Int J Earth Sciences (Geol Rundsch) (2001) 90 : 149–156
DOI 10.1007/s005310000161

ORIGINAL PAPER

M. R. Handy · J. Braun · M. Brown · N. Kukowski
M. S. Paterson · S. M. Schmid · B. Stöckhert
K. Stüwe · A. B. Thompson · E. Wosnitza

Rheology and geodynamic modelling: the next step forward

Received: 5 August 2000 / Accepted: 9 October 2000 / Published online: 9 March 2001

Abstract The application of continuum mechanics and microstructural analysis to geological studies over the past 30 years has spurred earth scientists to reassess fundamental tectonic processes such as subduction, collision and rifting in terms of dynamics. Armed with new analytical methods, geologists have returned to the field to look at rock structures with more mechanistic eyes. The advent of sophisticated computers, programs, and laboratory deformation equipment has facilitated the simulation of geodynamic processes that range in scale from the grain to the lithosphere. The result has been specialization, with the concomitant opening of communication gaps between geodynamicists, field geologists and rock mechanicists. Partly, these gaps reflect differences of perception and approach. In order to bridge these gaps, a workshop was organized after the DRM conference to debate how field and laboratory studies of deformed rocks can improve our understanding of lithospheric rheology, and in turn, how this understanding can be used to refine dynamic models of orogenesis. The workshop hosted participants with backgrounds in structural geology, experimental rock mechanics, metamorphic petrology and both numerical and analogue modelling. This paper summarizes the main controversies and conclusions reached during the workshop. For the sake of brevity, referencing in this summary is restricted to literature referred to during the oral presentations and to comments made by speakers themselves (names italicized).

Keywords Subduction · Collision · Rifting · Lithosphere · Rheology · Orogenesis

M. R. Handy (✉)
Geowissenschaften, Freie Universität Berlin, 12249 Berlin, Germany
E-mail: mhandy@zedat.fu-berlin.de
Phone: +49-30-83870311
Fax: +49-30-7759078

J. Braun · M. S. Paterson
Research School of Earth Sciences, Australian National University, Canberra ACT 0200 Australia

M. Brown
Department of Geology, University of Maryland, Baltimore, 20742-4211 Maryland, USA

N. Kukowski
GeoforschungsZentrum, Telegrafenberg, 14473 Potsdam, Germany

S. M. Schmid
Geologisches-Paläontologisches Institut, Universität Basel, 4056 Basel, Switzerland

B. Stöckhert
Institut für Geologie, Ruhr-Universität Bochum, 44801 Bochum, Germany

K. Stüwe
Institut für Geologie und Paläontologie, University of Graz, 8010 Graz, Austria

A. B. Thompson
Department of Erdwissenschaften, ETH-Zürich, 8092 Zürich, Switzerland

E. Wosnitza
Geologisches Institut, University of Freiburg, 79104 Freiburg, Germany

Problems of specialization and professional bias

Communication gaps between field geologists, geodynamicists and rock mechanicists arise not only from the specialized knowledge which has become necessary for new breakthroughs in their respective fields, but also from differences of scientific experience and approach: most field geologists have a detailed, kinematic view of an orogen, grappling as they must with the complexities of outcrop-scale structures and the regional geology (Fig. 1a). In reconstructing the three-dimensional structural evolution of an orogen, they are confronted with the problem of distinguishing dif-

