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Lithospheric folding and sedimentary basin evolution:

a review and analysis of formation mechanisms

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24 ABSTRACT

Lithospheric folding is an important mode of basin formation in compressional intraplate settings. Basins formed by lithospheric folding are characterized by distinct features in subsidence history. A comparison with extensional basins, foreland basins, intracratonic basins and pull-apart basins provides criteria for the discrimination between these modes of basin formation. These findings are important in deciphering the feed-backs between tectonics and surface processes. In addition, inferences on accommodation space and thermal regime have important consequences for hydrocarbon maturity. Lithospheric folding is coupled to compressional basin and fault reactivation and, therefore, strongly affects reservoir characteristics of sedimentary basins.

35 INTRODUCTION

Over the last few years much progress has been made in understanding the mechanisms of the formation of sedimentary basins (Cloetingh & Ziegler, 2007; Roure et al., 2010). Quantitative models for several basin types have significantly advanced the understanding of extensional (McKenzie, 1978; Van Wees et al., 2009) and foreland basins (Beaumont, 1981; Naylor & Sinclair, 2008). The mechanisms of basin formation as a result of lithospheric folding have received considerably less attention (Cloetingh & Ziegler, 2007) except for some clear-cut cases, in particular in central Asia (Burov & Molnar, 1998; Thomas et al., 1999a,b). Many data exist on the geometry of sedimentary basins affected by large-scale compressional intraplate deformation (Cobbold et al., 1993; Lefort & Agarwal, 1996, 2000, 2002; Burov & Molnar, 1998). These observations and results of analytical, numerical (Martinod & Davy, 1992; Burov et al., 1993; Gerbault et al., 1998; Cloetingh et al., 1999) and analogue modelling (Martinod & Davy, 1994; Sokoutis et al., 2005) demonstrate that the thermo-mechanical age of the lithosphere exerts a prime control on the wavelength of lithospheric folds. These studies focused on the role of tectonic stress in lithospheric folding

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(Stephenson et al., 1990; Burov & Molnar, 1998; Stephenson & Cloetingh, 1991; Perez-Gussinye & Watts, 2005). However, surface processes also play a significant role in the mechanics of lithospheric folding. Erosion enhances the development of folding and has a pronounced effect on its wavelengths (Cloetingh et al., 1999; Burov & Toussaint, 2007), specifically in the short-wavelength domain. In addition, both sedimentation and erosion are likely to significantly prolong the lifetime of folding. Sedimentation decreases the effect of gravity by filling the downward flexed basins and thus reducing the isostatic restoring force, whereas erosion of the upward flexed basement unloads the lithosphere on the basins' uplifted flanks. Lithospheric folding has significant effects for the geometry of sedimentary sequences deposited on folded lithosphere. However, the effect has been studied little for several reasons (Cobbold et al., 1993). The hanging walls of emerging low-angle thrust faults tend to collapse rapidly, blurring the structural relationships at basin edges. Further, ongoing sedimentation may onlap and bury the hanging walls, whereas scarp erosion may destroy them. Seismic images of basin edges may not be well resolved where footwall conglomerates contain few reflecting horizons or when low-angle thrusts reflect and refract seismic waves, thus hampering the imaging of deeper parts of the basin. Downwarping of folded basins renders surface geological mapping difficult. At the same time, high plateaux are not at first sight linked with basins of high hydrocarbon potential, leaving them in many places without well control or seismic constraints on basin architecture.

In this paper we review evidence from a number of basins in Eurasia, where folding has been documented and alternative basin formation mechanisms such as lithospheric extension or foreland flexure cannot fully explain key features of basin geometry and evolution. We restrict ourselves to cases where independent evidence points to a major role for lithospheric folding in the basin development. We demonstrate that an interpretation in terms of folding can sometimes resolve inconsistencies resulting from interpreting basins origin via other mechanisms. For example (Ritzmann & Faleide, 2009), the formation mechanism of intracratonic basins is still a matter of debate. A striking feature of these basins is the prolonged intervals of low rate subsidence

alternating with episodic accelerations in subsidence rates, often coeval with orogenic activity at plate boundaries. Due to the frequent absence of evidence for thinned lithosphere below these basins, alternative models have been proposed. Stel et al. (1993) propose that non-extensional crustal thinning is induced by basaltic underplating, which causes thermal uplift and erosional thinning. Tectonic loading at plate boundaries is also proposed as a cause of or contributor to large-scale sag-type subsidence in plate interiors (Cloetingh, 1988; Leighton & Kolata, 1990; Quinlan, 1987) as well as phase transformations (Artyushkov, 2007). Ritzmann & Faleide (2009) point out that intraplate stresses and crustal inhomogeneities coupled with loading scenarios provide the best explanation for the Barents Sea Basin and other intracratonic basin evolution. The basins reviewed here, however, do not include cases such as the Barents Sea where the effect of stresses on topography and accommodation space has been considerably complicated by other basin formation processes.

We present inferences from thermo-mechanical models of lithospheric folding and a comparison with the results of analogue modelling of folding. We concentrate on the overall geometry of basins formed by lithospheric folding, the typical timescales intrinsic to their development and the brittle deformation patterns in the underlying lithosphere. We then discuss subsidence patterns and heat flow characteristics for such basins. We conclude by comparing key features of basin accommodation shape, vertical motions, thermal history and brittle deformation patterns of basins formed on folded lithosphere, foreland basins, intracratonic basins, extensional basins and pull-apart basins.

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97 LITHOSPHERIC FOLDING: AN IMPORTANT MODE OF INTRAPLATE BASIN FORMATION AND 98 DEFORMATION

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Folding of the lithosphere, involving positive and negative deflections (**Fig. 1**), appears to be of more importance in the large-scale deformation of intraplate domains than hitherto realized

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(Cloetingh et al., 1999). As has been shown (Burov et al., 1993; Nikishin et al., 1993; Cloetingh et al., 1999; Gerbault et al., 1998; Schmalholtz & Podladchikov, 2000), visco-elastic folding starts to develop from the onset of compression and in contrast to elastic folding, does not require specifically large intraplate stresses. Folding may continue until the back-ground strain-rate drops, for example due to the localization of deformation in a specific area or to reduction of far-field forces. Such intraplate deformation results from transmission of intraplate stress fields away from plate boundaries into continental forelands (Ziegler et al., 1998; Van der Pluijm et al., 1997; Tesauro et al., 2005).

In continental lithosphere, several wavelengths of folding are expected to develop as result of the presence of several rheologically competent layers decoupled from each other by weak layers (Fig. 1, Fig. 3). Crustal folding can be traced from the surface topography, neotectonic movements and free air gravity. The crust-mantle boundary (Moho) topography reflects mantle folding, because the mantle strength is highest at this interface and thus the geometry of Moho deflection is most likely to correspond to that of mantle folding. This wavelength can be traced from Bouguer gravity anomalies because the crust-mantle boundary represents also the main density contrast in the lithosphere. It is to be expected that the base of the Mechanical Lithosphere (MLB) has roughly the same geometry as the Moho deflection. Likely, the lithosphere-asthenosphere boundary (LAB) has little to do with mantle lithosphere folding except some specific cases, as the MLB (roughly corresponding to 700°C depth) is found well above the LAB (1330°C). Consequently, the low viscosity zone between MLB and LAB may damp the deformation in a way that the LAB remains even flat. Due to the low viscosity of gravitating mantle between the LAB and MLB, the LAB is not necessarily advected or down-warped with deflection of the MLB, or at least the relations between the deformation of the MLB and LAB are not straightforward. In particular, gravity instabilities or convective movements not related to folding can perturb the LAB without impact on the MLB.

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Folding has important implications for vertical motions, sedimentary basin architecture and the evolution of hydrocarbon systems (Ziegler et al., 1995; 1998). The large wavelength of vertical motions associated with lithospheric folding necessitates the integration of data from large areas (Elfrink, 2001; Allen & Davies, 2007), often beyond the scope of regional structural and geophysical studies that target specific structural provinces. Recent studies on the North German Basin show neotectonic reactivation by lithospheric folding (Marotta et al., 2000). Similarly, acceleration of the Plio-Pleistocene subsidence in the North Sea Basin is attributed to stress-induced buckling of its lithosphere (Van Wees & Cloetingh, 1996). Folding of the Variscan lithosphere has been documented for Brittany (Bonnet et al., 2000; Lagarde et al., 2000), the adjacent Paris Basin (Lefort & Agarwal, 1996) and the Vosges-Black Forest arch (Dèzes et al., 2004; Ziegler & Dèzes, 2007; Bourgeois et al., 2007). Lithospheric folding, therefore, appears to be a effective mechanism for the propagation of tectonic deformation from active plate boundaries far into intraplate domains (e.g. Stephenson & Cloetingh, 1991; Burov et al., 1993; Ziegler et al., 1995, 1998).

Folding can be observed at different spatial scales. At the scale of a micro-continent that was affected by a succession of collisional events, Iberia illustrates lithospheric folding and interplay between neotectonics and surface processes (Cloetingh et al., 2002). An important factor favouring a lithosphere-folding scenario for Iberia is the compatibility of the thermo-tectonic age of its lithosphere and the wavelength of observed deformations.

Other well-documented examples of continental lithospheric folding occur in other cratons (Fig. 2). Prominent examples occur in the Western Goby area, in the Ferghana basin, but also probably in the Tadjik basin of Central Asia, involving a lithosphere with thermo-tectonic ages from 150 Ma (Kazakh shield, Ferghana, Tadjik basins and likely Dzhungaria basin) to 400 Ma (Tarim basin). In this area, mantle and crustal wavelengths are 360 km and 50 km, respectively, with a shortening rate of ~10-20 mm/yr and 200-250 km of shortening during 10-15 Myr (Burov et al., 1993; Burov & Molnar, 1998). Three-dimensional fold structures in the Tibetan Plateau have been inferred from

152 GRACE satellite gravity data, demonstrating prevailing fold wavelengths of 300 to 420 km (Shin *et*153 *al*, 2009).

The inferred wavelength of these neotectonic lithosphere folds is consistent with the general relationship between the wavelength of lithospheric folds and the thermo-tectonic age of the lithosphere shown by a global inventory of lithospheric folds (**Fig. 3**) (Cloetingh & Burov, 1996; Cloetingh *et al.*, 2005). In some other areas of continental lithosphere folding, smaller wavelength crustal folds have also been detected (Burov *et al.*, 1993; Nikishin *et al.*, 1993). Thermal thinning of the mantle lithosphere, often associated with volcanism and doming, enhances lithospheric folding and appears to affect the wavelengths of folds (Burov & Cloetingh, 2009).

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OBSERVATIONS

Basins developed on folded continental lithosphere have characteristic features. We illustrate this by largely concentrating on the basins of Central Asia, the folded lithosphere of Iberia, the NW European platform, the Pannonian Basin and the South Caspian Basin.

167 Compressional basins of Central Asia (Tienshan, Ferghana)

The young Ferghana and Tadjik basins are compressional basins northwest and north of the Pamir and south of the Tien Shan ranges (Burov & Molnar, 1998). In this area lithosphere underwent Jurassic reactivation and is characterized by relatively young thermo-mechanical ages (175 Ma) (Burg *et al.*, 1994; Burov & Molnar, 1998). In addition to having undergone thermal weakening, the lithosphere underlying these basins also probably has a weak, quartz-dominated lower crustal rheology. This has resulted in low values for the effective elastic plate thickness (EET) of the order of 15 km (Burov & Molnar, 1998).

The Central Tien Shan consists of an alternation of ranges and basins separated by reverse faults.
 The Tien Shan terminates west of the right-lateral Talas-Ferghana fault by splaying into two narrow

mountain chains that surround the Ferghana Valley. A narrow belt of mountains parallel to the central section of the fault slopes downward to the Ferghana Valley, underlain by a deep basin with as much as 8 km of Cretaceous-Cenozoic sediments (Cobbold et al., 1993). Gravity data suggest that the Ferghana and Tadjik basins are gravitationally overcompensated. The negative Bouquer gravity anomalies indicate a Moho several km deeper than predicted by models of local Airy isostasy (Burov & Molnar, 1998). This points to Moho down-warping due to non-isostatic processes and most probably, in this compressional context, to lithospheric folding. As demonstrated by numerical-thermomechanical experiments (Burov & Molnar, 1998), the approximately north-south shortening of the relatively thin lithosphere could have created central down-warping and anticline mountains north and south of the basin. According to this model, this shortening has produced a folding instability in the mantle lithosphere that has warped the basement immediately surrounding the basin upward, and has forced the basin floor down beneath the Ferghana Valley (with an estimate for the folding wavelength of the order of 200-250 km), possibly also producing deep mantle faulting (seen as Moho offsets in seismic reflection data).

