implications: The evidence that active magma chambers, some > 500 km^2, can form at upper crustal levels have a number of implications for generating the magma

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volumes and characteristics recorded by erupted volcanic materials (Bachmann et al., 2005; Charlier et al., 2007; Barboni and Schoene, 2014). These observations support models in which convection, mixing, and varied internal physical and chemical gradients form significant compositional diversity at upper crustal levels. The extensive mixing recorded in the TIC brings into question the interpretation of whole rock elemental and isotopic data and certainly indicates that the characteristics of parental magmas can be significantly modified at shallow crustal levels. The extensive mixing and recycling present challenges for evaluating the magnitude of fractionation and original melt compositions. Finally the suggested removal of 35-55% of previous magmatic units in plutons will dramatically alter volume estimates at arc scales of magma added to crustal columns and of the volumes of magma migrating through individual magmatic systems. Both regional and local estimates of magma additions will influence our understanding of the thermal and geochemical evolution of the growth of plutons, whereas the former may dramatically change estimates of magma additions during flare-ups and the resulting crustal growth (e.g., Paterson and Ducea, 2015, Roberts et al., 2015). Accurate estimates of magma addition rates are also necessary if accurate estimates of crustal thickness, isostatic response, surface elevation and erosion rates of the arc are to be determined (e.g., Lee et al., 2015).

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Figure Captions

Figure 1: Geologic map of the Tuolumne Intrusive Complex showing the main compositional units and U/Pb zircon ages (after Memeti et al., 2010) and the styles of the main internal contacts. Red line shows geochronologic transect shown in Figure 2b. Boxes show locations of detailed maps in Figs 3-6.

Figure 2: (A) U/Pb CA-TIMS single zircon ages from sample KCL.536 (Kuna Crest Lobe) representative of outer TIC units as well as the southern lobes and sample CTP.11 (Cathedral Peak sample) representative of inner TIC units. Outer units display uniform clusters of autocrystic zircons with rare xenocrysts or antecrysts, whereas inner units contain zircons that are significantly older than the predominant zircon population and have ages that are similar to the ages of the older parts of the TIC (Memeti et al., 2010, 2014) (B) $^{40}Ar/^{39}Ar$ biotite ages from coarse and fine-grained biotites along a southern transect (red line in Fig. 1) after Matzel et al. (2005). Note the step in ages between two more gently dipping sections of the curve, the latter indicating large regions of similar biotite cooling ages in the Half Dome granodiorites. These cooling ages match the U/Pb zircon crystallization age of the Half Dome and Cathedral Peak magmas, respectively.

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Figure 3: Geologic map of the Glen Aulin area along the western edge of the TIC (Fig. 1). Mapping completed by Vali Memeti, Bob Miller, Scott Paterson and Jiří Žák. Contact symbols same as in Fig. 1. Note the discontinuous hybrid zone between the Glen Aulin tonalite (Kuna Crest) and equigranular Half Dome granodiorite and truncations of older by younger magmatic units. Locally the Cathedral Peak granite cuts across Half Dome units and directly intrudes the Glen Aulin tonalite. PCT = Pacific Crest Trail. Mapping in Figures 3, 4, 5, 6 all completed at 1:10,000 to 1:5,000 scales.

Figure 4: Geologic map of the Kuna Crest lobe (Fig. 1) after Memeti et al. (2010). Note truncation of compositionally defined zones KC-l and KC-ll at the northwest end of the lobe by east-west striking Kuna Crest units in the main magma chamber. Also note the hybrid unit (tHD = Kuna Crest-Half Dome mix) in the center of the lobe.

Figure 5: Geologic map of the Fletcher Peak area at the southern end of the TIC (Fig. 1) redrafted after Žák et al. (2005). Contact symbols same as in Fig. 1. The east-west striking Kuna Crest unit is probably connected to the east-west striking KK-lll unit in Figure 4. In the Fletcher Peak area this unit is bordered and locally truncated by a complex hybrid zone and in turn bordered by the porphyritic Half Dome. Except locally within the hybrid zone, the equigranular Half Dome unit is absent here.

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Figure 6: (A) Geologic map of part of the eastern margin of the TIC west of Saddlebag Lake (Fig. 1) and (B) the Sawmill Canyon area (mapped at 1:5,000 scale after Paterson et al., 2009). Mapping completed by Scott Paterson and Jiří Žák. Contact symbols are the same as in Fig. 1. In (A) note that the Kuna Crest unit does not extend north of the Sawmill Canyon area and only two pieces of the Half Dome unit are found north of Sawmill Canyon. In (B) complex, sheeted, east-west striking hybrids of porphyritic Half Dome and Cathedral Peak magmas cut across and truncate north-south striking equigranular Half Dome and Kuna Crest units. These hybrid units are truncated along their western ends by a hybrid, locally hornblende-bearing Cathedral Peak unit.

Figure 7: Photos of recycled materials in the Kuna Crest units. Figure a, shows an example of Paleozoic metasedimentary stoped blocks; Figure b shows older plutonic, volcanic and mafic enclaves in a large mafic enclave now in a Kuna Crest matrix; Figure c, d, and e show “composite” blocks of Jurassic volcanic blocks (c, e) and calc-silicate metasediments (d), respectively surrounded by mafic Kuna Crest tonalite in more felsic Kuna Crest granodiorite. Figure f displays numerous cognate inclusions, mafic enclaves and rare xenoliths in a hybrid Kuna Crest granodiorite.

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Figure 8: Photos of recycled materials in Half Dome units. Figures a and b, show examples of metavolcanic and metasedimentary xenoliths in Half Dome granodiorite. Figures c and d show composite cognate inclusions where former enclave swarms with local xenoliths are surrounded by a matrix of Kuna Crest magmas and now reside in equigranular Half Dome granodiorite. Note the sharp boundaries defining the rectangular shapes that cut and truncates individual enclaves indicating that a former more elongate enclave swarm (e.g., Tobisch et al., 1997) was broken apart into rectangular pieces presumably during movement of the Half Dome magma. Figure e displays a composite block of volcanic rock locally attached to and intruded by Kuna Crest magma now surrounded by a Half Dome matrix. Figure f shows a collection of xenoliths, a large cognate inclusion of layered plutonic rock and enclaves near the Kuna Crest – Half Dome contact.

Figure 9 shows examples of inclusions residing in the Cathedral Peak granodiorite. Figure a shows a folded metavolcanic xenolith, Figure b a cognate inclusion of Half Dome granodiorite, Figure c a layered cognate inclusion of porphyritic Half Dome magma and Figure d a composite inclusion of layered plutonic rock rimmed by Half Dome magma in the Cathedral Peak granodiorite. Figure e shows a schlieren-bound tube sharply truncated by Cathedral Peak magma and f shows an example of a composite inclusions with a small piece of layered

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plutonic rock in a matrix of layered Half Dome granodiorite (HD), now entirely surrounded by Cathedral Peak granodiorite.

Figure 10 shows examples of various pieces of the Cathedral Peak granodiorite in the Johnson granite. Figure a shows Cathedral Peak K-feldspar megacryst antecrysts, Figure b an angular piece of Cathedral Peak granodiorite, and Figures c and d rounded (partially melted?) pieces of Cathedral Peak granodiorite, each with one large K-feldspar megacryst. Photos c and d taken by Laura Bracciali (see also Bracciali et al., 2008).

Figure 11: Photos of enclaves and enclave swarms. Figure a is a mafic quartz diorite dike intruding Kuna Crest granodiorite that is the same composition of many enclaves in the Kuna Crest. Figure b a similar dike to Figure a, but now broken apart and forming enclaves. Figure c depicts dispersed elliptical enclaves in the Half Dome granodiorite. Figure d shows an enclave swarm with local cognate inclusions and xenoliths. Figure e shows a composite enclave with a finer grained mafic center, rimmed by a medium grained granodiorite, now residing in Kuna Crest granodiorite. Figure f depicts a cognate block of a former enclave swarm. Note sharp edge to the block that cuts across individual enclaves.

