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Lithosphere folding: Primary response to compression? (from central Asia to Paris basin)

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Abstract. We examine the role of lithosphere folding in the large-scale evolution of the continental lithosphere. Analysis of the record of recent vertical motions and the geometry of basin deflection for a number of sites in Europe and worldwide suggests that lithospheric folding is a primary response of the lithosphere to recently induced compressional stress fields. Despite the widespread opinion, folding can persist during long periods of time independently of the presence of many inhomogeneities such as crustal faults and inherited weakness zones. The characteristic wavelengths of folding are determined by the presence of young lithosphere in large parts of Europe and central Asia and by the geometries of the sediment bodies acting as a load on the lithosphere in basins. The proximity of these sites to the areas of active tectonic compression suggests that the tectonically induced horizontal stresses are responsible for the large-scale warping of the lithosphere. Wavelengths and persistence of folding are controlled by many factors such as rheology, faulting, time after the end of the major tectonic compression, nonlinear effects, and initial geometry of the folded area. In particular, the persistence of periodical undulations in central Australia (700 Myr since onset of folding) or in the Paris basin (60 Myr) long after the end of the initial intensive tectonic compression requires a very strong rheology compatible with the effective elastic thickness values of about 100 km in the first case and 50-60 km in the second case.

1. Introduction

The lithosphere can undergo tectonic shortening in three principal ways: (1) volumetric shortening by distributed or localized thickening of the lithosphere due to compression, (2) folding, when shortening is accommodated by unstable, subperiodical, vertical upward and downward bending (escaping) of the lithosphere, and (3) underthrusting, subduction, when shortening is accommodated by localized, stable, downward escape of the lithosphere, i.e., by underthrusting of one block or plate along large thrust faults. The dynamics of lithosphere folding forms an important but hitherto largely underestimated component in models developed to study the evolution of the lithosphere in compressional regimes: until recently large-scale folding was accepted only for a few areas of tectonic compression such as the northeastern Indian Ocean or central Australia [*Turcotte and Schubert*, 1982; *Fleitout and Froidevaux*, 1982; *Lambeck*, 1983]. However, a number of recent data sets suggest that lithospheric folding may be a much more widespread mode of deformation than previously thought, though it may take less obvious forms than in the well-known "distinct" cases [*Ziegler et al.*, 1995; *Cloetingh and Burov*, 1996]. For this reason, a better understanding of processes controlling surface expressions of folding is required to address some of the recently raised questions (e.g., commonly inferred relationships between the folding wavelength and the age of the continental lithosphere do not explain some data from recently indicated areas of compressional instabilities [*Ziegler et al.*, 1995]).

It should be noted that at a small-scale, folding or buckling of layered mechanical structures is a typical response to horizontal shortening. For example, folding is largely observed in sedimentary layers, in outcrops exposing ductile shear bands, in clays, and in other superficial structures [*Johnson*, 1980; *Smith*, 1975]. Since the mechanism of folding is largely scale-independent, it is quite reasonable to suggest that folding may be reproduced on a lithospheric scale [e.g., *Biot*, 1961; *Stephenson and Lambeck*, 1985; *Stephenson and Cloetingh*, 1991]. Detailed small-scale studies of folding in the sedimentary cover, clays, soils, and other superficial materials [*Smith*, 1975, 1977, 1979] often consider sophisticated nonlinear rheologies in many aspects similar to those of the deep lithospheric materials. Thus it would be logical to assume that the transition from a small scale to a large scale will not change the characteristic response of the modeled system. The only really important difference, which might be expected, is associated with the effect of gravity, which is negligible for small systems and significant for large ones.

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The potential gravity energy scales as $\rho g h^2$ per unit length, where $h(x)$ is surface elevation, ρ is the density, and g is the acceleration due to gravity, which requires a nonlinear increase in tectonic force needed to maintain continuously growing topographic uplift. Thus the influence of the gravity force on folding is negligible for the systems with $h < 10\text{-}20$ m but becomes important starting from $h > 50\text{-}100$ m. For this reason, small-scale folds may be of very large amplitude, a few times larger than the layer thickness and comparable to or exceeding the wavelength of folding. At lithospheric scale, upward elevations do not exceed 5 km, which is smaller than the thickness of the folded layers and less than 10% of the characteristic lithospheric folding wavelengths (30-600 km). Most lithospheric-scale studies used more simplified models than the small-scale studies; for example, a simplified elastic rheology was typically used as a first approximation of the lithospheric material properties. This approximation may work for the flexural models but predicts unrealistically high stresses for buckling models. Partly for this reason, models of lithospheric folding were put aside for quite a long time, just until the late seventies, when a number of authors [*McAdoo and Sandwell*, 1985; *Zuber*, 1987; *Turcotte*, 1979; *Fleitout and Froidevaux*, 1982, 1983] showed that the application of viscous or more realistic yield-stress rheologies can partly resolve the "high-stress" problems met in the elastic folding models. However, solution of the "high-stress" problem just freed space for another one, the "low-stress" problem. The latter is associated with the current wide-spread opinion that thrust faults, either preexisting or accompanying lithospheric shortening, may weaken the compressed layer so much that it will not be able to support stresses necessary to initialize and maintain folding (the latter requires a large competence contrast between the folded layer and the embedding).

Starting with early papers on the occurrence of lithospheric folding in the intracratonic lithosphere of central Australia [*Lambeck*, 1983] or on the folding of the oceanic lithosphere in the northeastern Indian Ocean [*Geller et al.*, 1983; *Stein et al.*, 1989], continental folding was subsequently recognized in the lithosphere of central Australia [*Stephenson and Lambeck*, 1985; *Beekman et al.*, 1997], central Asia [*Nikishin et al.*, 1993; *Burov et al.*, 1993; *Burg et al.*, 1994; *Cobbold et al.*, 1993; *Burov and Molnar*, 1998], Arctic Canada [*Stephenson et al.*, 1990], Iberia [*Waltham et al.*, 1999; S. Cloetingh et al., Late Cenozoic lithosphere folds in Iberia?, submitted to *Tectonophysics*, 1999, hereinafter referred to as Cloetingh et al., submitted manuscript, 1999] and the Paris basin [*Lefort and Agarwal*, 1996; J.-P. Brun, personal communication, 1994]. A surprising aspect of the outcome of these studies was that the occurrence of folding was not restricted to lithosphere with intrinsic zones of weakness but also occurred in areas characterized by the presence of cold and presumably strong lithosphere (see Figure 1 and Table 1).

More recently, evidence has also been put forward to support the occurrence of a component of lithosphere folding in some of the major extensional basins of Europe, such as the North Sea basin [*van Wees and Cloetingh*, 1996; *van Balen et al.*, 1998], the Pannonian basin [*Horváth and Cloetingh*, 1996; *van Balen et al.*, 1996; *Fodor et al.*, 1999; *Bada et al.*, 1998], and the Gulf of Lion's margin of the western Mediterranean [*Kooi et al.*, 1992; *Chamot-Rooke et al.*, 1999]. These basins are located on lithosphere with low rigidity, and the observed wavelengths seem at first hand to be at odds with these rheological contexts (see Figures 1 and 2).

However, the discrepancy with the theoretical predictions and the observations of the compressional deformation in the former extensional basins is not surprising, since in most of the "problematic" cases the geometry and other assumptions of the linear theory are poorly satisfied. The linear theory of folding developed in the sixties to seventies by *Biot* [1961], *Ramberg* [1961], *Smith* [1975], *Fletcher* [1974], and *Johnson* [1980] predicts development of sinusoidal undulations of a horizontally shortened competent layer embedded in less

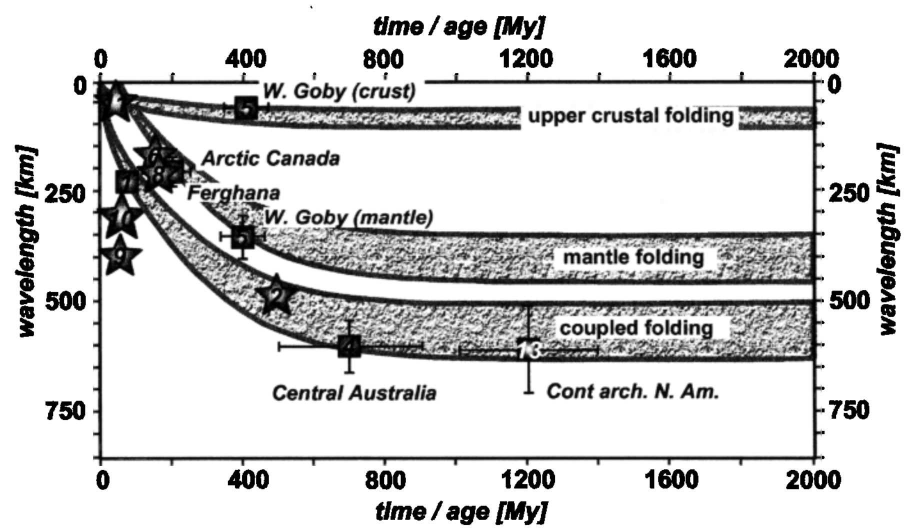

Figure 1. The observed wavelength of folding as a function of the thermal age (i.e., mechanical mantle thickness) calculated according to the model from *Burov et al.* [1993]. Numbers correspond to the ones used in Table 1. Squares show the cases of "regular" folding, whereas the stars mark "irregular" cases. Different theoretical curves correspond to crustal, mantle (supporting the presence of the decoupled rheology), and "welded" folding.

Table 1. Estimates for Wavelengths of Folding, Effective Elastic Thickness, Thermal Age, and the Onset of Folding

λ, km	EET, km	Thermal Age, Ma	t_0, Onset of Folding, Myr	Present State of Folding	Type
200-250 (1)	40	60	8	active deformation	B
500-600(2)	50	400-600	60	preserved	N
200 (3)	30	200	60	preserved	B
200 (present); > 400-500 (preserved) (4)	25 *	> 700	400-700	preserved	B
300-360 (5)	15	175 – 400	8-10	active deformation	B
100 –200 (6)	≥ 30	> 100	60	preserved	B/N
200 ? (7)	20-35	300	6	active subsidence	N
200–250 (8)	15	175	8-10	active deformation	B/N
350-400 (9)	6-9	< 20	4-6	active deformation	N
300 (10)	10-30	30	6-8	active deformation	B/N
40 (11)	20-25	< 20	6-8	active deformation	N
50 (12)	20-25	20	6-8	active subsidence	N
600 (13)	> 100	> 1200	1200	preserved ?	B
60 (14)	< 10	65	8-35	active deformation	B/N

EET, effective elastic thickness. Numbers in brackets refer to data sources: 1, Indian Ocean [*Cochran*, 1989; *Curray and Munasinghe*, 1989], (2) Russian platform [*Nikishin et al.*, 1997]; 3, Arctic Canada [*Stephenson et. al.*, 1990]; 4, central Australia [*Lambeck*, 1983; *Stephenson and Lambeck*, 1985; *Beekman et al.*, 1997]; 5, western Goby [*Nikishin et al.*, 1993; *Burov et al.*, 1993]; 6, Paris basin [*Lefort and Agarwal*, 1996]; 7, North Sea basin [*van Wees and Cloetingh*, 1996]; 8, Ferghana and Tadjik basins [*Burg et al.*, 1994; *Burov and Molnar*, 1998]; 9, Pannonian basin [*Horváth and Cloetingh*, 1996]; 10, Iberian continent [*Stapel et al.*, 1997; *Vegas et al.*, 1998; Cloetingh et al., submitted manuscript, 1999]; 11, southern Tyrrhenian Sea [*Mauffret et al.*, 1981]; 12, Gulf of Lion [*Kooi et al*, 1992]; 13, Transcontinental Arch of North America [*Ziegler et al.*, 1995]; 14, Norwegian sea [*Dore and Lundin*, 1996; *Vagnes et al.*, 1998]. "B" stands for regular folding style, "N" stands for "irregular" folding style, and "B/N" stands for the cases displaying both types of behavior.

