Special Section: «Keynote lectures of the 16th Conference on Deformation mechanisms, Rheology and Tectonics (DRT)»

Foreword

This Thematic Section collects original contributions (mainly short papers) offered by the invited keynote lecturers of the 16th issue of the Conference on Deformation mechanisms, Rheology and Tectonics (DRT), held for the first time in Italy at Dipartimento di Scienze della Terra «A. Desio», Facoltà di Scienze MFN, Università di Milano, from Sept. 24 to Oct. 2, 2007. As a background it is significant to recall that the DRT Conference series represents one of the most relevant biennial event for the international community that operates in the fields of structural geology, tectonics, geodynamics, modeling, experimental deformation and rheology. The DRT Conference series was promoted in 1976 in Leiden by Henk Zwart, Richard Lysle, Gordon Lister and Paul Williams (a history of the DRT meetings until 2001 may be found in the Preface of London Geological Society Spec. Publication series n° 200, 2002). More recent DRT Conferences, organized after year 2000, were held in Neustadt, Utrecht, St. Malo-Rennes, Zuerich and then Milano. Since the beginning, the series of DRT meetings managed to bring together structural geologists, material scientists and geophysicists devoted to theoretical and experimental work; successively field investigators joined the group, that continued to debate periodically updated advances in tectonics, gained at any scale.

In the Milano 2007 Conference, discussions on 10 topics were based on presentation of 14 keynote lectures, 38 oral communications and 135 posters. Subjects encompassed advances in the investigation of lithosphere-scale tectonic mechanisms, in connection with geological and geophysical results from various analytical scales. Besides these leading themes of oral and poster presentations, discussions during two field excursions and at the Workshop, spread over tectonic mechanisms of subduction-exhumation of ophiolites (Valle Po-Valle Varaita, excursion to Monviso) and of the continental crust (Oropa-Biella, excursion to Mucrone-Monte Mars metagranitoids) and on fitting of modelling predictions with structural and petrologic natural data on orogenic metamorphites. Workshop discussion helped to individuate non-consistent results and improvements, introduced by new approaches in gathering and processing field or laboratory data (P-T estimates, age data, tectonic units size, and others...) or by different modelling approaches.

With the aim of leaving a published trace in the Bollettino della Società Geologica Italiana, most Keynote Lecturers accepted to summarize the main points of their invited contributions in the following short papers, that represent the summary of the main stream of the scientific contents of the conference sessions. The scientific sessions were grouped by topics and themes therein covered: 1) crust and mantle rheology from micro- to mega- scale (strength contrast between crust and upper mantle, structure and rheology of lithosphere scale shear/fault zones); 2) numerical and analogical modeling of deformation processes (role of rheology in mechanical models, identification of rheological 'knowledge gaps', (up)scaling and scale dependence of rheological relations, studies on consistency of field observations and laboratory-derived flow laws); 3) absolute dating vs deformation: the rate of tectonics (advances in geochronology necessary to discern the micro-scale separation of isotopic imprints and the relationships between fabrics and mineral assemblages that reflect step-like evolutions related to displacement of tectonic units in active lithosphere zones); 4) deformation-metamorphism interaction: what does condition the memory of a rock? Insights from natural data, experiment and modeling (metamorphic reaction progress and deformation history confronted...
with the activation in adjacent rock volumes of contrasted deformation mechanisms and/or strain rates); 5) interaction between magmatism and deformation: field studies, numerical models and analogue experiments (deformation and melting processes, crystallization, segregation, transport and emplacement of melts or magmas, and the flow of two-phase materials with very contrasted rheology); 6) palaeorheology (estimates of rock rheology based on natural rocks and modeling); 7) the geophysical signature of deformation processes in crust and mantle (global dynamics, and methodologies for inferring the viscosity profile of the mantle, from GPS data constrains upon predictive geophysical modeling); 8) quantitative microstructure (microstructures and textures in rocks: image analysis, electron diffraction, X-ray diffraction, neutron diffraction; focus on microstructure and texture development, including polypore rocks, experimental microstructures and predictions); 9) brittle and ductile reactivation of compositional and structural heterogeneities (multi-scale reactivation of structures and strain localization; strain-stress patterns and rheological contrasts); 10) interaction between climate, erosion and tectonics (climate, surface processes and tectonics: search for testable predictions of models). In the post-conference Workshop in Oropa-Biella, conducted over one of the most intriguing subduction-exhumed Alpine rock associations, containing remnants of pre-Alpine continental crust, attention was driven to refinement of analytical strategies and techniques in Geology, that may improve realistic investigation of geophysical processes through numerical modeling.

Guido Gosso, Anna Maria Marotta, Roberto Sabadini and Maria Iole Spalla formed the Organising Committee; the Scientific Committee of the 16th DRT Conference was formed by Ulf Bayer, Jean Pierre Brun, Stephane Bonnet, Jean-Pierre Burg, Luigi Burlini, Daniel Chatteigner, Martyn Drury, Terry Engelder, Marnie Forster, Taras Gerya, Rob Govers, Harry W. Green II, Djordje Grujic, Gordon Lister, Giorgio Pennacchioni, Giorgio Ranalli, Claudio Rosenberg, Bernard Stoeckhert, Holger Stuenitz, Janos Urai, Jean Louis Vigneresse, Igor Villa, Paul F. Williams, and Michele Zucali; the post-Conference Workshop has been shaped by Daniele Castelli, Taras Gerya, Rob Govers, Anna Maria Marotta, and M. Iole Spalla; the field leaders of pre-Conference excursion were Daniele Castelli and Roberto Compagnoni and of post-Conference excursion were Daniele Castelli, Guido Gosso, Piergiorgio Rossetti, Maria Iole Spalla, Davide Zanoni and Michele Zucali. Field guides of pre- and post-conference excursions were published on volume 9 (2007) of Quaderni di Geodinamica Alpina and Quaternaria.

The Organising Committee of the 16th Conference of the DRT series thanks the keynote speakers for their contributions; support is here gratefully acknowledged from Società Geologica Italiana, Gruppo Italiano di Geologia Strutturale and the Section of Milano of CNR-IDPA (Istituto per la Dinamica dei Processi Ambientali), that respectively sponsored and financially sustained the present publication and the other editorial activities.

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The editors of the thematic section on 16th DRT Conference
Guido GOSSO (*), (**)  
Anna Maria MAROTTA (***)  
Maria Iole SPALLA (*), (***)

(Valle Po-Valle Varaita, Milano, Oropa-Biella, Sept. 24-Oct. 2, 2007)

(*) CNR - Istituto per la Dinamica dei Processi Ambientali, Sezione di Milano.  
(**) Dipartimento di Scienze della Terra «A. Desio», Università di Milano.
Rheology in numerical models of lithosphere deformation

SUSANNE J.H. BUITER (*)

ABSTRACT

Deformation of Earth's crust and lithosphere is characterised by elastic, viscous and brittle material behaviour. The implementation of this complex rheology is a challenge for numerical models owing to variations in laboratory and natural data, the choice of equations to describe the deformation processes and the different flavours of their numerical representation. I discuss some of the current issues associated with the role of elasticity in crust- to lithosphere-scale processes, the implementation of brittle behaviour in continuum models and the use of laboratory viscous flow laws. Results of numerical models show that the style of lithosphere deformation is strongly influenced by strength contrasts between materials. Numerical models can be used to evaluate the role of such strength contrasts and of the individual rheological components by testing model sensitivity to variations in parameter values.

KEY WORDS: Rheology, brittle failure, elasticity, flow law, numerical model.

INTRODUCTION

Numerical models are a useful tool to predict lithosphere deformation styles as a function of sensitivity to mechanical and thermal properties and driving forces. Since nature is more complex than can probably ever be captured in a numerical model, simplifications need to be made of geometry, driving forces (boundary conditions) and material behaviour. Deformation of the crust and lithosphere is characterised by a complex rheology with elastic, viscous and plastic (brittle) components (fig. 1). Laboratory measurements of the properties of rocks can be used to constrain the values of several of the material properties in models. In addition, models may not only use laboratory data in this 'passive' sense, but could also constrain rheological values by inversion of natural observations (e.g. KENIS et alii, 2005). The aim of this short paper is to discuss some of the choices and challenges associated with the implementation of an elasto-visco-plastic rheology in numerical models.

A DISCUSSION OF NUMERICAL RHEOLOGY

ELASTIC BEHAVIOUR

Elastic behaviour is characterised by a linear relationship between stress and strain (fig. 1):

\[ \sigma_{ij} = 2G\varepsilon_{ij} \]  

(1)

Here \( \sigma_{ij} \) is the stress tensor, G the shear modulus and \( \varepsilon_{ij} \) the strain tensor. Elastic stresses can grow with strain in an unlimited manner and will be released once strain is removed. This thus introduces a memory of deformation in materials. Elasticity is clearly important for processes on relatively short timescales (from seismic wave propagation to post-glacial rebound), but purely elastic models have also successfully been used to simulate longer-term processes such as the deflection of the lithosphere at a trench (TURCOTTE & SCHUBERT, 2002) and under the load of volcanic islands like Hawaii (WATTS, 2001). Elastic behaviour could be important for processes that have shorter duration than the Maxwell relaxation time, which is defined as the ratio of viscosity over shear modulus. For mantle processes the relaxation time is so small (on the order of 1000 yrs) that elasticity can safely be ignored. For processes on the scale of the lithosphere it may, however, be on the order of a million years, implying that elasticity may not always be neglected. A measure of the importance of elasticity is given by the Deborah number (REINER, 1965), which is defined by the ratio of Maxwell relaxation time to the characteristic deformation time. Small Deborah numbers imply viscous behaviour (though see its limitation in viscoelastic folding (SCHMALHOLZ & PODLADCHIKOV, 1995) and further discussion in MUHLHAUS & REGENAUER-LIEB (2005)).

The numerical implementation of elastic material behaviour has partly been hampered by technical challenges related to combining large deformations of materi-
als (requiring remeshing or an Eulerian approach) with a stress history (requiring a Lagrangian approach or tracking of stresses with particles). Developments in Earth Science numerical codes now allow explicit investigation of the role of elasticity in processes on the lithosphere to upper mantle scale (MORESI et alii, 2003; MÜHLHAUS & REGENAUER-LIEB, 2005). KAUS & BECKER (2007) show that elasticity has negligible effects on the dynamics of density-driven (Rayleigh-Taylor) lithospheric instabilities, but that viscoelastic models may locally result in different stresses than purely viscous models. Elasticity will thus not significantly affect the dynamics of mantle convection or lithospheric instabilities. However, it could play a role on local scales in processes associated with folding or subduction. Elasticity may also play a role in shear band formation where stresses build up to their maximum at which failure occurs and elastic unloading takes place outside the shear band (VERMEER, 1990). In these cases, the importance of elasticity should preferably be tested.

**Plastic behaviour**

Brittle behaviour (plasticity) limits stresses in regions in the upper and lower crust and upper mantle. Brittle stresses depend on the nature of the material, its water content, the stress regime (extension or compression) and whether material is newly fractured (failure) or sliding occurs on pre-existing failure planes (friction). The empirical Amonton’s law shows a linear relation between frictional shear stress ($\sigma_{\tau}$) and normal stress ($\sigma_{n}$) and can be written as:

$$\sigma_{\tau} = \mu \sigma_{n} (1 - \lambda) + C = \tan(\varphi) \sigma_{n} (1 - \lambda) + C$$ (2)

Here, $\mu$ is the friction coefficient, $\varphi$ the angle of internal friction, $\lambda$ pore fluid factor (ratio of pore fluid pressure over lithostatic pressure) and $C$ cohesion. This equation can be written in terms of the principal stresses ($\sigma_1$ and $\sigma_3$):

$$\frac{1}{2} (\sigma_1 - \sigma_3) = \frac{1}{2} (\sigma_1 + \sigma_3) (1 - \lambda) \sin(\varphi) + C \cos(\varphi)$$ (3)

The compilation by BYERLEE (1978) for dry materials ($\lambda = 0$) shows that $\mu$ = 0.85 for $\sigma_n < 200$ MPa and $\mu$ = 0.6 for 200 < $\sigma_n < 2000$ MPa. Rock cohesion in this compilation varies between 0 and 50 MPa. Using these values, extrapolation to depths below the Moho can lead to very high stresses (on the order of 1000 MPa for a compressional stress regime). These stresses are likely lower in nature owing to a different deformation mechanism on the transition between brittle and viscous behaviour (KOHLSTEDT et alii, 1995). Many numerical models need much lower...
values for the friction coefficient than reported by BYER-LEE (1978) to reproduce lithosphere deformation as seen in nature, especially along subduction faults (e.g., GERYA et alii, 2007). It is an open question whether these low numerical coefficients point to weak faults in nature (e.g., by foliation development, mineral transformations or high pore fluid pressures) or to a special feature of the models.

Some numerical (Lagrangian finite-element) models can implement a pre-existing failure plane along which the displacements are limited by the friction coefficient (e.g., MELOSH & WILLIAMS, 1989). The challenge of this approach is the need for remeshing as fault offsets become large. Alternatively, the implementation of plasticity in (finite-element or finite-difference) continuum models results in the formation of shear bands with a finite width. Mohr-Coulomb failure follows the above equations, but the values for $\psi$ and $C$ may differ from their values in friction. Construction of the Mohr circle at yield results in a prediction for the angle of shear zones with the direction of maximum compressive stress of $45° + \psi/2$. Shear zones in compression are therefore expected to have shallow dip angles ($30°$ for $\psi = 30°$), while extensional shear zones are steep ($60°$ for $\psi = 30°$). Measurements on shear bands in sand show, however, also the Roscoe angle $45° - \psi/2$ (ROSCEO, 1970; see also VERMEER, 1990) or an intermediate angle $45° - (\psi + \psi)/2$ (VARDOUKAS, 1980). Here $\psi$ is the dilation angle (the ratio of the rate of volumetric strain and the rate of shear strain). GERYA & YUEN (2007) obtain the intermediate ($45° - (\psi + \psi)/2$) shear zone angle in their dilatant finite-difference experiments. In these types of models, the dilation angle keeps the same value throughout the deformation history. However, sand shows changes in dilation during loading. Shear zone formation is associated with dilation which reaches its maximum rate at peak failure and thereafter tends to zero when stable sliding is achieved (LOHRMANN et alii, 2003). This behaviour can be captured with distinct element methods (EGHOLM, 2007) or sophisticated plasticity models (CROOK et alii, 2006). Many numerical models that do well in simulating large deformations are incompressible ($\psi = 0°$) and shear zone dip angles in these models could in theory range between 45° and the Mohr-Coulomb angle ($45° - \psi/2$). Incompressible viscoplastic models, however, often seem to result in 45° shear zone dip angles. For this reason, MORESI & MÜHLHAUS (2006) developed an anisotropic viscosity method that gives Mohr-Coulomb dip angles. It remains, however, to be established whether incompressible viscoplastic models may not intrinsically be able to result in Mohr-Coulomb angles and whether the obtained 45° shear zone angles point to something overlooked in these types of models.

