Do faults trigger folding in the lithosphere?

Muriel Gerbault

University of Montpellier, France.

Eugenii B. Burov

B.R.G.M. Orleans, France.

Alexei N.B. Poliakov, Marc Daignières

University of Montpellier, France.

Abstract. A number of observations reveal large periodic undulations within the oceanic and continental lithospheres. The question if these observations are the result of large-scale compressive instabilities, i.e. buckling, remains open. In this study, we support the buckling hypothesis by direct numerical modeling. We compare our results with the data on three most proeminent cases of the oceanic and continental folding-like deformation (Indian Ocean, Western Gobi (Central Asia) and Central Australia). We demonstrate that under reasonable tectonic stresses, folds can develop from brittle faults cutting through the brittle parts of a lithosphere. The predicted wavelengths and finite growth rates are in agreement with observations. We also show that within a continental lithosphere with thermal age greater than 400 My, either a bi-harmonic mode (two superimposed wavelengths, crustal and mantle one) or a coupled mode (mono-layer deformation) of inelastic folding can develop, depending on the strength and thickness of the lower crust.

Introduction

Large structures of apparent periodicity greater than 50-100 km are observed in many intra-plate compression areas of the oceanic as well as of continental lithosphere [*Stephenson and Cloetingh*, 1991]. In this paper we focus our attention on three distinct areas - the Indian Ocean (A), the Western Gobi (B), and Central Australia (C) - which main features are briefly reviewed below.

(A) The gravity and topography data acquired in the North part of the Indian Ocean (Figure 1a) revealed sub-parallel undulations of the oceanic basement with spacings from 100 km to 300 km, and amplitudes up to 1-2km in the early Bengal Fan sediments [*Weissel et al.*, 1980]. Seismological data (reflection profiles and seismicity patterns) show numerous crustal faults, some of which are treated as pre-existing spreading-center normal faults reactivated as thrust faults [*Chamot-Rooke et al.*, 1993]. Part of these faults may be traced down to 40 km.

(B) Since 10-15 My ago, the western part of the late Paleozoic Gobi region (Figure 1b) experiences active tectonic compression induced by the India-Eurasia collision, and resulting in deformation of the basement, downwarp basins and crustal faulting [*Nikishin et al.*, 1993]. Spectral analysis of the topography and gravity anomaliesreveal sub-parallel structures spreading north-eastward across the Dzungarian basin (Figure 1b), with two dominating harmonics of wavelengths of 50-60 km and 300-360 km [*Burov et al.*, 1993].

(C) The gravity field over central Australia exhibits series of east-west trending anomalies (extending over 600 km) having typical wavelengths of about 200 km (Figure 1c)[*Stephenson and Lambeck*, 1985]. The geology of the region is characterized by late Proterozoic to Carboniferous sedimentary basins [*Goleby et al.*, 1989]. Teleseismic travel times infer 20 km variations of the Moho depth, over horizontal distances less than 50 km.

These periodic structures can be explained as the signature of a small-scale asthenospheric convection [*Fleitout and Yuen*, 1984]. But even though this mechanism may work for the oceanic lithosphere, it appears unefficient for the thick continental lithosphere. The scenario of lithospheric folding may also have a number of objections, two main of which are: (a) if a lithosphere buckles prior to its brittle failure, then tectonic forces needed for buckling are too high compared to rocks strength; (b) if faulting occurs prior to buckling, then forces drop too low to initiate folding. Other questions are how large-scale continuous folding can be accompanied by discontinuous localized faulting, where and at which moment the faults start to form (before, after or simultaneously with folding), can folding "survive" faulting.

