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Lithospheric folding and sedimentary basin evolution: a review and analysis of formation mechanisms

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4 1 **Lithospheric folding and sedimentary basin evolution:**
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14 6 **Sierd Cloetingh^{a,*} and Evgenii Burov^b**
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24 **ABSTRACT**

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

35 **INTRODUCTION**

37 Over the last few years much progress has been made in understanding the mechanisms of the
38 formation of sedimentary basins (Cloetingh & Ziegler, 2007; Roure et al., 2010). Quantitative
39 models for several basin types have significantly advanced the understanding of extensional
40 (McKenzie, 1978; Van Wees *et al.*, 2009) and foreland basins (Beaumont, 1981; Naylor & Sinclair,
41 2008). The mechanisms of basin formation as a result of lithospheric folding have received
42 considerably less attention (Cloetingh & Ziegler, 2007) except for some clear-cut cases, in particular
43 in central Asia (Burov & Molnar, 1998; Thomas *et al.*, 1999a,b). Many data exist on the geometry of
44 sedimentary basins affected by large-scale compressional intraplate deformation (Cobbold *et al.*,
45 1993; Lefort & Agarwal, 1996, 2000, 2002; Burov & Molnar, 1998). These observations and results
46 of analytical, numerical (Martinod & Davy, 1992; Burov *et al.*, 1993; Gerbault *et al.*, 1998; Cloetingh
47 *et al.*, 1999) and analogue modelling (Martinod & Davy, 1994; Sokoutis *et al.*, 2005) demonstrate
48 that the thermo-mechanical age of the lithosphere exerts a prime control on the wavelength of
49 lithospheric folds. These studies focused on the role of tectonic stress in lithospheric folding

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3 50 (Stephenson *et al.*, 1990; Burov & Molnar, 1998; Stephenson & Cloetingh, 1991; Perez-Gussinye &
4
5 51 Watts, 2005). However, surface processes also play a significant role in the mechanics of
6
7 52 lithospheric folding. Erosion enhances the development of folding and has a pronounced effect on
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10 53 its wavelengths (Cloetingh *et al.*, 1999; Burov & Toussaint, 2007), specifically in the short-
11
12 54 wavelength domain. In addition, both sedimentation and erosion are likely to significantly prolong
13
14 55 the lifetime of folding. Sedimentation decreases the effect of gravity by filling the downward flexed
15
16 56 basins and thus reducing the isostatic restoring force, whereas erosion of the upward flexed
17
18 57 basement unloads the lithosphere on the basins' uplifted flanks. Lithospheric folding has significant
19
20 58 effects for the geometry of sedimentary sequences deposited on folded lithosphere. However, the
21
22 59 effect has been studied little for several reasons (Cobbold *et al.*, 1993). The hanging walls of
23
24 60 emerging low-angle thrust faults tend to collapse rapidly, blurring the structural relationships at
25
26 61 basin edges. Further, ongoing sedimentation may onlap and bury the hanging walls, whereas scarp
27
28 62 erosion may destroy them. Seismic images of basin edges may not be well resolved where footwall
29
30 63 conglomerates contain few reflecting horizons or when low-angle thrusts reflect and refract seismic
31
32 64 waves, thus hampering the imaging of deeper parts of the basin. Downwarping of folded basins
33
34 65 renders surface geological mapping difficult. At the same time, high plateaux are not at first sight
35
36 66 linked with basins of high hydrocarbon potential, leaving them in many places without well control or
37
38 67 seismic constraints on basin architecture.

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42 68 In this paper we review evidence from a number of basins in Eurasia, where folding has been
43
44 69 documented and alternative basin formation mechanisms such as lithospheric extension or foreland
45
46 70 flexure cannot fully explain key features of basin geometry and evolution. We restrict ourselves to
47
48 71 cases where independent evidence points to a major role for lithospheric folding in the basin
49
50 72 development. We demonstrate that an interpretation in terms of folding can sometimes resolve
51
52 73 inconsistencies resulting from interpreting basins origin via other mechanisms. For example
53
54 74 (Ritzmann & Faleide, 2009), the formation mechanism of intracratonic basins is still a matter of
55
56 75 debate. A striking feature of these basins is the prolonged intervals of low rate subsidence
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3 76 alternating with episodic accelerations in subsidence rates, often coeval with orogenic activity at
4
5 77 plate boundaries. Due to the frequent absence of evidence for thinned lithosphere below these
6
7 78 basins, alternative models have been proposed. Stel *et al.* (1993) propose that non-extensional
8
9 79 crustal thinning is induced by basaltic underplating, which causes thermal uplift and erosional
10
11 80 thinning. Tectonic loading at plate boundaries is also proposed as a cause of or contributor to large-
12
13 81 scale sag-type subsidence in plate interiors (Cloetingh, 1988; Leighton & Kolata, 1990; Quinlan,
14
15 82 1987) as well as phase transformations (Artyushkov, 2007). Ritzmann & Faleide (2009) point out
16
17 83 that intraplate stresses and crustal inhomogeneities coupled with loading scenarios provide the best
18
19 84 explanation for the Barents Sea Basin and other intracratonic basin evolution. The basins reviewed
20
21 85 here, however, do not include cases such as the Barents Sea where the effect of stresses on
22
23 86 topography and accommodation space has been considerably complicated by other basin formation
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25 87 processes.

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29 88 We present inferences from thermo-mechanical models of lithospheric folding and a comparison
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31 89 with the results of analogue modelling of folding. We concentrate on the overall geometry of basins
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33 90 formed by lithospheric folding, the typical timescales intrinsic to their development and the brittle
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35 91 deformation patterns in the underlying lithosphere. We then discuss subsidence patterns and heat
36
37 92 flow characteristics for such basins. We conclude by comparing key features of basin
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39 93 accommodation shape, vertical motions, thermal history and brittle deformation patterns of basins
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41 94 formed on folded lithosphere, foreland basins, intracratonic basins, extensional basins and pull-
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43 95 apart basins.

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49 97 **LITHOSPHERIC FOLDING: AN IMPORTANT MODE OF INTRAPLATE BASIN FORMATION AND**
50
51 98 **DEFORMATION**

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55 100 Folding of the lithosphere, involving positive and negative deflections (**Fig. 1**), appears to be of
56
57 101 more importance in the large-scale deformation of intraplate domains than hitherto realized
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3 102 (Cloetingh *et al.*, 1999). As has been shown (Burov *et al.*, 1993; Nikishin *et al.*, 1993; Cloetingh *et*
4
5 103 *al.*, 1999; Gerbault *et al.*, 1998; Schmalholtz & Podladchikov, 2000), visco-elastic folding starts to
6
7 104 develop from the onset of compression and in contrast to elastic folding, does not require
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9
10 105 specifically large intraplate stresses. Folding may continue until the back-ground strain-rate drops,
11
12 106 for example due to the localization of deformation in a specific area or to reduction of far-field
13
14 107 forces. Such intraplate deformation results from transmission of intraplate stress fields away from
15
16 108 plate boundaries into continental forelands (Ziegler *et al.*, 1998; Van der Pluijm *et al.*, 1997; Tesauro
17
18 109 *et al.*, 2005).

19
20 110 In continental lithosphere, several wavelengths of folding are expected to develop as result of the
21
22 111 presence of several rheologically competent layers decoupled from each other by weak layers (**Fig.**
23
24 112 **1, Fig. 3**). Crustal folding can be traced from the surface topography, neotectonic movements and
25
26 113 free air gravity. The crust-mantle boundary (Moho) topography reflects mantle folding, because the
27
28 114 mantle strength is highest at this interface and thus the geometry of Moho deflection is most likely to
29
30 115 correspond to that of mantle folding. This wavelength can be traced from Bouguer gravity anomalies
31
32 116 because the crust-mantle boundary represents also the main density contrast in the lithosphere. It is
33
34 117 to be expected that the base of the Mechanical Lithosphere (MLB) has roughly the same geometry
35
36 118 as the Moho deflection. Likely, the lithosphere-asthenosphere boundary (LAB) has little to do with
37
38 119 mantle lithosphere folding except some specific cases, as the MLB (roughly corresponding to 700°C
39
40 120 depth) is found well above the LAB (1330°C) . Consequently, the low viscosity zone between MLB
41
42 121 and LAB may damp the deformation in a way that the LAB remains even flat. Due to the low
43
44 122 viscosity of gravitating mantle between the LAB and MLB, the LAB is not necessarily advected or
45
46 123 down-warped with deflection of the MLB, or at least the relations between the deformation of the
47
48 124 MLB and LAB are not straightforward. In particular, gravity instabilities or convective movements not
49
50 125 related to folding can perturb the LAB without impact on the MLB.
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3 127 Folding has important implications for vertical motions, sedimentary basin architecture and the
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5 128 evolution of hydrocarbon systems (Ziegler *et al.*, 1995; 1998). The large wavelength of vertical
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7 129 motions associated with lithospheric folding necessitates the integration of data from large areas
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9
10 130 (Elfrink, 2001; Allen & Davies, 2007), often beyond the scope of regional structural and geophysical
11
12 131 studies that target specific structural provinces. Recent studies on the North German Basin show
13
14 132 neotectonic reactivation by lithospheric folding (Marotta *et al.*, 2000). Similarly, acceleration of the
15
16 133 Plio-Pleistocene subsidence in the North Sea Basin is attributed to stress-induced buckling of its
17
18 134 lithosphere (Van Wees & Cloetingh, 1996). Folding of the Variscan lithosphere has been
19
20 135 documented for Brittany (Bonnet *et al.*, 2000; Lagarde *et al.*, 2000), the adjacent Paris Basin (Lefort
21
22 & Agarwal, 1996) and the Vosges-Black Forest arch (Dèzes *et al.*, 2004; Ziegler & Dèzes, 2007;
23
24 136 Bourgeois *et al.*, 2007). Lithospheric folding, therefore, appears to be an effective mechanism for the
25
26 137 propagation of tectonic deformation from active plate boundaries far into intraplate domains (e.g.
27
28 138 Stephenson & Cloetingh, 1991; Burov *et al.*, 1993; Ziegler *et al.*, 1995, 1998).
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30 139
31 140 Folding can be observed at different spatial scales. At the scale of a micro-continent that was
32
33 141 affected by a succession of collisional events, Iberia illustrates lithospheric folding and interplay
34
35 142 between neotectonics and surface processes (Cloetingh *et al.*, 2002). An important factor favouring
36
37 143 a lithosphere-folding scenario for Iberia is the compatibility of the thermo-tectonic age of its
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39 144 lithosphere and the wavelength of observed deformations.
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42 145 Other well-documented examples of continental lithospheric folding occur in other cratons (**Fig. 2**).
43
44 146 Prominent examples occur in the Western Goby area, in the Ferghana basin, but also probably in
45
46 147 the Tadjik basin of Central Asia, involving a lithosphere with thermo-tectonic ages from 150 Ma
47
48 148 (Kazakh shield, Ferghana, Tadjik basins and likely Dzhungaria basin) to 400 Ma (Tarim basin). In
49
50 149 this area, mantle and crustal wavelengths are 360 km and 50 km, respectively, with a shortening
51
52 150 rate of ~10-20 mm/yr and 200-250 km of shortening during 10-15 Myr (Burov *et al.*, 1993; Burov &
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54 151 Molnar, 1998). Three-dimensional fold structures in the Tibetan Plateau have been inferred from
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3 152 GRACE satellite gravity data, demonstrating prevailing fold wavelengths of 300 to 420 km (Shin *et*
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5 153 *al.*, 2009).

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8 154 The inferred wavelength of these neotectonic lithosphere folds is consistent with the general
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10 155 relationship between the wavelength of lithospheric folds and the thermo-tectonic age of the
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12 156 lithosphere shown by a global inventory of lithospheric folds (**Fig. 3**) (Cloetingh & Burov, 1996;
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14 157 Cloetingh *et al.*, 2005). In some other areas of continental lithosphere folding, smaller wavelength
15
16 158 crustal folds have also been detected (Burov *et al.*, 1993; Nikishin *et al.*, 1993). Thermal thinning of
17
18 159 the mantle lithosphere, often associated with volcanism and doming, enhances lithospheric folding
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20 160 and appears to affect the wavelengths of folds (Burov & Cloetingh, 2009).

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24 25 162 **OBSERVATIONS**

26
27 163 Basins developed on folded continental lithosphere have characteristic features. We illustrate this
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29 164 by largely concentrating on the basins of Central Asia, the folded lithosphere of Iberia, the NW
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31 165 European platform, the Pannonian Basin and the South Caspian Basin.

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35 36 167 **Compressional basins of Central Asia (Tianshan, Ferghana)**

37
38 168 The young Ferghana and Tadjik basins are compressional basins northwest and north of the Pamir
39
40 169 and south of the Tien Shan ranges (Burov & Molnar, 1998). In this area lithosphere underwent
41
42 170 Jurassic reactivation and is characterized by relatively young thermo-mechanical ages (175 Ma)
43
44 171 (Burg *et al.*, 1994; Burov & Molnar, 1998). In addition to having undergone thermal weakening, the
45
46 172 lithosphere underlying these basins also probably has a weak, quartz-dominated lower crustal
47
48 173 rheology. This has resulted in low values for the effective elastic plate thickness (EET) of the order
49
50 174 of 15 km (Burov & Molnar, 1998).

51
52
53 175 The Central Tien Shan consists of an alternation of ranges and basins separated by reverse faults.

54
55 176 The Tien Shan terminates west of the right-lateral Talas-Ferghana fault by splaying into two narrow
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3 177 mountain chains that surround the Ferghana Valley. A narrow belt of mountains parallel to the
4
5 178 central section of the fault slopes downward to the Ferghana Valley, underlain by a deep basin with
6
7 179 as much as 8 km of Cretaceous-Cenozoic sediments (Cobbold *et al.*, 1993). Gravity data suggest
8
9 180 that the Ferghana and Tadjik basins are gravitationally overcompensated. The negative Bouguer
10
11 181 gravity anomalies indicate a Moho several km deeper than predicted by models of local Airy
12
13 182 isostasy (Burov & Molnar, 1998). This points to Moho down-warping due to non-isostatic processes
14
15 183 and most probably, in this compressional context, to lithospheric folding. As demonstrated by
16
17 184 numerical-thermomechanical experiments (Burov & Molnar, 1998), the approximately north-south
18
19 185 shortening of the relatively thin lithosphere could have created central down-warping and anticline
20
21 186 mountains north and south of the basin. According to this model, this shortening has produced a
22
23 187 folding instability in the mantle lithosphere that has warped the basement immediately surrounding
24
25 188 the basin upward, and has forced the basin floor down beneath the Ferghana Valley (with an
26
27 189 estimate for the folding wavelength of the order of 200-250 km), possibly also producing deep
28
29 190 mantle faulting (seen as Moho offsets in seismic reflection data).