*Value is for recent reheating at 200 Ma.

competent surroundings. This deformation is characterized by a single dominant wavelength between 3 and 50 km thickness of the competent layer. According to the linear theory, this dominant wavelength is time-independent and is mainly controlled by the thickness of the folded layer and much less by other parameters such as competence contrast or shortening rate. The linear analysis holds for many observed cases of folding but not for all. A number of very recent (generally small-scale) studies [*Davis*, 1994; *Zhang et al.*, 1996; *Mercier et al.*, 1997; *Bhalerao and Moon*, 1996; *Hunt et al.*, 1996; *Lan and Hudleston*, 1996] have shown that the wavelengths of folding may be significantly different from those inferred from the linear theory, if time-dependence and "non-sinusoidality" of deformation, realistic (elasto-plasto-ductile) rheologies, and layer geometries are taken into account. In particular, without any lateral inhomogeneities, depending only on the boundary conditions, time, rheology, and geometry, the wavelength of unstable plate deflections can significantly vary along the shortened plate, or even one single "megafold" can be formed [*Hunt et al.*, 1996; *Burg and Podladchikov*, 1999]. Regarding the frequently discussed influence of the preexisting faults on the development of folding, recent numerical experiments on folding of brittle-elasto-viscous lithosphere [*Gerbault et al.*, 1999] as well as previous analogue experiments on plasto-elastic lithosphere [*Shemenda*, 1989, 1992; *Burg et al.*, 1994; *Martinod and Davy*, 1994] have shown that (1) lithospheric folding and faulting can develop simultaneously and (2) preexisting crustal-scale faults do not prevent but participate in the development of folding instabilities. The instabilities, in turn, can provoke the formation of new faults at the inflection points of the folds. This disagreement with the usual expectations can be explained by the underestimated role of the gravity and friction in the behavior of the large-scale fault-and-fold systems. Yet most of the previous analogue studies were based on simplified pressure-, strain-, and temperature-independent rheologies, whereas the numerical studies employing nonlinear brittle-elasto-ductile rheology did not consider developed (large strain) stages of deformation. Consequently, a more thorough study is needed to verify the previous preliminary results and the hypothesis on folding and faulting interactions.

Based on the above mentioned new observational data sets and theoretical developments, the primary goal of our study is to demonstrate that lithospheric folding is much more widespread than is currently assumed. We propose that folding represents a rather typical initial and intermediate stage of tectonic compression, which may take quite different forms, which complicates its identification. For this reason, the cases of "irregular" folding were not recognized before. To explore this concept, we will discuss in the following the role of the nonlinear rheology, faulting, and other factors such as sediment loading and prefolding plate geometry in modifying the inferred wavelengths of lithosphere folds. In addition, we investigate further the role of coupled versus decoupled rheologies in the response of the lithosphere to large-scale compressional deformation. We will discriminate between the "typical," "regular," or "distinct" folding characterized by periodical deformation, which can be explained by the common linear theory, and "irregular" folding, which does not fit in the conventional theoretical scheme; for example, it may be aperiodical, polyharmonic, or have a much shorter or longer wavelength compared to the "linear" predictions.

2. Folding as a Mechanism of Shortening: Origin of Different Forms of Folding

Folding is commonly associated with periodical deformation of layered structures, and thus the periodicity is

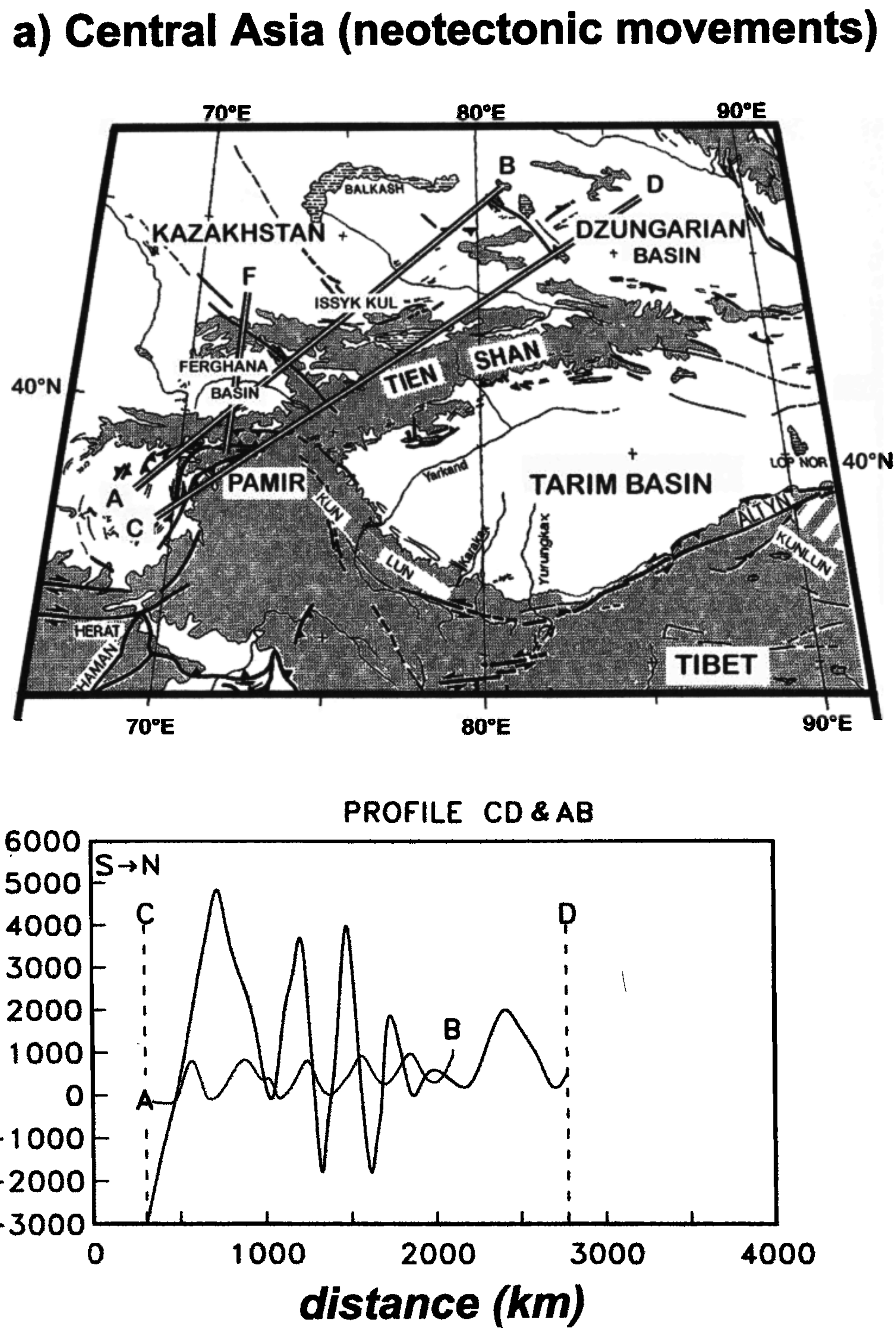

Figure 2. (a) Typical case of "linear" folding (profile of neotectonic undulations across the western Goby, central Asia [*Burov et al.*, 1993], (b) Typical case of "irregular" folding (Pannonian basin, simplified stratigraphic crosssections, recent vertical movements, and basement deflection profiles from *Horváth and Cloetingh* [1996], *Horváth* [1993], *Jaó* [1992], and *Bada et al.* [1998]), (c) Intermediate case of folding (gravity-based Moho deflection profile in the Ferghana basin, from *Burov and Molnar* [1998]).

often believed to be its primary, almost synonymous identifying feature. However, from a mechanical point of view, the periodicity is not a necessary requirement. For the mechanical theory, folding is just a compressional instability developing in stiff layers embedded in a weaker background, and the periodic, time-independent solution of the complete governing equilibrium and conservation equations is only the simplest one among the other possible solutions.

The physical mechanism of folding is well understood. In a continuum layered medium the stresses and strains must be

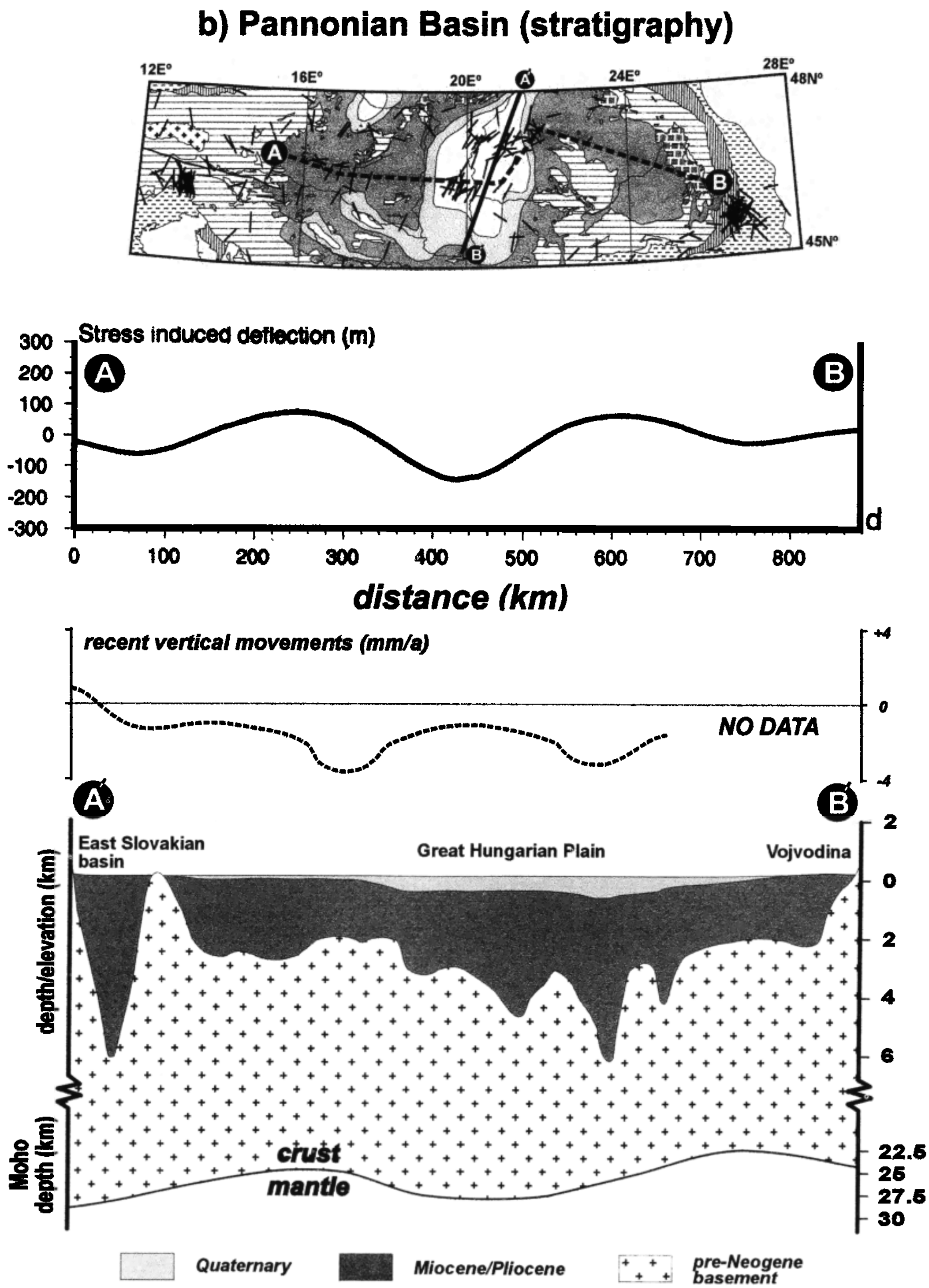

Figure 2. (continued)

continued across the interfaces between the layers. In compressed multilayers with contrasting mechanical properties (i.e., strong and weak layers), this requirement may be difficult to satisfy, because the same amount of shortening in a stiffer layer would require much larger stress than in the neighboring weaker layer. The system becomes unstable and, "trying" to reduce the stress or strain unconformities at the interfaces between the layers, starts to fold (buckle) in response to even negligibly low perturbations. In nonelastic media with strain-dependent properties, locally increased flexural strain at the fold limbs can create weakened zones significantly facilitating further deformation. These weak or softened plastic or viscous zones are often referred to as inelastic "hinges," since the system easily folds at such