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Figure 12: Photos of truncated and likely eroded/reintruded and partially recycled magmatic structures. Figures a and b show repeated mafic schlieren defining two sets of Cathedral Peak magmatic troughs with examples of sharp truncations of older schlieren layers. These truncations are visible at the centimeter scale and sharply cut across both mafic and felsic layers as well as subparallel magmatic mineral foliations with no deflection of the older layers. A new magmatic foliation is subparallel to the new, cross-cutting schlieren layer. Figure c shows a package of mafic-felsic schlieren layers of Half Dome granodiorite that show excellent grading of mafic minerals and are sharply truncated by younger batches of magma. Figure d shows schlieren in porphyritic Half Dome granodiorite sharply truncated by younger schlieren layers. Figure e shows an example of a “bifurcated” schlieren-bound migrating tube(s) that show repeated truncations of older tube margins by younger margins and now surrounded by and locally reintruded by Cathedral Peak granodiorite. Figure f displays an example in Cathedral Peak granodiorite of planar schlieren and subparallel magmatic foliation truncated by, and presumably removed during formation of a schlieren bound trough in which a new schlieren parallel foliation is formed.

Figure 13: Photos of likely recycled antecrystic minerals. Figures a and b are former large euhedral hornblendes, typical of the Half Dome units, now replaced by biotite and surrounded by Cathedral Peak granodiorite, Figures c and d show

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large, zoned, isolated K-feldspar megacrysts, some of which have been interpreted to form in porphyritic Half Dome magmas and are now recycled into Cathedral Peak granodiorite.

Figure 14: Cartoon showing our interpretation of the former minimum (stippled) and maximum (no pattern) extents (in yellow) of removed parts of the TIC for Kuna Crest (A) and Half Dome (B) units resulting in the present configuration (C). For each panel, black ellipses show estimated size, position and age of inferred Kuna Crest (A), Half Dome (B) and Cathedral Peak (C) magma chambers based on age distributions and the minimum extents of removed materials. See text for discussion of evidence of these magma chambers.

Figure 15: A comparison of magma additions (in km^2) versus age (95 – 85 Ma) in the TIC based on the extent of present-day exposures and inferred former extents of units prior to removal. If estimates of removed material are approximately correct, then the areas or volumes of Kuna Crest and Half Dome magmas were equal to or greater than the Cathedral Peak magmas. Either curve also implies that the magma addition rates to the TIC were episodic and that three prominent increases in magma additions occurred around 94-93 Ma, 91-90 Ma, and 88-87 Ma.

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Figure 16: Cartoon of an idealized magma chamber showing a number of potential processes that would lead to erosion of older units, mixing of older and younger crystal mushes, recycling of older materials into younger batches, and formation of local magmatic structures. See text for discussion.

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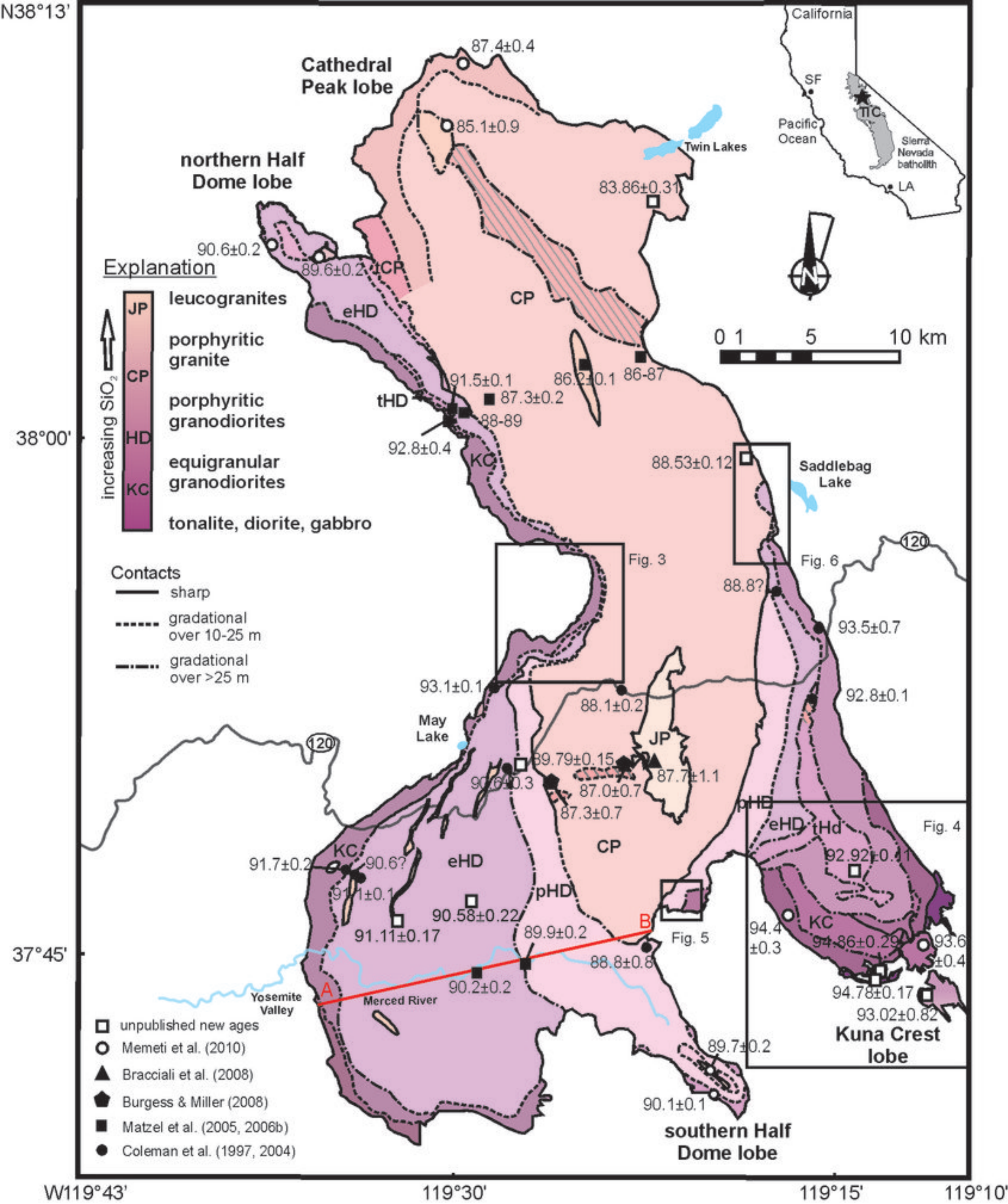
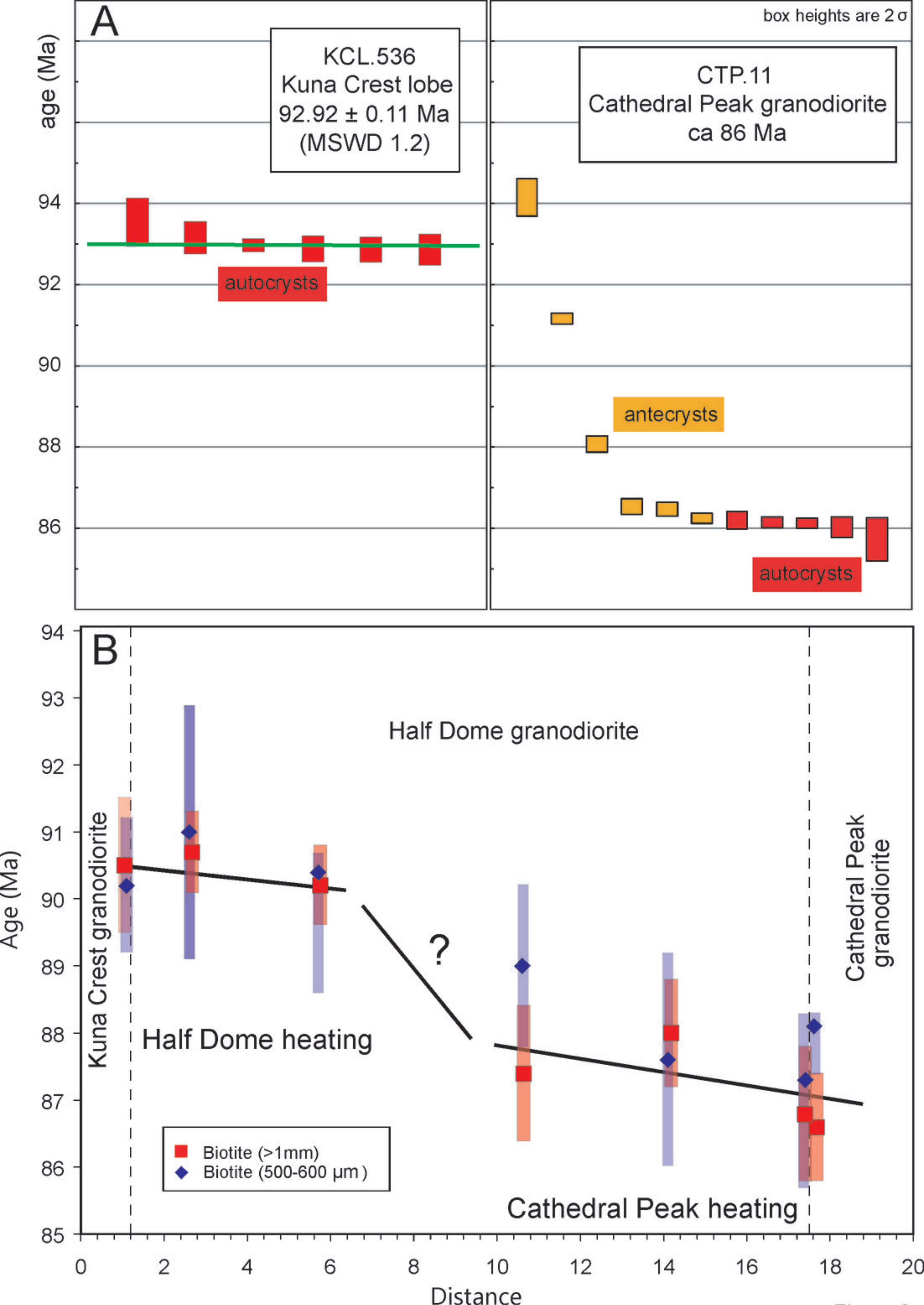


