GEOPHYSICAL RESEARCH LETTERS, VOL. 35, L07304, doi:10.1029/2007GL032860, 2008

Click Here for Full Article

Role of Korteweg stresses in geodynamics

Gabriele Morra[1,2] and David A. Yuen[3]

Received 21 December 2007; revised 26 February 2008; accepted 29 February 2008; published 3 April 2008.

[1] It has come to our attention that the constitutive relationship used in the modeling of geodynamical flow problems with strongly variable physical properties, should have additional terms in the stress tensor, known in the literature as Korteweg stresses (K-stresses). These stresses arising at diffuse interfaces, which can best be explained in terms of density gradients, have already been mentioned in the literature for more than one hundred years, but have not received attention recently until a combination of experimental and numerical evidence have confirmed their existence. We will discuss the important potential role these new terms have for geophysics. It has many ramifications in geodynamics, ranging from mantle convection to earthquakes and magma fragmentation. **Citation:** Morra, G., and D. A. Yuen (2008), Role of Korteweg stresses in geodynamics, *Geophys. Res. Lett.*, *35*, L07304, doi:10.1029/2007GL032860.

1. Introduction

[2] Stresses generated in geodynamics are strongly influenced by density gradients along interfaces, such as phase transitions and compositional gradients. The nature of the interface between two different miscible fluids has been the topic of intense study for more than 150 years in the fields of physics and chemistry (e.g. Gibbs, 1876). Most geoscientists have not scrutinized in detail the precise nature of the interface and the stresses produced there by the density gradients. *Korteweg* [1901] has proposed a rheological relationship for a capillary type of stresses, based on the density and its spatial gradients. They have to be known as Korteweg stresses (σ_K), or K-stresses, and are illustrated in Figure 1a. In brief, they represent additional terms in the constitutive relationship, which may be important in geodynamics. *Joseph* [1990, 1996], *Mungall* [1994], *Petitjeans and Maxworthy* [1996], *Chen and Meiburg* [1996, 2002], *Anderson et al.* [1998], *Brenner* [2005], *Pojman et al.* [2006], *Chen et al.* [2006], and others have called attention to the role played by K-stresses but, up to now, its significance has not been properly appreciated among earth scientists. In this letter we will draw attention to this kind of capillary stresses associated with diffuse interfaces and evaluate its potentially important role for geophysics, since it has great impact on many areas.

2. Experimental Indications

[3] Two seminal experiments for understanding the dynamics of light and low viscous diapirs were carried out more than twenty years ago by *Olson and Singer* [1985] (Figure 1b) and *Griffiths* [1986], who found similar results for rising cavity plumes. A particular detail, well documented in both works, is the unexplained departure from the Stokes terminal velocity of the plumes. *Olson and Singer* [1985] found systematically a lower velocity of a cavity plume for all parameter changes tested the role of the wall effects but they found it to be negligible and quantified a scaling difference from an expected $v_{exp} \propto t^{2/3}$, where v is the ascent rate, to and this was observed to be $v_{obs} \propto t^{2/5}$. *Griffiths* [1986] also observed the same lower rising velocity and assessed this to be 22%. He sought an explanation from the influences of the boundary conditions, although he also stated somewhat paradoxically that the rising velocity of the plume remains constant for a distance longer than two diameters of the plume head (r) and a large lateral size of the container (l) with r/l = 0.07.

[4] *Joseph* [1990] proposed a new interpretation of the experimental results by putting forward the role of K-stresses at the miscible interface zone between plume and surrounding fluid. Although he could not quantify the force involved, he displayed a long list of laboratory experiments, carried out between 1870 and 1930, in which the "membrane" character of the miscible interface was investigated and demonstrated experimentally. Most strikingly, he recovered an empirical expression for the stresses that one would expect to be generated at a smooth infinitesimally thin interface, originally proposed by *Korteweg* [1901], which depends on density and density gradient terms, formally analogous to an elastic membrane. However, although he showed beyond any doubt the similarity between miscible interface and interface membrane, he could not definitively demonstrate the precise role of K-stresses in the above experiments, because quantification of the K-stresses required the physical determination of unknown constants in front of the expression to be evaluated.