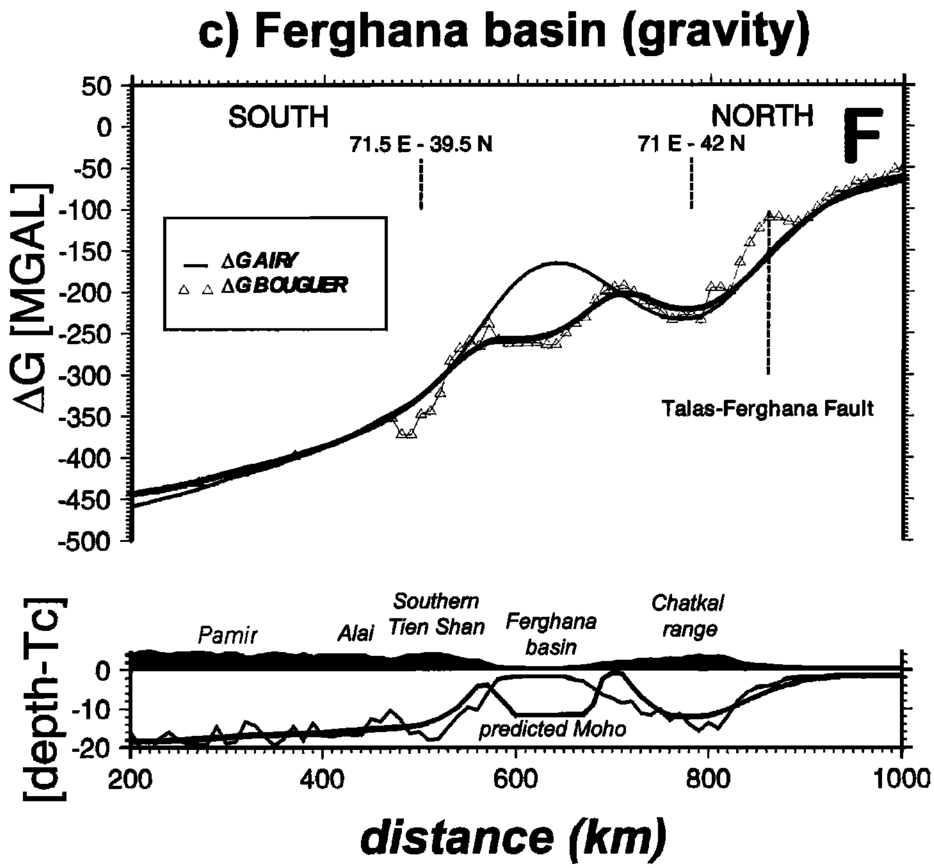

Figure 2. (continued)

weakened zones, similar to a carpenter's folding rule, under much lower compressional force than in the "normal" case. Homogeneous shortening of the lithosphere under horizontal compression requires a larger amount of work than that required by an equivalent amount of shortening by folding, whereas "shortening" by underthrusting/subduction of the lithosphere may be more advantageous then folding but cannot start immediately after the onset of shortening. For this reason, folding is likely to be a primary and even "standard" response to tectonic compression, which probably may continue, in an attenuated form, even after the beginning of subduction. As was noted in section 1, we subdivide the cases of lithosphere folding into two large, partly overlapping classes.

2.1. Linear, or Regular, Folding

"Biot's," or linear, folding encompasses the cases where compression of the lithosphere takes place from the beginning, leading to the formation of alternating basins and highs, and where the conditions of the linear theory are more or less satisfied (thin layer approximation, no strain-softening, plain layers, etc.). In the case of linear folding in a Newtonian media, an asymptotic relation derived from the thin layer equilibrium equation is quite simple [*Biot*, 1961; *Ramberg*, 1961]:

$$\lambda_l = 2\pi h\,(\mu_{l1}/6\mu_{l2})^{1/3}. \qquad (1)$$

Here λ_l is the "Laplacian" dominant wavelength of folding in the absence of gravity, h is the thickness of the competent layer (crustal or mantle), and (μ_{l1} and μ_{l2} are the effective viscosities (or competencies) of the strong layer and weak surrounds, respectively. Equation (1), derived assuming no gravity, gives estimates of λ_l/h = 20-40 for typical competence contrasts. This certainly does not hold for most lithospheric-scale cases, where λ/h = 4 - 6 is more common due to the participation of the gravity-dependent terms. In a simplest case of a single stiff layer embedded in inviscid medium, the dominant gravity-dependent harmonics can be accounted as [e.g., *Burov and Molnar*, 1998]

$$\lambda_g = 2\pi(2\,\dot{\varepsilon}\,\mu_{\mathrm{eff}}h^3/3n\,\Delta\rho g)^{1/4}, \qquad (2)$$

where $\Delta\rho$ is the density contrast, $\dot{\varepsilon}$ is the strain rate, μ_{eff} is the effective viscosity, n is the power law exponent (equating 1 in the Newtonian case), and h is the thickness of the competent layer. In a more general case, the term $2\dot{\varepsilon}\mu_{\mathrm{eff}}h^3/3n$ can be replaced with the term $G_{\mathrm{eff}}\,h^3/3$, where G_{eff} is the effective shear moduli, which can be used for any rheology. For example, in the elastic case, G_{eff} = $3D/h$, where $D = ET_e^3/12(1-\nu^2)$ is the flexural rigidity of the plate, E is Young's modulus, and ν is Poisson's ratio. For typical lithospheric h, $\Delta\rho$, $\dot{\varepsilon}$, and power law rheologies, the ratio λ_g/h is 3 - 6, which is compatible with many observations. For lithospheric parameters, λ_g grows faster than λ_l does, but

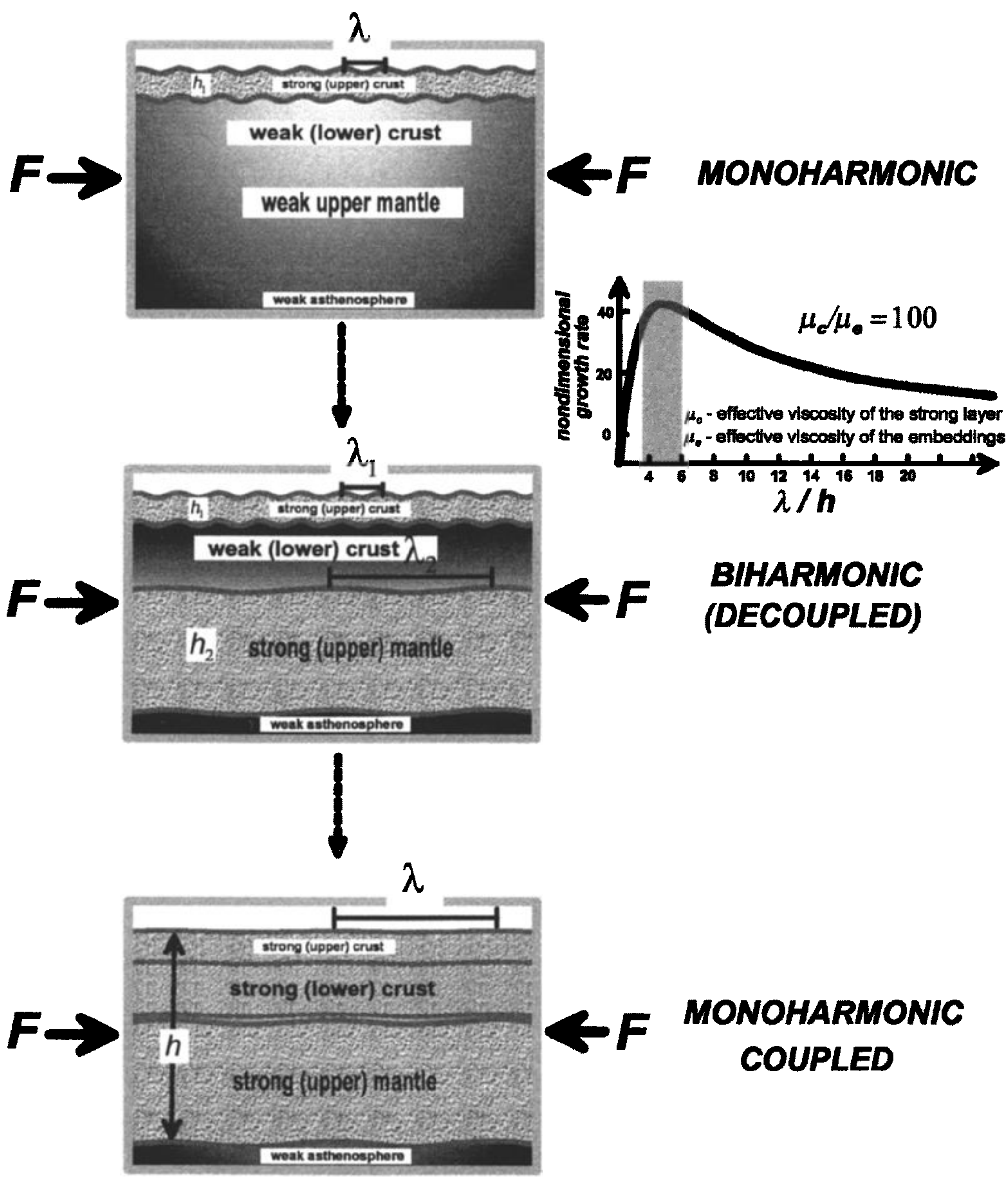

Figure 3a. Sketch of typical folding models (h_1 and h_2 are thicknesses of the competent crust and mantle, respectively). The system is submitted to compression by horizontal tectonic force F. In the case when the lower crust is weak, the upper crust may fold independently of the mantle part (wavelength λ_2), with a wavelength λ_1 (decoupled, or biharmonic folding). Young (< 150 Ma) and very old (> 1000 Ma) lithospheres (single competent layer or coupled crust and mantle) develop monoharmonic folding only. Inset shows the analytical estimate for the growth rate of strongly non-Newtonian folding (coupled layers, non-Newtonian rheology) as a function of λ/h for a typical ratio of the effective viscosities of the competent layer and embeddings (100). The ratio of the effective non-Newtonian power exponents is 100 [after *Burov et al.*, 1993]). Shaded rectangle shows the range of the dominating λ/h ratios (4-6).

naturally streams to infinity (flat surface) for small density contrasts. For more realistic cases, complex analytical relations based on the solution of full stress equations or at least taking account of gravity-load-dependent terms, multilayer structure and strongly non-Newtonian rheologies were proposed [*Smith*, 1975, 1977, 1979; *Martinod and Davy*, 1992, 1994; *Burov et al.*, 1993] (Figure 3a). However, even the most sophisticated analytical solutions are basically inadequate for large strain cases and faulted composite lithosphere. For these cases, pure numerical or laboratory experiments are indispensable [e.g., *Cobbold*, 1975, 1977; *Shemenda*, 1989; 1992; *Beekman*, 1994; *Burov and Molnar*, 1998; *Gerbault et al.*, 1999] (Figure 3b).