The width of shear bands depends on the numerical resolution. This implies that shear bands can become extremely narrow for very fine grids and for this reason some models have introduced intrinsic minimum length scales (de BORST & SLUYS, 1991). The use of mean stress (or dynamic pressure) instead of lithostatic pressure in the numerical implementation of Mohr-Coulomb plasticity (see equation 3) improves localisation behaviour (e.g., BUTTER et alii, 2006). In numerical models stresses are followed while they build up until the yield surface is reached. This stress build-up phase will be different for viscoelastic and viscous models. In nature, stresses will never exceed the yield stress, but in numerical models a stress overshoot can occur. Stresses then need to be brought back to yield and different techniques exist to do this (e.g., viscosity iteration or return mapping). It has not yet been clearly established if these differences in techniques could have a significant effect on numerical shear zones (see also BUTTER et alii, 2005). After yielding, associated ($\psi = \psi$) or non-associated ($\psi \neq \psi$) plastic flow occurs. This continued deformation after failure distinguishes many Earth Sciences problems from engineering-type applications.
Viscous behaviour of crustal and upper mantle rocks is described by an empirical flow law, which relates strain-rate to stress:

\[
\dot{\varepsilon} = A \sigma^n d^p \exp \left( -\frac{Q + PV}{RT} \right)
\] (4)

Here \( A \) is the pre-exponent, \( n \) the stress exponent, \( d \) grain size, \( p \) grain size exponent, \( Q \) activation energy, \( P \) pressure, \( V \) activation volume, \( R \) the gas constant and \( T \) temperature. The pre-exponent \( A \) may include melt fraction, oxygen fugacity and water content. Usually the strain-rate (\( \dot{\varepsilon} \)) is uniaxial and stress (\( \sigma \)) is the differential stress. Measured flow laws need, therefore, to be converted to effective stress and strain-rate for a general implementation in numerical models (e.g., Ranalli, 1987). At low-stress conditions, grain boundary diffusion creep (e.g. Coble creep) is in general important, whereas deformation by movement of dislocations (dislocation creep) is more characteristic for high-stress conditions. Diffusion creep is characterised by \( n = 1 \) and is grain size dependent (\( p = 3 \) for olivine (Hirth & Kohlstedt, 2003)). Dislocation creep is grain size independent (\( p = 0 \)), while \( n \gg 3 \). Diffusion and dislocation creep may occur simultaneously (they act in parallel), in which case the contributions from both mechanisms need to be taken into account.

Flow laws are determined by measuring stresses for varying strain-rate and at different temperatures. Ideally, the conditions should be such that deformation occurs by one deformation mechanism and at steady state. The laboratory conditions imply that measurements are made on small rock samples (mm), at low strains and high strain-rates (10\(^{-3}\) – 10\(^{-6}\) s\(^{-1}\)) and need to be extrapolated to geological conditions. Many of the issues associated with this extrapolation and the implementation of laboratory flow laws in models are discussed in, among others, Paterson (1987, 2001), Rutter & Brodie (1991) and Handy et alii (2001). The uncertainty involved in the extrapolation to low strain rates (10\(^{-14}\) – 10\(^{-16}\) s\(^{-1}\)), potentially different grain sizes and high strains is essentially unknown. However, support to the large extrapolations is given by similarities in microstructures between naturally and experimentally deformed materials (e.g., as discussed for quartzite by Hirth et alii (2001)). Current challenges are to formulate flow laws for polymineral rocks and flow laws that quantify the effects of water (Korenaga & Karato, 2008) and melt content.

Choosing the flow law to use in a geodynamic model is not straightforward (Ranalli, 2003; Burov, 2003). Extrapolated published flow laws for crustal and mantle rocks show a large variation in strength, thus giving a choice of weak or strong materials in models. The best approach to dealing with this uncertainty in flow law data in numerical modelling is to treat flow laws as a variable and to examine model sensitivity to viscous strength variations. Simple two-layer models of a brittle...
upper layer bonded to a linear viscous lower layer show, for example, that the number of shear zones in the brittle layer decreases as the strength contrast between the two layers increases (fig. 2) (Morese & Muhlhäusl, 2006; Butter et alii, 2008). The rheology of subducted material and the surrounding mantle has been shown to influence the dynamics of subduction zones (e.g., Billen & Hirth, 2005; 2007). Burow & Watts (2006) show that in their models a weak olivine mantle (which I infer to be wet 'heim dunite by Chopra & Paterson (1981)) results in deformation styles which are incompatible with observed lithosphere stability and deformation at subduction zones, whereas a strong upper mantle (inferred to be dry olivine by Hirth & Kohlstedt (1996)) explains the persistence of mountains and the integrity of subducting slabs (fig. 3).

CONCLUDING REMARKS

Models of deformation processes in the crust and lithosphere require a sophisticated rheology description with elastic, viscous and plastic components. I have touched upon some of the open questions associated with especially numerical plasticity and the use of laboratory flow laws in models. Plastic behaviour in continuum models results in the formation of grid-size dependent shear bands. Their dip angle has been shown to vary between the Roscoe, Mohr-Coulomb or an intermediate angle and it has until now not been clearly established whether one of these representations should be preferred. Elasticity may play a role in deformation processes on the scale of the lithosphere (e.g. subduction) and its importance should be tested for crust- to lithosphere-scale models. Extrapolated published viscous flow laws for crustal and mantle rocks show a large variation in strength, giving modellers a choice between weak and strong numerical materials. Results of numerical models show that in many cases deformation depends on the strength contrasts between layers and less on the absolute values of the material strengths (e.g., Burow & Watts, 2006; Morese & Muhlhäusl, 2006). A useful approach is to treat rheology in numerical models as a variable and to test model sensitivity to reasonable variations in the values of the rheological components.

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Modelling intrusion of mafic and ultramafic magma into the continental crust: numerical methodology and results

Jean-Pierre Burg (*) & Taras V. Gerya (*)

INTRODUCTION

Field studies and geophysical imaging indicate that granitic and non-granitic plutons have both very variable and comparable shapes and sizes. We simulated numerically intrusion of partially molten mantle rocks from a sub-lithospheric magmatic source region (SMSR). Our systematic numerical modelling results show that intrusion typically spans a few hundred kyr spanning three stages: (1) magmatic channel spreading, (2) emplacement and (3) post-intrusive subsidence and cooling. The duration of each of these stages strongly depends on the viscosity of ascending magma. Upward magma transport from sublithospheric depth is driven by the positive buoyancy of the partially-molten rocks with respect to the overriding colder mantle lithosphere. By systematically varying the model parameters we document variations in intrusion dynamics and geometry that range from funnel- and finger-shaped bodies (pipes, dikes) to deep seated balloon-shaped intrusions. We simulated numerically intrusion of partially molten mafic and ultramafic bodies into the lower crust (e.g. Rudnick & Gao, 2004) and (3) how such ultramafic/mafic magmas?

MODELLING TECHNIQUES

We decided to take advantage of recent progress in hardware and software capabilities to generate two-dimensional visco-elastic-numerical models of mafic-ultramafic intrusion emplacement incorporating in particular the temperature-sensitive properties of both magma and country rocks. Thermomechanical modelling of magma intrusion is numerically challenging because it involves simultaneous and intense deformation of materials with very contrasting rheological properties: the country, crustal rocks are visco-elastic while the intruding magma is a low viscosity, complex fluid (e.g.
We employ the 2-D code I2ELVIS (Gerya & Yuen, 2003a, 2007), which is based on finite-differences with a marker-in-cell technique. The code allows for the accurate conservative solution of the governing equations on a rectangular fully staggered Eulerian grid. New developments allow for both large viscosity contrasts and strong deformation of visco-elasto-plastic multiphase flow. The code was tested for a variety of problems by comparing results with both analytical solutions and analogue sandbox experiments (Gerya & Yuen, 2003, 2007).

We simulated numerically intrusion of partially molten mantle rocks from a sub-lithospheric magmatic source region (SMSR, fig. 1, 0 Kyr). Developments introduced for intrusion simulation allow for both large viscosity contrasts and strong deformation of visco-elasto-plastic multiphase flow, incorporating temperature-dependent rheologies of both intrusive molten rocks and host rocks (Gerya & Burg, 2007). A magmatic channel is a vertical, 1.5 km wide zone characterised by a wet olivine rheology and a low 1 MPa plastic strength throughout the lithospheric mantle. The initial thermal structure of the lithosphere is as usually assumed, with a 35 km thick crust (fig. 1, 0 Kyr) corresponding to a sectioned linear temperature profile limited by 0°C at the surface, 400°C at the bottom of the crust and 1300°C at 195 km depth. The temperature gradient in the asthenospheric mantle is 0.6°C/km below 195 km depth.

The code grey (code colour in the coloured version) identifying rock types is given in figure 1. The discrimination between «peridotite» and «molten peridotite» is thermal, separating material points (pixels) above/below the wet solidus temperature of peridotite at a given pressure. The melt fraction is strongly changing with water content, variations within few % of melt fraction at given pressure-temperature condition are possible. Therefore, it

Fig. 1 - Enlarged 20-50 × 215 km areas of the original 1100 km × 300 km reference model. Distribution of rock layers in the intrusion area during emplacement of the ultramafic body into the crust from below the lithosphere via the magmatic channel. Legend: 1) weak layer (air, water); 2) sediments; 3, 4) upper crust (3 - solid, 4 - molten); 5, 6) lower crust (5 - solid, 6 - molten); 7, 8) mantle (7 - lithospheric, 8 - asthenospheric); 9, 10) peridotite (9 - molten, 10 - crystallized); 11, 12) gabbro (11 - molten, 12 - crystallized). Time (kyr) is given in the figures. White numbered lines are isotherms in °C. Vertical scale: depth below the upper boundary of the model. Initial numerical setting of this study is shown on the leftmost section of the model (0 Myr). The lithospheric and asthenospheric mantles have the same physical properties, different grey tones are used for a better visualization of deformation and structural development. This is also true for the passive colour-layering in the upper and the lower crust. Initial and boundary conditions are detailed in (Gerya & Burg, 2007).

– Porzione estesa 20-50 × 215 km del modello di riferimento originale di dimensioni 1100 km × 300 km. Distribuzione dei livelli di roccia nell’area d’intrusione durante la messa in posto del corpo ultra-femico nella crosta da livelli sub-litosferici attraverso il canale magm atico. LEGENDA: 1) strato debole (aria, acqua); 2) sedimenti; 3, 4) crosta superiore (3 - solida, 4 - fusa); 5, 6) crosta inferiore (5 - solida, 6 - fusa); 7, 8) mantello (7 - litosferico, 8 - astenosferico); 9, 10) peridotite (9 - fusa, 10 - cristallizzata); 11, 12) gabbro (11 - fuso, 12 - cristallizzato). Il tempo (in migliaia di anni) è indicato nelle figure. Le linee bianche numerate sono isoterme in °C. Scala verticale: profondità sotto il bordo superiore del modello. La configurazione numerica iniziale di questo studio è mostrata nella sezione a sinistra del modello (0 Ma). I mantelli litosferico e astenosferico hanno le stesse proprietà fisiche; differenti toni di grigio sono utilizzati per visualizzare meglio lo sviluppo della deformazione e delle strutture. Ciò è vero anche per la stratificazione passiva della crosta superiore e inferiore. Le condizioni iniziali e al contorno sono dettagliate in Gerya & Burg (2007).
is illusory to predict the exact melt fraction at any point of the models, in particular because the simplified linear melting model implemented here (GERYA & BURG, 2007) does not allow a very high precision on this question.

**DISCUSSION**

Modelling results (cf. GERYA & BURG, 2007, for details of experiments) show that intrusion typically lasts a few hundred kyr spanning three stages: (1) magmatic channel spreading (fig. 1, 0-16 Kyr), (2) emplacement (fig. 1, 22-41 Kyr, fig. 2) and (3) post-intrusive subsidence and cooling (fig. 1, 71-1171 Kyr, fig. 3). The duration of each of these stages strongly depends on the viscosity of ascending magma.

Upward magma transport from sublithospheric depth is driven by the positive buoyancy of the partially-molten rocks with respect to the overriding colder mantle lithosphere. The gravitational balance controls the height of the
column of molten rock but not the volume of magmatic rocks below and above the Moho. The molten rocks are pooling along the crust/mantle boundary only if the lower crust is ductile and very weak (fig. 4, deep grey field or red field in the colour version), which may be expected at the base of island arcs. It seems natural that otherwise, basic – ultrabasic magma is injected into the crust, most commonly as a finger/pipe-shaped body (fig. 4, intermediate grey field or pink field in the coloured version).

Emplacement within the crust exploits the space opened by the displacement of tectonic crustal blocks bounded by localized zones of intense plastic deformation. Temperature is the important player in controlling crustal viscosities, hence either viscous or elasto-plastic mechanisms of crustal deformation, which defines modes and rates of emplacement. Early normal faults (fig. 1, 16 Kyr) produce early surface subsidence in grabens but rapidly become inverted into thrusts (fig. 1, 22 Kyr) responsible for surface uplift while the within-crust pluton inflates and rises in the crust (fig. 2).

Late emplacement phases are responsible for cooling and subsiding of the magmatic body and partial return magma flow back into the magmatic channel below the Moho (fig. 3). This event is linked to subsidence of the surface.

By systematically varying the model parameters we document variations in intrusion dynamics and geometry that range from funnel- and finger-shaped bodies (pipes, dikes) to deep seated balloon-shaped intrusions and flattened shallow magmatic sills (fig. 4). Relatively cold elasto-plastic crust ($T_{\text{Moho}} = 400^\circ\text{C}$) promotes a strong upward propagation of magma due to the significant decrease of plastic strength of the crust with decreasing confining pressure (fig. 4, intermediate and light grey fields or pink and blue fields in the coloured version). Emplacement in this case is controlled by crustal faulting and subsequent block displacements. Warmer crust ($T_{\text{Moho}} = 600^\circ\text{C}$) triggers lateral spreading of magma above the Moho, with emplacement being accommodated by coeval viscous deformation of the lower crust and fault tectonics in the upper crust (fig. 4, deep grey field or red field in the coloured version).

CONCLUSION

Emplacement of high density, mafic and ultramafic magma into low-density rocks is a stable mechanism for a wide range of model parameters that match geological settings in which partially molten mafic-ultramafic rocks are generated below the lithosphere. We expect this process to be particularly active beneath subduction-related magmatic arcs where huge volumes of partially molten rocks produced from hydrous cold plume activity accumulate below the overriding lithosphere (GERYA & YUEN, 2003b).

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Magma-controlled tectonics in compressional settings: insights from geological examples and experimental modelling

Olivier Galland (*), (**) , Peter R. Cobbold (**), Erwan Hallot (*) & Jean de Bremond d’Ars (*)

ABSTRACT

Magmatic activity tends to concentrate at tectonic plate boundaries. At rapidly convergent margins, such as the Andes, intense magmatic activity is coeval with strong tectonic shortening, and some volcanoes and magmatic intrusions have been emplaced near active compressional structures, usually major thrust faults. In order to understand the links between magmatic systems and compressional deformation structures in the upper crust, we describe the structure of an active volcano (Tromen, Argentina) and an exhumed intrusion (Boulder Batholith, U.S.A.) emplaced during compressional deformation. In those examples, magmatic systems and thrust faults exhibit geometrical and chronological relationships. We also present results of experimental modelling of magma emplacement during compression. The comparison between geological examples and experiments show close similarities. That suggests that the presence of magma influences the deformation pattern in the brittle crust. The influence of deep magma bodies is also to be explored at the scale of the whole crust during the development of active margins.

KEY WORDS: magma-controlled tectonics, compressional tectonics, Tromen volcano, experimental modelling.

