Formation and transfer of stoped blocks into magma chambers: The high-temperature interplay between focused porous flow, cracking, channel flow, host-rock anisotropy, and regional deformation

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ABSTRACT

Magmatic stoping, i.e., the formation, transfer into, and movement through magma of older plutonic and metamorphic host-rock xenoliths, was widespread in the Mesozoic Sierra Nevada batholith (California, United States). However, the prevailing view that stoped blocks form by rapid thermal shattering and collapse into chambers may not be the dominant process of block formation and displacement into chambers in the Sierra Nevada. In detailed studies in and around the Tuolumne Batholith and Jackass Lakes pluton, we found evidence for the following history of block formation in slightly older, fairly isotropic plutonic host rocks: (1) low stress sites developed, leading to planar zones of increased porosity; (2) focused porous flow of first felsic melts followed by intermediate melts led to growth of magma fingers, which in turn led to increased porosity and loss of host-rock cohesion; and (3) connection of magmatic fingers resulted in the formation of dike-like channels in which fluid facilitated removal of all host-rock material in these planar zones. Once formed, blocks were initially displaced by repeated magma injections along these channels, often resulting in unidirectional growth in these zones creating local magmatic sheeted complexes along block margins. Free block rotation occurred when sufficient nonlayered magma surrounded the host block; in some cases, segments of former sheeted zones remain attached to rotated blocks.

In anisotropic metamorphic host rocks, focused porous flow may have locally played a role, but the dominant processes during initial block formation were cracking, parallel and at high angles to anisotropy, and intrusion of magma by channel flow. Subsequent initial block displacement and eventual rotation are identical to those in the nearly isotropic host rock. The driving forces for the development of low-stress sites, cracking, dilation, and magma flow remain uncertain, but likely reflect the interplay between regional stress, magma buoyancy stresses, thermal gradients, and host-rock properties, and not simply rapid heating and thermal expansion cracking. Thus a number of processes may drive block formation, some of which are rapid (thermal shattering, roof collapse) whereas others occur over longer durations (incremental magmatic pulsing and formation of sheeted complexes, regional deformation).

INTRODUCTION

Magmatic stoping, i.e., the formation and transport of host-rock pieces into a magma chamber, was recognized long ago (Goodchild, 1892; Daly, 1903), and stoped blocks have been described from many plutons worldwide (Daly, 1933; Buddington, 1959; Cobbing and Pitcher, 1972; Pitcher and Berger, 1972; Myers, 1975; Barnes et al., 1986, 2003; Becker and Brown, 1985; Clarke and Clarke, 1998; Paterson and Miller, 1998; Miller and Paterson, 2001; Pignotta et al., 2001a, 2001b; Yoshinobu et al., 2003a; 2003b; 2009; Hawkins and Wiebe, 2004; Žák et al., 2006; Farris and Paterson, 2007; Lipman, 2007; Pignotta and Paterson, 2007). The extent and importance of stoping in magma chambers have been matters of vigorous scientific debate, as exemplified by recent suggestions by Glazner et al. (2004), Coleman et al. (2004), and Glazner and Bartley (2006), who suggested that magmatic stoping is rare. Evaluating magmatic stoping is important given that stoping may (1) operate as an important material transfer process (Paterson and Fowler, 1993) during final emplacement of plutons (Fowler, 1996; Žák et al., 2006; Paterson and Farris, 2008); (2) remove large sections of aureoles in host rock that preserved evidence for earlier material transfer processes and thus significantly reduce the information available about the growth of chambers (Fowler, 1996; Wetmore et al., 2003; Yoshinobu et al., 2003a); (3) contribute to vertical exchange of mass and heat within the crust (Marsh, 1982; Furlong et al., 1991; Paterson et al., 1996; Farris and Paterson, 2007); and (4) contribute to the chemical contamination and cooling of magmas during their ascent and emplacement (Marsh, 1982; Furlong and Myers, 1985; McLeod et al., 1998; Barnes et al., 2004; Beard et al., 2005; Clarke, 2007; Erdmann et al., 2009). Whether common or rare, magmatic stoping is also of interest because it allows us to examine other long-standing issues. (1) How do fractures form in rocks at high temperatures (Rubin, 1993; Pignotta et al., 1999; Farris and Paterson, 2007; Pignotta and Paterson, 2007)? (2) How do dikes form in these environments during or after cracking (Rubin, 1995)? (3) What role does anisotropy play in cracking and block formation (Clarke et al., 1998)? (4) What other processes occur in these high-temperature, melt-preserved conditions (Paterson and Miller, 1998b; Brown and Solar, 1998; Barnes et al., 2004; Beard et al., 2005; Clarke, 2007; Yoshinobu et al., 2009)?

Some progress has been made on questions concerning when and where stoping occurs in magmatic systems (Buddington, 1959; Paterson et al., 1991; Marko et al., 2005), how to distin-
guish host-rock rafts from stoped blocks (cf. Glazner and Bartley, 2006; Paterson et al., 2008, for the use of these terms), and the behavior of blocks once within a magma chamber (Paterson and Miller, 1998b; Wolak, 2004; Farris and Paterson, 2007; Yoshinobu et al., 2009). Our study begins by focusing on the issue of how blocks initially form and how they are displaced into magma chambers, and then returns to some of the above issues. Block formation, and transfer into chambers, has been traditionally viewed as a rather dramatic process of thermal shattering and rapid collapse of host-rock blocks into the chamber (Marsh, 1982; Furlong and Myers, 1985; Clarke et al., 1998; Pignotta et al., 2001a). We suggest that this is not the only, or in some cases, even most common mechanism.

For a study on the initiation of stoped blocks, we can temporarily put aside some of the above debates because it is of less relevance how widespread stopping is, as long as at least a few staked blocks formed. The distinction between a raft (commonly defined as a nondisplaced piece of host rock surrounded by magma) versus a staked block (commonly defined as a detached and displaced piece of host rock surrounded by magma) is also less important (e.g., Paterson et al., 2008), as the formation of rafts requires processes similar to block formation, i.e., at least the development of two fractures and magma pulses to form along the raft’s sides, and either some translation of rafts or stopping between rafts (Paterson et al., 2008). Here we document that the initial formation of rafts involves many of the same processes as the formation of staked blocks.

We examine three general settings in the central Sierra Nevada (California, United States), where the processes of crack initiation, initial melt flow, disking, and block formation are well preserved. The first is along contacts of one batch of magma intruded into older, either partially or completely solidified plutonic material. In these settings the host rock may have small amounts of interstitial melt remaining during block formation and at least in a rheological sense at outcrop scale is weakly anisotropic and relatively homogeneous. The second setting is at pluton margins, where magma pulses intrude either metasedimentary or metavolcanic host rocks that are both anisotropic and heterogeneous. The third setting is host-rock blocks already in chambers where blocks were injected by melts during continued block disaggregation (see also Clarke et al., 1998; Farris and Paterson, 2007); the latter example provides additional information regarding the development of high-temperature cracks and the growth and/or propagation of dikes in already hot host rocks.

For each of these settings we describe the observed structures and the inferred processes accompanying crack formation, injection of melt, and formation of blocks. Our foremost conclusion is that blocks formed by an incremental process involving the following steps: (1) dilation and/or microcracking of the host rock resulting in a zone of increased porosity; (2) injection by focused porous flow of felsic melts followed by intermediate melts along these permeable zones forming magma fingers (Pollard et al., 1975; Schofield et al., 2010); (3) continued injection and channelized flow of magma as these fingers widen into planar dikes; and (4) further widening and block displacement through continued injection of numerous pulses of magma and channelized flow forming layered sequences somewhat analogous to layered veins formed by a crack-seal mechanism (Ramsay, 1980; Clarke and Clarke, 1998; Paterson et al., 2007a). Block formation in highly anisotropic host rock is dominated by cracking and channelized magma flow. In both cases blocks begin to freely rotate only when a critical amount of fairly unlayered, thus lower viscosity, magma collects around the block.

**PLUTONS EXAMINED**

The plutons in which block initiation was examined are the Tuolumne Batholith and Turner and Jackass Lakes plutons in the central Sierra Nevada, California (Fig. 1). The Tuolumne Batholith is an ~1100 km², Late Cretaceous composite batholith emplaced into Early Cretaceous granitoids and amphibolite facies metasedimentary rocks of the Kings Sequence to the west (Bateman and Chappell, 1979; Huber et al., 1989; Memeti et al., 2010a) and older plutonic and gneissic to amphibolite facies metavolcanic rocks of Triassic to Cretaceous age to the east (Fig. 1; Huber et al., 1989). Contacts of the Tuolumne Batholith with host rock dip steeply and are generally discordant to the host-rock structures, although rare 100 km domains occur in which older structures are deformed into a margin-parallel foliation and steeply plunging lineation. Estimates of the depth of the currently exposed surface of the Tuolumne Batholith, based on Al-in-hornblende barometry (Ague and Brimhall, 1988; Webber et al., 2001; Gray et al., 2008; Anderson et al., 2008; Memeti et al., 2009), indicate an emplacement depth of ~6–9 km, consistent with the occurrence of andalusite and sillimanite in the surrounding metasedimentary host rock (Memeti et al., 2005).