The "linear" cases generally involve strong lithosphere, the behavior of which is less affected by inhomogeneities, faulting, surface loads, and structural peculiarities. In these

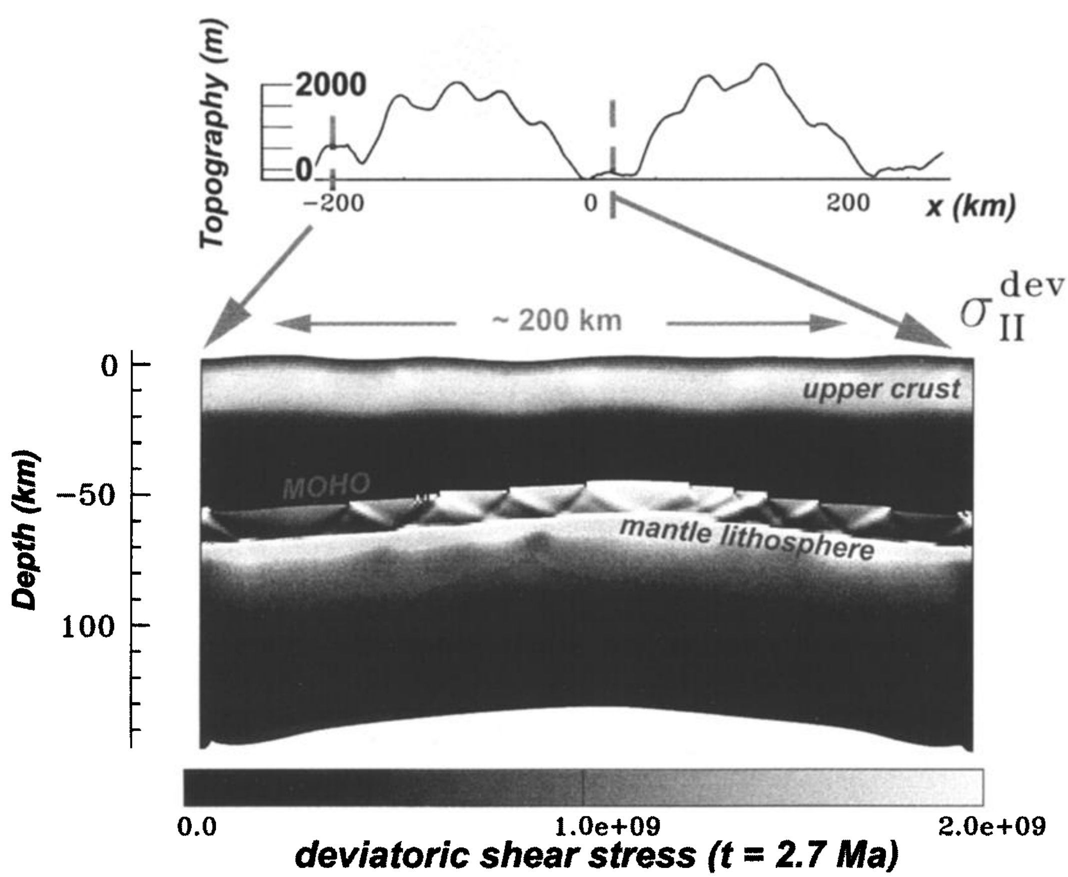

Figure 3b. Topography undulations and a "zoom" showing pattern of the second invariant of deviatoric shear stress (pascals) for the case of decoupled (biharmonic) folding of the middle-aged lithosphere (thermal age 250 Ma) with initial crustal thickness of 45 km, shortened at a rate of 2 cm/yr. Quartz (crust) -olivine (mantle) rheology is chosen (see Table 2 and the appendix for the parameters). Note the different wavelengths of crustal (40 - 50 km) and mantle (200 - 250 km) folding, as well as the intensive faulting in the upper mantle part. Crustal faults are as intensive as the mantle ones but not well imaged in Figure 3b, since the associated stresses are much lower than the mantle ones. See the strain rate patterns in Figure 4, which depict crustal faulting better than the stress patterns do.

cases, folding can be detected by matching the observed wavelengths of deformation and datings of the onset of folding (t_0) with the theoretical predictions (Indian Ocean, t_0=8 Myr, [*Cochran*, 1989; *Curray and Munasinghe*, 1989]; Russian platform, t_0=60 Myr, [*Nikishin et al.*, 1997]; Arctic Canada, t_0=60 Myr, [*Stephenson et al.*, 1990]; central Austaralia, t_0=400 Myr, [*Lambeck*, 1983; *Stephenson and Lambeck*, 1985; *Beekman et al.*, 1997]; western Goby, t_0=8-10 Myr, [*Nikishin et al.*, 1993; *Burov et al.*, 1993]; and Paris basin, t_0=60 Myr, [*Lefort and Agarwàl*, 1996]).

2.2. Irregular, or Nonlinear, Folding

The second class of folding refers to less distinguishable cases basically not covered by Biot's linear theory. Biot's theory assumes that the governing differential equations are linear and have a wavelength-dependent solution corresponding to time-independent dominant wavelength. Actually, (especially young) lithosphere may exhibit a significantly nonlinear response to the horizontal compression, resulting from various factors such as faulting, softening, strain rate dependence of the yield strength, or, finally, three-dimensional (3-D) effects. The last factor can not only affect the wavelength of folding but also lead to the appearance of a second dominant wavelength associated with out-of-plane deformation. All these factors would result in spatial distribution of folding, aperiodical time-dependent spacing of fold limbs, and strong lateral amplitude variations. The deviations from the assumptions of Biot's (or Ramberg's or Smith's) theory also include complex situations where folding (1) was followed or already preceded by distributed faulting [e.g., *Gerbault et al.*, 1999], (2) has been imposed on previously extended, predeformed lithosphere [e.g., *Brun and Nalpas*, 1996], (3) involved deformed layers that were too short compared to their thickness [e.g., *Burov and Molnar*, 1998], or (4) involved a part of the lithosphere that had too many lateral inhomogeneities, affected by significant sedimentation and erosion [e.g., *Cloetingh et al.*, submitted manuscript, 1999]. In these cases, folding instabilities can be poliharmonic or aharmonic or may exhibit much longer or shorter wavelengths than those of the theoretical predictions. For example, loading of a very young lithosphere by horizontally spreading sediments may increase the wavelength of the deformation (which is probably the case for the Pannonian basin). For these reasons, such folding instabilities are difficult to discriminate and to interpret using conventional criteria based on spectral analysis. Therefore we have to investigate some secondary characteristics of folding, such as, for example, an unusually high amplitude of undulations and accelerations of vertical motions (North Sea basin, 6 Myr, [*van Wees and Cloetingh*, 1996], Ferghana and Tadjik basins, 8-10 Myr, [*Burg et al.*, 1994; *Burov and Molnar*, 1998], Pannonian basin, 4-6 Myr, [*Horváth and Cloetingh*, 1996]; Iberian continent, 6-8 Myr, [Cloetingh et al., submitted manuscript, 1999; *Waltham et al.*, 1999]), southern Tyrrhenian Sea, 6-8 Myr, [*Mauffret et al.*, 1981], and Gulf of Lion, 6-8 Myr, [*Kooi et al*, 1992]). The use of these characteristics is based on the notion that vertical undulations due to compressional instability develop much faster (several times) than those produced by other mechanisms such as volumetric thickening and thermally or conventionally induced vertical movements [e.g., *Gerbault et al.*, 1999]. In most typical cases, only a few percent of shortening is sufficient to produce unstable undulations with an amplitude of a few kilometers, whereas other mechanisms would require more than 10% shortening. This is the case for the 3 - 5 km high Tien Shan range in central Asia, which was formed within approximately 10-15 Myr as a result of about 250 km plate shortening induced by the Indian–Eurasian collision, whereas the first buckling-induced neotectonic movements with amplitudes of a few kilometres developed in this area already after approximately 50 km shortening, several megayears before the creation of the mountain range [*Nikishin et al.*, 1993].

3. Geological and Geophysical Records of Folded Areas

3.1. New Observational Methods of Identification of Folding

The geological and geophysical recognition of folding requires a multidisciplinary approach. With the exception of the ocean lithosphere, folding is rarely reflected directly in the topography. Because of morphogenic activity modifying the surface topography and the possibility of crustal-mantle decoupling, folding is most times better reflected in the sedimentary records, differential subsidence and uplift data, fault spacing, seismic refraction and reflection profiles, and in the data on Moho topography (known from seismic reflection, gravity, and other data). The mantle deformation has normally very long wavelength characteristics requiring long geophysical and geological crosssections obtained using deep penetration techniques. For shallow crustal folding, high-resolution shallow seismics has proven to be a very important new diagnostic tool, additionally constraining the geometry of folding-associated crustal faulting. New petroleum industry data sets based on deep seismic reflection profiling such as those carried out along the Norwegian margin have also underlined the importance of the large-scale compressional deformation in passive margin settings, i.e., areas previously considered tectonically quiet [*Vagnes et al.*, 1998]. It should be realized that although the subtle geomorphic expressions of large-scale tilting have been recognized frequently by geomorphologists, the link with the intraplate deformation on the lithosphere and the crustal scale was often not made, since the supporting evidence from the geological and geophysical data sets were till recently lacking [*Horváth and Cloetingh*, 1996]. Only over the last few years, an integration of evidence from different data sets at different temporal and spatial scales has been made, which has provided a growing basis in support of folding as a key mechanism of continental intraplate deformation. Important differences were recently recognized between the wavelengths and durations of folding. In section 3.2 we will examine the different types of folding, particularly focusing on the atypical expressions of folding and the role of rheology and mechanical structure on the preservation of the folds.

3.2. Regular and Irregular Lithospheric Folding: Natural Examples and Predictions

In continental areas a number of examples of prominent periodical folding are found (which we call distinct or regular), such as Arctic Canada [*Stephenson and Cloetingh*, 1991], central Australia [*Stephenson and Lambeck*, 1985], central Asia [*Martinod and Davy*, 1992, 1994; *Nikishin et al.*, 1993; *Burov et al.*, 1993; *Cobbold et al.*, 1993], Tibetan plateau [*Burg et al.*, 1994], and Paris basin [*Lefort and Agarwal*, 1996]. Evidence for unstable lithospheric deformation in these areas is inferred from the presence of periodical undulations of the basement and Bouguer gravity anomalies, by indication of spatially periodic tectonic movements (with wavelengths matching the predictions of the linear theory of folding) and by rapidity of the growth of the vertical undulations. In the cases referred to as irregular or nonlinear, folding is aperiodic or has wavelengths significantly differing from the predictions of the linear theoretical model. Figures 1 and 2 and Table 1 summarize the regular and irregular cases of folding. Figure 1 displays observed wavelengths of folding plotted against the theoretical dependence between the wavelength of linear folding and thermotectonic age of the lithosphere, which reflects the thickness of the competent crust and mantle. As shown, most cases of irregular folding correspond to young, weak lithosphere, which is more affected by the side effects than the old and strong lithosphere is. Figure 2 demonstrates some representative data crosssections for regular and irregular folding. In the case of western Goby [*Burov et al.*, 1993], presented in Figure 2a, the wavelength of folding is close to the analytical solution for power law rheologies derived on the basis of the linear analysis by *Smith* [1975,1977,1979]. The young Pannonian basin [*Horváth and Cloetingh*, 1996] exhibits extremely long-wavelength deformation (300 km) compared with 50 km to the maximum 100 km predicted from the linear theory. The "intermediate" Ferghana basin in central Asia [*Burov and Molnar*, 1998] has a "correct" wavelength of deflection, but the ratio plate length/wavelength is too small to satisfy conditions of thin plate/layer approximation. The gravity signal over the Ferghana basin also has some peculiar short-wavelength features, which can result from plastic hinging or faulting on the sides of the basin.

In the present study we demonstrate the relative importance of lithospheric parameters on compressional basin deformation in terms of preexisting rheology, strain rate history, and the forces driving compression. To this aim, we have selected a number of areas in the Africa/Arabia -Eurasia plate collision zone and in the Alpine foreland (see Table 1 for the key parameters of folding) for a closer investigation of parameters characterizing their large-scale compressional deformation.

In section 3.2.1 we summarize geophysical evidences for folding in young extensional basins affected by late stage compression: the Gulf of Lion and the Tyrrhenian, Pannonian, and Transylvanian basins. We also describe in sections 3.2.2 – 3.2.6 the data related to other cases of irregular folding: the Iberia basin, the North Sea basin, the Helland-Hansen Arch, the Russian platform, and the Ferghana and Tadjik basins in central Asia.

3.2.1. Western Mediterranean and Pannonian basin. On the basis of seismic profiles in the western Mediterranean, *Mauffret et al.* [1981] pointed to strong evidence for late stage compressional basin deformation in this area. *Cloetingh and Kooi* [1992] noticed the simultaneous occurrence of accelerations in Pliocene-Quaternary subsidence with the onset of late stage compression, proposing large-scale downwarping as a mechanism. A more extensive investigation, backed up by large-scale modeling and gravity studies, was recently done for the Pannonian basin [*Vackarcs et al.*, 1994; *Horváth and Cloetingh*, 1996] and the Gulf of Lions [*Kooi et al.*, 1992; *Chamot-Rooke et al.*, 1999]. A modeling study, constrained by more recent seismic evidence for compressional deformation in the Tyrrhenian Sea [*Pepe,* 1998] is yet to come.

