2. Frameworks of deformation, scaling relations and dependence on parameter influence

The imbalance between a) tectonic forces that are characterized by convergence and coupling between upper and lower plate or the coupling between the lithosphere and the underlying mantle, and b) buoyancy forces that are caused by the lateral and vertical density contrasts within the lithosphere, is often held responsible for orogen dynamics (e.g., McKenzie, 1969; Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Richardson et al., 1979; Dewey, 1980; Ranalli, 1987), more references in Sengör, 1990). The resulting force first drives the activation of faults in the upper brittle crust that deforms according to Byerlee's law (Byerlee, 1978); the remaining force is then released as deformation of the ductile lower lithosphere, depending on rheology (and therefore temperature) and strain rate (e.g., Ranalli, 1987; Thatcher, 1995; Royden, 1996; Zoback and Townsend, 2001).

These features are subject to the three common deformation frameworks (summarized by Ben-Zion and Sammis, 2003), which differ according to the mode of strain accumulation and their dominant structures. The first is the "continuum-Euclidean" framework, in which the effect of strain softening decreases the yield strength of the upper brittle crust and strain localizes at fault planes that have simple "Euclidean", i.e., planar geometries. These faults are evenly distributed in an elastic continuum accumulating different amounts of strain (Thatcher, 1995; Behn et al., 2002; Ben-Zion and Sammis, 2003). The continuum has been approximated by the simplified "thin viscous sheet model" (England and McKenzie, 1982), assuming the lithosphere to be viscous, rheologically homogeneous and isostatically compensated (Thatcher, 1995; Townsend and Sonder, 2001).

The second deformation framework is the "block model", describing strain accumulation at the margins of rigid blocks that do not deform internally (e.g., Jackson, 2002). Strain accumulation is diffuse and occurs in broad shear zones (King et al., 1994), where faults are numerous, but small. This is further reinforced by the effect of strain hardening, which is characteristic for the block model.

In the third model, strain accumulates along complex structures that follow fractal patterns and/or power law relationships (e.g., Turcotte, 1992; Bak, 1996). The processes of strain softening and strain hardening balance each other so that the system is in a "critical state". Deformation structures are scale-invariant and self-similar (e.g., Turcotte, 1992; Bak, 1996), so that the structures show neither spatial nor temporal patterns (be it regional fault networks, single faults or any other structures within fault zones) (e.g., Ben-Zion and Sammis, 2003).

One good example for a power law relation is the Gutenberg-Richter relation, in which the frequency of seismic events correlates with their magnitude (Bak and Chang, 1989; Kagan and Jackson, 1991; Bak, 1996). Similarly, the particle size in shear zones correlates with the duration of fragmentation processes (Billi and Storti, 2003), whereas the displacement of faults in analogue models correlates with their size, and the size of faults with slip frequency (Bellahsen et al., 2003). Fractal patterns exist for particle size in fault cataclasites (Billi and Storti, 2003), length of fault segments (Okubo and Aki, 1987; Marrett and Allmendinger, 1994), and size of crustal blocks of the block model (Gallagher, 1981; Nur et al., 1989).

Ben-Zion and Sammis (2003) argue that fractal patterns and power law relations are not exclusively indicative of complex structures, as they can also be found when structures generally have Euclidean geometries. In a similar fashion, Turcotte and Glasscoe (2004) point out that deformation in the crust is not linearly viscous according to the "viscous sheet model" (England and McKenzie, 1982), but according to a power law relation which correlates stress and strain rate, and that faults of various sizes equally accumulate strain. However, they concede that these arguments do not preclude the "continuum model".

Zoback et al. (2002) give another example in which various models occur in close spatial proximity, namely along the North American plate margin (fore-arc and adjacent foreland basin). High heat flow is responsible for low viscosities in the lower crust supporting ductile deformation. When heat flow is high, only small stresses are necessary to cause brittle deformation at the same time, which is consistent with the continuum model. In contrast, the heat flow in the adjacent foreland basin is very low, causing the crust to behave as a rigid block, not deforming internally.

This is characteristic for the block model.

Given these examples, it is likely that models co-exist both in space and time, and that characteristic structures overprint each other. Ben-Zion and Sammis (2003) argue that strain initially accumulates in diffuse shear zones during a period of strain hardening or along complex structures. Subsequently, characteristic structures are dominantly strain softening and strain localizes along structures. Complex structures that have played an important role in the initial stage become more and more planar with increased slip.

Changes from initially diffuse strain distribution in broad shear zones to strain localization along planar faults have been demonstrated in analogue models with granular media like sand (Adam et al., 2004). Simulations by Tchalenko (1970) show that deformation initially accumulates along Riedel shears, then along P- and Y-shears, and eventually forms narrow shear zones.

To date, the spatial distribution of characteristic structures has been better examined on various scales (from millimeter scale to fault networks up to plate margins) than the temporal pattern. This is due to the fact that different stages, during which characteristic deformation structures are formed, cannot be easily differentiated in the field, as structures overprint each other. Therefore, we lack the details of the deformational patterns over time, e.g., regarding the number of deformation stages and their duration, which again depends on the spatial scale of the observed structures. In contrast, sedimentary patterns and stratigraphic sequences have been studied with much higher temporal resolution in the outcrop and in seismic sections.

As far as the temporal aspect of strain accumulation is concerned, previous studies mainly focused on the comparison of deformation rates on the geological long-term scale of several million years and current GPS rates. In this regard, Leffler et al. (1997) and Liu et al. (2000) have shown that, e.g., for the Central Andes, shortening rates of both scales are different. Klosko et al. (2002) note that such a comparison is not straightforward, because GPS rates also include the elastic component. The works of Friedrich et al. (2003, 2004) in the Basin and Range province, USA, additionally cover time scales over four orders of magnitude in between and can therefore quantify the duration of deformation activity from sudden seismic events over kiloyears, to millions of years. Their results also show that displacement rates are different for each of the studied scales. This suggests the existence of characteristic deformation patterns in time. Yet, the characteristics of such patterns remain to be identified over the scales.

Further complications result from the diverse impact of parameters that are either 1) intrinsic such as rheological properties with lateral and vertical mechanical anisotropies (e.g., Thatcher, 1995; Townsend and Sonder, 2001; Jackson, 2002; Pysklywec et al., 2002; Klepeis et al., 2004); buoyancy effects caused by density and viscosity contrasts (Townsend and Sonder, 2001; Jackson, 2002); coupling effects between upper and lower crust or between lithosphere and mantle (Vanderhaeghe and Teyssier, 2001; Klepeis et al., 2004); and thermal effects (e.g., a. heat flow, Zoback et al., 2002; b. delamination processes, Corti et al., 2003; Babeyko et al., 2004a; c. magmatism, Corti et al., 2003; Klepeis et al., 2004; Trumbull et al., 2006), or 2) influenced by external factors on the deformation system including convergence rate, subduction angle, degree of coupling between the upper and the lower plates, effect of an indenter during collision (Pysklywec et al., 2002), and climatic effects (Schlunegger and Willett, 1999; Zeitler et al., 2001; Hoth et al., 2004).

Each of these parameters acts on a particular scale or range of scales in time as well as in space. However, their impact is likely to extend to other scales as well. Additionally, parameters are coupled variously to each other, changing their influence on a system and thus leading to an increased complexity of the relation between causes and effects, both in time and space.