N38°13'
38°00'
37°45'
W119°43'
119°30'
119°15'
119°10'
Cathedral Peak lobe
northern Half Dome lobe
southern Half Dome lobe
Kuna Crest lobe
California
SF
Pacific Ocean
TIC
Sierra Nevada batholith
LA
Twin Lakes
Saddlebag Lake
May Lake
Yosemite Valley
Merced River
0 1 5 10 km
Explanation
increasing SiO2
JP
CP
HD
KC
leucogranites
porphyritic granite
porphyritic granodiorites
equigranular granodiorites
tonalite, diorite, gabbro
Contacts
sharp
gradational over 10-25 m
gradational over >25 m
unpublished new ages
Memeti et al. (2010)
Bracciali et al. (2008)
Burgess & Miller (2008)
Matzel et al. (2005, 2006b)
Coleman et al. (1997, 2004)
Fig. 3
Fig. 4
Fig. 5
Fig. 6
87.4±0.4
85.1±0.9
83.86±0.31
90.6±0.2
89.6±0.2
91.5±0.1
87.3±0.2
86.2±0.1
86-87
88-89
92.8±0.4
88.53±0.12
88.8?
93.5±0.7
92.8±0.1
93.1±0.1
88.1±0.2
89.79±0.15
90.6±0.3
87.0±0.7
87.7±1.1
87.3±0.7
91.7±0.2
90.6?
91.1±0.1
91.11±0.17
90.58±0.22
89.9±0.2
88.8±0.8
90.2±0.2
92.92±0.11
94.4 ±0.3
94.86±0.29
93.6 ±0.4
94.78±0.17
93.02±0.82
89.7±0.2
90.1±0.1
tCP
CP
eHD
tHD
KC
JP
pHD
tHd
A
B
120

Figure 1



A
box heights are 2 σ
age (Ma)
KCL.536
Kuna Crest lobe
92.92 ± 0.11 Ma
(MSWD 1.2)
CTP.11
Cathedral Peak granodiorite
ca 86 Ma
94
92
90
88
86
autocrysts
antecrysts
autocrysts
B
94
93
92
91
90
89
88
87
86
85
Half Dome granodiorite
Kuna Crest granodiorite
Cathedral Peak granodiorite
?
Half Dome heating
Cathedral Peak heating
Biotite (>1mm)
Biotite (500-600 μm)
Age (Ma)
0
2
4
6
8
10
12
14
16
18
20
Distance