The Pannonian basin, formed by Neogene stretching and thinning of thickened Alpine lithosphere [*Horváth*, 1993], is presently probably the hottest basin of Europe. The crustal structure and subsidence history of the basin have been extensively documented through a large array of deep seismic reflection profiles, industry seismics [e.g., *Posgay et al.*, 1996], and boreholes [*Horváth*, 1993]. The Pannonian basin underwent a Pliocene-Quaternary reactivation, leading to large-scale warping of the lithosphere with a characteristic wavelength of several hundred kilometres [*Vackarcs et al.*, 1994; *Horváth and Cloetingh*, 1996]. The deformation has a clear expression in the topography of the area and has been recorded in tilted river terraces [*Horváth and Cloetingh*, 1996]. Seismic high-resolution shallow profiling has extensively documented the fine structure of the deformation [*Horváth and Cloetingh*, 1996]. The present-day stress field of the Pannonian basin has been studied extensively through focal mechanism studies and borehole breakout studies [*Gerner et al.*, 1999], demonstrating that the anomalous subsidence and uplift patterns occur in a regime of present-day compression. As illustrated by Figures 1-3, the first-order patterns of lithospheric deflection are consistent with an explanation in terms of late stage compression. Superimposed on the intralithospheric folding are contributions by slab detachment processes at the margins of the Pannonian basin, recorded by anomalous Pliocene-Quaternary uplift in the Styrian basin [*Sachsenhofer et al.*, 1997] and the Transylvanian basin. Within the Transylvanian basin itself, also a late stage deepening has been observed compatible with the documented stress field in the area [*Negut et al.*, 1997; *Ciulavu*, 1999]. As one can see on the profiles intersecting the Pannonian basin, there is no principal difference between the two orthogonal directions even though the major tectonic compression is roughly northward. This can be explained by important transpressional deformation of this three-dimensional basin, thus justifying the initial hypothesis of a folding mechanism. Of course, slab detachment or surface processes could also be responsible for the rapid vertical movements of the Pannonian basement but quite unlikely in both orthogonal directions. In contrast, a compressional instability in three dimensions would be a plausible mechanism, since, assuming "rigid" borders of the surrounding region, it may cause "circular" folding even due to unidirectional deformation.

3.2.2. Iberian continent. Another irregular case, Paleozoic Iberian continent is located at a very close distance to the African-Eurasian plate boundary. Iberia is surrounded by zones of a high level of recent tectonic activity, with late Neogene opening of extensional basins in the adjacent western Mediterranean [*Janssen et al.*, 1993; *Docherty and Banda*, 1995], the Betic and Pyrenees orogenies at its southern and northern margins [*Verges et al.*, 1995], and the Gibraltar-Azores deformation area in the southwest [*Masson et al.*, 1994]. The plate tectonic activity at Iberia's margins has gone through a time sequence starting in Eocene times in the north [*Roest and Srivastava*, 1991] and continuing until the present day at the southern plate boundary. Recently, the lack of data for the Iberian part of the World Stress Map has been compensated by a substantial body of new stress indicator data from borehole breakouts and structural analysis (see for an overview *Ribeiro et al.* [1996], *Jurado and Mueller* [1997], and *de Vicente et al.* [1996]). These data show the existence of consistently oriented stress fields in Iberia, with an orientation consistent with the plate interactions in the area. The central part of Spain occupies a position at the crossroads of the stresses propagated into the plate interior and provides a natural laboratory to examine the fine structure of intraplate deformation. Analysis of gravity, topography, and stratigraphy has shown the existence of large-wavelength folding with typical wavelengths of 300 km [*Stapel et al.*, 1997; *Waltham et al.*, 1999]. It is noteworthy that *Cobbold et al.* [1993] have already suggested possible folding scenarios for the deformation of the intra-Iberian Ebro basin, whereas the new data confirm that the same scenario can be applicable to the whole continent.

3.2.3. North Sea basin. The North Sea basin has, upon its formation by extension in Permo-Triassic time, been subjected to several phases in its basin evolution. The subsidence history is marked by a first initial phase of subsidence, compatible with predictions from stretching models, followed by a phase of thermal subsidence, interrupted by phases of atypical subsidence and uplift, with a timing and spatial distribution suggesting a far-field control by plate boundary forces operating at the Alpine collision zone [*Ziegler et al.*, 1995]. Particularly striking is the Pliocene-Quaternary acceleration of tectonic subsidence in the central part of the North Sea. As in the Pannonian basin, this acceleration in subsidence occurs at a time when according to predictions from stretching models subsidence should have decayed to essentially zero [*Cloetingh et al.*, 1990; *Cloetingh and Kooi*, 1992]. Two-dimensional (2-D) [*Kooi et al.*, 1991] and 3-D modeling studies [*van Wees and Cloetingh*, 1996] have demonstrated that the observed large-scale warping is compatible with present-day compression in NW Europe [*Muller et al.*, 1992]. The 3-D modeling has demonstrated that for stresses equivalent to those induced by ridge push forces, 700 m of accelerated subsidence in the North Sea basin can be induced [*van Wees and Cloetingh*, 1996]. Along with this downwarping in the central North Sea, an upwarping is generated at the margins of the North Sea basin, amplifying postglacial rebound, consistent with recent data [*Japsen*, 1998].

3.2.4. Helland-Hansen Arch. As a result of intensive petroleum exploration, detailed evidence has been found for the widespread occurrence of large-scale Cenozoic compressional domes associated with intensive crustal faulting on the margins of the northern Atlantic [*Dore and Lundin*, 1996]. The characteristic wavelength of these undulations over the southern part of the Helland-Hansen Arch is about 60 km, indicating that folding occurs in the upper crust only, as a result of crust-mantle decoupling due to very intensive thermal regime [*Vagnes et al.*, 1998].

3.2.5. Russian platform. With the advent of more detailed information of the crustal and sedimentary record of the Russian platform [*Nikishin et al.*, 1996] and the application of backstripping analyses techniques to the sedimentary record, it has become clear that the Russian platform over its long geological history has recorded important changes in stress propagated into this vast area from the plate boundaries. The present accelerations in neotectonic subsidence recorded in, for example, the Moscow basin appear to a large extent to be associated with the plate coupling processes in the Caucasus region [*Nikishin et al.*, 1996] and are possibly related to large-scale spatially periodic instabilities.

3.2.6. Ferghana and Tadjik basins, central Asia. The young Ferghana and Tadjik basins are two compressional basins northwest and north of Pamir and south of the Tien Shan ranges [*Burov and Molnar*, 1998]. They both underwent Jurassic rejuvenation and are characterized by very young thermal age (175 Ma) [*Burg et al.*, 1994; *Burov and Molnar*, 1998]. In addition to thermal weakening, these basins also probably have a weak lower crustal rheology (quartz), resulting, together with young thermal age, in low effective elastic thickness (EET) values of 15 km [*Burov and Molnar*, 1998]. The gravity data suggest that these two compressional basins are gravitationally overcompensated (i.e., the gravity anomalies indicate a several kilometres deeper Moho than that predicted by local isostatic models [*Burov and Kogan*, 1990; *Burg et al.*, 1994; *Burov and Molnar*, 1998]), which has lead to a hypothesis that these basins may result from unstable downwarping, possibly associated with deep mantle faulting.

It should be noted that the subdivision into regular and irregular folding is very approximate. For example, folding in the Ferghana and Tadjik basins can be treated as both regular (the wavelength roughly matches the linear theory predictions) and irregular (the area is faulted and probably was formed in a zone of preexisting weakness; formally, it is also too short to be analyzed using thin layer approximation, see Figure 2).

4. Mechanical Properties of the Lithosphere in Compressional Zones

Lithospheric folding is largely controlled by rheological and thermal structure. Effects of different initial crustal thickness and thermal state of the prefold lithosphere determine the effectiveness of far-field and near-field stresses and associated strain rates in the process of compressional basin deformation. Both these parameters control the prefold rheology, the material parameters of which are constrained by results of experimental rock mechanics data adopting the yield strength envelope concept [*Carter and Tsenn*, 1987; *Cloetingh and Banda*, 1992; *Ranalli*, 1994; *Burov and Diament*, 1995; *Cloetingh and Burov*, 1996]. The synthetic strength profiles derived for a representative background strain rate provide a qualitative picture of the distribution of lithospheric strength with depth, and they give a measure for the maximum stress levels to be supported by the lithosphere. In reality, strain rate may vary laterally and with depth, and for this reason, direct thermomechanical numerical calculations based on the explicit form of the rheological laws can give a more precise idea of strength and stress distribution in the lithosphere. Composition and temperature are controlling factors on bulk lithospheric strength [*Kusznir and Park*, 1987], reflected also in indirect observables such as EET [*Burov and Diament*, 1995] and distribution of intraplate seismicity [e.g., *Cloetingh and Banda*, 1992]. Important differences can be expected in the starting conditions for folding in relatively warm Alpine lithosphere such as that encountered in Iberia or the Pannonian basin, as opposed to cold foreland lithosphere such as that encountered in the North Sea basin and the Russian platform. The evidence for the rheological state of the lithosphere comes not only from these strength profiles but is also available through an extensive set of modeling studies carried out in Europe's main basins, formed by extensional processes in Mesozoic and Tertiary times but deformed by late stage compression only very lately in Pliocene-Quaternary times [e.g., *Horváth and Cloetingh*, 1996; *van Wees and Cloetingh*, 1996]. The kinematic models for extensional basin formation for the Alpine/Mediterranean and North Sea basins [e.g., *Kooi et al.*, 1992; *Cloetingh et al.*, 1995] and their validation through testing with data sets from rifted basins have yielded important constraints on the mechanics of the prefolding stage. In addition, questions have risen concerning the mantle part of the inferred strength distributions, suggesting that mantle strengths are possibly overestimated. For example, intraplate seismicity distributions do not show earthquake foci at the levels of the subcrustal mantle lithosphere, where rheological profiles predict considerable strength [*Cloetingh and Banda*, 1992].

Within the European lithosphere, considerable spatial variations in lithosphere rheology occur [*Cloetingh and Burov*, 1996]. In general, these variations follow the predictions of thermal models for continental lithosphere evolution, showing a clear increase of the bulk strength of the lithosphere with increasing thermotectonic age. As demonstrated by *Cloetingh and Burov* [1996], the predictions from strength profiles constructed from extrapolation of rock mechanics data are compatible with the outcomes of flexural studies of Europe's foreland basins. On a more regional scale, strength profiles have been constructed for a number of regions of Europe, including the Romanian part of the Pannonian basin and the Transylvanian basin [*Lankreijer et al.*, 1997], the Mediterranean margins of Iberia [*Cloetingh et al.*, 1992], and the Tyrrhenian Sea [*Spadini et al.*, 1995]. In general, the strength distribution patterns suggest a relatively low strength of the lithosphere in the near-field areas close to the African-European plate boundary.

Basically, the major differences in prefold rheology are the result of a cool versus a warm prefold lithosphere in the areas analyzed. Whereas Mesozoic and Paleozoic rifting and subsequent neotectonic folding probably took place under relatively stable thermal conditions, the Alpine/Mediterranean setting can be characterized by a transient thermotectonic regime. As observed from a comparison of the Alpine/Mediterranean basins and the cratonic basins, folding durations tend to be longer for the cratonic basins, probably reflecting a larger integrated strength for this class of basins.

5. Mechanical Model of Folding

In this section we conduct a number of numerical experiments allowing us to investigate the development of folding instabilities in "realistic" conditions, i.e., explicitly taking into account most of the factors that are thought to be responsible for deviations from the linear behavior. These factors include brittle-elasto-ductile rheology explicitly taken from the laboratory data, temperature, large strains, horizontally and vertically variable strain rates, and the possibility of lithospheric faulting during deformation.