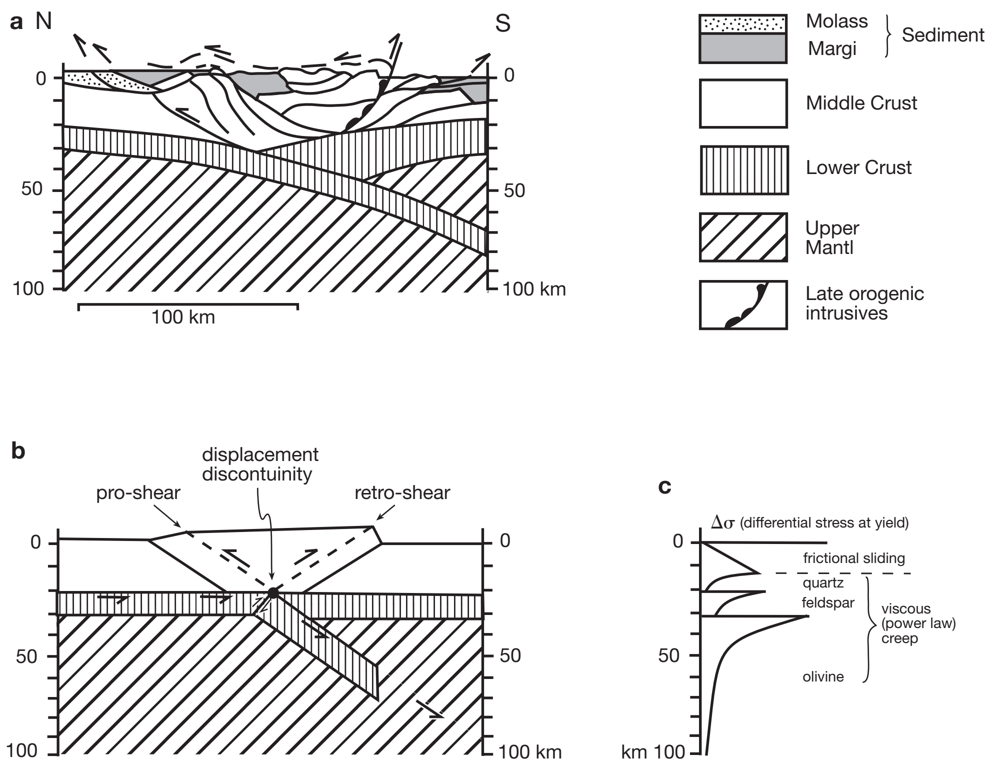

Fig. 1a–c Current abstractions of a single orogenic reality. **a** An orogen as seen by the geologist/geophysicist (inspired by the Swiss NFP-20 eastern transect across the Alps). **b** An orogen as conceived by a geodynamicist/modeller (inspired by models of C. Beaumont and co-workers). **c** The strength of the lithosphere ($\Delta\sigma$) as a function of depth (z), as envisioned by rock mechanicists (Brace and Kohlstedt 1980; Ranalli and Murphy 1987). The shape of the curves leads some to refer to the diagram as a Christmas tree

ferently aged structures, strata and metamorphic mineral assemblages, some of which formed during unrelated tectonothermal events. Dating structures is a problem, especially if reliable markers are rare or missing. From the geometry of structures, geologists can infer the relative strengths of rocks and minerals. Under ideal circumstances, they can even determine the local direction and magnitude of finite strain axes. Unfortunately, even good structural information in nature reveals very little about the stress history of the orogenic lithosphere. In the absence of a stress-strain curve, field geologists can only infer the rheology of rocks by comparing natural structures and microstructures with those of tiny rock specimens deformed in the laboratory at much higher temperatures and strain rates. The geodynamicist's view of the lithosphere is perforce a large-scale geometrical and dynamic abstraction. Most dynamic orogenic models are two-dimensional and make plane-strain or thin-sheet approximations. At best, their geometry approximates that of lithosphere-scale structures obtained from geophysical images of mountain belts (Fig. 1b). Current numerical models of orogenesis are based on continuum mechanical principles and entail the assumption of an initially rheologically stratified lithosphere that undergoes homogeneous, steady-state deformation. Abstractions like this are certainly necessary if a model is to deliver results. Yet such abstractions evoke scepticism among some geologists because they are difficult to reconcile with the fact that deformational and metamorphic features in real rocks are often discontinuous in both time and space, at least on some scales. Interestingly, even a simple, two-dimensional model orogen requires that the deformation of the lithosphere be transient and heterogeneous at a displacement discontinuity or singularity defining subduction (Fig. 1b), especially if this discontinuity migrates in space and time during orogenesis.

The rock mechanicists' view of Earth's lithosphere is primarily one-dimensional and is perhaps best exemplified with the simple strength vs depth diagram (Fig. 1c). Rock strength is inferred to increase with depth where deformation is by pressure-dependent fracture and frictional sliding, and, because of the positive dT/dz of geotherms, decreases with depth where deformation occurs by temperature- and strain-rate-dependent creep. Alternatively, mechanicists think in multivariable space, constructing deformation regime maps for competing grain-scale processes. Models like this are empirically based and derive from the extrapolation of steady-state, laboratory flow laws to natural temperatures and strain rates. Yet, the only real justification for extrapolating steady-state flow laws to natural conditions is the similarity of microstructures in experiments with microstructures formed under dynamic conditions in naturally deformed rocks. Extrapolating experimental flow laws involves making numerous questionable assumptions, e.g. that real rocks deform at thermal, mechanical and structural steady state on the time scale of orogenesis, or that deformation occurs at conditions of uniform strain rate or uniform stress. The latter conditions only pertain if mechanical anisotropies in rocks are perfectly planar and remain, respectively, perpendicular or parallel to the shearing plane during deformation. In addition, experimental flow laws are often derived for very fine-grained aggregates, and extrapolating these flow laws entails the tacit assumption that grain-boundary effects are the same in nature as in the experiment, despite the fact that natural grain sizes are significantly larger than experimental ones.

Both the geodynamic and rock mechanical views are simple, based as they are on many assumptions that are untested, and a few that are false. Yet to an extent, the models they engender are realistic and have become paradigms.