Pre-existing thermal structure and variations in crustal thickness have played a major control on the styles and distribution of faulting in this region (Fig. 4), (see also Burov & Molnar 1998; Cobbold et al., 1993). Primary faults appear before the folding develops with a spacing that is proportional to brittle layer thickness. Subsequently the two processes, faulting and folding, co-exist in such a way that faults localize at the inflection points of folds. At this stage, the appearance of faults does not significantly influences the wavelength of folding. Because of the weakness of the lower crust, the upper crust should be completely decoupled from the mantle and interact with it only by flow in the lower crust. Lower crustal flow changes the wavelength and amplitude of the surface folding, which is terminated by the development of a single down-warped megafold (Cloetingh et al., 1999). Liu et al. (2007) reported evidence for lithospheric folding in the Central/ North China Tarim Basin region, initiated during Pliocene times, with good time constraints provided by Ar/Ar dating. Two wavelengths are found of 30 and 400 km. The thermo-mechanical age of the lithosphere is

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estimated to be 250 Ma, as the main thermal perturbation was associated with a major orogenic phase at 250 Ma ago. These findings are consistent with models in which decoupled lithosphere folds with 30 km and 400 km wavelength correspond to, respectively, crustal and mantle folding. Deep seismic data for basins north of and within the Tibetan plateau (Li *et al.*, 2006; Liu *et al.*, 2006; Zhao *et al.*, 2006) show that folding is not limited to the Tibetan plateau (see also Shin et al, 2009).

209 Folded lithosphere of Iberia

10 As reviewed by Cloetingh et al. (2002), Tertiary lithospheric folding of Iberia occurs in Variscan 11 lithosphere with wavelengths of 300 km, leading to the development of a system of parallel trending 12 basins and highs (Fig. 5a). This folding generated the distribution of basins and mountain chains, 13 bounded by folds and faults (Cloetingh et al., 2002; De Vicente et al., 2007). The regularity of the 14 fluvial network pattern of the central-western part of Iberia also indicates large-scale lithospheric 15 folding. Alpine (mainly Cantabrian-Pyrenean related) compressional tectonics was the principal 16 factor for closing internally drained sedimentary basins, including the Ebro, Duero and Tagus basins 17 (Casas-Sainz & De Vicente, 2009).

18 A large body of geophysical and geological observations is available on the crustal geometries and 19 stress regimes of Iberia (De Vicente et al., 2007, 2008). This is supplemented by thermo-20 geochronology data (De Bruijne & Andriessen, 2002; Ter Voorde et al., 2004), demonstrating 21 accelerated uplift of the highs during Late Miocene-Pliocene times. In spite of the existence of 22 recent localized volcanism in the inner part of the Iberian plate, an important contribution from 23 thermal uplift is questionable due to the distribution of crustal thickness (Casas-Sainz & De Vicente, 24 2009). Fernandez-Lozano et al. (2010) have constructed a three-layer analogue model consisting of 25 brittle upper crust, ductile lower crust and ductile upper mantle for Iberia. These models show that 26 folding is associated with deformation of narrow mountain ranges separated by basins (see Fig. 27 **5b)**. The narrow mountain ranges representing upper crustal pop-ups form the main topographic 28 reliefs. Shortening is accommodated within the viscous crust beneath the pop-ups leading to lateral

thickness variations of the ductile crust. These results are consistent with seismic and gravity data
collected in the Cantabrian mountains, the Spanish Central System, and the Toledo Mountains (see
also Fig. 5a).

The earliest sediments in the terrestrial basins (Calvo et al., 1993) were deposited during the Eocene, with very well developed continuous successions, dominated by continental facies with high sedimentation rates. The Duero and Tagus basins began to be individualized at the onset of uplift and the formation of a topographic range in the Central System (Casas-Sainz & De Vicente, 2009). In contrast with this early more homogeneous stage, the Oligocene-Lower Miocene stratigraphy of the basins is characterized by several tectono-stratigraphic units, separated by unconformities representing short periods of time. The number and specific ages of these units varies between basins, indicating variable activity at basin borders. The fact that the largest number of breaks in the sedimentary succession is concentrated during this time interval points to a tectonic control (De Vicente & Vegas, 2009).

The final stages in the evolution of these onshore basins are characterized by widespread Miocene lacustrine sedimentation in the endorheic basins and during Pliocene and Quaternary times, by the beginning of river capture and exorheism in the Duero, Ebro and Tagus basins (Garcia-Castellanos et al., 2003). Alluvial deposits were deposited by the fluvial network developed on the ancient, internally drained basins (Casas-Sainz & De Vicente, 2009). During the endorheic-exorheic transition, paleo-karstification processes and erosion occurred in the Late Miocene lacustrine limestones. Two main types of high plains can be distinguished on the Iberian Peninsula. The coupled origin of both types of surfaces can be related to uplift/sedimentation histories along the Tertiary compressional basin margins in the Iberian Peninsula (Casas-Sainz & De Vicente, 2009). The first are high plains with Tertiary sedimentary basins, containing well preserved Late Miocene to Pliocene strata with subhorizontal bedding. Typical examples are the Duero and Tagus basins. The second consists of eroded sediments of Mesosoic and Paleozoic rocks with planation surfaces developed on the Paleogene-Lower Miocene ranges created by fold and thrust systems and

basement uplifts. Different patterns in evolution and stratigraphy between the eastern and western Tagus basin may be related to the mechanical decoupling between Iberia and Africa during the Miocene emplacement of the Alboran Domain in the southern part of the plate. This generated extensional stresses in the eastern Iberian plate, coeval with compression and lithospheric folding in the western part of the plate (De Vicente *et al.*, 2008; De Vicente & Vegas, 2009).

In general the dimensions of peneplains linked to extensional basins and passive margins are larger, and their evolution involves time scales much longer than those characteristic for high plains of the Iberian plate (Casas-Sainz & De Vicente, 2009). Peneplains associated with compressional or convergent regimes (e.g. Anatolia, Central Iran and Morocco) are characterized by spatial dimensions comparable with Iberia. In addition, their tectonic setting in the hanging wall of the African-Eurasian convergence zone is also similar. Thus the Iberian plate may be an analogue for other present-day intra-mountain basins.

The European Alpine foreland

Geophysical and geomorphological studies (e.g. Lefort & Agarwal, 1996; 2000; 2002; Marotta et al., 2000) indicate large-scale compressional deformation of the European Alpine foreland lithosphere. Although different in structural grain and rheological structure (Tesauro et al., 2007; Cloetingh et al., 2005) inherited from differences in their Paleozoic-Mesozoic history, different segments of the lithosphere in the northern Alpine European foreland share many similarities in their Cenozoic intraplate deformation. Seismic and geological evidence shows that the North German Basin cannot be explained by classical models of basin formation (Marotta et al., 2000). The basin underwent a polyphase history (Scheck & Bayer, 1999; Littke et al., 2008) of extension in the late Triassic followed by lithospheric scale folding in the late Cretaceous-Early Cenozoic (Mazur et al., 2005; Marotta et al., 2000). 3D volumetric analysis shows that the compressional event was followed by faster subsidence during the Cenozoic (Scheck & Bayer, 1999). A similar observation was made by Van Wees & Cloetingh (1996) for the North Sea Basin. Lithospheric folding modified a basin formed

initially by deep seated thermal perturbations (Scheck & Bayer, 1999) with a thinned crust and
shallow Moho. The Moho updoming with a high of 3 km for the height of the bulge in the southern
part of the North German basin appears to be the consequence of flexural buckling of the previously
thinned lithosphere by a compressive stress perpendicular to the strike of the basin.

From earliest Triassic to Late Jurassic, the Paris Basin (Fig. 6a) subsided in an extensional framework and was larger than the present basin (Guillocheau et al., 2000). The western margin of the Paris Basin and the rifted Atlantic margin of France were subject to thermal rejuvenation during Mesozoic extension related to North Atlantic rifting (Ziegler & Dèzes, 2006; Robin et al., 2003). Subsequent compressional intraplate deformation (Ziegler et al., 1995) also affected the Paris Basin (Lefort & Agarwal, 1996). Numerical modelling indicates that the amplitude of the compression induced vertical deflection in the Paris Basin is relatively small, requiring additional mechanisms such as flexure due to the load by the Alpine system (Bourgeois et al., 2007; Ziegler & Dèzes, 2007). However, periodic deformation at the same wavelength occurs northwest of the Paris Basin (Fig. 6 a, Bourgeois et al., 2007), which would not be the case if the north-eastern part of the basin was flexed down by the load of the Alps. Bourgeois et al. (2007) carefully analysed basement geometry and timing of vertical motions in the NW European platform. They separate the contributions from the European Cenozoic Rift System (ECRIS) and the contributions from long-wavelength folding striking NE and located between the Alpine front and the North Sea. According to their analysis, the ECRIS developed mostly between 37 and 17 Ma, whereas lithospheric folds developed between 17 Ma and present, with a wavelength of 270 km and amplitude of 1500 m (see Fig. 6b).

Quaternary folding of the Variscan lithosphere in the area of the Armorican Massif (Bonnet *et al.*, deformation. The timing and spatial pattern of uplift inferred from river incision studies in Brittany is incompatible with a glacio-eustatic origin. Therefore, Bonnet *et al.* (2000) related the vertical motions to deflection of the lithosphere by the present-day NW-SE directed compressional intraplate 1

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stress field. Stress-induced uplift of the area appears to control fluvial incision rates and the position
 of the main drainage divides. Leveling studies (LeNotre *et al.*, 1999) indicate its ongoing
 deformation.

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2 311 Pannonian Basin System

Recent studies of the Pannonian Basin system - formed as a Miocene back-arc basin behind the Carpathian arc - show that active crustal deformation resulted in significant differential vertical motions during Late Neogene times (Horvath & Cloetingh, 1996; Cloetingh *et al.*, 2005, 2006; Horvath *et al.*, 2006; Matenco *et al.*, 2007; **Fig. 7a**). The river network in the area is affected by differential neotectonic crustal motions (Necea *et al.*, 2005; Dombradi *et al.*, 2010).

317 The Late Neogene evolution of the Pannonian Basin is largely controlled by its interaction with the 318 Carpathian arc and the Adriatic indentor impinging the Pannonian basin lithosphere (e.g. Bada et al, 319 2001). As a result of the consumption of subductable lithosphere of the European foreland, the 320 Pannonian basin became locked. At the same time, the continuous N-NE directed indentation of the 321 Adriatic microplate built up a high level of compressional stresses in the Pannonian Basin. The 322 structural inversion of the basin is a direct consequence of the temporal changes in the stress field 323 from extension to compression. Recent studies of the stress field together with an extensive array of 324 geological and geophysical data have shown that horizontal stresses are transmitted from the plate 325 boundaries into the interior of the Pannonian Basin (Bada et al., 2007).

The Pannonian Basin is underlain by what is probably the hottest lithosphere of continental Europe, characterized by a very low lithospheric strength (Tesauro et al., 2007). Miocene extension resulted in a high level of mantle stretching, leading to a strength distribution with practically zero strength in the subcrustal lithosphere and lower crust (Cloetingh *et al.*, 2005), making it prone to tectonic reactivation. Therefore, in this area relatively low compressional stresses induced mainly by plate boundary interactions, and further amplified by topography-induced stresses (Bada *et al.*, 2001), could initiate significant intraplate deformation.

It should be noted that in the Pannonian basin system, seismic tomography (Wortel & Spakman, 2000) supports slab detachment under the adjacent Romanian Carpathians. At the same time, the seismic tomography shows the presence of a hot upper mantle below the folded lithosphere in the Pannonian basin. In this area, therefore, the upper mantle structure reflects more the back-arc extension prior to the folding than the consequences of folding on the density stratification of the mantle. The reverse is true for the crustal configuration. The geometries imaged by the crustal scale deep seismic reflection profiling are clearly dominated by compressional structures resulting from folding (**Fig. 7b**).

Quaternary patterns of differential uplift and subsidence with a timing coeval with an increased level of compressional stresses are interpreted as caused by lithospheric folding (Horvath & Cloetingh, 1996; Cloetingh et al., 2006; Horvath et al., 2006; Bada et al., 2007). Late-stage compressional deformation of the Pennonian basin is demonstrated to be of key importance for the geometry of basin fill (Sacchi et al., 1999, see Fig. 7b) and for its reservoir structures, both crucial in assessment of its hydrocarbon potential (Horváth, 1995). Analogue tectonic experiments by Dombradi et al. (2010) further examine accumulation and transmission of stresses in the hot and weak back-arc lithosphere of the Pannonian basin and the interplay between lithospheric processes and the intrabasinal topography development. Based on available rheological and seismotectonic constraints, the basin system is represented by a two-layer model with a 10-15 km thick brittle upper crust. The lower crust and the mantle are represented by a low-strength layer of 45 km thickness (Dombradi et al., 2010). The strain rates applied were constrained by GPS velocities (Grenerczy et al., 2005). In the experiments lithospheric folds develop with wavelengths equivalent to 200-250 km (Fig. 7c). The experiments also show that crustal thickness variations largely determine the localisation of stresses. Topographic profiles extracted by laser scans of the model show that surface topography reflects long wavelength buckling of the lithosphere. These results are compatible with Pliocene-Quaternary subsidence, which occurs in the centre of the Pannonian basin, simultaneously with accelerated uplift at the basin margins. The latter has lead to elevated

topography of up to 1 km in areas at the rim of the Pannonian Basin such as the Styrian, Zala and Transylvanian basins (Cloetingh et al., 2006).

