Do faults trigger folding in the lithosphere ?

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Abstract. Large-scale periodic undulations within the oceanic and continental lithospheres revealed by a number of observations, are often treated as compressive instabilities, i.e. lithospheric buckling or folding. These undulations are normally associated with intensive faulting, which raises questions on the role of faulting in the folding process, and even on the possibility of folding in highly faulted media. In this study, we demonstrate that folding can "survive" faulting and that both processes may develop concurrently, so that faulting may serve as a mechanism of folding in the brittle domain. We support this hypothesis by direct numerical modeling. The results are compared with the data on three most prominent and well-known cases of the oceanic and continental folding-like deformation ("Indian Ocean" type, "Western Gobi (central Asia)" type and "Central Australian" type). We demonstrate that under reasonable tectonic stresses, folds can develop from faults cutting through the brittle parts of the lithosphere. The predicted wavelengths and finite growth rates are in agreement with the observations. We also show that within a continental lithosphere with thermotectonic 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.

1. Introduction

Large structures with apparent periodicity from 50 to 600 km are observed in many areas of large-scale tectonic compression [*Stephenson and Cloetingh*, 1991]. In this paper, we focus our attention on three well-studied areas with distinctively different folding scenarios (suggested in the previous studies) - the Indian Ocean (A), the Western Gobi (B), and Central Australia (C). The main features of these areas are briefly described below.

(A) The gravity and topography data acquired in the North part of the Indian Ocean (Figure 1a) revealed sub-parallel basement undulations 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 detect numerous crustal faults, some of which are treated as pre-existing spreading-center normal faults reactivated as deep thrust faults traced down to 40 km depth [*Chamot-Rooke et al.*, 1993].

(B) Since 10-15 My ago, the western part of the late Paleozoic Gobi region (Figure 1b) experiences active tectonic compression induced by the northward India-Eurasia collision and associated with periodic basement deformation, downwarp basins and crustal faulting [*Burov et al.*, 1993]. Neotectonic vertical movements, topography and spectral analysis of gravity data reveal sub-parallel structures spreading northeastward across the basins adjacent to the Tien Shan ranges (Figure 1b), with two dominating harmonics of wavelengths of 50-60 km and 300-360 km [*Burov et al.*, 1993]. The direction of the most prominent undulations is slightly incompatible with that of the main tectonic compression, probably due to transpression and eastward reduction of the lithospheric strength .

(C) The gravity field over central Australia exhibits series of east-west trending anomalies (extending over 600 km) having typical wavelengths of more than 200 km (Figure 1c); the tectonic data also suggest earlier (400-500 My ago) basement undulations with two times larger wavelength [*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 sharp 20 km variations of the Moho depth over horizontal distances less than 50 km possibly associated with large-scale faults.

Earlier explanations treating such periodic structures as signatures of a small-scale asthenospheric convection [*Fleitout and Yuen*, 1984] failed to provide convincing arguments for the appearance of regular lineations in the areas of known large-scale compressive stresses. The scenario of lithospheric folding also has met a number of objections, two "traditional" of which are:

(a) if the lithosphere folds/buckles prior to its brittle failure, then forces needed for folding may be too high compared to reasonable estimates of the tectonic forces; (b) if faulting occurs prior to buckling, then forces may drop too low to initiate folding, and folding will never develop (or will stop if faulting occurs during folding).

The first question was partly resolved by introducing viscous yield envelope instead of a uniform strength elastic plate (see, e.g., [*Zuber and Parmentier,* 1996]). Though the use of a constant strain rate envelope for both plastic and ductile lithosphere can be criticized, the question (a) now turns to a rather secondary problem. Thus in our study we will focus on the question (b) which was never answered. Our general goal is to clarify the role of faulting in the folding process, e.g. 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), and how folding can "survive" faulting. With this aim, we performed direct numerical simulations and verified if calculated compressional forces, wavelengths, and timing of folding are comparable to nature.