[5] Since that time several works have attempted to find evidence of the role of this force in analogue laboratory experiments, analyzed in concert with numerical models [e.g., *Petitjeanes and Maxworthy*, 1996; *Chen and Meiburg*, 1996]. These works confirmed unequivocally a departure from the usual Stokes prediction of a diapir tip velocity for capillary systems, where the surface tension is expected to have a major role. More recently [*Chen and Meiburg*, 2002] developed a numerical code that solved both the Stokes equation and the K-stresses and explicitly showed how the Korteweg terms in a capillary geometry would exert a significant effect on the speed of the tip of a rising diapir (Figure 1c). Finally, *Pojman et al.* [2006] found the first direct experimental evidence and explicit quantification of

[1]Department of Geological Sciences, University Roma Tre, Rome, Italy.
[2]Also at Geophysics Department, ETH Zurich, Zurich, Switzerland.
[3]Department of Geology and Geophysics and Minnesota Supercomputing Institute, University of Minnesota, Minneapolis, Minnesota, USA.

0094-8276/08/2007GL032860$05.00

an effective interfacial tension (EIT) between two miscible fluids.

[6] Independently, a set of experimental works with mixing fluids close to critical point in a Hele-Shaw domain [*Vailati and Giglio*, 1997; *Cicuta et al.*, 2000, 2001] intentionally aimed at understanding the arising of giant fluctuations in the long term during the mixing of two miscible fluids, have shown how the capillary stresses are expected to grow in the beginning of a free diffusion phase and disappear when the coherence of the interface is destroyed by nonequilibrium fluctuations. Here the K-stresses were measured indirectly through a relationship between the scaling length of the interface and the associated stress [*Cicuta et al.*, 2000; *Vailati and Giglio*, 1997]. Following this approach, we estimated bounds on the K-stresses in the Earth's mantle (section 4).

[7] These many experiments strongly encourage further investigations into the role played by the K-stresses on the evolution of a miscible interface. Nevertheless we want to point out that it is still very difficult to directly measure them in the laboratory experiments above mentioned, therefore other unknown causes associated to K-stresses might lead to the observed existent discrepancies [*Ribe et al.*, 2007].

3. Mathematical Formulation of the Korteweg Stresses

[8] When inertial forces can be neglected as in the mantle dynamics and magma chambers, the fluid dynamics is described by the equilibrium between external forces (buoyancy $\Delta\rho\mathbf{g}$ in geodynamics) and the internal response of the system described in general by the divergence of a tensor of the deviatoric stresses ($\nabla\cdot\tau$). Most stress tensor models depend from velocity gradient (strain rate) and/or displacement gradient (strain). The traditional Newtonian rheology, where a linear viscosity term μ expresses the ratio between shear deformation rate and stresses:

$$\sigma_{ij}^S = \mu\left(\partial_i u_j + \partial_j \mu_i\right) - 2/3\lambda\delta_{ij}\partial_i u_i. \tag{1}$$

While much effort has been expended in proposing a nonlinear viscosity based on strain-rate dependent formulations of the stress tensor, no attention at all has been paid toward formulating a theory that involves density and its gradients terms. Such a formulation is not new and has been already put forth by *Korteweg* [1901] and echoed by *Joseph* [1990] and *Brenner* [2005] in a mathematical form analogous to the elastic stresses at the boundary between two immiscible fluids:

$$\sigma_{ij}^K = \delta_{ij}\left(\alpha\nabla^2\rho + \beta\nabla\rho\nabla\rho\right) + \delta\left(\partial_i\rho\right)\left(\partial_j\rho\right) + \gamma\partial_i\left(\partial_j\rho\right) \tag{2}$$

where the constants α, β, δ, and γ need to be determined from theory or observation. The term between parentheses on the left is the bulk expression of the K-stresses, while the terms on the right express their shear component. Therefore the Stokes equation can be recast in a new format where now the stress term depends separately on strain rates and density gradients:

$$0 = \rho g_i + \partial_j\left(\sigma_{ij}^S + \sigma_{ij}^K\right) \tag{3}$$

Although their nature is fundamentally different, K-stresses can be represented analogously to the surface tension that appears at the boundary between two immiscible fluids. This analogy has been exploited by *Pojman et al.* [2006] and *Zoltowski et al.* [2007] for extracting K-stresses from laboratory experiments and more recently by *Chen et al.* [2006], who numerically explore the entire range of viscosity and density parameters by means of a non-dimensional formulation, showing the scale-free nature of Korteweg stresses.

