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Reply to comments of Jon Olson and Richard Schultz

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Olson and Schultz (2011) have somewhat misrepresented my line of argument by stating that in Scholz (2010) I presumed that fracture toughness k_c scales as $\sqrt{L}$. What I actually did was to point out that their data (my Fig. 1) shows that $k_c \propto \sqrt{L}$ (my Fig. 2). With my eqn. (3) this requires that $d_{\max} \propto L$ which contradicts their central argument that $d_{\max} \propto \sqrt{L}$. The difference is fundamental: in the first case the crack is stable under stress boundary conditions, in the second it is not. Olson and Schultz point out that rock type, temperature etc. can result in variations in k_c. This is true but only at the level of a factor of five or so (e.g. Peck et al., 1985), not the factor of 500 seen in their data.

I then went on to point out several errors in data handling or interpretation that led to the data giving the illusion of being better fit by a square root line. The first was mixing data for single segment veins with multiple segment veins. These two are not self-similar and should not be scaled together. The second set of problems has to do with interpreting multiple segment fractures. If they are an echelon set they are likely to be the result of mixed mode III + I fracture. In that case the length of the segments are not causally related to the opening of the master crack and so the two should not be related (as is the case for the Shiprock dikes). In the more general case, segments of multiple segmented cracks have, as a result of their stress interactions, scaling of the form $d_{\max} \propto L^n$, where $n < 1$ and decreases as the degree of interaction between the cracks increases. This multiple segment scaling may approach a square root dependency, but this cannot be interpreted in terms of the propagation of an individual crack and thereby conclude that the crack is stable only under constant displacement boundary conditions, as was done by Olson (2003).

Olson and Schultz go on to argue that the evidence from other sources for $\sqrt{L}$ scaling of k_c is mixed. Labuz et al. (1987), in lab experiments on granite, found $k_c \propto \sqrt{L}$ scaling out to crack lengths of 50 mm (I could not find their statement, cited by Olson and Schultz, that k_c is constant from 80 to 160 mm). This type of scaling of k_c, termed *R*-curve behavior, is typical of low porosity rock-like materials such as polycrystalline ceramics (e.g. Evans, 1990). Olson and Schultz noted that tuff, a vesicular rock with a porosity of about 25%, exhibits a different form of scaling of k_c. The fracture process in such high porosity rock is fundamentally different from that in low porosity rock. For example, the process zone associated with faulting of high porosity rock is characterized by compaction due to porosity collapse, rather than the dilatancy from microcracking that occurs in low porosity rock such as granite. As a result, deformation bands are formed rather than faults (e.g. Fossen et al., 2007). We therefore should not consider the behavior of tuff to be particularly relevant to that of low porosity rock, which is the main interest here.

In Scholz (2010) I discussed the scaling of process zone width out to field scales. This is a critical point that was not addressed by Olson and Schultz, so I take the opportunity here to make the more complete argument. In Fig. 1 is shown a log–log plot of the process zone width W_p vs d, the aperture of joints and dikes. The estimated length of the fractures, when available, are shown in brackets. The fit to the data yields a slope of 0.9. This is an underestimate of the slope because the apertures of the two smallest data sets (A & B) are for unfilled joints and are themselves underestimates. These data are most consistent with linear scaling between W_p and d, with a typical ratio $W_p/d \approx 10$. With the assumption that the process zone is the region within which the local stress exceeds the tensile strength T of the host rock, Pollard and Segall (1987) derived the expression

$$W_p = \frac{1}{4}L[\Delta\sigma/(T - \sigma_r)]^2 \quad (1)$$

where $\Delta\sigma$ is the crack driving stress and σ_r the regional stress. Because this says that $W_p \propto L$ and the observation of Fig. 1 indicate that $W_p \propto d$, it follows that $d \propto L$. To put this in the framework of the Griffith energy balance, note that the fracture energy per unit crack