INTRODUCTION

Magmatic activity mostly occurs at plate boundaries, where tectonic deformation also concentrates. Because magmatic bodies and their country rock have very contrasting rheological properties, one might expect deformation to be influenced by the presence of magmatic bodies at depth (e.g. Burov et alii, 2003). Although this problem has been addressed in the lower crust (e.g. Hollister & Crawford, 1986; Davidson et alii, 1992; Brown & Solar, 1998; Brown & Solar, 1999; Rosenberg & Handy, 2000; Barraud et alii, 2001), very little is known about the processes of such an interaction between magmatism and country rock deformation in the brittle upper crust. Which one comes first? Which one controls the other? Most of previous research has focused on deformation controlling magmatism (e.g. Hubbert & Willis, 1957; Marrett & Emerman, 1992). Here we also attempt to consider the opposite mechanism, i.e. magma-controlled deformation.

At rapidly convergent margins, such as the Andes, one might expect that horizontal compression prevents the rise of magma through the brittle upper crust (Hubbert & Willis, 1957; Hamilton, 1995). Nevertheless, volcanic activity is also common in compressional environments. Such a contradiction highlights the lack of understanding of the mechanical interplay between magmatism and deformation. We therefore address the processes of magma-controlled deformation in compressional settings. We first describe two geological examples of magmatic complexes emplaced in such settings, Tromen volcano, Neuquén basin, Argentina, and the Boulder batholith, Montana, USA. Subsequently, we present results of experimental modelling of magma emplacement during shortening. Thus, we show that magma can transport in a shortening crust, and that magma-controlled deformation processes can play an important role in the structural development of the upper crust.

GEOLOGICAL OBSERVATIONS

It is well known that magmatic activity is common at convergent margins. However, only a few studies have addressed the association between magmatic complexes and thrust faults (e.g. Hollister & Crawford, 1986; Parry et alii, 1997; Tibaldi, 2005). Noticeable examples are Tromen volcano, Neuquén province, Argentina (Galland et alii, 2007b), and the Boulder batholith, Montana, USA (Kalayk et alii, 2001). Tromen is a Pleistocene-Holocene back-arc volcano, located in the northern segment of the Southern Andes (fig. 1). It lies in a thick-
Fig. 1 - Two geological examples of magmatic complexes emplaced in compressional tectonic setting: a) Simplified geological map of Tromen volcano, Neuquén basin, Argentina. Tromen is Andean alkaline back-arc Quaternary volcano, located on top of arcuate east-verging thrust. It built up during thrusting deformation. Modified after GALLAND et alii (2007b); b) Simplified geological map of Boulder batholith, Montana, USA. Boulder batholith was emplaced into Sevier fold-and-thrust belt, during thrusting deformation. Locally around Boulder batholith, thrust front exhibits strongly arcuate trace (Helena salient). Modified after KALAKAY et alii (2001). Structures of both Tromen volcano and Boulder batholith suggest control of magmatic complexes on deformation.


Fig. 3 - a) Photograph of map view of typical model without injection. Piston (left) deformed model made of compacted silica powder. Straight thrusts accommodated shortening. Straight line locates cross section; b) Photograph and corresponding schematic drawing of cross section. Offset of horizontal markers locates faults (thrusts). Straight thrusts form at base of piston; c) Photograph of map view of typical model with injection. Straight and arcuate thrusts accommodated shortening. Poorly deformed plateau laid between straight and arcuate thrusts. Injected molten oil erupted along trace of arcuate thrust; d) Photograph and corresponding schematic drawing of cross section. Intruding oil (gray) forms basal sill. Straight thrusts form at base of piston. Arcuate thrust nucleate at leading edge of sill. Plateau lies above sill.

- a) Immagine fotografica dall’alto di un tipico modello senza iniezione. A sinistra un modello deformato a pistone, composto di silice in polvere. I sovrascorrimenti rettilinei hanno accodato il raccorciamento. La linea retta individua la sezione verticale; b) Fotografia e disegno schematico della sezione verticale. L’olio iniettato è eruttato lungo le tracce dei sovrascorrimenti arcuati; c) Immagine fotografica dall’alto di un tipico modello con iniezione. I sovrascorrimenti rettilinei e arcuati hanno accodato il raccorciamento. Tra i sovrascorrimenti rettilinei e arcuati si trovano piattaforme poco deformate. Il liquido iniettato è eruttato lungo le tracce dei sovrascorrimenti arcuati; d) Fotografia e disegno schematico della sezione verticale. L’olio che s’intrude (grigio) forma sill basali. Alla base del pistone si formano sovrascorrimenti rettilinei. Alla terminazione frontale del sill nucleano sovrascorrimenti arcuati. Il plateau si trova al di sopra del sill.
skinned fold-and-thrust belt in the western margin of the Neuquén basin (COBBOLD & ROSSELLO, 2003). Its volcanic products are unconformable upon Mesozoic strata of the basin. It built up above the hanging-wall of a major eastward-verging thrust fault (fig. 1). The Boulder batholith is a Cretaceous intrusive complex, east of the major Idaho-Bitterroot batholith (KALAKAY et alii, 2001). It was emplaced in the upper crust, within the Sevier fold-and-thrust belt (fig. 1). It has a flat-lying tabular shape, and an estimated thickness between 5 and 12 km.

Both Tromen volcano and the Boulder batholith have close chronological and structural relationships with their substrata (fig. 1):

1) They lie close to major thrust faults.
2) Their emplacement was coeval with thrusting.
3) The thrust fronts have strongly arcuate shapes around the volcano or batholith.

Geological observations on Tromen volcano and the Boulder batholith show close relationships between thrusting and magmatism (KALAKAY et alii, 2001; LAGE-SON et alii, 2001; GALLAND et alii, 2007b). They show that magma can ascend and be emplaced in compression and that thrust faults are likely to control magma transport. In addition, the arcuate thrusts around the volcano or batholith may result from the influence of magma upon the deformation pattern. The following experimental results illustrate how magmatic activity may control compressional deformation.

Fig. 2 - Schematic drawing of experimental setup (see text for explanations).
– Disegno schematico dell'impianto sperimentale (vedi il testo per le spiegazioni).

Fig. 3.
EXPERIMENTAL MODELLING

In order to study the mechanical interactions between compressional deformation and magmatic intrusion, we resorted to laboratory experiments, in which an analogue of the brittle crust shortened, while melt was intruding (fig. 2). We used (1) a cohesive fine-grained silica powder to represent the brittle crust, and (2) a molten low-viscosity vegetable oil to represent magma (Galland et alii, 2006). In the experiments, horizontal shortening and injection were coeval but independent. Shortening resulted in thrust faults, while overpressured oil formed tabular intrusions.

In those experiments where there was no injection, shortening resulted in a classical thrust wedge, in which thrusts had straight traces and were 5-6 cm apart (fig. 3; Galland et alii, 2003; Galland et alii, 2007a); the apical angle of the wedge was about 15°. In the other experiments, where there was injection, oil formed a basal sill, and the structure of the wedge was very different. Once in place, the sill lubricated the base of the model, so that arcuate thrusts formed at the leading edge of the sill (fig. 3). The distance between thrusts increased, defining a non-deformed plateau. The apical angle of the wedge was smaller than 10°. Uplift of the plateau promoted further intrusion of oil at depth. In general, the pattern of deformation and intrusion depended on the kinematic ratio R between rates of shortening and injection (Galland et alii, 2007a). The lengths of the basal sill and plateau increased with decreasing R. Thus, from our experiments we infer that a small amount of magma in a deforming brittle crust strongly modifies the deformation pattern. Intrusions control the formation of arcuate thrusts and slightly deformed plateaus by lubricating their bases.

DISCUSSION AND CONCLUSIONS

There are close similarities between Tromen or the Boulder batholith and our experimental results. According to the geological observations, melts rose and was emplaced during thrusting. In addition, thrusts have similar arcuate shapes around the magmatic complexes, which are in the hanging walls of the arcuate thrusts. Thus we infer that arcuate structures around Tromen volcano and the Boulder batholith have resulted from the interaction between compressional deformation and non-solidified magma. Similar relationships between thrusts and active volcanoes exist at Guagua Pichincha, Ecuador (Legrand et alii, 2002), Socompa, Chile (van Wyk de Vries et alii, 2001), and Taapaca, Chile (Clavero et alii, 2004). We therefore suspect that similar processes were at work in those volcanoes.

Our experimental results show that magmatic systems submitted to compression can control the formation and the shape of thrust faults in a upper brittle crust. Such magma-controlled processes are likely to be of first-order importance in the development of compressional active margins, such as the Andes, and possibly beyond the scale of the upper crust. At large scale, the potential mechanical impact of deep magmatic intrusions should be explored in models of active margins.

REFERENCES


Subduction zone earthquake mechanisms and the H$_2$O content of subducting lithosphere

H.W. Green, II (*)

ABSTRACT

Brittle fracture and frictional sliding are impossible below a few tens of km yet earthquakes occur in subducting lithosphere to ~700 km. Experimental work shows that at high pressure a small amount of low-viscosity «fluid» must be generated to enable shear failure and comparison with the earthquake distribution in the upper 300 km of subducting slabs strongly indicates that the method of earthquake initiation is dehydration of hydrous phases. In contrast, the earthquake distribution below 400 km shows no correlation with quake initiation is dehydration of hydrous phases. Indeed assuming that hydrous phases are present predicts earthquakes in places where they are not observed. Thus, the earthquake distribution suggests that H$_2$O-releasing reactions do not take place in the mantle transition zone. In addition, metastable olivine has now been detected by seismic means in 2 subduction zones. Such metastable olivine can explain the earthquake distribution and also is incompatible with the presence of H$_2$O, even in very small amounts. I conclude that subduction zones are essentially dry below 400 km.

KEY WORDS: antigorite serpentinite, dehydration embrittlement, metastable olivine.

Unassisted brittle shear failure and/or frictional sliding on pre-existing faults, the mechanisms by which materials fail by shearing at low pressure are impossible at depths in excess of a few tens of km in Earth because brittle failure and friction are strongly inhibited by increasing pressure and plastic flow is enhanced exponentially by increasing temperature (see reviews by Green & Paterson, 1965; Green, 2007). Laboratory experiments show that shearing instabilities at pressures above ~3 GPa only occur in the presence of a small amount of a phase that has an effective viscosity very much lower than the dominant material. Generation of such «fluid» can be a natural consequence of the rising temperature and/or pressure in subducting material. Examples are: (a) dehydration embrittlement (Raleigh & Paterson, 1965); (b) transformation-induced faulting (Green & Burnley, 1989; Green et alii, 1990); (c) thermal runaway leading to melting (Karato et alii, 2001; Green & Marone, 2002). The requirement of presence of a small amount of «fluid» combined with the observed distribution of earthquakes at high pressure (restriction to subduction zones) indicates that such «fluid»-producing mineral reactions must be occurring at sites of earthquake generation.

Dehydration embrittlement has been demonstrated in a variety of hydrous phases (e.g. Raleigh & Paterson, 1965; Jung et alii, 2004) and is strongly implicated as the trigger mechanism of intermediate-depth earthquakes (70-300 km) (Peacock, 2001; Hacker et alii, 2003). Antigorite serpentinite is capable of initiating such a shearing instability during dehydration under stress at pressures from 0.1 to 6 GPa in the laboratory (Jung et alii, 2004), a range over which the volume change accompanying dehydration changes from positive to negative, yet the shearing instability occurs under all conditions. The instability does not disappear when the net volume change of reaction becomes negative (ΔVreaction < 0) because the instability is not dependent upon the net volume change but rather upon the ΔV of fluid and solid components independently; under all conditions the fluid remains less dense than the solid matrix (ΔVfluid > 0) and the nanocrystalline solid reaction products remain more dense (ΔVsolid < 0); rather than canceling each other out, they both participate in the instability via formation of microcracks and microanticracks, respectively (Jung et alii, 2004). Exsolution of very small quantities of H$_2$O from nominally anhydrous phases can also trigger instability in the laboratory (Zhang et alii, 2004). It is thus highly likely that dehydration under stress of any reasonably abundant phase in subducting lithosphere can trigger earthquakes.

Here I use this logic to argue that subducting lithosphere is progressively «wrung dry» over the depth interval
50-400 km and that only very small amounts of H$_2$O exist in such lithosphere below that depth (fig. 1). The evidence is the following: (1) Earthquake frequency declines exponentially between 70 and 300 km (suggesting that the cause of the instability is being exhausted) (fig. 2); (2) the resurgence of earthquakes in the transition zone could, in the text-favors the latter. Modified after GREEN (2005).

mineral reactions (indicating that dehydration of these phases is unlikely to be involved in deep earthquakes); (3) significant amounts of DHMS can be stable only if the highly abundant phases wadsleyite and/or ringwoodite (collectively referred to as spinel in fig. 1) are fully saturated with H$_2$O, which would require more water at depths of 200-300 km than could be consistent with the known mineral possibilities in the lithosphere and their seismic properties (CHEN & BRUDZINSKI, 2001; BRUDZINSKI & CHEN, 2003) (hence saturation is extremely unlikely); (4) tectonic stresses are not necessary for earthquakes to occur (BRUDZINSKI & CHEN, 2003) – local stresses generated by the ΔV between adjacent regions of unreacted and reacted mineral assemblages are sufficient (hence the ad hoc argument that earthquakes disappear because there are no stresses is unlikely to be valid); (5) if significant H$_2$O is present in ringwoodite, wherever lithosphere passes through into the lower mantle there should be an abundance of earthquakes where that H$_2$O is released during ringwoodite breakdown (the lack of such earthquakes implies that even small amounts of H$_2$O in ringwoodite are unlikely); (6) if H$_2$O gets passed from hydrous ringwoodite to phase D at the top of the lower mantle (hence avoiding constraint #5), there should be a
flurry of earthquakes in the lower mantle during the dehydration of phase D, the last of the DHMS phases (such earthquakes are absent, strongly suggesting that phase D is also absent); (7) there is now strong seismic evidence for the presence of metastable olivine in two deep slabs (Tonga (e.g. CHEN & BRUDZINSKI, 2001) and Mariana (KANESHIMA et alii, 2007)), requiring that slabs in those subduction zones be essentially dry (because if significant H$_2$O is present, the kinetics of the ol–spinel reactions would be enhanced sufficiently that metastable olivine would not be preserved (DU FRANE & SHARP, 2007).

It could be argued that dehydration of antigorite at depths of 200-250 km should lead to enhanced dissolution of H$_2$O into olivine and pyroxenes, with that H$_2$O then carried into the transition zone. However, the empirical evidence cited above strongly suggests that doesn’t happen. One possible explanation is that the recent evidence for likely low oxygen fugacity at depths in excess of ~250 km (ROHRBACH et alii, 2007) drastically reduces the H$_2$O fugacity, subverting its solubility in silicates and/or its catalytic effect on mineral reactions.

In summary, the exponential decline in earthquake frequency between 70 and 300 km suggests exhaustion of a critical factor in their generation. The only factor that seems a logical possibility is exhaustion of the availability of hydrous phases to initiate the earthquakes. Despite the experimentally-demonstrated water-carrying capacity of the nominally anhydrous upper mantle phases and the DHMS, the abundance and distribution of earthquakes bears no recognizable relationship to their experimentally-determined properties and phase boundaries; there is a lack of earthquakes in locations where they would be expected if H$_2$O is significantly present and being liberated, and an abundance of earthquakes in locations where H$_2$O, if present, would not be expected to be liberated. In contrast, there are earthquakes where independent seismic evidence strongly suggests the presence of metastable olivine which is incompatible with the presence of significant H$_2$O. These observations singly and in concert imply that (i) dehydration embrittlement is at best a minor trigger of earthquakes in the mantle transition zone and (ii) subduction zones deeper than ~400 km are essentially dry. A corollary is that subduction does not significantly recycle H$_2$O into the deep mantle, at least not at this time.