So why improve such models? There are two main reasons: Firstly, we will understand the dynamics of lithospheric processes, such as orogenesis, better if we conduct quantitative forward modelling to test how varying rheological parameters affect the geometry and evolution of mountain belts and rifts. Rheology is only one parameter but an influential one that is poorly constrained at present. Modelling offers us a possibility to place additional constraints on the rheology of rocks. This approach complements laboratory experiments and field work, which face limitations of their own related to problems of time, scale, and drilling depths (to name only few). Secondly, there is something disturbing about models that are simple but nevertheless predict natural geometries. Perhaps the models are only apparently successful, i.e. they may be telling us right things based on some wrong assumptions. New perspectives open up when fundamental assumptions (e.g. steady state) are questioned and discrepancies between nature and experiment are noted.

Current approaches, results, questions and debates

To focus debate, four questions were posed at the beginning of the workshop:

1. What is the effect of rheology on numerically modelled orogenesis and rifting?
2. What does nature tell us about rock rheology?
3. What are the problems in extrapolating experimental flow laws to natural strain rates and temperatures?
4. What can we learn from analogue models?

These questions were addressed by invited speakers (*Chris Beaumont, Mervyn Paterson, Stefan Schmid*) and then opened to discussion between shorter oral presentations from other workshop participants. Herein we summarize some of the results and debates centred on these questions as a prelude to a venue for future work.

Modelling studies show that although rheology certainly affects the style and evolution of lithospheric deformation during both orogenesis and rifting, it is only one parameter among other important factors (e.g. imposed tectonic forces, force of gravity, denudation rate). Several presentations demonstrated how velocity boundary conditions can be varied in numerical and physical models to investigate the deformational styles of crustal layers with various mechanical and thermal properties. As mentioned in the previous section, rheology is simulated by extrapolating experimentally derived constitutive equations of flow to natural temperatures and strain rates. In finite-element models of lithospheric subduction beneath an orogen (Beaumont et al. 1994), altering the rheological parameters in the constitutive equations affects orogenic style, but these changes in style reflect the relative magnitude of crustal strength, the driving forces and the force of gravity rather than the absolute strength or material properties of the layers (*Beaumont*). Modelling strain localization and strain-dependent changes in rheology is in its infancy, partly because the size of finite-element grids in such models places a lower limit on the scale of strain heterogeneities that can be modelled, and partly because we do not know yet over what scales the rheological parameters are invariant. Therefore, the finite-element approach may not be suitable for modelling strain localization if localization involves the propagation of instabilities nucleated on very small scales.

One approach to simulating strain-dependent changes in rheology entails parametrizing the transition between two deformation mechanisms (*Braun*). Numerical analysis of a rheological system in the vicinity of a deformation mechanism boundary reveals that such systems behave in a very complex manner which is not observed on either side of this boundary. In Braun's model of the switch from grain-size insensitive (dislocation) creep to grain-size sensitive (diffusional) creep, for example, even minor perturbations in the initial grain-size distribution within the mantle can induce runaway strain localization along fine-grained zones. Localization patterns during rifting propagate to progressively larger scales and eventually affect the entire lithosphere. Thus, current numerical methods are capable of reproducing strain localization on the same range of scales as those observed in nature. This raises the following question: What scale(s) of inherent heterogeneities engender and/or inhibit the growth of mechanical instabilities which lead to strain localization on the lithospheric scale?

Field studies of natural structures have confirmed some, but not all, of the detachment horizons predicted in dynamic models (*Schmid*). Detailed structural studies of both orogenic and rifted continental crust indicate that weak zones of detachment develop with progressive strain at the following locations (Fig. 2): (a) at or near the brittle-to-ductile transition; (b) at the base of hydrous, quartz- and melt-rich layers in the intermediate and lower crust; and (c) possibly also at the top of the mantle lithosphere. These weak zones and detachments resulted from various mechanism transitions which are either strain-dependent at