4. Dynamical Consequences From Forces due to Density Gradients

[9] The diagram displayed in Figure 2 illustrates an overall view of the scales in which K-stresses are involved. Global geodynamics triggers K-stresses producing effects at all length scales, depending on the physics dictating each scale. We consider three major lengths (micro, meso and macro scales) in an attempt to offer a first estimate of their contribution.

[10] In the lower mantle compositional fluctuations are broader and less steep, strain-rates are low except for ascending plumes and descending slabs. At the macro-scale, the downwelling of cold slabs and the upwelling of hot plumes in the mantle environment naturally generate steep temperature and compositional gradients. Also phase changes, being possibly very sharp (O(10 km)), produce estimated steep density gradient due to density steps of 1–10%. Thus stresses associated with small density gradients may exert significant effects only over larger distances. Therefore, we believe that an effort should be engendered from the mineral physics community to assess the probable values of K-stresses under deep Earth conditions, especially in view of the post-perovskite phase transition [*Murakami et al.*, 2004] and the high-spin to low-spin transition of Fe in the mid-mantle [*Crowhurst et al.*, 2008].

[11] At an intermediate scale between grain size up to km, deep mantle rocks are expected to display layering, mixing, eventually melting. K-stresses appear at each density transition; therefore layering will be also characterized by an oscillatory presence of Korteweg stresses. The shear K-stresses are proportional to the curvature of the boundary, therefore they will influence the size of fingering and grain growth dynamics, as shown numerically and experimentally by *Chen et al.* [2005] and *Pojman et al.* [2006]. Non-dimensional quantities have been put forward and can be re-employed also for estimating the role of K-stresses in geodynamics (Bond number $Bo = \rho g l^2/\tau$ and Capillary number $Ca = \mu u/\tau$, where τ are K-stresses, instead of surface tension).

[12] At the large end of the spectrum shown in Figure 2, the effects of K-stresses are expected to behave similarly to the experiments of *Olson and Singer* [1985] and *Griffiths* [1986]. In this case a Rayleigh-Taylor instability trigger a rising mantle anomaly delimited by a diffuse interface that has an asymptotic thickness, which depends on the diffusion coefficient and the velocity of the process.

[13] We try here to assess the magnitude of the K-stresses in geodynamics, following the analytical work of *Vailati and Giglio* [1997] and the experimental results of *Cicuta et al.* [2000, 2001]. As noted above, there is a physical