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Non-equilibrium thermodynamics
and the coupling between deformation and metamorphism

Bruce Hobbs (*) & Alison Ord

ABSTRACT

The concept of non-equilibrium phase diagrams is explored and an example presented for the quartz-coesite reaction. The equilibrium phase boundary is relevant for systems undergoing mineral reactions only for very high temperatures and very low strain rates. The influence of strain rate is to modify the equilibrium Clausius-Clapeyron slope by a factor that is similar in magnitude to that slope. Increases in strain rate remove the non-equilibrium phase boundary further from the equilibrium phase boundary. This explains the common observation that mineral reactions proceed to completion in shear zones rather than in adjacent undeformed material.

KEY WORDS: non-equilibrium phase diagrams, entropy production, deformation enhanced reactions.

INTRODUCTION

Non-equilibrium thermodynamics increasingly has very wide application in many fields of science and engineering (Coussy, 2004; Ottinger, 2005) but apart from a flurry of activity in the 1960’s to 1980’s (Kamb, 1959; Korzhinskii, 1959; Paterson, 1973; Fisher, 1973) there has been relatively little application within geology. Recently there has been a resurgence of interest in non-equilibrium thermodynamics with respect to damage mechanics in seismology (Lyakhovsky & Ben Zion, 1997) and in structural geology/geodynamics (Rege-NAUER-LIEB & Yuen, 2003; Hobbs et alii, 2007a; Hobbs et alii, in press). In this abstract we set out to discuss some important applications of non-equilibrium thermodynamics to deforming, reacting metamorphic systems. We make a distinction between classical chemical thermodynamics (CEM) where minimisation of the Gibbs Free Energy defines the stable states and non-equilibrium thermodynamics where either minimisation or maximisation of the entropy production rate defines the stable phases.

The problem in applying non-equilibrium thermodynamics to geological problems derives from the apparent lack of a set of guiding principles that would allow progress. In any system, whether at equilibrium or not, one can define a function, The Gibbs Free Energy. This function is minimised at equilibrium and so one can proceed to define equilibrium assemblages of minerals. Another function, the entropy, is maximised at equilibrium. For non-equilibrium systems, it has never been clear, until recently, that a similar principle was available. In fact, two apparently opposing views seemed to emerge in the literature. One is due to Prigogine (1955) who claimed that in non-equilibrium systems, the rate of entropy production is minimised. The other view is due to Zeigler (1980) who claimed that the rate of entropy production is maximised in non-equilibrium systems. This apparent paradox is resolved when one understands that the Prigogine principle holds for linear steady state systems whereas the Zeigler principle holds for systems that are not constrained to be at steady state. This now opens the way to describe the evolution of geological systems that are forced out of equilibrium by continued deformation, fluid flow, heat flow and chemical reactions. We employ the Prigogine principle below to understand the construction of non-equilibrium phase diagrams and why deformation enhances the progress of metamorphic reactions.

To focus the discussion we concentrate on metamorphic rocks undergoing only deformation and chemical reactions and exclude the effects of fluid transport; an introduction to this area is given by Coussy (2004). As such, the energy dissipated during deformation and metamorphism consists of four parts: (i) That due to mechanical processes; this comprises dissipation arising from deformation and from introducing chemical components into a deforming system by chemical reaction. (ii) That arising from the flux of chemical components across gradients in chemical potential and temperature. (iii) That arising from chemical reactions and (iv) That arising from thermal conduction.

(*) CSIRO Exploration and Mining, Perth, Australia; University of Western Australia, Perth, Australia. Corresponding author - Telephone: +61 418 395 545. E-mail: bruce.hobbs@csiro.au
In many metamorphic rocks there is evidence of non-equilibrium in the form of partially reacted mineral assemblages and/or melting. An example is the Monte Murcione in the Italian Alps (Zucali et alii, 2002). It is also commonly observed in such areas that in immediately adjacent rocks, the metamorphic reactions reach completion only in the highly deformed shear zones. The two important questions are: (i) In regions where the metamorphic reactions have not proceeded to completion, are estimates of P, T conditions derived from equilibrium theory relevant and/or accurate? and (ii) What role does deformation play in promoting metamorphic reactions?

**THE NON-EQUILIBRIUM CHEMICAL POTENTIAL**

The chemical potential of a component is a quantity that measures the energy required to insert 1 mole of that component into a system under adiabatic conditions. This definition is relevant whether the system is or is not at equilibrium. Consider a system with K components. We define the non-equilibrium chemical potential, \( \mu^K \), of the \( K^{th} \) component inserted into a deforming, chemically reactive system as (Kondepudi & Prigogine, 1998; Coussy, 2004):

\[
\mu^K = \mu^K(\sigma_\eta, P, T, \xi^K) \tag{1}
\]

where \( \sigma_\eta \) is the deviatoric stress, \( P \) is the pressure, in this case, the mean stress, \( T \) is the absolute temperature and \( \xi^K \) is the extent of the chemical reaction that produces the \( K^{th} \) component. Explicit forms of this state equation are given by Paterson (1973) and Shimizu (1997). To be explicit here the pressure is

\[
\frac{P}{3} (\sigma_{11} + \sigma_{22} + \sigma_{33})
\]

Even in a system not at equilibrium, at a phase boundary the difference in the sum of the chemical potentials of the phases on either side of the boundary is zero, as is also the difference in the affinities of the reactions involved. An important difference between the non-equilibrium and the classical chemical potential is that the pressure for the non-equilibrium situation is measured by the mean stress. As the stress relaxes to hydrostatic and the chemical reactions proceed to completion, the non-equilibrium chemical potential evolves to become the classical chemical potential. From equation (1),

\[
d\mu^K = \frac{\partial \mu^K}{\partial \sigma_\eta} d\sigma_\eta + \frac{\partial \mu^K}{\partial P} dP + \frac{\partial \mu^K}{\partial T} dT + \frac{\partial \mu^K}{\partial \xi^K} d\xi^K \tag{2}
\]

or,

\[
d\mu^K = \epsilon^K_{\eta} d\sigma_\eta + V^K dP - S^K dT + A^K d\xi^K \tag{3}
\]

where \( \epsilon^K_{\eta} \) is the elastic strain of component \( K \), \( V^K \) is the volume of component \( K \), \( S^K \) is the entropy of component \( K \) and \( A^K \) is the affinity of the reaction that produces \( K \).

At this stage we focus in on a particular simple kind of chemical reaction, namely,

\[
A \rightleftharpoons B \tag{4}
\]

where \( A \) is, for example, quartz or graphite and \( B \) is coesite or diamond respectively.

Then, for example, equation (3) becomes

\[
d\mu^{coesite} = \epsilon^{coesite}_{\eta} d\sigma_\eta + V^{coesite} dP - S^{coesite} dT + A^{coesite} d\xi^{coesite} \tag{5}
\]

We define \( A = A^{coesite} - A^{quartz} \). Then arguments presented by Kondepudi & Prigogine (1998) mean that for the entropy production rate to be a minimum,

\[
A^{coesite} = \frac{L^{coesite}}{(L^{coesite} + L^{quartz})} \text{ and } A^{quartz} = \frac{L^{quartz}}{(L^{coesite} + L^{quartz})} \tag{6}
\]

where \( L^K \) is the coefficient that links the extent of reaction \( K \) to the affinity of that reaction (Kondepudi & Prigogine, 1998).

**NON EQUILIBRIUM PHASE DIAGRAMS**

From equation (3) we obtain:

\[
\frac{d(\Delta \mu)}{d\xi^K} = \frac{\Delta \xi^K}{\Delta V} \frac{d\mu^{coesite}}{d\xi^K} \tag{7}
\]

where \( \Delta(\cdot) \) is the change in \( (\cdot) \) during the chemical reaction and deformation. That is \( \Delta S = S^{coesite} - S^{quartz} \) and so on.

At a phase boundary, \( \frac{d(\Delta \mu)}{d\xi^K} = 0 \) and \( \Delta \xi^K = 0 \) hence,

\[
\frac{dP}{dT} = \frac{\Delta S}{\Delta V} \frac{\Delta \xi^K}{\Delta V} \frac{d\sigma_\eta}{dT} \tag{8}
\]

Thus, at a phase boundary the classical Clausius-Clapeyron slope at equilibrium, \( \Delta S/\Delta V \) is modified in the non-equilibrium case by a term involving the elastic strain contrast between the two phases and the temperature derivative of the deviatoric stress tensor. If the strains are elastic then this latter term involves only the temperature dependence of the elastic moduli. If the total strains arise from steady state power-law creep of the form \( \sigma_{\eta,j} = L^{-1/N} \epsilon_{\eta,j} (J_2)^{(N-1)} \exp(Q/RT) \) then,

\[
\frac{dP}{dT} = \frac{\Delta S}{\Delta V} \frac{\Delta \sigma_{\eta,j}}{\Delta V} \frac{Q}{RT} L^{-1/N} \epsilon_{\eta,j} (J_2)^{(N-1)} \exp(Q/RT) \tag{9}
\]

at a phase boundary. Hence, for a fixed strain rate, \( dP/dT \) approaches the non-equilibrium Clausius-Clapeyron slope at high temperatures but at low temperatures, the second term on the right hand side of (9) can be of similar magnitude to the classical Clausius-Clapeyron slope, \( \Delta S/\Delta V \). As the strain rate decreases at constant temperature, \( dP/dT \) approaches the classical equilibrium Clausius-Clapeyron \( dP/dT \) slope. This behaviour is illustrated in fig. 1.

The non-equilibrium phase boundary between two phases, A and B, represents the boundary between two regions where A is stable on one side and B on the other so long as the stress is maintained on the system. Although these states are stable they are not stable equilibrium states; nor are they unstable equilibrium states so that terms such as «overstepping» should not be used to describe these states.

**INFLUENCE OF DEFORMATION ON THE EXTENT OF A CHEMICAL REACTION**

Following Coussy (2004) we can write

\[
\epsilon^{K} = \frac{\partial \Psi(\epsilon_{\eta,j}, P, T, A^K)}{\partial A^K}
\]
where $\Psi$ is the Helmholtz Free Energy. We then obtain:

$$d\xi^K = \frac{\partial \Psi}{\partial A^K} d\varepsilon_{ij} + \frac{\partial \Psi}{\partial P} dP + \frac{\partial \Psi}{\partial T} dT + \frac{\partial \Psi}{\partial K^2} dA^K$$ (10)

At constant pressure and temperature this reduces to

$$d\xi^K = \alpha d\varepsilon_{ij} + \beta dA^K$$ (11)

where $\alpha$ is the parameter that measures how the extent of a reaction changes with strain and $\beta$ is a parameter that measures how the extent of a reaction depends on the affinity of that reaction. For the reaction described by (4) KONDEPUDI & PRIGOGINE (1998) discuss the form of $\beta$ for a situation corresponding to minimum entropy rate production. The relation between the extent of a reaction and the strain is discussed by COUSSY (2004).

Without proceeding to detail here, equation (11) says that the extent of a reaction is increased by increases in strain and by increases in the affinity of the reaction. Thus increased strain enhances the progress of a chemical reaction although we need to point out that this is true so long as $\alpha \neq 0$. This means that the reaction must contribute to the strain either through a volume change or through the preferred diffusion of chemical components.

However there is another important factor involved in the enhancement of chemical reaction by deformation. Fig. 2 shows the influence of strain rate on the position of the non-equilibrium phase boundary. At low strain rates a particular $P$, $T$ environment may be below a phase boundary for low strain rates but be above the phase boundary for higher strain rates. Thus a region may be such that a phase such as coesite or diamond is not stable at the ambient strain rate but is stable within shear zones in that same environment.

(i) In regions where the metamorphic reactions have not proceeded to completion, estimates of $P$, $T$ conditions derived from equilibrium theory are relevant only at high temperatures and low strain rates. If reactions have not proceeded to completion, the relevant measure of pressure is the mean stress and not the lithostatic pressure.

(ii) Deformation plays an important role in promoting metamorphic reactions through a direct influence on the extent of the reaction. High strain rates displace the non-equilibrium stability field for a particular reaction from the equilibrium field so that reactions commonly are observed to proceed to completion within shear zones and not in adjacent relatively undeformed rocks.

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Interaction between brittle fracture and ductile flow during crustal deformation

Neil S. Mancktelow (*)

ABSTRACT

Most theoretical models of crustal deformation assume that rocks either fracture according to some yield criterion (usually taken as a simple linear Mohr-Coulomb dependence on pressure) or flow in a viscous manner due to crystal plasticity or diffusion. However, direct field observation shows a much more intimate interplay between fracture and flow, with precursor fractures localizing subsequent ductile shear zones that may in turn be overprinted by discrete fractures, implying multiple cycles of brittle-ductile behaviour. Fracture and flow may be coeval over small distances, with the perturbation flow surrounding active fractures producing characteristic flanking structures. Rheological models must include this linked brittle-ductile behaviour if they are to realistically model rock deformation.

KEY WORDS: deformation, rheology, crust, lithosphere, brittle, ductile, faults, shear zones.

INTRODUCTION

The «yield-strength envelope» (Goetze & Evans, 1979) is a simple 1D, constant strain rate (CSR) concept that has been very commonly applied in numerical, analogue and conceptual models of crustal, and on a larger scale, lithospheric deformation (e.g. Ranalli & Murphy, 1987). This representation of the mechanical behaviour of rocks with increasing depth implies distinct «brittle» and «ductile» rheological layers, corresponding to Mohr-Coulomb failure or viscous flow respectively (fig. 1) although, depending on the assumed geotherm, a compositionally layered lithosphere could have several such brittle-ductile transitions (e.g. Ranalli & Murphy, 1987). A «Christmas-tree» envelope as shown in fig. 1 predicts very high differential stress at the depth of the brittle ductile transition (at least for relatively dry rocks and high strain rates), which is an artefact of the simplifying assumptions of constant strain rate and constant linear dependence of Mohr-Coulomb yield on pressure. Models involving constant force (CF) or strain-rate-dependent force (SRDF) as boundary conditions overcome the problem of unrealistically high stress levels (e.g. Porth, 2000) as does recent evidence for high-pressure brittle fracture with a weaker dependence on the confining pressure (Shimada, 1993; Zang et alii, 2007). Nevertheless, these models still imply that large regions of the crust deform exclusively by either brittle fracture or viscous crystal-plastic flow.

Increasing the strain rate (fig. 1a) will shift the brittle-ductile transition to greater depth and this effect may be enhanced if the pore fluid pressure is also increased (fig. 1b). Such a shift in the depth of the transition could occur at the tip of a downward propagating seismic fault localized in the upper crust (e.g. Ellis & Stockhert, 2004). Cycles of seismic reactivation and intervening aseismic creep could thereby lead to periodic brittle and ductile behaviour in the middle to lower crust. However, in this simple conceptual model, the major part of the lithosphere away from the relatively narrow brittle-ductile transition zone is still considered to be either brittle or ductile at any particular depth and time.