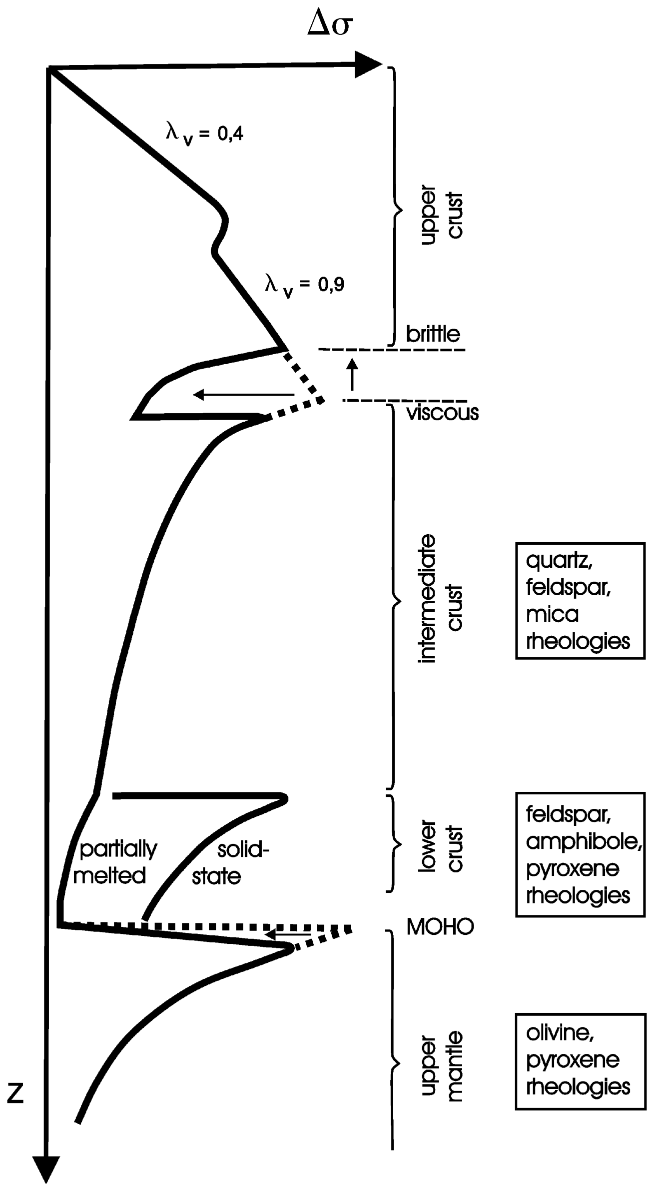

Fig. 2 Generic strength vs depth diagram for orogenic lithosphere showing location of strength maxima and minima inferred from natural and experimental studies of deformed rocks (compare with Fig. 1c; modified from Handy 1989; Handy et al., 2001). *Dotted line* and *solid line* show low-strain and high-strain parts of strength-depth curves, respectively. *Small arrows* indicate movement of curves during progressive deformation. λ_v is the pore-fluid factor in the upper crust, defined as the ratio of the pore-fluid pressure to the lithostatic pressure (P_f/P_l)

initially high differential stresses (e.g. transition from cataclasis or dislocation creep to diffusion creep in (a) and (c)) or enhanced by syntectonic metamorphic reactions (Handy 1989).

Experimentally derived constitutive equations are relatively well constrained for power-law (dislocation) creep of polycrystalline aggregates of quartz, calcite, halite and olivine (*Paterson* (1987); Carter and Tsenn 1987; Kohlstedt et al. 1995). This does not mean, however, that their extrapolation to natural conditions necessarily yields realistic lithospheric strengths (Paterson 1987). Extrapolated mechanical data from the laboratory are only geologically relevant if the deformation processes in the Earth are really reproduced on the laboratory scale. Even then, there is the problem that the laboratory data pertain to a small sample of rock. This sample does not necessarily represent fully the geological rock mass because of larger-scale heterogeneities (e.g. joints, shear zones) in the rock. Furthermore, the regional-scale geological body may contain different rock types. The rheological properties that are appropriate for geodynamical modelling are those of a "representative elementary volume" (REV) of the rock mass or regional body which averages over the heterogeneities and rock types; therefore, it is in general also necessary to undertake a modelling to obtain the rheological parameters appropriate to the REV from the parameters determined in the laboratory on individual rock types. This rheological modelling will be analogous to that involved in modelling the properties of a polymineralic rock in terms of the properties of the individual phases. Another problem is whether steady-state flow laws are appropriate for simulating the rheology of transient, high-strain deformation in the lithosphere. The assumption of mechanical steady state may only be applicable to nature if the rate of microstructural equilibration during changing environmental conditions is fast with respect to the duration of deformation.

Despite problems of equilibration rate, detailed microstructural and palaeopiezometric studies in field areas with a well-constrained thermal and kinematic history (e.g. the sheared contact aureole of syn-tectonic granitoids) indicate that, at least in the case of quartz, the differential stresses obtained from grain-size palaeopiezometers (Christie and Ord 1980; Etheridge and Wilkie 1981) are consistent with the stresses independently predicted by the experimental flow laws for dislocation creep at the temperatures and strain rates of deformation (*Schmid*). This lends semi-quantitative credence to the extrapolatability of steady-state flow laws for power-law creep of quartzite. In fact, the relative strengths of several silicate minerals inferred from microstructures in naturally deformed rocks (Handy and Zingg 1991) is in surprisingly good agreement with relative strengths of these mineral aggregates obtained by extrapolating laboratory flow laws to a range of natural strain rates and temperatures (*Schmid*). Moreover, the minimum temperature predicted for the onset of intracrystalline plasticity in major rock-forming minerals is similar to that estimated by extrapolating laboratory flow laws for power-law creep and frictional sliding to natural conditions (e.g. Brace and Kohlstedt 1980). For example, van Daalen et al. (1999) estimated a minimum temperature of 270 °C for the onset of intracrystalline plasticity in natural quartzite layers.