This batholith consists of three inward-nested, inward- and northward-younging, and progressively more evolved intrusive units called the Kuna Crest, Half Dome, and Cathedral Peak, and a scattering of leucogranite bodies in all of these units, such as the Johnson Granite Porphyry (Bateman, 1992; Memeti et al., 2010b). Geochronologic studies (Kistler and Fleck, 1994; Coleman and Glazner, 1997; Coleman et al., 2004; Matzel et al., 2005, 2006; Miller et al., 2007; Memeti et al., 2010b) indicate that the batholith was constructed over an ~10 m.y. duration between ca. 95 and ca. 85 Ma, and youngs both inward and to the northeast. Several large lobes or fingers of magma that are normally zoned and young inward (Memeti et al., 2010b) extend outward from the main Tuolumne Batholith body. One of these lobes, consisting of porphyritic Half Dome Granodiorite, extends southward from the main batholith and intrudes into slightly older granitic units, the Red Devil Lake and Turner Lake granites.

Immediately south of the Tuolumne Batholith, the Cretaceous Jackass Lakes pluton (Fig. 1), a ~13 km by ~17 km rectangular body, is separated into 4 domains by slightly older metavolcanic and plutonic pendants (Peck, 1980; Pignotta et al., 2010). The U/Pb zircon geochronology provides a ca. 97–98 Ma age of crystallization for the pluton (McNulty et al., 1996). The Jackass Lakes pluton is locally intruded to the south by the ca. 90 Ma Mount Givens pluton (McNulty et al., 2000) and to the north by the 95 ± 2 Ma Red Devil Lake pluton (Tobisch et al., 1995). The Jackass Lakes pluton was interpreted by Peck (1980) to be a relatively shallow, magmatic body that intruded coeval volcanic and subvolcanic plutonic rocks of the ca. 98–101 Ma Minarets caldera sequence (Stern et al., 1981; Fiske and Tobisch, 1978, 1994). Our recent studies generally support this interpretation, although Al-in-hornblende analyses indicate a depth of emplacement between ~8 and 11 km (Coyne et al., 2004; Pignotta et al., 2010).

Mapping across the northern half of the Jackass Lakes pluton suggests that units change intrusive geometry from sheet-like bodies along the eastern margin to more irregular bodies in the interior with composition varying from diorite to leucogranite (Pignotta and Paterson, 2003; Coyne et al., 2004; Pignotta et al., 2010). Abundant host-rock xenoliths occur through-
out the Jackass Lakes pluton and vary from kilometer-scale pendants to centimeter-scale blocks. The large metatelluric rocks have lithologies similar to those in the Minarsells caldera sequence exposed in the Ritter Range pendant to the east. Smaller stoped blocks or rafts are particularly common near the margins of these large pendants, but occur throughout the pluton and range in size (centimeter to decimeter scale) and rock type (metatelluric, less commonly metasedimentary; Yoshinobu et al., 2009; Pignotta et al., 2010).

We examined the process of stoping in a number of settings in and around the Tuolumne Batholith (Fig. 1). In this paper we focus on four areas: (1) the margin of the Half Dome Granodiorite lobe where it intrudes the Turner Lake granite (Figs. 2 and 3; Economos et al., 2009); (2) a gradational internal contact near Potter Point, where the Half Dome Granodiorite intruded the Kuna Crest Granodiorite (Figs. 4 and 5; Zák and Paterson, 2010); (3) in and around the metasedimentary May Lake pendant near the western margin of the batholith (Figs. 6 and 7); and (4) in the Jackass Lakes pluton, where we focus on block formation near pendants and continued disaggregation of individual stoped blocks once they were engulfed by magma in the magma chamber (Figs. 8, 9, and 10; Pignotta et al., 2010). We first examine block initiation in the Turner Lake granite and Potter Point area in the Tuolumne Batholith. These two regions both provide a similar story of block initiation in weakly anisotropic granitic rocks and initial block translation and/or rotation. We then examine block formation and disaggregation of the May Lake pendant and pendants in the Jackass Lakes pluton, which provide examples of stoping of moderately to strongly anisotropic metasedimentary and metatelluric rocks.

FIELD EVIDENCE OF BLOCK FORMATION

Formation of Turner Lake Granite Blocks During Half Dome Granodiorite Intrusion

Near the southern border of Yosemite National Park, a lobe of porphyritic Half Dome Granodiorite extends southward ~15 km from the main batholith into the host rock (Figs. 1 and 2A). This lobe strikes ~130° and intrudes the Turner Lake biotite granite and farther south the Jackass Lakes pluton. The lobe of Half Dome Granodiorite is internally zoned from a hornblende-rich margin of equigranular granodiorite to a central, more leucocratic, porphyritic granodiorite and granite with local leucogranite lenses (Fig. 2A; Economos et al., 2009). At the southern margin of the lobe, the marginal hornblende-rich facies of the equigranular Half Dome Granodiorite intrudes the Turner Lake granite, resulting in two mappable zones therein: an inner magmatic zone, where mingling between the Turner Lake granite and equigranular Half Dome Granodiorite occurs along with the local formation of dikes and schlieren-bounded tubes and troughs; and an outer more brittle zone, commonly with dikes and veins of equigranular Half Dome Granodiorite intruding fractured Turner Lake granite (Fig. 2A). In this region the Turner Lake granite has a magmatic foliation with an average orientation of 303°/81° (azimuthal strike/dip) and a steep magmatic lineation plunging 345°/87° (azimuthal trend/plunge). In the Half Dome Granodiorite lobe the average orientation of the magmatic foliation is 292°/81° and the magmatic lineation plunges 019°/86°.

A detailed examination of block formation in the Turner Lake granite during intrusion by Half Dome Granodiorite magma, ~300 m south of the nearest Half Dome lobe margin, involved mapping the region at a scale of 1:24,000 (Fig. 2A; Economos et al., 2009) followed by the construction of two 1:100-scale grid maps ~100 m apart on glacially polished surfaces (Figs. 2B, 2C). We interpret grid map 1 (Fig. 2B) to reflect the initiation of block formation and grid map 2 (Fig. 2C) to show a more advanced stage of Half Dome Granodiorite intrusion and Turner Lake granite block rotation. The average magmatic foliation in the Turner Lake granite in these grid-mapped areas is 129°/83°, roughly parallel to the nearby Half Dome lobe margin, which strikes ~120°. Magmatic lineation consistently plunges steeply, and thus was not easily measured on the horizontal and glaciated but locally weathered and jointed rock slabs.

Grid map 1 consists of Turner Lake granite with ~1 cm large quartz and 1–2 cm-long K-feldspar phenocrysts, and biotite as the only mafic phase that was intruded by a network of thin (millimeters to centimeters in thickness), quartz-poor, hornblende-rich, equigranular Half Dome Granodiorite dikes and/or veins containing no K-feldspar phenocrysts (Figs. 2B and 3). The K-feldspar phenocrysts form clusters in the Turner Lake granite, which are shown in Figure 2B. Thus, in the grid-mapped area, hornblende is unique to the Half Dome magmas, whereas K-feldspar phenocrysts and large quartz pools are unique to the Turner Lake granite. This distinction allowed the identification and separation of the two rock types even at the mineral scale during grid mapping.

Upon casual inspection the Half Dome Granodiorite in this area appears to form a pattern of thin dikes and veins with a dominant orientation of 141°/89°, and secondary bodies at other orientations (Figs. 2B and 3). These dike-like bodies create a general pattern of rectangular and less common triangular Turner Lake granite blocks (Figs. 2B and 3B). Their dominant orientation is clearly not parallel to the magmatic foliation in the Turner Lake granite, and instead typically occurs at ~15°–20° clockwise from this foliation (Fig. 2B). The tip region and some walls of these dike-like zones are composed of leucocratic material dominated by quartz with minor biotite and feldspars. Cracks are sometimes found in front of dike tips and elsewhere cracks occur without any new material associated with them. A number of crack tips are terminated where they intersect clusters of K-feldspar phenocrysts.

Upon closer inspection a number of fascinating features indicate that these zones initially formed as zones or fingers of focused porous flow (Korenaga and Kelemen, 1998; Kelemen et al., 2000), which then evolved into dikes of channelized melt flow (Pollard et al., 1975; Schofield et al., 2010). For example, when the dike-like zones are narrow (<5 cm), they are composed of a framework of K-feldspar phenocrysts and quartz pools (formed in the Turner Lake granite, but not in the equigranular Half Dome Granodiorite in the grid map area) surrounded by local pools of Half Dome Granodiorite or single hornblended, typical for the Half Dome Granodiorite (Fig. 3F). Elsewhere, small linear segments of these focused porous flow zones are separated from one another by completely intact bridges of Turner Lake granite (Fig. 3F). Some of these segments have blunt rounded tips consistent with porous flow, but unusual for dike tips. However, the focused porous flow segments form fairly linear zones, and in grid map 1 evolve along strike into dikes along block margins (Figs. 2B and 3G).

