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Reply to the comment by Scott Paterson on "Relationships between melt-induced rheological transitions and finite strain: Observations from host rock pendants of the Tuolumne Intrusive Suite, Sierra Nevada, California" by Markus Albertz

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Paterson (2007) commented on my recent article on finite strain and microstructural analyses in the eastern wall rocks (Saddlebag Lake pendant, SLP) of the Tuolumne Intrusive Suite (TIS) that describes an increase in finite strain towards the pluton contact and an associated transition in deformation mechanisms from dislocation creep to melt-assisted granular flow (Albertz, 2006). Below, I condense and address the points in the order that they are raised in the discussion.

(1) The data presented (i.e., Fig. 7 in Albertz, 2006) do not support that the finite strain increases drastically or abruptly toward the eastern margin of the TIS.

Viewing Fig. 7 (strain transects perpendicular to the eastern TIS margin, SLP) in conjunction with the accompanying text suggests that the characterization of the strain gradient in the SLP as drastic and abrupt is adequate:

> Within ca. 100 m to the contact, strain intensities increase abruptly and consistently from an average background value of ca. 43% to ca. 70% (Fig. 7A) and ca. 85% (Fig. 7B), respectively. 82–83% shortening is also observed between the strain transects (Albertz et al., 2005). Closest to the pluton, low to intermediate strain intensities are absent and in general the rocks are so highly shortened that strains often cannot be quantified fully because outcrop heights are shorter than long axes of strain markers. Thus the high proximal strains are minimum estimates (Albertz, 2006, p. 1431).

This excerpt shows that: (1) 82–83% shortening near the contact is also observed between the strain transects; (2) only high strains occur near the pluton margin; there are no low to intermediate strains; and (3) the highest strains near the TIS margin exceed 85% shortening.

Paterson (2007) argues that a strain increase of only ca. 5–10% can be verified. A thorough assessment of the data (Fig. 7, Albertz, 2006), however, shows that this statement applies only when the high proximal (near the pluton) amount of z-axis shortening is selectively compared against the highest distal (background) amount of z-axis shortening. In transect A (Fig. 7A; Albertz, 2006), the high proximal and highest distal amounts of z-axis shortening are 70% and 65%, respectively. In transect B (Fig. 7B; Albertz, 2006), the corresponding values are 85% and 65%, respectively. Hence, the increase in z-axis shortening toward the pluton margin is 5–20% but this comparison ignores that the lower amounts of distal z-axis shortening also increase to 85% toward the TIS margin, and this increase is more than that for the highest distal strains. The lowest amounts of distal z-axis shortening along transects A and B, respectively, are 20 and 25% which yields an increase in z-axis shortening of 50–60%. Both

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approaches are to be avoided, however, because neither provides representative values. In my article, I chose to use the more meaningful mean value of the amount of distal z-axis shortening which is ca. 43% and it means that the average increase in z-axis shortening toward the TIS margin is ca. 27–42% (Albertz, 2006).

Whereas 27% z-axis shortening (transect A) is significantly less than 42% (transect B), the field relationships suggest that an equal amount of shortening probably occurred in transect A but was subsequently removed:

> … a portion of the inner aureole, including stratigraphy and the corresponding finite strains, is missing. Notice that rocks of the Horse Canyon sequence as well as older TIS units are truncated at a high angle to strike by Cathedral Peak granodiorite (Fig. 3A). As a result, z-axis shortening at the margin in transect A is 70% (Fig. 7A) whereas 85% shortening was preserved in transect B (Fig. 7B). Also notice that the width of the domain of high-temperature microstructures along transect A is only ca. 1/3 the equivalent width in transect B (Fig. 9) (Albertz, 2006, p. 1440).

(2) Regarding the strain gradient measured in the eastern TIS aureole, the following two processes are downplayed: (1) strain caused by movement in regional shear zones (e.g., Greene and Schweickert, 1995; Tikoff and de Saint Blanquat, 1997), and (2) tectonic strain occurring during or after pluton emplacement as indicated by magmatic fabrics within the plutonic rocks.

The shear zones described in the papers quoted by Paterson (2007) are the Gem Lake (Greene and Schweickert, 1995) and Rosy Finch (Tikoff and de Saint Blanquat, 1997) shear zones which lie in the Northern Ritter Range pendant and the Mono Pass Intrusive Suite, respectively, both of which are located south of the SLP. Pertinent to the SLP, the Cascade Lake shear zone is proposed to be associated with dextral strike-slip displacement and shallowly (Tikoff and Greene, 1997) to moderately (Tikoff et al., 2005) plunging lineations. Together, these shear zones are believed to be part of a through-going shear zone (Sierra Crest Shear Zone System, Tikoff and de Saint Blanquat, 1997).

However, my field observations are inconsistent with dextral shear in the SLP in the manner proposed by the authors. For example, lineations in the SLP are predominantly subvertical, the kinematic indicators in the contact zone in the SLP are highly variable and neither consistent with dextral nor with sinistral shear (Albertz and Paterson, 2002; Albertz, 2006), and there is no evidence for syn-intrusive strike-slip shear zones immediately north and south of the type locality at Cascade Lake (Albertz, 2006, p. 1442). If future work provides more convincing evidence for a crustal-scale dextral strike-slip shear zone through the SLP (although I have examined the entire contact zone and have not found this evidence) the strain measurements (Albertz, 2006) are equally valid. Then, however, it becomes relevant to examine feedback processes between melt and deformation and contemplate whether the presence of melt in the SLP triggered shearing or whether a crustal shear zone accommodated channeling of melt, or perhaps both.