5.1. Model Setup and Numerical Experiments on Regular Folding in Realistic Conditions

Figure 3 presents setups for both linear analytical and numerical models of folding. Figure 3a shows a sketch of typical monoharmonic and biharmonic folding models with an analytical estimate for the growth rate dependence on the λ/h ratio (λ is the wavelength of folding, and h is the thickness of the competent layer) [*Burov et al.*, 1993]. The brittle-elasto-ductile lithospheric rheology and lateral discontinuities strongly affect the dynamics of the lithospheric deformation [*Vilotte et al.*, 1993; *Burov et al.*, 1993]. For this reason, to investigate the development of folding instabilities in brittle-elasto-ductile multilayers, we adopted the Fast Lagrangian Analysis of Continua (FLAC)-based [*Cundall*, 1989] finite element code Paravoz [*Poliakov et al.*, 1993], which has a good record in modeling of folding and buckling instabilities in nonlinear media [*Burov and Molnar*, 1998; *Gerbault et al.*, 1999] (see appendix for the details on the numerical method).

The finite amplitude of deformation in an unstable compressive regime can be influenced by many usually negligible factors. Therefore our primary goal was only to obtain a reasonable amplitude of deformation applying reasonable forces (i.e., not exceeding mechanical strength of the lithosphere). For the layers with EET exceeding 15-20 km, the wavelength of the deformation is seemingly stable and is not significantly affected by variations in the loads (Figure 1). Thus the EET usually controls well the wavelengths that we can observe, but the magnitudes of deflections and corresponding gravity anomalies depend also on density contrasts and the magnitude of the horizontal force.

During our numerical experiments we have tested a significant number of situations encompassing most of the possible scenarios starting from relatively young continental lithosphere (thermotectonic age less than 150 Ma) and ending with very old lithosphere (thermotectonic age 2000 Ma). Figure 3b presents results of one of the series of experiments with non inear brittle-elastic-ductile rheology derived from rock mechanics data (Table 2, appendix. Figure 3b demonstrates the development of pronounced decoupled folding in a central Asia-type lithosphere (thermotectonic age from 175 to 450 Ma). These experiments were completed using quartz-olivine rheology from a previous study *by Burov et al.* [1993] (Table 2) that infers a weak lower crust promoting partial mechanical decoupling between the competent upper crust and the upper mantle. As can be seen, folding of a lithosphere with such a mechanically weak lower crust (of which there is a number of examples in central Asia, [*Sabitova*, 1986; *Roecker et al.*, 1993; *Burov and Molnar*, 1998]), is characterized by two characteristic wavelengths. The short one corresponds to crustal folding (30-60 km), and the larger one (200-350 km) corresponds to mantle folding. Figure 3b presents results for the thermotectonic age of 250 Ma, yielding 40-50 and 200-250 km wavelengths for the crust and mantle, respectively. Assumption of an older age (400 Ma) results in larger respective wavelengths of 60 and 350 km. In spite of the nonlinear rheology and intensive faulting, this experiment demonstrates good correspondence with the Biot - Smith's theory ($h_1 \approx$ 10-15 km, and $h_2 \approx$ 50-60 km; Figures 1, 2, and 3a), which confirms our recent result [*Gerbault et al.*, 1999] that faulted layers can effectively transmit horizontal stress. The next experiment (Figure 4) considers a coupled typical case of cold lithosphere (e.g., Australian craton before rejuvenation, age > 700 Ma; Table 1). The wavelength of the deformation is much larger (> 500 km) than that in the case of Figure 3b, since the crust is mechanically welded with the mantle lithosphere leading to total effective mechanical thickness of around 120 km. It can be seen that folding may activate intensive faulting in the crust and mantle and then continue for a very long time. In our longest experiments, folding persisted for 25 Myr before the final localization of the deformation on one single fault. Surface processes (erosion and sedimentation) may significantly prolong the lifetime of folding, since they decrease the effect of gravity by (1) filling the downward flexed basins and thus reducing the restoring force and (2) cutting the upward flexed basement and thus unloading the lithosphere in the uplifted areas.

5.2. Experiments on Irregular Nonlinear Folding

As was noted in the previous sections, various factors, such as nonlinear rheology, large surface loads, lateral inhomogeneities, and large strains may lead to significant

Table 2. Parameters of Dislocation Creep for Lithospheric Rocks and Minerals

Mineral/Rock	A, Pa^{-n}s^{-1}	H, kJ mol^{-1}	n
quartzite (dry)	5×10^{-12}	190	3
diorite (dry)	5.01×10^{-15}	212	2.4
diabase (dry)	6.31×10^{-20}	276	3.05
Olivine/dunite (dry)*	7×10^{-14}	520	3

Parameter values correspond to the lower bounds on the rock strength [*Brace and Kohlstedt*, 1980; *Carter and Tsenn*, 1987; *Tsenn and Carter*, 1987; *Kirby and Kronenberg*, 1987]. The elastic moduli used for all materials throughout this paper are E (Young's) modulus = 0.8 GPa and ν (Poisson's ratio) = 0.25. The brittle properties are represented by Mohr-Coulomb plasticity with friction angle 30° and cohesion 20 MPa [*Gerbault et al.*, 1999]. For olivine at σ_1-$\sigma_3 \geq 200$ MPa (Dorn's dislocation glide), $\dot{\varepsilon} = \dot{\varepsilon}_0 \exp\{-H^*[1-(\sigma_1-\sigma_3)/\sigma_0]^2/RT\}$, where $\dot{\varepsilon}_0 = 5.7\times10^{11}$ s^{-1}, $\sigma_0 = 8.5\times10^3$ Mpa, and H = 535 kJ/mol.

*Parameter values for dislocation climb at σ_1-$\sigma_3 < 200$ MPa deviations of folding behavior from that predicted by the linear theory [*Davis*, 1994; *Zhang et al.*, 1996; *Mercier et al.*, 1997; *Bhalerao and Moon*, 1996; *Hunt et al.*, 1996; *Lan and Hudleston*, 1996]. Of course, this especially concerns young, thin lithosphere and much less old, strong continental plate segments. In the case of irregular folding (generally very young, weak lithosphere), the deformation may be very much affected by various factors (gravity-driven lateral spreading of the sedimentary infill, inhomogeneities, and nonlinear behavior). Consequently, the wavelength of the deformation can be much longer or shorter than that predicted by Biot's theory. The deformation may be highly aperiodic as well

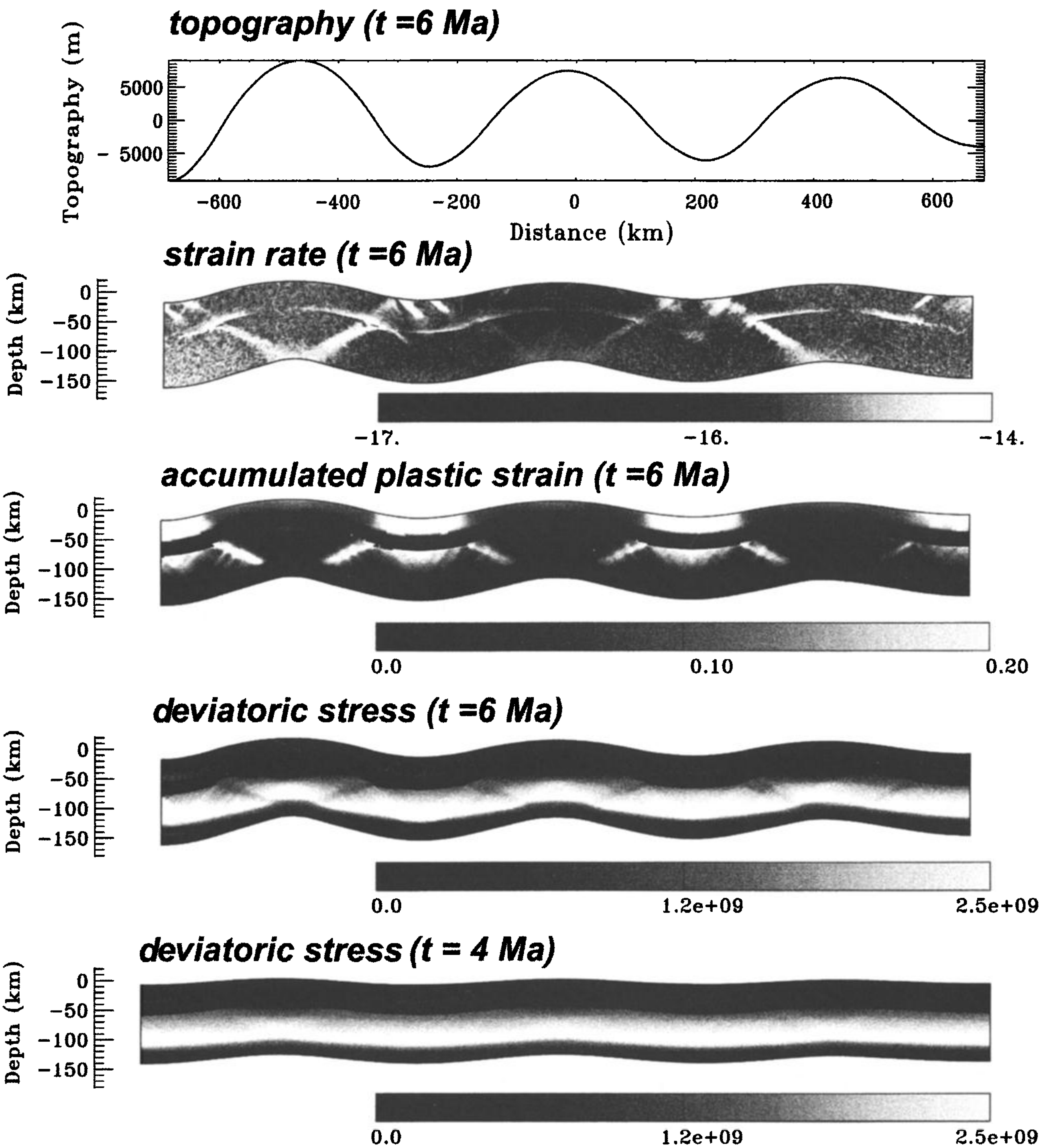

Figure 4. Coupled folding (old cratonic lithosphere with thermal age > 700 Ma and strong diabase lower crustal rheology; Table 2). Note the crustal and mantle faulting and a larger wavelength of the deformation (500 km). Shortening is at a rate of 1.5 cm/yr.