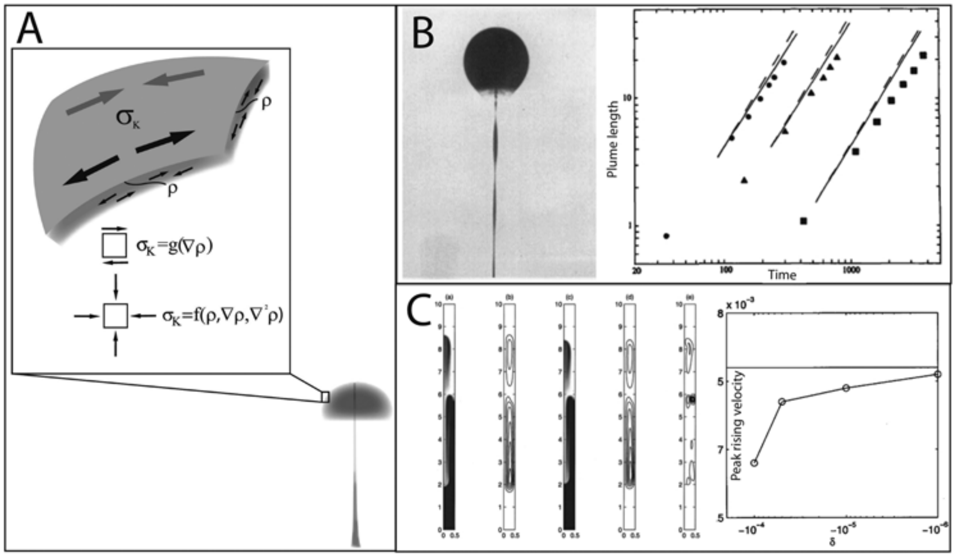

Figure 1. (a) When two miscible fluids are mixed dynamically, for example, through a rising plume (bottom right) a diffused boundary across the shell with a density (ρ) gradient forms, analogous to a membrane structure. Korteweg stresses (K-stresses) appear at the central part of the "membrane", where the gradient reaches its peak. The bulk component is a function of density, density gradient and the Laplacian of the density, while the shear contribution depends on the quadractic density gradient. (b) Snapshot of the rising of a miscible cavity diapir performed in a laboratory with two miscible fluids [from *Olson and Singer*, 1985]. The velocity of the diapir in function of a dimensionless time for three sets of experiments with three diapirs with different sets of discharge and three buoyancies, respectively (1.8×10^{-2}, 3.8×10^{-3}, 3.3×10^{-4}) mL/s and (291, 196, 80) cm/s^2. Dashed and continuous lines represent the expected Stokes rising velocity without and with wall correction, respectively. All points show a systematic consistent negative departure from the expected trends, following a different power law ($t^{5/3}$ instead of $t^{7/5}$). (c) Numerical results (pattern and streamlines) of the growth of a miscible diapir in a capillary tube [from *Chen and Meiburg*, 2002], performed solving Stokes law (examples a and b) and adding the K-stresses (examples c and d) with $\delta = 10^{-4}$. The last column (example e) shows the difference between the two models: the main difference is at the tip of the plume. The plot on the right shows that the sudden decrease of the peak velocity appears for values of delta close to $\delta = 10^{-4}$.

relationship between non-equilibrium density fluctuations and the K-stresses: when the thickness of the layer grows, interface excitations also grow, increasing the role of capillary forces, and it is only when the non-equilibrium fluctuations enter into play that K-stresses decay. This is not only clearly illustrated from a theoretical point of view but has been also demonstrated in laboratory experiments, as shown by *Cicuta et al.* [2000] and its implications in global geodynamics should be further studied.

[14] First, we calculate major length scales associated to the mantle density. When two fluids characterized by different densities are allowed to mix together, density anomalies go through a time interval in which non-equilibrium fluctuations appear and survive up to a length-scale

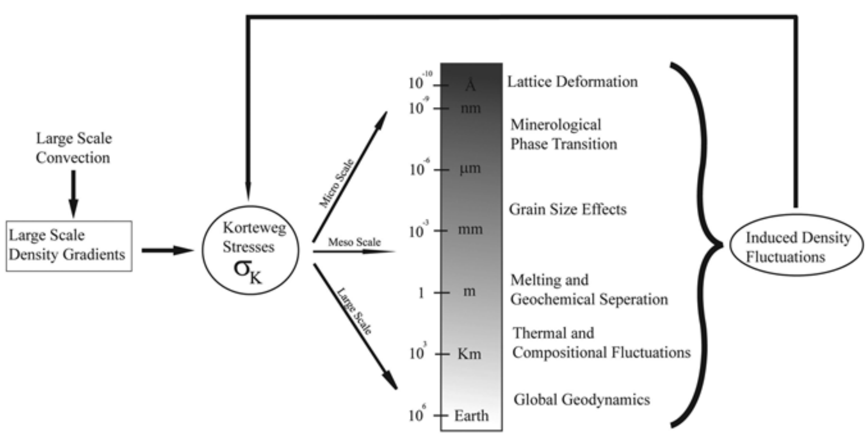

Figure 2. When large-scale geodynamics induces density gradients in some regions of the mantle, they provoke K-stresses that will modify the momentum equation at all scales. At each scale a different physical consequences will produce local density fluctuations. The associated K-stresses will contribute to the momentum equation and potentially trigger a positive feedback mechanism.

called the roll-off that can be determined from the wave vector q_{RO} [*Vailati and Giglio*, 1997]:

$$qRO = \left(\frac{g\partial_z\rho}{\mu D}\right)^{1/4} \tag{4}$$

where the associated wave-length is then $\lambda_{RO} = 2\pi/q_{RO}$. For lower mantle conditions for example, taking (g = 10 m/s^2, D = 10^{-6} m^2/s, $\mu = 10^{22}$ Pas), λ_{RO} oscillates between 200km for a sharp plume head boundary ($\Delta\rho$ = 100 Kg/m^3 in Δz = 10 km) and 2000 km for a very mild density fluctuation ($\Delta\rho$ = 1 Kg/m^3 in Δx = 1000 km).