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32
33 191 Pre-existing thermal structure and variations in crustal thickness have played a major control on the
34
35 192 styles and distribution of faulting in this region (**Fig. 4**), (see also Burov & Molnar 1998; Cobbold *et*
36
37 193 *al.*, 1993). Primary faults appear before the folding develops with a spacing that is proportional to
38
39 194 brittle layer thickness. Subsequently the two processes, faulting and folding, co-exist in such a way
40
41 195 that faults localize at the inflection points of folds. At this stage, the appearance of faults does not
42
43 196 significantly influence the wavelength of folding. Because of the weakness of the lower crust, the
44
45 197 upper crust should be completely decoupled from the mantle and interact with it only by flow in the
46
47 198 lower crust. Lower crustal flow changes the wavelength and amplitude of the surface folding, which
48
49 199 is terminated by the development of a single down-warped megafold (Cloetingh *et al.*, 1999). Liu *et*
50
51 200 *al.* (2007) reported evidence for lithospheric folding in the Central/ North China Tarim Basin region,
52
53 201 initiated during Pliocene times, with good time constraints provided by Ar/Ar dating. Two
54
55 202 wavelengths are found of 30 and 400 km. The thermo-mechanical age of the lithosphere is
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3 203 estimated to be 250 Ma, as the main thermal perturbation was associated with a major orogenic
4
5 204 phase at 250 Ma ago. These findings are consistent with models in which decoupled lithosphere
6
7 205 folds with 30 km and 400 km wavelength correspond to, respectively, crustal and mantle folding.
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10 206 Deep seismic data for basins north of and within the Tibetan plateau (Li *et al.*, 2006; Liu *et al.*, 2006;
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12 207 Zhao *et al.*, 2006) show that folding is not limited to the Tibetan plateau (see also Shin *et al.*, 2009).
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14 208

16 209 **Folded lithosphere of Iberia**

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18 210 As reviewed by Cloetingh *et al.* (2002), Tertiary lithospheric folding of Iberia occurs in Variscan
19
20 211 lithosphere with wavelengths of 300 km, leading to the development of a system of parallel trending
21
22 212 basins and highs (**Fig. 5a**). This folding generated the distribution of basins and mountain chains,
23
24 213 bounded by folds and faults (Cloetingh *et al.*, 2002; De Vicente *et al.*, 2007). The regularity of the
25
26 214 fluvial network pattern of the central-western part of Iberia also indicates large-scale lithospheric
27
28 215 folding. Alpine (mainly Cantabrian-Pyrenean related) compressional tectonics was the principal
29
30 216 factor for closing internally drained sedimentary basins, including the Ebro, Duero and Tagus basins
31
32 217 (Casas-Sainz & De Vicente, 2009).
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35
36 218 A large body of geophysical and geological observations is available on the crustal geometries and
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38 219 stress regimes of Iberia (De Vicente *et al.*, 2007, 2008). This is supplemented by thermo-
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40 220 geochronology data (De Bruijne & Andriessen, 2002; Ter Voorde *et al.*, 2004), demonstrating
41
42 221 accelerated uplift of the highs during Late Miocene-Pliocene times. In spite of the existence of
43
44 222 recent localized volcanism in the inner part of the Iberian plate, an important contribution from
45
46 223 thermal uplift is questionable due to the distribution of crustal thickness (Casas-Sainz & De Vicente,
47
48 224 2009). Fernandez-Lozano *et al.* (2010) have constructed a three-layer analogue model consisting of
49
50 225 brittle upper crust, ductile lower crust and ductile upper mantle for Iberia. These models show that
51
52 226 folding is associated with deformation of narrow mountain ranges separated by basins (see **Fig.**
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54
55 227 **5b**). The narrow mountain ranges representing upper crustal pop-ups form the main topographic
56
57 228 reliefs. Shortening is accommodated within the viscous crust beneath the pop-ups leading to lateral
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3 229 thickness variations of the ductile crust. These results are consistent with seismic and gravity data
4
5 230 collected in the Cantabrian mountains, the Spanish Central System, and the Toledo Mountains (see
6
7 231 also **Fig. 5a**).

8
9 232 The earliest sediments in the terrestrial basins (Calvo et al., 1993) were deposited during the
10
11 233 Eocene, with very well developed continuous successions, dominated by continental facies with
12
13 234 high sedimentation rates. The Duero and Tagus basins began to be individualized at the onset of
14
15 235 uplift and the formation of a topographic range in the Central System (Casas-Sainz & De Vicente,
16
17 236 2009). In contrast with this early more homogeneous stage, the Oligocene-Lower Miocene
18
19 237 stratigraphy of the basins is characterized by several tectono-stratigraphic units, separated by
20
21 238 unconformities representing short periods of time. The number and specific ages of these units
22
23 239 varies between basins, indicating variable activity at basin borders. The fact that the largest number
24
25 240 of breaks in the sedimentary succession is concentrated during this time interval points to a tectonic
26
27 241 control (De Vicente & Vegas, 2009).

28
29 242 The final stages in the evolution of these onshore basins are characterized by widespread Miocene
30
31 243 lacustrine sedimentation in the endorheic basins and during Pliocene and Quaternary times, by the
32
33 244 beginning of river capture and exorheism in the Duero, Ebro and Tagus basins (Garcia-Castellanos
34
35 245 *et al.*, 2003). Alluvial deposits were deposited by the fluvial network developed on the ancient,
36
37 246 internally drained basins (Casas-Sainz & De Vicente, 2009). During the endorheic-exorheic
38
39 247 transition, paleo-karstification processes and erosion occurred in the Late Miocene lacustrine
40
41 248 limestones. Two main types of high plains can be distinguished on the Iberian Peninsula. The
42
43 249 coupled origin of both types of surfaces can be related to uplift/sedimentation histories along the
44
45 250 Tertiary compressional basin margins in the Iberian Peninsula (Casas-Sainz & De Vicente, 2009).
46
47 251 The first are high plains with Tertiary sedimentary basins, containing well preserved Late Miocene to
48
49 252 Pliocene strata with subhorizontal bedding. Typical examples are the Duero and Tagus basins. The
50
51 253 second consists of eroded sediments of Mesozoic and Paleozoic rocks with planation surfaces
52
53 254 developed on the Paleogene-Lower Miocene ranges created by fold and thrust systems and
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3 255 basement uplifts. Different patterns in evolution and stratigraphy between the eastern and western
4
5 256 Tagus basin may be related to the mechanical decoupling between Iberia and Africa during the
6
7 257 Miocene emplacement of the Alboran Domain in the southern part of the plate. This generated
8
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10 258 extensional stresses in the eastern Iberian plate, coeval with compression and lithospheric folding in
11
12 259 the western part of the plate (De Vicente *et al.*, 2008; De Vicente & Vegas, 2009).

13
14 260 In general the dimensions of peneplains linked to extensional basins and passive margins are
15
16 261 larger, and their evolution involves time scales much longer than those characteristic for high plains
17
18 262 of the Iberian plate (Casas-Sainz & De Vicente, 2009). Peneplains associated with compressional
19
20 263 or convergent regimes (e.g. Anatolia, Central Iran and Morocco) are characterized by spatial
21
22 264 dimensions comparable with Iberia. In addition, their tectonic setting in the hanging wall of the
23
24 265 African-Eurasian convergence zone is also similar. Thus the Iberian plate may be an analogue for
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26 266 other present-day intra-mountain basins.

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30 31 268 **The European Alpine foreland**

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33 269 Geophysical and geomorphological studies (e.g. Lefort & Agarwal, 1996; 2000; 2002; Marotta *et al.*,
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35 270 2000) indicate large-scale compressional deformation of the European Alpine foreland lithosphere.
36
37 271 Although different in structural grain and rheological structure (Tesauro *et al.*, 2007; Cloetingh *et al.*,
38
39 272 2005) inherited from differences in their Paleozoic-Mesozoic history, different segments of the
40
41 273 lithosphere in the northern Alpine European foreland share many similarities in their Cenozoic
42
43 274 intraplate deformation. Seismic and geological evidence shows that the North German Basin cannot
44
45 275 be explained by classical models of basin formation (Marotta *et al.*, 2000). The basin underwent a
46
47 276 polyphase history (Scheck & Bayer, 1999; Littke *et al.*, 2008) of extension in the late Triassic
48
49 277 followed by lithospheric scale folding in the late Cretaceous-Early Cenozoic (Mazur *et al.*, 2005;
50
51 278 Marotta *et al.*, 2000). 3D volumetric analysis shows that the compressional event was followed by
52
53 279 faster subsidence during the Cenozoic (Scheck & Bayer, 1999). A similar observation was made by
54
55 280 Van Wees & Cloetingh (1996) for the North Sea Basin. Lithospheric folding modified a basin formed
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3 281 initially by deep seated thermal perturbations (Scheck & Bayer, 1999) with a thinned crust and
4
5 282 shallow Moho. The Moho updoming with a high of 3 km for the height of the bulge in the southern
6
7 283 part of the North German basin appears to be the consequence of flexural buckling of the previously
8
9 284 thinned lithosphere by a compressive stress perpendicular to the strike of the basin.

10 285 From earliest Triassic to Late Jurassic, the Paris Basin (**Fig. 6a**) subsided in an extensional
11
12 286 framework and was larger than the present basin (Guillocheau *et al.*, 2000). The western margin of
13
14 287 the Paris Basin and the rifted Atlantic margin of France were subject to thermal rejuvenation during
15
16 288 Mesozoic extension related to North Atlantic rifting (Ziegler & Dèzes, 2006; Robin *et al.*, 2003).
17
18 289 Subsequent compressional intraplate deformation (Ziegler *et al.*, 1995) also affected the Paris Basin
19
20 290 (Lefort & Agarwal, 1996). Numerical modelling indicates that the amplitude of the compression
21
22 291 induced vertical deflection in the Paris Basin is relatively small, requiring additional mechanisms
23
24 292 such as flexure due to the load by the Alpine system (Bourgeois *et al.*, 2007; Ziegler & Dèzes,
25
26 293 2007). However, periodic deformation at the same wavelength occurs northwest of the Paris Basin
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28 294 (**Fig. 6 a**, Bourgeois *et al.*, 2007), which would not be the case if the north-eastern part of the basin
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30 295 was flexed down by the load of the Alps. Bourgeois *et al.* (2007) carefully analysed basement
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32 296 geometry and timing of vertical motions in the NW European platform. They separate the
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34 297 contributions from the European Cenozoic Rift System (ECRIS) and the contributions from long-
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36 298 wavelength folding striking NE and located between the Alpine front and the North Sea. According
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38 299 to their analysis, the ECRIS developed mostly between 37 and 17 Ma, whereas lithospheric folds
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40 300 developed between 17 Ma and present, with a wavelength of 270 km and amplitude of 1500 m (see
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42 301 **Fig. 6b**).

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44 302 Quaternary folding of the Variscan lithosphere in the area of the Armorican Massif (Bonnet *et al.*,
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46 303 2000) developed folds with a wavelength of 250 km, pointing to a mantle-lithospheric control on
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48 304 deformation. The timing and spatial pattern of uplift inferred from river incision studies in Brittany is
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50 305 incompatible with a glacio-eustatic origin. Therefore, Bonnet *et al.* (2000) related the vertical
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52 306 motions to deflection of the lithosphere by the present-day NW-SE directed compressional intraplate
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3 307 stress field. Stress-induced uplift of the area appears to control fluvial incision rates and the position
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5 308 of the main drainage divides. Leveling studies (LeNotre *et al.*, 1999) indicate its ongoing
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7 309 deformation.
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11 311 **Pannonian Basin System**

12 312 Recent studies of the Pannonian Basin system - formed as a Miocene back-arc basin behind the
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14 313 Carpathian arc - show that active crustal deformation resulted in significant differential vertical
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16 314 motions during Late Neogene times (Horvath & Cloetingh, 1996; Cloetingh *et al.*, 2005, 2006;
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18 315 Horvath *et al.*, 2006; Matenco *et al.*, 2007; **Fig. 7a**). The river network in the area is affected by
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20 316 differential neotectonic crustal motions (Necea *et al.*, 2005; Dombradi *et al.*, 2010).
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23 317 The Late Neogene evolution of the Pannonian Basin is largely controlled by its interaction with the
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25 318 Carpathian arc and the Adriatic indenter impinging the Pannonian basin lithosphere (e.g. Bada *et al.*,
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27 319 2001). As a result of the consumption of subductable lithosphere of the European foreland, the
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29 320 Pannonian basin became locked. At the same time, the continuous N-NE directed indentation of the
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31 321 Adriatic microplate built up a high level of compressional stresses in the Pannonian Basin. The
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33 322 structural inversion of the basin is a direct consequence of the temporal changes in the stress field
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35 323 from extension to compression. Recent studies of the stress field together with an extensive array of
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37 324 geological and geophysical data have shown that horizontal stresses are transmitted from the plate
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39 325 boundaries into the interior of the Pannonian Basin (Bada *et al.*, 2007).
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44 326 The Pannonian Basin is underlain by what is probably the hottest lithosphere of continental Europe,
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46 327 characterized by a very low lithospheric strength (Tesauro *et al.*, 2007). Miocene extension resulted
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48 328 in a high level of mantle stretching, leading to a strength distribution with practically zero strength in
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50 329 the subcrustal lithosphere and lower crust (Cloetingh *et al.*, 2005), making it prone to tectonic
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52 330 reactivation. Therefore, in this area relatively low compressional stresses induced mainly by plate
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54 331 boundary interactions, and further amplified by topography-induced stresses (Bada *et al.*, 2001),
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56 332 could initiate significant intraplate deformation.
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3 333 It should be noted that in the Pannonian basin system, seismic tomography (Wortel & Spakman,
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5 334 2000) supports slab detachment under the adjacent Romanian Carpathians. At the same time, the
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7 335 seismic tomography shows the presence of a hot upper mantle below the folded lithosphere in the
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9 336 Pannonian basin. In this area, therefore, the upper mantle structure reflects more the back-arc
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11 337 extension prior to the folding than the consequences of folding on the density stratification of the
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13 338 mantle. The reverse is true for the crustal configuration. The geometries imaged by the crustal scale
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15 339 deep seismic reflection profiling are clearly dominated by compressional structures resulting from
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17 340 folding (**Fig. 7b**).

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19
20 341 Quaternary patterns of differential uplift and subsidence with a timing coeval with an increased level
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22 342 of compressional stresses are interpreted as caused by lithospheric folding (Horvath & Cloetingh,
23
24 343 1996; Cloetingh *et al.*, 2006; Horvath *et al.*, 2006; Bada *et al.*, 2007). Late-stage compressional
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26 344 deformation of the Pannonian basin is demonstrated to be of key importance for the geometry of
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28 345 basin fill (Sacchi *et al.*, 1999, see **Fig. 7b**) and for its reservoir structures, both crucial in
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30 346 assessment of its hydrocarbon potential (Horváth, 1995). Analogue tectonic experiments by
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32 347 Dombradi *et al.* (2010) further examine accumulation and transmission of stresses in the hot and
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34 348 weak back-arc lithosphere of the Pannonian basin and the interplay between lithospheric processes
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36 349 and the intrabasinal topography development. Based on available rheological and seismotectonic
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38 350 constraints, the basin system is represented by a two-layer model with a 10-15 km thick brittle upper
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40 351 crust. The lower crust and the mantle are represented by a low-strength layer of 45 km thickness
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42 352 (Dombradi *et al.*, 2010). The strain rates applied were constrained by GPS velocities (Grenerczy *et*
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44 353 *al.*, 2005). In the experiments lithospheric folds develop with wavelengths equivalent to 200-250 km
45
46 354 (**Fig. 7c**). The experiments also show that crustal thickness variations largely determine the
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48 355 localisation of stresses. Topographic profiles extracted by laser scans of the model show that
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50 356 surface topography reflects long wavelength buckling of the lithosphere. These results are
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52 357 compatible with Pliocene-Quaternary subsidence, which occurs in the centre of the Pannonian
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54 358 basin, simultaneously with accelerated uplift at the basin margins. The latter has lead to elevated
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3 359 topography of up to 1 km in areas at the rim of the Pannonian Basin such as the Styrian, Zala and
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5 360 Transylvanian basins (Cloetingh *et al.*, 2006).
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9 362 **The South Caspian Basin**

10 363 The South Caspian Basin with an estimated sedimentary thickness up to 20 km (Brunet *et al.*, 2003;
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12 364 Egan *et al.*, 2009), is probably one the deepest basins in the world. It is flanked by the Alborz
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14 365 Mountains to the south, with an elevation of 2-4 km (see **Fig. 8**). The transition between basin and
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16 366 flanking high is abrupt and coincides with a coastline bounded by a major fault system. The
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18 367 formation of the South Caspian Basin is enigmatic. Attempts to explain it as an extensional basin or
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20 368 as a foreland basin have been unsuccessful. There is a lack of evidence for basin extension and
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22 369 both stretching models and topographic loads and slab pull operating on down flexed lithosphere in
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24 370 foreland basin systems cannot explain differential vertical motions of this order of magnitude. As a
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26 371 result, phase changes in continental crust have been proposed as a mechanism for the formation of
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28 372 this superdeep basin (Artyushkov, 2007). A limitation in conclusively resolving the basin formation
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30 373 mechanism is the lack of multi-channel seismic reflection data capable of imaging crustal structure
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32 374 below the thick sedimentary sequences. Over the last few years, however, other constraints on the
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34 375 area have become available (e.g. Guest *et al.*, 2007).
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40 376 **Figure 8** presents a cross section of northern Iran and the South Caspian Basin (Guest *et al.*,
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42 377 2007). As shown, wavelengths for compressional folding of the lithosphere are typically in the range
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44 378 of 400 km. Although the actual basin formation mechanism for the Caspian Basin is not well
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46 379 resolved (e.g. Artyushkov, 2007), late Neogene folding has been affecting a lithosphere probably
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48 380 thermally reset by middle Late Jurassic marginal basin formation (Guest *et al.*, 2007) with a thermo-
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50 381 mechanical age at the onset of collision of 130-150 Ma. Satellite data demonstrate an exceptionally
51
52 382 large gravity anomaly over the basin (Kaban *et al.*, 2002). Seismic tomography shows slab
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54 383 detachment under the Iranian plateau (Hafkenscheid *et al.*, 2006; Alinaghi *et al.*, 2007) beginning at
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56 384 about 10-15 Ma.
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3 385 Basin analysis has shown that basement inversion has occurred 20 Ma ago, simultaneously with
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5 386 the slowing of Arabian/Eurasian plate convergence and the onset of accumulation of Neogene
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7 387 clastics in foreland basins (e.g. Fakhari et al., 2008). By 10 Ma, contraction occurred by
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9 388 underplating of the Arabian crustal units beneath the Iranian plate. Estimated shortening rates from
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11 389 current geodetic surveys (Vernant *et al.*, 2004) are 7 mm/yr. Results from geothermochronology
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13 390 demonstrate rapid uplift of the Alborz mountains, with rates of 0.7 km/my exhumation between 6
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15 391 and 4 Ma (Axen *et al.*, 2001), implying approximately 10 km of uplift. This uplift was nearly
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17 392 synchronous with rapid South Caspian subsidence (Nadirov *et al.*, 1997) and subsequent folding
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19 393 (Devlin *et al.*, 1999). South Caspian sedimentation rates locally increased more than tenfold at ca 6
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21 394 Ma, with more than 10 km of sediments deposited since then. Axen *et al.* (2001) argue that if
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23 395 approximately 10 km of post 6 Ma sediments are present in this basin, then as much as 20 km
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25 396 (equivalent to 80 %) of the structural relief of about 25 km between the high Alborz and the
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27 397 southernmost Caspian basement may be younger than 6 Ma.