In contrast to quartzite, there are still no adequate flow laws for feldspar aggregates that can be used to estimate the rheology of the lower crust. Numerical models of orogenesis which utilize the few available feldspar flow laws yield anomalous orogenic styles which are diagnostic of an unrealistically weak lower crust (*Ellis*). This is clearly inconsistent with field evidence that the lower crust is relatively flow resistant (*Schmid*). Large segments of the lower crust are inferred to be stronger than both the intermediate crust and the upper mantle, based on the geometry of lower crustal reflectors in seismic reflection profiles across the Central Alps (NFP20-East profile; Schmid et al. 1996). The opposite conclusion is reached by Reston (1990) based on the geometry of seismic reflectors in the northeastern Atlantic passive continental margin, where a relatively weak lower crust is believed to have decoupled the upper crust from the upper (lithospheric) mantle. However, *Schmid* pointed out that the same reflection seismic data can be used to infer detachment of the entire crust from the upper mantle along the MOHO and/or within the base of the lower crust; therefore, even in continental margins undergoing extension, the lower crust may be relatively strong compared with the upper crust and upper mantle.

Orogens with well-constrained structures and histories are natural laboratories on the largest scale and can be used to infer stresses in the lithosphere (e.g. Fleitout and Froidevaux 1983; Molnar and England 1990; England and Molnar 1991; Stüwe 1998). These and many other studies show that the average shear stress acting on the base of the continental lithosphere is of the order of 100 MPa. Due to thermal and lithological heterogeneities, however, it is likely that some parts of the lithosphere are much stronger than this value (several hundred MPa), whereas other parts are much weaker (tens of MPa). These values fall within the range of differential stresses obtained from laboratory-calibrated palaeopiezometers applied to dynamically recrystallized aggregates in natural mylonites (Etheridge and Wilkie 1981). They are also consistent with flow stresses obtained from flow laws for solid-state creep of two-phase aggregates (Handy et al. 1999).

New flow laws for high-pressure and ultrahigh-pressure minerals in crustal rocks (e.g. coesite; Stöckhert and Renner 1998) provide upper bounds on the shear stresses in subduction and collision zones (*Stöckhert*). The preservation of undeformed to weakly deformed high- and ultrahigh-pressure metamorphic rocks in mountain belts and the notable absence of evidence therein for dislocation creep at ultrahigh-pressure conditions indicate that deformation during subduction is localized within a low-viscosity channel at the top of the downgoing slab (Stöckhert et al. 1997, 1999; Stöckhert and Renner 1998). This suggests effective mechanical decoupling at the plate interface. It further indicates that shear heating does not contribute significantly to the heat budget of subduction zones (Stöckhert and Renner 1998). Certainly, differential stresses in low-viscosity channels are lower than those required in computer simulations (Peacock 1996) to raise the local subduction geotherm to levels necessary for the formation of commonly observed (ultra)high-pressure mineral assemblages at these high pressures. Dissolution–precipitation creep is suggested to be the dominant deformation mechanism in subduction shear zones, accounting for the low shear stresses and high strain rates characteristic of most subduction zones. Conventional-strength profiles (so-called Christmas trees; Ranalli and Murphy 1987; Fig. 1c) based on extrapolated flow laws for grain-size insensitive, power-law creep are therefore inappropriate to simulate the rheology of subduction zones and some collisional zones.

In other tectonic settings, large-scale heat sources affect crustal rheology. For example, advected heat associated with syn-tectonic granitic intrusions in the intermediate crust is expected to have a profound softening effect on the rheology of continental transpressive zones (Brown and Solar 1999). This effect generates feedback relations between deformation and melt movement (Brown and Solar 1998) and limits the contribution of deformation to the heat budget. Similarly, the duration of deformation with respect to the characteristic thermal diffusion time in deforming rock determines the evolution of the relative viscous strength of different layers in the lithosphere during orogenesis (*Schulmann, Thompson*, Thompson et al. 1997, 2001).

In affecting the local heat budget, deformation perturbs chemical equilibrium on the grain scale (Stünitz 1998). This in turn may affect the interpretation of pressure–temperature paths in metamorphic rocks which are used to constrain geodynamic models (Stüwe and Sandiford 1994). Barr and Houseman (1996) studied the dynamics of brittle fracture development in rocks and concluded that because of a singularity condition near tips of faults, mineral parageneses in fractured metamorphic rocks may record pressures exceeding the actual lithostatic pressure. Provided that the rate of strain energy dissipatation is sufficiently high, the temperature-sensitive mineral parageneses may reflect the level of differential stress level (e.g. England and Molnar 1991; Stüwe 1998); however, few studies thus far have provided unambiguous evidence that metamorphic minerals record the differential stress during deformation.