[15] A quantity analogous to the roll-off is available for capillary stresses [*Cicuta et al.*, 2000] that links the density variation, gravity and surface tension with the capillary wave vector:

$$q_{cap} = \left(\frac{g\Delta\rho}{\tau}\right)^{1/2} \tag{5}$$

where τ is the integral of the longitudinal stress along the membrane. Although we don't have a handle on the surface stress expected for Earth conditions, we can invert (5) in order to calculate upper bounds, following the approach of *Cicuta et al.* [2000]. The diffusion between two miscible fluids is defined by a transition from an initial phase in which capillary stresses are dominating to a second stage in which the non-equilibrium fluctuations overwhelm the capillary effects. The transition from the first to the second stage occurs after a cross-over time [*Cicuta et al.*, 2001]

$$t_{co} = \frac{1}{Dq^2}\frac{1}{1+\left(\frac{qRO}{q}\right)^4} \tag{6}$$

which is extremely large under mantle conditions O(10^{15}s) until $q > q_{RO}$, i.e. $\lambda < \lambda_{RO}$. Beyond this threshold t_{co} decays extremely fast (fourth power of q). We can therefore use the relationships (4), (5) and (6) for estimating a lower bound for q_{cap}, being $q^{min}{}_{cap} = q_{RO} = 2\pi/(200$ km). An upper bound for q_{cap} can be set more simply considering the location of the steepest macroscopic gradient at the geodynamic scale, that we assume to be the boundary over the head of a plume Δz = 10 km, therefore $q^{max}{}_{cap} = 2\pi/(10$ km).

[16] Employing our estimate for q_{cap}, we can invert (5) obtain an estimate for the lower and upper bound of the forces at the boundary ($\tau = g\Delta\rho/q^2_{cap}$). Assuming $\Delta\rho$ = 100 Kg/m^3, one finds $\tau_{min} \approx 2{\cdot}10^9$ N/m and $\tau_{max} \approx 8{\cdot}10^{11}$ N/m. For a the peak case of the surface over the head of a plume with compositional thickness of 10 km, and assuming a uniform shear stress through its all thickness, we get $\sigma^{peak}{}_{min} \approx$ 200 KPa and $\sigma^{peak}{}_{max} \approx$ 80 MPa, while for the very smooth background compositional gradient related to the 200km large fluctuations, one obtains $\sigma^{bg}{}_{min} \approx$ 10 KPa and $\sigma^{bg}{}_{max} \approx$ 4 MPa. The order of magnitude represents an important contribution that the K-stresses are expected to apply in the mantle.

5. Discussions and Ramifications

[17] In this work we have shown from several lines of experimental evidence demonstrating the inherent inconsistencies between Stokes law predictions and laboratory observations. These would argue for the neglect of a fundamental type of force in geodynamics, called Korteweg or capillary, which may be important. Although it is today difficult to evaluate the exact magnitude of K-stresses under lower mantle conditions, we put forward a very first estimate of the stresses created by density gradients. For a more precise assessment, the constants associated with K-stresses may be evaluated from first-principles calculations, taking into account effects of grain-boundary which will allow for cross-scaling from micro to mesocales.

[18] Although they are formally similar, Korteweg and elastic stresses are physically distinct, therefore they contribute an additional stress to the Earth, which might play a role in creating the conditions for earthquakes release for example, inside a subducting plate, where temperature gradient are maximum, and at phase transitions, where they might represent an important contribution to the local stress, for example at the 660 km transition [*Morris*, 1992]. Because K-stresses will be proportional to the local curvature of the phase transition, their estimates in mantle convection models require accurate assessment of the phase transition near a mantle phase boundary [*Richter*, 1973]. Although the high-low spin transition is estimated to be very smooth and through several hundred kilometers, its global contribution integrated through the Earth sphericity should be assessed in order to rule out its contribution. All fluctuations of chemical, phase and thermal layering also induce alternating density gradients producing fluctuations in the stresses. Besides the giant fluctuations phenomena, the equivalent of a spinodal decomposition immiscible fluid, observed in the laboratory by *Vailati and Giglio* [1997], can make it possible for background density anomalies with a length-scale O (100 to 500 km) to exist in the mantle. They must be sought out by accurate analysis of seismic data or by seismic imaging.