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31 398 Folds in the southernmost Caspian Basin (Devlin *et al.*, 1999) and in the Neogene of the Northern
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33 399 Alborz foothills imply contraction. The lack of a crustal root under the Alborz Mountains also points
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35 400 to a flexural support by the South Caspian basement (Axen *et al.*, 2001). The simultaneous reversal
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37 401 of Zagros strike slip and extrusion of central Iran, coarse Zagros molasse deposition, Dead Sea
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39 402 transform reorganisation, Red Sea oceanic spreading (Chu & Gordon, 1998) all initiated around $5 \pm$
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41 403 2 Ma (e.g. Axen et al., 2001; Smit et al., 2010), was interpreted by these authors in terms of a
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43 404 widespread tectonic event. Slab detachment could explain part of the recent uplift of Iran, but does
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45 405 not directly explain the simultaneous dramatic subsidence in the adjacent Caspian Basin. However,
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47 406 slab detachment could have an indirect effect if reducing slab pull forces acting on the downgoing
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49 407 lithosphere promote the development of compressional stresses in both the downgoing and
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51 408 overriding plates (Cloetingh *et al.*, 2004; Matenco *et al.*, 2007).

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55 409 Guest *et al.* (2007) argue in favour of compressional deformation of the South Caspian Basin/Alborz
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57 410 mountains. This interpretation was largely based on available industrial seismic reflection data and
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3 411 gravity data for the basin and followed a similar proposition made for the late-stage deformation of
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5 412 the adjacent Black Sea (Nikishin *et al.*, 2003; Cloetingh *et al.*, 2008). Guest *et al.* (2007) present
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7 413 kinematic models, without addressing the reason for the difference in the order of magnitude of the
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9 414 basin depression and the adjacent high. Analogue experiments (Sokoutis *et al.* 2005) show that
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11 415 when compression acts on two blocks with contrasting thickness (and rigidity), a major syncline will
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13 416 develop on top of the suture zone separating these blocks, flanked by an anticline of much lower
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15 417 amplitude. These results are consistent with predictions from folding theory that as a result of the
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17 418 acting gravity field, down-warping is mechanically more effective than up-warping against the action
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19 419 of gravity forces. The presented numerical experiments illustrate this (see for example **Fig. 9**).
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25 421 **Basins on folded lithosphere in Eurasia: differences and similarities in lithospheric**
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27 422 **deformation and basin habitat**

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29 423 Comparison of lithospheric deformation in Central Asia, Iberia, the NW European Alpine foreland,
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31 424 the Pannonian Basin and the South Caspian Basin demonstrates common features compatible with
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33 425 an interpretation in terms of basins developed on folded lithosphere. Amongst these are their
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35 426 symmetrical shape, dimensions and the temporal association of their basin formation history with
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37 427 the build up of intraplate stresses. Important differences can be observed between the Central
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39 428 Asian and Iberian basins and the South Caspian Basin on one side and the European Alpine
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41 429 foreland and the Pannonian Basin on the other side.

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44 430 The Central Asian Basins and the basins of the Iberian plate were initiated by Tertiary folding,
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46 431 whereas the European Alpine foreland underwent a polyphase history, following Carboniferous
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48 432 and Triassic stretching (Scheck & Bayer, 1999; Le Solleuz *et al.*, 2004), prior to Late Cretaceous-
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50 433 Early Cenozoic folding. Similarly, the Pannonian Basin, formed by back-arc extension in Miocene
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52 434 times, was subsequently compressionaly reactivated (Cloetingh *et al.*, 2008). The origin of the
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54 435 South Caspian Basin is not well resolved, but km scale vertical motions are compatible with a
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56 436 megafold as a prime control on present-day basin geometry. These differences in evolution have a
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3 437 pronounced impact on basin dimensions and on the nature of basin fill. The Central Asian basins,
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5 438 the folded basins of Iberia, and the South Caspian Basin have a wavelength compatible with
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7 439 theoretical predictions. In contrast, the fold wavelengths of the European Alpine foreland and the
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9 440 Pannonian Basin appear to be primarily controlled by the pre-existing geometry of the sediment-
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11 441 basement interface and pre-existing Moho topography inherited from the pre-folding stage.
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13 442 Differences occur also in the development of the ranges flanking the basins. In Central Asia and
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15 443 Iberia, these highs have amplitudes and widths comparable to the prediction for periodic folding,
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17 444 whereas in the Paris Basin and the North German Basin these highs are less well developed. The
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19 445 ratio of the depth of the South Caspian Basin to the height of the flanking Alborz Mountains is
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21 446 compatible with a megafold under this area of dramatic differential topography. The large
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23 447 amplitudes of folding at the flanks of the Pannonian Basin system are compatible with its extremely
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25 448 weak rheology. Differences in the amplitude of the highs might also partly result from the differences
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27 449 in neotectonic shortening. The highs bounding the basins of Iberia, Pannonian Basin, South
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29 450 Caspian Basin and Central Asia, located relatively close to the site of ongoing convergence, are still
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31 451 experiencing active uplift. This is much less the case for the low strain deforming areas of the NW
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33 452 European platform and the North German basin where the peak compression occurred during Late
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35 453 Cretaceous-Early Cenozoic. Another difference occurs in the depositional regime prevailing in these
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37 454 basins. The closed basin systems of Central Asia, Iberia and the intra-Carpathian Pannonian Basin
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39 455 are dominated by continental and lacustrine deposits, whereas capture is usually a late-stage
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41 456 feature of these basins. The Paris basin of the NW European platform has a basin fill largely
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43 457 dominated by marine sediments deposited during the pre-folding stage of their evolution, followed
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45 458 by marine Tertiary successions in the basin centre with truncated basin margins. The creation of
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47 459 basin highs in this case was in general not capable of interrupting open connections to marine
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49 460 environments.

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51 461 A general observation for all these basin systems is the flatness of sedimentary sequences in the
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53 462 depocentres. Faulting plays a minor role in the basin development. Faulting appears to be
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3 463 concentrated at the basin margins, with a highly variable depth extent, varying from upper crustal
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5 464 scale faults bounding pop-up structures in the Central System of Iberia and the Pannonian Basin to
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7 465 minor faulting in the basement of the North German Basin. The thermal regimes of the basins are
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9 466 similar with thermal histories strongly affected by subsidence and sediment deposition on down
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11 467 flexed lithosphere, pre-conditioned by its thermo-mechanical age.
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16 469 **THE ROLE OF LITHOSPHERIC RHEOLOGY**

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18 470 Folding is commonly associated with periodic deformation of layered structures with contrasting
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20 471 mechanical properties. As a consequence, the harmonic spatial periodicity is often believed to be its
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22 472 primary, almost synonymous, identifying factor. However, from a mechanical point of view, constant
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24 473 wavelength is not a necessary requirement. Folding is a compressional instability developing in stiff
25
26 474 layers embedded in a weaker material. The harmonic solution of the equilibrium and conservation
27
28 475 equations is a possible solution (Muehlhaus *et al.*, 1998; Burov & Molnar, 1998; Cloetingh *et al.*,
29
30 476 1999). The physical mechanism of folding is well understood. In a continuous layered medium the
31
32 477 stresses and strains must be continued across the interfaces between the layers. In multi-layer
33
34 478 systems with contrasting mechanical properties (e.g. strong and weak layers) this requirement may
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36 479 be difficult to satisfy, because the same amount of shortening in a stiffer layer would require much
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38 480 larger stress than in the neighbouring weaker layer. The system becomes unstable and, in
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40 481 attempting to reduce the stress and strain unconformities at the interfaces between the layers, starts
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42 482 to fold (or buckle) in response to very small perturbations. In non-elastic media with strain-
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44 483 dependent properties, locally increased flexural strain at the fold limbs can create weakened zones
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46 484 that significantly facilitate further deformation. These weak or softened plastic or viscous zones are
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48 485 often referred to as inelastic hinges, since the system easily folds at such weakened zones.
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50 486 Homogeneous shortening of the lithosphere under horizontal compression requires more work than
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52 487 required by an equivalent amount of shortening by folding. However, shortening by
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54 488 underthrusting/subduction of the lithosphere may be more mechanically efficient than folding but
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3 489 may be blocked or unable to start immediately after the onset of shortening. For this reason, folding
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5 490 is likely to be a primary and common response to tectonic compression. Folding probably may
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7 491 continue, in an attenuated form, even after the beginning of subduction (Cloetingh *et al.*, 1999), or
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10 492 re-appear when the subduction channel is locked or during the aftermath of collision (Cloetingh *et*
11
12 493 *al.*, 2004; Matenco *et al.*, 2007).

13
14 494 The rheological structure of the lithosphere (Watts & Burov, 2003) is a key factor in lithospheric
15
16 495 folding. Compressional stresses must build up to a level comparable with the integrated strength of
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18 496 the lithosphere in order to induce folding. Prior to the advent of rheological models based on
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20 497 extrapolation of laboratory experiments (Goetze & Evans, 1979; Carter & Tsenn, 1987) lithospheric
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22 498 stresses were thought to be incapable of reaching the failure levels required for folding (Turcotte &
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24
25 499 Schubert, 2002).

26
27 500 The simplest expression of the folding process is in oceanic lithosphere, characterized by an
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29 501 absence of a rheological stratification, and with thermo-mechanical ages spanning a limited age
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31 502 window with a maximum of 200 Myr (Geller *et al.*, 1993; Zitellini *et al.*, 2010). Because weak
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33 503 lithosphere is easier to fold than strong lithosphere, one would expect the first observations of
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35 504 lithospheric folding from the rheologically weak continental plate interiors. However, the observation
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37 505 of folding of the relatively strong lithosphere of the central Indian Ocean with a thermo-mechanical
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39 506 age of 80 Ma and with wavelengths of 200-250 km (Geller *et al.*, 1983; McAdoo & Sandwell, 1985;
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41 507 Stein *et al.*, 1989; Bull & Scrutton, 1992) triggered the interest in this mode of deformation. There
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43 508 may be two reasons for this. First, the recognition of folding, by mapping of basin reflectors and
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45 509 gravity anomalies requires spatial scales of several hundreds of kilometres. This is feasible with
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47 510 marine geophysical surveys, but generally in excess of the spatial scales covered by land surveys.
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49 511 An exception has been the systematic mapping of basins in Central Australia, located on strong
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51 512 lithosphere with a thermo-mechanical age of approximately 700 Ma (Lambeck, 1983; Stephenson &
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53 513 Lambeck, 1985). The Indo-Australian plate is under an exceptionally high level of compression
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55 514 (Stein *et al.*, 1989), as a result of its collision with the Eurasian plate. Under these conditions high
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3 515 stresses might have been more important than the rheology of the plate interior (Beekman et al.,
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5 516 1996; Gerbault *et al.*, 1998; Gerbault, 2000).

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7 517 Secondly, folding in oceanic lithosphere is not affected by surface erosion, making it relatively easy
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10 518 to recognize. The presence of only one dominant folding wavelength characteristic of non-stratified
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12 519 oceanic lithosphere also facilitated a quantitative interpretation. Following the early studies in the
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14 520 Indian Ocean and in Central Australia in the rear of the Indian-Eurasian collision, attention shifted to
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16 521 the other side of the plate contact in Central Asia. Large scale folding patterns were recognized
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18 522 through analysis of geophysical data, including gravity and topography (Burov *et al.*, 1993; Nikishin
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20 523 *et al.*, 1993; Burov & Molnar, 1998) as well as detailed geological studies (Cobbold *et al.*, 1993).

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22 524 On the basis of an inventory of currently available results of studies of folded continental
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24 525 lithosphere, it appears that in general the observed wavelengths of folding follow theoretical
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26 526 predictions for wavelengths as a function of thermo-mechanical age remarkably well (see **Fig. 3**).

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33 529 **NUMERICAL EXPERIMENTS FOR SELECTED THERMO-MECHANICAL AGES OF**
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36 530 **LITHOSPHERE**

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40 532 Folding is an unstable deformation that may develop in stratified media with significant competence
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42 533 contrasts, as occur in the continental lithosphere, in response to horizontal loading (Biot, 1961;
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44 534 Ramberg, 1961). These instabilities may develop under compression or basal shear as a result of
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46 535 strain/stress incompatibility on the interfaces between layers of different competence. In the initial
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48 536 stages of deformation, compressional instabilities cause periodic folds with an exponential growth
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50 537 rate and a dominant wavelength λ roughly proportional to 5 -10 times the thickness(es) of the
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52 538 competent layer(s). In the lithosphere, the layer thickness is that of the competent core h . So, $h =$
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55 539 10-100 km, gives rise to $\lambda = 50-1000$ km.
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3 540 "Biot's", or linear, folding encompasses the cases where compression of the lithosphere takes place
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5 541 from the beginning, leading to the formation of alternating (synclinal) basins and (anticlinal) highs,
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7 542 and where the conditions of the linear theory are more or less satisfied (e.g. thin layer
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9 543 approximation, no strain-softening, plane layers). For linear folding in a Newtonian medium, an
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11 544 asymptotic relationship derived from the thin layer equilibrium equation is (Biot, 1961; Ramberg,
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13 545 1961):

$$\lambda_1 = 2\pi h (\mu_{11}/6\mu_{12})^{1/3} \quad (1)$$

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17
18 546 Here λ_1 is the "Laplacian" dominant wavelength of folding in the absence of gravity, h is the
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20 547 thickness of the competent layer (crustal or mantle), and μ_{11} and μ_{12} are the effective viscosities (or
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22 548 competencies) of the strong layer and weak surrounding material, respectively. The above
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24 549 equation, derived assuming zero gravity, provides estimates of $\lambda_1/h = 20-40$ for typical competence
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26 550 contrasts. This does not hold for most lithospheric-scale cases, where $\lambda/h = 4 - 6$ is more common
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28 551 due to the influence of the gravity-dependent terms. A reduction of the normalized characteristic
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30 552 wavelength to values around 4 is also predicted for power-law rheology or plastic layers (Smith,
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32 553 1979). For a single stiff viscous layer on top of an inviscid medium, the dominant gravity-dependent
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34 554 wavelength is (Burov & Molnar, 1998):

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

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42 555 where $\Delta\rho$ is the density contrast, $\dot{\epsilon}$ is the strain rate, μ_{eff} is the layers' effective ductile viscosity
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44 556 (see Appendix A), n is the power-law exponent, and h is the thickness of the competent layer.