Scaled analogue models have been used to simulate a wide variety of tectonic processes and settings, e.g. deformation in fold-and-thrust belts (e.g. Malavieille 1984), mass transfer and deformation at convergent margins (e.g. Dahlen 1990; Gutscher et al. 1998), extensional fault geometry in core complexes (e.g. Brun et al. 1994), subduction and exhumation of high-pressure rocks (Chemenda et al. 1995), pluton emplacement mechanisms (Soula 1982; Román-Berdiel et al. 1995, 1997; Benn et al. 1998), block rotation within strike-slip and oblique-convergent fault zones (Schreurs 1994). A wide range of analogue materials with both cohesive frictional (Navier-Coulomb: e.g. sand) and linear viscous (Newtonian: e.g. silicon putty) rheologies have been used to reproduce the rheology of geological materials (review by Koyi 1997). The authenticity of such models depends strongly on their "scaleability" for kinematic, dynamic and rheological similarity with nature. Sandbox models of accretionary wedges that employ dry quartz sand simulate dilatant behaviour and have coefficients of cohesion and internal friction that are similar to those of crustal rock undergoing brittle deformation. Because the scaling factors for cohesion and geometry are the same between nature and experiment, sandbox simulations of mass-transfer modes, deformation patterns and force balancing that employ dry quartz sand or other granular materials, such as micro-glass beads or mortar, yield predictable and quantitative results. The applicability of such scaled models to natural systems is limited because the models are dry (i.e., there is no pore-fluid pressure) and because the scaling factors for the elastic parameters between sand and natural rocks differ from the scaling factors for cohesion and geometry by at least two orders of magnitude. For this reason, numerical models are an important supplement to analogue models (*Kukowski*). Initial results of thermo-mechanical modelling demonstrate the possibility of maintaining and measuring temperature gradients during deformation on a scale that allows one to simulate large-scale tectonic processes (*Wosnitza* et al., 2001). The advantage of analogue models is that they allow direct observation of the evolution of large-scale tectonic processes. In addition, the controlled kinematic, dynamic and rheological boundary conditions imposed in analogue models render them ideal for testing and calibrating numerical dynamic models.

Venue for future work

The workshop concluded with recommendations on how future research could usefully be directed in order to resolve the questions raised herein. Recommendations were made for field studies, rock deformation experiments and geodynamic modelling, although it was agreed that future projects will benefit from a combination of these approaches.

Field studies

Starting with large-scale tectonics, more integrated geological–geophysical studies of different types of mountain belts are needed to establish the relationship between orogenic architecture, kinematic framework and the pre-existing (i.e. pre-orogenic) lithospheric configuration.

A key to understanding crustal dynamics certainly lies in understanding mantle dynamics, particularly the role of the asthenosphere during orogenesis and rifting. Field studies on exhumed ultramafic rocks, as for example in Oman or in the circum-Mediterranean mountain belts, may yield insight into the mechanisms of deformation and localization in the mantle lithosphere (e.g. Vissers et al. 1995).

Detailed structural studies are required to determine the length scales of deformational heterogeneity, e.g. the spacing and/or length of shear zones on different scales of observation. Are these length scales dependent on the conditions and/or mechanisms of deformation? Such studies would help define the "representative elementary volume" of rock mentioned previously as a possible requisite for scaling laboratory rheologies to natural rheologies.

Additional field studies in well-known geological settings (i.e. where differential stress, temperature and/or strain rate can be reasonably well constrained) are needed to test laboratory flow laws for various creep mechanisms. Field studies of natural tectonites can also be used to derive constitutive laws for grain growth under both static and dynamic conditions, and to establish melt distribution in both annealed and dynamically recrystallized, anatectic rocks. On an even smaller scale, element distribution patterns in porphyroclasts may yield a new way of determining the magnitudes of shear stresses (*Stüwe*) provided that the kinematic and thermal history of the samples is simple and well established. In general, nature can be used as a laboratory if the kinematic and metamorphic histories are sufficiently well known to provide tight constraints on the physical conditions of deformation (temperature, effective pressure, strain rate, strain, differential stress).

A search for and study of transient phenomena in naturally deformed rocks (e.g. at the brittle–ductile transition) may help to characterize some of the underlying causes of episodicity and cyclicity in orogenic processes. Most, if not all, geological processes are temporally discontinuous and are often perceived as "events" in the rock record (e.g. earthquakes, melt segregation, convection, erosion). A given process appears to occur only if a threshhold value of a parameter(s) related to that process is exceeded. Episodicity in orogenesis may also be related to complex, non-linear feedback mechanisms, as for example in the relationship between the change in strain rate along large shear zones as a function of viscous shear stress (by a power-law relation), temperature (by an exponential function), syn-tectonic grain size (by a power-law relation), erosion rate and crustal thickness (by a quadratic function). Constraining threshhold values and non-linear feedback mechanisms in naturally and experimentally deformed rocks may help to simulate the role of transient rheologies in orogenesis.