The existing analytical studies are limited by consideration of only infinitesimal strains and do not handle explicitly the faulting process. These models predict dominant wavelength-to-thickness ratios (λ/h) between 4 and 6 (if the gravity term is included) [*Biot,* 1961; *Zuber*, 1987;*Martinod and Davy*, 1992; *Burov et al.*, 1993]. Analogue experiments handle large deformation, but 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 of folding with faulting were proposed by *Beekman et al.* [1994] and *Wallace and Melosh* [1994], but there the faults are *explicitely* introduced into the model in order to trigger folding. Thus this approach is inadequate to our problem.

In this study, we show that folding and faulting may develop *simultaneously*, and that it is quite misleading to separate these phenomena. We thus consider folding as a *mode* of deformation, and faulting as its *mechanism* in the brittle domain. We also numerically verify the idea of bi-harmonic folding suggested by *Burov et al* [1993] for the middle-aged continental lithosphere (200 - 500 Ma), and compare the predicted λ/h ratios with other the previous models [*Martinod and Davy*, 1992].

2. Model

We use a finite-element FLAC-like code PAROVOZ [*Poliakov et al.*, 1993] which allows for realistic visco-elasto-plastic rheologies. Pre-defined faults are also not needed, as they can form themselves in a self-consistent way during loading. Similar to *Buck and Poliakov* [1998], the elastic and brittle deformations are approximated by the Mohr-Coulomb non-associated elasto-plasticity with friction angle 30 and cohesion 20 MPa, while the intracrystalline plasticity (ductile creep) is approximated by a non-newtonian fluid (Table 1). The temperature field is calculated using heat conduction equations from *Burov and Diament* [1995], with parameters therein. The lithosphere is stratified onto different lithologies (upper, lower crustal and mantle layers of thickness h_c, h_l, h_m, respectively) with quartz, diabase, or dry olivine dominant rheologies (Table 1).

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

For the 60 Ma oceanic lithosphere (case A), rheological and thermal structure yields a single 40 km thick competent layer. In the continental cases (B,C), two (crustal and mantle) competent layers appear. Their effective thicknesses are mainly temperature-controlled [*Burov et al.*, 1993]). If the lower crust is weak and thick (case B), it forms a weak ductile channel *decoupling* the upper crust from the mantle and favoring bi-harmonic folding with two characteristic wavelengths (about 4-6 times the thickness of the crust and mantle respectively), as predicted by *Burov et al.*[1993]. When the crust is strong (case C), either due to older thermal age ($>$ 700 My) or due to a stronger lithology (e.g. diabase), the deformation of the crust and mantle may be *coupled*, resulting in a larger values of λ corresponding to 4-6 times the sum of the crustal and mantle thicknesses [*Martinod and Davy*, 1992].

2.1. Experiments on single layer folding (A) - "Indian Ocean" type.

The deformation (detailed illustration is to be published elsewhere) develops as following. At the onset of loading, distributed faulting initiates at the surface, where the yield strength is minimum, and gradually propagates downwards. When the whole competent layer is at the yield state (3% shortening, 3My), folding starts to develop very rapidly, with $\lambda \sim 220$ - 250 km, and concurrent with distributed faulting in the hinges of folds (faults may be normal or inverse depending on the sense of curvature). At later stages, inverse localized faults cutting the whole layer take place in the inflection points, and accompany folding until at least 10% of shortening. At 7 My, 6 % (73 km) shortening is reached (Figure 2a,3a), which well coincides with the data for the Indian Ocean (around 7 My and 5% respectively) [*Cochran et al.*, 1987; *Chamot-Rooke et al.,* 1993].