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47 557 At large amounts of shortening this deformation may become a-periodic (Hunt *et al.*, 1996) so that
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49 558 the wavelength and amplitude of folding vary along the deformed transect. Plume-induced periodic
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51 559 deformation (Burov & Cloetingh, 2009) also includes folding or boudinage caused by shear on the
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53 560 base of the lithosphere. However, the prime cause for folding of continental lithosphere is tectonic
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55 561 loading by horizontal far-field forces. For characteristic parameters of various layers within the
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3 562 continental lithosphere ($h = 10 - 120$ km, $\mu = 10^{19} - 10^{26}$ Pa s; $\Delta\rho = 10 - 600$ kg m⁻³), equations (1-
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5 563 2) predict folding wavelengths that may vary from 30 km (upper crustal folding) to as high as 600-
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7 564 800 km. The findings of numerical models (Cloetingh *et al.*, 1999) are in excellent agreement with
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9 565 analogue tectonic experiments (Sokoutis *et al.*, 2005). Numerical experiments show that the
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11 566 wavelength and amplitude of folding is largely controlled by the thermo-mechanical age and less by
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14 567 the rate of shortening (Burov & Cloetingh, 2009). Lithospheres older than 500 Ma develop nearly
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16 568 the same wavelength of folding as 1000 Ma old lithosphere because of the similar nearly stationary
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18 569 thermal state. Here, therefore, we restrict ourselves to a set of experiments for ages of 150 and 300
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21 570 Ma and convergence rates of 1.5 and 3 cm/yr.

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23 571 Surface topography and sedimentary deposition is described through diffusion-like erosion laws and
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25 572 fluvial transport laws (Gossman, 1976; Kirkby, 1986; Leeder, 1991; Beaumont *et al.*, 1992; Kooi and
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27 573 Beaumont, 1994):

$$dh/dt = \text{div}(k^*(x,h,\nabla h) \nabla h) \quad (\text{slope erosion})$$

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31
32 574 where h is topography, t is time, x is the horizontal coordinate, k^* is an experimentally established
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34 575 scale-dependent coefficient of erosion (500 – 1000 m²/y). For simplicity, we used zero order linear
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37 576 diffusion (k^* constant) for the short-range erosion and flat deposition for long-range fluvial transport
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39 577 (Avouac & Burov, 1996). The hill-slope erosion law ensures conservation of matter but holds only
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41 578 for small-scales. At large scales it is no longer applicable, and conservation of matter is not
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43 579 observed, because some sediments are lost due to transport out of the system by long-range
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46 580 processes.

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49 581 **Figure 9** shows the results of the experiments for relatively young (150 Ma) lithosphere compressed
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51 582 at rates of 1.5 and 3 cm/yr. Strongly decoupled low-amplitude short wavelength (mantle wavelength
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53 583 $\lambda = 250$ km) folding developed leading to crustal thickening above a synclinal mantle fold. At later
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55 584 stages, harmonic folding was followed by mega-folding (Cloetingh *et al.*, 1999; Burg &
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57 585 Podladchikov, 1999). For higher rates of shortening these mantle synclines developed in a mode

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3 586 resembling symmetric subduction. For lower convergence rates (**Fig. 9a**) folding does not develop
4
5 587 due to the low Peclet number of the system, leading to heat diffusion and weakening of folds.
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8 588 Folding is well developed for medium (300 Ma) age lithosphere (**Fig. 10**). This case is characterized
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10 589 by long mantle wavelengths ($\lambda=360$ km) and high surface amplitudes (2000 m after 10 Myr). These
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12 590 compare well with observations from Iberia (Fernandez-Lozano *et al.*, 2010), Western Goby and the
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14 591 Ferghana basin in Central Asia (Burov *et al.*, 1993; Burov & Molnar, 1998). At late stages (10 - 26
15
16 592 Myr since the onset of shortening for 3 cm/y or 20- 50% of shortening), folding becomes aperiodic,
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18 593 leading to mega-folding (Cloetingh *et al.*, 1999; Burg & Podladchikov, 1999) and subsequent
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20 594 formation of high-amplitude crustal down-warps. The amplitude of vertical movements may reach 20
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22 595 km (± 10 km) or even more. As pointed out earlier, such high amplitudes of vertical motions are
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24 596 observed in the South Caspian Basin (Guest *et al.*, 2007) and the Barents Sea (Ritzmann &
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26 597 Faleide, 2009). However, it may be relatively rare for folding to continue for periods in excess of 10
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28 598 Myr. More typically is that at a certain stage deformation localizes along single major fault zones
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30 599 (Cloetingh *et al.*, 1999; Gerbault *et al.*, 1998).
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34 600 As was discussed by Bird (1991), Avouac & Burov (1996) and Cloetingh *et al.* (1999), large-scale
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36 601 undulations of the lithosphere cannot be preserved for a long time (longer than 10 Myr) in the
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38 602 absence of sufficient compression, except for plates with very strong (especially lower crustal)
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40 603 rheology. Otherwise, the folds either will be flattened by gravity-driven crustal flow associated with
41
42 604 the large crust-mantle density contrast at the Moho, or deformation will localize along some of the
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44 605 faults created at the inflection points of folds.
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47 606 The fact that folding with wavelengths of 400-500 km and 500-600 km persists long after cessation
48
49 607 of the compression in the presumably high-rigidity Australian craton and Russian platform points to
50
51 608 a strong rheology yielding a high value for the effective elastic thickness EET (in excess of 60 km),
52
53 609 as expected from its thermo-mechanical age (Beekman *et al.*, 1997). The long wavelengths of 400-
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55 610 600 km of crust-mantle down-warping in the eastern Barents Sea, where Mid-Cenozoic
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3 611 compression has affected a basin system formed during Late Permian-Triassic time (Ritzmann &
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5 612 Faleide, 2009), with a total infill of sediments of the order of 15-20 km is another impressive
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7 613 example of compressional deformation on relatively cold lithosphere.

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10 614 Most of the areas with present-day long-wavelength continental lithosphere folding formed during or
11
12 615 short after the Alpine collision 60 Myr ago. It thus seems that 60 Ma is a characteristic timescale of
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14 616 gravity collapse of the large-scale folds in such intermediate-age lithospheres. In the case of very
15
16 617 weak quartz-dominated lower crust, folds may disappear rapidly within 8-15 Myr. Experiments
17
18 618 (Cloetingh *et al.*, 1999) for 1000 Ma old lithosphere, with a diabase lower crust, demonstrate that
19
20 619 folding-induced deformation persists for 10-50 Myr following the cessation of compression. During
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22 620 this period the amplitude of the folds decreased by less than 10 percent. At this stage, crustal faults
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24 621 may accelerate the gravity collapse of folds, creating inverted basins.

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30 31 624 **BASIN GEOMETRY AND ACCOMMODATION SPACE**

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34 35 626 **Basin shape**

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37 627 Folding of the lithosphere leads to a symmetrical pattern of down-warped areas (synclines) flanked
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39 628 by highs (anticlines) of similar amplitude and wavelength, like for instance the Ferghana basin in
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41 629 Central Asia (Burov & Molnar, 1998).

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44 630 This symmetry is in marked contrast to the asymmetrical shape of foreland basins, formed by
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46 631 flexure in front of an orogenic wedge, which are flanked by a flexural bulge of an amplitude that is
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48 632 only up to 10 percent of the maximum depth of the foreland depression (Royden, 1988; Zoetemeijer
49
50 633 *et al.*, 1999). Both types of compressional basins show linear nature alignment of parallel
51
52 634 depocentres and flanking highs. An important difference is the presence of parallel trending
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54 635 depocentres for lithospheric folding, whereas foreland flexural basins have a single depositional
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56 636 system. In addition, basins developed on folded lithosphere have a static location of the axis of their

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3 637 depocentres, in contrast to foreland basins, where the axis of depocentres can migrate with time
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5 638 (e.g. Zoetemeijer *et al.*, 1993).

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7 639 In both cases, the integrated strength of the lithosphere (Watts and Burov, 2003; Tesauro *et al.*,
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9 640 2007) defines the characteristic width of the basins. For folding, basin width roughly corresponds to
10
11 641 one half of the fold wavelength (λ), which equals 5 to 10 times the thickness of the strong core of
12
13 642 the lithospheric plate, h ($h = 10$ to 100 km, i.e. $\lambda = 50 - 1000$ km). In the case of crust-mantle
14
15 643 decoupling, which occurs in relatively young plates, two dominant folding wavelengths develop
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17 644 (typically $250 - 400$ km and $50-100$ km). Thus two basin populations may be observed, one
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19 645 imbricated within the other. In the case of flexural foreland deformation, the width of the basin is
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21 646 controlled by the flexural parameter and is thus of the same order as the largest fold wavelength.
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28 648 **Subsidence patterns**

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30 649 Lithospheric folding results from an instability in the lithosphere due to stress/strain incompatibilities
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32 650 that develop in rheologically stratified layers under compressive strain. This process typically
33
34 651 operates with timescales of the order of 1-10 Myr (Cloetingh *et al.*, 1999). The preservation of
35
36 652 lithospheric folds appears to depend strongly on the thermo-mechanical age of the underlying
37
38 653 lithosphere (Cloetingh *et al.*, 1999). Folds in young lithosphere will not be preserved upon relaxation
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40 654 of the stress field, whereas ones in cratonic lithosphere will be well preserved (Cloetingh *et al.*,
41
42 655 1999).

43
44 656 **Figure 11a** displays characteristic basin subsidence patterns predicted by folding of 150 and 300
45
46 657 Ma old lithosphere due to a 3 cm/year shortening rate. As illustrated (**Figure 11b**), the differential
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48 658 motions occur in three distinct phases:

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51 659 **Stage 1** is the basin formation phase, coinciding with the initiation and development of folding,
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53 660 which is marked by an acceleration of subsidence in the basin and uplift in the flanking highs. As a
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55 661 result of the time lag between sediment supply and the creation of accommodation space, during
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57 662 the few Myr needed to form the basin, deposition cannot keep up with subsidence causing
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3 663 sediment-starved basins. A similar pattern occurs in pull-apart basins (Pitman & Golovchenko,
4
5 664 1983) which are also associated with ultra rapid subsidence in their formation stage also lasting only
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7 665 a few Myr.

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10 666 **Stage 2** is the basin preservation stage, where equilibrium develops between sediment supply and
11
12 667 sediment deposition. As a result the basin will be rapidly filled to overfilled.

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14 668 **Stage 3** is the basin destruction phase, characterized by basin capture and removal of sediments to
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16 669 areas outside the folding system. During this phase, the size of the accommodation space is
17
18 670 reduced and erosion occurs on both flanks and depocentre. Thus, the net effect of these three
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20 671 stages is that lithospheric folding will lead to the development of distinct depositional-erosional
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22 672 cycles.

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25 673 Characteristic patterns for vertical motions for folding of continental lithosphere of 150 and 300 Ma
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27 674 old (**Fig. 11**) demonstrate a remarkably short time scale in which substantial amounts of tectonic
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29 675 subsidence are induced by the process of lithospheric folding. As shown by **Figure 11**, within 1 Myr
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31 676 after the initiation of folding induced by shortening the lithosphere at a rate of 3 cm/yr, subsidence
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33 677 rates are in the order of 5-15 km/Myr, depending on the thermo-mechanical age of the lithosphere.
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35 678 Thus, folding appears to be more effective in middle aged lithosphere of 300 Ma than in lithosphere
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37 679 of younger ages (Cloetingh *et al.*, 1999; Burov & Cloetingh, 2009). After 2 Myr, a slow uplift phase
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39 680 starts of relatively minor magnitude of the order of a few hundreds of meters to a km. The predicted
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41 681 subsidence in **Figure 11** is for the centre of the syncline. As pointed out above, erosion of the
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43 682 uplifted flanks is taken into account, adopting a diffusive equation approach to erosion (see Burov &
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45 683 Cloetingh, 1997; Cloetingh *et al.*, 1999, for further details).

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48 684 The alignment of parallel highs and basins of similar dimensions has an important consequence for
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50 685 the areal extent of the source areas for sediments available for deposition in the folded depressions.

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52 686 In comparison with foreland basins the distribution of sources is more symmetrical, superseding the
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54 687 volume of sediments that can be eroded from, for example, the flexural foreland bulge. As noted
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56 688 earlier, erosion reduces the contribution of gravity-dependent terms and accelerates local
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3 689 deformation. Erosion, therefore, has an important feed-back with the geometry of the
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5 690 accommodation space in changing the spectrum of wavelengths (Cloetingh *et al.*, 1999). Erosion
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7 691 acts as a filter, suppressing the shorter wavelengths in folded basin topography. Strong erosion,
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9 692 insufficiently compensated by tectonic deformation, can even wipe out most of the topography.
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11 693 However, if the erosion is tuned to the average elevation rates (**Fig. 12**), it may dramatically
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13 694 accelerate folding (Cloetingh *et al.*, 1999).

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16 695 There has been a rapid development and application of stratigraphic modelling packages to
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18 696 sequence stratigraphy over the last two decades (e.g. Cross, 1989; Alzaga-Ruiz *et al.*, 2008).
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20 697 Among the different types of modelling programs, the forward stratigraphic modelling approach has
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22 698 been applied extensively by basin modellers from academia and petroleum industry (e.g. Lawrence
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24 699 *et al.*, 1990; Granjeon, 2009). Forward stratigraphic modelling commonly starts with an estimate of
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26 700 the initial conditions related to an adopted basin formation mechanism and controlling parameters,
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28 701 followed by a forward prediction in time. Most of these software packages are based on sequence
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30 702 stratigraphic concepts (Wilgus *et al.*, 1988). Assumptions for initial subsidence are either ad-hoc or
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32 703 related to stretching (e.g. Lawrence *et al.*, 1990) or foreland flexure (Garcia-Castellanos *et al.*,
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34 704 1997), but to date do not include lithospheric folding as a basin formation mechanism.
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40 706 **Marine versus continental deposits and closed basins versus capture of drainage**

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42 707 Erosion of uplifted areas and sedimentation in the depressions created by folded lithosphere is a
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44 708 self reinforcing process promoting continuing uplift of the highs and subsidence in the depocentres.
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46 709 Thus, such basins will remain closed during much of their evolution. Folded basins will hence be
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48 710 predominantly characterized by continental deposits with only minor deposition of marine sediments
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50 711 during marine incursions or during basin capture. This seems to be the case for folded basins in
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52 712 Iberia (e.g. Duero Basin) and Central Asia (Ferghana, Tarim Basin, Lake Issyk-Kul) (Cobbold *et al.*,
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54 713 1993). As has been pointed out for flexural foreland basins (Garcia-Castellanos *et al.*, 2003), the
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56 714 capture time for opening basins decreases with increasing flexural rigidity. The flexural upwarp to
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3 715 restoring the topography removed by erosion will be almost instantaneous for very weak
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5 716 lithosphere, but requires up to 50-100 Myr for high lithospheric rigidities (Garcia-Castellanos *et al.*,
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7 717 2003). Thus, basins created by folding of cratonic lithosphere probably have a shorter capture time
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9 718 than ones formed in younger lithosphere.

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14 720 **THERMAL REGIME, STRESS REGIME AND STYLE OF FAULTING**

16 721

18 722 **Thermal regime**

20 723 Lithospheric folding is controlled by the interplay of lithospheric stresses and inherited strength of
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22 724 the lithosphere. The thermal regime controls the rheological profile and differs between folded
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24 725 basins developed in young lithosphere and basins in cratonic lithosphere. The latter are associated
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26 726 with much lower thermal gradients than basins developed on young continental lithosphere. As for
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28 727 foreland basins, the initiation of folding is not associated with a thermal instability, unless folding
29
30 728 occurs in conjunction with plume activity (Burov & Cloetingh, 2009). Following the deposition of
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32 729 radiogenic sediments in the folded depression, sediment blanketing will affect the heat flow
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34 730 (Stephenson *et al.*, 1990; Lavier & Steckler, 1997; Van Wees *et al.*, 2009), modifying the surface
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36 731 heat flow in the basin centre. As pointed out by Lavier & Steckler (1997) and Ziegler *et al.* (1998),
37
38 732 the effect of sediment fill is to weaken the lithosphere. The low thermal conductivities of the
39
40 733 sediments lead to high temperatures in the upper lithosphere and consequently low local yield
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42 734 strength. During basin subsidence sediments deposited in the central parts of the basin might be
43
44 735 exposed to temperature windows corresponding to hydrocarbon generation. At the same time,
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46 736 sediments pre-dating the folding may have undergone extra burial due to syn-folding sedimentation.
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48 737 An example is the Paris basin, where source rocks of Toarcian age were folded in Cenozoic time
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50 738 (Guillocheau *et al.*, 2000; Le Solleuz *et al.*, 2004).