Rock deformation studies

New experiments are needed to investigate the rheology of polycrystalline mineral aggregates during strain localization, syntectonic reaction and syntectonic partial melting. In particular, high-strain (torsional) experiments designed to allow the extraction of time- and strain-dependent flow parameters may be used to investigate strain localization. Although such experiments are less direct in providing rheological data because of the inhomogeneous deformation in a torsion specimen, with appropriate analysis they may provide modellers with flow laws that are more applicable to the deformational processes during subduction, orogenesis and rifting than current flow laws derived from relatively low strain, homogeneous deformation. Feldspar and amphibole are abundant constituents of the lower continental crust; thus, better flow laws for dislocation and diffusion creep of these minerals are obviously desirable for extrapolation to natural conditions in dynamic models.

There is a great need for experimental flow laws for very fine-grained rocks, especially those rocks which comprise two or more dispersed minerals. These should be tested against a new generation of theoretical flow laws that describe the rheology of aggregates undergoing deformation by more than one mechanism (e.g. dislocation and diffusion creep) operating in series and/or in parallel.

Dynamic modelling

In the absence of reliable flow laws for some important minerals (e.g. feldspar) and deformation mechanisms (e.g. grain-size-sensitive creep of mono- and polymineralic aggregates), dynamic modellers can conduct parameter studies to test the robustness of their models to changes in rheology with respect to other factors. Strain localization over geologically significant time periods might be simulated by algorithms which emulate coupled processes (e.g. grain-size reduction and grain growth, localization and melt segregation).

On a different scale, dynamic models should be tested against well-studied type areas for subduction, orogenesis (e.g. the Central and Western Alps) and rifting (e.g. Galicia passive margin). Finally, the code used in the dynamic models can be improved by comparing experimental numerical models against analogue model experiments.

Even with a plethora of information, we can only shed more light on geodynamic processes if we realize that different approaches are not mutually exclusive, no matter how different the results of modelling, laboratory and field work appear to be in some instances. It was in this spirit of reaching conciliation through discrepancy that we sought to make progress in this workshop.

References

Barr TD, Houseman G (1996) Deformation fields around a fault embedded in a non-linear ductile medium. Geophys J Int 125:473–490

Beaumont C, Fullsack P, Hamilton J (1994) Styles of crustal deformation in compressional orogens caused by subduction of the underlying lithosphere. Tectonophysics 232:119–132

Benn K, Odonne F, Saint Blanquat M (1998) Pluton emplacement during transpression in brittle crust: new views from analogue experiments. Geology 26:1079–1082

Brace WF, Kohlstedt DL (1980) Limits on lithospheric stress imposed by laboratory experiments. J Geophys Res 85:6248–6252

Brown M, Solar GS (1998) Shear zone systems and melts: feedback relations and self-organization in orogenic belts. J Struct Geol 20:211–227

Brown M, Solar GS (1999) The mechanism of ascent and emplacement of granite magma during transpression: a syntectonic granite paradigm. Tectonophysics 312:1–33

Brun JP, Sokoutis D, Driessche J (1994) Analogue modeling of detachment fault systems and core complexes. Geology 22:319–322

Carter NL, Tsenn MC (1987) Flow properties of continental lithosphere. Tectonophysics 136:27–63

Chemenda AI, Mattauer M, Malavieille J, Bokun AN (1995) A mechanism for syn-collisional rock exhumation and associated normal faulting: results from physical modelling. Earth Planet Sci Lett 132:225–232

Christie JM, Ord A (1980) Flow stress from microstructures of mylonites: example and current assessment. J Geophys Res 85:6253–6262

Dahlen FA (1990) Critical taper model for fold-and-thrust belts and accretionary wedges. Annu Rev Earth Planet Sci 18:55–99

England P, Molnar P (1991) Inferences of deviatoric stress in actively deforming belts from simple physical models. Phil Trans R Soc Lond 337:151–164

Etheridge MA, Wilkie JC (1981) An assessment of dynamically recrystallised grain size as a paleopiezometer in quartz-bearing mylonite zones. Tectonophysics 78:475–508

Fleitout L, Froidevaux C (1983) Tectonic stress in the lithosphere. Tectonics 2:315–324

Gutscher M-A, Kukowski N, Lallemand SE, Malavieille J (1998) Material transfer in accretionary wedges from analysis of a systematic series of analogue experiments. J Struct Geol 20:407–416

Handy MR (1989) Deformation regimes and the rheological evolution of fault zones in the lithosphere: the effects of pressure, temperature, grain size, and time. Tectonophysics 163:119–152

Handy MR, Zingg A (1991) The tectonic and rheologic evolution of an attenuated cross section of the continental crust: Ivrea crustal section, southern Alps, northwestern Italy and southern Switzerland. Geol Soc Am Bull 103:236–253