Once fingers of Half Dome Granodiorite are connected and enlarged (typically to widths of 5–25 cm) they contain several features indicative of channelized flow in dikes. For example, in places Half Dome Granodiorite hornblended define arcuate-shaped magmatic foliation from dike walls to the center. Minerals formed in the Half Dome Granodiorite dominate the matrix in these dikes and only rare xenocrysts of Turner Lake granite remain. Furthermore, local xenoliths and microgranitoid enclaves can be found in these zones, but are absent from zones interpreted as focused porous flow. In one case an enclave extends from wall to wall of a dike and shows signs of distortion (compaction) along both walls. We interpret this to indicate that the dike was once wider, that the enclave flowed into the dike during channelized flow, and that the distance between dike walls decreased as magma flow slowed, eventually distorting the enclave (e.g., Figs. 3F, 3G).
Figure 1. Map of Tuolumne Batholith redrafted after Huber et al. (1989). Boxed areas are locations discussed in text of completed studies of stoping. Sp—areas of stoping of plutonic units; Sh—areas of stoping of host-rock units. BLP—Benson Lake pendant; CLP—Cinko Lake pendant; GA—Glen Aulin; HDL—Half Dome lobe; JL—Jackass Lakes pluton; MLP—May Lake pendant; SLP—Soldier Lake pluton; SM—Sawmill Canyon; PP—Potter Point; WLP—Waugh Lake pluton. Inset is California map showing outline of Sierra Nevada arc and location of Tuolumne Batholith. SF—San Francisco; TB—Tuolumne Batholith; LA—Los Angeles.
Figure 2. (A) Half Dome lobe map (see Fig. 1) intruding older granites and showing internal zonation and overprinting magmatic fabrics. (B) Grid map 1 (location is shown by red star in A). Dikes and dikelets of Cretaceous Half Dome (K_{HD}) Granodiorite and associated melts along cracks in the host Turner Lake (K_{TL}) granite. See key for colors and symbols of magmatic foliations, K-feldspar clusters, and mafic enclaves in grid map. (C) Grid map 2 ~100 m south of B. Amoeboid distribution of Half Dome Granodiorite in Turner Lake granite with accumulation of mafic minerals, microgranitoid enclaves, and Turner Lake granite blocks.
In wider dikes (typically 15 cm wide or larger), we find examples indicating multiple pulses of magma (Fig. 3H; e.g., Clarke and Clarke, 1998). Separate pulses are recognized by their slightly different compositions and microstructures, and are often bounded by graded schlieren. These schlieren in places define troughs, which are truncated by younger schlieren-bounded troughs, indicating erosion of older schlieren during flow and redeposition of magmatic minerals (Fig. 3H). These trough cutoffs provide a local younging or growth direction, and where multiple cutoffs are preserved in a single sheeted zone, they almost invariably show a consistent younging direction from one side of the zone to the other. This suggests that younger pulses consistently intruded along the same margin of the sheeted zones.

In contrast, grid map 2 is dominated by Half Dome Granodiorite with an unusually large amount of coarse mafic minerals resulting in hornblende-biotite cumulates. These cumulates formed irregularly within the Turner Lake granite, initially forming dikes and then entirely encompassing pieces of Turner Lake granite, resulting in both angular and rounded blocks rotated within the cumulate. These pieces could locally be reassembled using matching boundaries. The overall pattern of the areas occupied by Half Dome magmas in this grid is more irregular to rounded, with fewer angular block boundaries and no evidence of repeated pulsing along a single pathway as seen in grid map 1. Magmatic foliation in both the Half Dome and the Turner Lake granite units in grid map 2 are steep and typically parallel, forming a strike maximum toward the northwest and smaller maxima toward the southeast and the southwest.

### Intrachamber Stoping at Potter Point, Tuolumne Batholith

An intriguing example of intrachamber block formation was discovered in the eastern part of the Tuolumne Batholith along a gradational...
Stoped block formation

The domain of block formation is exposed in a flat, glacially polished, ~300 × 100 m outcrop located in Lyell Canyon, ~1 km southeast of Potter Point (Fig. 4B). The domain occurs at the inner edge of an ~300-m-wide transitional zone between the Kuna Crest and equigranular Half Dome Granodiorites, and near where the transitional zone is in contact with equigranular Half Dome Granodiorite (Fig. 4B; Žák and Paterson, 2005, 2010; Memeti and Paterson, 2006, 2010b). This transitional Kuna Crest–Half Dome Granodiorite unit is characterized by a modal decrease in the percentage of mafic minerals, increase in the size and abundance of euhedral hornblende, and a modal decrease in titanite. This rock type is herein referred to as the transitional Kuna Crest Granodiorite, and it defines a zone that strikes ~50° across Lyell Canyon, changes to ~340°, and continues northward along the east side of the canyon (Fig. 4B).

At the Potter Point locality, the transitional Kuna Crest Granodiorite has been intruded by numerous mafic (quartz diorites, locally with hornblende-biotite cumulates), felsic (aplite, aplopegmatitic, and trondhjemitic compositions), and mafic-felsic sheets. Mingling, enclave swarms, ladder dikes, or schlieren tubes (Weinberg et al., 2001; Paterson, 2009) and schlieren-bounded troughs, along with numerous blocks marked with dashed lines. Half Dome sheets show crosscutting relationships indicating sheets young to left in photo.

Figure 3 (continued). Field photos of Turner Lake granite block formation during intrusion of Half Dome Granodiorite. (E) Dike with felsic margin in grid map 1 with lots of Turner Lake granite-derived crystals in dike. Felsic margin may be former dike tip as shown in C. Knife is 6 cm long. (F) Turner Lake granite bridges between fingers (e.g., white arrows) of focused porous flow of hornblende-bearing Half Dome Granodiorite in grid map 1. Ruler is 15 cm. (G) More continuous Half Dome Granodiorite channels with regions of enclave-rich magma (southwest of ruler) in Turner Lake granite. (H) Magmatic layers formed by repeated Half Dome pulsing in zone between two Turner Lake granite blocks (block margins marked with dashed lines). Half Dome sheets show crosscutting relationships indicating sheets young to left in photo.
Figure 4. (A) Map showing distribution of internal units (I-V$_{\text{Vhybrid}}$; V$_{\text{hybrid}}$ = transitional HD and KC unit) in the Kuna Crest Granodiorite and in the adjacent Half Dome Granodiorite (eHD and pHD; e = equigranular, p = porphyritic) in the Kuna Crest lobe region of the Tuolumne Batholith (location in Fig. 1). (B) Map showing complex mingling/mixing zone (mHD) along the Half Dome-V$_{\text{Vhybrid}}$ contact (location = box in Fig. 4A) and orientation of structures in the Potter Point area, Lyell Canyon; (C) Grid map #3 (location = red star in Fig. 4B) showing angular blocks of mHD, quartz dioritic enclaves in eHD, and cumulates of mafic minerals.
Grid map 3 (Fig. 4C) was constructed using a rectangular 9 m × 7 m grid on a flat glacially polished surface. Photographs were taken of each grid cell, put together into a photomosaic, and redrawn into a highly detailed outcrop map (Fig. 4C). The mapped zone has outer boundaries featuring mostly straight, sharp, approximately northeast-southwest–trending, steeply dipping, synmagmatic fractures. However, some segments have a more irregular to lobate geometry. Inside the zone, abundant blocks of the transitional Kuna Crest Granodiorite and felsic dikes are surrounded by compositionally heterogeneous intrusive material, typically more mafic than the host granodiorite. This matrix contains abundant accumulated mafic microgranitoid enclaves or larger irregular blobs of granodiorite and quartz diorite (Figs. 4 and 5). In some places the enclaves are closely packed, touch, or impinge on and deform one another. The more mafic intrusive material is associated with synmagmatic quartz diorite units seen elsewhere in this lobe of the batholith (Memeti et al., 2007, 2010b). The more felsic intrusive magma is likely derived from equigranular Half Dome magma, because it is clearly younger than the transitional Kuna Crest Granodiorite and contains large euhedral hornblendes and microgranitoid enclaves that are similar to those found in the Half Dome Granodiorite. However, further geochemical and petrologic study is needed to precisely define all the units exposed in the Potter Point area.