The South Caspian Basin

The South Caspian Basin with an estimated sedimentary thickness up to 20 km (Brunet et al., 2003; Egan et al., 2009), is probably one the deepest basins in the world. It is flanked by the Alborz Mountains to the south, with an elevation of 2-4 km (see Fig. 8). The transition between basin and flanking high is abrupt and coincides with a coastline bounded by a major fault system. The formation of the South Caspian Basin is enigmatic. Attempts to explain it as an extensional basin or as a foreland basin have been unsuccessful. There is a lack of evidence for basin extension and both stretching models and topographic loads and slab pull operating on down flexed lithosphere in foreland basin systems cannot explain differential vertical motions of this order of magnitude. As a result, phase changes in continental crust have been proposed as a mechanism for the formation of this superdeep basin (Artyushkov, 2007). A limitation in conclusively resolving the basin formation mechanism is the lack of multi-channel seismic reflection data capable of imaging crustal structure below the thick sedimentary sequences. Over the last few years, however, other constraints on the area have become available (e.g. Guest et al., 2007).

Figure 8 presents a cross section of northern Iran and the South Caspian Basin (Guest et al., 2007). As shown, wavelengths for compressional folding of the lithosphere are typically in the range of 400 km. Although the actual basin formation mechanism for the Caspian Basin is not well resolved (e.g. Artyushkov, 2007), late Neogene folding has been affecting a lithosphere probably thermally reset by middle Late Jurassic marginal basin formation (Guest et al., 2007) with a thermo-mechanical age at the onset of collision of 130-150 Ma. Satelite data demonstrate an exceptionally large gravity anomaly over the basin (Kaban et al., 2002). Seismic tomography shows slab detachment under the Iranian plateau (Hafkenscheid et al., 2006; Alinaghi et al., 2007) beginning at about 10-15 Ma.

Basin analysis has shown that basement inversion has occurred 20 Ma ago, simultaneously with the slowing of Arabian/Eurasian plate convergence and the onset of accumulation of Neogene clastics in foreland basins (e.g. Fakhari et al., 2008). By 10 Ma, contraction occurred by underplating of the Arabian crustal units beneath the Iranian plate. Estimated shortening rates from current geodetic surveys (Vernant et al., 2004) are 7 mm/yr. Results from geothermochronology demonstrate rapid uplift of the Alborz mountains, with rates of 0.7 km/my exhumation between 6 and 4 Ma (Axen et al., 2001), implying approximately 10 km of uplift. This uplift was nearly synchronous with rapid South Caspian subsidence (Nadirov et al., 1997) and subsequent folding (Devlin et al., 1999). South Caspian sedimentation rates locally increased more than tenfold at ca 6 Ma, with more than 10 km of sediments deposited since then. Axen et al. (2001) argue that if approximately 10 km of post 6 Ma sediments are present in this basin, then as much as 20 km (equivalent to 80 %) of the structural relief of about 25 km between the high Alborz and the southernmost Caspian basement may be younger than 6 Ma.

Folds in the southernmost Caspian Basin (Devlin et al., 1999) and in the Neogene of the Northern Alborz foothills imply contraction. The lack of a crustal root under the Alborz Mountains also points to a flexural support by the South Caspian basement (Axen et al., 2001). The simultaneous reversal of Zagros strike slip and extrusion of central Iran, coarse Zagros molasse deposition, Dead Sea transform reorganisation, Red Sea oceanic spreading (Chu & Gordon, 1998) all initiated around 5 ± 2 Ma (e.g. Axen et al., 2001; Smit et al., 2010), was interpreted by these authors in terms of a widespread tectonic event. Slab detachment could explain part of the recent uplift of Iran, but does not directly explain the simultaneous dramatic subsidence in the adjacent Caspian Basin. However, slab detachment could have an indirect effect if reducing slab pull forces acting on the downgoing lithosphere promote the development of compressional stresses in both the downgoing and overriding plates (Cloetingh et al., 2004; Matenco et al., 2007).

Guest et al. (2007) argue in favour of compressional deformation of the South Caspian Basin/Alborz mountains. This interpretation was largely based on available industrial seismic reflection data and

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59 60 411 gravity data for the basin and followed a similar proposition made for the late-stage deformation of 412 the adjacent Black Sea (Nikishin et al., 2003; Cloetingh et al., 2008). Guest et al. (2007) present 413 kinematic models, without addressing the reason for the difference in the order of magnitude of the 414 basin depression and the adjacent high. Analogue experiments (Sokoutis et al. 2005) show that 415 when compression acts on two blocks with contrasting thickness (and rigidity), a major syncline will 416 develop on top of the suture zone separating these blocks, flanked by an anticline of much lower 417 amplitude. These results are consistent with predictions from folding theory that as a result of the 418 acting gravity field, down-warping is mechanically more effective than up-warping against the action 419 of gravity forces. The presented numerical experiments illustrate this (see for example Fig. 9).

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421 Basins on folded lithosphere in Eurasia: differences and similarities in lithospheric 422 deformation and basin habitat

423 Comparison of lithospheric deformation in Central Asia, Iberia, the NW European Alpine foreland, 424 the Pannonian Basin and the South Caspian Basin demonstrates common features compatible with 425 an interpretation in terms of basins developed on folded lithosphere. Amongst these are their 426 symmetrical shape, dimensions and the temporal association of their basin formation history with 427 the build up of intraplate stresses. Important differences can be observed between the Central 428 Asian and Iberian basins and the South Caspian Basin on one side and the European Alpine 429 foreland and the Pannonian Basin on the other side.

430 The Central Asian Basins and the basins of the Iberian plate were initiated by Tertiary folding, 431 whereas the European Alpine foreland underwent a polyphase history, following Carboniferous 432 and Triassic stretching (Scheck & Bayer, 1999; Le Solleuz et al., 2004), prior to Late Cretaceous-433 Early Cenozoic folding. Similarly, the Pannonian Basin, formed by back-arc extension in Miocene 434 times, was subsequently compressionally reactivated (Cloetingh et al., 2008). The origin of the 435 South Caspian Basin is not well resolved, but km scale vertical motions are compatible with a 57 436 megafold as a prime control on present-day basin geometry. These differences in evolution have a 58

pronounced impact on basin dimensions and on the nature of basin fill. The Central Asian basins, the folded basins of Iberia, and the South Caspian Basin have a wavelength compatible with theoretical predictions. In contrast, the fold wavelengths of the European Alpine foreland and the Pannonian Basin appear to be primarily controlled by the pre-existing geometry of the sediment-basement interface and pre-existing Moho topography inherited from the pre-folding stage. Differences occur also in the development of the ranges flanking the basins. In Central Asia and Iberia, these highs have amplitudes and widths comparable to the prediction for periodic folding, whereas in the Paris Basin and the North German Basin these highs are less well developed. The ratio of the depth of the South Caspian Basin to the height of the flanking Alborz Mountains is compatible with a megafold under this area of dramatic differential topography. The large amplitudes of folding at the flanks of the Pannonian Basin system are compatible with its extremely weak rheology. Differences in the amplitude of the highs might also partly result from the differences in neotectonic shortening. The highs bounding the basins of Iberia, Pannonian Basin, South Caspian Basin and Central Asia, located relatively close to the site of ongoing convergence, are still experiencing active uplift. This is much less the case for the low strain deforming areas of the NW European platform and the North German basin where the peak compression occurred during Late Cretaceous-Early Cenozoic. Another difference occurs in the depositional regime prevailing in these basins. The closed basin systems of Central Asia, Iberia and the intra-Carpathian Pannonian Basin are dominated by continental and lacustrine deposits, whereas capture is usually a late-stage feature of these basins. The Paris basin of the NW European platform has a basin fill largely dominated by marine sediments deposited during the pre-folding stage of their evolution, followed by marine Tertiary successions in the basin centre with truncated basin margins. The creation of basin highs in this case was in general not capable of interrupting open connections to marine environments.

461 A general observation for all these basin systems is the flatness of sedimentary sequences in the 462 depocentres. Faulting plays a minor role in the basin development. Faulting appears to be Page 19 of 91

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59 60 463 concentrated at the basin margins, with a highly variable depth extent, varying from upper crustal 464 scale faults bounding pop-up structures in the Central System of Iberia and the Pannonian Basin to 465 minor faulting in the basement of the North German Basin. The thermal regimes of the basins are 466 similar with thermal histories strongly affected by subsidence and sediment deposition on down 467 flexed lithosphere, pre-conditioned by its thermo-mechanical age.

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THE ROLE OF LITHOSPHERIC RHEOLOGY

470 Folding is commonly associated with periodic deformation of layered structures with contrasting 471 mechanical properties. As a consequence, the harmonic spatial periodicity is often believed to be its 472 primary, almost synonymous, identifying factor. However, from a mechanical point of view, constant 473 wavelength is not a necessary requirement. Folding is a compressional instability developing in stiff 474 layers embedded in a weaker material. The harmonic solution of the equilibrium and conservation 475 equations is a possible solution (Muehlhaus et al., 1998; Burov & Molnar, 1998; Cloetingh et al., 476 1999). The physical mechanism of folding is well understood. In a continuous layered medium the 477 stresses and strains must be continued across the interfaces between the layers. In multi-layer 478 systems with contrasting mechanical properties (e.g. strong and weak layers) this requirement may 479 be difficult to satisfy, because the same amount of shortening in a stiffer layer would require much 480 larger stress than in the neighbouring weaker layer. The system becomes unstable and, in 481 attempting to reduce the stress and strain unconformities at the interfaces between the layers, starts 482 to fold (or buckle) in response to very small perturbations. In non-elastic media with strain-483 dependent properties, locally increased flexural strain at the fold limbs can create weakened zones 484 that significantly facilitate further deformation. These weak or softened plastic or viscous zones are 485 often referred to as inelastic hinges, since the system easily folds at such weakened zones. 486 Homogeneous shortening of the lithosphere under horizontal compression requires more work than 487 required by an equivalent amount of shortening by folding. However, shortening bv 57 488 underthrusting/subduction of the lithosphere may be more mechanically efficient than folding but 58

may be blocked or unable to start immediately after the onset of shortening. For this reason, folding is likely to be a primary and common response to tectonic compression. Folding probably may continue, in an attenuated form, even after the beginning of subduction (Cloetingh et al., 1999), or re-appear when the subduction channel is locked or during the aftermath of collision (Cloetingh et al., 2004; Matenco et al., 2007). The rheological structure of the lithosphere (Watts & Burov, 2003) is a key factor in lithospheric folding. Compressional stresses must build up to a level comparable with the integrated strength of Schubert, 2002).

the lithosphere in order to induce folding. Prior to the advent of rheological models based on extrapolation of laboratory experiments (Goetze & Evans, 1979; Carter & Tsenn, 1987) lithospheric stresses were thought to be incapable of reaching the failure levels required for folding (Turcotte & The simplest expression of the folding process is in oceanic lithosphere, characterized by an absence of a rheological stratification, and with thermo-mechanical ages spanning a limited age window with a maximum of 200 Myr (Geller et al., 1993; Zitellini et al., 2010). Because weak lithosphere is easier to fold than strong lithosphere, one would expect the first observations of lithospheric folding from the rheologically weak continental plate interiors. However, the observation of folding of the relatively strong lithosphere of the central Indian Ocean with a thermo-mechanical age of 80 Ma and with wavelengths of 200-250 km (Geller et al., 1983; McAdoo & Sandwell, 1985; Stein et al., 1989; Bull & Scrutton, 1992) triggered the interest in this mode of deformation. There may be two reasons for this. First, the recognition of folding, by mapping of basin reflectors and gravity anomalies requires spatial scales of several hundreds of kilometres. This is feasible with marine geophysical surveys, but generally in excess of the spatial scales covered by land surveys. An exception has been the systematic mapping of basins in Central Australia, located on strong lithosphere with a thermo-mechanical age of approximately 700 Ma (Lambeck, 1983; Stephenson & Lambeck, 1985). The Indo-Australian plate is under an exceptionally high level of compression (Stein et al., 1989), as a result of its collision with the Eurasian plate. Under these conditions high

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515 stresses might have been more important than the rheology of the plate interior (Beekman et al., 516 1996; Gerbault et al., 1998; Gerbault, 2000).

517 Secondly, folding in oceanic lithosphere is not affected by surface erosion, making it relatively easy 518 to recognize. The presence of only one dominant folding wavelength characteristic of non-stratified 519 oceanic lithosphere also facilitated a quantitative interpretation. Following the early studies in the 520 Indian Ocean and in Central Australia in the rear of the Indian-Eurasian collision, attention shifted to 521 the other side of the plate contact in Central Asia. Large scale folding patterns were recognized 522 through analysis of geophysical data, including gravity and topography (Burov et al., 1993; Nikishin 523 et al., 1993; Burov & Molnar, 1998) as well as detailed geological studies (Cobbold et al., 1993).

524 On the basis of an inventory of currently available results of studies of folded continental 525 lithosphere, it appears that in general the observed wavelengths of folding follow theoretical 526 predictions for wavelengths as a function of thermo-mechanical age remarkably well (see Fig. 3).