51 739 During basin capture overall cooling takes place. This effect is illustrated by characteristic thermal
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53 740 evolution for different thermo-tectonic ages of 150 and 300 Ma of the folded continental lithosphere,
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3 741 incorporating radioactive heat production in the sediments filling in the synclinal depression, shown
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5 742 in **Figure 13**. Subsidence induced by folding is calculated for the centre of the basin. Basins
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7 743 developing on folded lithosphere are characterized by their relatively low heat flow at the onset of
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9 744 folding, followed by a steady increase with time, doubling the heat flow over a time interval of the
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11 745 order of 5 Myr following the cessation of shortening. This increase is primarily due to the
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13 746 contribution of radioactive heat production in the very substantial pile of sediments accumulating in
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15 747 a relatively short time interval. The patterns of heat flow are similar for different thermo-mechanical
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17 748 ages of the lithosphere, with the oldest lithosphere having the lower heat flow. Older ages enhance
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19 749 the accommodation space and the contribution of the sediments to heat flow but are compensated
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21 750 by a larger decrease in heat flow with aging lithosphere. Another important factor is the mode of
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23 751 folding. Crustal scale folding, which is characteristic for Lake Issyk-Kul, will be associated with
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25 752 shorter wavelengths, shallower basin depths and thinner sequences of heat producing sediments
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27 753 than for mantle lithospheric folding. This effect might explain the relatively low heat flow observed in
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29 754 Lake Issyk-Kul (Vermeesch *et al.*, 2004).
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34 756 **Modes of brittle deformation and faulting**

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37 757 The role of pre-existing faults on the development of folding has been investigated through
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39 758 numerical experiments on folding of brittle-elasto-viscous lithosphere (Beekman *et al.*, 1996;
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41 759 Gerbault *et al.*, 1998; Cloetingh *et al.*, 1999) and through analogue experiments on plastic-elastic
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43 760 lithosphere (e.g. Martinod & Davy, 1994; Sokoutis *et al.*, 2005). These studies show that lithospheric
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45 761 folding and faulting can develop simultaneously and that pre-existing crustal scale faults do not
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47 762 prevent, but instead promote the development of folding instabilities. The instabilities, in turn, can
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49 763 initiate the formation of new faults at the inflection points of the folds. Although the usual expectation
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51 764 is that faulting prevents folding, this view neglects the role of the gravity and friction on faults in the
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53 765 behaviour of the large-scale fault-and-fold systems. As also observed in analogue experiments
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55 766 (Martinod & Davy, 1994), this continuous behaviour of faulted lithosphere can be explained by fault
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3 767 locking due to gravity and friction. After some sliding (uplift) on a fault, the potential gravitational
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5 768 energy working against friction generated by horizontal shortening forces becomes too high. As a
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7 769 result, the fault locks and the rock mass transmits horizontal stresses as a continuous medium. As
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10 770 the compression continues, one of the folds finally starts to grow faster than others, resulting in a
11
12 771 loss of the periodicity and the formation of a mega-fold (see also the discussion in the context of the
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14 772 Ferghana and Tadjik basins of Central Asia).

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16 773 Following the initiation of folding, and the further build up of compressional stresses, folding will,
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18 774 therefore, usually be accompanied by brittle deformation and faulting in the crust and upper
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20 775 lithosphere. The surface expression of folding in uniform lithosphere is frequently in the form of
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22 776 fault-controlled pop-up and pop-down structures and inverted basins (e.g. Cobbold *et al.*, 1993),
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25 777 accompanied by development of imbrications of bivergent thrust wedges at the areas where
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27 778 shortening is applied (Sokoutis *et al.*, 2005). These findings (see **Fig. 14**) are corroborated by
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29 779 observations in the Central System of Iberia, where folding of Variscan lithosphere with a thermo-
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31 780 mechanical age of 280 Ma occurs (Cloetingh *et al.*, 2002). Here the brittle deformation associated
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33 781 with crustal scale folding expressed by a folded Moho is mechanically decoupled from folding of the
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35 782 mantle lithosphere (Fernandez-Lozano *et al.*, 2010; see **Fig. 5b**).

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37 783 Crustal and mantle faults may develop as a result of folding (Burov & Molnar, 1998) (**Fig. 15**).
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39 784 Folding can continue after these faults develop and folding and faulting can co-exist for times of
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41 785 several Myr (**Fig. 5b**). For crust-mantle decoupling, a complex distribution of crustal faults and
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43 786 surface undulations can be expected, with crustal faults not necessarily connected to mantle faults.
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46 787 Numerical experiments suggest that mantle faults are unlikely to be longer than a few km in depth
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48 788 (Burov & Molnar, 1998). Such faults may also not develop at longer wavelengths of folding
49
50 789 connected with stronger lithosphere. However, deep seated faults in mantle lithosphere can develop
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52 790 when cratonic lithosphere is folded (Cloetingh *et al.*, 1999). The mode of faulting, depth extent and
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54 791 spacing of faults depend on the thermo-mechanical age of the lithosphere. Once the development
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56 792 of folding begins in a basin, the intensity and depth penetration of faulting will accelerate. The mode
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3 793 of decoupling between crustal and mantle folding is crucial. Discontinuities within the folded
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5 794 lithosphere will localize the development of deep synclines, flanked by more modest anticlines
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7 795 (Sokoutis *et al.*, 2005) and affect the distribution of faulting (**Fig. 14**).

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10 796 During the next phase of basin evolution, when a steady state is reached between the creation of
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12 797 accommodation space and sediment supply, only minor faulting will occur. This is in marked
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14 798 contrast with the later phase of basin capture, which is in general characterized by erosion and
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16 799 narrowing of the basin. Due to erosional unloading, stress relaxation will occur, manifested in
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18 800 upward propagating faults.

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22 802 **Interaction of folding with other tectonic processes**

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25 803 Folding, in addition to being a basin forming mechanism, frequently interacts with other tectonic
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27 804 processes. As pointed out by Cloetingh (1988), intraplate compression can modify pre-existing
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29 805 basins. As discussed above, the Pannonian Basin of Central Europe created by Miocene back-arc
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31 806 extension appears to be an example of such a configuration, characterized by a-typical a-periodic
32
33 807 folding. Other examples are cratonic sag basins where intraplate compression (Cloetingh, 1988) is
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35 808 thought to occur in interaction with phase changes in the lithosphere (Artyushkov, 2007). Examples
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37 809 of these might be the Barents Sea (Ritzmann & Faleide, 2009) and possibly the South Caspian
38
39 810 Basin, although in the latter case pre-orogenic extension cannot be ruled out (Guest *et al.*, 2007).
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41 811 Basins exposed to changes in tectonic regime will have a polyphase record (Cloetingh & Ziegler,
42
43 812 2007), characterized by a superposition of more than two regimes, such as pre-orogenic extension,
44
45 813 foreland flexure and late-stage folding. This sequence appears to be characteristic for some very
46
47 814 deep foreland basins, such as the Focsani depression of the Romanian Carpathians, with more
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49 815 than 16 km of sediments, which was affected by extension due to the opening of the Black Sea
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51 816 Basin, followed by foreland flexure and overprinted by Late Miocene compression (Tarapoanca *et*
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53 817 *al.*, 2003). As pointed out above (Ziegler & Dèzes, 2007), the Northwestern European foreland
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55 818 indicates folding, overprinting, rifting and foreland flexure. Quantification of the topographies created
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3 819 by rifting and foreland flexure and correcting for them is essential to reconstruct the shape of the
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5 820 additional accommodation shape created by subsequent lithospheric folding (Bourgeois *et al.*,
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7 821 2007).

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10 822 The impingement of plumes on the base of continental lithosphere can induce differential
11
12 823 topography similar to the surface deflections characteristic for lithosphere folds (Guillou-Frottier *et*
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14 824 *al.*, 2007; Burov *et al.*, 2007). Plumes and lithospheric folds frequently interact in space and time in
15
16 825 the geological record (Ziegler & Dèzes, 2007; Burov & Cloetingh, 2009). Therefore, the evolution of
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18 826 folded lithosphere basins can be overprinted by the signatures of a plume or vice-versa. This will
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20 827 lead to an amplification of the induced vertical motions, particularly significant for young and
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22 828 intermediate age lithosphere. Emplacement of hot upper mantle material will raise temperatures in
23
24 829 the lithosphere and increase heat flow. Due to the lag time in the propagation of heat, the effect at
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26 830 the surface might become manifest only after several tens of Myr. Plume emplacement might also
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28 831 weaken the lithosphere, making it more prone to compressional fault reactivation after cessation of
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30 832 the folding.

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33 833 Interplay of lithosphere folding and plume impingement on the continental lithosphere occurred
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35 834 probably almost simultaneously in late Neogene times in the Northwestern European foreland. Two
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37 835 examples are the Eifel and Massif Central areas of the Alpine foreland of NW Europe (Cloetingh &
38
39 836 van Wees, 2005; Ziegler & Dèzes, 2007). Both areas are sites of main Late Neogene volcanic
40
41 837 activity in the European Cenozoic rift system (ECRIS). Seismic tomography (Ritter *et al.*, 2001)
42
43 838 shows finger shaped baby plumes with a characteristic spatial diameter of 100 km, extending
44
45 839 downward to 400 km. Ziegler & Dèzes (2007) propose that plume activity occurs simultaneously
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47 840 with recent compressional deformation in the Massif Central area (Guillou-Frottier *et al.*, 2007) and
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49 841 the Ardennes/Eifel area. Geomorphological studies constrain the recent uplift of the Ardennes and
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51 842 the Eifel area. An order in magnitude difference occurs between the uplift of the Eifel area, underlain
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53 843 by a plume and the adjacent Ardennes area where evidence for a plume is lacking. The patterns of
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55 844 uplift appear to be radial, superimposed by a linear NE-SW trend perpendicular to the main axis of
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3 845 compression. Baby plumes primarily develop in the anticlines of lithospheric folds (Burov &
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5 846 Cloetingh, 2009). The plume activity presumably accelerated the rate of uplift by a factor of 3-5
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7 847 (Cloetingh & Ziegler, 2007). According to presently available data, the plumes in this segment of the
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9 848 European Cenozoic Rift System arrived about 1 Myr ago. This activity was preceded and followed
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11 849 by lithospheric folding that continues to present-day since 17 Myr ago (Bourgeois *et al.*, 2007).
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13 850 Thermo-mechanical modelling illustrates the relative effectiveness of amplification of lithosphere
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15 851 deformation and topographic effects induced by plumes through folding and vice-versa. Burov &
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17 852 Cloetingh (2009) examined the response times and time-lags involved and whether these baby
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19 853 plumes were more efficient in localizing deformation than large plumes. Plume-affected folding
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21 854 appears to accelerate surface uplift, whereas folding goes into saturation and stagnates when
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23 855 plastic hinges form. For the discrimination of plumes and folding it is critical to access constraints on
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25 856 the presence or absence of radial versus linear symmetry, heat flow anomalies, gravity and geoid
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27 857 data. Plume activity facilitates folding, by dramatically lowering the stress levels required (Burov &
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29 858 Cloetingh, 2009). Plume impact also reduces the fold wavelength and localizes folding above the
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31 859 plume impact area. A general outcome of the modelling and observations is that lithospheric folding
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33 860 as a mechanism for producing thermal perturbations in the lithosphere/upper mantle system is a
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35 861 less feasible scenario.
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42 863 **FOLDING AND OTHER MODES OF BASIN FORMATION: DIFFERENCES AND SIMILARITIES**
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44 864 **IN STRUCTURE AND EVOLUTION**

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46 865 Sedimentary basins formed through lithospheric folding have characteristic features including
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48 866 vertical motions, basin architecture, thermal regime and fault activity, distinguishing them from other
49
50 867 basin types (Cloetingh & Ziegler, 2007; Xie & Heller, 2009). **Figure 1b** and **Table 2** illustrate
51
52 868 differences and similarities between basins developed on folded lithosphere (FLB), foreland basins
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54 869 (FB), intracratonic basins (ICB), extensional basins (EB), and pull-apart basins (PAB).
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3 870 Sedimentary basin systems are, by their nature, prone to tectonic reactivation and, therefore,
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5 871 frequently characterised by a poly-phase evolution (Cloetingh & Ziegler, 2007). Extensional basins
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7 872 are, for example, often formed in areas previously thickened by tectonic compression and
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10 873 subsequently often subject to late-stage compression (Lundin & Dorè, 1997). Examples include the
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12 874 Pannonian basin of Hungary (Horváth & Cloetingh, 1996) and the North Atlantic rifted margins
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14 875 (Cloetingh *et al.*, 2008). Similarly, foreland basins are frequently characterised by pre-orogenic
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16 876 extension. Examples include the Carpathian foredeep (e.g. Tarapoinca *et al.*, 2003) and the
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18 877 Aquitaine basin of Southern France, which was the retro-arc foreland basin of the Pyrenees
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20 878 (Desegaulx *et al.*, 1991). The actual subsidence patterns of these poly-phase systems are often
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22 879 more complex than predicted by the end-member models discussed below that only consider the
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24 880 basin formation mechanism. Retro-arc foreland basins are characterised by a prolonged subsidence
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26 881 history with a distinct thermal signature including the effects of thermal cooling inherited from pre-
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28 882 orogenic extension. This is sometimes neglected in comparing pro-arc and retro-arc foreland basins
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30 883 (Naylor & Sinclair, 2007).

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33 884 Basin infill records allow main features, including geometry, vertical motions and faulting
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35 885 characteristics to be unravelled (Zoetemeijer *et al.*, 1993). However, heat flow histories need to be
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37 886 determined through kinematic reconstruction and forward modelling. Results of heat flow history
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39 887 modelling for FB and EB settings (Van Wees *et al.*, 2009) can be subsequently compared to FLB
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41 888 results presented in this paper (**Fig. 13**). For ICB and PAB we have refrained from a quantitative
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43 889 approach for heat flow predictions as vertical motions driving subsidence cannot be linked in a
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45 890 straightforward way to a single or uniformly distributed lithosphere process. As a consequence, for
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47 891 these settings tentative heat flow patterns have been drawn.

50 892 *Folded lithosphere basins* (FLB) that develop as a result of periodic folding are highly symmetrical,
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52 893 with dimensions that can vary from 50 km to 600 km, depending on lithospheric age and shortening
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54 894 rate. These basins can accumulate thick sedimentary sequences up to the order of 20 km. The time
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56 895 scales associated with this process of basin formation are very short, typically a few Myr. The
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3 896 subsidence is so fast that erosion and sediment supply at the basin formation stage typically lags
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5 897 behind, leading to the development of starved basins, followed by shallowing upward sequences
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7 898 after stress relaxation. Subsidence patterns are characteristically convex upward with time. A
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9 899 noticeable feature is the development of significant topography of the order of several km flanking
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11 900 the synclinal depression. No initial heating is predicted, with an increase in heat flow with time
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13 901 following the basin formation. Significant brittle deformation is expected to occur in the folded
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15 902 basement, accelerating during the folding phase.

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18 903 *Foreland basins* (FB) are strongly asymmetrical, with depths usually less than 10 km, and have
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20 904 modest flexural bulges of a few hundred meters (Royden, 1993; DeCelles & Gilles, 1996;
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22 905 Zoetemeijer *et al.*, 1999). Important differences occur in the nature of the subsidence of pro- and
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24 906 retro-foreland basins (Naylor & Sinclair, 2007). The first are typically associated with short life spans
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26 907 and convex up subsidence patterns with time, whereas the latter show more prolonged subsidence
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28 908 histories and concave up subsidence patterns. Subsidence patterns will be interrupted during
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30 909 thrusting phases, also leading to uplift of the flexural marginal bulge (Quinlan & Beaumont, 1984;
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32 910 Zoetemeijer *et al.*, 1993). The predicted heat flow pattern of a pro-foreland setting shown for the
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34 911 Ebro foreland basin (well Jabali from Verges, 1999), marked by rapid basement subsidence and
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36 912 sedimentary infill of ca 4 km in ca 10 My period (other parameters as in Van Wees *et al.*, 2009).
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38 913 Flexure and rapid sediment infill is marked by a reduction of heat flow, followed by a progressive
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40 914 increase related to radiogenic heat production of foreland sediments. Basement fracturing is
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42 915 concentrated at the points of maximum curvature in the flexed lithosphere, at the maximum points of
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44 916 upward and downward deflection, respectively, with intensity higher than inferred for FLB.