Paterson (2007) claims that I concluded that aureole strain largely reflects emplacement. This is an incorrect statement. I explicitly discuss possible causes for the strain gradient in the SLP and ultimately conclude that the data are best consistent with regional deformation, that is, regional contraction and/or weakly dextral transpression (Albertz, 2006, pp. 1439–1440; Albertz, 2004, unpublished doctoral dissertation). With regard to the role of emplacement-related strain, the strain data are compared against the exact requirements of emplacement-related strain (to the extent that this is justifiable using previous studies that convincingly identified emplacement-related strain in aureoles) and typical emplacement-related flattening strain is in fact excluded. However, I discuss that if a flattened aureole once existed it may have been removed by stoping and that, under the limiting condition that emplacement-related strain (if ever developed) occurred under plane strain conditions, it would not be possible to separate it from the regional (plane) strain on the basis of finite strain analysis (Albertz, 2006, p. 1440).

Paterson (2007) suggests that regional strain affected the TIS during and after its emplacement and that this strain also affected the weaker rocks in the contact aureole. In view of my conclusion that the strain gradient in the SLP is most likely associated with regional deformation (Albertz, 2006, pp. 1439–1440), and that emplaced yet unsolidified magma clearly continued to deform (magmatic foliations in the Kuna Crest, Half Dome and Cathedral Peak granodiorites, Albertz, 2006), the purpose of raising this issue is unclear.

The above comment is based on a definition of emplacement that post-dates the publication of my article, and as such is not a valid critique. Žák et al. (2007) define emplacement as:

> the construction of the full dimensions of a pluton (i.e., all host rock has been displaced), which, however, does not necessarily imply the completion of internal processes operating within the remaining magma chamber prior to its full crystallization.

However, emplacement can also be understood as the sum of all processes that occur until a magma chamber is fully solidified. For example, magma injection by diking often continues through the latest stages of an intrusion such as the TIS, and strictly, this is also a form of emplacement by Žák et al's (2007) definition, although at a smaller scale (e.g., magmatically folded dikes in the TIS, Albertz et al., 2005). Furthermore, emplacement of magma into the crust is also associated with the emplacement of a heat anomaly which in turn affects the rheology of the adjacent rocks. Consider that magma is emplaced in a regional contractional strain field and assume that the magma chamber does not submit any significant stresses to the surrounding rocks. Even in such a scenario, the magma-induced thermal gradient would result in higher levels of finite strain near the margin of the intrusion because the rates of viscous flow increase as the temperature rises. In any case, the exact definition of emplacement is inconsequential with regard to the inferred cause for the strain gradient in the SLP, that is, regional deformation. To put it another way, there is clear evidence for syn-intrusive (or syn-magmatic) regional deformation but the presence of magmatically folded dikes and melt-bearing microstructures (Albertz et al., 2005; Albertz, 2006) exclude significant post-intrusive regional deformation which would have resulted in a sub-solidus overprint of the magmatic microstructures.

(3) Melt, and in particular in situ melting of the host rock, is unlikely to have played the most important role in weakening the strength of the host rock.

Based on the observation that the first appearance of melt-bearing microstructures coincides with increasing strain towards the TIS contact I infer in my article (Albertz, 2006, p. 1440) that the presence of melt had a first-order effect on the strength of the host rock (also Albertz, 2004, unpublished doctoral dissertation, pp. 126–127). Paterson (2007) argues that a number of (additional) parameters and processes, such as temperature, grain size, anisotropy, fluids, the presence of melt derived from the pluton, and in situ melting contribute to weakening of host rocks. I regard all of these additional effects to be potentially important but likely to be of second order and/or causatively related. For example, increasing temperatures toward the TIS margin are consistent with partial melting (Albertz et al., 2005, p. 453) and the generation of partial melt may have reduced grain size by cracking (Albertz, 2006, pp. 1438–1439).

Paterson (2007) states that his outcrop scale examination of melt-veins in the host rock indicates that they are derived from the batholith, rather than by in situ melting, whereas the microscopically documented widespread occurrence of former interstitial melt pockets, melt films that coat grains, and rounded grains that resemble corroded reactant minerals suggest in situ partial melting (Fig. 8, Albertz, 2006). The truth remains to be determined by additional field, microscopic and geochemical work. Regardless of the outcome, however, the role of melt in weakening the host rock by triggering melt-assisted granular flow (Albertz, 2006) will be equally valid. I view the introduction of melt to any deforming system as a critical event that profoundly changes the rheological behavior regardless of how the melt was actually delivered to the system. However, it may be appropriate to adjust Fig. 10 in my article (Albertz, 2006) by replacing ''partial melting'' with ''introduction of melt'' so that both processes are equally considered.

(4) Uncertainty as to what type of markers were used in Fry analysis.

Paterson (2007) quotes from my article (Albertz, 2006, pp. 1427 and 1429) that:

> Fry analyses were used in samples with no markers.

This citation is inaccurate. The correct quotes are:

> Fry's (1979) center-to-center technique was used to calculate two-dimensional strain ellipses in rocks lacking strain markers (Albertz, 2006, p. 1427).

and

> The data in domain (2) were retrieved from Fry analysis on samples lacking strain markers (Albertz, 2006, p. 1429).

Fry's (1979) center-to-center technique is based on the principle that the distribution of particle centers (which were originally statistically nearest neighbors) in deformed rocks can be used to analyze the finite strain. Whereas most metavolcanic host rock units in the SLP contain lithic fragments that can be used as strain markers by measuring their aspect ratios on principal surfaces (Albertz, 2006, p. 1427), some of these rocks lack such strain markers. However, these rocks display porphyritic textures and I treated euhedral quartz and/or feldspar phenocrysts as particle centers in the sense of Fry (1979) to analyze the finite matrix strain.

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