Figure 2

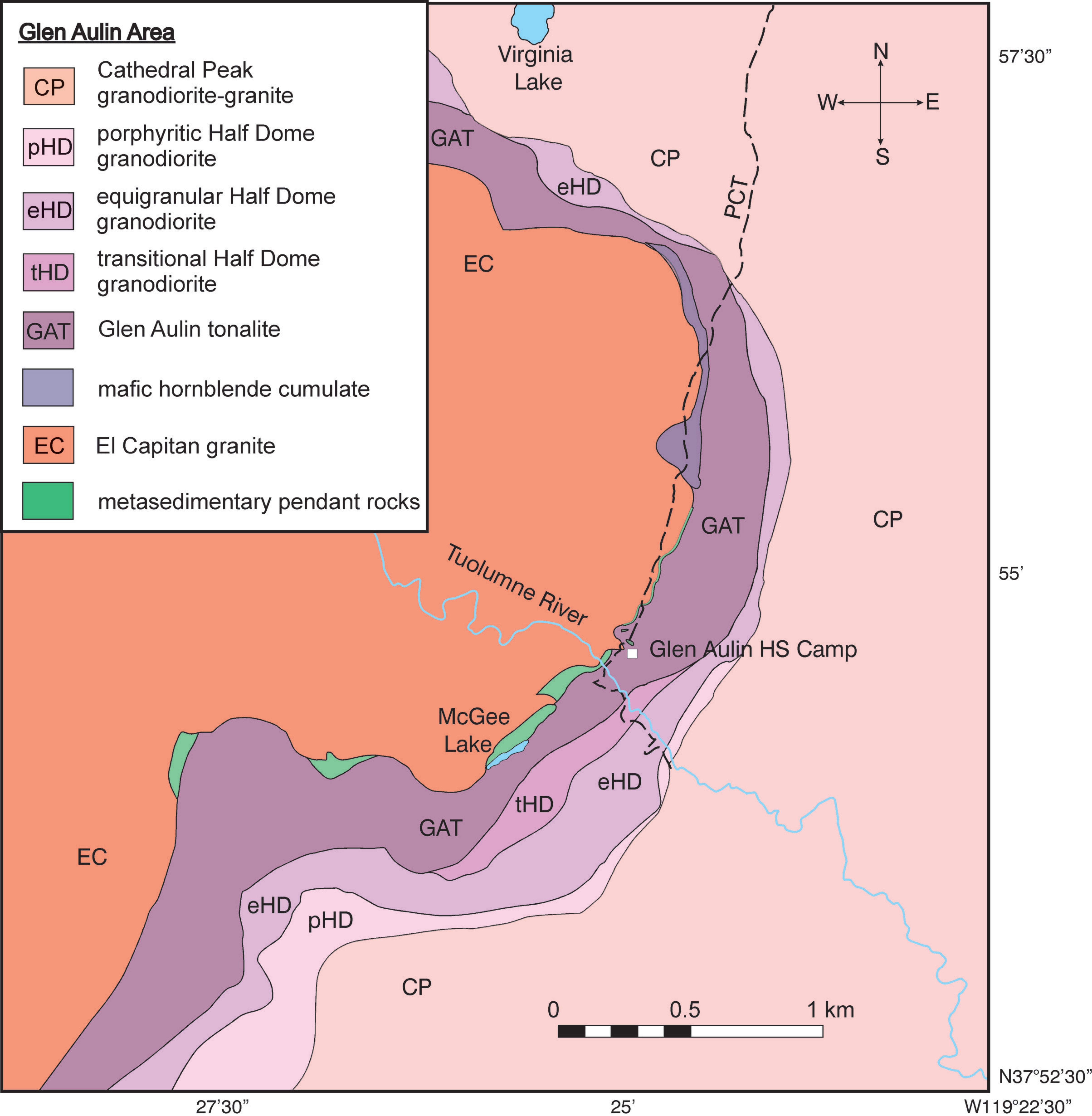

Glen Aulin Area
CP
Cathedral Peak granodiorite-granite
pHD
porphyritic Half Dome granodiorite
eHD
equigranular Half Dome granodiorite
tHD
transitional Half Dome granodiorite
GAT
Glen Aulin tonalite
mafic hornblende cumulate
EC
El Capitan granite
metasedimentary pendant rocks
Virginia Lake
N
W
E
S
57'30"
CP
GAT
eHD
PCT
EC
GAT
CP
55'
Tuolumne River
Glen Aulin HS Camp
McGee Lake
tHD
eHD
GAT
EC
eHD
pHD
CP
0
0.5
1 km
N37°52'30"
27'30"
25'
W119°22'30"

Figure 3

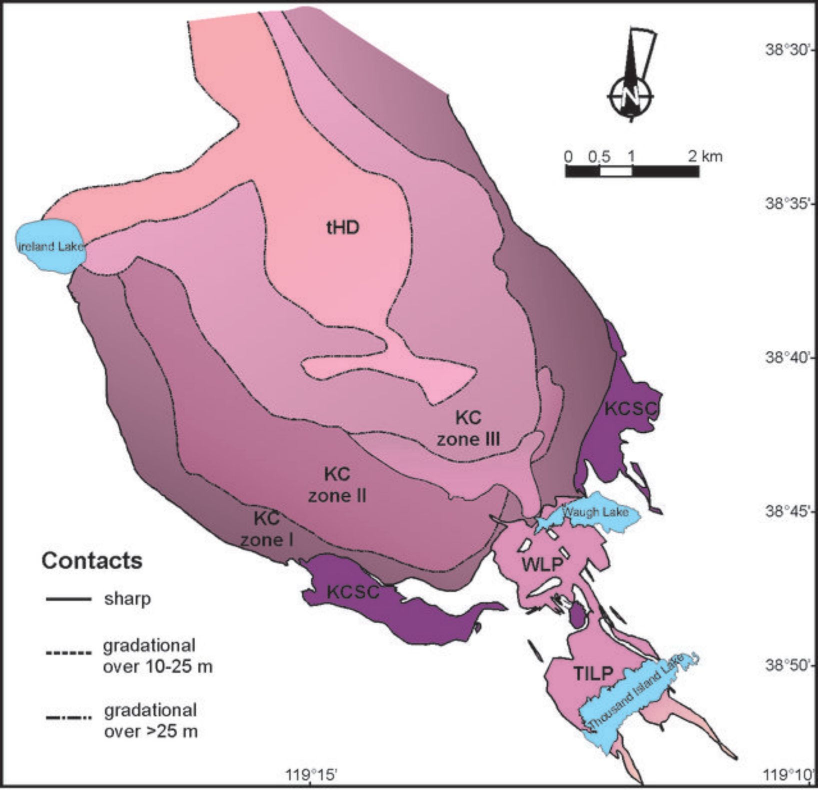
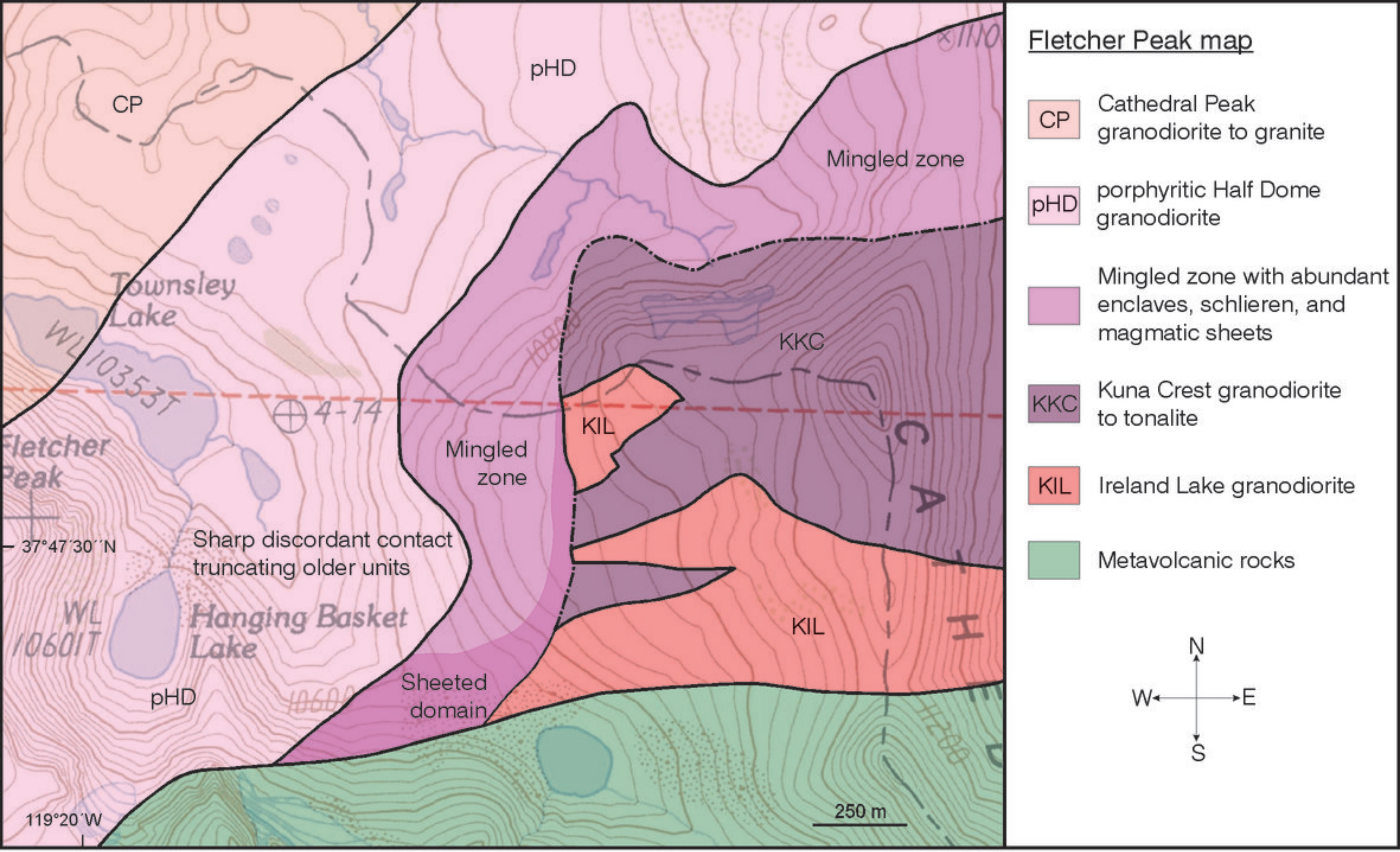