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47 917 *Intracratonic basins* (ICB) share their symmetrical shape and great thickness of their sedimentary
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49 918 basin fill with FLB. An important difference is their evolution in time, characteristically associated
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51 919 with very long time spans (typically of the order of 500-800 Myr) with long phases of low subsidence
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53 920 rate punctuated by phases of subsidence acceleration. As pointed out above, this is why ICB are
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55 921 probably affected by short-term phases of enhanced intraplate compression superimposed on other
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3 922 mechanisms operating on longer time scales. The large dimensions of these basins point to
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5 923 formation in high rigidity lithosphere, that freezes in basin deformation over long time scales. The
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7 924 long time scales involved also imply that the highs induced by compressional upwarping will be
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9 925 commonly eroded. A major difference between ICB and FLB is that current hypotheses for ICB
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11 926 formation are casted only in terms of the tectonic processes operation at their earlier stage such as
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13 927 phase changes in the lithosphere. It appears, therefore, that lithospheric stress fields and the
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15 928 lithosphere rheology (inferred from gravity and flexure) are far more diagnostic than the lithosphere-
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17 929 asthenosphere boundary, which is only giving information on the thermal state of the lithosphere.
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19 930 Heat flow patterns of ICB will be dominated by sediment infill. ICB are in general characterized by a
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21 931 low intensity of faulting. However, in the absence of a rigorous understanding of the formation
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23 932 mechanisms of ICB, predictions for heat flow are often not available. As a consequence, their heat
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25 933 flow is most often described primarily in terms of the thermal consequences of their burial history.

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27 934 *Extensional basins* (EB) have received major attention from basin modellers, both because of their
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29 935 relative abundance and importance and because of the availability of a relatively simple quantitative
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31 936 model for their formation (McKenzie, 1978). The latter can be relatively easily be incorporated in
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33 937 forward modelling of their subsidence and heat flow histories (Van Wees *et al.*, 2009). The basin
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35 938 shape is symmetrical or asymmetrical depending on a pure shear or simple shear mode of
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37 939 extension. Rifts can be narrow or wide, depending on the mode of rifting and the rheology of the
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39 940 extending lithosphere (Huisman and Beaumont, 2002). Dependent on the rigidity of the extending
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41 941 lithosphere, rift shoulders with asymmetrical shape can be formed and be prone to erosion (Van
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43 942 Balen *et al.*, 1995, Van der Beek *et al.*, 1995). The development of rift shoulders and their erosion
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45 943 can lead to feed back with lower crustal flow and faulting in the extending lithosphere (Burov &
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47 944 Cloetingh, 1997). During the post-rift phase, flexural widening of the basin commonly occurs. A
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49 945 characteristic feature of EB predicted by simple stretching models (McKenzie, 1978) is the
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51 946 succession of rapid initial subsidence in the syn-rift phase and exponentially decaying subsidence in
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53 947 their post-rift phase with a typical thermally controlled time constants of the order of 70 Myr. Feed
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3 948 backs between rift shoulder erosion and lower crustal flow (Burov & Cloetingh, 1997; Burov &
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5 949 Poliakov, 2001), have pronounced effects on subsidence in the final phases of rifting and in the
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7 950 early post-rift phase. Both can be influenced by changes in the mechanical properties of the
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9 951 lithosphere and its strengthening upon rifting. Multiple rifting phases is another common feature of
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11 952 EB. The duration of the syn-rift phase can vary from quasi-instantaneous (shorter than 20 Myr) to
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13 953 several hundreds of Myr (Cloetingh and Ziegler, 2007). In the latter case basin migration is a
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15 954 common feature (Corti *et al.*, 2003). Extensional faulting occurs during the rift phase, with absence
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17 955 of faulting in the post-rift phase. Heat flow reaches a peak during the syn-rift phase, followed by a
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19 956 rapid decrease in heat flow (Van Wees *et al.*, 2009). The syn-rift peak in heat flow is low or almost
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21 957 absent in continental extensional settings as a consequence of removal of heat producing elements
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23 958 in the crust and cooling by sediment infill. **Fig. 16** shows such a setting, which is based on the
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25 959 benchmark model presented in Van Wees *et al.* (2009), As a consequence of the loss of radiogenic
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27 960 heat production in the crust at the end of the post-rift stage values, heat flow evolves to values
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29 961 considerably lower than the initial ones. The EB heat flow pattern is, therefore, the reverse of that in
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31 962 FB where syntectonic heat flow is first lowered due to sediment infilling (Kombrink *et al.*, 2008), but
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33 963 eventually increases due to thermal relaxation and additional radiogenic heat production in the
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35 964 sediments.

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39 965 *Pull-apart basins* (PAB) are characterized by a narrow fault bounded geometry (Smit *et al.*, 2008),
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41 966 developing over a very short time span. Models have been developed to calculate subsidence
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43 967 histories and heat flow histories for these basins (Pitman & Golovchenko, 1983), incorporating
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45 968 lateral heat transport in stretching models. These models predict rapid subsidence of several km's
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47 969 over time intervals of the order of 0.1 – 1 Myr, followed by a decay in the subsidence with a concave
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49 970 shape upward. Erosion and sediment flux can not keep pace with the subsidence at the basin
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51 971 formation phase, leading initially to a starved basin, followed by a shallowing upward sequence
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53 972 (Pitman & Golovchenko, 1983). These basins are intensively faulted, also due to their strike-slip
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55 973 origin, whereas heat flow decays on very short time scales of the order of a few Myr. Although this
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3 974 scenario is a good first order description, many PAB show a more punctuated fault controlled
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5 975 subsidence history, often terminating by late-stage uplift, as observed for Late Neogene pull-apart
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7 976 basins in the internal zone of the Betics in SE Spain (Cloetingh *et al.*, 1992) and for the flanks of
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10 977 pull-apart basins in the Dead Sea Fault System (Smit *et al.*, 2008, 2010).

11 978 From the above overview, it is apparent that FLB have in common with FB a compressional origin
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14 979 and thermal histories controlled to a large extent by burial histories. At the same time, pronounced
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16 980 differences occur in the duration of subsidence, varying from ultra short (FLB 1-10's Myr), medium
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18 981 (FB, tens of Myr) and ultra long (ICB, hundreds of Myr). An important difference occurs in the spatial
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20 982 scales of the topography generated by folding and foreland fold-and-thrust belt deformation. In the
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22 983 latter case a much steeper differential topography is generated over typical horizontal differences of
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24 984 the order of 10 km between peak and trough. In contrast, in the case of lithospheric folding
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26 985 horizontal distances between the topographic highs and the axes of the depocentres are typically of
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28 986 the order of hundreds of kilometres. As a result, diffusive transfer of mass will occur at a much
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30 987 slower rate of sediment supply in the case of folded lithosphere than rates of sediment supply
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32 988 locally generated by thrusting. This together with a very rapid initial subsidence of folded lithosphere
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34 989 basins has important consequences for the paleobathymetry of these systems.

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37 990 FLB contrast with extensional and pull-apart basins in the shape of the subsidence curves, being
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39 991 convex up for FLB and concave for EB and PAB. The heat flow histories differ also. EB and PAB
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41 992 are characterized by a phase of initial heating, whereas FLB and FB are characterized by an
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43 993 absence of initial heating. A common feature of FLB and PAB formation is their capability to cause
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45 994 dramatic differential vertical motions in a short time interval, predicting shallowing upward
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47 995 sequences. FLB are to some extent unique in terms of the magnitude of the uplifted topography
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49 996 surrounding the down warped depocentres. Similar magnitudes are only reached at rift shoulders of
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51 997 basins developing in intracratonic lithosphere (Van der Beek *et al.*, 1995). FLB can be associated
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53 998 with localized brittle deformation and pop-up structures in the upper crust and widespread brittle
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55 999 deformation at deeper crustal and upper mantle levels.
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7 1002 **CONCLUSIONS**8
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10 1004 Sedimentary basins controlled by lithospheric folding are characterized by a number of features in
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12 1005 their basin architecture, subsidence history, thermal evolution and faulting patterns making them
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14 1006 distinctly different from other basin systems. Of particular importance are the relatively short
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16 1007 temporal scales involved in their evolution, whereas their spatial scales vary from tens to several
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18 1008 hundreds of km, depending on the rheological stratification of the lithosphere and its thermo-
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20 1009 mechanical age. Subsidence patterns are characterized by an acceleration of subsidence in the
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22 1010 depocentres during compressional basin formation, occurring simultaneously with accelerated uplift
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24 1011 of and erosion from flanking basin highs. In contrast with asymmetrical foreland basins, basins on
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26 1012 folded lithosphere are symmetrical with equal dimensions for linearly trending topographic highs and
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28 1013 lows.

29 1014 Basins on folded lithosphere have a strong tendency to remain closed systems during a major part
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31 1015 of their evolution, with capture times decreasing with thermo-mechanical age from several tens of
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33 1016 Myr to less than 10 Myr. The thermal history of these basins is not characterized by an initial
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35 1017 thermal perturbation, unless folding occurs in interaction with plume activity. In the absence of an
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37 1018 interplay with lithospheric scale thermal perturbations, their thermal evolution will be primarily
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39 1019 controlled by the burial history of the sediments deposited in the accommodation space created by
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41 1020 the folding process.

42 1021 Intrabasinal faulting and the formation of pop-up structures and inverted basins at upper crustal
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44 1022 levels are important in the structural evolution of these basin systems. The intensity of the brittle
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46 1023 deformation peaks with ongoing shortening of the lithosphere during folding but faulting activity is
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48 1024 dramatically reduced after the cessation of folding. At this stage, with an equilibrium between
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50 1025 sedimentation and erosion, minor faulting occurs as a result of stress relaxation due to erosion of
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3 1026 folded highs. During capture and overall uplift of basins on folded lithosphere, their accommodation
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5 1027 space is further reduced by unflexing and associated stress release through faulting.
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For Review Only

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3 1600 **APPENDIX A: MODEL DESCRIPTION**
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7 1602 **A1. Equations.** The Flamar-Para(o)voz is a “2.5 D” FLAC-like code (Cundall, 1989). It has a mixed
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9 1603 finite-difference/finite element numerical scheme, in which the coordinate frame is Cartesian 2D, but
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11 1604 stress/strain relations are computed in a 3D formulation. The Lagrangian mesh of Paravoz is
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13 1605 composed of quadrilateral elements subdivided onto 2 couples of triangular sub-elements with tri-
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15 1606 linear shape functions. Para(o)voz is a large strain fully explicit time-marching algorithm. It locally
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17 1607 solves full Newtonian equations of motion in continuum mechanics approximation:
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19

$$\langle \rho \ddot{\mathbf{u}} \rangle - \text{div} \boldsymbol{\sigma} - \rho \mathbf{g} = 0 \quad (\text{A1})$$

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21
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24 1608 coupled with constitutive equations:
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$$\frac{D\boldsymbol{\sigma}}{Dt} = F(\boldsymbol{\sigma}, \mathbf{u}, \nabla \dot{\mathbf{u}}, \dots, T \dots) \quad (\text{A2})$$

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31 1609 and with equations of heat transfer (heat advection $\mathbf{u} \nabla T$ in the equation below is solved
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33 1610 separately):
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$$\rho C_p D T / D t + \dot{\mathbf{u}} \nabla T - k \text{div}(\nabla T) - H_r = 0 \quad (\text{A3})$$

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

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41 1611 Here \mathbf{u} , $\boldsymbol{\sigma}$, \mathbf{g} , k are the respective terms for the displacement, stress, acceleration due to body
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43 1612 forces and thermal conductivity. The triangular brackets in (A1) specify conditional use of the related
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45 1613 term (in quasi-static mode inertial terms are dumped using inertial mass scaling (Cundall, 1989)).
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47 1614 The terms t , ρ , C_p , T and H_r designate respectively time, density, specific heat, temperature and
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49 1615 internal heat production. The terms D/Dt , $D\boldsymbol{\sigma}/Dt$, F are a time derivative, an objective (Jaumann)
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51 1616 stress time derivative and a functional, respectively. In the Lagrangian method, incremental
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53 1617 displacements are added to the grid coordinates allowing the mesh to move and deform with
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55 1618 material. This enables solution of large-strain problems locally using a small-strain formulation. On
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3 1619 each time step the solution is obtained in local coordinates, which are then updated in a large strain
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5 1620 mode.

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8 1621 Solution of (A1) provides velocities at mesh points used for computation of element strains and of
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10 1622 heat advection $\mathbf{u} \nabla T$. These strains are used in (A2) to calculate element stresses and equivalent
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12 1623 forces as input for the computation of the velocities for the next time step. Due to the explicit
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14 1624 approach, there are no convergence issues, which is rather a common problem of implicit methods
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16 1625 in case of non-linear rheologies. The algorithm automatically checks and adopts the internal time
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18 1626 step using 0.1 – 0.5 of Courant's criterion for propagation of information, which warrants a stable
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20 1627 solution.

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25 26 1629 **A2. Explicit EVP rheology.**

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29 1630 We use a serial (Maxwell type) body (eq. 1), in which total strain increment in each numeric element
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31 1631 is defined by a sum of elastic, viscous and brittle strain increments. Consequently, in contrast to
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33 1632 fluid dynamic approaches, where non-viscous rheological terms are simulated using pseudo-plastic
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35 1633 and pseudo-elastic viscous terms (e.g. Bercovici *et al.*, 2001; Solomatov & Moresi, 2000), our
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37 1634 method explicitly treats all rheological terms. The parameters of elastic-ductile-plastic rheology laws
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39 1635 for crust and mantle come from rock mechanics data (**Tables A1 and A2**; Kirby & Kronenberg,
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41 1636 1987; Kohlstedt *et al.*, 1995).

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46 47 1638 **Plastic (brittle) behavior.**

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49 1639 The brittle behavior of rocks is described by Byerlee's law (Byerlee, 1978; Ranalli, 1995), which
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51 1640 corresponds to Mohr-Coulomb material with a friction angle $\phi = 30^\circ$ and cohesion $|C_0| < 20$ MPa
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53 1641 (e.g., Gerbault *et al.*, 1998):

$$\tau = C_0 - \sigma_n \tan\phi \quad (A4)$$

1642 where σ_n is normal stress $\sigma_n = \frac{1}{3}\sigma_1 + \sigma_{II}^{\text{dev}} \sin\phi$, $\frac{1}{3}\sigma_1 = P$ is the effective pressure, σ_{II}^{dev} is the
 1643 second invariant of deviatoric stress, or effective shear stress. The condition of transition to brittle
 1644 deformation (function of rupture f) reads as: $f = \sigma_{II}^{\text{dev}} + P \sin\phi - C_0 \cos\phi = 0$ and $\partial f/\partial t = 0$. In terms of
 1645 principal stresses, the equivalent of the yield criterion (5) reads as:

$$\sigma_1 - \sigma_3 = -\sin\phi (\sigma_1 + \sigma_3 - 2C_0/\tan\phi) \quad (A5)$$

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1647 **Elastic behavior.**

1648 The elastic behavior is described by the linear Hooke's law:

$$\sigma_{ij} = \lambda \varepsilon_{ii} \delta_{ij} + 2G \varepsilon_{ij} \quad (A6)$$

1649 where λ and G are Lamé's constants. Repeating indexes mean summation and δ is Kronecker's
 1650 operator.

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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:

$$e_{ij}^d = A (\sigma_1 - \sigma_3)^n \exp(-Q/R^*T) \quad (A7)$$

1657 Where $e_{ij}^d = \dot{\varepsilon}_{ij}^d$ is shear strain rate, A is material constant, n is the power law exponent, Q is the
 1658 activation enthalpy, R is Boltzmann's gas constant, and T is temperature in K°, σ_1 and σ_3 are the
 1659 principal stresses. The effective viscosity μ_{eff} for this law is defined as:

$$\tau_{ij} \equiv \mu_{eff} e_{ij}^d$$

1660 which yields:

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

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

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

1663 where

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

1664 parameters A, n, Q are experimentally determined material constants (**Table A1**). Using olivine
1665 parameters (**Table A1**), one can verify that the predicted effective viscosity at the base of the
1666 lithosphere is $10^{19} - 5 \times 10^{19}$ Pa s matching post-glacial rebound data (Turcotte & Schubert, 2002).
1667 In the depth interval of 200 km - 0 km the effective viscosity grows from 10^{19} to $10^{25} - 10^{27}$ Pa s with
1668 decreasing temperature. Within the adiabatic temperature interval at depth, the dislocation flow law
1669 (A8) is replaced by nearly Newtonian diffusion creep, which results in a quasi-constant mantle
1670 viscosity of $10^{19} - 10^{21}$ Pa s (e.g., Turcotte & Schubert, 2002). Weinberg & Podladchikov (1994)
1671 have also shown that the effective viscosity in close vicinity of an ascending diapir is influenced by
1672 the local strain rate field and partly by heat exchanges between the diapir and surrounding rock,
1673 which suggests possible changes.