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529 NUMERICAL **EXPERIMENTS** FOR SELECTED **THERMO-MECHANICAL** AGES OF 530 LITHOSPHERE

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532 Folding is an unstable deformation that may develop in stratified media with significant competence 533 contrasts, as occur in the continental lithosphere, in response to horizontal loading (Biot, 1961; 534 Ramberg, 1961). These instabilities may develop under compression or basal shear as a result of 535 strain/stress incompatibility on the interfaces between layers of different competence. In the initial 536 stages of deformation, compressional instabilities cause periodic folds with an exponential growth 537 rate and a dominant wavelength λ roughly proportional to 5 -10 times the thickness(es) of the 538 competent layer(s). In the lithosphere, the layer thickness is that of the competent core h. So, h =539 10-100 km, gives rise to $\lambda = 50-1000$ km.

(1)

"Biot's", or linear, folding encompasses the cases where compression of the lithosphere takes place from the beginning, leading to the formation of alternating (synclinal) basins and (anticlinal) highs, and where the conditions of the linear theory are more or less satisfied (e.g. thin layer approximation, no strain-softening, plane layers). For linear folding in a Newtonian medium, an asymptotic relationship derived from the thin layer equilibrium equation is (Biot, 1961; Ramberg, 1961): $\lambda_1 = 2\pi h \left(\mu_{l1} / 6 \mu_{l2} \right)^{1/3}$ Here λ_1 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 surrounding material, respectively. The above equation, derived assuming zero gravity, provides estimates of $\lambda_1/h = 20-40$ for typical competence contrasts. This does not hold for most lithospheric-scale cases, where $\lambda/h = 4 - 6$ is more common due to the influence of the gravity-dependent terms. A reduction of the normalized characteristic wavelength to values around 4 is also predicted for power-law rheology or plastic layers (Smith, 1979). For a single stiff viscous layer on top of an inviscid medium, the dominant gravity-dependent

wavelength is (Burov & Molnar, 1998):

$$\lambda_g \sim 2\pi (2h \dot{\varepsilon} \mu_{eff} / (\Delta \rho g))^{1/2}$$
(2)

where $\Delta \rho$ is the density contrast, $\dot{\varepsilon}$ is the strain rate, μ_{eff} is the layers' effective ductile viscosity (see Appendix A), *n* is the power-law exponent, and *h* is the thickness of the competent layer.

At large amounts of shortening this deformation may become a-periodic (Hunt et al., 1996) so that the wavelength and amplitude of folding vary along the deformed transect. Plume-induced periodic deformation (Burov & Cloetingh, 2009) also includes folding or boudinage caused by shear on the base of the lithosphere. However, the prime cause for folding of continental lithosphere is tectonic loading by horizontal far-field forces. For characteristic parameters of various layers within the

continental lithosphere (h = 10 - 120 km, $\mu = 10^{19} - 10^{26}$ Pa s: $\Delta \rho = 10 - 600$ kg m⁻³), equations (1-2) predict folding wavelengths that may vary from 30 km (upper crustal folding) to as high as 600-800 km. The findings of numerical models (Cloetingh et al., 1999) are in excellent agreement with analogue tectonic experiments (Sokoutis et al., 2005). Numerical experiments show that the wavelength and amplitude of folding is largely controlled by the thermo-mechanical age and less by the rate of shortening (Burov & Cloetingh, 2009). Lithospheres older than 500 Ma develop nearly the same wavelength of folding as 1000 Ma old lithosphere because of the similar nearly stationary thermal state. Here, therefore, we restrict ourselves to a set of experiments for ages of 150 and 300 Ma and convergence rates of 1.5 and 3 cm/yr.

571 Surface topography and sedimentary deposition is described through diffusion-like erosion laws and 572 fluvial transport laws (Gossman, 1976; Kirkby, 1986; Leeder, 1991; Beaumont *et al.*, 1992; Kooi and 573 Beaumont, 1994):

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dh/dt = div(k^*(x,h,\nabla h) \nabla h) (slope erosion)
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574 where h is topography, t is time, x is the horizontal coordinate, k^* is an experimentally established 575 scale-dependent coefficient of erosion (500 – 1000 m²/y). For simplicity, we used zero order linear 576 diffusion (k^* constant) for the short-range erosion and flat deposition for long-range fluvial transport 577 (Avouac & Burov, 1996). The hill-slope erosion law ensures conservation of matter but holds only 578 for small-scales. At large scales it is no longer applicable, and conservation of matter is not 579 observed, because some sediments are lost due to transport out of the system by long-range 580 processes.

Figure 9 shows the results of the experiments for relatively young (150 Ma) lithosphere compressed at rates of 1.5 and 3 cm/yr. Strongly decoupled low-amplitude short wavelength (mantle wavelength $\lambda = 250$ km) folding developed leading to crustal thickening above a synclinal mantle fold. At later stages, harmonic folding was followed by mega-folding (Cloetingh *et al.*, 1999; Burg & Podladchikov, 1999). For higher rates of shortening these mantle synclines developed in a mode

resembling symmetric subduction. For lower convergence rates (Fig. 9a) folding does not develop
 due to the low Peclet number of the system, leading to heat diffusion and weakening of folds.

Folding is well developed for medium (300 Ma) age lithosphere (Fig. 10). This case is characterized by long mantle wavelengths (λ =360 km) and high surface amplitudes (2000 m after 10 Myr). These compare well with observations from Iberia (Fernandez-Lozano et al., 2010), Western Goby and the Ferghana basin in Central Asia (Burov et al., 1993; Burov & Molnar, 1998). At late stages (10 - 26 Myr since the onset of shortening for 3 cm/y or 20- 50% of shortening), folding becomes aperiodic, leading to mega-folding (Cloetingh et al., 1999; Burg & Podladchikov, 1999) and subsequent formation of high-amplitude crustal down-warps. The amplitude of vertical movements may reach 20 km (± 10 km) or even more. As pointed out earlier, such high amplitudes of vertical motions are observed in the South Caspian Basin (Guest et al., 2007) and the Barents Sea (Ritzmann & Faleide, 2009). However, it may be relatively rare for folding to continue for periods in excess of 10 Myr. More typically is that at a certain stage deformation localizes along single major fault zones (Cloetingh et al., 1999; Gerbault et al., 1998).

As was discussed by Bird (1991), Avouac & Burov (1996) and Cloetingh *et al.* (1999), large-scale undulations of the lithosphere cannot be preserved for a long time (longer than 10 Myr) in the absence of sufficient compression, except for plates with very strong (especially lower crustal) rheology. Otherwise, the folds either will be flattened by gravity-driven crustal flow associated with the large crust-mantle density contrast at the Moho, or deformation will localize along some of the faults created at the inflection points of folds.

The fact that folding with wavelengths of 400-500 km and 500-600 km persists long after cessation of the compression in the presumably high-rigidity Australian craton and Russian platform points to a strong rheology yielding a high value for the effective elastic thickness EET (in excess of 60 km), as expected from its thermo-mechanical age (Beekman *et al.*, 1997). The long wavelengths of 400-600 km of crust-mantle down-warping in the eastern Barents Sea, where Mid-Cenozoic

compression has affected a basin system formed during Late Permian-Triassic time (Ritzmann & Faleide, 2009), with a total infill of sediments of the order of 15-20 km is another impressive example of compressional deformation on relatively cold lithosphere. Most of the areas with present-day long-wavelength continental lithosphere folding formed during or short after the Alpine collision 60 Myr ago. It thus seems that 60 Ma is a characteristic timescale of gravity collapse of the large-scale folds in such intermediate-age lithospheres. In the case of very weak guartz-dominated lower crust, folds may disappear rapidly within 8-15 Myr. Experiments (Cloetingh et al., 1999) for 1000 Ma old lithosphere, with a diabase lower crust, demonstrate that folding-induced deformation persists for 10-50 Myr following the cessation of compression. During this period the amplitude of the folds decreased by less than 10 percent. At this stage, crustal faults may accelerate the gravity collapse of folds, creating inverted basins. **BASIN GEOMETRY AND ACCOMMODATION SPACE** Basin shape Folding of the lithosphere leads to a symmetrical pattern of down-warped areas (synclines) flanked by highs (anticlines) of similar amplitude and wavelength, like for instance the Ferghana basin in

629 Central Asia (Burov & Molnar, 1998).

This symmetry is in marked contrast to the asymmetrical shape of foreland basins, formed by flexure in front of an orogenic wedge, which are flanked by a flexural bulge of an amplitude that is only up to 10 percent of the maximum depth of the foreland depression (Royden, 1988; Zoetemeijer et al., 1999). Both types of compressional basins show linear nature alignment of parallel depocentres and flanking highs. An important difference is the presence of parallel trending depocentres for lithospheric folding, whereas foreland flexural basins have a single depositional system. In addition, basins developed on folded lithosphere have a static location of the axis of their

637 depocentres, in contrast to foreland basins, where the axis of depocentres can migrate with time 638 (e.g. Zoetemeijer *et al.*, 1993).

In both cases, the integrated strength of the lithosphere (Watts and Burov, 2003; Tesauro et al., 2007) defines the characteristic width of the basins. For folding, basin width roughly corresponds to one half of the fold wavelength (λ), which equals 5 to 10 times the thickness of the strong core of the lithospheric plate, h (h = 10 to 100 km, i.e. $\lambda = 50 - 1000$ km). In the case of crust-mantle decoupling, which occurs in relatively young plates, two dominant folding wavelengths develop (typically 250 - 400 km and 50-100 km). Thus two basin populations may be observed, one imbricated within the other. In the case of flexural foreland deformation, the width of the basin is controlled by the flexural parameter and is thus of the same order as the largest fold wavelength.

Subsidence patterns

Lithospheric folding results from an instability in the lithosphere due to stress/strain incompatibilities that develop in rheologically stratified layers under compressive strain. This process typically operates with timescales of the order of 1-10 Myr (Cloetingh *et al.*, 1999). The preservation of lithospheric folds appears to depend strongly on the thermo-mechanical age of the underlying lithosphere (Cloetingh *et al.*, 1999). Folds in young lithosphere will not be preserved upon relaxation of the stress field, whereas ones in cratonic lithosphere will be well preserved (Cloetingh *et al.*, 1999).

Figure 11a displays characteristic basin subsidence patterns predicted by folding of 150 and 300 Ma old lithosphere due to a 3 cm/year shortening rate. As illustrated (**Figure 11b**), the differential motions occur in three distinct phases:

Stage 1 is the basin formation phase, coinciding with the initiation and development of folding, which is marked by an acceleration of subsidence in the basin and uplift in the flanking highs. As a result of the time lag between sediment supply and the creation of accommodation space, during the few Myr needed to form the basin, deposition cannot keep up with subsidence causing Page 27 of 91

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sediment-starved basins. A similar pattern occurs in pull-apart basins (Pitman & Golovchenko,
1983) which are also associated with ultra rapid subsidence in their formation stage also lasting only
a few Myr.

666 Stage 2 is the basin preservation stage, where equilibrium develops between sediment supply and
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 2 667 sediment deposition. As a result the basin will be rapidly filled to overfilled.

Stage 3 is the basin destruction phase, characterized by basin capture and removal of sediments to areas outside the folding system. During this phase, the size of the accommodation space is reduced and erosion occurs on both flanks and depocentre. Thus, the net effect of these three stages is that lithospheric folding will lead to the development of distinct depositional-erosional cycles.

Characteristic patterns for vertical motions for folding of continental lithosphere of 150 and 300 Ma old (Fig. 11) demonstrate a remarkably short time scale in which substantial amounts of tectonic subsidence are induced by the process of lithospheric folding. As shown by **Figure 11**, within 1 Myr after the initiation of folding induced by shortening the lithosphere at a rate of 3 cm/yr, subsidence rates are in the order of 5-15 km/Myr, depending on the thermo-mechanical age of the lithosphere. Thus, folding appears to be more effective in middle aged lithosphere of 300 Ma than in lithosphere of younger ages (Cloetingh et al., 1999; Burov & Cloetingh, 2009). After 2 Myr, a slow uplift phase starts of relatively minor magnitude of the order of a few hundreds of meters to a km. The predicted subsidence in Figure 11 is for the centre of the syncline. As pointed out above, erosion of the uplifted flanks is taken into account, adopting a diffusive equation approach to erosion (see Burov & Cloetingh, 1997; Cloetingh et al., 1999, for further details).

The alignment of parallel highs and basins of similar dimensions has an important consequence for the areal extent of the source areas for sediments available for deposition in the folded depressions. In comparison with foreland basins the distribution of sources is more symmetrical, superseding the volume of sediments that can be eroded from, for example, the flexural foreland bulge. As noted earlier, erosion reduces the contribution of gravity-dependent terms and accelerates local

> deformation. Erosion, therefore, has an important feed-back with the geometry of the accommodation space in changing the spectrum of wavelengths (Cloetingh *et al.*, 1999). Erosion acts as a filter, suppressing the shorter wavelengths in folded basin topography. Strong erosion, insufficiently compensated by tectonic deformation, can even wipe out most of the topography. However, if the erosion is tuned to the average elevation rates (**Fig. 12**), it may dramatically accelerate folding (Cloetingh *et al.*, 1999).