FIELD OBSERVATIONS

Nevertheless, it is becoming increasingly clear from field observation that in reality there is an intimate interplay in space and time between precursor heterogeneities (either structural or compositional), brittle fracture, fluid-rock interaction and more distributed «ductile flow» (e.g. Segall & Simpson, 1986; Guermani & Pennacchioni, 1998; Mancktelow & Pennacchioni, 2005). In particular, there are now many well-documented examples of brittle precursors localizing subsequent ductile deformation under metamorphic conditions ranging from upper greenschist to granulite facies, conditions that are typical of the middle to lower crust. Fig. 2 is an example of a «paired shear zone» from the Neves area of the Tauern window in the eastern Alps (Mancktelow & Pennac-
Fig. 1 - Simple 1D, constant strain rate ‘yield-strength envelope’ for a 30 km thick crustal section approximated by brittle Mohr-Coulomb fracture with internal friction angle of 30° and power-law viscous flow of ‘average wet quartzite’ according to PATERSON & Luan (1990). The effect of increased strain rate (a) and a combination of increased strain rate and pore fluid pressure (b) on the depth of the brittle-ductile transition in a compressive tectonic regime is shown in cartoon form.

– Semplice profilo di resistenza monodimensionale per una sezione di crosta spessa 30 km, approssimato da frattura fragile Mohr-Coulomb con angolo di frizione interna di 30° e da una legge esponenziale di flusso viscoso per una «quarzite idrata media» secondo PATERSON & Luan (1990). L’effetto dell’aumento della velocità di deformazione (a) e della velocità di deformazione combinata alla pressione dei fluidi nei pori (b) sulla profondità della transizione fragile-duitile in un regime tettonico compressivo è mostrata in forma schematica.
CHIONI, 2005; PENNACCHIONI & MANCKTELOW, 2007). An initial precursor fracture has allowed fluid infiltration, with the development of a thin epidote-rich vein flanked by a bleached zone to either side as the result of fluid-rock interaction. Most of these precursor structures are sealed joints and show no discernible shear offset prior to reactivation, even at the microscopic scale. Such joints, with lengths on the order of tens of metres and widths less than a millimetre, can only develop by extensional failure and not by localization of crystal plastic deformation. During dextral reactivation under amphibolite facies conditions, heterogeneous ductile shearing was localized on the boundaries of this bleached zone to develop the characteristic paired geometry. In rare cases, subsequent straight discrete fractures offset such ductile shear zones, and these fractures may themselves in turn be loci for localized shearing. These field relationships provide evidence for cycles of discrete brittle fracture and more distributed ductile shearing within a small area, although the time scale involved cannot be determined.

Distributed ductile deformation and localized slip on discrete fractures can occur synchronously. A good example of this is seen in fig. 3, from the same general area as fig. 2. Precursor discrete fractures were sealed with newly grown quartz, plagioclase, biotite, and garnet indicating metamorphic temperatures consistent with regional peak metamorphic temperatures of around 550-600°C. These fractures show a left-stepping geometry typical of Riedel fault development in an overall dextral shear. The sealed fractures have subsequently been reactivated, again in dextral shear, with the formation of a compressive bridge in the left-stepping zone. The compressive bridge develops a distributed foliation – a typical «ductile» structure – whereas slip on the precursor fractures remains localized (at least initially) on the fracture itself. Completely analogous structures were described by PENNACCHIONI (2005) from the Adamello tonalite, with the interplay between slip on discrete fractures and distributed ductile strain in the stepovers also occurring under amphibolite facies conditions during cooling of the pluton. It is not necessarily the case that there are distinct periods of brittle and ductile behaviour, as would be implied by the models of fig. 1. Flanking structures developed around brittle faults (e.g. PASSCHIER, 2001; GRASEMANN & STUWE, 2001; EXNER et alii, 2004; KOCHER & MANCKTELOW, 2005) are particularly clear examples of interacting brittle-ductile deformation, because their geometry can only be explained if discrete slip occurred synchronously.
with the more distributed surrounding ductile flow. In fact, models assuming perfectly free slip on an isolated fracture within a viscous surrounding matrix best explain the observed flanking geometry and can be used to estimate both the amount of general shear and the kinematic vorticity number (Kocher & Mancktełow, 2005). Examples of flanking structures developed in calcite marbles under amphibolite facies conditions (e.g. fig. 4) demonstrate that brittle fracturing can play an important role even in weak rocks at high temperature conditions generally taken to imply exclusively ductile or viscous behaviour.

The interplay between fracture and flow can still occur at great depth, even under (ultra-) high pressure conditions. Initial seismic faulting under eclogite facies conditions in the Bergen Arcs of western Norway allowed water infiltration into otherwise dry rocks, localizing the transformation to eclogites and also localizing ductile shear zones on these precursor fractures (e.g. Boundy et alii, 1992). Both brittle fracture and ductile shearing occurred under the same metamorphic conditions, with the crucial factor being the influence of water and fluid-rock interaction. A recent study by Fusseis et alii (2006) has also shown that shear zone localization in strongly anisotropic schists can also be controlled by brittle precursors, that shear zones lengthened by a combination of fracturing and mylonitic shearing, and that the overall geometry strongly reflects the interplay between brittle fracture and ductile flow.

CONCLUSIONS

Field-based studies in granitoids, schists, and even eclogite facies gneisses have established that brittle precursors and fluid-rock interaction may be critical for the initiation and localization of «ductile» shear zones in otherwise relatively homogeneous rocks. It follows that natural deformation structures and the bulk rheology and dynamics of the crust (and lithosphere) cannot be understood in terms of simple, non-interacting brittle and ductile models. More realistic models of lithospheric deformation must involve a combined elasto-visco-plastic rheology, also considering the importance influence of fluid-rock interaction and compositional heterogeneity on strain localization.

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Relict meso- and micro-structures in orogenic garnet peridotites as tracers of mantle dynamics and metasomatism at convergent plate margins

M. Scambelluri (*), H.L.M. Van Roermund (**) & T. Pettke (***)

ABSTRACT

At subduction zones, a number of geologic processes are caused by influx in the supra-subduction mantle wedge of fluid phases released by the subducting plates. The distribution of fluids in such settings affects the mineralogical, chemical and structural transformation of rocks and, hence, the survival of relict minerals and structures of previous events. These features can be investigated by means of field-based studies of high and ultrahigh-pressure (HP-UHP) orogenic terrains that contain mantle wedge materials texturally sampled by the subducting plates. Here we review two examples of garnet peridotites hosted in HP-UHP continental crust, which record different P-T stories: (i) shallow spinel-facies lithospheric mantle wedge down-dragged to depth during subduction and recrystallized to garnet + amphibole assemblages due to the infiltration of crust-derived fluids (Western Gneiss Region, Norway); (ii) transition-zone mantle upwelled and accreted to cratonic roots, and involved in subduction-zone recrystallization at 200 km depth enhanced by crustal fluids (UHP garnet peridotites, Western Gneiss Region, Norway). Our textural and petrologic study shows that the water distribution controls development of the new assemblages and the metasomatic imprints of these rocks, independently on the depth and degree of metamorphism. We conclude that mantle re-fertilization by crust-derived subduction fluids is an effective mechanism working on a 100-200 km depth range.

KEY WORDS: Convergent margins, mantle wedge, fluid, UHP metamorphism, mantle metasomatism.

INTRODUCTION

Convergent plate margins are highly evolving environments, where significant physical and chemical changes affect the subducting plates and the overlying mantle wedges. At subduction zones, mass transfer, hydration and melting of the supra-subduction mantle domains is caused by the influx of fluid phases released by the subducting plates. The distribution of fluids and of deformation in such settings affects the extent of mineralogical and structural transformation of rocks and, hence, the survival of relict minerals and structures of previous events.

The interplay of deformation, metamorphism and fluid infiltration at convergent margins can be investigated through field-based studies of the high (HP) and ultrahigh-pressure (UHP) rocks exposed in mountain buildings, which represent excellent natural observatories on the Earth interiors and on subduction dynamics (Chopin, 2003; with references). An increasing number of studies has recently shown that, besides the oceanic and continental lithosphere recording prograde subduction metamorphism, the HP-UHP terrains contain mantle wedge materials textonically sampled by the subducting crust (Brueckner, 1998; Nimis & Morten, 2000; Zhang et alii, 2000). Such mantle rocks may preserve phase transitions attained at exceptional depths, i.e. in the range of 200 to 350 km, representing the deepest transformations discovered in mantle rocks textonically exposed at the surface (Dobrzynetska et alii, 1996; Van Roermund & Drury, 1998; Spengler et alii, 2006; Song et alii, 2004; Scambelluri et alii, 2008).

So far, much research has been focussed on the slabs and still few are the observations of mantle wedge peridotites, which are the least known pieces of the subduction process.
factory. Information can be achieved from orogenic garnet peridotites of mantle wedge origin, which may be viewed as metre to kilometre-scale tectonic ‘xenoliths’ sampled at different depths by the subducted continental plates (fig. 1). These peridotites can preserve old events, pre-dating their engagement in the subducted crust and enabling to design the long-term mantle dynamics at convergent settings.

Here we review two case-studies of garnet peridotites hosted in subducted continental basements, which record different P-T, i.e. physical, trajectories prior to their uptake in the crust. The two examples correspond to (fig. 1): (i) shallow lithospheric mantle down-dragged to depth by corner-flow motion in the mantle wedge (Ulten Zone garnet peridotites, Eastern Alps, Italy); (ii) transition-zone mantle upwelled and accreted to cratonic roots (UHP garnet peridotites, Western Gneiss Region, Norway). In both cases the mantle rocks were flushed and metasomatized by incompatible element-rich fluids sourced from the continental crust, prior to or during their uptake in the subducting slabs. Such fluids are crucial to rock recrystallization and to the preservation of former structures.

FIELD-BASED CASE STUDIES

THE HP GARNET PERIDOTITES FROM THE ULTEN ZONE, ITALIAN EASTERN ALPS

The Ulten Zone peridotite bodies are hosted by Variscan high-pressure migmatites (Godard et alii, 1996). They are porphyroclastic spinel peridotites (T=1200°C; P=1.5 GPa) recrystallized into fine-grained garnet +
amphibole peridotites \( (T = 850°C; P_{\text{max}} = 3 \text{ GPa}) \) in response to corner-flow inside a mantle wedge and slicing into a subducted continental slab (Obata & Morten, 1987; Nimis & Morten, 2000; Tumiati et alii, 2003). The rock textures change from coarse porphyroclastic in the spinel-facies high-temperature domain of the mantle wedge, to coronitic and mylonitic in the lower-temperature (garnet-facies) hydrated region over the slab (fig. 2). The coronitic garnet peridotites display relict porphyroclastic textures where spinel (the black mineral spots in fig. 2A) is contoured by coronitic garnet associated with minor amounts of amphibole. The highly sheared garnet peridotite mylonites (fig. 2B) are significantly enriched in amphibole (up to 20\% modal amphibole), suggesting an open-system fluid influx in the highly deformed zones. These features indicate that localized aqueous fluid infiltration in such wedge domains highly enhanced deformation development and chemical changes in rocks.

The incompatible element-enriched signature of the garnet + amphibole peridotites clearly indicate that the incoming aqueous fluids determined a new, metasomatic, geochemical imprint of the garnet peridotites (Rampone & Morten, 2001; Scambelluri et alii, 2006). In particular, the significant enrichment in Sr, Pb and H\(_2\)O of the garnet + amphibole peridotites (fig. 3) indicate that the fluid phase carried crust-derived components.

In the Uiten Zone peridotites, the heterogeneity in the fluid flow patterns enabled survival of the precursor...
anhydrous spinel-facies domains unaffected by fluid influx and by garnet-facies recrystallization aside of highly sheared mylonitic hydrated garnet peridotites. Once engaged in the crust, the peridotite lenses behaved as rigid bodies which escaped the exhumation tectonics that mostly involved the surrounding softer gneisses.

THE UHP GARNET PERIDOTITES AND WEBSTERITES FROM WESTERN NORWAY

The UHP gneisses of the Western Gneiss Region (Norway) record subduction to the coesite and to the diamond stability fields (Smith, 1984; Dobrzhinskaya et alii, 1995; Van Roermund et alii, 2002). The diamond-facies gneisses host garnet peridotites and websterites recording uprising from extraordinary depths prior to uptake in the continental slab. These ultramafic rocks (exposed in the islands of Otrøy and Bardane) derive from depleted Archean transition-zone mantle upwelled and accreted to a cratonic lithosphere (Van Roermund & Drury, 1998; Spengler et alii, 2006). Evidence for this Archean story are decimetric garnets preserved in Otrøy, hosting orthopyroxene and clinopyroxene exsolved from precursor ultradepth majoritic garnet (up to 20 volume % pyroxene component). Majoritic garnets form above 5 GPa through the progressive, pressure-dependent, incorporation of pyroxene into garnet, leading to formation of supersilicic garnets with Si exceeding 3 atoms per formula unit (Ringwood & Major, 1971; Akaoji & Akimoto, 1977; see Griffin, 2008 for a short review). The Archean garnets from Otrøy contained up to 20 volume % pyroxene, now exsolved as intercrystalline grains and as coarse exsolution lamellae inside garnet. Fig. 4A reports such pyroxene lamellae (20-30 µm thick, hundreds µm long lamellae), exsolved under high-temperatures, as shown by the garnet/cpx REE distribution (Spengler et alii, 2006). The high amounts of pyroxene exsolutions in this garnet indicates provenance from the transitions-zone, 350 km deep, mantle. These pyroxenes and garnets display REE-depleted compositions, indicating that the original majorite crystallized in extremely refractory peridotite after high degrees of partial melting during the Archean upwelling history.

This ultradepth mantle was involved in the 430 Ma-old Scandian subduction cycle, forming new clinopyroxene + orthopyroxene + phlogopite + garnet + spinel + carbonate, which host microdiamond-bearing inclusions precipitated by circulating COH silicate fluids (Van Roermund et alii, 2002; Carswell & Van Roermund, 2005). This stage is mostly recorded in the island of Bardane. The circulating subduction fluids also crystallized new majoritic garnet at grain boundaries and in microveins. This new majoritic garnet hosts maximum 1.5 volume % thin px needles (5 mm thick, 100 mm long; fig. 4B) exsolved under low-temperatures, as indicated by the garnet/cpx REE distribution. The amounts of pyroxene needles exsolved indicate that the new majoritic garnet formed at 7 Gpa and 900-100°C (Scambelluri et alii, 2008). Pictures in fig. 4 are taken at the same magnification and refer to the Archean high temperature majorite (fig. 4A) and to the Scandian subduction majorite (fig. 4B) from Western Norway. They emphasize
the different size of pyroxene exsolutions in these garnets, which may be taken as textural evidence for distinct exsolution temperatures and geologic environments.

The majorities of fig. 4 also display significantly different trace element compositions. The subduction majorite has flat REE patterns (fig. 5); this contrasts with the REE depleted composition of the Archean majorite and indicates re-fertilization of the starting depleted peridotite by crust-derived fluid at 200 km depth. Distinct generations of majoritic garnet thus survive in the same terrain, displaying distinct textures, compositions, and exsolution temperatures.

The majorite microstructures and compositions enable to discriminate between different crystallization environments: hot sub-cratonic lithosphere vs. colder subduction-zones. Crystallization of the new majoritic assemblage in Bardane was fluid induced, the archean transition zone majorites in Otrøy likely escaped fluid infiltration and survived the subduction event.