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3 1675 **FIGURE CAPTIONS**
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8 1677 **Figure 1.** Concepts of folding in rheologically stratified lithosphere and feedback with sedimentation
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10 1678 in downfolded areas and erosion of adjacent highs. Different simultaneously occurring wavelengths
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12 1679 of crustal and mantle folding are a consequence of the rheological stratification of the lithosphere.
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14 1680 Surface wavelengths can be affected also by feedback with surface processes.
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18 1682 **Figure 2.** Map with examples of well-documented areas affected by lithospheric folding. Areas:
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20 1683 Central Indian Ocean Basin (CIOB) (Geller *et al.*, 1983 ; Stein *et al.*, 1989) ; NW European platform
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22 1684 (Bourgeois *et al.*, 2007 ; Lefort & Agarwal, 1996; Marotta *et al.*, 2000); Pannonian Basin (Dombradi
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24 1685 *et al.*, 2010); Iberia (Cloetingh *et al.*, 2002; De Vicente *et al.*, 2007; Fernandez-Lozano *et al.*, 2010);
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26 1686 Central Asia (Burov *et al.*, 1993; Nikishin *et al.*, 1993; Burov & Molnar, 1998); Tibet/ Himalayan
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28 1687 syntaxis belt (Burg & Podlachikov, 1999; Shin *et al.*, 2009); Central Australia (Stephenson &
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30 1688 Lambeck, 1985; Lambeck, 1983); Arctic Canada (Stephenson *et al.*, 1990); Transcontinental Arch
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32 1689 of North America (Ziegler *et al.*, 1995); South Caspian Basin (Guest *et al.*, 2007); Laramide
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34 1690 Foreland (USA) (Tikoff & Maxson, 2001); Barents Sea (Ritzmann & Faleide, 2009) (map credit:
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36 1691 NASA).
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42 1693 **Figure 3.** Relationship between wavelengths of lithospheric folding and thermo-tectonic age of
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44 1694 lithosphere. The grey zones correspond to theoretical folding wavelengths derived for the upper
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46 1695 crust, mantle and coupled whole mantle folding, based on an analytical model that accounts for
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48 1696 strength variations as a function of thermo-mechanical age (Cloetingh *et al.*, 1999). Black squares
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50 1697 are estimates for wavelengths of differential topography and thermo-tectonic ages inferred from
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52 1698 observational studies. Numbers refer to sites listed in Table 1.
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3 1700 **Figure 4.** (a) Map showing major tectonic features of Tien Shan area of Central Asia , with location
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5 1701 of folding profiles discussed in Burov *et al.* (1993), Burov & Molnar (1998) and Cloetingh *et al.*
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7 1702 (1999). (b) Neotectonic movements along the profiles AD and CD (Burov *et al.*, 1993). Modified
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9 1703 after Cloetingh *et al.* (1999). (c) Measured and calculated Bouguer gravity anomalies for lithosphere
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11 1704 under horizontal compression along profile F across the Ferghana basin with corresponding
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13 1705 topography profile (d) (after Burov & Molnar, 1998). Predicted topography and Moho geometry for
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15 1706 Airy and best fitting folding scenario. During compression the amplitude of plate deflection changed
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17 1707 significantly while the wavelength of folding stayed constant (Burov & Molnar, 1998). Red lines are
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19 1708 Bouguer anomaly predicted by folding model and Moho depression, respectively. Solid grey line is
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21 1709 Airy isostasy. Solid black line is Airy Moho. Triangles indicate observed Bouguer anomaly. Green is
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23 1710 topography above sea level. FB: Ferghana Basin.

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27 1712 **Figure 5.** First order features of basin geometry of folded lithosphere of Iberia. (a) Topography
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29 1713 displaying cylindrical patterns of alternating parallel trending highs and lows and corresponding
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31 1714 gravity anomalies (red line) (see Cloetingh *et al.*, 2002). (b) Results from analogue modelling
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33 1715 experiments, displaying pop-up structures underlying topographic highs, accompanied by folding of
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35 1716 the Moho (Fernandez-Lozano *et al.*, 2010). Top panel shows final top view of the analogue model
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37 1717 and location of the cross-section shown below. Arrow indicates the direction of the moving wall. λ is
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39 1718 the wavelength of folding.

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43 1720 **Figure 6.** (a) Map of elevation of basement in the NW Alpine foreland with rift signature removed
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45 1721 from the map of actual basement (from Bourgeois *et al.*, 2007). Dashed lines show axes of Late
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47 1722 Neogene lithospheric folds, with a succession of anticlines and synclines: the Normandy-
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49 1723 Vogelsberg anticline (NVA), Sologne-Franconian Basin Syncline (SFS) and the Burgundy-Swabian
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51 1724 Jura Anticline (BSJA). LBR: London Brabant Ridge with thick crust. BF: Black Forest; FB
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53 1725 Franconian Basin; E: Eifel; H: Hunsruck; S: Sauerland; T: Taunus; V: Vosges. (b) Cross sections of
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3 1726 top basement in the NW Alpine foreland with rift signatures removed (Bourgeois *et al.*, 2007; see for
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5 1727 location Fig. 6a). Thin line: actual topography surface. Bold line: elevation of top basement as
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7 1728 predicted in the absence of rifting. Fold signatures are similar with a wavelength λ of 270 km and
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10 1729 amplitude A of 1500 m in the Franconian Basin (Cross section G-G') and in the upper Rhine Graben
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12 1730 area (Cross section H-H'). In the Paris basin, development of folds with assumed same wavelength
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14 1731 and amplitude (stippled bold line in cross section C-C'), implies subsidence of Sologne and uplift of
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16 1732 Normandy.
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21 1734 **Figure 7.** (a) Schematic E-W cross section through the Pannonian-Carpathian basin system
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23 1735 (Cloetingh *et al.*, 2008), displaying anomalous Late Neogene acceleration of the subsidence
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25 1736 between isochrons 2 and 0 Ma in the centre of the Pannonian basin and the Focsani depression of
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27 1737 the Southern Carpathian foredeep, flanked by uplifting highs. Wavelengths of the differential vertical
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29 1738 motions are typically of the order of 250 km. (b) Interpreted seismic section in the central part of the
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31 1739 Pannonian Basin (Sacchi *et al.*, 1999) showing evidence for late Neogene crustal scale folding at
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34 1740 scales of 20-40 km. Note syn-rift grabens formed during the Miocene back-arc extension of the
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36 1741 basin. (c) Results of analogue tectonic experiments carried out to examine the patterns of
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38 1742 differential vertical motions in the northern part of the Pannonian Basin and surrounding areas
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40 1743 (Dombradi *et al.*, 2010), demonstrating a-typical folding of variable amplitude with a wavelength
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42 1744 equivalent to 200-250 km.
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46 1746 **Figure 8.** Schematic north-south trending cross-section illustrating the basic tectonic setting for
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48 1747 northern Iran and the South Caspian Basin (Guest *et al.*, 2007), illustrating basin inversion of the
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50 1748 central Iranian basin, pronounced differential topography at the transition of the Alborz mountains,
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52 1749 with a topographic high of several km's and the South Caspian Basin with about 20 km of
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55 1750 sedimentary infill.
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3 1752 **Figure 9.** Characteristic patterns of continental lithosphere deformation induced by folding of the
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5 1753 lithosphere for with time increasing amounts of horizontal shortening for a 150 Ma old lithosphere
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7 1754 with convergence rates of **(a)** 1.5 cm/yr and **(b)** 3 cm/yr. From left to right: material phase
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10 1755 distribution (with blue for the crust, orange for the mantle lithosphere and red for the upper mantle
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12 1756 below the lithosphere), temperature, effective viscosity, and topography.

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16 1758 **Figure 10.** Characteristic patterns of continental lithosphere deformation induced by folding of
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18 1759 lithosphere with time. Increasing amounts of horizontal shortening for a 300 Ma old lithosphere with
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20 1760 convergence rates of **(a)** 1.5 cm/yr and **(b)** 3 cm/yr. From left to right: phase distribution,
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22 1761 temperature, effective viscosity, and topography.

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24 1762
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27 1763 **Figure 11. (a)** Characteristic subsidence patterns in centre of synclinal depression for thermo-
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29 1764 mechanical age of 150 and 300 Ma, respectively. Shortening rate is 3 cm/year. **(b)** Characteristic
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31 1765 stages in the evolution of a basin formed by lithospheric folding. Stage 1: Basin formation stage,
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33 1766 acceleration of subsidence and uplift during folding. Stage 2: Steady-state: equilibrium between
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35 1767 tectonic subsidence and sediment supply from eroding highs. Stage 3: Capture of folded basin;
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37 1768 overall uplift and erosion. Wiggled waved blue line marks position of base level.

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42 1770 **Figure 12.** Illustration of the effect of erosion, which acts as a filter suppressing the short
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44 1771 wavelengths. In this case, the wavelength and amplitude vary along the cross section at different
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46 1772 stages of deformation due to a partial crust-mantle coupling and strain localization for a 400 Ma old
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48 1773 lithosphere with weak quartz -dominated rheology. After 5 percent shortening (top), after 25 percent
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50 1774 shortening (middle), after 25 percent shortening, strong zero order diffusional erosion (Avouac &
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52 1775 Burov, 1996) tuned to keep mean elevation at the level of 3000 m. Erosion reduces the contribution
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54 1776 of gravity-dependent terms (middle wavelength) and accelerates local deformation. Strong erosion,
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3 1777 insufficiently compensated by the tectonic deformation wipes out most of the topography (after
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5 1778 Cloetingh *et al.*, 1999).
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10 1780 **Figure 13.** Thermal consequences of lithosphere folding, and associated basin evolution depicted in
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12 1781 Fig. 11. The heat flow response is marked by an initial decrease in heat flow as a consequence of
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14 1782 rapid sedimentation, followed by a progressive long-term heat flow increase after termination of
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16 1783 shortening. Thermal conductivity of sediments is $2.5 \text{ W m}^{-1} \text{ C}^{-1}$. Surface radiogenic heat production
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18 1784 H_s is $9.5 \times 10^{-10} \text{ W kg}^{-1}$. Concentration of radiogenic heat sources within sedimentary basin fill is
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20 1785 homogeneous. Exponential decay of heat production is assumed for underlying crustal rocks. Note
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22 1786 that the figure shows only 7 Ma of thermal evolution. The heat flow does not increase anymore after
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24 1787 some tens of Ma and subsequently decreases at later stages.
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29 1789 **Figure 14.** Results of analogue tectonic experiments for lithospheric folding (after Sokoutis *et al.*,
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31 1790 2005) (Top): folding of uniform lithosphere. Cross section demonstrating pop-up structures in the
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33 1791 upper crust above highs induced by lithospheric folding. λ_1 and λ_2 indicate first-order and second-
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35 1792 order wavelengths, respectively. Imbrication of bivergent thrust wedges is responsible for the
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37 1793 relatively high topography belt (Σ -belt) at the right-hand of the model and closure of prismatic basins
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39 1794 with a rather undisturbed bottom. The progressive sinking of the prismatic basin allows transferring
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41 1795 of the crustal material (i.e. cover sediments) that was initially situated at the surface of the model
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43 1796 down to significant depth. Left-side (a) final top view of model after 34% of bulk shortening. Right
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45 1797 side (b) cross-section located on the top view. Arrows indicate the direction of the moving wall. Note
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47 1798 the similarity to Central Iberia (**Fig. 5**). (Bottom): Geometries predicted by analogue experiment,
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49 1799 shortening a lithosphere with two contrasting blocks, bounded by suture. A deep synclinal
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51 1800 depocentre develops over the suture zone, flanked by an anticline of smaller amplitude. Left-side (c)
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53 1801 final top view of model after 26% of bulk shortening. Right side (d) cross-section located on the top
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3 1802 view. λ_1 indicates a first-order wavelength of the folding. Note the striking similarity to the
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5 1803 configuration of the South Caspian Basin/ Alborz Mountain system displayed in **Fig. 8**.

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10 1805 **Figure 15.** Numerical experiment on faulting developed in folded continental lithosphere of 150 Ma
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12 1806 (Burov & Molnar, 1998). Note different wavelengths of the crustal and mantle part of the lithosphere
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14 1807 and differences in spacing and character of faults that are conditioned by folding. In each layer, two
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16 1808 fault spacings occur: a shorter one proportional to layer thickness (also controlled by the strength
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18 1809 contrast between the stiff layer and embeddings), and the longer wavelength controlled by the
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20 1810 wavelength of folding. The insert shows plastic (brittle) strain localization characterizing initialization
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22 1811 of mantle faulting.

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27 1813 **Figure 16.** Comparison of main characteristics of basins developed on folded lithosphere (FLB with
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29 1814 a shortening rate of 3 cm/yr) with other basin types, including foreland basins (FBr = retro-arc
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31 1815 foreland basin; FBp = pro-arc foreland basin), intracratonic basins (ICB), extensional basins (EB)
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33 1816 and pull-apart basins (PAB). Different panels give theoretical predictions. (Top): Basin subsidence
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35 1817 history in basin centre. (Bottom): Thermal history. See text for explanation.

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TABLE 1

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 <i>et al.</i> , 1994
2	Western Goby	175-400	300-360	Nikishin <i>et al.</i> , 1993; Burov <i>et al.</i> , 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 <i>et al.</i> , 2007 ; Dombradi <i>et al.</i> , 2010
11	NW European platform	180-230	270	Bourgeois <i>et al.</i> , 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 <i>et al.</i> , 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

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
EB	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
PAB	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**

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

11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32
Parameter	Values and units	Definition																			
σ, τ	Pa, MPa	stress (complete, deviatoric)																			
P	Pa, MPa	pressure																			
u	m, mm	displacement vector																			
v	$m s^{-1}$, $mm an^{-1}$	velocity vector																			
$\dot{\epsilon}$	s^{-1}	strain rate																			
μ	10^{19} - 10^{25} Pa s	effective viscosity ($\tau/\dot{\epsilon}$)																			
T	$^{\circ}C$, K	temperature																			
h_c	40 km	Moho depth																			
h_l	100 - 250 km	thickness of lithosphere																			
D	100 - 200 km	plume diameter																			
ρ	$3330 kg m^{-3}$	density of mantle lithosphere																			
ρ_m	$3340 kg m^{-3}$	reference deep mantle density																			
ρ_p	$\rho_m + \Delta\rho_{ch} + \alpha\rho_m\Delta T$	plume density																			
$\Delta\rho_{ch}$	0-25 $kg m^{-3}$	chemical density contrast																			
g	$9.8 m s^{-2}$	acceleration due to gravity																			
C_p	$10^3 J Kg^{-1} ^{\circ}C^{-1}$	specific heat																			
ΔT	$250^{\circ}C$	initial temperature contrast plume - background																			
α	$3 \times 10^{-5} ^{\circ}C^{-1}$	thermal expansion coefficient at 0 km depth																			
β	$8 \times 10^{-12} Pa^{-1}$	isothermal compressibility																			

Table A2. Specific rheology and related thermal parameters. Compilation by Burov *et al.* (2001). ρ is density; Q , n , A are material-dependent parameters of ductile flow laws (Kirby & Kronenberg, 1987; Kohlstedt *et al.*, 1995). Other parameters from Turcotte & 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
<u>Specific upper or weak (quartz) lower-crust properties</u>	
ρ (upper crust)	2700 kg m ⁻³
ρ (lower crust)	2900 kg m ⁻³
n	2.4
A	6.7×10^{-6} MPa ⁻ⁿ ·s ⁻¹
Q	1.56×10^5 kJ·mol ⁻¹
<u>Specific strong lower crust properties (diabase or basalt)</u>	
ρ	2980 kg m ⁻³
n	3.4
A	2×10^{-4} MPa ⁻ⁿ ·s ⁻¹
Q	2.6×10^5 kJ·mol ⁻¹
<u>Specific mantle properties (olivine)</u>	
ρ (lithosphere)	3330 kg m ⁻³
n	3
A	1×10^4 MPa ⁻ⁿ ·s ⁻¹
Q	5.2×10^5 kJ·mol ⁻¹
<u>Thermal model</u>	
Surface temperature (0 km depth)	0 °C
Temperature at the bottom of thermal lithosphere	1330 °C
Temperature at 660 km depth	2000 °C
Thermal conductivity of crust k	2.5 Wm ⁻¹ °C ⁻¹
Thermal conductivity of mantle k	3.5 Wm ⁻¹ °C ⁻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 h_r	10 km
Thermo-tectonic age of the lithosphere a	60 (young) to 1000 Ma (old)