There has been a rapid development and application of stratigraphic modelling packages to sequence stratigraphy over the last two decades (e.g. Cross, 1989; Alzaga-Ruiz *et al.*, 2008). Among the different types of modelling programs, the forward stratigraphic modelling approach has been applied extensively by basin modellers from academia and petroleum industry (e.g. Lawrence *et al.*, 1990; Granjeon, 2009). Forward stratigraphic modelling commonly starts with an estimate of the initial conditions related to an adopted basin formation mechanism and controlling parameters, followed by a forward prediction in time. Most of these software packages are based on sequence stratigraphic concepts (Wilgus *et al.*, 1988). Assumptions for initial subsidence are either ad-hoc or related to stretching (e.g. Lawrence *et al.*, 1990) or foreland flexure (Garcia-Castellanos *et al.*, 1997), but to date do not include lithospheric folding as a basin formation mechanism.

706 Marine versus continental deposits and closed basins versus capture of drainage

Erosion of uplifted areas and sedimentation in the depressions created by folded lithosphere is a self reinforcing process promoting continuing uplift of the highs and subsidence in the depocentres. Thus, such basins will remain closed during much of their evolution. Folded basins will hence be predominantly characterized by continental deposits with only minor deposition of marine sediments during marine incursions or during basin capture. This seems to be the case for folded basins in lberia (e.g. Duero Basin) and Central Asia (Ferghana, Tarim Basin, Lake Issyk-Kul) (Cobbold *et al.,* 1993). As has been pointed out for flexural foreland basins (Garcia-Castellanos *et al.,* 2003), the capture time for opening basins decreases with increasing flexural rigidity. The flexural upwarp to Page 29 of 91

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59 60 restoring the topography removed by erosion will be almost instantaneous for very weak
lithosphere, but requires up to 50-100 Myr for high lithospheric rigidities (Garcia-Castellanos *et al.*,
2003). Thus, basins created by folding of cratonic lithosphere probably have a shorter capture time
than ones formed in younger lithosphere.

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720 THERMAL REGIME, STRESS REGIME AND STYLE OF FAULTING

722 Thermal regime

723 Lithospheric folding is controlled by the interplay of lithospheric stresses and inherited strength of 724 the lithosphere. The thermal regime controls the rheological profile and differs between folded 725 basins developed in young lithosphere and basins in cratonic lithosphere. The latter are associated 726 with much lower thermal gradients than basins developed on young continental lithosphere. As for 727 foreland basins, the initiation of folding is not associated with a thermal instability, unless folding 728 occurs in conjunction with plume activity (Burov & Cloetingh, 2009). Following the deposition of 729 radiogenic sediments in the folded depression, sediment blanketing will affect the heat flow 730 (Stephenson et al., 1990; Lavier & Steckler, 1997; Van Wees et al., 2009), modifying the surface 731 heat flow in the basin centre. As pointed out by Lavier & Steckler (1997) and Ziegler et al. (1998), 732 the effect of sediment fill is to weaken the lithosphere. The low thermal conductivities of the 733 sediments lead to high temperatures in the upper lithosphere and consequently low local yield 734 strength. During basin subsidence sediments deposited in the central parts of the basin might be 735 exposed to temperature windows corresponding to hydrocarbon generation. At the same time, 736 sediments pre-dating the folding may have undergone extra burial due to syn-folding sedimentation. 737 An example is the Paris basin, where source rocks of Toarcian age were folded in Cenozoic time 738 (Guillocheau et al., 2000; Le Solleuz et al., 2004).

⁵⁵ 739 During basin capture overall cooling takes place. This effect is illustrated by characteristic thermal
 ⁵⁷ 740 evolution for different thermo-tectonic ages of 150 and 300 Ma of the folded continental lithosphere,

incorporating radioactive heat production in the sediments filling in the synclinal depression, shown in Figure 13. Subsidence induced by folding is calculated for the centre of the basin. Basins developing on folded lithosphere are characterized by their relatively low heat flow at the onset of folding, followed by a steady increase with time, doubling the heat flow over a time interval of the order of 5 Myr following the cessation of shortening. This increase is primarily due to the contribution of radioactive heat production in the very substantial pile of sediments accumulating in a relatively short time interval. The patterns of heat flow are similar for different thermo-mechanical ages of the lithosphere, with the oldest lithosphere having the lower heat flow. Older ages enhance the accommodation space and the contribution of the sediments to heat flow but are compensated by a larger decrease in heat flow with aging lithosphere. Another important factor is the mode of folding. Crustal scale folding, which is characteristic for Lake Issyk-Kul, will be associated with shorter wavelengths, shallower basin depths and thinner sequences of heat producing sediments than for mantle lithospheric folding. This effect might explain the relatively low heat flow observed in Lake Issyk-Kul (Vermeesch et al., 2004).

756 Modes of brittle deformation and faulting

The role of pre-existing faults on the development of folding has been investigated through numerical experiments on folding of brittle-elasto-viscous lithosphere (Beekman et al., 1996; Gerbault et al., 1998; Cloetingh et al., 1999) and through analogue experiments on plastic-elastic lithosphere (e.g. Martinod & Davy, 1994; Sokoutis et al., 2005). These studies show that lithospheric folding and faulting can develop simultaneously and that pre-existing crustal scale faults do not prevent, but instead promote the development of folding instabilities. The instabilities, in turn, can initiate the formation of new faults at the inflection points of the folds. Although the usual expectation is that faulting prevents folding, this view neglects the role of the gravity and friction on faults in the behaviour of the large-scale fault-and-fold systems. As also observed in analogue experiments (Martinod & Davy, 1994), this continuous behaviour of faulted lithosphere can be explained by fault

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 100 locking due to gravity and friction. After some sliding (uplift) on a fault, the potential gravitational 1768 energy working against friction generated by horizontal shortening forces becomes too high. As a 1769 result, the fault locks and the rock mass transmits horizontal stresses as a continuous medium. As 1760 the compression continues, one of the folds finally starts to grow faster than others, resulting in a 1771 loss of the periodicity and the formation of a mega-fold (see also the discussion in the context of the 1772 Ferghana and Tadjik basins of Central Asia).

Following the initiation of folding, and the further build up of compressional stresses, folding will, therefore, usually be accompanied by brittle deformation and faulting in the crust and upper lithosphere. The surface expression of folding in uniform lithosphere is frequently in the form of fault-controlled pop-up and pop-down structures and inverted basins (e.g. Cobbold et al., 1993). accompanied by development of imbrications of bivergent thrust wedges at the areas where shortening is applied (Sokoutis et al., 2005). These findings (see Fig. 14) are corroborated by observations in the Central System of Iberia, where folding of Variscan lithosphere with a thermo-mechanical age of 280 Ma occurs (Cloetingh et al., 2002). Here the brittle deformation associated with crustal scale folding expressed by a folded Moho is mechanically decoupled from folding of the mantle lithosphere (Fernandez-Lozano et al., 2010; see Fig. 5b).

Crustal and mantle faults may develop as a result of folding (Burov & Molnar, 1998) (Fig. 15). Folding can continue after these faults develop and folding and faulting can co-exist for times of several Myr (Fig. 5b). For crust-mantle decoupling, a complex distribution of crustal faults and surface undulations can be expected, with crustal faults not necessarily connected to mantle faults. Numerical experiments suggest that mantle faults are unlikely to be longer than a few km in depth (Burov & Molnar, 1998). Such faults may also not develop at longer wavelengths of folding connected with stronger lithosphere. However, deep seated faults in mantle lithosphere can develop when cratonic lithosphere is folded (Cloetingh et al., 1999). The mode of faulting, depth extent and spacing of faults depend on the thermo-mechanical age of the lithosphere. Once the development of folding begins in a basin, the intensity and depth penetration of faulting will accelerate. The mode

of decoupling between crustal and mantle folding is crucial. Discontinuities within the folded lithosphere will localize the development of deep synclines, flanked by more modest anticlines (Sokoutis et al., 2005) and affect the distribution of faulting (Fig. 14).

During the next phase of basin evolution, when a steady state is reached between the creation of accommodation space and sediment supply, only minor faulting will occur. This is in marked contrast with the later phase of basin capture, which is in general characterized by erosion and narrowing of the basin. Due to erosional unloading, stress relaxation will occur, manifested in upward propagating faults.

Interaction of folding with other tectonic processes

Folding, in addition to being a basin forming mechanism, frequently interacts with other tectonic processes. As pointed out by Cloetingh (1988), intraplate compression can modify pre-existing basins. As discussed above, the Pannonian Basin of Central Europe created by Miocene back-arc extension appears to be an example of such a configuration, characterized by a-typical a-periodic folding. Other examples are cratonic sag basins where intraplate compression (Cloetingh, 1988) is thought to occur in interaction with phase changes in the lithosphere (Artyushkov, 2007). Examples of these might be the Barents Sea (Ritzmann & Faleide, 2009) and possibly the South Caspian Basin, although in the latter case pre-orogenic extension cannot be ruled out (Guest et al., 2007). Basins exposed to changes in tectonic regime will have a polyphase record (Cloetingh & Ziegler, 2007), characterized by a superposition of more than two regimes, such as pre-orogenic extension, foreland flexure and late-stage folding. This sequence appears to be characteristic for some very deep foreland basins, such as the Focsani depression of the Romanian Carpathians, with more than 16 km of sediments, which was affected by extension due to the opening of the Black Sea Basin, followed by foreland flexure and overprinted by Late Miocene compression (Tarapoanca et al., 2003). As pointed out above (Ziegler & Dèzes, 2007), the Northwestern European foreland indicates folding, overprinting, rifting and foreland flexure. Quantification of the topographies created

 by rifting and foreland flexure and correcting for them is essential to reconstruct the shape of the
additional accommodation shape created by subsequent lithospheric folding (Bourgeois *et al.*,
2007).

The impingement of plumes on the base of continental lithosphere can induce differential topography similar to the surface deflections characteristic for lithosphere folds (Guillou-Frottier et al., 2007; Burov et al., 2007). Plumes and lithospheric folds frequently interact in space and time in the geological record (Ziegler & Dèzes, 2007; Burov & Cloetingh, 2009). Therefore, the evolution of folded lithosphere basins can be overprinted by the signatures of a plume or vice-versa. This will lead to an amplification of the induced vertical motions, particularly significant for young and intermediate age lithosphere. Emplacement of hot upper mantle material will raise temperatures in the lithosphere and increase heat flow. Due to the lag time in the propagation of heat, the effect at the surface might become manifest only after several tens of Myr. Plume emplacement might also weaken the lithosphere, making it more prone to compressional fault reactivation after cessation of the folding.

Interplay of lithosphere folding and plume impingement on the continental lithosphere occurred probably almost simultaneously in late Neogene times in the Northwestern European foreland. Two examples are the Eifel and Massif Central areas of the Alpine foreland of NW Europe (Cloetingh & van Wees, 2005; Ziegler & Dèzes, 2007). Both areas are sites of main Late Neogene volcanic activity in the European Cenozoic rift system (ECRIS). Seismic tomography (Ritter et al., 2001) shows finger shaped baby plumes with a characteristic spatial diameter of 100 km, extending downward to 400 km. Ziegler & Dèzes (2007) propose that plume activity occurs simultaneously with recent compressional deformation in the Massif Central area (Guillou-Frottier et al., 2007) and the Ardennes/Eifel area. Geomorphological studies constrain the recent uplift of the Ardennes and the Eifel area. An order in magnitude difference occurs between the uplift of the Eifel area, underlain by a plume and the adjacent Ardennes area where evidence for a plume is lacking. The patterns of uplift appear to be radial, superimposed by a linear NE-SW trend perpendicular to the main axis of

compression. Baby plumes primarily develop in the anticlines of lithospheric folds (Burov &
Cloetingh, 2009). The plume activity presumably accelerated the rate of uplift by a factor of 3-5
(Cloetingh & Ziegler, 2007). According to presently available data, the plumes in this segment of the
European Cenozoic Rift System arrived about 1 Myr ago. This activity was preceded and followed
by lithospheric folding that continues to present-day since 17 Myr ago (Bourgeois *et al.*, 2007).

Thermo-mechanical modelling illustrates the relative effectiveness of amplification of lithosphere deformation and topographic effects induced by plumes though folding and vice-versa. Burov & Cloetingh (2009) examined the response times and time-lags involved and whether these baby plumes were more efficient in localizing deformation than large plumes. Plume-affected folding appears to accelerate surface uplift, whereas folding goes into saturation and stagnates when plastic hinges form. For the discrimination of plumes and folding it is critical to access constraints on the presence or absence of radial versus linear symmetry, heat flow anomalies, gravity and geoid data. Plume activity facilitates folding, by dramatically lowering the stress levels required (Burov & Cloetingh, 2009). Plume impact also reduces the fold wavelength and localizes folding above the plume impact area. A general outcome of the modelling and observations is that lithospheric folding as a mechanism for producing thermal perturbations in the lithosphere/upper mantle system is a less feasible scenario.