**CONCLUSIVE REMARKS**

Our study shows that continental crustal slabs subducted to variable depths entrain mantle wedge peridotites, the relict structures of which emphasize the stories and fate of the subcontinental mantle through time. The water distribution controls development of the new assemblages and the preservation of relics, independently on the depth and degree of metamorphism. Comparison of the trace element compositions of clinopyroxenes pertaining to the metasomatic HP and UHP subduction assemblages in Ulten and Bardane emphasizes a strong similarity in the Light Rare Earth Elements and in the Large Ion Lithophile Elements of such phases (fig. 6; data from Scambelluri et alii, 2006; 2008). Since the clinopyroxenes exchanged components with the incoming metasomatic fluids, this similarity indicates that the fluid phase compositions did not change dramatically with depth and implies that mantle re-fertilization by crust-derived subduction fluids is an effective mechanism working on a 100-200 km depth range.

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Estimation of palaeorheology from buckle-fold geometries

Stefan M. Schmalholz (*) & Neil S. Mancktelow (*)

ABSTRACT

The geometry of a natural single-layer fold train is investigated to estimate the effective viscosity ratio between the folded layer and the surrounding medium. Different methods based on analytical solutions for folding yield consistent estimates for the range of viscosity ratios of between 20 and 70 and for values of the power-law exponent of the layer of between 1.8 and 5. The error range for such viscosity ratio estimates is roughly on the order of a factor of 2.

KEY WORDS: buckling, folding, power-law rheology, strain estimation, palaeorheology.

INTRODUCTION

Geological structures as observed in the field can be regarded as results of natural rock deformation experiments. However, in contrast to laboratory or numerical rock deformation experiments, in nature only the final geometry is directly measurable and the initial and boundary conditions, as well as the rock rheology at the time of deformation, are unknowns that cannot be immediately established. Indeed, one of the main aims of structural geologists is to reconstruct both the kinematics and dynamics of the natural deformation history from the observed structures.

This study focuses on buckle-fold structures and aims to assess the rheology and effective strength ratio (or «competence» contrast) between the folded layer and the embedding matrix at the time of folding. In the framework of continuum mechanics, as applied in this study, rheology means the constitutive equations relating stress tensor components to strain or strain rate tensor components (e.g. Johnson & Fletcher, 1994). In this sense, this study aims to estimate the effective rheology on the scale of observation (i.e. the folded layer and the embedding medium). This effective rheology may differ from the rheology on the micro-scale (i.e. on the scale of individual small crystals building the rock layer).

There are several rheologies, such as elastic or viscous, that can potentially describe the deformation behaviour of folded layers (fig. 1). However, a purely elastic rheology is unlikely to be appropriate for buckle-fold formation, because elastic strains are very small (<<1%). An elastoplastic rheology is also unlikely to be dominant during buckle-fold formation, because elastoplastic deformation causes localized shear bands (i.e. bifurcation, e.g. Vermeer & de Borst, 1990) and not distributed deformation as observed in natural buckle-folds. The remaining and most likely candidates for the rheology generating buckle-folds are viscoelastic (Schmalholz & Podladchikov, 1999; Mancktelow, 1999), linear viscous (Biot, 1961) and power-law (Fletcher, 1974) rheologies. This study focuses on folding of power-law layers (e.g. Fletcher, 1974; Johnson & Fletcher, 1994) embedded in a viscous (Newtonian) matrix. The power-law rheology includes the Newtonian case when the power-law exponent is 1.

The aims of this study are (i) to present methods for estimating the effective viscosity ratio and power-law exponent of a folded layer and (ii) to discuss the accuracy and reliability of such palaeorheology estimates.

METHOD AND RESULTS

The two interfaces of the natural buckle-fold train were digitized using MATLAB (fig. 1). The slope of the two interfaces (i.e. the derivative of the Y-coordinate with respect to the X-coordinate, fig. 2) was analyzed with an algorithm that determines the sign of the slope. Every position on the interface at which the slope of the interface changes its sign was identified and marked with a diamond-shaped symbol (fig. 2). These locations represent potential fold hinges. Due to natural irregularities, sometimes more than one hinge position was identified (fig. 2). Representative fold hinge positions were determined by personal interpretation and five individual buckle-folds were identified within the fold train. The ratio of fold-span to layer thickness of the individual folds varies between 4.6 and 10.6. The average ratio is about 8. In addition, the Fourier spectrum was calculated (using the MATLAB fast fourier transform, fft, function) for the top, bottom and averaged layer interface (fig. 2). The
Fig. 1 - The photograph shows a natural folded quartz vein embedded in shale from Vale Figueiras, SW Portugal, with the digitized top and bottom interface of the folded quartz vein given below.

La fotografia mostra una vena naturale di quarzo piegata, incassata in argillite, a Vale Figueiras, Portogallo SO; nell’immagine sottostante sono rappresentate le superfici digitalizzate di tetto e di letto della vena di quarzo piegata.

Fig. 2.
three Fourier spectra are significantly different. The best Fourier representation of the layer shape is presumably the Fourier spectrum of the averaged interface, because the averaged interface is least sensitive to natural interface irregularities around the fold hinges (because such irregularities are smoothed in the process of averaging). The Fourier spectrum of the averaged layer interface yields the largest amplitude at a ratio of wavelength to layer thickness of about 9. This value is close to the value of 8 obtained by using the distances between fold hinges, i.e. the fold-span.

The ratio of fold-span to layer thickness and the ratio of wavelength to layer thickness can be used to estimate the viscosity ratio and the power-law exponent of the layer. In the current analysis, the analytical solution of FLETCHER (1974) is used. This provides values for the ratio of dominant wavelength to thickness and for the maximal growth rate (normalized against the background shortening rate) as a function of the viscosity ratio and the power-law exponent of the layer (the embedding matrix is assumed to be Newtonian, fig. 3). The measured ratios of fold-span to thickness (about 8) and of wavelength to thickness (about 9) are used as lower and upper bounds for the ratio of dominant wavelength to thickness. No correction for the shortening of the dominant wavelength, as described in SHERWIN & CHAPPLE (1968), has been applied. Furthermore, analytical folding solutions (e.g. JOHNSON & FLETCHER, 1994) and numerical simulations of folding show that values of the maximal growth rate should be greater than about 10, in order to generate observable buckle-folds with a more or less constant layer thickness. The analytical solution shows that values of the power-law exponent between 1.8 and 5 and viscosity ratios between 20 and 70 yield values for the ratio of dominant wavelength to thickness between 8 and 9 and values of the maximal growth rate larger than 10 (gray patch in fig. 3).

Additionally, the values of the amplitude, A, wavelength, $\lambda$ (i.e. horizontal hinge distance), and thickness, H, of the five individual folds shown in fig. 2 have been measured and the corresponding ratios $H/\lambda$ and $A/\lambda$ plotted on the strain map developed by SCHMALHOLZ & POHLADCHIKOV (2001). This strain map can be used to estimate the amount of shortening and the viscosity ratio from buckle-fold geometries and includes a correction for the shortening of $\lambda$ and thickening of H during folding. The individual values are distributed and the average value for $H/\lambda$ and $A/\lambda$ is close to the dashed line for a viscosity ratio of 50 (plus symbol in fig. 4). This is in agreement with the viscosity ratio estimates using the analytical solution of FLETCHER (1974), with values between 20 and 70 (fig. 3).
DISCUSSION AND CONCLUSIONS

The analytical solution for folding of a power-law layer embedded in a Newtonian matrix (fig. 3, Fletcher, 1974) shows that there is no unique solution for the dominant wavelength because it depends on both the viscosity ratio and the power-law exponent. The range of possible solutions can be reduced by using the fact that fold growth rates should be at least ten times larger than the shortening rate in order to generate observable folds with more or less constant layer thickness. Smaller growth rates produce fold shapes with strongly varying layer thickness due to deformation that is significantly affected by layer thickening (e.g. Johnson & Fletcher, 1994).

Potentially the best method to quantify a fold shape is by calculating the Fourier spectrum of the averaged interface of the folded layer, because no interpretations and decisions concerning the positions of fold hinges have to be made. However, several individual folds should be present within a fold train to yield a representative Fourier spectrum. Also, if the folds have overthrown limbs, i.e. more than one vertical coordinate corresponds to one horizontal coordinate, then it is not possible to calculate a Fourier spectrum. In such cases, the more interpretative method of defining fold hinge positions has to be used.

The analysis presented here shows that effective viscosity ratios and power-law exponents can be estimated from buckle-fold geometries, but the error range is roughly on the order of a factor of 2 (i.e. between 25 and 100 for an estimate of 50). Better accuracy is difficult to obtain due to (i) the natural irregularities of fold shapes that are considerably affected by the initial perturbation geometry of the layer (e.g. Mancktełow, 1999, 2001) and (ii) by the non-uniqueness of the buckling process, which means that different combinations of material parameters can generate the same fold shape with a particular dominant wavelength.

The estimates for the effective viscosity ratio between 20 and 70 and for the power-law exponent of the layer between 1.8 and 5 are well within the range of experimentally confirmed values for mechanically strong quartz within mechanically weak shale (e.g. Carter & Tsen, 1984). Additional constraints on the rheology may be obtained from microstructural observations of the folded vein, but such observations were not available for the investigated fold.

REFERENCES

Shallow earth rheology from glacial isostatic adjustment constrained by GOCE

L.L.A. Vermeersen (*) & H.H.A. Schotman (**), (*)

ABSTRACT

The Earth’s asthenosphere and lower continental crust can regionally have viscosities that are one to several orders of magnitude smaller than typical mantle viscosities. As a consequence, such shallow low-viscosity layers could induce high-harmonic (spherical harmonics 50-200) gravity and geoid anomalies due to remaining isostasy deviations following Late-Pleistocene glacial isostatic adjustment (GIA). Such high-harmonic geoid and gravity signatures would depend also on the detailed ice and meltwater loading distribution and history. ESA’s GOCE satellite mission, scheduled for launch summer 2008, is designed to map the quasi-static geoid with centimeter accuracy and gravity anomalies with milligal accuracy at a resolution of 100 kilometers or better. This might offer the possibility of detecting gravity and geoid effects of low-viscosity shallow earth layers and differences of the effects of various Pleistocene ice decay scenarios. For example, our predictions show that for a typical low-viscosity crustal zone GOCE should be able to discern differences between ice-load histories down to length scales of about 150 km. One of the major challenges in interpreting such high-harmonic, regional-scale, geoid signatures in GOCE solutions will be to discriminate GIA-signatures from various other solid-earth contributions. It might be of help here that the high-harmonic geoid and gravity signatures form quite characteristic 2-D patterns, depending on both ice load and low-viscosity zone model patterns.

KEY WORDS: Glacial Isostatic Adjustment, Crust and Mantle Rheology, GOCE, Gravity Anomalies and Geoid.

INTRODUCTION

In Summer 2008 ESA’s Gravity and steady-state Ocean Circulation Explorer (GOCE) satellite will be launched. GOCE will observe the Earth’s gravity field with unprecedented resolution down to 100 km and accuracies down to 1-2 cm in geoid height and down to 1 mgal in gravity anomaly (e.g., Visser et alii, 2002). Such high resolution and accuracies, with almost uniform coverage, have some interesting prospects for the solid-earth sciences, notably for the shallow parts of the Earth. One example of this is glacial isostatic adjustment (GIA). In earlier studies (Vermeersen 2003; van der Wal et alii, 2004; Schotman & Vermeersen, 2005; Schotman et alii, 2007a) we have shown that crustal and asthenospheric low-viscosity zones can induce geoid and gravity signatures that are above accuracy and resolution thresholds of expected GOCE performance. Here we will concentrate on the question whether it might be possible to discern the effects of lateral variations in earth structure, including regional crustal and asthenospheric low-viscosity zones. Continental crust can have zones of low viscosity for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension. In order to give an impression of the perturbations that such low-viscosity zones can give on present-day GIA-induced geoid anomalies we model resulting GIA geoid anomalies we model resulting GIA geoid anomalies for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension. In order to give an impression of the perturbations that such low-viscosity zones can give on present-day GIA-induced geoid anomalies we model resulting GIA geoid anomalies for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension. In order to give an impression of the perturbations that such low-viscosity zones can give on present-day GIA-induced geoid anomalies we model resulting GIA geoid anomalies for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension. In order to give an impression of the perturbations that such low-viscosity zones can give on present-day GIA-induced geoid anomalies we model resulting GIA geoid anomalies for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension. In order to give an impression of the perturbations that such low-viscosity zones can give on present-day GIA-induced geoid anomalies we model resulting GIA geoid anomalies for regions with a larger than average heat flow. Generally such areas can be expected to occur in regions that are under extension.
Fig. 1 - Spherical earth model.
– Modello di Terra sferica.

Fig. 2 - a) Difference in geoid anomalies triggered by the standard earth model of fig. 1, that has no low-viscous lower crust, and the earth model that has the low-viscous lower crust for Northern Europe; b) Difference in geoid anomalies with respect to fig. 2a, assuming that the Baltic Shield is not underlain by the low-viscosity zone of the inset in fig. 1.
– a) Differenza nelle anomalie del geoide indotte dal modello di Terra standard mostrato in fig. 1, senza crosta inferiore a bassa viscosità, e il modello di Terra che ha una crosta inferiore a bassa viscosità nel Nord Europa; b) Differenza nelle anomalie del geoide rispetto alla fig. 2a, assunto che sotto lo Scudo Baltico non ci sia la zona a bassa viscosità indicata in fig. 1.
erally homogeneous one, we use the earth model as depicted in fig. 1 as standard model. This model consists of an inviscid core, viscoelastic lower and upper mantle and elastic upper part. In this elastic upper part a low-viscosity lower crust is sandwiched between the elastic upper crust and the elastic lithosphere below. Here we will not model the indicated asthenosphere explicitly, but consider it to be part of the upper mantle with the same viscosity as the upper mantle. Results from asthenospheric low-viscosity zones can be found in Schotman et alii (2008). Elastic parameters and radial density profile are based on PREM (Dziewonski & Anderson, 1981). For laterally homogeneous, self-gravitating, spherical earth models we use the normal mode technique as described in Sabadin & Vermeersen (2004). The rheology is a simple linear Maxwell viscoelastic one. Lower mantle viscosity is five times the Haskell value, while upper mantle viscosity is half the Haskell value. Values for other parameters and variables are indicated in the figure. Computations with the laterally varying earth model are performed by means of the finite-element package ABAQUS (e.g., Wu et alii, 2005). The earth model we use is a viscoelastic halfspace model with the same layering as the laterally homogeneous spherical earth model, but it is not self-gravitating. However, it has been shown in, e.g., Schotman et alii (2008) that the lack of self-gravitation is partly compensated by the lack of sphericity. Furthermore, long-wavelength differences largely cancel out for the small-scale perturbation signatures related to the shallow crustal low-viscosity zone that we are interested in here. A validation of using finite elements for computing geoid height perturbations can be found in Schotman et alii (2008). It is assumed that the complete region has the same homogeneous lithospheric thickness as in the standard earth model in these finite-element computations. For modeling results with varying thicknesses we refer again to Schotman et alii (2008). The Late-Pleistocene ice mass decay model is based on ICE-5G of Pelteir (2004), although we have also considered other ice models like RSES of Lambeck et alii (1998). It turned out, however, that the background ice decay model has a negligible influence on the spectral characteristics associated with the crustal low-viscosity zone contributions to geoid and gravity anomalies (Schotman et alii, 2008). Spatial patterns of these geoid and gravity anomalies can differ considerably, of course, as individual ice sheets from various ice models can differ in position.