Whereas the numerous existing models have improved our understanding of folding, they could not handle the above questions. The analytical studies consider only infinitesimal strains and do not consider explicitely the faulting process. In general these models predict dominating wavelength-to-thickness ratio (λ/h) between 4 and 6 [*Biot*, 1961; *Ricard and Froidevaux*, 1986; *Zuber*, 1987; *Martinod and Davy*, 1992; *Burov et al.*, 1993]. Analogue experiments handle large deformation, but they use either oversimplified rheologies (pressure-temperature independent), or a not-to-scale ratio of yield stresses to elastic modulus [*Martinod and Davy*, 1994]. Numerical models were proposed by *Beekman et al.* [1994] and *Wallace and Melosh* [1994], where faults were *explicitely* introduced into the model, in order to trigger folding. However, this approach is inadequate to our problem because the location and geometry of faults are pre-defined in their models.

In this study, we want to demonstrate that folding and faulting may happen *simultaneously*: folding may be acco-

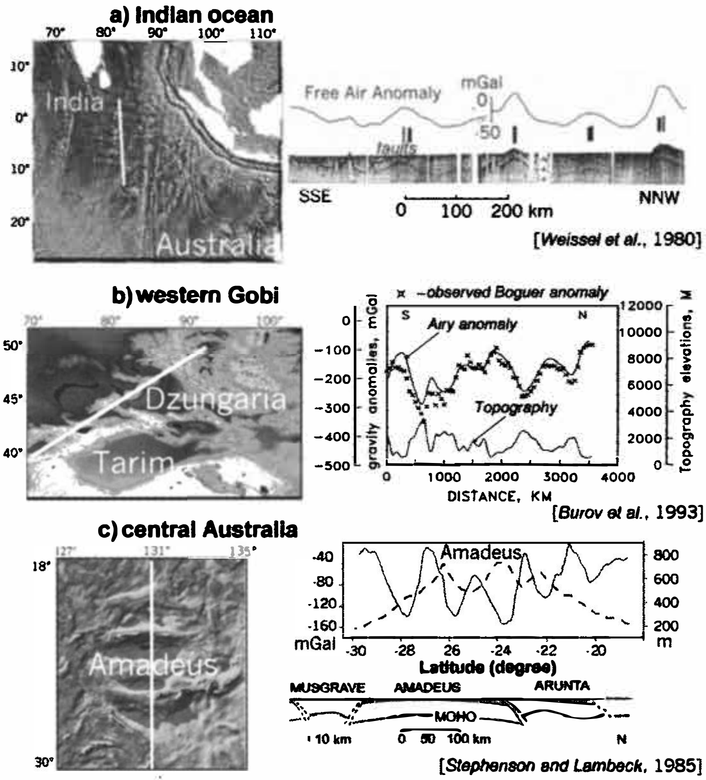

Figure 1. a) Map of free-air gravity anomalies in Central Indian Ocean with seismic travel time profile and the corresponding gravity anomaly. b) Map of topography in Central Asia. Along the white line, observed Bouguer gravity (crossed circles), topography and theoretical anomaly for Airy isostasy (solid lines) are shown. c) Map of Bouguer anomalies in Central Australia, with a north-south profile : topography (in dashed lines) and Bouguer anomalies (in solid lines). Schematic crustal structure across the basins.

modated by brittle faults and it is misleading to separate these phenomenas. We consider folding as a *mode* of deformation, and faulting as a *mechanism* of brittle deformation. To support this point of view, we perform direct numerical simulations and verify if calculated compressional forces, wavelengths, and timing of folding are comparable to nature.

We also numerically verify the idea of bi-harmonic folding predicted by *Burov et al* [1993] for intermediate-aged continental lithosphere, and check the ratio λ/h of the folding layers predicted by previous models [*Martinod and Davy*, 1992; *Ricard and Froidevaux*, 1986].

Model

We use a numerical code PAROVOZ derived from the FLAC method [*Poliakov et al.*, 1993]. This method allows for realistic visco-elasto-plastic rheologies. Another important advantage is that faults are not pre-defined but form themselves during loading, in a self-consistent way. The lithosphere is stratified as layers of either quartz, diabase, or olivine dominant rheology (Table 1). Elastic and brittle deformations are approximated by the Mohr-Coulomb non-associated elasto-plasticity, while the intracrystalline plasticity is approximated by a non-newtonian fluid, with parameters similar to *Buck and Poliakov* [1998]. The temperature field is calculated according to the heat conduction equation similar to *Burov and Diament* [1995], with parameters therein.