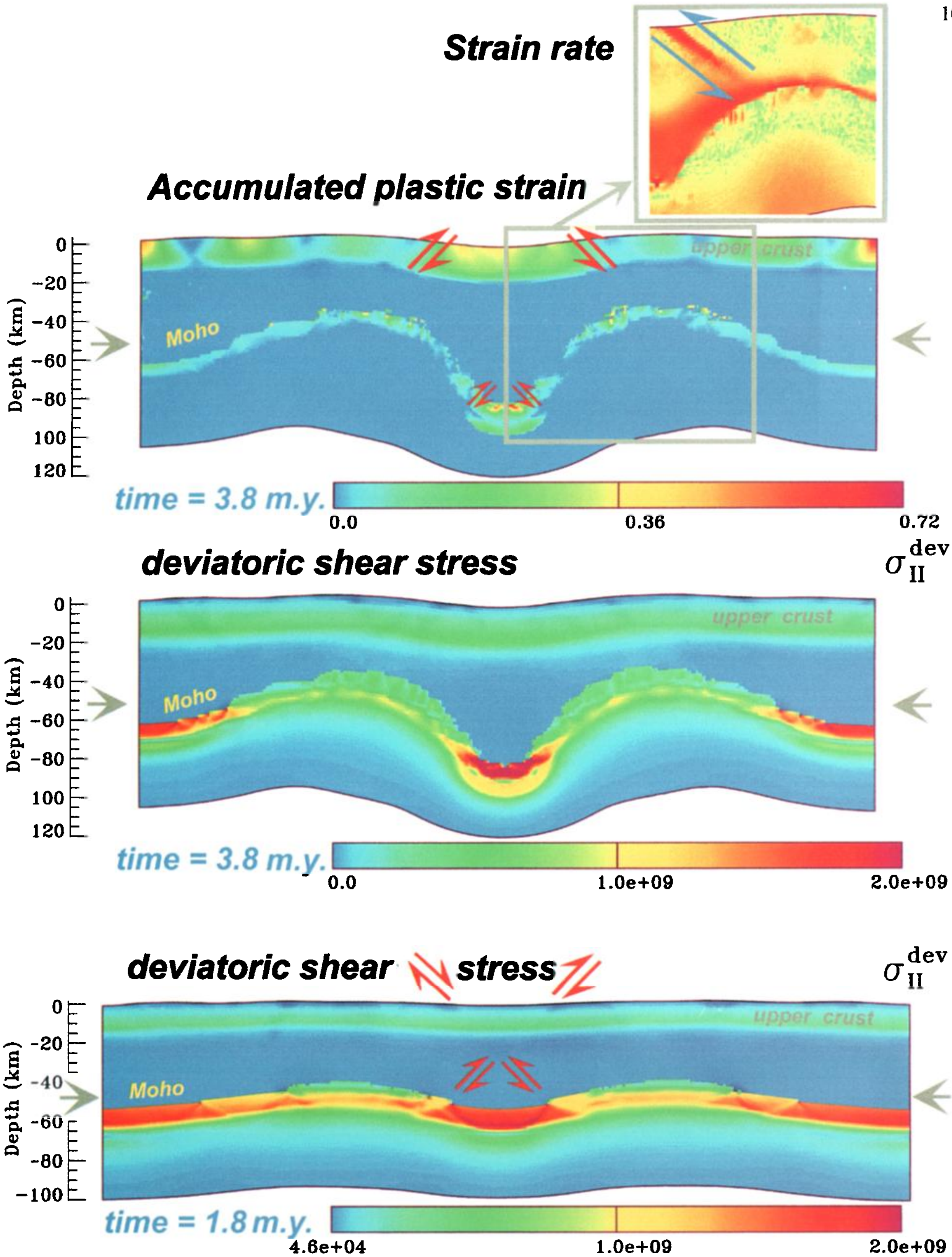

Plate 1. Numerical experiment demonstrating nonlinear, irregular folding: folding and faulting experiment (Ferghana/Tadjik basin type lithosphere, central Asia, quartz crustal rheology, olivine mantle rheology (Table 2), thermal age 175 Ma. The faults appear before the folding develops, but then the two processes, faulting and folding, can co-exist in such a way that folding is accommodated by faulting. Because of the weakness of the lower crust, the upper crust is completely decoupled from the mantle and interacts with it only by flow in the lower crust. The participation of the flow changes the wavelength and amplitude of folding, which finishes by the development of a single downwarped "megafold."

[*Hunt et al.*, 1996]. As noted in section 3.2, a new database including such irregular folding cases is now available for large areas of the European foreland and its margins as a result of intensive geophysical and geological studies carried out during the last decade, which have significantly enhanced the current understanding of the basin (de)formation mechanisms operating in these areas.

Plate 1 demonstrates the case of inverse "megafolding", the result of the prolongation of the numerical experiments considering simultaneous development of folding and faulting in a young Ferghana/Tadjik type basin (Figure 2c, [*Burov and Molnar*, 1998], which, as was mentioned above, possibly results from unstable downwarping, and may be associated with deep mantle faulting. Actually, for Ferghana-Tadjik settings our numerical experiments reproduce the formation of active crustal and mantle faults with characteristic spacing corresponding to the thicknesses of the brittle crustal and mantle domains. Surprisingly, the appearance of faults does not significantly influence the wavelength of folding: both processes continue to coexist for a long time, so that faulting serves as a mechanism of folding in the brittle domain [*Gerbault et al.*, 1999]. Also observed in some analogue experiments [*Martinod and Davy*, 1994], this "continuous" behavior of faulted lithosphere can be explained by fault locking due to gravity and friction: after some sliding (uplift) on the fault, the potential gravity energy and thus work against friction to be done by the forces of horizontal shortening become too high, and the fault locks and transmits the horizontal stress as a continuum medium. As soon as the compression continues, one of the folds finally starts to grow faster than the others, resulting in a loss of periodicity and the formation of a mega-fold (Plate 1), which can finish up by initialization of subduction and mountain building. In the case presented in Plate 1, large horizontal shortening also resulted in the decrease of the observed wavelength of mantle folding from the initial value of 250 km to approximately 150 km, and the increase of the crustal wavelength from 40-50 to 100 km. Note that the experiments shown in Figure 5 present a "developed" case with respect to that demonstrated by *Burov and Molnar* [1998]. Consequently, these results can be regarded as a possible scenario of the evolution of the deformation in the recently inverted basins.

As another possible example of irregular folding, Figure 5 demonstrates a case of demicoupled/demidecoupled lithosphere (similar to Figure 3b but with colder geotherm of 400 Ma). In the intermediate cases (between the coupled and decoupled state) the mantle lithosphere can be in some places coupled or decoupled with the upper crust, depending on the stress and strain rate. In this case, some parts of the plate may deform in a biharmonic mode whilst others will exhibit longer-wavelength monolayer folding.

5.3. Experiments on Preserved Folding

As was discussed by *Bird* [1991] and *Avouac and Burov* [1996], large-scale undulations of the lithosphere in the absence of sufficient compression cannot be preserved for a long time (> 10 Myr), except for very strong (especially low crustal) lithospheric rheology. Otherwise, they will be flattened owing to the gravity-driven crustal flow associated with the omnipresent large crust-mantle density contrast at the

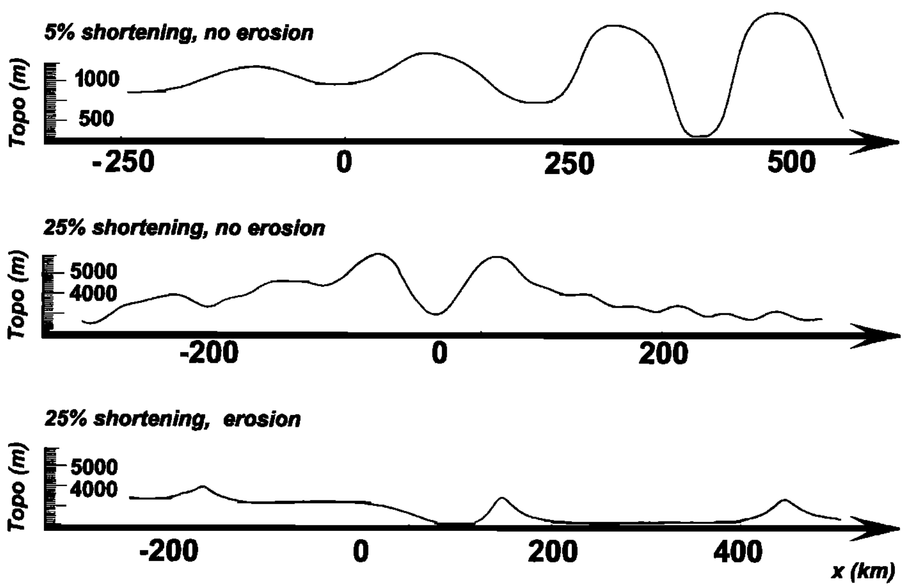

Figure 5. Numerical experiment demonstrating nonlinear, irregular folding. In some cases, the instabilities can be quite chaotic. Figure 5 demonstrates different cases of irregular folding with wavelength and amplitude varying along the plate on different stages of deformation due to partial crust-mantle coupling and strain localizations (400 Ma lithosphere with weak quartz-dominated crustal rheology; Table 2). After 5% shortening, (top), after 25% shortening, (middle), after 25% shortening (bottom), strong zero-order diffusional erosion [*Avouac and Burov*, 1996] tuned to keep mean elevations at the level of 3000 m. Erosion reduces the contribution of gravity-dependent terms (middle wavelength) and accelerates local deformations. Strong erosion, insufficiently compensated by the tectonic deformation (bottom), wipes out most of the topography. Yet, if the erosion is tuned to the average elevation rates, it may dramatically accelerate folding.

Moho boundary. Near the Moho, some parts of the folded crust occur at the same depth as the folded dense mantle, resulting in remarkable pressure differences (5 MPa per 1 km of Moho depression plus a contribution from the hydrostatically disbalanced part of the surface depression). These pressure differences in most cases are sufficient to overcome the yielding stress of the lower crust at Moho depth levels [*Bird*, 1991]. As a result, the crust would flow and flatten the bent layer. The occurrence of long - timescale preservation of folding after cessation of the compression (Figure 6) in the presumably strong Australian craton, Parisian basin or Russian platform points to a quite strong rheology yielding high EET values (> 60 km) and consequently confirming experimental estimates listed in Table 2. Yet the amplitude of the vertical deflection in the Paris basin is quite small, allowing for other mechanisms such as simple flexure due to the load by the Alpine system. However, the periodic deformation at the same wavelength is more or less well expressed to the northwest of the area, which would not be the case if the northeastern part of the basin was simply flexed down by the load of the Alps.

Most of the presently observed areas of folds coincide in time with the Alpine collision 60 Myr ago. It is thus reasonable to assume that the most typical characteristic timescale of the gravity collapse of the large-scale folds in the intermediate-age lithospheres is limited by this time, though in the cases of very weak quartz-dominated lower crust the folds may disappear within the following 8-15 Myr (analogously to the estimates made by *Bird* [1991] for the mountains). The experiment of Figure 6 (old 1000 Ma lithosphere with diabase lower crust; Table 2) demonstrates the preservation of folding-induced deformation for 10 Myr (potentially > 50 Myr) following the cessation of the tectonic compression. In this experiment we stopped the horizontal compression after the first 14 Myr of shortening and then studied the asymptotic behavior of the system in the next 10 Myr. During this period the amplitude of folds decreased by less than 10%, which yields at least 50 Myr decay time as an asymptotic prediction. The presence of the crustal faults, of course, may accelerate the gravity collapse of the folds, resulting in the creation of the inverted basins. For example, the experiment of Figure 6 predicts reactivation and inverted activity of some faults after the cessation of the tectonic compression.

6. Discussion

Up till now, we have discussed various parameters characterizing the rheological state of the folded lithosphere and the associated deformation signature. The initiation of folded basin formation depends on the interplay of forces operating on the continental lithosphere and the spatial distribution of the lithospheric strength. In platform settings a large distance away from plate boundaries, far-field lithospheric stresses should be relatively constant [*Zoback*,

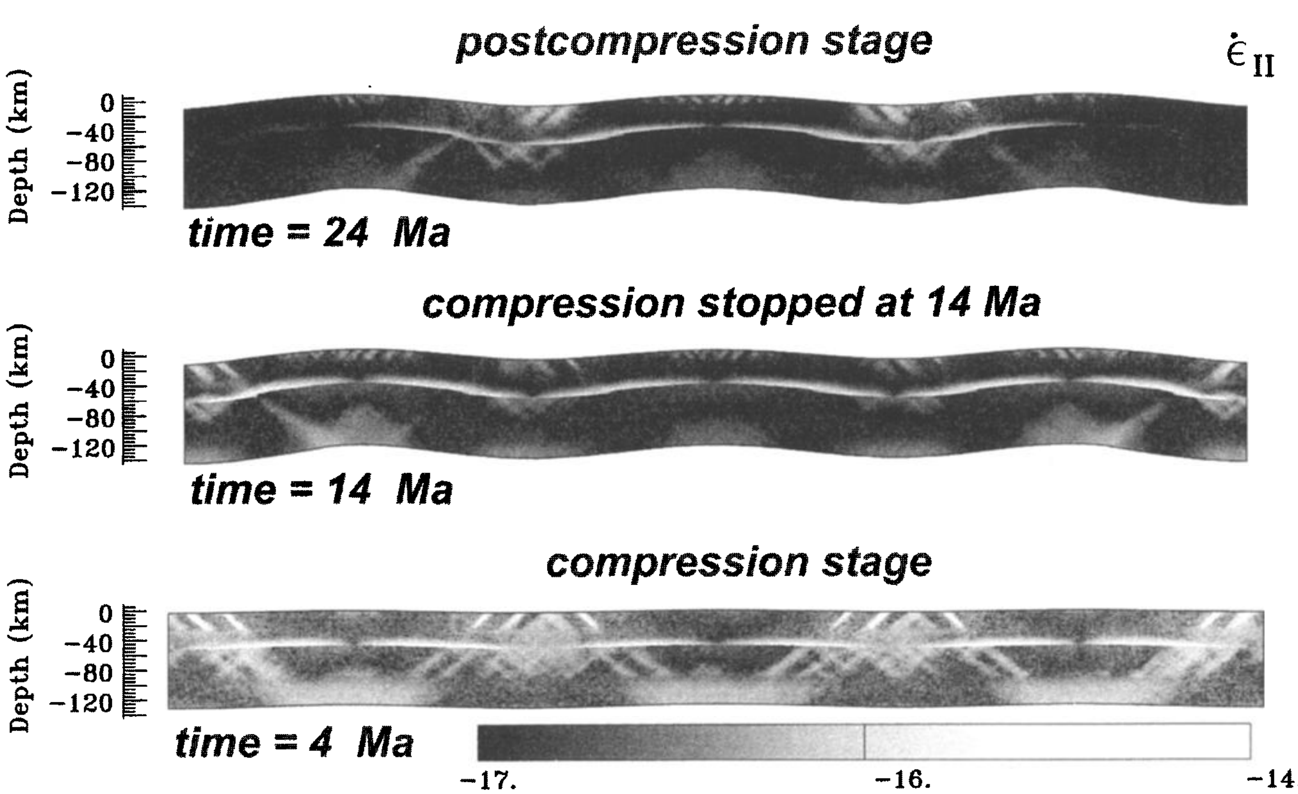

Figure 6. Strain rates for preserved folding. Preserved folding is an indicator of a strong lithospheric rheology (same parameters as those for the Figure 4, but with horizontal compression stopped after 14 Ma). It can be observed that for a strong cratonic-type rheology, folding may be frozen -in for a very long time.