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863 FOLDING AND OTHER MODES OF BASIN FORMATION: DIFFERENCES AND SIMILARITIES 864 IN STRUCTURE AND EVOLUTION

Sedimentary basins formed through lithospheric folding have characteristic features including vertical motions, basin architecture, thermal regime and fault activity, distinguishing them from other basin types (Cloetingh & Ziegler, 2007; Xie & Heller, 2009). **Figure 1b** and **Table 2** illustrate differences and similarities between basins developed on folded lithosphere (FLB), foreland basins (FB), intracratonic basins (ICB), extensional basins (EB), and pull-apart basins (PAB).
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Sedimentary basin systems are, by their nature, prone to tectonic reactivation and, therefore, frequently characterised by a poly-phase evolution (Cloetingh & Ziegler, 2007). Extensional basins are, for example, often formed in areas previously thickened by tectonic compression and subsequently often subject to late-stage compression (Lundin & Dorè, 1997). Examples include the Pannonian basin of Hungary (Horváth & Cloetingh, 1996) and the North Atlantic rifted margins (Cloetingh et al., 2008). Similarly, foreland basins are frequently characterised by pre-orogenic extension. Examples include the Carpathian foredeep (e.g. Tarapoanca et al., 2003) and the Aquitaine basin of Southern France, which was the retro-arc foreland basin of the Pyrenees (Desegaulx et al., 1991). The actual subsidence patterns of these poly-phase systems are often more complex than predicted by the end-member models discussed below that only consider the basin formation mechanism. Retro-arc foreland basins are characterised by a prolonged subsidence history with a distinct thermal signature including the effects of thermal cooling inherited from pre-orogenic extension. This is sometimes neglected in comparing pro-arc and retro-arc foreland basins (Naylor & Sinclair, 2007).

Basin infill records allow main features, including geometry, vertical motions and faulting characteristics to be unravelled (Zoetemeijer et al., 1993). However, heat flow histories need to be determined through kinematic reconstruction and forward modelling. Results of heat flow history modelling for FB and EB settings (Van Wees et al., 2009) can be subsequently compared to FLB results presented in this paper (Fig. 13). For ICB and PAB we have refrained from a quantitative approach for heat flow predictions as vertical motions driving subsidence cannot be linked in a straightforward way to a single or uniformly distributed lithosphere process. As a consequence, for these settings tentative heat flow patterns have been drawn.

Folded lithosphere basins (FLB) that develop as a result of periodic folding are highly symmetrical,
with dimensions that can vary from 50 km to 600 km, depending on lithospheric age and shortening
rate. These basins can accumulate thick sedimentary sequences up to the order of 20 km. The time
scales associated with this process of basin formation are very short, typically a few Myr. The

subsidence is so fast that erosion and sediment supply at the basin formation stage typically lags behind, leading to the development of starved basins, followed by shallowing upward sequences after stress relaxation. Subsidence patterns are characteristically convex upward with time. A noticeable feature is the development of significant topography of the order of several km flanking the synclinal depression. No initial heating is predicted, with an increase in heat flow with time following the basin formation. Significant brittle deformation is expected to occur in the folded basement, accelerating during the folding phase.

Foreland basins (FB) are strongly asymmetrical, with depths usually less than 10 km, and have modest flexural bulges of a few hundred meters (Royden, 1993; DeCelles & Gilles, 1996; Zoetemeijer et al., 1999). Important differences occur in the nature of the subsidence of pro- and retro-foreland basins (Naylor & Sinclair, 2007). The first are typically associated with short life spans and convex up subsidence patterns with time, whereas the latter show more prolonged subsidence histories and concave up subsidence patterns. Subsidence patterns will be interrupted during thrusting phases, also leading to uplift of the flexural marginal bulge (Quinlan & Beaumont, 1984; Zoetemeijer et al., 1993). The predicted heat flow pattern of a pro-foreland setting shown for the Ebro foreland basin (well Jabali from Verges, 1999), marked by rapid basement subsidence and sedimentary infill of ca 4 km in ca 10 My period (other parameters as in Van Wees et al., 2009). Flexure and rapid sediment infill is marked by a reduction of heat flow, followed by a progressive increase related to radiogenic heat production of foreland sediments. Basement fracturing is concentrated at the points of maximum curvature in the flexed lithosphere, at the maximum points of upward and downward deflection, respectively, with intensity higher than inferred for FLB.

Intracratoric basins (ICB) share their symmetrical shape and great thickness of their sedimentary basin fill with FLB. An important difference is their evolution in time, characteristically associated with very long time spans (typically of the order of 500-800 Myr) with long phases of low subsidence rate punctuated by phases of subsidence acceleration. As pointed out above, this is why ICB are probably affected by short-term phases of enhanced intraplate compression superimposed on other

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 mechanisms operating on longer time scales. The large dimensions of these basins point to formation in high rigidity lithosphere, that freezes in basin deformation over long time scales. The long time scales involved also imply that the highs induced by compressional upwarping will be commonly eroded. A major difference between ICB and FLB is that current hypotheses for ICB formation are casted only in terms of the tectonic processes operation at their earlier stage such as phase changes in the lithosphere. It appears, therefore, that lithospheric stress fields and the lithosphere rheology (inferred from gravity and flexure) are far more diagnostic than the lithosphere-asthenosphere boundary, which is only giving information on the thermal state of the lithosphere. Heat flow patterns of ICB will be dominated by sediment infill. ICB are in general characterized by a low intensity of faulting. However, in the absence of a rigorous understanding of the formation mechanisms of ICB, predictions for heat flow are often not available. As a consequence, their heat flow is most often described primarily in terms of the thermal consequences of their burial history.

Extensional basins (EB) have received major attention from basin modellers, both because of their relative abundance and importance and because of the availability of a relatively simple quantitative model for their formation (McKenzie, 1978). The latter can be relatively easily be incorporated in forward modelling of their subsidence and heat flow histories (Van Wees et al., 2009). The basin shape is symmetrical or asymmetrical depending on a pure shear or simple shear mode of extension. Rifts can be narrow or wide, depending on the mode of rifting and the rheology of the extending lithosphere (Huismans and Beaumont, 2002). Dependent on the rigidity of the extending lithosphere, rift shoulders with asymmetrical shape can be formed and be prone to erosion (Van Balen et al., 1995, Van der Beek et al., 1995). The development of rift shoulders and their erosion can lead to feed back with lower crustal flow and faulting in the extending lithosphere (Burov & Cloetingh, 1997). During the post-rift phase, flexural widening of the basin commonly occurs. A characteristic feature of EB predicted by simple stretching models (McKenzie, 1978) is the succession of rapid initial subsidence in the syn-rift phase and exponentially decaying subsidence in their post-rift phase with a typical thermally controlled time constants of the order of 70 Myr. Feed

backs between rift shoulder erosion and lower crustal flow (Burov & Cloetingh, 1997; Burov & Poliakov, 2001), have pronounced effects on subsidence in the final phases of rifting and in the early post-rift phase. Both can be influenced by changes in the mechanical properties of the lithosphere and its strengthening upon rifting. Multiple rifting phases is another common feature of EB. The duration of the syn-rift phase can vary from quasi-instantaneous (shorter than 20 Myr) to several hundreds of Myr (Cloetingh and Ziegler, 2007). In the latter case basin migration is a common feature (Corti et al., 2003). Extensional faulting occurs during the rift phase, with absence of faulting in the post-rift phase. Heat flow reaches a peak during the syn-rift phase, followed by a rapid decrease in heat flow (Van Wees et al., 2009). The syn-rift peak in heat flow is low or almost absent in continental extensional settings as a consequence of removal of heat producing elements in the crust and cooling by sediment infill. Fig. 16 shows such a setting, which is based on the benchmark model presented in Van Wees et al. (2009), As a consequence of the loss of radiogenic heat production in the crust at the end of the post-rift stage values, heat flow evolves to values considerably lower than the initial ones. The EB heat flow pattern is, therefore, the reverse of that in FB where syntectonic heat flow is first lowered due to sediment infilling (Kombrink et al., 2008), but eventually increases due to thermal relaxation and additional radiogenic heat production in the sediments.

Pull-apart basins (PAB) are characterized by a narrow fault bounded geometry (Smit et al., 2008). developing over a very short time span. Models have been developed to calculate subsidence histories and heat flow histories for these basins (Pitman & Golovchenko, 1983), incorporating lateral heat transport in stretching models. These models predict rapid subsidence of several km's over time intervals of the order of 0.1 - 1 Myr, followed by a decay in the subsidence with a concave shape upward. Erosion and sediment flux can not keep pace with the subsidence at the basin formation phase, leading initially to a starved basin, followed by a shallowing upward sequence (Pitman & Golovchenko, 1983). These basins are intensively faulted, also due to their strike-slip origin, whereas heat flow decays on very short time scales of the order of a few Myr. Although this

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 974 scenario is a good first order description, many PAB show a more punctuated fault controlled
975 subsidence history, often terminating by late-stage uplift, as observed for Late Neogene pull-apart
976 basins in the internal zone of the Betics in SE Spain (Cloetingh *et al.*, 1992) and for the flanks of
977 pull-apart basins in the Dead Sea Fault System (Smit *et al.*, 2008, 2010).

From the above overview, it is apparent that FLB have in common with FB a compressional origin and thermal histories controlled to a large extent by burial histories. At the same time, pronounced differences occur in the duration of subsidence, varying from ultra short (FLB 1-10's Myr), medium (FB, tens of Myr) and ultra long (ICB, hundreds of Myr). An important difference occurs in the spatial scales of the topography generated by folding and foreland fold-and-thrust belt deformation. In the latter case a much steeper differential topography is generated over typical horizontal differences of the order of 10 km between peak and trough. In contrast, in the case of lithospheric folding horizontal distances between the topographic highs and the axes of the depocentres are typically of the order of hundreds of kilometres. As a result, diffusive transfer of mass will occur at a much slower rate of sediment supply in the case of folded lithosphere than rates of sediment supply locally generated by thrusting. This together with a very rapid initial subsidence of folded lithosphere basins has important consequences for the paleobathymetry of these systems.

FLB contrast with extensional and pull-apart basins in the shape of the subsidence curves, being convex up for FLB and concave for EB and PAB. The heat flow histories differ also. EB and PAB are characterized by a phase of initial heating, whereas FLB and FB are characterized by an absence of initial heating. A common feature of FLB and PAB formation is their capability to cause dramatic differential vertical motions in a short time interval, predicting shallowing upward sequences. FLB are to some extent unique in terms of the magnitude of the uplifted topography surrounding the down warped depocentres. Similar magnitudes are only reached at rift shoulders of basins developing in intracratonic lithosphere (Van der Beek et al., 1995). FLB can be associated with localized brittle deformation and pop-up structures in the upper crust and widespread brittle deformation at deeper crustal and upper mantle levels.

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CONCLUSIONS

4 Sedimentary basins controlled by lithospheric folding are characterized by a number of features in 5 their basin architecture, subsidence history, thermal evolution and faulting patterns making them 6 distinctly different from other basin systems. Of particular importance are the relatively short 7 temporal scales involved in their evolution, whereas their spatial scales vary from tens to several 8 hundreds of km, depending on the rheological stratification of the lithosphere and its thermo-9 mechanical age. Subsidence patterns are characterized by an acceleration of subsidence in the 0 depocentres during compressional basin formation, occurring simultaneously with accelerated uplift 1 of and erosion from flanking basin highs. In contrast with asymmetrical foreland basins, basins on 2 folded lithosphere are symmetrical with equal dimensions for linearly trending topographic highs and 3 lows.

Basins on folded lithosphere have a strong tendency to remain closed systems during a major part of their evolution, with capture times decreasing with thermo-mechanical age from several tens of Myr to less than 10 Myr. The thermal history of these basins is not characterized by an initial thermal perturbation, unless folding occurs in interaction with plume activity. In the absence of an interplay with lithospheric scale thermal perturbations, their thermal evolution will be primarily controlled by the burial history of the sediments deposited in the accommodation space created by the folding process.

Intrabasinal faulting and the formation of pop-up structures and inverted basins at upper crustal levels are important in the structural evolution of these basin systems. The intensity of the brittle deformation peaks with ongoing shortening of the lithosphere during folding but faulting activity is dramatically reduced after the cessation of folding. At this stage, with an equilibrium between sedimentation and erosion, minor faulting occurs as a result of stress relaxation due to erosion of

1 2 3	1026	folded highs. During capture and overall uplift of basins on folded lithesphere, their accommodation
4	1020	Tolded highs. During capture and overall upint of basins of folded innosphere, their accommodation
6	1027	space is further reduced by unflexing and associated stress release through faulting.
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APPENDIX A: MODEL DESCRIPTION

A1. Equations. The Flamar-Para(o)voz is a "2.5 D" FLAC-like code (Cundall, 1989). It has a mixed finite-difference/finite element numerical scheme, in which the coordinate frame is Cartesian 2D, but stress/strain relations are computed in a 3D formulation. The Lagrangian mesh of Paravoz is composed of quadrilateral elements subdivided onto 2 couples of triangular sub-elements with trilinear shape functions. Para(o)voz is a large strain fully explicit time-marching algorithm. It locally solves full Newtonian equations of motion in continuum mechanics approximation:

$$\left\langle \rho \mathbf{i} \mathbf{i} \right\rangle - di v \mathbf{\sigma} - \rho \mathbf{g} = 0 \tag{A1}$$

1608 coupled with constitutive equations:

$$\frac{D\sigma}{Dt} = F(\sigma, \mathbf{u}, \nabla \dot{\mathbf{u}}, ...T...)$$
(A2)

1609 and with equations of heat transfer (heat advection $\boldsymbol{u} \nabla T$ in the equation below is solved 1610 separately):

$$\rho C_p DT/Dt + \dot{\mathbf{u}} \nabla T - k \operatorname{div}(\nabla T) - H_r = 0$$

$$\rho = \rho_0 (1 - \alpha T)$$
(A3)

Here **u**, σ , **g**, k are the respective terms for the displacement, stress, acceleration due to body forces and thermal conductivity. The triangular brackets in (A1) specify conditional use of the related term (in quasi-static mode inertial terms are dumped using inertial mass scaling (Cundall, 1989)). The terms t, ρ , C_{ρ} , T and H_r designate respectively time, density, specific heat, temperature and internal heat production. The terms D/Dt, D_{σ}/Dt , F are a time derivative, an objective (Jaumann) stress time derivative and a functional, respectively. In the Lagrangian method, incremental displacements are added to the grid coordinates allowing the mesh to move and deform with material. This enables solution of large-strain problems locally using a small-strain formulation. On

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1619 each time step the solution is obtained in local coordinates, which are then updated in a large strain1620 mode.