The size of the transitional Kuna Crest Granodiorite blocks varies from several decimeters to 5 m (Fig. 4C and 5). In map view, the blocks are triangular or rectangular with either blunt or very irregular corners (Figs. 5C, 5E–5H). Margins to these blocks are typically sharp and straight, but
locally also lobate or diffuse. Structures such as deformed layering, enclaves, and sheets around the blocks indicate that the blocks have been rotated and/or displaced relative to each other in the horizontal section (Figs. 4C and 5E–5H). Blocks of both the host transitional Kuna Crest Granodiorite and the more mafic Half Dome-derived magma have been overprinted by a late ~290°, steeply dipping magmatic foliation (Type 4 fabric of Žák et al., 2007). This late fabric is preserved inside the blocks, overprints their margins, and is also preserved in younger magma pulses around the blocks (Fig. 4C).

A variety of small-scale structures, similar to those described in the Turner Lake granite, preserved in this and nearby zones provides further evidence for the mechanisms of crack initiation and breakup of the Kuna Crest Granodiorite host (both normal and transitional), and in some places, of felsic dikes. In places, thin (0.5–1 cm), submagmatic cracks in the host granodiorite and inside some previously formed blocks were intruded by leucocratic, aplitic to trondhjemitic melt to form veins (Figs. 5A, 5B). These leucocratic veins are very fine-grained, have diffuse margins and tips, and dissipate into the host granodiorite. Bridges of host material between crack segments, abundant host minerals in veins, and broad, rounded tips of veins all indicate that the initial cracking and vein formation occurred by growth of fingers during focused porous flow of melts into Kuna Crest Granodiorite.

Cracks that grew into wider veins (10–20 cm = dikes) during channelized flow in this area are filled by quartz diorite and granodioritic magmas. These dikes have sharp margins against the host granodiorite, contain few to no host-rock minerals, and often show evidence for multiple pulses of magma intrusion resulting in larger, composite, sheet-like bodies 0.5 m to several meters thick (Figs. 5D–5F). In map view, these mafic dikes are straight, but some have also been folded into open magmatic folds with a prominent approximately northwest-southeast to west-northwest–east-southeast magmatic

Figure 5 (continued). Field photos of Potter Point area. (E) Sheeted zone forming along growing block margin. (F) Connected sheeted zone forming along two margins at block corner. Block is underneath erratic boulder. (G, H) Rotated and collected stoped blocks with fairly sharp margins and rectangular shapes, sometimes with accumulations of microgranitoid enclaves in surrounding magma. H is photo from grid map 3 area shown in Figure 4C. The hammers in E, F, G, H are all approximately ~30 cm long.
foliation parallel to their axial planes. Locally, a few thin leucocratic veins derived from the surrounding granodiorite intruded across the mafic dikes and were magmatically folded.

A more geographically extensive examination of the dikes and larger, internally layered sheets in the Potter Point area indicates that they form two systematic sets based upon their orientation and crosscutting relationships (see Žák and Paterson, 2010): (1) an older, ~290°–310° striking, steeply dipping, predominantly dioritic set, and (2) a younger, ~30°, steeply dipping, predominantly felsic set. However, several cases of reverse or more complex crosscutting relationships were observed. The average spacing (the shortest distance perpendicular to sheet margins) of the sheets is 20–30 m. The individual sheets are typically 1–2 m thick and have sharp and straight outer boundaries. Commonly, the sheets are strongly internally layered with alternating dioritic to granitic layers that frequently crosscut each other. In some cases sheets preserve more mafic interiors relative to felsic rims. Timing relationships in composite mafic-felsic sheets indicate that younger mafic magmas intruded the central parts of felsic sheets. Kinematics associated with approximately north-east-southwest mafic sheets, observed in a horizontal surface and determined by offsets of older structures along sheet margins, are predominantly right lateral, whereas approximately north-east-southwest felsic sheets are mostly dominantly right lateral, whereas approximately of older structures along sheet margins, are pre-

A horizontal surface and determined by offsets (5) late crack-normal shortening or shrinkage occurs, as recorded by minor magmatic folds and the packing and deformation of enclaves. A great majority of the above structures were overprinted by a late ~290° regional magmatic fabric, suggesting that melt was present during and after the growth of layered zones and block formation in the host granodiorite.

Formation of Metasedimentary Blocks Around the May Lake Pendant

Several metasedimentary pendants are preserved along the western margin of the Tuolumne Batholith and eastern margin of the ca. 102 Ma El Capitan batholith (Fig. 1; Huber et al., 1989; Schweickert and Lahren, 1991; Memeti et al., 2010a). One of these, the northeast-trending May Lake inter plutonic screen, consists of ~3000 m x 750 m of deformed quartzites, metapelites, and less common calc-silicate rocks, marbles, and biotite-hornblende gneisses (Figs. 6 and 7). Most compositional units are interbedded, whereas the <5-m-thick biotite-hornblende gneisses crosscut layers and are interpreted as former mafic dikes. The dominant structures in the May Lake pendant are open to tight folds (F3; Fig. 7A) that refold both smaller scale isoclinal folds (F2; Fig. 7B) with amplitudes and wavelengths <2 m and a layer-parallel axial planar foliation (S3). Younger kink and box folds (F4) with locally developed, west-northwest–striking axial planar fracture cleavage are developed in domains and fold all previous structures. Outside of fold hinge regions, these structures form a strong anisotropy (bedding plus foliation) striking approximately northeast-southwest in the pendant.

Contacts between the May Lake pendant and the adjacent batholiths are irregularly stepped (Figs. 6A, 6B). Step lengths vary between centimeters and tens of meters. Lit-par-lit contacts between metasedimentary and magmatic rocks are locally developed particularly at the corners of stepped margins, and are most common in the marginal 2 m of the pendant (Fig. 6B). Dikes of both the El Capitan Granite and Tuolumne Batholith intrude farther into the pendant but are also most common near the pendant margins.

Along the southeast margin of this pendant, the Tuolumne Batholith consists of the ca. 94–92 Ma Kuna Crest tonalite to granodiorite, a 100–250 m transitional zone with complex schlieren, layering, and migling, and the ca. 92–88 Ma Half Dome Granodiorite to the east (Fig. 6A). Within the Kuna Crest, a 5–150-m-wide zone of leucocratic sheets dominated by fine- to coarse-grained quartz and K-feldspar and little plagioclase and biotite is present. Xenoliths from the May Lake pendant occur in the Kuna Crest unit, are particularly common in the leucocratic sheets (Figs. 7C, 7D), and are present but decrease rapidly in abundance in the transitional zone and the Half Dome Granodiorite.

Block Initiation in Pendant

In contrast to the previous Turner Granite and Potter Point examples, there is less evidence preserved of focused porous flow and growth of magmatic fingers during block formation in the May Lake inter plutonic screen. Discontinuous melt leucosomes (Figs. 7A, 7B) between larger veins and dikes may represent such focused porous flow, but are difficult to separate from the alternatives of in situ melting and local channel flow from nearby melt zones. Dike-filled fractures formed by channelized magma flow in both the pendant and in disaggregating blocks adjacent to the pendant are common. At least three sets of plutonic dikes or veins occur in this inter plutonic screen that are compositionally distinct: one is derived from the El Capitan Granite, one was likely generated by in situ melting during either El Capitan or Tuolumne Batholith emplacement, and a third is derived from the Tuolumne Batholith (Fig. 7). All three sets represent examples of host-rock cracking and magma injection, and thus may play a role in block formation, although here we focus on just the latter two. The melt leucosomes within the metasemipelites and metapelites comprise, to various degrees, thinly laminated mesosomes with a characteristic assemblage of biotite, quartz, plagioclase, K-feldspar, ± andalusite, ± cordierite, and ± garnet, and leucosomes dominated by quartz and K-feldspar (commonly partially replaced by muscovite) and rare plagioclase. The leucosomes are dominantly oriented parallel to the compositional layering, although abundant leucosomes occur oblique to and cut across the compositional layering. In these cases, leucosomes are often located parallel to fold axial planes, along shear planes, or filled in boudin necks, which are low-stress sites (Fig. 7).
Figure 6 (on this and following page). (A) Simplified geologic map of the May Lake pendant, located along the western margin of the Tuolumne Batholith and eastern margin of the El Capitan Granite (see Fig. 1 for location). Note collections of stoped blocks along the eastern margin of the pendant with blocks now residing in the Kuna Crest unit of the Tuolumne Batholith.
Figure 6 (continued). (B) Grid map #4 of rotated stoped metasedimentary block along the northeastern edge of the pendant in Kuna Crest magma.
Layer-parallel leucosomes are commonly rimmed by <2-mm-thick, fine- to medium-grained biotite-rich melanosomes and are typically thickened in fold hinges; leucosomes oriented at high angles to fold limbs are pytymatically folded. Mesosomes and leucosomes in fold limbs commonly show pinch-and-swell structures.

The dikes from the Tuolumne Batholith are typically fine- to coarse-grained leucocratic granite, but tonalite and granodiorite injections occur locally. They show geometrical relationships identical to those of the leucosomes noted, i.e., they most commonly intruded parallel to compositional layering, but in some places intruded at high angles to this layering. In the latter case, they are often subparallel to the axial planar cleavage of F2 and F3 fold sets (Fig. 7B). The development of these dikes records the initial formation of host-rock block margins: their geometry and timing relationships indicate that the strong anisotropy in the metasedimentary rocks, plus active deformation, played a critical role in dike orientation and resulting block formation.