[19] It is almost impossible to assess the influence of K-stresses without quantifying the smallest scale effects. Evaluation of the consequences of one scale to another will require accurate numerical modelling, using different techniques. They might also act and influence phenomena beyond geodynamics, such as core dynamics, where there is evidence for density stratification at the base of the outer core [*Souriau and Poupinet*, 1991].

[20] It has also been experimentally observed the presence of K-stresses in silicates melts [*Mungall*, 1994]. K-stresses might help to explain magma fracturing [*Papale*, 1999]. A recent review [*Zhang et al.*, 2007] has shown that the most important parameter for triggering magma fragmentation seems to be critical differential pressure between the bubbles immersed in the magma and the ambient pressure, which controls the dynamics of the breakup of bubbles with dire environmental consequences. Experimental data [*Spieler et al.*, 2004] can also fit with the interpretation that the critical tensile stress at the outer wall of the melt shell, combined to vesicularity, would explain magma fragmentation. We want to stress here that, because local fluctuations of density will naturally create local K-stresses changes, combined with the critical stresses in proximity to the bubbles, this might enhance conditions for dramatic plastic behavior [*Alidibirov and Dingwell*, 1996].

[21] **Acknowledgments.** We thank "Disco" Dan Joseph for invaluable discussions in 1990 and 2007. We are grateful for continuous encouragement from Klaus Regenauer-Lieb. We are grateful to A. Vailati and M. Giglio for sharing their knowledge, to Artem Oganov for interesting discussions, to Hans Jurgen Hermann for his insights on relating surface tension to a free energy formulation, and to Melanie Ross for technical assistance. This work was also supported by the ITR program of National Science Foundation. As part of the Eurohorcs/ESF - European Young Investigators Awards Scheme, it was also supported by funds from the National Research Council of Italy and other National Funding Agencies participating in the 3rd Memorandum of Understanding, as well as from the EC Sixth Framework Programme.

References

Alidibirov, M., and D. B. Dingwell (1996), Magma fragmentation by rapid decompression, *Nature*, *380*, 146–149.

Anderson, D. M., G. B. McFadden, and A. A. Wheeler (1998), Diffuse-interface methods in fluid mechanics, *Annu. Rev. Fluid Mech.*, *30*, 139–165.

Brenner, H. (2005), Navier-Stokes revisited, *Physica A*, *349*, 60–132.

Chen, C. Y., and E. Meiburg (1996), Miscible displacements in a capillary tube. Part 2. Numerical simulations, *J. Fluid Mech.*, *326*, 57–90.

Chen, C. Y., and E. Meiburg (2002), Miscible displacements in capillary tubes: Influence of Korteweg stresses and divergence effects, *Phys. Fluids*, *14*, 2052–2058.

Chen, C. Y., C. H. Chen, and J. A. Miranda (2005), Numerical study of miscible fingering in a time-dependent gap Hele-Shaw cell, *Phys. Rev. E*, *71*, 056304, doi:10.1103/PhysRevE.71.056304.

Chen, C. Y., C. H. Chen, and J. A. Miranda (2006), Numerical study of pattern formation in miscible rotating Hele-Shaw flows, *Phys. Rev. E*, *73*, 046306, doi: 10.1103/PhysRevE.73.046306.

Cicuta, P., A. Vailati, and M. Giglio (2000), Equilibrium and nonequilibrium fluctuations at the interface between two fluid phases, *Phys. Rev. E*, *62*, 4920–4926.

Cicuta, P., A. Vailati, and M. Giglio (2001), Capillary-to-bulk crossover of nonequilibrium fluctuations in the free diffusion of a near-critical binary liquid mixture, *Appl. Opt.*, *40*, 4140–4145.