Handy MR, Wissing S, Streit JE (1999) Strength and structure of mylonite with combined frictional-viscous rheology and varied bimineralic composition. Tectonophysics 303:175–192

Handy MR, Mulch A, Rosenau M, Rosenberg CL (2001) The role of fault zones and melts as agents of weakening, hardening and differentiation of the continental crust: a synthesis. In: Holdsworth BE, Strachan RA, Magloughlin J, Knipe RJ (eds) The nature and tectonic significance of fault weakening. Geol Soc Spec Publ (in press)

Kohlstedt DL, Evans B, Mackwell SJ (1995) Strength of the lithosphere: constraints imposed by laboratory experiments. J Geophys Res 100 (B9):587–602

Koyi H (1997) Analogue modelling: from a qualitative to a quantitative technique: a historical outline. J Geol 20:223–238

Malavieille J (1984) Modélisation expérimentale des chevauchements imbriqués: application aux chaines de montagnes: Bull Soc Géol France 7:129–138
Molnar P, England P (1990) Temperatures, heat flux, and frictional stress near major thrust faults. J Geophys Res 95:4833–4856
Paterson MS (1987) Problems in the extrapolation of laboratory rheological data. Tectonophysics 133:33–43
Peacock SM (1996) Thermal and petrological structure of subduction zones. In: Bebout GE, Scholl DW, Kirby SH, Platt JP (eds) Subduction: top to bottom. American Geophysical Union, Washington, D.C., pp 119–134
Ranalli G, Murphy DC (1987) Rheological stratification of the lithosphere. Tectonophysics 132:281–298
Reston TJ (1990) Shear in the lower crust during extension: not so pure and simple. Tectonophysics 173:175–183
Román-Berdiel T, Gapais D, Brun JP (1995) Analogue models of laccolith formation. J Struct Geol 17:1337–1346
Román-Berdiel T, Gapais D, Brun JP (1997) Granite intrusion along strike slip shear zones in nature and experiment. Am J Sci 297:651–678
Schmid SM, Pfiffner OA, Froitzheim N, Schönborn G, Kissling E (1996) Geophysical–geological transect and tectonic evolution of the Swiss–Italian Alps. Tectonics 15:1036–1064
Schreurs G (1994) Experiments on strike-slip faulting and block rotation. Geology 22:567–570
Soula JC (1982) Characteristics and mode of emplacement of gneiss domes and plutonic domes in central-eastern Pyrenees. J Struct Geol 4:313–342
Stöckhert B, Renner J (1998) Rheology of crustal rocks at ultrahigh pressure. In: Hacker BR, Liou JG (eds) When continents collide: geodynamics and geochemistry of ultrahigh-pressure rocks.Kluwer, Dordrecht, pp 57–95
Stöckhert B, Massonne HJ, Nowlan EU (1997) Low differential stress during high-pressure metamorphism: the microstructural record of a metapelite from the Eclogite Zone, Tauern Window, Eastern Alps. In: Schreyer W, Stöckhert B (eds) High pressure metamorphism in nature and experiment. Lithos 41:103–118
Stöckhert B, Wachmann M, Küster M, Bimmermann S (1998) Low effective viscosity during high-pressure metamorphism due to dissolution–precipitation creep: the record of HP–LT metamorphic carbonates and siliciclastic rocks from Crete. Tectonophysics 303:299–319
Stünitz H (1998) Syndeformational recrystallization: dynamic or compositionally induced? Contrib Mineral Petrol 131:219–236
Stüwe K (1998) Heat sources for Eoalpine metamorphism in the Eastern Alps. A discussion. Tectonophysics 287:251–269
Stüwe K, Sandiford M (1994) On the contribution of deviatoric stresses to metamorphic PT paths: an example appropriate to low-P, high-T metamorphism. J Metam Geol 12:445–454
Thompson AB, Schulmann K, Jezek J (1997) Thermal evolution and exhumation in obliquely convergent (transpressive) orogens. Tectonophysics 280:171–184
Thompson AB, Schulmann K, Jezek J, Tolar V (2001) Thermally softened continental extensional zones (arcs and rifts) as precursors to thickened orogenic belts. Tectonophysics (in press)
Van Daalen MR, Heilbronner R, Kunze K (1999) Orientation analysis of localized shear deformation in quartz fibres at the brittle–ductile transition. Tectonophysics 303:83–108
Vissers RLM, Drury MR, Hoogerduijn Strating EH, Spiers EH, van der Wal D (1995) Mantle shear zones and their effect on lithosphere strength during continental break-up. Tectonophysics 249:155–171
Wosnitza E, Grujic D, Hofmann R, Behrmann JH (2001) A new apparatus for thermomechanical analogue modeling. In: Koyi HA, Mancktelow NS (eds) Tectonic modeling: a volume in honor of Hans Ramberg. Geol Soc Am Mem 193, chap 18 (in press)