Solution of (A1) provides velocities at mesh points used for computation of element strains and of heat advection $u\nabla T$. These strains are used in (A2) to calculate element stresses and equivalent forces as input for the computation of the velocities for the next time step. Due to the explicit approach, there are no convergence issues, which is rather a common problem of implicit methods in case of non-linear rheologies. The algorithm automatically checks and adopts the internal time step using 0.1 - 0.5 of Courant's criterion for propagation of information, which warrants a stable solution.

,

A2. Explicit EVP rheology.

We use a serial (Maxwell type) body (eq. 1), in which total strain increment in each numeric element is defined by a sum of elastic, viscous and brittle strain increments. Consequently, in contrast to fluid dynamic approaches, where non-viscous rheological terms are simulated using pseudo-plastic and pseudo-elastic viscous terms (e.g. Bercovici *et al.*, 2001; Solomatov & Moresi, 2000), our method explicitly treats all rheological terms. The parameters of elastic-ductile-plastic rheology laws for crust and mantle come from rock mechanics data (**Tables A1 and A2**; Kirby & Kronenberg, 1987; Kohlstedt *et al.*, 1995).

. 5 1637

Plastic (brittle) behavior.

1639 The brittle behavior of rocks is described by Byerlee's law (Byerlee, 1978; Ranalli, 1995), which 1640 corresponds to Mohr-Coulomb material with a friction angle $\phi = 30^{\circ}$ and cohesion $|C_0| < 20$ MPa 1641 (e.g., Gerbault *et al.*, 1998):

$$|\tau = C_0 - \sigma_r \tan \phi$$
(A4)
1642 where σ_n is normal stress $\sigma_n = \frac{1}{2}\sigma_1 + \sigma_n^{-1}e^{i\sigma_n} \sin \phi$, $\frac{1}{2}\sigma_1 = P$ is the effective pressure, $\sigma_n^{-1}e^{i\sigma_n}$ is the
1643 second invariant of deviatoric stress, or effective shear stress. The condition of transition to brittle
1644 deformation (function of rupture h reads as: $f = \sigma_n^{-1}e^{i\sigma_n} + P \sin \phi - C_n \cos \phi = 0$ and $\partial ll \partial l = 0$. In terms of
1645 principal stresses, the equivalent of the yield criterion (5) reads as:
1646
1647 **Elastic behavior**.
1648 The elastic behavior.
1649 where λ and G are Lame's constants. Repeating indexes mean summation and δ is Kronecker's
operator.
1651
1652 **Viscous (ductile) behavior**.
1653 Mantle convection and a part of lithospheric deformation is controlled by thermally activated creep
1654 (Ranalli, 1995; Kirby & Kronenberg, 1987). Within deep lithosphere and underlying mantle regions,
1655 the creeping flow is non-Newtonian since the effective viscosity can vary within 10 orders of
1656 magnitude as a function of differential stress:
 $e^{\sigma_1} = A(\sigma_1 - \sigma_2)^n exp(-(\Omega R^{-1}\Gamma^{-1}))$ (A7)
1657 Where $\sigma_{\pi}^2 = \delta_{\pi}^2$ is shear strain rate, A is material constant, *n* is the power law exponent, *Q* is the
activation enthalpy, *R* is Boltzman's gas constant, and *T* is temperature in K^{*}, σ_1 and σ_2 are the
1659 principal stresses. The effective viscosity μ_m for this law is defined as:

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 1660 which yields:

$$\mu_{eff} = e^{d}_{ij} \,^{(1-n)/n} \, A^{-1/n} \, exp \, (Q \, (nRT)^{-1})$$
(A8)

1661 For non-uniaxial deformation, the law (A8) is converted to a triaxial form, using invariant of strain 1662 rate and geometrical proportionality factors:

 $\mu_{eff} = e^{d}_{ij} (1-n)/n A^{-1/n} exp (Q (nRT)^{-1})$

 $\tau_{ij} \equiv \mu_{eff} \, \mathbf{e}^{d}_{ij}$

1663 where

$$e^{d}_{II} = (Inv_{II} (e_{ij}))^{\frac{1}{2}}$$
 and $A^{*} = \frac{1}{2}A \cdot 3^{(n+1)/2}$ (A9)

parameters A, n, Q are experimentally determined material constants (Table A1). Using olivine parameters (Table A1), one can verify that the predicted effective viscosity at the base of the lithosphere is $10^{19} - 5 \times 10^{19}$ Pa s matching post-glacial rebound data (Turcotte & Schubert, 2002). In the depth interval of 200 km - 0 km the effective viscosity grows from 10¹⁹ to 10²⁵ - 10²⁷ Pa s with decreasing temperature. Within the adiabatic temperature interval at depth, the dislocation flow law (A8) is replaced by nearly Newtonian diffusion creep, which results in a quasi-constant mantle viscosity of 10¹⁹ – 10²¹ Pa s (e.g., Turcotte & Schubert, 2002). Weinberg & Podladchikov (1994) have also shown that the effective viscosity in close vicinity of an ascending diapir is influenced by the local strain rate field and partly by heat exchanges between the diapir and surrounding rock. which suggests possible changes.

1675 FIGURE CAPTIONS

Figure 1. Concepts of folding in rheologically stratified lithosphere and feedback with sedimentation
in downfolded areas and erosion of adjacent highs. Different simultaneously occurring wavelengths
of crustal and mantle folding are a consequence of the rheological stratification of the lithosphere.
Surface wavelengths can be affected also by feedback with surface processes.

Figure 2. Map with examples of well-documented areas affected by lithospheric folding. Areas: Central Indian Ocean Basin (CIOB) (Geller et al., 1983 ; Stein et al., 1989) ; NW European platform (Bourgeois et al., 2007; Lefort & Agarwal, 1996; Marotta et al., 2000); Pannonian Basin (Dombradi et al., 2010); Iberia (Cloetingh et al., 2002; De Vicente et al., 2007; Fernandez-Lozano et al., 2010); Central Asia (Burov et al., 1993; Nikishin et al., 1993; Burov & Molnar, 1998); Tibet/ Himalayan syntaxis belt (Burg & Podlachikov, 1999; Shin et al., 2009); Central Australia (Stephenson & Lambeck, 1985; Lambeck, 1983); Arctic Canada (Stephenson et al., 1990); Transcontinental Arch of North America (Ziegler et al., 1995); South Caspian Basin (Guest et al., 2007); Laramide Foreland (USA) (Tikoff & Maxson, 2001); Barents Sea (Ritzmann & Faleide, 2009) (map credit: NASA).

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Figure 3. Relationship between wavelengths of lithospheric folding and thermo-tectonic age of lithosphere. The grey zones correspond to theoretical folding wavelengths derived for the upper crust, mantle and coupled whole mantle folding, based on an analytical model that accounts for strength variations as a function of thermo-mechanical age (Cloetingh et al., 1999). Black squares are estimates for wavelengths of differential topography and thermo-tectonic ages inferred from observational studies. Numbers refer to sites listed in Table 1.

Figure 4. (a) Map showing major tectonic features of Tien Shan area of Central Asia, with location of folding profiles discussed in Burov et al. (1993), Burov & Molnar (1998) and Cloetingh et al. (1999). (b) Neotectonic movements along the profiles AD and CD (Burov et al., 1993). Modified after Cloetingh et al. (1999). (c) Measured and calculated Bouquer gravity anomalies for lithosphere under horizontal compression along profile F across the Ferghana basin with corresponding topography profile (d) (after Burov & Molnar, 1998). Predicted topography and Moho geometry for Airy and best fitting folding scenario. During compression the amplitude of plate deflection changed significantly while the wavelength of folding stayed constant (Burov & Molnar, 1998). Red lines are Bouquer anomaly predicted by folding model and Moho depression, respectively. Solid grey line is Airy isostasy. Solid black line is Airy Moho. Triangles indicate observed Bouguer anomaly. Green is topography above sea level. FB: Ferghana Basin.

Figure 5. First order features of basin geometry of folded lithosphere of Iberia. (**a**) Topography displaying cylindrical patterns of alternating parallel trending highs and lows and corresponding gravity anomalies (red line) (see Cloetingh *et al.*, 2002). (**b**) Results from analogue modelling experiments, displaying pop-up structures underlying topographic highs, accompanied by folding of the Moho (Fernandez-Lozano *et al.*, 2010). Top panel shows final top view of the analogue model and location of the cross-section shown below. Arrow indicates the direction of the moving wall. λ is the wavelength of folding.

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Figure 6. (a) Map of elevation of basement in the NW Alpine foreland with rift signature removed from the map of actual basement (from Bourgeois *et al.*, 2007). Dashed lines show axes of Late Neogene lithospheric folds, with a succession of anticlines and synclines: the Normandy-Vogelsberg anticline (NVA), Sologne-Franconian Basin Syncline (SFS) and the Burgundy-Swabian Jura Anticline (BSJA). LBR: London Brabant Ridge with thick crust. BF: Black Forest; FB Franconian Basin; E: Eifel; H: Hunsruck; S: Sauerland; T: Taunus; V: Vosges. (b) Cross sections of

top basement in the NW Alpine foreland with rift signatures removed (Bourgeois *et al.*, 2007; see for location Fig. 6a). Thin line: actual topography surface. Bold line: elevation of top basement as predicted in the absence of rifting. Fold signatures are similar with a wavelength λ of 270 km and amplitude *A* of 1500 m in the Franconian Basin (Cross section G-G') and in the upper Rhine Graben area (Cross section H-H'). In the Paris basin, development of folds with assumed same wavelength and amplitude (stippled bold line in cross section C-C'), implies subsidence of Sologne and uplift of Normandy.

Figure 7. (a) Schematic E-W cross section through the Pannonian-Carpathian basin system (Cloetingh et al., 2008), displaying anomalous Late Neogene acceleration of the subsidence between isochrons 2 and 0 Ma in the centre of the Pannonian basin and the Focsani depression of the Southern Carpathian foredeep, flanked by uplifting highs. Wavelenghs of the differential vertical motions are typically of the order of 250 km. (b) Interpreted seismic section in the central part of the Pannonian Basin (Sacchi et al., 1999) showing evidence for late Neogene crustal scale folding at scales of 20-40 km. Note syn-rift grabens formed during the Miocene back-arc extension of the basin. (c) Results of analogue tectonic experiments carried out to examine the patterns of differential vertical motions in the northern part of the Pannonian Basin and surrounding areas (Dombradi et al., 2010), demonstrating a-typical folding of variable amplitude with a wavelength equivalent to 200-250 km.

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Figure 8. Schematic north-south trending cross-section illustrating the basic tectonic setting for northern Iran and the South Caspian Basin (Guest et al., 2007), illustrating basin inversion of the central Iranian basin, pronounced differential topography at the transition of the Alborz mountains, with a topographic high of several km's and the South Caspian Basin with about 20 km of sedimentary infill.

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752 Figure 9. Characteristic patterns of continental lithosphere deformation induced by folding of the 753 lithosphere for with time increasing amounts of horizontal shortening for a 150 Ma old lithosphere 754 with convergence rates of (a) 1.5 cm/yr and (b) 3 cm/yr. From left to right: material phase 755 distribution (with blue for the crust, orange for the mantle lithosphere and red for the upper mantle 756 below the lithosphere), temperature, effective viscosity, and topography.

758 Figure 10. Characteristic patterns of continental lithosphere deformation induced by folding of 759 lithosphere with time. Increasing amounts of horizontal shortening for a 300 Ma old lithosphere with 760 convergence rates of (a) 1.5 cm/yr and (b) 3 cm/yr. From left to right: phase distribution, 761 temperature, effective viscosity, and topography.

763 Figure 11. (a) Characteristic subsidence patterns in centre of synclinal depression for thermo-764 mechanical age of 150 and 300 Ma, respectively. Shortening rate is 3 cm/year. (b) Characteristic 765 stages in the evolution of a basin formed by lithospheric folding. Stage 1: Basin formation stage, 766 acceleration of subsidence and uplift during folding. Stage 2: Steady-state: equilibrium between 767 tectonic subsidence and sediment supply from eroding highs. Stage 3: Capture of folded basin; 768 overall uplift and erosion. Wiggled waved blue line marks position of base level.