One block at the immediate northeast boundary of the May Lake pendant was examined in detail. Figure 6B shows grid map 4 of this metasedimentary block enclosed in the Kuna Crest Granodiorite to leucogranite, covering an area of >13 m × 6 m. The northwestern margin of the block is ~1 m away from the contact with in situ rocks of the May Lake screen. Decimeter-
to meter-thick, medium- to coarse-grained quartzite layers, intercalated with centimeter- to decimeter-thick metapelitic layers, dominate the block. The thicker layers are continuous and most of the thinner layers are discontinuous. Aligned metamorphic minerals define a foliation, \( S_2 \), parallel to the compositional layering, cut by centimeter- to decimeter-scale faults, in both the block and the in situ rocks of the pendant. The orientations of structures between these two sites (primarily compositional layering and foliation) differ on average by \( \sim 35^\circ \) in strike, whereas the dips remain consistently steep.

The southern contact of the block is subparallel to the compositional layering and \( S_2 \), but all other block margins are oriented approximately perpendicular or at high angles to the compositional layering and \( S_2 \). The layering and foliation of the in situ pendant rocks are sharply truncated by the magmatic rocks at a high angle. All margins are sharp and irregularly stepped on a scale of millimeters to decimeters; the locations of stepped contacts are typically associated with changes in layer composition (Fig. 6B).

The host around this metasedimentary block is a medium- to coarse-grained leucogranite with \( \leq 5\% \) modal volume biotite \( \pm \) hornblende. Patches of pegmatite are typically oriented parallel to the contact of magmatic and metasedimentary rocks. The foliation within the leucogranite is weak, but where visible, it is subparallel to the margin of the metasedimentary block and in situ host rock. A few biotite clusters and microgranitoid enclaves are also oriented subparallel to the margins of the metasedimentary block and in situ host.

Small-scale blocks of centimeter to decimeter size are irregularly distributed around the large metasedimentary block (Fig. 6B). Their orientations are either random to each other (e.g., small blocks along the southwestern and northern-western margin of the large block), or oriented subparallel to the margin of the larger block, the magmatic layering, and the magmatic foliation. Several of the small blocks, which show compositional layering and \( S_2 \) at an angle to the adjacent large block and the in situ metasedimentary rocks of the pendant, are also not of the same rock type. For example, blocks of metapelitic rock occur along the strike of quartzitic layers of the large block and the in situ rocks, and quartzite blocks occur along the strike of a metapelitic layer of the large block.

Dikes (<10 cm wide) and irregular magma pods occur primarily within the large block, and to a lesser extent in the in situ metasedimentary rocks. Most dikes are oriented approximately oblique to the compositional layering and \( S_2 \) of the larger block. The dikes are leucogranitic to pegmatitic, and originate in the magmatic host (i.e., no crosscutting relationships are visible macroscopically where the dikes enter the block), and some dikes only contain quartz at their tips. The pods of pegmatitic material have highly irregular shapes in the block, but the compositional layering and \( S_2 \) of the metasedimentary rocks commonly define one of the dominant contact surfaces. Some dikes and pegmatitic pods contain small-scale fragments of the metasedimentary host.

The magmatic host shows some layering and irregular patches of fine-grained to coarse-grained leucogranite and pegmatite along the block margins, possibly indicating that several injections of magma occurred along the fracture between the large block and the in situ rocks of the May Lake pendant. The quartz-filled tips of veins and dikes within the block suggest that the injected melts reached volatile saturation, and that the quartz crystallized from a supersaturated fluid. Clarke et al. (1998) suggested that the exsolution of a fluid from the magma might have triggered further opening or widening of a preexisting crack.

This metasedimentary xenolith of the May Lake pendant (grid map 4) is interpreted as a stope block, given the evidence for a \( \sim 35^\circ \) rotation in a counterclockwise direction. Other structures, such as an open or kink fold that could explain the geometric relationship between the block and the in situ rocks, where the core of the fold would have been obscured by the injection of the host magma, can be ruled out because no folds with the required geometric relationship and orientation exist at the mapped locality or in the entire May Lake pendant. For several of the smaller stopped blocks, the variable orientations over a short distance and the mismatch in rock type along strike show clearly that the rock fragments are blocks that have been rotated and transported for at least some distance in the magmatic host.

**Metasedimentary Block Disintegration in Tuolumne Batholith**

Metasedimentary blocks, most of which compositionally and structurally match units in the May Lake pendant, are common (<1–5 vol%) in the <500-m-wide zone of Kuna Crest Granodiorite, are less abundant in the Kuna Crest–Half Dome hybrid zone (\( \leq 2\% \)), and rare in the Half Dome Granodiorite (\( \leq 1\% \); Figs. 6, 7C, and 7D). Blocks of metapelitic, quartzitic, and calc-silicate rocks and rarer metavolcanic rocks and hornblende gneiss can reach sizes of \( <150 \) m \( \times \) \( 40 \) m in two dimensions, but more typically range from \( <5 \) cm to \( <5 \) m in size. The larger blocks occur as either isolated entities or with smaller blocks around their margins. The smaller blocks occur both as isolated pieces and in clusters (Figs. 6B and 7D). Contacts between metasedimentary blocks and host granitoids are typically sharp and straight, to irregularly stepped or slightly curved. Only a few blocks show contacts that are gradational on a millimeter scale, characterized by a mixture of fine-grained block material and material of the granitic host (Fig. 7C).

Commonly, blocks with compositional layering have their long axis oriented parallel to this anisotropy in the block. The long axis and thus the bedding of many blocks are oriented subparallel to the local strike of the magmatic foliation, which in places is subparallel to the contact of the May Lake pendant (e.g., in Fig. 6 orientations are visible for the larger blocks). Thus, these blocks may in places appear to be rafts (sensu Glazner and Bartley, 2006), but systematic changes in their long axes orientations closely match changes in magmatic fabric orientations and not necessarily always in situ host-rock structure, indicating that they are displaced blocks, not rafts. In support of this, other metasedimentary blocks have random orientations both with respect to each other and to the main magmatic foliation, and are clearly rotated. In detail, the magmatic foliations in the granodiorite may be (1) subparallel to the contact with the metasedimentary blocks, thereby wrapping around the margins of the blocks (observed mostly around the large blocks), (2) subparallel or oblique, without changing its strike around the block, or (3) irregular (mostly observed around the small- to medium-size blocks).

These metasedimentary blocks are commonly immersed in, rimmed by, and injected by aplite leucogranite. Thin dikes of Kuna Crest or Half Dome Granodiorite occur rarely within the blocks. Veins and dikes in the blocks are typically parallel to metamorphic layering (i.e., transposed bedding), but also cut across this layering at high angles and exerted some control on the continued disintegration of the blocks (Figs. 7C, 7D).

**Block Formation and Disaggregation in Jackass Lakes Granodiorite**

A number of roof and/or intraplutonic metavolcanic, and less commonly metasedimentary, pendants exist in the Jackass Lakes pluton (Fig. 8A). Each pendant typically features numerous host-rock xenoliths nearby (Figs. 8B–8D and 9) that decrease in number with distance from the pendant (McNulty et al., 1996; Krueger et al., 2004; Wolak, 2004; Wolak et al., 2005; Krueger, 2005; Pignotta et al., 2010). The distribution and characteristics of stope blocks preserved around one of the large pendants are the focus of previous studies (Wolak, 2004; Krueger, 2005; Yoshinobu et al., 2009).
Figure 8 (on this and following page). (A) Map of Jackass Lakes pluton after Pignotta et al. (2010). See Figure 1 for location. Note large pendants and smaller blocks in the Jackass Lakes pluton. UTM—Universal Transverse Mercator coordinates.
Figure 8 (continued). (B) Grid map of metavolcanic block formation in Jackass Lakes granodiorite showing vein and dike growth and multiple intrusions into dikes along block margins. JLP—Jackass Lakes pluton. (C) Photo of metavolcanic block in Jackass Lakes pluton with magmatic layering along block margins. Hammer is ~30 cm long. (D) Highlighting of C showing metavolcanic block (green), leucocratic granitic material along block margin (orange), and two sets of magmatic layering that formed along block margins (dashed and dotted black lines). Some erosion of older layers and crosscutting relationships indicate that dotted layers formed first, then the dashed. Both sets of magmatic layering contain pieces of partially melted (?) metavolcanic block material forming diffuse metavolcanic fragments. Graded magmatic layers form where more mafic material accumulated at the bottom of the layer. Magmatic foliation crosscuts all structure (lithologic contacts, magmatic layering) and strikes approximately north-south in photo.
Here we focus on the intrusion of magma into pendants (Figs. 8B–8D and 9), formation and initial movement of blocks along pendant boundaries (Figs. 8B–8D), and the continued disintegration and new block formation of a metavolcanic xenolith in the Jackass Lakes pluton ~100 m away from the nearest pendant (Fig. 10).