**MODELLING RESULTS**

Fig. 2a shows the difference in geoid anomalies triggered by the standard earth model of fig. 1 that has no low-viscous lower crust and the earth model that has the low-viscous lower crust for Northern Europe. Here the low-viscosity zone is taken as a laterally homogeneous layer, so also the Baltic Shield is (unrealistically) presumed to have this low-viscosity crustal layer. The geoid height perturbations are clearly above the expected accuracy level of 1 cm of maps that GOCE will deliver, at many places the differential signal will even be more than an order of magnitude larger than this 1 cm level. The contours with number «2» or «–2» signify differences between modelling results obtained with the (spherical earth, self-gravitating) normal mode method and the (halfspace, non-self-gravitating) finite element model. The numbers are the difference in cm between the two modelling results, showing that the modelling results differences between those obtained by the normal mode approach and the finite element model are only slightly larger than the expected uncertainties in the GOCE data. This result illustrates what has already been mentioned in the former section: effects of self-gravitation and sphericity partly annihilate one another, specifically for small-scale signatures in geoid anomaly. Fig. 2b shows the effects of lateral heterogeneities. Depicted is the difference in geoid anomalies with respect to fig. 2a assuming that the Baltic Shield is now not underlain by the low-viscosity zone of the inset in fig. 1. It is obvious from fig. 2b that differences between the lateral and non-lateral results are generally small outside the Baltic Shield area and become most prominent prominent underneath those regions (i.e., the Baltic Shield) that do not have the crustal low-viscosity zone any longer. The differences are large enough compared to the expected performance of GOCE, even up to one order of magnitude, that the effects of lateral variations on the geoid might become discernable in GOCE data. Also the resolution should not be a problem: most of the patches in figs. 2a and 2b extend over more than 100 km up to even hundreds of km for the more elongated structures. These spatial geoid anomaly patterns, apart from spectral signatures, might help in identifying GIA-induced contributions coming from the effects of lateral, low-viscous, shallow crustal zones and discern them from other contributions like internal mass anomalies and topographic features from tectonic or geomorphological origins. Finally, it should be emphasized that the modelling results of figs. 2a and 2b are only meant to give an indication about what might possibly be deduced from GOCE data concerning (crustal) lateral variations and shallow low viscosity zones. More detailed earth models, based on detailed structural, compositional and rheological (also non-linear) data, seismic tomography, electrical conductivity studies, etc., are necessary before realistic comparisons can be made with data from GOCE. For further details we refer to Schotman et alii (2008).

**CONCLUSIONS**

GIA model simulations indicate that information on shallow low viscosity zones might be deduced from GOCE geoid solutions, although uncertainties in both ice load history and earth structure could hamper unique interpretations to some extent. Combining spectral information with spatial patterns could reduce these uncertainties, whereby the range of possible ice and earth models is already constrained through other geodetic and geophysical data (e.g., GPS, ice load dynamics, tide gauge records). Lateral variations in earth structure, specifically with respect to occurrence of low-viscosity zones as a function of tectonic province, do have discernable effects on geoid anomalies, although they appear to be constrained to those regions that do not have a low-viscosity zone (compared to the laterally homogeneous low-viscosity zone case) and to their immediate surroundings.
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Instabilities development in partially molten rocks

Jean-Louis Vigneresse (**), Jean-Pierre Burg (**) & Jean-François Moyen (***)

ABSTRACT

Partially molten rocks (PMR) are characterized by specific and contrasting rheological properties. At a macro scale, single or multiple melt inclusions are well known to be present in the rocks. These positive or negative buoyancy manifestations generate strong strain partitioning. At a meso scale, structures are consistently oriented in a migmatitic body with those of the surroundings, indicating that the migmatites were deformed as a whole. By contrast, ubiquitous strain partitioning and melt distribution are widely present in the same migmatitic body, reflecting highly heterogeneous strain and intrinsic rheological instabilities. A continuous transition from a liquid-like to a solid-like rheology, as many averaging processes implicitly assume, cannot explain this two-fold information. We develop a full analysis, considering the stress and strain rate, and the relative proportion of melt and solid phases. Temperature varies from $T_{solidus}$ to $T_{liquidus}$ in a PMR. We also assume that the transition to melting is not dual to crystallization. However, we prefer using the viscosity rather than the stress, since the former is better constrained from experiments. The viscosity of the matrix, which deforms according to a power law, shows shear thinning, whereas that of the melt remains constant. The viscosity contrast between the two phases thus varies with strain rate. The lower the strain rate, the higher is the viscosity contrast, hence instabilities development is controlled by the rheology. The path followed during a transition also controls the intermediate state, and may lead to instabilities, resulting from mechanical reasons or from the respective amount in each phase. In the last case, the concentration in one phase induces instabilities. A surface describing viscosity in a 3D diagram (strain rate-amount of phase-viscosity) is constructed, that presents a cusp shape for low strain rates. The diagram depicts two types of behaviour and a critical state. At high strain, the viscosity contrast between melt and matrix is lowest. The rock behaves as a near-homogeneous body and a continuous description of its rheology may be estimated. Instabilities lead to fabric development resulting from crystals alignment. At low strain rate, three domains are separated by a critical state. When the proportion of one phase is very small, the material behaves as the other end-member. For intermediate proportions, the cusp indicates three possible viscosity values. Two are metastable, whereas the third is virtual. Hence, the viscosity of the mixture jumps back and forth from the viscosity of one phase to that of the other. A similar process occurs for temperature, since the cusp in the viscosity profile has also implications in a diagram linking temperature and stress. Different behaviours result, depending on whether the deformation takes place under a fixed content in each phase, a common stress, a common strain rate or common temperature. We list several implications for partially molten rocks that may explain fabric development, contact melting between crystals, strain localization, mineral banding, shear heating, welding, stick-slip-like melt extraction, magma fragmentation or formation of strong or fragile glass. A phase diagram that incorporates temperature, stress and concentration is constructed for PMR that bears much similitude with those issued for other soft materials.

KEY WORDS: rheology, two-phase material, migmatites.

RIASSUNTO

Sviluppo di instabilità in rocce parzialmente fuse.

Le rocce parzialmente fuse (PMR) sono caratterizzate da comportamenti specifici e contrastati. Ad esempio in un corpo di migmatiti le strutture delle fasi solide e piccola scala sono orientate coerentemente con quelle delle rocce circostanti e ciò indica che le migmatiti sono state deformate come un unico insieme. Al contrario, le ubiquitarie ripartizione della distorsione e distribuzione del fusibile, diffuse nelle stesse migmatiti, possono essere interpretate come risultato di una serie di cause meccaniche oppure di diversa quantità relativa delle fasi. Nell’ultimo caso, la concentrazione di una delle fasi induce l’instabilità. Viene qui costruita una superficie che descrive la viscosità in un diagramma tridimensionale (velocità di distorsione-quantità della fase-viscosità) e che presenta una forma a cuspide a basse velocità di distorsione. Due are metastable, whereas the third is virtual. Hence, the viscosity of the mixture jumps back and forth from the viscosity of one phase to that of the other. A similar process occurs for temperature, since the cusp in the viscosity profile has also implications in a diagram linking temperature and stress. Different behaviours result, depending on whether the deformation takes place under a fixed content in each phase, a common stress, a common strain rate or common temperature. We list several implications for partially molten rocks that may explain fabric development, contact melting between crystals, strain localization, mineral banding, shear heating, welding, stick-slip-like melt extraction, magma fragmentation or formation of strong or fragile glass. A phase diagram that incorporates temperature, stress and concentration is constructed for PMR that bears much similitude with those issued for other soft materials.

KEY WORDS: reologia, materiali bifasici, migmatiti.

(*) Nancy-Université, G2R, BP 23, F-54501 Vandoeuvre Cedex, France, jean-louis.vigneresse@2r.ushp-nancy.fr
(**) Geologisches Institut, ETH-Zentrum, Leonhardstrasse 19, 8093 Zürich, Switzerland, jpb@erdw.ethz.ch
(***) Department of Geology, Stellenbosch University, Private Bag X1, Stellenbosch 7802, South Africa, jfmoyen@gmail.com
Most materials that constitute our direct environment are composed of several phases that all behave differently when submitted to stress. Rheology and continuum mechanics are usually the field of investigation for such behaviour. However, the basics hypotheses assume that the material presents continuous, or not too contrasted, properties between phases. Thus can be the case in solid rocks, where the minerals react similarly to a bulk stress. It is no more the case when one phase is solid, or highly viscous, and when the other phase is a liquid or a gas. For instance, sand usually flows under the wind, resulting in booming dunes whereas those keep a bulk pile shape. Conversely, mud saturated with water flows and spreads. Lavas adopt a similar behaviour: When a high temperature, they flow over kilometres, without any real structural control, except those imposed by the surrounding topography. Conversely, internal structures develop that can be used to infer flow directions and internal stress pattern. Such situations are hard to describe with usual methods, and any kind of averaging from the laws governing the end-members usually fail to describe instabilities that soon develop. Those are specifically observed into partially molten rocks, here referred to as PMR.

The present paper started from field observations on migmatites and crystallizing magmas. Migmatites are rocks that were partially molten rocks before they crystallized in their actual state (Menhert, 1968; Ashworth, 1985). In contrast, fabrics in magma record the shear flow of the melt during emplacement (Paterson et alii, 1998). PMRs can accordingly be envisioned as two-phase materials. One phase is solid (the rock that melts or crystallizes in magma); it is hereafter referred to as matrix. Melt is the other phase, here, essentially referring to felsic melts, though general term of granitic melt should not be restricted to any specific composition.

We develop a description of the PMR rheology that takes into consideration:

1) The amount of the solid phase ($\Phi$), ranging from 0 to 1. It is similar to a volume.
2) The intrinsic viscosity $h$ of each phase, intimately linked with the strain rate ($\gamma$).
3) The applied stress ($\sigma$).
4) The temperature ($T$) interval between solidus and liquidus.

The choice of the viscosity is for convenience, because it is better constrained by experiments than stress or strain rate. In consequence, after selection of stress as the intrinsic variable, a full description could be represented into a 3D diagram with coordinates stress, temperature and volume.

The constraints taken into account relate to field observations. They consist in:

1) The changing viscosity contrast with strain rate.
2) The non-linear aspect of melting rate.
3) The different evolution of viscosity with temperature for melt and matrix
4) The difference between melting and crystallization.
5) The bulk motion «en bloc» at the scale of a magmatic body and the small-scale heterogeneous motion with instabilities.

The present paper combines information about parameters identified in previous studies with important review papers about silicate melts (Mysen & Richet, 2005), granite rheology (Petford, 2003), pastes (Coussot, 2007), polymers (De Gennes, 1979), foams (Kraynik, 1979), dense suspensions (Stickel & Powell, 2005), analogue deformation (Rosenberg, 2001), friction (Persson, 2000), granular flow (Jaeger et alii, 1996) and wet granular flow (Mitari & Nori, 2006). Previously, we focused on identifying:

1) The evidence of two thresholds during melting and crystallization (Vigneresse et alii, 1996).
2) The non-duality between melting and crystallization (Vigneresse et alii, 1996).
3) The importance of strain partitioning between phases (Vigneresse & Tikoff, 1999).
4) The non-linear behaviour of the melting rate and melt distribution (Burg & Vigneresse, 2002).
5) The rheological contrast between melt and matrix (Burg & Vigneresse, 2002).
6) The presentation and solution of a double system of equations for melt extraction (Rabinowicz & Vigneresse, 2005).
7) The necessity of including pure and simple shear for melt extraction (Rabinowicz & Vigneresse, 2004; Vigneresse & Burg, 2005).
9) The cusped shape of the viscosity as a function of strain rate (Vigneresse & Burg, 2004).
10) The discontinuities the cusp shape induces on a stress-strain rate diagram (Vigneresse et alii, 2007).
11) The role of nonlinear melting in the melt production, i.e. on the phase proportion (Vigneresse et alii, 2007).
12) The importance of mapping those parameters for identifying instabilities development (Vigneresse et alii, 2007).

RHEOLOGY OF THE TWO END-MEMBERS OF A PMR

Rheology commonly describes the relation between shear stress ($\sigma$) and shear strain ($\gamma$), whereas time dependent effects imply a strain rate ($\dot{\gamma}$) response to stress. We use a shear strain rather than a plane strain ($\varepsilon$) since most magmatic flows develop under shear.

Melt and its matrix are the two end-members of the system. The melt behaves as a Newtonian body for moderate to low strain rates. A constant viscosity relates linearly strain rate to stress. Within the temperature range of melting (650-900°C), calc-alkaline granitic melts present viscosity value around $10^6$ Pa.s (Clemens & Petford, 1999). It exponentially decreases with temperature, in function of the activation energy $E$, with a typical value about 300 kJ/mole (Malloé, 1985). Around 800°C, viscosity decreases by 2.5-3.0 orders of magnitude for an increase of 100°C.

In contrast, crustal rocks brought at the same temperature range (650-900°C) deform in a ductile manner. We adopt the case of dislocation creep of a single crystal, with a power law exponent of 3 (Nicolas & Poirier, 1976). Experimentally obtained values for amphibolites, with values log A = -4.9 and Q = 243 kJ/mole (Kirby & Kronenberg, 1987), are used as a proxy for the restitic matrix of PMR, yielding a melt of granitic composition. The effective viscosity is estimated from the local tangent to the stress-strain rate curve.
Under those assumptions, the preceding numerical values provides the equations for the melt

$$\log \eta = 6$$

and for the matrix

$$\log \eta = 10.66 - 2/3 \log \gamma^o$$

The rheology of mixed melt and matrix (PMR) cannot be simply defined as the combination of those two end-members, depending on their relative proportion (fig. 1). During crystallisation, the solid particles interact which each other, leading to the Einstein-Roscoe law (EINSTEIN, 1906; ROSCOE, 1952; ARZI, 1978):

$$\eta = \eta_0 (1 - \Phi/\Phi_{max})^{-n_e}$$

in which \(\eta_0\) is the initial melt viscosity, \(\Phi_{max}\) is the maximum packing assemblage, and \(n_e\) an experimentally determined coefficient (LEJEUNE & RICHET, 1995). It has been experimentally validated up to 0.40 of solid phase, less than maximum packing, about 0.75 (ROGERS et alii, 1994). Particle interactions become important at higher concentrations, changing the exponent into \(-n_e\Phi_{max}\). This reduces the exponent value from 2.5 to about 1.8 (KRIEGER & DOUGHERTY, 1959). However, the viscosity increases by 4 to 5 orders of magnitude near maximum packing. Indeed, the mixture becomes thixotropic (BARNES, 1997) with departures from non-linearity in case of crystallization and pseudo-plastic in case of melting. Nevertheless, the viscosity contrast between melt and matrix ranges from 10 to 14 orders of magnitude (BURG & VIGNERESSE, 2002).

**PMR SPECIFICITIES**

A PMR combines three possibilities to develop instabilities. One is mechanical or rheological, owing to the large viscosity contrast between melt and matrix. The second is driven by the respective amount of each phase. The third is chemical and relates to temperature, especially during the interval between melting and crystallization. A 3D diagram combining stress, temperature and the volume of one phase is suggested that would provide a complete mapping of the complex PMR rheology.