The dimensions of the problem, rheological boundaries (Moho), thermal age and applied strain rate (from which velocities are calculated) are given in Table 2. Hydrostatic boundary conditions are used at the bottom of the lithosphere, the upper surface is stress free and horizontal convergent velocities are applied at the lateral boundaries.

For the modelling of the oceanic lithosphere (case A), we choose a thermal age of 60 My yielding one 40 km thick competent layer. In the cases of continental folding (B,C), there are two (crustal and mantle) competent layers, which effective thicknesses are mainly temperature-controlled (250^oC isotherm roughly limits the competent upper crust whereas 750^oC limits the upper mantle [*Burov et al.*, 1993]). If the lower crust is weak and thick (case B), it can act as a ductile channel *decoupling* the upper crust and the upper mantle. Then, a biharmonic folding with two characteristic wavelengths may develop (about 4-6 times the thickness of the crust and mantle respectively), as predicted *Burov et al.*[1993]. When the crust is strong (case C), either because of its thermal age (> 700 My) or because of its rheological composition (e.g. diabase dominant crust), the deformation of the crust and mantle may be *coupled*, resulting in a larger wavelength corresponding to 4-6 times the sum of the crustal and mantle thicknesses [*Martinod and Davy*, 1992].

Single layer folding (A) - Indian Ocean

Our numerical experiment has shown the following evolution of deformation (detailed illustration is to be published elsewhere). At the onset of loading, distributed faulting initiates at the surface (where the yield strength is minimum) and propagates downwards with time. After a shortening of

Table 1. Rheological parameters (source from *Ranalli and Murphy* [1986]) used in the creep law [*Chen and Morgan*, 1990]: $\mu = \frac{1}{4}(\frac{4}{3A})^{\frac{1}{n}}\dot{\varepsilon}^{\frac{1}{n}-1}exp^{\frac{H}{nRT}}$.

rock	density kg/m^3	n	Activ. energy $Jmol^{-1}$	A $Pa^{-n}s^{-1}$
quartz (q1)	2700	2.7	1.3410^5	1.2610^{-7}
quartz (q2)	2700	2.4	1.5610^5	6.810^{-6}
diabase (di)	2800	3.4	2.610^5	210^{-4}
olivine (ol)	3200	3.	5.210^5	710^4

approximately 3% (i.e. 3 My), when the whole competent layer is at the yield state, folding rapidly develops with a dominant wavelength of approximately 210 km. We observe distributed faulting in the hinges of folds, which are normal or reverse depending on the sense of curvature of the layer. However, at a late stage of deformation, faults cutting the whole layer take place in the inflection points (Figure 2a). These 'localized' faults accompany folding until at least 10% of the shortening. After 7 My of compression (Figure 2a) the shortening is 6 % (73 km). This coincides with the data for the Indian Ocean (around 7 My and 5% respectively) [*Cochran et al.*, 1987; *Chamot-Rooke et al.*, 1993].

Biharmonic folding (B) - Central Asia

The numerical model shows the following evolution: after approximately 4% of shortening, the upper crust reaches the yield state and undergoes folding (accomodated by brittle faults) with a wavelength of 60 km and amplitudes up to 300 m. After 10% of shortening, mantle folding is significant, and the long wavelength undulations ($\sim$ 350 km) are *superimposed* on the short wavelength undulations (Figure 2b). The maximum topography reaches 4000 m. Intense viscous deformation concentrates within the lower crust, while brittle deformation localizes at the trough and inflection points of the folds (Figure 3b). Despite the presence of faults, the wavelengths of folding suggest that the lithosphere behaves as a system of strong layers.

Coupled folding (C) - Central Australia

In this case the crust is stronger: a 6 km layer of quartz rheology is underlain by 29 km of diabase rheology (see Table 1 and 2 for other parameters). We do not detect any crustal mode of folding. When both upper mantle and upper crust are at the yield state, folding of the upper mantle *coupled* with the upper crust occurs, with a wavelength of around 400 km. After 24% of shortening (Figure 2c), brittle shear zones start to localize in the mantle in the inflection points.