38°30'
38°35'
38°40'
38°45'
38°50'
119°15'
119°10'
0 0.5 1 2 km
N
tHD
Ireland Lake
KC
zone III
KC
zone II
KC
zone I
KCSC
KCSC
Waugh Lake
WLP
TILP
Thousand Island Lake
Contacts
sharp
gradational
over 10-25 m
gradational
over >25 m

Figure 4



Fletcher Peak map
CP
Cathedral Peak granodiorite to granite
pHD
porphyritic Half Dome granodiorite
Mingled zone with abundant enclaves, schlieren, and magmatic sheets
KKC
Kuna Crest granodiorite to tonalite
KIL
Ireland Lake granodiorite
Metavolcanic rocks
N
W
E
S
CP
pHD
Mingled zone
KKC
KIL
Mingled zone
Sharp discordant contact truncating older units
Sheeted domain
KIL
pHD
Townsley Lake
Fletcher Peak
Hanging Basket Lake
4-14
37°47´30´´N
119°20´W
250 m

Figure 5

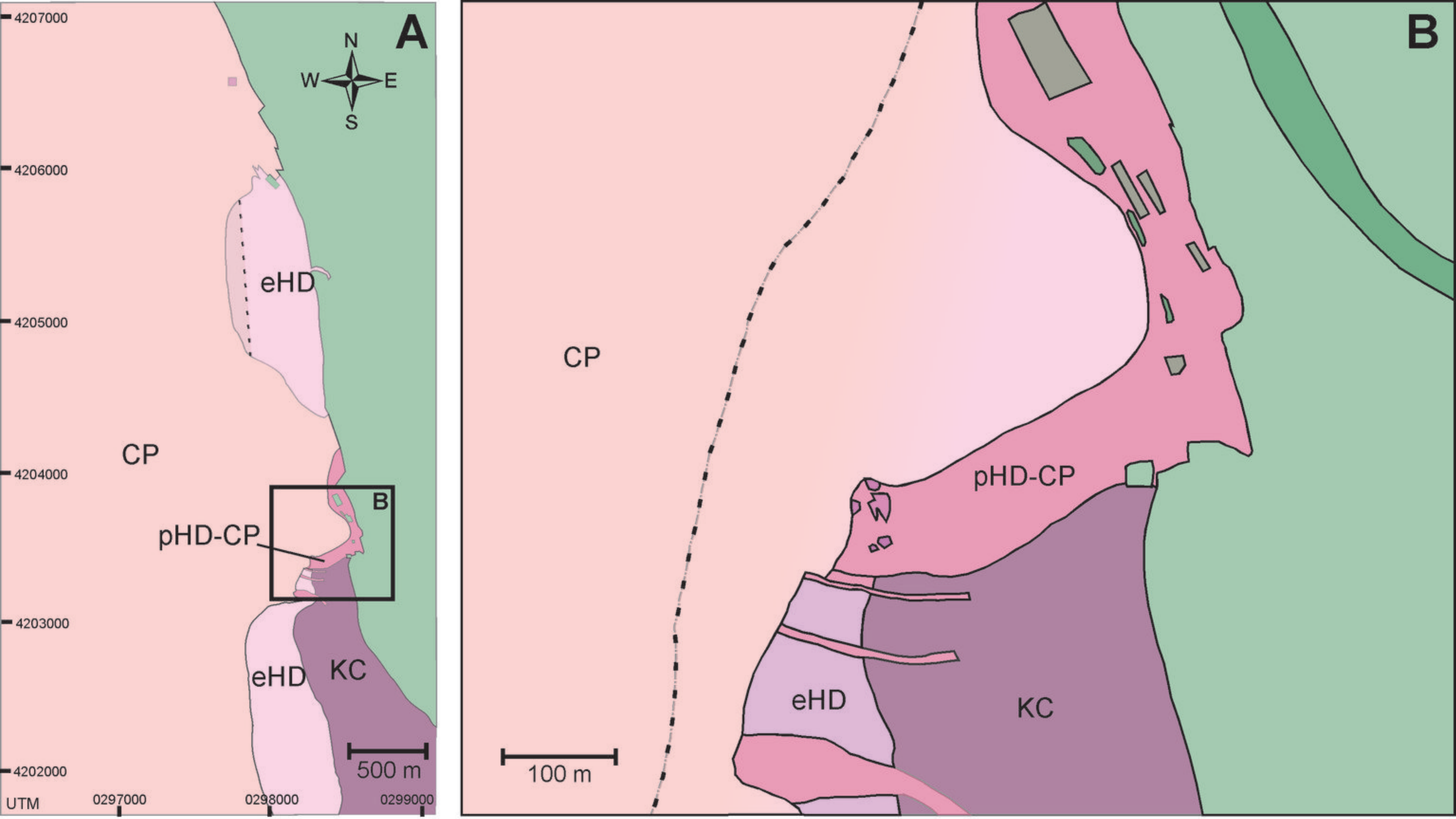

A
N
W
E
S
4207000
4206000
4205000
4204000
4203000
4202000
UTM
0297000
0298000
0299000
eHD
CP
pHD-CP
B
eHD
KC
500 m
B
CP
pHD-CP
eHD
KC
100 m