However, before constructing this diagram, one should take into account the specific points that characterize PMR rheology. Those are the existence of two thresholds during the transition between the end-members (VIGNERESSE et alii, 1996), strain partitioning (VIGNERESSE & TIKOFF, 1999) and feedback loops that develop due to nonlinear processes (BURG & VIGNERESSE, 2002). The link between the rheology of a strong matrix and that of a concentrated suspension, drawn from Einstein-Roscoe equation (RENNER et alii, 2000; ROSENBERG, 2001) is seriously questioned since it does not allow any instability to develop (BURG & VIGNERESSE, 2002).

The range of threshold values for melting and crystallization overlapped. Thus, a definite rheology cannot be ascertained in that domain, that sees overlapping of two behaviours, each being related to one end-member.

In addition, this domain, with two metastable states varies in size depending on the strain rate or stress acting on the system. Instabilities develop during melting or crystallization, when the slope of the flow curve relating the transition from one phase to the other has becomes negative (SPENLEY et alii, 1993). In case of a system under common stress, fluid decomposes into a layered structure, with alternate layers of high and low strain rate. Conversely, in case of deformation under common stress, shear localization develops (fig. 2).

The bulk rheology of a PMR should be examined in a 3D (\(\sigma - \gamma^o - \Phi\)) diagram. However, the pair \(\sigma - \gamma^o\) is poorly determined from experiments, that often develop under constant and fast strain rate. Hence, they are limited by the total duration of the experiments. We prefer adopting a 3D (\(\eta - \gamma^o - \Phi\)) diagram because the pair \(\eta - \gamma^o\) is experimentally constrained.

We start with the state equations for the melt and its matrix (Eqs. 1 and 2). Owing to large variations in viscosity, the strain rate response to stress plots in a log-log diagram. A line with constant slope represents the melt, whereas another line represents the matrix. In between, the Einstein-Roscoe curve is not strain rate dependent. The two surfaces constructed from the two end-members overlap over a wide range of \(\Phi\) (0.50 to 0.75). The connection between the two end-members takes the form of a cusp surface in the (\(\eta - \gamma^o - \Phi\)) diagram.

Temperature has a differential effect on the viscosity of melt and matrix, resulting from the activation energy values for those phases. They respectively plot as two lines with different slope on a semi-log diagram as a function of temperature. The viscosity for the transitional state must be computed for fixed values of strain rate from the 3D diagram (\(\eta - \gamma^o - \Phi\)).
Under high strain situation, the transition from the solid to the weak phase is monotonous, giving place to a smooth viscosity variation. In contrast, the low strain rate case has to take into account the cusp that develops in the \((\eta - \Phi)\) diagram. Cusp occurs in between 30 and 60\% of the solid phase. The transition in viscosity also adopts a cusp shape within this range of temperature. Instabilities may develop depending on the followed path, i.e. constant stress or constant temperature, identically to the instabilities with strain rate.

All parameters are now settled to build a phase diagram that would determine the limits of PMR rheology, in function of the phase amount, viscosity and temperature. The strain rate should be introduced to determine the respective occurrence of instabilities. The basic ingredients to construct a 3D diagram (\(\Phi - \eta - T\)) are the preceding diagrams (fig. 3). For a better readability, we use \(\phi = 1 - \Phi\), the amount of liquid phase, and because it is better constrained, we use the viscosity instead of stress.

The three axes (\(\phi - \eta - T\)) determine the range of occurrence of PMR. Whereas the amount of melt ranges from 0 to 1, the temperature ranges from \(T_{\text{solidus}}\) to \(T_{\text{liquidus}}\), and the viscosity, which is plotted in a log scale ranges from the viscosity of the melt at \(T_{\text{liquidus}}\) to the value for the matrix at \(T_{\text{solidus}}\). The resulting diagram adopts the shape of a quarter of quasi-cylindrical body, the concavity of which faces the origin. This shape results from the two limiting values in temperature and phase amount, whilst the concave pattern results from the melting curve. In case of cusp development, the quasi-cylindrical body also presents a cusped surface, toward the origin.

Such mapping is useful as much as it can prompt the design for new experiments through predicting the behaviour of a studied system. A first attempt has been to classify the instabilities in a two-phase material according to the shape of the flow curve. It corresponds to considering the concentration, spinodal decomposition, or strain rate, i.e. essentially adopting a mechanical point of view (OLMSTED & LU, 1999).

**GEOLOGICAL IMPLICATIONS**

**MAGMA EMPLACEMENT AND STRUCTURES**

During felsic body emplacement, the strain rate is commonly higher than \(10^{-12}\) s\(^{-1}\), implying a stress level over 10 MPa (HARRIS et alii, 2000; VIGNERESSE, 2005; HAWKESWORTH et alii, 2004). The viscosity contrast between melt and matrix is the lowest, thus relaxation times for both phases have similar amplitude. The bulk material responds as a single-phase body with a bulk viscosity. Two situations can be observed that relate the strain rate and the ability of PMR to flow. Migmatitic bodies present the same structural trends as surrounding rocks (NZENTI et alii, 1988) documenting «en masse» deformation of the PMR massif.

Decreasing the strain rate implies increasing the viscosity contrast between melt and matrix. In a PMR, the rotation of the first formed crystals results in a fabric (BOUCHEZ, 1997; ARBARET et alii, 2000). When crystals interactions develop, it can lead to particle segregation, controlled by the concentration, as it has been described as Bagnold segregation (BAGNOLD, 1954). Conversely, when the strain rate locally exceeds the ability of a PMR to deform viscously, then it breaks into fragments like during brittle deformation (PAPALE, 1999), as observed during volcanic eruptions. Experiments on brittle fragmentation of magmatic melts suggest strain rates ranging from 50 to 150 s\(^{-1}\) (BÜTTNER et alii, 2006).
Mineral banding is one way to accommodate velocity continuity between phases, as described in flowing liquid crystals (Bonn & alii, 1998). It manifests in PMR through schlieren and melt-rich segregation (Clarke & Clarke, 1998; Weinberg et alii, 2001; Clarke et alii, 2002). In this case, it manifests through crystal sorting by size or by composition. In obsidian, it also takes the form of alternating bands of different colour some tens of microns to decimeters in width (Swanson et alii, 1989; Smith, 2002). The occurrence of shear bands due to strain localisation in plastic material results from deformation concentration on planes. In PMR, the different viscosity between the two phases leads to strain partitioning (Vigneresse & Tikoff, 1999). The discrete distribution of localised shear zones with only a few cm in width profoundly differs from the usual observation that ductile rocks should present diffuse deformation. In crystallizing magma, strain localisation develops within a non-yet consolidated framework of touching crystals, leading to formation of dilatant proto-faults (Guineberteau et alii, 1989; Pons et alii, 1995; Smith, 2000).

**Melt segregation**

Melt segregation at incipient melting results when both pure and shear stress apply on a PMR (Rabinowicz & Vigneresse, 2004). A compaction length describes the resulting space and time discontinuities. Melt-rich bands form at low angles (within 20°C) when observed on analogue material (Rosenberg & Handy, 2000; Barraud et alii, 2004) and natural samples (Katz et alii, 2006). They occur both during partial melting (Marchildon & Brown, 2002) and crystallization (Gourlay & Dahle, 2007). Instabilities in time result in cyclic periods of segregation, driven by the amount of melt (Rabinowicz & Vigneresse, 2004; Vigneresse & Burg, 2005).

Grain boundaries wetting by incipient melt is due to progressive depinning of the melt along the boundary surface, bearing relation to stick-slip motion observed during friction. Sliding motion is discontinuous and depends on the differential velocity between the two surfaces in contact leading to stick-slip motion (Scholz, 1990; Thompson & Robbins, 1990). It results from a competition between nucleation and growth rate of the pinning zones on one hand and the sliding velocity on the other hand. Indeed, stick-slip vanishes as the velocity overcomes a critical value, just because pinning has no more chance to develop.

At the end of crystallisation, the high proportion of the solid phase drastically reduces the melt mobility, isolating small-scale closed systems. The strain rate variation within the solid phase is analogue to pressure dissolution, resulting in important stress gradient between touching crystals. The gradient relaxes by dissolving one crystal to the benefit of another one (Grinfeld, 1993), leading to crystal impingement (Means & Park, 1994; Park & Means, 1996) in analogue experiments or in natural examples described in a crystallizing gabbro (Nicolas & Ildefonse, 1996; Rosenberg, 2001). At a larger scale, similar observations have been realised in metamorphic aureoles induced by granitic intrusions (Marchildon & Brown, 2002).

Sintering and high-pressure aggregation of particles into a solid bloc is observed in tuff welding (Grunder & Russell, 2005). Competition between compaction and viscous flow results in sintering, adhesion of molten fragments and deformation of glassy clasts (Smith, 1960). Superplasticity is been widely observed as related either to microgran or microstructural behaviour. It is interpreted as a transition between creep at low stresses and plastic flow near the yield stress. Viscous heating may lead to tachylites or pseudo-tachylites formation (Spray, 1995).

Dilatancy is a volume expansion in response to an applied stress, also synonymous with shear thickening, induced by the increasing viscosity of the crystallising magma. Nevertheless some dilatant veins also show internal brecciation (Smith, 1996) indicating that still present melt overcame the brittle/ductile transition. Dilatant regions are a sink for the residual melt in a flowing magma has been widely recognised by a more abundant glassy material (Smith, 2000).

**MEMORY EFFECTS**

Most of the instabilities above described present hysteresis, i.e. memory effect. It means that the transition from one state to the other is not dual to the reverse transition in terms of energy balance. Hysteresis is commonly described for induced magnetization (Bertotti, 1998), but also for plastic deformation (Prandtl, 1928). Indeed, a plastic body retains some strain (Bridge, 1950) when stress returns to its initial state. It profoundly contrasts with elastic deformation during which the strained body returns to its initial state when the stress is no more applied.

Hysteresis is the manifestation of stored energy. The return to initial conditions requires additional forces. This is the case for plastic deformation, or magnetism through the magnetic coercive field. In the transition to melting, the additional energy takes the form of the latent heat. During crystallisation of viscous material, there is a continuous reduction in the mobility of elements, manifested by the viscosity increase. Energy is thus continuously released between the liquidus and the solidus, corresponding to the entropy step due to latent heat when considering the temperature. The correlation between latent heat and viscosity is linear (Garai, 2004) for materials that show a good Arrhenius behaviour. This would correspond to a well-defined heat capacity gap between the liquid and the solid state, that is, to contrasted values of entropy of structural configuration (Bottina, 1994). When this is not the case, as for instance in fragile glass material, the number of intermediate structural configurations is large allowing intermediate metastable states, hence departure to Arrhenian behaviour, and non-Arrhenian viscosity (Angell, 1995) and consequently larger hysteretic loop. Indeed, hysteretic flow curves have been observed for non-Newtonian flows (Bonn et alii, 1998).

The memory effect or hysteresis in PMR is observed during successive phases of heating and cooling silicate melts above their liquidus temperature (Yue, 2004). The repeated heating and cooling phase lead to a gradual transition from non-equilibrium to equilibrium states. An ordered structure is observed up to 70°C above Tliquidus. The conversion from multi-crystalline phases to a single phase indicates that the liquid remembers the structures previously formed (Yue, 2004). Indeed, glassmakers use
cycles of rapid heating and cooling to transform a fragile crystalline phase into a stronger one (Conradt, 2004).

In our suggested model, hysteresis should be understood as a dissipation mechanism unable to return to its initial state without the addition of extra energy. However, repeated cycles of straining could lead to unexpected large strain, especially when the material has not the time to completely relax and return toward a state near its initial conditions. This is obviously the case when seismic waves, which are successive cycles of compression and extension, interact with a two-phase material. Nonlinear effects develop that indicate no return to initial conditions before the material is strained again. It usually leads to soil liquefaction when seismic waves propagate through saturated sediments (Ishihara, 1993).

The preferential reusing of a vein by new magma is also a sign of hysteresis. Tubes offer a pathway for lava to flow over large distances (Petersen et alii, 1994; Calvari & Pinkerton, 1994).

Finally, the reusing of already formed plastic shear zones is also a consequence of the memory effect. Grain reduction in a shear band or weakened material due to a former heating are potential sites to localise strain for a future deformation cycle. In that sense shear heating (Scholz, 1980) could provide natural conditions for rapidly deforming magma-present material.

IMPLICATIONS FOR EXPERIMENTAL DEFORMATION

The present paper offers an explanation for the development of much instability observed in natural conditions. However, it should be better regarded as a short review on the conditions under which those instabilities are produced. It is a former guide for designing experimental or numerical studies in order to address such instabilities.

One problem with experiments performed on natural rocks of analogue materials is the duration of the experiment. It directly points to the effect of strain rate. Adopting the time scale for a one year experiment, a long time indeed when considering the stability of one experiment, implies a strain rate of at least $3.2 \times 10^{-8}$ s$^{-1}$. Each additional order of magnitude implies a factor of 10, that would results in a maximum strain rate of $10^{-10}$ s$^{-1}$ obtained during a single experimental life.

One possibility to overpass this difficulty would be to change the material for some analogue material, resulting in the application of more reasonable strain rate. Adopting a common value of $10^{-5}$ s$^{-1}$, which is in use in many experimental press systems, limits in turn the viscosity contrast between a two-phases material.

The idea of the paper started from a different point of view. Provided experiments are not able to address the development of instabilities in terms of viscosity, strain rate and stress, it should possible to design some specific experiment that would be designed to address only one type of instability, depending on the temperature, viscosity and relative percentage of each phase (fig. 3).

CONCLUSIONS

Partially molten rocks (PMR) are commonly described as inhomogeneous, with a locally variable and unpredictable amount of melt. They also show local heterogeneities in strain distribution, with a neat predominance of non-coaxial deformation and shear. In contrast, at a large scale, their internal structures are concordant with those of the surrounding. PMR are by evidence a place where instabilities develop. Examining the rheology of PMR, we suggest three types of instabilities, one related to mechanical reasons, as shear zone localization or stick-slip motion, one linked with the concentration of solid phase, as banding or dilatant zones and a third one linked to temperature occurs when melting rate overcomes the rate of melt extraction.

Our model of two-phase rheology presented through a 3D diagram ($\gamma^e - \Phi - \eta$) shows a transition between a low strain rate regime during which the transition from one phase to the other is continuous in terms of rheology. It corresponds to a bulk motion of magma as a solid body, as exemplified during magma crystallisation or when migmatitic bodies are tectonically deformed. In contrast, at low strain rate, a cusp develops within the surface that represents the effective viscosity. It is the place of successive jumps between the rheology of each phase. It is naturally also the place where instabilities develop, depending on whether they develop under common shear rate or common stress. Crystal impingement during crystallisation reflects the progressive jamming. Melt extraction is also unstable, leading to competition between discontinuous melt production and melt extraction. All those instabilities strongly depend on the path adopted to go from one rheology to the other, resulting in strain localisation or phase banding. At moderate strain rate, crystals orientate toward an equilibrium position, giving place to a fabric in the magma.

The construction of a phase diagram allows designing specific experiments for better understanding the onset of those instabilities. It is essentially controlled by the amount of phase, the available stress and the temperature.

REFERENCES