Table 2. Input parameters and results for models A, B and C.

	A	B	C
dimensions [km]	$1200*60$	$1200*120$	$720*120$
rheology	ol	q1+ol	q2+di+ol
Moho [km]	-	45	35
strain-rate $[s^{-1}]$	3.10^{-16}	5.10^{-16}	1.510^{-15}
thermal age[My]	60	450	700
$\frac{\lambda}{h}$	$\frac{210}{40} \sim 5$	$\frac{\lambda_c}{h_c} = \frac{60}{15} \sim 5$	-
$\frac{\lambda_m}{h_c+h_l+h_m}$	-	$\frac{350}{100} = 3.5$	$\frac{400}{100} = 4$
$\frac{\lambda_m}{\sqrt{h_l^2+h_m^2}}$	-	$\frac{350}{57} = 6.1$	$\frac{400}{87} = 4.7$
Force [N/m]	3.10^{13}	5.410^{13}	10^{14}

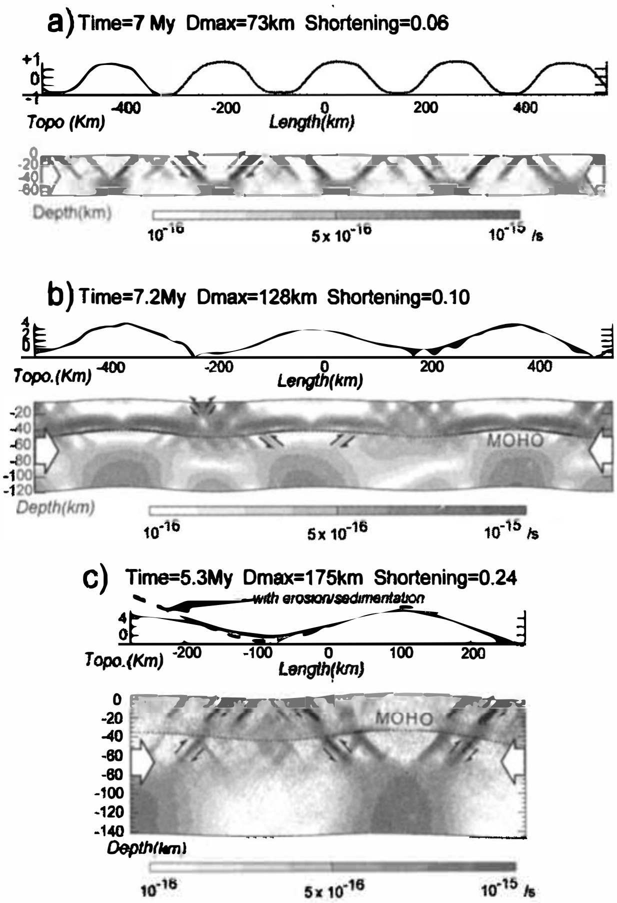

Figure 2. Numerical models of folding: oceanic lithosphere single layer mode (**a**), continental biharmonic mode (**b**), and continental coupled mode (**c**). Snap-shots of topography and strain-rate (maximum rate of deformation in the darkest areas) after several million years of compression. The applied velocities and hence the timing differ in each case.

Crustal faults accumulate at the trough of the folds, as a continuation of mantle faults to the surface (Figure 3c).

Discussion and Conclusions

We support the idea that the periodic structures observed in many compressional zones are the result of large-scale lithospheric folding, developping in a few stages: at the early stages, diffused faulting propagates downwards as compressive stresses build up. When the whole competent layer is at the yield state, folding starts to grow rapidly with a stable dominant wavelength. Thus faults do trigger folding, but only once they cut through the whole layer. Although pre-existing zones of weakness are certainly present in nature, it is not necessary to introduce them into the models to trigger development of folding.