Figure 6

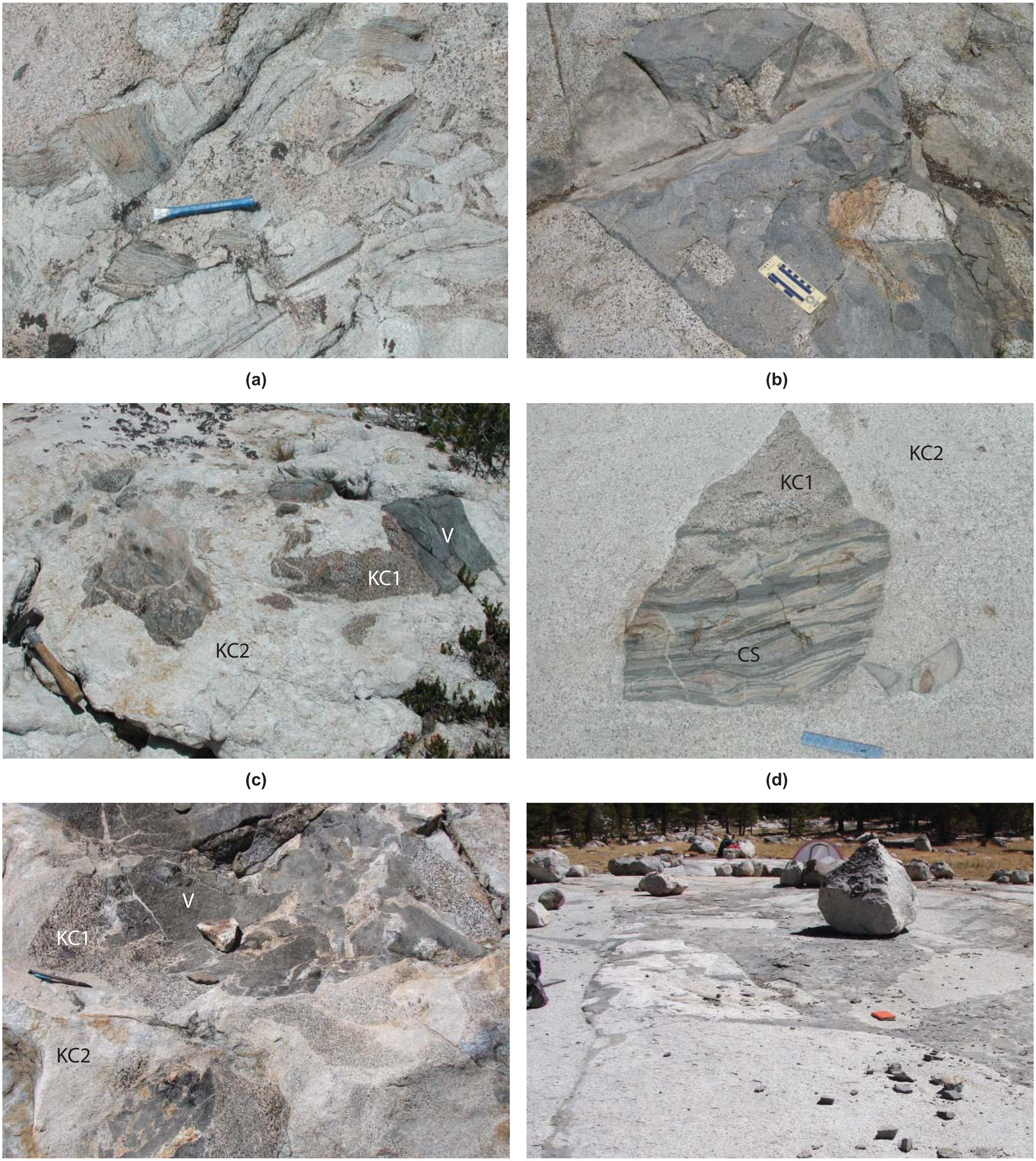


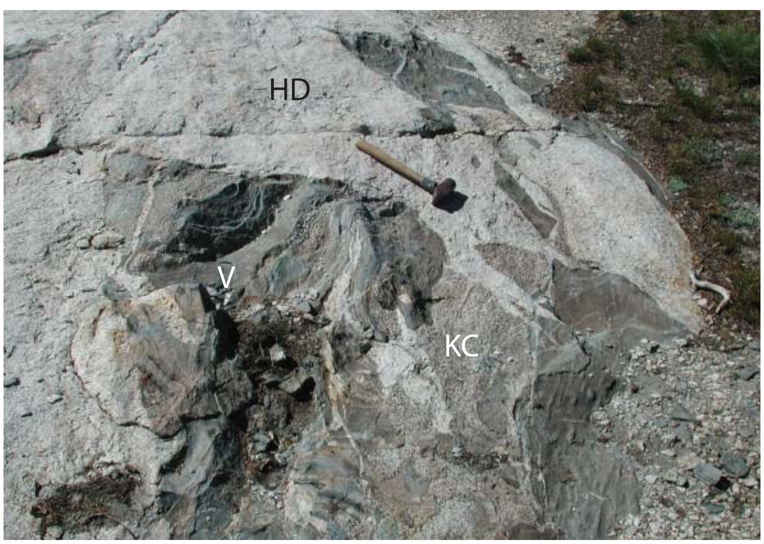
Figure 7

(a)

(b)

(c)

(d)



(e)

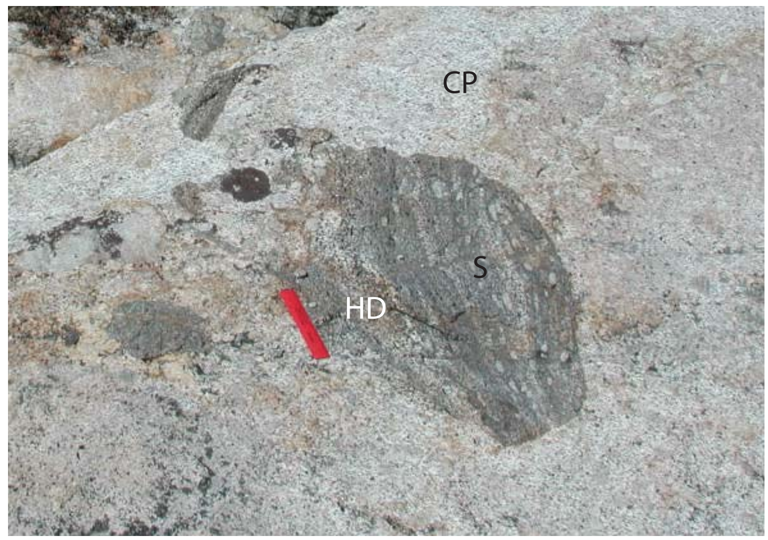
(f)

Figure 8

(a)

(b)

(c)

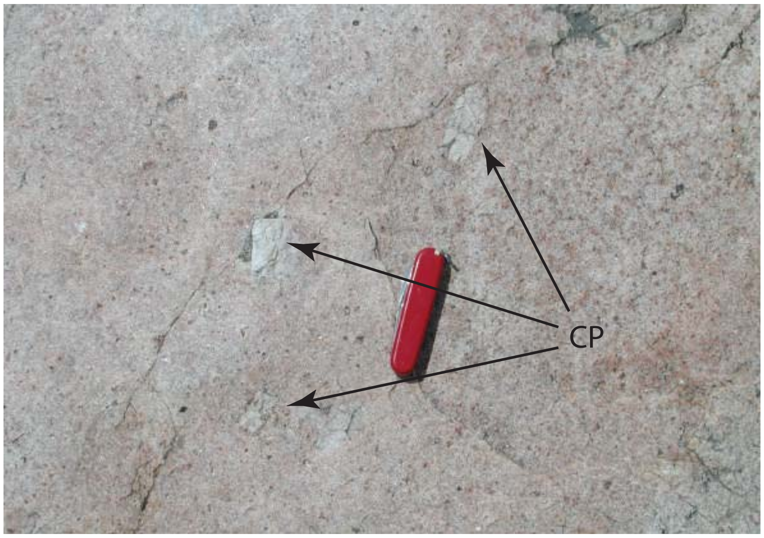


(d)

(e)

(f)

Figure 9



(a)

(b)

(c)

(d)

Figure 10

(a)

(b)

(c)

(d)

(e)

(f)

Figure 11

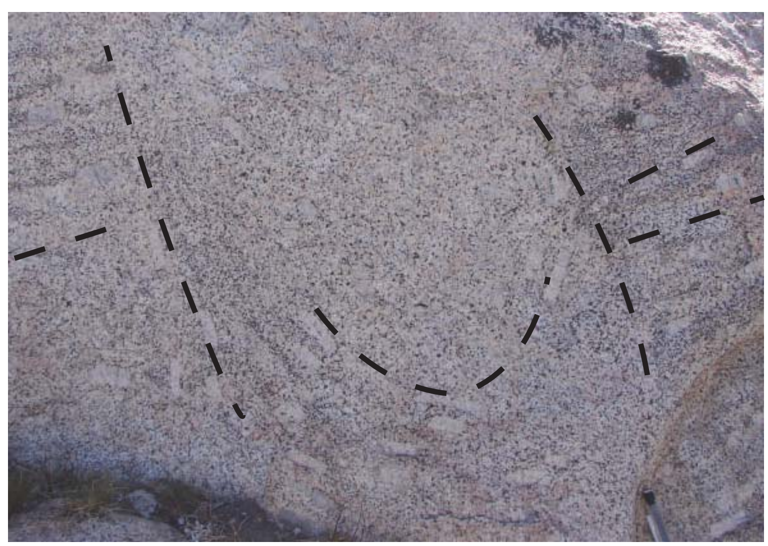
(a)

(b)