1992], and therefore lithospheric deformation is expected to be primarily controlled by the initial distribution and subsequent evolution of lithospheric strength, possibly amplified by local sources of stresses. At plate margins, close to the main sources of intraplate stress, large horizontal and vertical fluctuations occur. For intraplate settings and adopting constant intraplate stresses, localization of extensional basin deformation is in agreement with a predicted reduction of extensional strength of the basin center relative to its margins [e.g., *Buck*, 1991; *Bassi*, 1995; *van Wees and Stephenson*, 1995].

Differences in the mode of folding between platforms and Alpine/Mediterranean basins appear to be primarily controlled by differences in (transient) thermal regime during the prefold phase and the duration of prefold lithospheric thinning, affecting the level of external stress required to result in folding. The latter is directly coupled to the presence of near-field versus far-field regional stresses.

6.1. Folding Wavelength

Folding of multilayers with lithospheric rheology can demonstrate good correspondence with Biot's linear theory only for small strains and relatively strong plates (EET > 15 km). Nevertheless, most of the cases presented in Figures 1 and 2 demonstrate more or less good correlation between the folding wavelength and crustal and lithospheric thickness (thermotectonic age). Inspection of the related field data reveals a particularly strong inverse correlation between the prefold crustal structure and wavelength in the Alpine/Mediterranean basins with an increasing thickness of the prefold crust, wavelengths are decreasing. This can be explained by the inverse dependence of the integrated lithospheric strength on the crustal thickness demonstrated by *Burov and Diament* [1995]. Thus lithosphere segments with thick crust are weaker and easier to fold. A less pronounced but similar correlation can also be observed between the lithospheric thickness and wavelength of folding. We infer from these findings that crustal control on folding wavelength prevails over the lithospheric control on this parameter, suggesting rheological decoupling in the prefold lithosphere. In contrast, and although characterized by a much larger scatter in the inferred parameter values, the case histories for foreland folding discussed here indicate a trend of an increase of the wavelength of folding with increasing lithospheric thickness, which may suggest that folding is primarily controlled by lithospheric thermal structure in these cases. This, however, does not apply to very young, weak basins, where the folding wavelength can also be strongly affected by factors other than the lithospheric strength (e.g., laterally spreading sedimentary infill, gravity-induced flattening of Moho due to weakly resisting lithosphere, heterogeneities, and basin geometry). In such areas the wavelength of folding can vary significantly, from a few to many hundreds of kilometers.

The data also show the absence of a correlation between prefold crustal thickness and wavelength in the Alpine foreland and platform areas. As pointed out by *Burov and Diament* [1995] and *Cloetingh and Burov* [1996], the highest lithospheric thickness values most likely reflect low mantle heat flow and a thermal structure, which would suppress intralithosphere decoupling during folding.

6.2. Folding Forces

Differences between the magnitudes of the driving forces for lithospheric folding can also be inferred from observations and models of folding. Whereas the Alpine foreland seems to have been predominantly affected by far-field stresses, in some cases possibly in combination with plume activity, the basins that formed in a regime of general convergence have mostly been affected by near-field stresses resulting from slab retreat [e.g., *Bassi and Sabadini*, 1994]. However, the estimation of actual forces may be complicated owing to the presence of various supplementary controls on their magnitudes. For example, sedimentary fill in the basins reduces the effect of gravity by decreasing the difference in density between the overlying and underlying material. As a result, buckling of a layer overlain by sediment that accumulates as buckling occurs can be initiated with a horizontal force lower than that for a layer that buckles into air. The presence of preexisting heterogeneities may also significantly reduce folding forces. The horizontal force required to buckle a plate with realistic brittle-elasto-ductile rheological structure is much lower than that which can be expected for folding of a plate with a linear rheology. In the experiments on the central Asian folding we concluded that folding of a brittle-elasto-ductile lithospheric plate with a Jurassic geotherm requires a horizontal force per unit length of the order of 10^{12} N/m, which is about one order lower than that in the conventional elastic models [e.g., *McAdoo and Sandwell*, 1985]. Finally, it should be noted that the 2-D models do not take into account out-of-plane stresses, which may lead to 2-3 times overestimated horizontal stresses [*van Wees and Cloetingh*, 1996].

In addition, rheological behavior at depth may exhibit some currently poorly constrained features such as strain softening and reduction of the brittle strength at significant depth. For example, the compressional brittle strength at the depth of 40-50 km may be 2-3 times lower than that induced from Byerlee's law [*Kirby et al.*, 1991; *Ranalli*, 1995]. However, a significant sharp competence contrast (> 10) between the stiff layer and embedding is always needed to maintain folding, which suggests the presence of strong lithospheric or crustal cores in folded areas (rather than strength distributed over the whole lithospheric thickness). For these reasons, further rheological investigations are needed to constrain horizontal tectonic forces operating in nature.

7. Conclusions

1. Folding (along with crustal and possibly mantle faulting) is a "standard" response of the lithosphere to tectonic compression, which is normally followed by localization of the deformation leading to orogeny and subduction. On a large scale, faulting does not prevent folding but actually serves as a mechanism of the unstable deformation in the brittle domain. This result is quite different from the small-scale studies of folding and faulting [e.g., *Johnson*, 1980] where the faults cannot be locked, owing to the little importance of gravity, and folding stops as soon as the system becomes faulted, just because sliding on the faults requires less energy than folding does. On a lithospheric scale, upward sliding on the faults is severely limited by the work against the gravity, friction, and

resistance of the embeddings. Thus the developed faults soon become locked, new faults form, and the system behaves as an essentially solid layered media.

We demonstrated the possibility of large-scale mantle faulting. This result is highly dependent on the applicability of the existing rheology laws at great depth. The indirect rheological data [*Kirby et al.*, 1991; *Ranalli*, 1995] suggest that the brittle strength might be much lower at the depths in excess of 40-50 km than the predictions of Byerlee's law. This boosts the importance of the brittle behavior on the sub-Moho depths and potentially resolves the problem of high horizontal stress required for the initialization of folding when commonly inferred Byerlee's brittle rheology is assumed.

2. The wavelength of the pronounced folding (strong lithosphere with competent layers with EET > 15 km) on the first stages follows the linear theory.

3. When irregular folding takes place, the wavelength of the "megafolds" is essentially determined by the geometry of the extensional basins; For example, weak, large basins may produce large-wavelength folding whilst strong, narrow basins may produce small-wavelength folding, with an important influence of the sedimentary load.

4. Developed stages of folding may be associated with lateral variations of the wavelength, irregular changes in the amplitude, and formation of megabuckles, etc. In the case of decoupled folding (weak lower crust), partial coupling with the mantle lithosphere may temporarily occur, owing to the ductile flow induced by folding of the mantle. This partial coupling may affect the wavelength of crustal folding and the amplitude of the basement subsidence.

5. Preservation of folding over large time spans > 20 Myr requires sufficiently strong rheology equivalent to that obtained for thermal age > 700-1000 Ma for the common rheological parameters from Table 2 (diabase lower crust). This explains why most of the known cases of folding are relatively recent. The relationship between the duration of preservation of folding and the age or EET of the lithosphere is nonlinear and is associated with short-lifetime spans of folds in the intermediate-age lithosphere and large-lifetime spans of folds in old lithosphere (cratons).

Appendix: Numerical Model

The plasto-elasto-viscous finite element code Paravoz [e.g., *Poliakov et al.*, 1993] is a fully explicit large-strain time-marching scheme, which solves the classical full Newton equations of motion (in a form adopted in the continuous mechanics) using fast Lagrangian analysis of continua (FLAC) [*Cundall*, 1989]:

$$\rho\, \partial v_i/\partial t - \partial\sigma_{ij}/\partial x_j - \rho g_i = 0\,,$$

where v is velocity, g is the acceleration due to gravity, and ρ is the density. This equation is written in small-strain formulation, but it can be used for large-strain problems if the local coordinates are dynamically updated to satisfy small strain conditions at grid points (large-strain deformation is modeled via sets of incremental small-strain deformations). This is achieved using the Lagrangian moving mesh method. In this method the numerical mesh moves with the material, and at each time step the new positions of the mesh grid nodes are calculated and updated in large-strain mode from the current velocity field using an explicit procedure (two-stage Runge-Kutta). Solution of these equations provides velocities at mesh points, which allows us to calculate element strains ε_{ij}. These strains are used in the constitutive relations to calculate element stresses σ_{ij} and equivalent forces $\rho\partial v_i/\partial t$, which form the basic input for the next calculation cycle. For elastic and brittle materials the constitutive relations have a linear form:

$$\varepsilon_{ij} = A\sigma_{ij} + A_0\,,$$

with A and A_0 being the constitutive parameters matrixes.

For the ductile rheology the constitutive relations become more complex:

$$\dot{\varepsilon}_{ij} = A\sigma^{n-1}\sigma_{ij}\,,$$

where $\dot{\varepsilon}_{ij}$ is the strain rate, and $\sigma=(1/2\ \sigma_{ij}\sigma_{ij})^{1/2}$ is the effective stress (second invariant). The variables n (the effective stress exponent) and A (constitutive parameter) describe the properties of a specific material (Table 2). For ductile materials, n usually equals 2-4, and A is depth- and temperature-dependent. For the brittle and elastic materials, A is usually only depth-dependent. Yet A and A_0 can be functions of strain or stress for softening or hardening materials. To allow for explicit solution of the governing equations, the FLAC method employs a dynamic relaxation technique based on the introduction of artificial inertial masses in the dynamic system. Adaptive remeshing technique developed by A.N.B. Poliakov and Yu. Podladchikov [*Poliakov et al.*, 1993] permits us to resolve strain localizations, leading to formation of the faults. The solver of the FLAC method does not imply any inherent rheology assumptions, in contrast with the most common finite element techniques based on the displacement method.

For the elastic rheology we adopted the following values of the constitutive parameters: E (Young's) modulus = 0.8 GPa and ν (Poisson's ratio) = 0.25. The brittle behavior is modeled by Mohr-Coulomb plasticity with friction angle 30° and cohesion 20 MPa [*Gerbault et al.*, 1999].

Since some of the rheological parameters are temperature-dependent, the equations of motion are coupled with the heat transport equations:

$$\rho C_p\, \partial T/\partial t - div(\mathbf{k}\nabla T) + \mathbf{v}\nabla T = H\,,$$

where $\mathbf{v}$ is the velocity tensor, C_p is the specific heat, $\mathbf{k}$ is the thermal conductivity tensor, and H is the radiogenic heat production per unit volume (here we use the commonly inferred values adopted, e.g., by *Burov et al.* [1993]. The size of the finite elements was between 2.5x5 and 5x7.5 km.

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