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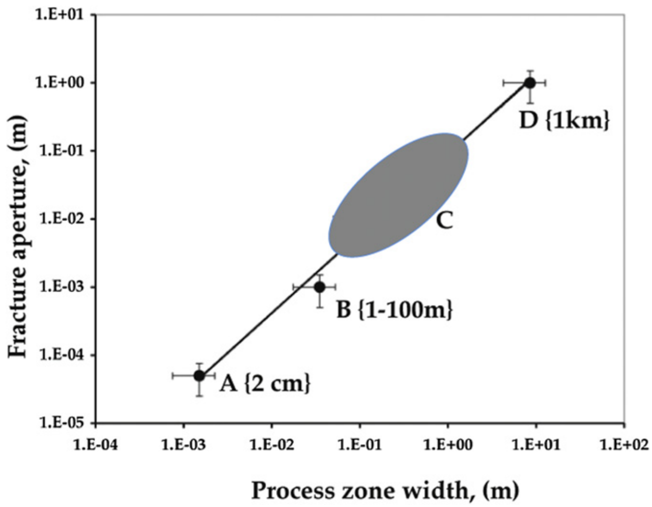

Fig. 1. A log–log plot of process zone width vs aperture for joints and dikes. Fracture length given in brackets. Data sources: (A) lab experiment on opening mode cracks in granite (Swanson, 1987), (B) joints in granite (Segall, 1984; Segall and Pollard, 1983), (C) dikes in gneisses (Engvik et al., 2005), (D) dikes in sandstone (Delaney et al., 1986).

length G_c must equal the surface energy of all the cracks in the process zone. Therefore $G_c \propto W_p$ and from eqn. (1), $G_c \propto L$. If the latter is used in the Griffith energy balance instead of Griffith's assumption that G_c is scale independent, it is found that the crack is stable under stress loading. And because $G_c \propto k_c^2$, we obtain $k_c \propto \sqrt{L}$, as before.

References

Delaney, P.T., Pollard, D.D., Ziony, J.I., McKee, E.H., 1986. Field relations between dikes and joints – Emplacement processes and paleostress analysis. J. Geophys. Res.-Solid Earth Planets 91 (B5), 4920–4938.
Engvik, A.K., Bertram, A., Kalthoff, J.F., Stockhert, B., Austrheim, H., Elvevold, S., 2005. Magma-driven hydraulic fracturing and infiltration of fluids into the damaged host rock, an example from Dronning Maud Land, Antarctica. J. Struct. Geol. 27 (5), 839–854.
Evans, A.G., 1990. Perspective on the development of high-toughness ceramics. J. Am. Ceramics Soc. 73, 187–206.
Fossen, H., Schultz, R.A., Shipton, Z.K., Mair, K., 2007. Deformation bands in sandstone: a review. J. Struct. Geol. 164, 755–769.
Labuz, J.F., Shah, S.P., Dowding, C.H., 1987. The fracture process zone in granite – Evidence and effect. Int. J. Rock Mech. Mining Sci. Geomech. Abs. 24 (4), 235–246.
Olson, J.E., 2003. Sublinear scaling of fracture aperture versus length: an exception or the rule? J. Geophys. Res.-Solid Earth 108 (B9).
Olson, J.E., Schultz, R.A., 2011. Comments on "A note on the scaling relations for opening mode fractures in rock" by C.H. Scholz. J. Struct. Geol.
Peck, L., Barton, C.C., Gordon, R.B., 1985. Microstructure and the resistance of rock to tensile fracture. J. Geophys. Res.-Solid Earth Planets 90 (NB13), 1533–1546.
Pollard, D.D., Segall, P., 1987. Theoretical displacements and stresses near fractures in rock: with applications to faults, joints, dikes, and solution surfaces. In: Atkinson, B.K. (Ed.), Fracture Mechanics of Rock. Academic Press, London, pp. 277–348.
Scholz, C.H., 2010. A note on the scaling relations for opening mode fractures in rock. J. Struct. Geol. 32, 1485–1487.
Segall, P., 1984. Formation and growth of extensional fracture sets. Geol. Soc. Am. Bull. 95 (4), 454–462.
Segall, P., Pollard, D., 1983. Joint formation in granitic rock of the Sierra Nevada. Geol. Soc. Am. Bull. 94, 563–575.
Swanson, P.L., 1987. Tensile fracture resistance mechanisms in brittle polycrystals: an ultrasonic and microscopic investigation. J. Geophys. Res. 92, 8015–8036.