(c)

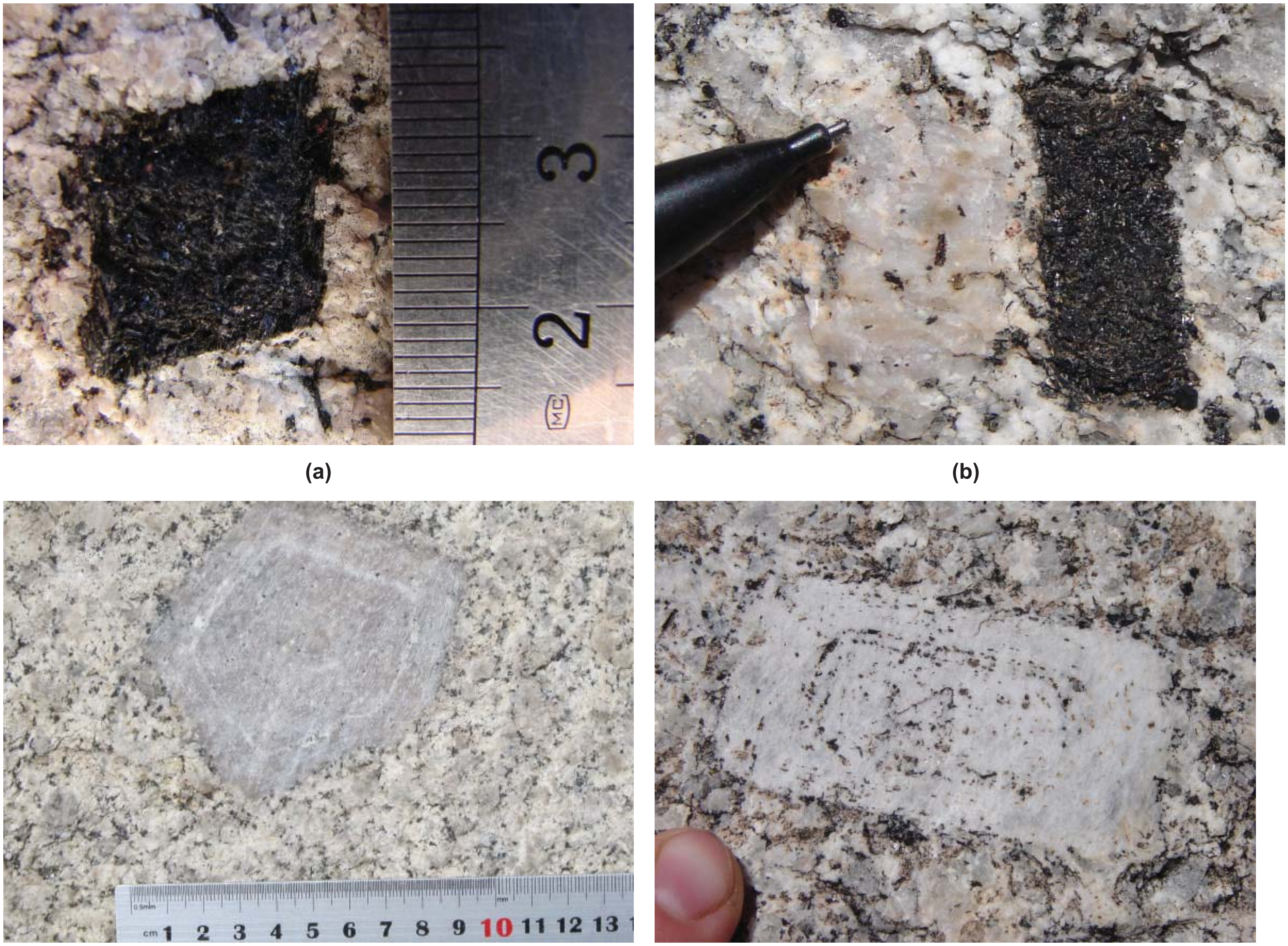
(d)

(e)



(f)

Figure 12



(a) (b) (c) (d)

Figure 13

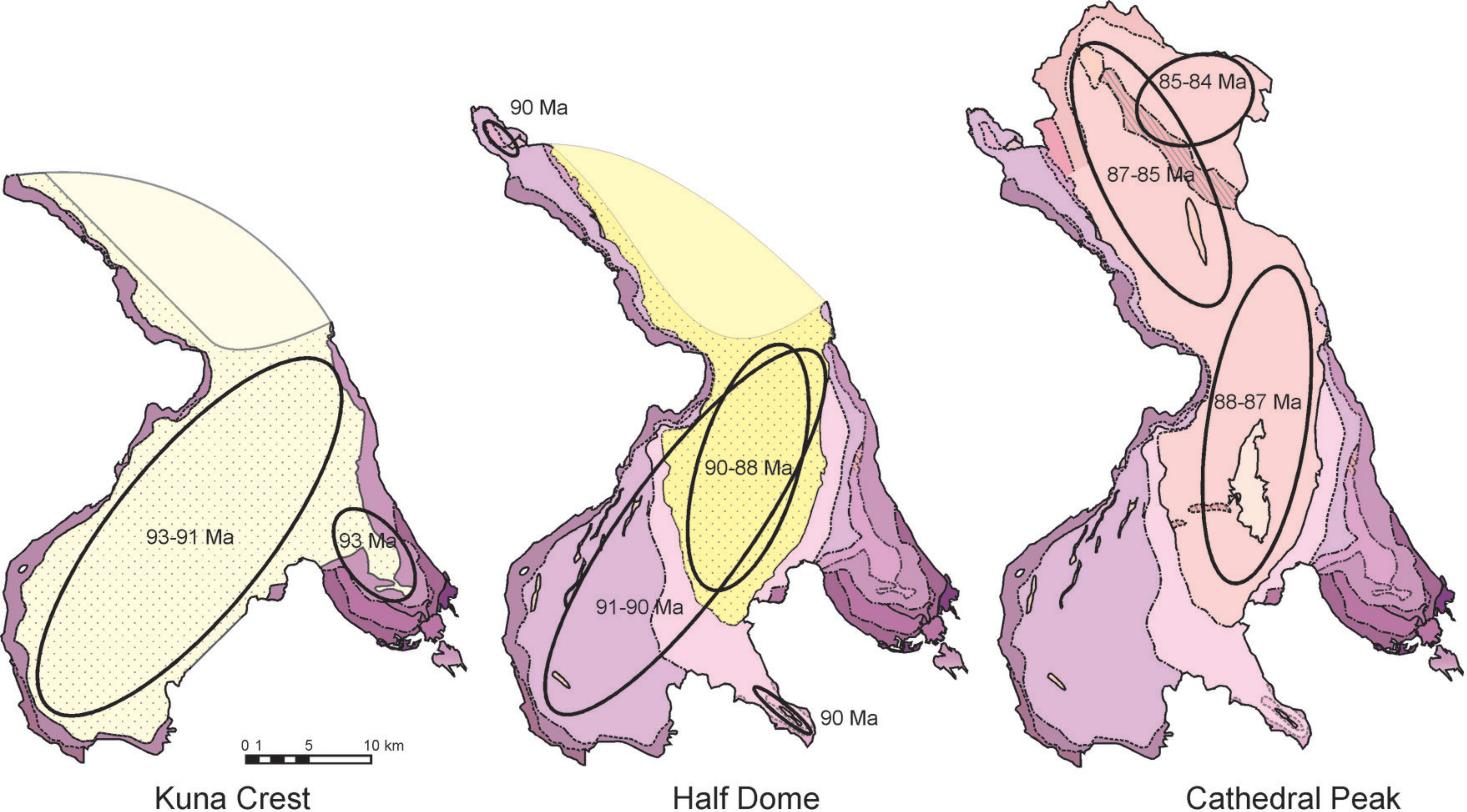
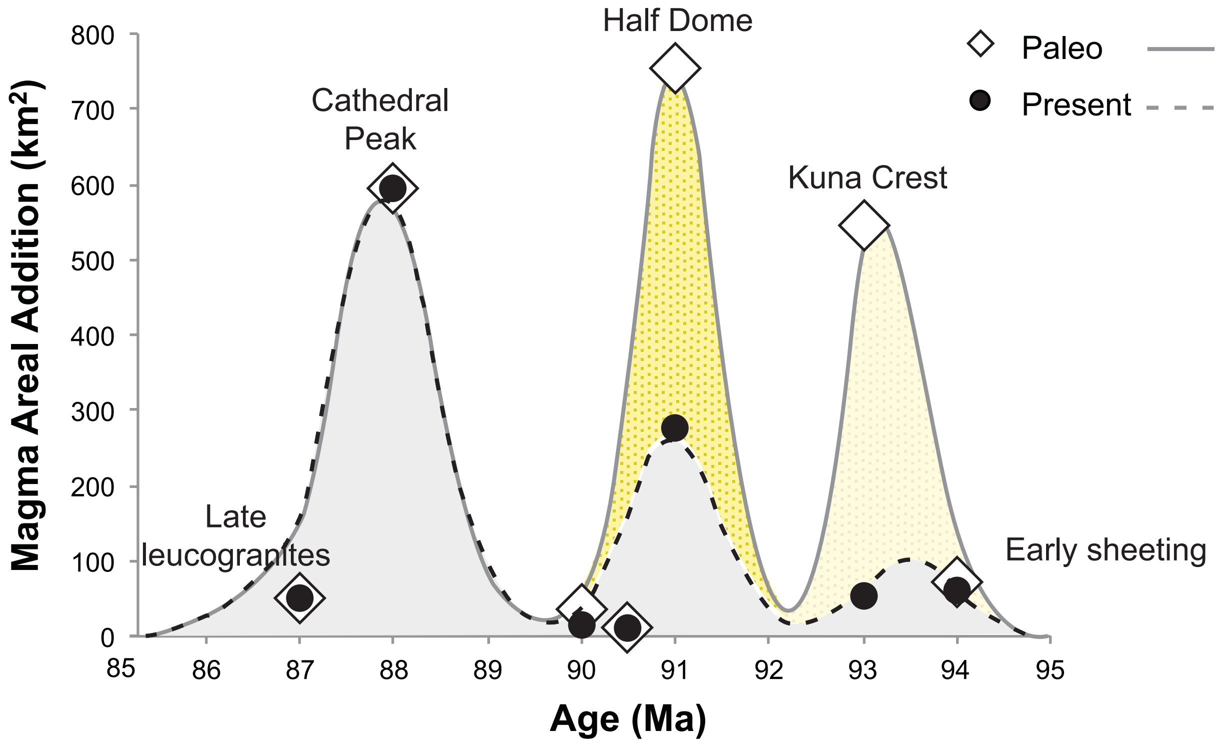