Crowhurst, J. C., J. M. Brown, A. F. Goncharov, and S. D. Jacobsen (2008), Elasticity of (Mg, Fe) O through the spin transition of iron in the lower mantle, *Science*, *319*, 451–453.

Griffiths, R. W. (1986), Thermals in extremely viscous fluids, including the effects of temperature-dependent viscosity, *J. Fluid Mech.*, *166*, 115–138.

Joseph, D. D. (1990), Fluid dynamics of two miscible liquids with slow diffusion and Korteweg stresses, *Eur. J. Mech. B*, *9*, 565–596.

Joseph, D. D. (1996), Non-solenoidal velocity effects and Korteweg stresses in simple mixtures of incompressible liquids, *Physica D*, *97*, 104–125.

Korteweg, D. J. (1901), Sur la forme que prennent les equations du mouvements des fluides si l'on tient compte des forces capillaires causees par des variations de densite considerables mais continues et sur la theorie de la capillarite dans l'hypothese d'une variation continue de la densite, *Arch. Neerl. Sci. Exactes Nat., Ser. II*, *6*, 1–24.

Morris, S. (1992), Stress relief during solid-state transformations in minerals, *Proc. R. Soc. London, Ser. A*, *436*, 203–216.

Mungall, J. E. (1994), Interfacial tension in miscible two-fluid systems with linear viscoelastic rheology, *Phys. Rev. Lett.*, *73*, 288–291.

Murakami, M., K. Hirose, K. Kawamura, N. Sata, and Y. Ohishi (2004), Post-perovskite phase transition in MgSiO3, *Science*, *304*, 855–858, doi:10.1126/science.1095932.

Olson, P., and H. Singer (1985), Creeping plumes, *J. Fluid Mech.*, *158*, 511–531.

Papale, P. (1999), Strain-induced magma fragmentation in explosive eruptions, *Nature*, *397*, 425–428.

Petitjeans, P., and T. Maxworthy (1996), Miscible displacements in capillary tubes. Part 1. Experiments, *J. Fluid Mech.*, *326*, 37–56.

Pojman, J. A., C. Whitmore, M. L. TurcoLiveri, R. Lombardo, J. Marszalek, R. Parker, and B. Zoltowski (2006), Evidence for the existence of an effective interfacial tension between miscible fluids: Isobutyric acid-water and 1–butanol-water in a spinning-drop tensiometer, *Langmuir 2006*, *22*, 2569–2577.

Ribe, N., A. Davaille, U. Christensen, (2007), Fluid dynamics of mantle plumes, in *Mantle Plumes: A Multidisciplinary Approach*, chap. 1, pp. 1–48, Springer, New York.

Richter, F. M. (1973), Finite amplitude convection through a phase boundary, *Geophys. J. R. Astron. Soc.*, *35*, 265–276.

Souriau, A., and G. Poupinet (1991), A study of the outermost liquid core using differential travel times of the SKS, SKKS and S3KS phases, *Phys. Earth Planet. Inter.*, *68*, 183–199.

Spieler, O., B. Kennedy, U. Kueppers, D. B. Dingwell, B. Scheu, and J. Taddeucci (2004), The fragmentation threshold of pyroclastic rocks, *Earth Planet. Sci. Lett.*, *226*, 139–148.

Vailati, A., and M. Giglio (1997), Giant fluctuations in a free diffusion process, *Nature*, *390*, 262–265.

Zhang, Y., Z. Xu, M. Zhu, and H. Wang (2007), Silicate melt properties and volcanic eruptions, *Rev. Geophys.*, *45*, RG4004, doi:10.1029/2006RG000216.

Zoltowski, B., Y. Chekanov, J. Masere, J. A. Pojman, and V. Volpert (2007), Evidence for the existence of an effective interfacial tension between miscible fluids. 2. Dodecyl acrylate-poly (dodecyl acrylate) in a spinning drop tensiometer, *Langmuir 2007*, *23*, 5522–5531.

G. Morra, Department of Geological Sciences, University Roma Tre, I-00146 Roma, Italy. (gabriele.morra@erdw.ethz.ch)

D. A. Yuen, Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455–0219, USA.