**Block Formation**

As in the metasedimentary blocks at May Lake, we see less preserved evidence of focused porous flow in these metavolcanic pendants, although some features in Figures 9C, 9E, and 9F may reflect former examples of this. More commonly cracking and melt intrusion by channelized magma flow resulting in vein and dike networks are preserved. Dike intrusion into metavolcanic rocks has formed both rectangular (Figs. 9A, 9C, 9D) and triangular (Figs. 9B, 9D) blocks identical to those seen in all previously discussed localities (e.g., cf. Figs. 3, 5, and 7). Vein and dike networks in metavolcanic rocks show intimate relationships to either previous host-rock structures or structures actively forming at the time of melt injection, in which case the veins typically intrude along and at high angles to shear surfaces (Figs. 9E, 9F), or are associated with folds (Figs. 9G, 9H). In most cases, thin veins (less than several centimeters wide) typically consist of fine-grained, leucocratic, quartz-rich material (Figs. 9E–9H). As veins widen into dikes, the granitic material coarsens and becomes more mafic due to the addition of biotite and hornblende (Figs. 9A–9D). Thin, fine-grained, leucocratic margins to these thicker dikes (e.g., Fig. 9A) may represent original leucocratic veins that initially intruded along the cracks. The breaking off and movement of small host-rock xenoliths began during channelized magma flow in dikes (Figs. 9C, 9D).

We have examined in detail two cases of the initial displacement of metavolcanic blocks into the Jackass Lakes pluton magma chamber (Figs. 8B–8D). Figure 8B shows a map of a...
Stoped block formation

The smallest veins in the pendant are dominated by leucogranite and intrude subparallel, and at high angles, to the metamorphic foliation in the metavolcanics. These leucocratic veins grade into mafic and/or felsic layered sequences in the largest dike in the pendant, and into similar layered sequences on three sides of the block. Small metavolcanic xenoliths are locally preserved in these layered sequences. Topographically above the block, some of the mafic layers show magmatic load casts and draped layers, indicating that these layers were locally sinking due to gravity relative to nearby layers (e.g., Wiebe and Collins, 1998). The load casts presumably reflect a density difference between the mafic and felsic layers; the draped layering may reflect downward warping caused by motion of the block.

Some metavolcanic blocks entirely enclosed by Jackass Lakes plutonic rock also have spectacular layered sequences preserved along their margins (Figs. 8C, 8D) similar to those described in the Half Dome lobe and Potter Point examples. Leucogranite is locally preserved along the block margins and in one vein in the block (Figs. 8C, 8D). Local erosion of older layers and crosscutting relationships indicate largely unidirectional growth (and block movement) during formation of the layered sequences.

The presence of layered magmatic sequences in settings, where blocks have initially been cracked and only locally translated, suggests that the layered sequences now preserved around rotated blocks away from pendant margins also record the initial formation of these blocks, after which the layers remained welded to the blocks. If so, then the example in Figures 8C and 8D shows the incremental displacement of the block initially to the left in the photo, and then simultaneously both downward and left during formation of the second layer set.

Figure 9 (continued). Field photos from Jackass Lakes area. Hammers (30 cm long) or rulers (15 cm long) for scale. (E) Small leucocratic veins of Jackass Lakes pluton melts in metavolcanic host rocks (near eastern margin of Jackass Lakes pluton) in which veins were intruded during active deformation of the host rock. (F) Small leucocratic veins in metavolcaniclastic host rock that was undergoing folding. (G) Granodiorite (eastern margin of Jackass Lakes pluton) full of metavolcanic host-rock fragments. (H) Large magmatically folded dike of Jackass Lakes pluton in deformed metavolcanic block in Jackass Lakes pluton.
Block Disintegration and/or Formation in the Magma Chamber

We have examined cracking, vein formation, and further block development in one meta-andesitic block located ~100 m from the margin of the Post Peak pendant (Figs. 8 and 10A), and some of the rheological implications of this work (presented by Yoshinobu et al., 2009). The ~45 m × 25 m block is only exposed on a glacially polished subhorizontal surface, although jointing provides local three-dimensional control on the centimeter to 0.5 m scale. On the subhorizontal exposure, the long axis of the block strikes ~335°. The block is metamorphosed and deformed with a well-developed foliation oriented ~325°/83° defined by aligned porphyroclasts of plagioclase and hornblende, and by aligned biotite and quartz in the fine-grained groundmass.

Particular attention was paid to the margins of the block (Figs. 10B–10D). In general, at the meter to centimeter scale, these margins are defined by straight to slightly curved segments connected by gentle to angular corners where the margin abruptly changes directions (Figs. 10B, 10C); in places the corners approach 90°. Both types of margins are suggestive of the formation and removal of smaller irregular pieces from the larger block. Some of these smaller pieces are locally preserved and can be fit back to the larger block (Fig. 10D). The straight margin segments in both the larger block and smaller pieces are in places parallel to, but also cut across, the subsolidus foliation in the block, indicating that during cracking and formation of both the original block and smaller pieces, the anisotropy (foliation and lineation) in the block did not exert a first order control on crack orientation.

Figure 10 (on this and following page). Field photos of disaggregation of meta-andesite block near Post Peak pendant (see Fig. 8A for location). Orange ruler for scale is 15 cm long. (A) Overview of southern margin of Post Peak pendant and nearby isolated block in Jackass Lakes pluton (looking west in photo). Isolated block is 45 × 25 m. (B) Stepped margin of metavolcanic block in leucocratic Jackass Lakes pluton magma. (C) Close-up of stepped margin with leucocratic Jackass Lakes granodioritic magma in steps. Note small vein of leucocratic material in metavolcanic block that initiated at block corner. (D) Small stoped blocks ripped off margin of larger block during intrusion of Jackass Lakes pluton magma. Figure adapted from Yoshinobu et al. (2009).
In general, the host granodiorite does not record a dramatic change in composition or microstructures as the block is approached. However, locally more leucocratic patches occur along outward-facing steps or in thinly layered (defined by modal differences in mineralogy) sequences in small zones along straight margins (Fig. 10C). These layers are different from the veins that now intrude the block; the veins gradually taper to thin tips (Figs. 10C, 10G, 10H). These veins can be traced continuously across the block-magma margin and merging with Jackass Lakes pluton matrix. Also note cuspate-lobate metavolcanic margin of block. (G) Three different vein sets of Jackass Lakes pluton melts in meta-andesite block. Note example near ruler where vein changes direction from one set to another. (H) Same veins in G showing boudinage or folding as function of orientation relative to foliation in metavolcanic block. Figure adapted from Yoshinobu et al. (2009).

Figure 10 (continued). Field photos of disaggregation of meta-andesite block near Post Peak pendant (see Fig. 8A for location). Orange ruler for scale is 15 cm long. (E) In situ leucocratic vein that drained from Jackass Lakes pluton into metavolcanic block, implying that block did not move after vein formed. (F) Small leucocratic veins cutting across block-magma margin and merging with Jackass Lakes pluton matrix. Also note cuspate-lobate metavolcanic margin of block. (G) Three different vein sets of Jackass Lakes pluton melts in meta-andesite block. Note example near ruler where vein changes direction from one set to another. (H) Same veins in G showing boudinage or folding as function of orientation relative to foliation in metavolcanic block. Figure adapted from Yoshinobu et al. (2009).
veins to be separated into three groups, although there are gradations in each (Figs. 10G, 10H). The oldest set is parallel to the foliation in the block, the second is perpendicular to this foliation, and the third is at a high angle, but often not exactly perpendicular, to the foliation. The third set is also less common and tends to be thicker, in one case ranging to 15 cm thick. Vein tips (or fractures in front of tips) curve away from their normal orientations in some places (Figs. 10G, 10H), and in extreme cases single veins of the first two sets abruptly change from foliation-parallel to foliation-perpendicular orientations or vice versa (Fig. 10G), and form bridging veins. Veins range from strongly deformed, largely in the magmatic state, to relatively undeformed (Fig. 10), the style of deformation (boudinage versus folding) being a function of the vein orientation relative to that of the block foliation. This deformation of veins in the block, continuity of veins into the surrounding plutonic groundmass around the block, and presence of small pieces of the block along its margins establish that the block was continuing to deform internally and disintegrate along its margins while in a crystal mush.