At the advanced stages of deformation, small diffused faults remain in the hinges, because these are compatible with the bending strains. At the same time, large faults progressively stabilize at the inflection points of the folds, because these are kinematically compatible with the overall compression [*Gerbault et al.*, 1998]. These faults may explain pop-up structures observed in many places (i.e. in

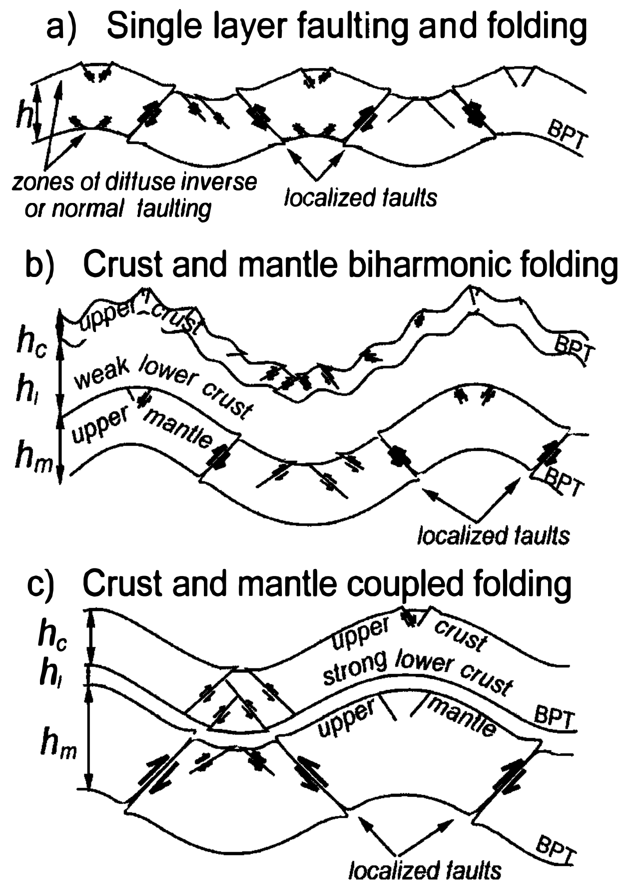

Figure 3. Summary cartoon illustrating how folds are accomodated by faults.

Tibet [*Burg et al.*, 1994]). This result suggests that faulted heterogeneous lithosphere maintains significant horizontal strength and can effectively behave as a strong layered media.

The obtained compressive forces (see Table 2) are consistent with the estimates of plate tectonics stresses *Cloetingh and Wortel*, 1986]. These stresses may be lower if erosion and sedimentation processes are taken into account (reducing the effect of gravity, unpublished results). The Mohr-Coulomb rheology we used for mantellic material also gives an upper value for the compressive forces.

The obtained values of *the ratio* λ/h (calculated by different methods see Table 2) range from 4 to 6, in agreement with the previous analytical models (we estimate the thickness of competent layers h by assuming that their base can sustain deviatoric stresses greater than 5% of the hydrostatic pressure at that depth).

References

Beekman, F., J.M. Bull, S. Cloetingh, R.A. Scrutton, Crustal fault reactivation facilitating lithospheric folding in the Central Indian Ocean, *Geol. Soc. Spec. Pub., 99*, 251-263, 1996.

Biot, M.A., Theory of folding of Stratified viscoelastic Media and its implications in tectonics and orogenesis, *Geol. Soc. Am. Bull., 72*, 1595-1620, 1961.

Buck, W.R., and A.N.B. Poliakov, Abyssal hills formed by stretching oceanic lithosphere, *Nature, 392*, 272-275, 1998.

Burg, J-P., P. Davy, J. Martinod, Shortening of analogue models of the continental lithosphere: new hypothesis for the formation of the tibetan plateau, *Tectonics, 13*, 475-483, 1994.

Burov, E.B., and M. Diament, Effective elastic thickness (Te) of continental lithosphere: what does it really mean? *J. Geophys. Res., 100*, 3905-3927, 1995.

Burov, E.B., L.I. Lobkovsky, S. Cloetingh, and A.M. Nikishin, Continental lithosphere folding in Central Asia (II): constraints from gravity and topography, *Tectonophysics, 226*,73-87, 1993.