90 Ma
85-84 Ma
87-85 Ma
88-87 Ma
90-88 Ma
93-91 Ma
93 Ma
91-90 Ma
90 Ma
0 1 5 10 km
Kuna Crest
Half Dome
Cathedral Peak

Figure 14



Half Dome
Cathedral
Peak
Kuna Crest
Late
leucogranites
Early sheeting
Paleo
Present
Magma Areal Addition (km²)
Age (Ma)
800
700
600
500
400
300
200
100
0
85
86
87
88
89
90
91
92
93
94
95

Figure 15

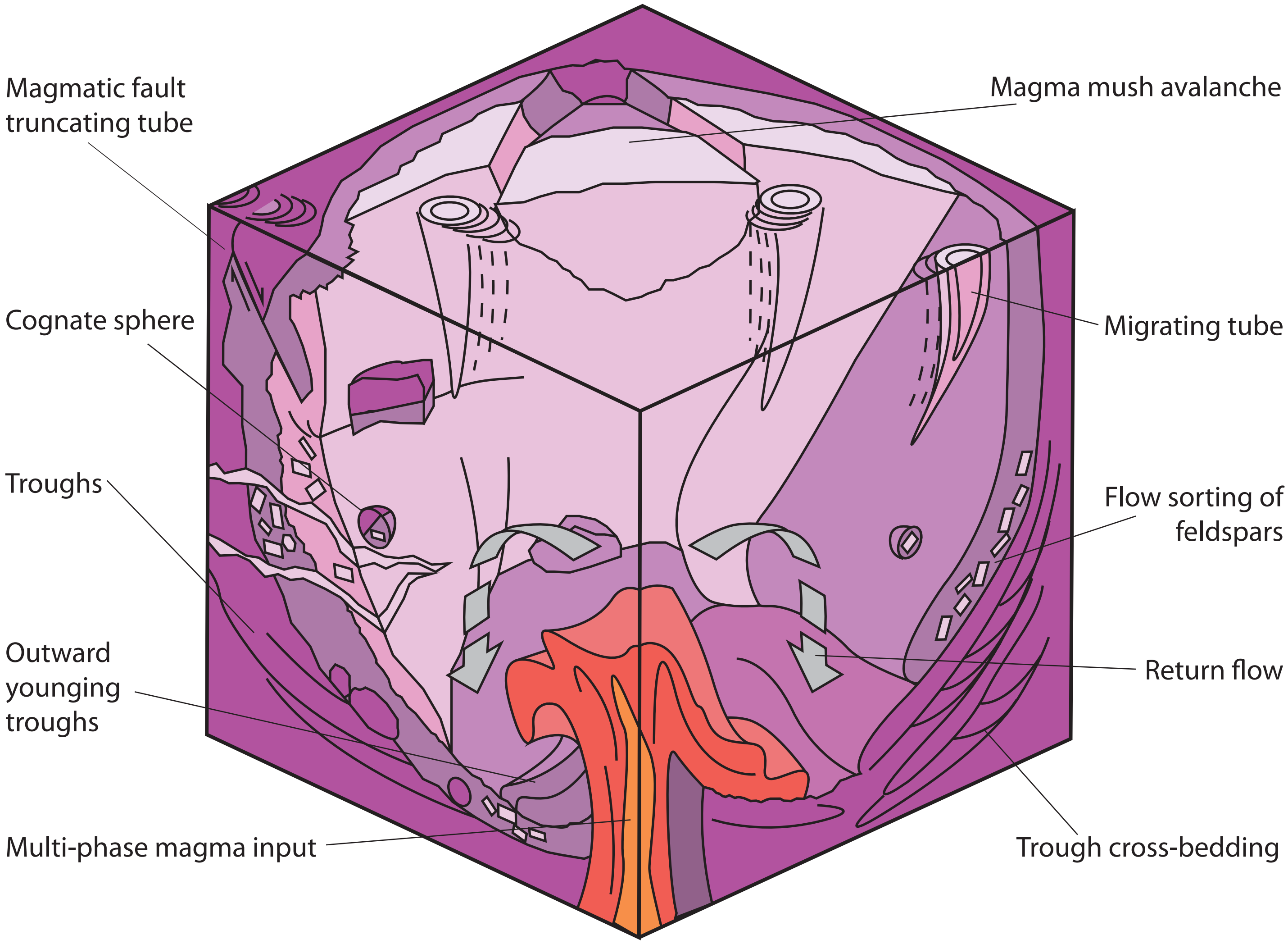

Magmatic fault
truncating tube
Magma mush avalanche
Cognate sphere
Migrating tube
Troughs
Flow sorting of
feldspars
Outward
younging
troughs
Return flow
Multi-phase magma input
Trough cross-bedding

Figure 16