**DISCUSSION**

The most commonly proposed process for stoped block formation is differential heating and thermal expansion cracking. As the host rock is heated, the ensuing thermal gradient results in stresses as high as 0.2–0.4 GPa, which may lead to cracking (Marsh, 1982; Furlong and Myers, 1985; Clarke et al., 1998). In isotropic material, these cracks would typically form parallel to the host rock–magma contact (i.e., perpendicular to thermal gradients). Clarke et al. (1998) noted that cracks in other orientations may form due to differential heating controlled by various anisotropies. Alternatively, cracks at high angles to the magma–host rock contact may form as a result of thermal expansion leading to host-rock buckling, which controls cracking (Chyi et al., 1985; Spence and Turcotte, 1985; Clarke et al., 1998; Pignotta et al., 1999; Dumond et al., 2005; Farris and Paterson, 2007). Thermally driven cracking may be a rapid and potentially catastrophic process, particularly in cases where colder blocks are rapidly transported into hotter portions of a magma chamber (Clarke et al., 1998; Pignotta et al., 2001a; Hawkins and Wiebe, 2004; Farris and Paterson, 2007). Finite difference thermal modeling (Paterson and Okaya, 1999; Pignotta et al., 1999, 2001a; Paterson et al., 2007b) indicates that cracks parallel to the magma–host rock contacts, greatest dilation toward the magma chamber, and the rectangular to triangular shapes of blocks are all permissive of thermal stress–driven cracking. The rates of conductive and convective block heating indicate that this process of thermal cracking, in most cases, must have occurred in less than hundreds of days, and possibly was very rapid (Paterson et al., 2007a).

However, our observations indicate that simple thermal shattering is not the only, and possibly not the most common, mechanism of stoped block formation, and in the following we summarize an alternative model.

**Cracking and Initial Vein Formation**

In regions in the central Sierra Nevada where we examined block initiation, one of two sequences occurred, depending on host-rock lithology, during initial vein formation around developing host-rock blocks. In plutonic host rocks relatively planar zones of microcracking and/or increased porosity resulted in the growth of magma fingers by focused porous flow (Fig. 11). These magma fingers grew and increasingly intersected along the planar zones, eventually leading to continuous magma channels along block margins through which magma moved (channel flow; Figs. 3, 5, and 10). In some cases where cracks were already well developed, leucocratic melts may have intruded directly by channel flow (Fig. 3).

In metavolcanic and metasedimentary host rocks, we see only minor evidence of focused porous flow having formed magma fingers. Instead, many veins seem to have formed by channel flow along fractures (Figs. 7, 9, and 10). In both sequences, relatively leucocratic melts almost always formed the veins or magma fingers (now preserved in thin vein tips and along the margins of dikes; see Figs. 3, 5, 7, 9, and 10), followed by the intrusion of more intermediate magma compositions with greater amounts of biotite, hornblende, and plagioclase.

We suggest that the main differences between these two sequences result from the presence of small amounts of melt in the host plutonic rocks (in both examples the host granitoids are only slightly older than the intruding magma and either had remaining melt or were potentially reheated above their solidus), their weak anisotropy, and possibly their coarser grain sizes. In contrast, metamorphic host rocks have moderate to strong anisotropies due to compositional layering, moderately to intensely developed foliations, folding and shearing, and finer grain sizes. Locally we see evidence for the formation of small amounts of melt in some of the metamorphic host rocks (particularly metapelites) as small leucosome lenses aligned with the foliation or as small pools in low stress sites (Figs. 7 and 9). Whether focused porous flow played a role in the collection of these melts is unclear.

Determining what drove crack formation and/or increased porosity is a challenging problem, requiring that the following field observa-

**Figure 11. Suggested block formation. (A) Zones of focused porous flow of melts (pink) into host rock with intact bridges of host rock between melt channels. (B) Connected channel flow formed by expansion and connection of porous flow zones shown in A. (C) Formation of multiple magma sheets and/or dikes as host blocks are displaced laterally. Growth directions of sheeted complexes along block boundaries are typically unidirectional. (D) Rotated host-rock block surrounded by unlayered magma. Former sheeted zones and internal dikes are now frozen to block and thus are also rotated. New melt veins (not shown) may form as block stops moving (e.g., Figs. 10E, 10F).**
tions be explained: (1) dilation of the host rock in three dimensions, resulting in a local volume increase, but with the largest direction of expansion typically toward the magma chamber; (2) the formation of rectangular-to-triangular-shaped blocks in map view; (3) the presence of many fractures at various angles to the foliation in moderate to weakly anisotropic rocks, and more often parallel and perpendicular to the foliation in strongly anisotropic rocks; (4) the ample evidence of active host-rock deformation during block formation; and (5) the formation of small melt lenses in some host rocks, probably caused by local melting, during block formation.

We discuss several likely mechanisms of initial crack formation, in addition to the thermal heating discussed herein, with the caveat that they all possibly played some role during host-rock cracking, leading to complex interactions between them. We also note that host rocks had preexisting cracks and/or heterogeneities, which provided weaknesses that any of these processes may have selectively utilized.

Regional ductile deformation was active during block formation in all four of our examined areas, and in many cases continued to affect the newly formed blocks in the magma chambers (Figs. 7, 9, and 10). Evidence of regional deformation in the plutonic host rocks includes (1) regional magmatic foliations and steep lineations that overprint internal contacts with the foliations being axial planar to folded dikes and igneous layering; (2) local magmatic folds and faults with consistent orientations; (3) magmatically folded and boudinaged veins and dikes in stope blocks; and (4) boudinage of host rock in directions parallel to magmatic lineation (see also Fowler and Paterson, 1997; Culshaw and Bhatnagar, 2001; Yoshinobu et al., 2009). Metamorphic host rocks preserve evidence of folding, boudinage, and shear that in some cases involve dikes of the nearby magma source and/or locally formed leucosome material. Some of these structures are older, but the orientation, geometry, and timing of others are usually continuous with nearby magmatic structures and thus are interpreted to be coupled, implying deformation of the magma and host rocks during block formation (Paterson et al., 1998; Žák et al., 2007).

There are a number of interesting implications of active regional deformation during block formation. The first is that it clearly played an important role in controlling the development of melt pathways through the host rock, and in the breaking apart of host rock (e.g., to form boudins in pendants and smaller block fragments in plutons; Figs. 7, 9, and 10). That is, regional deformation aided block formation. Second, as regional deformation is a relatively slow process, occurring over longer durations than, for example, thermal heating, block formation is not always a rapid process. We have estimated strains associated with syn-emplacement regional folding and boudinage, and determined that as much as 30% shortening took place before the magma froze, a process that probably occurred over $10^4$ to $10^5$ years, or at least significantly longer than the days to years over which steep thermal gradients could have been maintained (Paterson et al., 2007; Paterson et al., 2009). In cases where regional deformation in host blocks and magmas were coupled, we can infer that the magmas transmitted deviatoric tectonic stresses and did not have significant viscosity contrasts with the host-rock blocks, and thus were fairly crystal rich. Therefore, stope block formation and disaggregation were processes that continued even though magmas had high crystal percent. Particularly clear examples of this exist in the Mitchell Peak pluton (Pignotta and Paterson, 2007) and the Jackson Lakes pluton (Yoshinobu et al., 2009).

Locally, heating and regional deformation worked together to form lenses of leucosomeal melts, which led to the loss of cohesion along planes of compositional layering (e.g., between metapelitic and quartzite layers in the May Lake screen) and along actively forming structures (fold axial planes, shear zones; Figs. 7 and 9). This aided in both block formation from in situ host rock and block disaggregation in magma chambers. Nevertheless, it appears to have been a fairly uncommon process in the areas examined.

Another likely process for stope block formation is magma injection during crack propagation, driven by changes in magma pressure in chambers (Rubin, 1993, 1995). Magma pressure is likely to increase and decrease as new magma is injected into a chamber, as magma is removed (transferred to other chambers or lost during volcanic eruptions), and as the magnitude of regional stress on a chamber fluctuates. For example, Hawkins and Wiebe (2004) suggested that stoping events were related to a cycle of mafic magma injections leading to volcanic eruptions followed by a reduction in magma chamber pressure. We certainly see widespread evidence of diking at all scales, and a number of dikes can be traced into chambers where they merge with the host granitoid, indicating that melts drained from the chamber into the host rock by channel flow via these dikes (Figs. 3, 7, 9, and 10). However, it remains difficult to determine if a change in magma pressure drove dike formation, or if another process formed low-stress sites, which in turn drew in melts (e.g., Clarke et al., 1998); large dikes that propagate tens to hundreds of meters and crosscut a number of features probably represent the former, whereas thin, discontinuous veins may represent the latter (Figs. 7, 9, and 10). Rubin (1993, 1995) concluded that host-rock cracking is easiest when a higher thermal gradient exists, whereas initiation and propagation of a dike along a chamber margin is easiest when a lower thermal gradient exists (i.e., when the host rock is hot, growing dike tips will not freeze rapidly). Thus, thermally driven cracking is likely to be the dominant process during initial high thermal gradients, but is replaced by magma pressure-driven diking as the host rocks thermally equilibrate. Where magma flow in dikes or chambers occurs (Figs. 5, 7, and 9), we suggest that viscous shear coupling between the flowing magma and host rock can also lead to new block formation.