2.2. Experiments on biharmonic folding (B) - "central Asia" type.

In this case, after approximately 4% of shortening, the upper crust reaches the yield state and undergoes folding accomodated by brittle faults, with λ=60 km and amplitudes up to 300 m. After 10% of shortening, mantle folding with $\lambda \sim$ 350 km becomes significant, and is *superimposed* on the short wavelength undulations (Figure 2b). The maximum basement undulations reach 4000 m. Intense viscous shear strain concentrates in the lower crust, while the brittle deformation localizes at the troughs and inflection points of the folds (Figure 2b, 3b). Despite the large faults, the wavelengths of folding indicate that the lithosphere still behaves as a system of strong continuous layers.

2.3. Experiments on coupled folding (C) - "central Australia" type.

For this case, the crustal rheology is stronger: a 15 km quartz layer is underlain by 20 km of diabase layer (see Table 1 and 2 for parameters), and folding of the upper mantle becomes *coupled* with that of the upper crust, with larger $\lambda \sim$ 400 km. At 24% shortening (Figure 2c), mantle shear zones start to localize in the inflection points. Crustal faults concentrate at the troughs of the folds, as an upward continuation of the mantle faults (Figure 3c). During this experiment, we also verified that time-dependent sedimentary loading, which must be important in the area, could not significantly influence the wavelength of folding (Figure 2c, dashed line).

3. Discussion and Conclusions

The large-scale lithospheric folding develops in several stages starting from diffused faults propagating downwards when the compressive stresses build up. The faults may trigger folding only once they cut through the whole layer. Consequently, although the pre-existing zones of weakness present in nature, they may be not "dangerous" for folding, as well as are not needed to trigger its development.

At later stages of compression, small diffused normal or inverse faults may remain in the hinges as far as they stay compatible with the bending strains, whereas large inverse faults stabilize at the inflection points of the folds [*Gerbault et al.*, 1998]. This may explain pop-up and inverse structures frequently observed in Tibet and Central Asia.

Estimated horizontal forces (Table 2) are consistent with the upper estimates of the tectonic forces [*Cloetingh and Wortel*, 1986], but could be much lower if erosion and sedimentation processes (reducing the effect of gravity) , and a weaker brittle rheology were taken into account.

The estimated values of *the ratio* λ/h (Table 2) range between 4 and 6, in a surprising agreement with the previous continuous models. This suggests that even completely faulted heterogeneous lithosphere can maintain significant horizontal strength and behave as a strong layered media.

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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.

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.

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.

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.

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.

Zuber, M.T., and E.M. Parmentier, Finite amplitude folding of a continuously viscosity-stratified lithosphere, *J. Geophys. Res., 101*, 5489-5498, 1996.

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Figure 1. **a)** Map of FAA gravity anomalies in Central Indian Ocean with seismic travel time profile and corresponding gravity anomaly. **b)** Topography map of central Asia with a profile of observed Bouguer gravity (crossed circles), topography and theoretical Airy anomaly (solid lines) . **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.

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Figure 2. Numerical models of folding: oceanic lithosphere single layer mode **(a)**, continental biharmonic mode **(b)**, and continental coupled mode **(c)**. Snap-shots of basement topography and strain-rate at developed stages of compression (see Table 2 for parameters). The dashed line in Figure 3c corresponds to solution obtained for diffusional erosion/sedimentation at average 0.5 mm/y rate.

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

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

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

Table 2. Input parameters and results for models A, B and C (thickness of competent layers h corresponds to the zones where deviatoric stresses are greater than 5% of the hydrostatic pressure)

	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
$\frac{\lambda}{h}$	$\frac{210}{40} \sim 5$	$\frac{\lambda_c}{h_c} = \frac{60}{15} \sim 5$	-
$\frac{\lambda_m}{h_c+h_l+h_m}$	-	$\frac{350}{100} = 3.5$	$\frac{400}{100} = 4$
$\frac{\lambda_m}{\sqrt{h_c^2+h_m^2}}$	-	$\frac{350}{57} = 6.1$	$\frac{400}{87} = 4.7$
Force [N/m]	3.10^{13}	5.410^{13}	10^{14}

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