Chamot-Rooke, N., F. Jestin, B. de Voogd, and Phèdre working group, Intraplate shortening in the Central Indian Ocean determined from a 2100-km-long north-south deep seismic reflection profile, *Geology, 21*, 1043-1046, 1993.

Chen, Y., and W.J. Morgan, A nonlinear rheology model for mid-ocean ridge axis topography. *J. Geophys. Res., 95*, 17583-17604, 1990.

Cloetingh, S., and R. Wortel, Stress in the Indo-Australian plate, *Tectonophysics, 132*, 49-67, 1986.

Cochran, J.R. et al., Intraplate deformation and Bengal Fan sedimentation, in *Proc. ODP Int. Rep., 116*, 3-11, 1988.

Fleitout, L., and D. Yuen, Secondary convection and the growth of the oceanic lithosphere, *Phys. Earth Planet. Int., 36*, 181-212, 1984.

Gerbault, M., A. Poliakov, M. Daignières, Prediction of faulting from the theories of elasticity and plasticity, what are the limits? *J. Struct. Geol., 20*, 301-320, 1998.

Goleby, B.R., R.D. Shaw, C. Wright, B.L.N. Kennet, K. Lambeck, Geophysical evidence for "thick-skinned" crustal deformation in Central Australia, *Nature, 337*, 325-330, 1989.

Martinod, J., and P. Davy, Periodic Instabilities during compression of the lithosphere 1. Deformation modes from an analytical perturbation method, *J. Geophys. Res., 92*, 1999-2014, 1992.

Martinod, J., and P. Davy, Periodic Instabilities during compression of the lithosphere 2. Analogue experiments, *J. Geophys. Res., 99*, 12057-12069, 1994.

Nikishin, A.M., S. Cloetingh, L.I. Lobkovsky, E.B. Burov, A.L. Lankreijer, Continental lithosphere folding in Central Asia (I): constraints from rheological observations, *Tectonophysics, 226*, 59-72, 1993.

Poliakov, A.N.B., P. Cundall, Y. Podladchikov, and V. Laykhovsky, An explicit inertial method for the simulation of viscoelastic flow: an evaluation of elastic effects on diapiric flow in two- and three-layers models, in *Flow and Creep in the Solar System: Observations, Modelling and Theory*, D.B. Stone and S.K. Runcorn (editors), Kluwer, Holland, 175-195, 1993.

Ranalli, G., and D.C. Murphy, Rheological stratification of the lithosphere, *Tectonophysics, 132*, 281-295, 1987.

Ricard, Y., and C. Froidevaux, Stretching instabilities and lithospheric boudinage, *J. Geophys. Res., 91*, 8314-8324, 1986.

Stephenson, R.A., and S. Cloetingh, Some examples and mechanical aspects of continental lithospheric buckling, *Tectonophysics, 188*, 27-37, 1991.

Stephenson, R.A., and K. Lambeck, Isostatic Response of the lithosphere with in-plane stress: application to Central Australia, *J. Geophys. Res., 90*, 8581-8588, 1985.

Wallace, M., and H. Melosh, Buckling of a pervasively faulted lithosphere, *Pageoph., 142*, 239-261, 1994.

Weissel, J., R.N. Anderson, and C. Geller, Deformation of the Indo-Australian plate, *Nature, 287*, 284-291, 1980.

Zuber,M.T., Compression of oceanic lithosphere: analysis of intraplate deformation in the Central Indian basin, *J. Geophys. Res., 92*, 4817-4825, 1987.

M. Gerbault, A.N.B. Poliakov, M. Daignières, Lab. Géophysique et Tectonique (CNRS UMR 5573), U.S.T.L. 4 pl. Bataillon, 34090 Montpellier, France. (e-mail: gerbault@dstu.univ-montp2.fr)

E.B. Burov, Research Department, B.R.G.M. Orleans, France.

(Received August 6, 1998; revised October 5, 1998; accepted November 3, 1998.)