Clarke et al. (1998) explored the potential importance of host-rock anisotropy, at both microscale and macroscale, during thermal heating and block formation. Our studies indicate that the relatively weak anisotropy resulting from mineral alignment in both plutonic and metamorphic host rocks was not a first-order control on block formation, as initial cracks and final block margins were commonly formed at angles to these mineral foliations (Figs. 3, 5, 9, and 10). However, where foliation intensities increase, and particularly where metamorphic layering exists (often transposed and recrystallized bedding), anisotropy became an important factor and block margins formed either parallel or perpendicular to these anisotropies (Fig. 7). We suspect that the differential thermal expansion and different effective viscosities of these metamorphic layers played an important role during block formation.

In summary, we see evidence for a number of block formation processes (heating, deformation, local melting, dike injections) that were locally affected by the presence of strong anisotropies (metamorphic layering and mineral alignments). These processes temporally overlapped, but operated at different rates to form cracks and initial melt channels along block boundaries.

**Growth of Sheeted Zones and Resulting Block Translation**

Some of the described processes may result in the rapid formation and displacement of stope blocks, such as thermal shattering or viscous flow coupling. However, we found a number of cases where complex sheeted zones first formed along block margins (Figs. 3H, 5E, 5F, 8C, and 8D), thus supporting the incremental pulsing of magma into growing cracks and/or dikes as blocks were incrementally translated. Growth
directions of sheeted zones tended to remain consistent (new pulses were either emplaced always along the host rock or the block margin), analogous to a crack-seal mechanism of vein formation (Ramsay, 1980; Cox, 1987). The presence of schlieren-bounded troughs, truncation of these troughs by erosion during magma flow, and redeposition resulting in mineral grading in schlieren, all point toward a complex interplay between the intrusion of multiple pulses, mineral sorting, and melt fractionation within pulses during the growth of these zones. These sheeted zones sometimes remained welded to block margins as the blocks begin to rotate in and move through magma chambers. Alternatively, sheeted zones may continue to form around isolated blocks, i.e., at significant distances from any preserved in situ host rock (Fig. 8C).

These sheeted zones also indicate that initial block displacements typically occurred incrementally over some extended duration, in contrast to the rapid or catastrophic process associated with block formation by thermal shattering or sudden roof collapse. Sheeted zones vary in width from tens of centimeters to 0.5 km (Figs. 3H, 5E, 5F, and 8C; see also Clarke et al., 1998; Paterson et al., 2009), suggesting that initial magnitudes of incremental displacement around some blocks vary tremendously before these blocks freely move and rotate in chambers. These rafts clearly evolved into rotated stoped blocks in the regions we examined; the best examples are given by blocks in which both the block structures and some preserved sheeting along block margins were truncated by the surrounding magma, these structures being at angles to comparable structures in nearby in situ host rocks. Even in cases where sheet-bounded host rocks are less clearly rotated, we do not view them as in situ rafts since the development of the sheeted zones requires displacement of the host rocks (Paterson et al., 2008). We can only think of one way to form nondisplaced rafts now surrounded by magma: i.e., to remove vertically or laterally host rocks between the rafts out of the present plane of exposure, which could only occur by melting (an unlikely process in most cases) or by a stoping-like process.

**Block Detachment and Rotation**

Some small host-rock blocks appear to have immediately started to rotate in cases where they formed during channel flow in dikes or along boundaries of larger blocks isolated in magma chambers (Figs. 3, 7, and 10). However, the fact that sheeted zones formed along many block margins suggests that other blocks did not immediately freely rotate in magma, and rotation here is first observed only when a fairly large (relative to the block size) pool of magma formed around blocks (Figs. 2C, 4C, and 5H). These cases of block detachment and rotation likely represent areas where the flux of magma around the developing blocks was high, prohibiting the formation of layered sheeted zones, and/or the surrounding magmas were hot and crystal-poor and thus had low effective viscosities.

**CONCLUSIONS**

Stoping is commonly associated with the growth of magma chambers in the Sierra Nevada batholith, although its magnitude in any one system remains poorly constrained. Through an examination of the initial formation of stopped blocks, we record the interplay between focused porous flow, high-temperature cracking, channelized magma flow, and host-rock deformation. The Tuolumne Batholith and Jackass Lakes pluton both preserve evidence for the following evolution of block formation (Fig. 11).

1. In fairly isotropic plutonic host rocks, low-stress sites developed leading to planar zones of increased porosity in which focused porous flow resulted in magma fingers of first felsic, volatile-rich(?) melts, followed by more intermediate composition melts. Growth of fingers along fairly planar regions resulted in amalgamation into dike-like zones, in which channelized magma flow occurred, and sheeted zones formed along block margins. Weak host-rock anisotropies in these settings did not play a significant role in block formation.

2. In anisotropic metamorphic host rocks, cracking and channel flow, and less obviously focused porous flow, led to block formation. Host-rock anisotropy (e.g., subparallel mineral alignment and metamorphic layering) only played an important role when strongly developed.

3. Block shapes are usually rectangular, less commonly triangular, and vary greatly in size. During initial block formation, three-dimensional dilation occurred, although the dominant direction of expansion was inward into the nearby magma chamber.

4. The driving forces for dilation remain uncertain, but likely reflect the interplay between thermal stress gradients formed during heating, regional deviatoric stress leading to host-rock strain at slower rates than heating, and fluctuating magma stresses. Host-rock properties only played a secondary role and were most important in strongly layered host rocks.

5. Once formed, some blocks began to move immediately if involved in flow in dikes or during block disintegration in chambers. However, many blocks initially were incrementally displaced by repeated magma injections along their margins, resulting in sheeted complexes akin to crack-seal vein growth in metamorphic rocks. These sheeted complexes typically have consistent growth directions and preserve evidence of magma erosion and redeposition during injection, suggestive of an incremental process that occurred over an extended duration. Block rotation in this case only occurred once sufficient amounts of magma surrounded the host block. In some cases segments of the sheeted zones remain attached to these rotated blocks.

6. Many blocks continued to disintegrate, and in doing so formed new smaller blocks by the same block-formation processes discussed here. The common pattern of decreasing size and number of blocks with distance from the in situ host rock (see also Wolak, 2004; Krueger, 2005) is a reflection of this process of block disintegration and results in a fractal distribution of block sizes (Farrirs and Paterson, 2007).

This study has several important implications. First, it adds detail to a number of clear examples of stopping both within and along pluton margins in the central Sierra Nevada and thus to the growing number of publications that clearly indicate that stopping is an important process during magma chamber growth. It also further establishes the occurrence of stopping along internal contacts within magma chambers (e.g., Paterson et al., 2008); this may be significant giving the growing popularity of forming magma bodies through the incremental addition of many magma pulses and the recognition of mixed crystal populations in magmatic and volcanic systems (Davidson et al., 2007; Lackey et al., 2008; Krause et al., 2009). The magnitude of these processes and thus the effects on the evolving magma systems remain poorly constrained; however, the widespread evidence of stopping certainly indicates that it is important to understand these effects. As noted by Gerbi et al. (2004), even if only ~20% of the removal of host rock required during magma emplacement occurs by stopping, it may largely remove the very host rock that had preserved information about other emplacement processes.

Second, block formation may not always be a rapid thermally driven process, but instead may involve longer durations of the incremental formation of blocks and incremental displacement into the magma chambers, during which local magmatic sheeted complexes formed around blocks, and regional deformation of both the host rocks and magmas occurred. The absolute time scales remain uncertain. However, the observations that formation of the sheeted complexes involves growth of magma fingers leading to repeated channel flow events, significant crystallization, erosion of crystal mushes, and continued flow and redeposition of crystals, plus the
recognition that regional deformation can play a role in block formation, indicate that these durations are likely $10^4$ to $10^5$ years, or at least much longer than the days to hundreds of years over which a steep thermal gradient could have been maintained in these settings. One important implication of this is that the metamorphic host rock or slightly older plutonic rock being processed and moved into magma pulses is already heated, and will thus not have as direct an effect on cooling on the magma pulse it is in, as implied by strictly thermally driven stoping. Another implication is that it allows for more possibilities regarding when and where stoping may occur in magmatic systems, since steep thermal gradients are not required.

Third, our studies indicate that a very complex interplay between both brittle and ductile processes resulting from regional tectonism and more localized magmatic processes occurred during these high-temperature and moderate-pressure conditions. Our studies add to the growing awareness that dike-like or sill-like bodies often grow or expand through the formation and amalgamation of magma fingers (e.g., Pollard et al., 1975; Schofield et al., 2010). Host-rock properties also influence this complex interplay, the three most important being bulk composition, presence of melt in the host rock, and intensity of anisotropy. To fully understand the effects on the physical and chemical evolution of magmatic systems of the style of stoping described herein, both between magmatic units within plutons and of metamorphic host rocks along pluton margins, we need to better understand the behavior of host-rock blocks in chambers. This study indicates that the continued mechanical breakdown of these blocks is the dominant process (see also Clarke et al., 1998), and that thermal cooling in these multipulsed magmatic systems may be less a factor than previously believed.

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