770 Figure 12. Illustration of the effect of erosion, which acts as a filter suppressing the short 771 wavelengths. In this case, the wavelength and amplitude vary along the cross section at different 772 stages of deformation due to a partial crust-mantle coupling and strain localization for a 400 Ma old 773 lithosphere with weak guartz -dominated rheology. After 5 percent shortening (top), after 25 percent 774 shortening (middle), after 25 percent shortening, strong zero order diffusional erosion (Avouac & 775 Burov, 1996) tuned to keep mean elevation at the level of 3000 m. Erosion reduces the contribution 776 of gravity-dependent terms (middle wavelength) and accelerates local deformation. Strong erosion,

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insufficiently compensated by the tectonic deformation wipes out most of the topography (afterCloetingh *et al.*, 1999).

Figure 13. Thermal consequences of lithosphere folding, and associated basin evolution depicted in Fig. 11. The heat flow response is marked by an initial decrease in heat flow as a consequence of rapid sedimentation, followed by a progressive long-term heat flow increase after termination of shortening. Thermal conductivity of sediments is 2.5 Wm⁻¹C⁻¹. Surface radiogenic heat production H_s is 9.5x10⁻¹⁰ Wkg⁻¹. Concentration of radiogenic heat sources within sedimentary basin fill is homogeneous. Exponential decay of heat production is assumed for underlying crustal rocks. Note that the figure shows only 7 Ma of thermal evolution. The heat flow does not increase anymore after some tens of Ma and subsequently decreases at later stages.

Figure 14. Results of analogue tectonic experiments for lithospheric folding (after Sokoutis et al., 2005) (Top): folding of uniform lithosphere. Cross section demonstrating pop-up structures in the upper crust above highs induced by lithospheric folding. λ_1 and λ_2 indicate first-order and second-order wavelengths, respectively. Imbrication of bivergent thrust wedges is responsible for the relatively high topography belt (Σ -belt) at the right-hand of the model and closure of prismatic basins with a rather undisturbed bottom. The progressive sinking of the prismatic basin allows transferring of the crustal material (i.e. cover sediments) that was initially situated at the surface of the model down to significant depth. Left-side (a) final top view of model after 34% of bulk shortening. Right side (b) cross-section located on the top view. Arrows indicate the direction of the moving wall. Note the similarity to Central Iberia (Fig. 5). (Bottom): Geometries predicted by analogue experiment, shortening a lithosphere with two contrasting blocks, bounded by suture. A deep synclinal depocentre develops over the suture zone, flanked by an anticline of smaller amplitude. Left-side (c) final top view of model after 26% of bulk shortening. Right side (d) cross-section located on the top

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- 3 4 5 6	1802	view. λ_1 indicates a first-order wavelength of the folding. Note the striking similarity to the			
	1803	configuration of the South Caspian Basin/ Alborz Mountain system displayed in Fig. 8.			
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$\begin{array}{c}1\\2\\3\\4\\5\\6\\7\\8\\9\\10\\11\\2\\13\\14\\15\\16\\17\\8\\9\\21\\22\\3\\4\\25\\6\\7\\8\\9\\0\\1\\2\\2\\3\\3\\4\\3\\3\\3\\4\\3\\3\\3\\4\\3\\3\\3\\4\\3\\3\\3\\4\\3\\3\\3\\3\\4\\3$	1805	Figure 15. Numerical experiment on faulting developed in folded continental lithosphere of 150 Ma			
	1806	(Burov & Molnar, 1998). Note different wavelengths of the crustal and mantle part of the lithosphere			
	1807	and differences in spacing and character of faults that are conditioned by folding. In each layer, two			
	1808	fault spacings occur: a shorter one proportional to layer thickness (also controlled by the streng			
	1809	contrast between the stiff layer and embeddings), and the longer wavelength controlled by the			
	1810	wavelength of folding. The insert shows plastic (brittle) strain localization characterizing initialization			
	1811	of mantle faulting.			
	1812				
	1813	Figure 16. Comparison of main characteristics of basins developed on folded lithosphere (FLB with			
	1814	a shortening rate of 3 cm/yr) with other basin types, including foreland basins (FBr = retro-ar			
	1815	foreland basin; FBp = pro-arc foreland basin), intracratonic basins (ICB), extensional basins (E			
	1816	and pull-apart basins (PAB). Different panels give theoretical predictions. (Top): Basin subsidence			
35 36 27	1817	history in basin centre. (Bottom): Thermal history. See text for explanation.			
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TABLE 1

2 Wavelengths and ages of folded lithosphere displayed in Fig. 3 (see Fig. 2 for location)

	Area	Thermo- tectonic age (Ma)	Folding wavelength λ (km)	References
1	Tien Shan	175	200-250	Burov & Molnar, 1998; Burg et al., 1994
2	Western Goby	175-400	300-360	Nikishin et al., 1993; Burov et al., 1993
3	Central Asia	370-430	50-70 (crust) 300-400 (litho-mantle)	Nikishin <i>et al.</i> , 1993; Burov <i>et al.</i> , 1993
4	Himalayan syntaxis belt	8-10	150	Burg & Podlachikov, 1999
5	Central Australia	500-900	550-650	Lambeck, 1983; Stephenson & Lambeck, 1985; Beekman <i>et al.</i> , 1997
6	Russian platform	400-600	500-600	Nikishin <i>et al.</i> , 1997
7	South Caspian Basin	125-155	350-450	Guest <i>et al.</i> , 2007
8	Eastern Black Sea	40-80	50-100 (crust) 100-150 (litho-mantle)	Cloetingh <i>et al.</i> , 2008
9	Western Black Sea	75-125	50-100 (crust) 100-200 (litho-mantle)	Cloetingh <i>et al.</i> , 2008
10	Pannonian Basin System	20	200-250	Horvath & Cloetingh, 1996; Matenco et al., 2007; Dombradi et al., 2010
11	NW European platform	180-230 🦷	270	Bourgeois et al., 2007
12	Brittany	210-290	225-275	Bonnet <i>et al.</i> , 2000
13	Iberia	330-370	40-80 (crust) 125-275 (litho-mantle)	Cloetingh <i>et al.</i> , 2002
14	Barents Sea	215-245	550-650	Ritzmann & Faleide, 2009
15	Canadian Arctic	150-250	170-230	Stephenson et al., 1990
16	Transcontinental Arch of North America	1000-1400	500-700	Ziegler <i>et al.</i> , 1995
17	Laramide foreland (USA)	175-225	190	Tikoff & Maxson, 2001

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6	TABLE 2. Key characteristic features of different classes of sedimentary basins

Basin type	Basin shape	Dimensions	Subsidence characteristics	Thermal evolution	Faulting history
FLB	Symmetrical	Width: 50-600 km; depth: up to 20 km	Accelerated subsidence, time scales 1-10 Myr; simultaneously uplift at highs	No initial heating; heat flow increases in first 10's Myr with sediment deposition and burial, and subsequently decreases	Intensive deformation, accelerating trough basin formation phase
FB	Asymmetrical	Width: 50-250 km; depth: up to 10 km	Linear subsidence, interrupted by short-lived faster subsidence during thrusting phases; timescales 10-100 Myr; simultaneously uplift from thrust front and foreland bulge	No initial heating; heat flow increases with time with sediment deposition and burial	Faulting and thrusting limited to orogenic wedge; minor deformation through faulting in foreland
ЕВ	Symmetrical (pure shear); asymmetrical (simple shear)	Width: 30-500 km; depth: up to 10 km	Rapid initial subsidence, followed by post-rift decay in subsidence; in case of multiple rifting repeated accelerated subsidence; time scales for post-rift subsidence of the order of 70 Myr; At rifting stage development takes place of asymmetrical rift shoulder topography, flanking the rift	Initial heating event, followed by decaying heat flow	Extensional faulting limited to basin formation phase; during post-rift absence of faulting
РАВ	Symmetrical	Width: 20-50 km; depth: up to 10 km	Very rapid initial subsidence, followed by post-rift decay in subsidence, time scales: 1-10 Myr In early stage, sedimentation cannot keep up with subsidence	Initial heating event, followed by very rapid decay in heat flow	Extensional faulting limited to basin formation phase; during post-rift absence of faulting
ICB	Symmetrical	Width: 500-1000 km; depth: up to 10 km	Slow subsidence, punctuated by periods of faster subsidence, possibly related to large-scale tectonic events; Characteristic time scales: 500-800 Myr	In absence of understanding of basin formation mechanism no prediction for initial heat flow; Heat flow at later stage dominated by burial history	Very low level of faulting throughout tectonic history

8 Table A1

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9 Notations and physical values common for all experiments (Turcotte and Schubert, 2002; Schubert 10 *et al.*, 2001)

11 **Parameter** Values and units Definition Pa, MPa 12 stress (complete, deviatoric) σ, τ Ρ 13 Pa, MPa pressure 14 m. mm displacement vector u m s⁻¹, mm an⁻¹ 15 velocity vector v s⁻¹ strain rate 16 Ė 10¹⁹-10²⁵ Pa s 17 effective viscosity $(\tau/\dot{\varepsilon})$ μ 18 Т °C, K temperature Moho depth $h_{\rm c}$ 19 40 km 20 100 - 250 km thickness of lithosphere $h_{\rm I}$ 100 - 200 km 21 D plume diameter 3330 kg m⁻³ density of mantle lithosphere 22 ρ 3340 kg m⁻³ 23 reference deep mantle density $\rho_{\rm m}$ 24 plume density $\rho_{\rm m} + \Delta \rho_{\rm ch} + \alpha \rho_{\rm m} \Delta T$ ho_{p} 0-25 kg m⁻³ 25 chemical density contrast $\Delta \rho_{ch}$ 9.8 m s^{-2} 26 acceleration due to gravity g 10³ J Kg⁻¹ ℃⁻¹ $C_{\rm p}$ 27 specific heat 250°C 28 ΔT initial temperature contrast plume - background 3×10⁻⁵ ℃⁻¹ 29 thermal expansion coefficient at 0 km depth α 8×10⁻¹² Pa⁻¹ isothermal compressibility 30 β 31

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is density; Q, n, A are material-dependent parameters	s of ductile flow laws (Kirby & I
1987; Kohlstedt et al., 1995). Other parameters from Tur	rcotte & Schubert (2002).
Parameter	Value
All rocks	
λ , G Lamé elastic constants ($\lambda = G$)	30 GPa (above 250 km depth)
λ , G Lamé elastic constants ($\lambda = G$)	60 GPa (below 250 km depth)
ϕ friction angle (Mohr-Coulomb criterion)	30°
C_0 cohesion (Mohr-Coulomb criterion)	20 MPa
Specific upper or weak (quartz) lower-crust properti	ies
ρ (upper crust)	2700 kg m ⁻³
ρ (lower crust)	2900 kg m ⁻³
n	2.4
Α	6.7 × 10 ^{−6} MPa ^{−n} ·s ^{−1}
Q	$1.56 \times 10^5 \text{ kJ} \cdot \text{mol}^{-1}$
Specific strong lower crust properties (diabase or basalt)	
ρ	2980 kg m ⁻³
'n	3.4
A	2 × 10 ⁻⁴ MPa ⁻ⁿ ·s ⁻¹
Q	$2.6 \times 10^5 \text{ kJ} \cdot \text{mol}^{-1}$
Specific mantle properties (olivine)	
ρ (lithosphere)	3330 kg m ⁻³
n	3
A	$1 \times 10^4 \text{ MPa}^{-n} \cdot \text{s}^{-1}$
Q	5.2 × 10 ⁵ kJ·mol ^{−1}
Thermal model	
Surface temperature (0 km depth)	O°O
Temperature at the bottom of thermal lithosphere	1330 <i>°</i> C
Temperature at 660 km depth	2000 °C
Thermal conductivity of crust k	2.5 Wm ⁻¹ ℃ ⁻¹
Thermal conductivity of mantle k	3.5 Wm ⁻¹ ℃ ⁻¹
Thermal diffusivity of mantle χ	10 ⁻⁶ m ² .s ⁻¹
Surface radiogenic heat production H_{s}	9.5×10⁻¹⁰ W kg⁻¹
Radiogenic heat production decay depth hr	10 km
Thermo-tectonic age of the lithosphere a	60 (young) to 1000 Ma (old)













Figure 3













Figure 6a



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H'

N

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C

Figure 6b G +2 km Swabian Jura Molasse Basin Netherlands Sauerland Tai Spessart Franconian Basin 0 old signature (= top of basement after A=1500 m removal of synthetic rift signature) $\lambda = 270 \text{ km}$ 200 km -5 km H +2 km Lower Rhine Eifel Hunsrück Lorraine URG Black Forest Swabian Molasse Alps Graben & Flanders Jura Basin 0 Fold signature (= top of basement after removal of synthetic rift signature) $\lambda = 270 \text{ km}$ 200 km -5 km С $\lambda = 270 \text{ km}$ +2 km Molasse Jura Morvan Bresse Paris Basin Sologne Normandy Basin , 0 A=1500 m Fold signature Top of basement after removal of synthetic rift signa 200 km -5 km

Figure 7a





Figure 7b













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