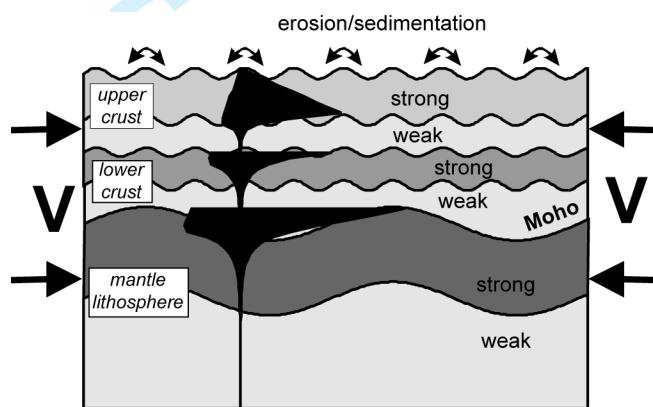


Figure 1

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Figure 2

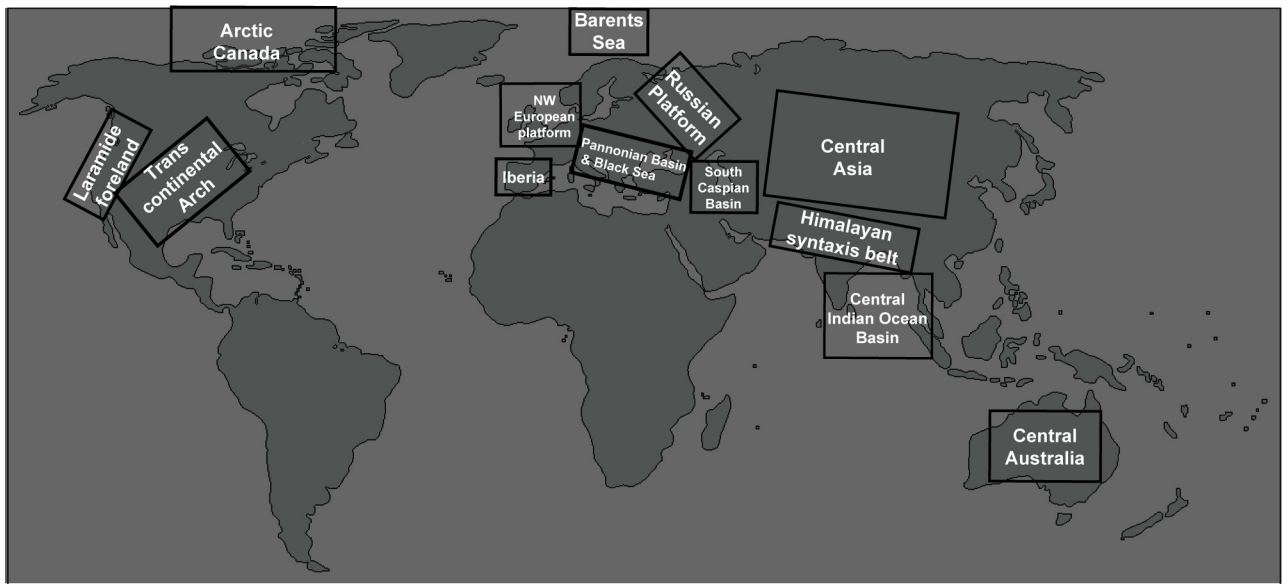
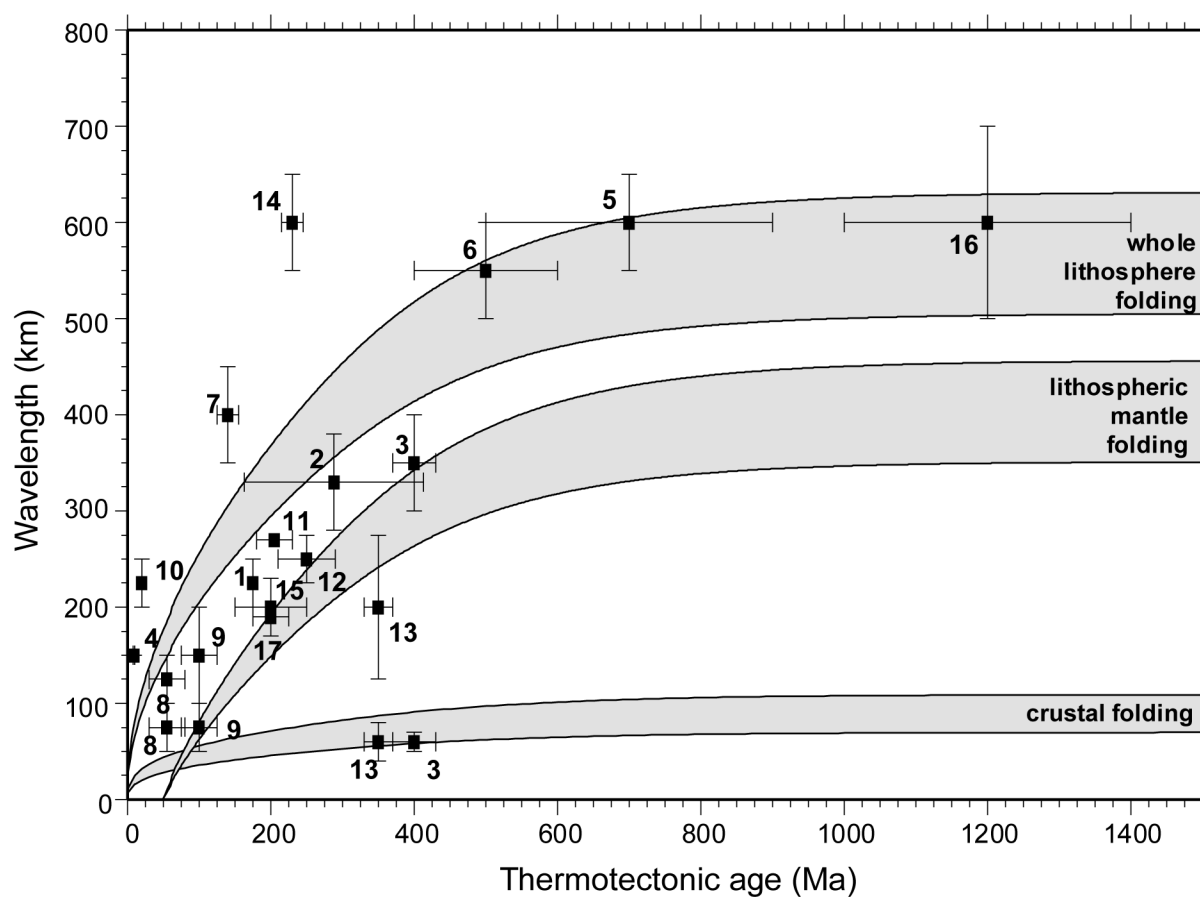
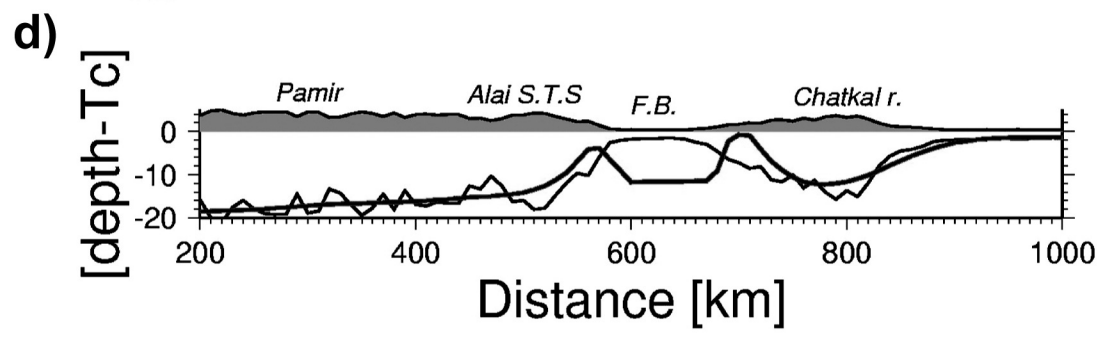
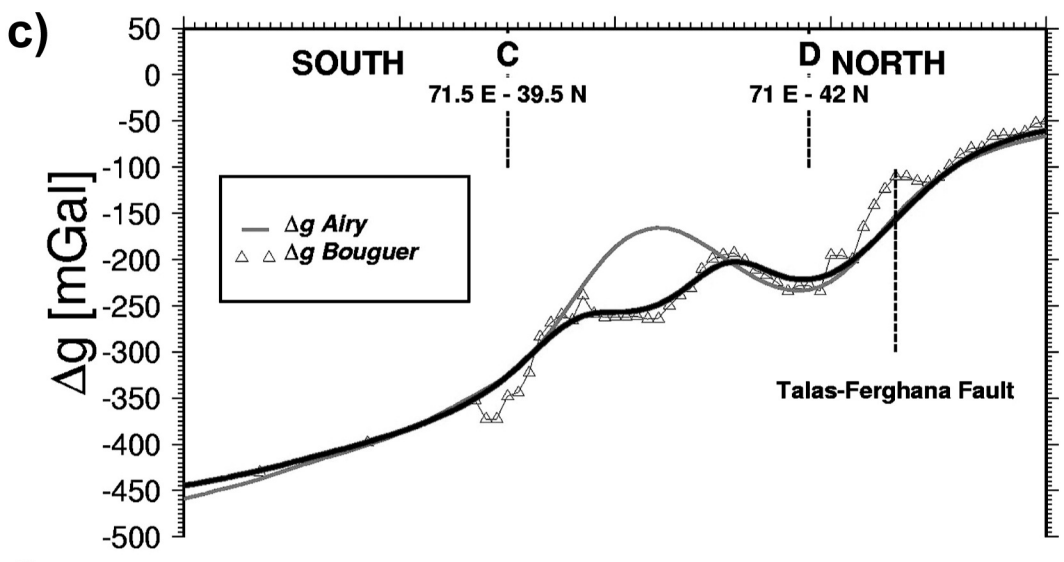
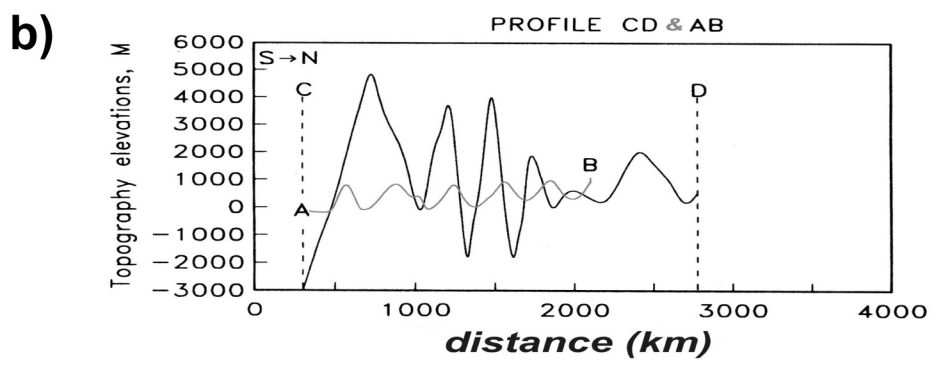
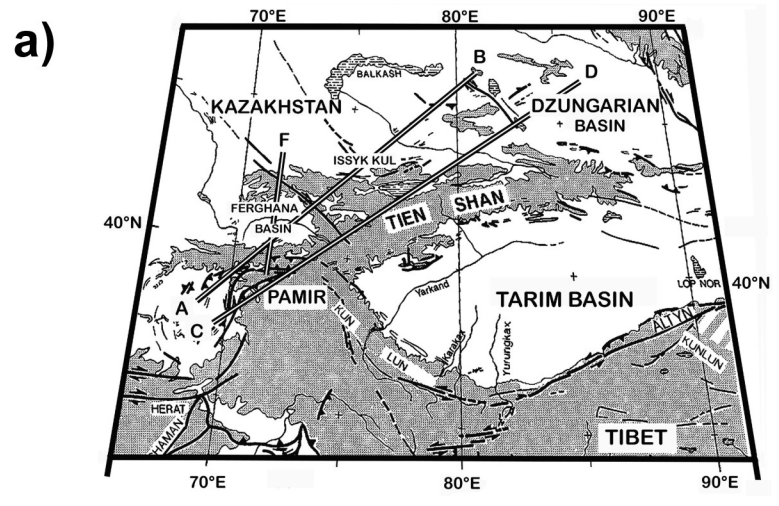


Figure 3



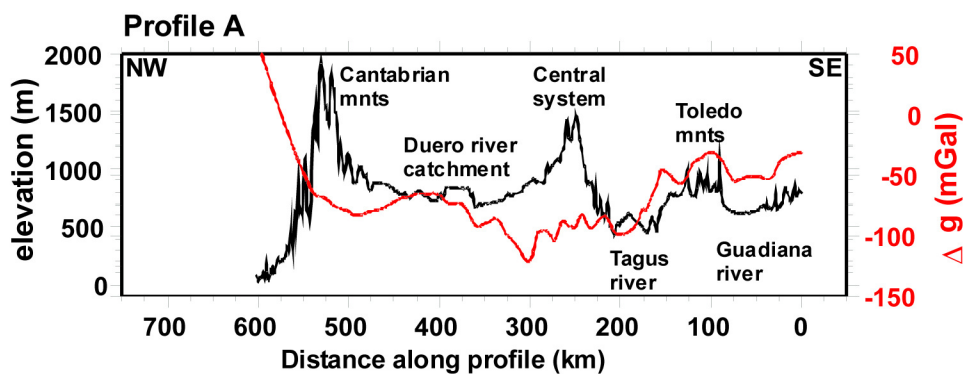
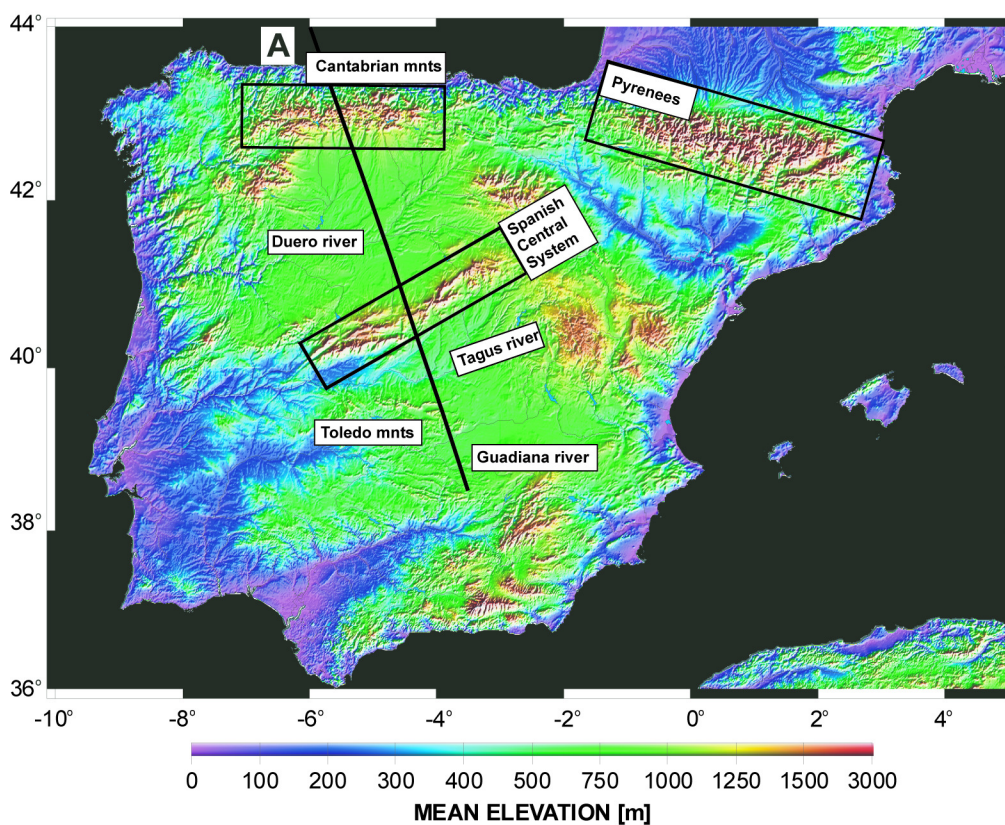
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Figure 5a



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Figure 5b

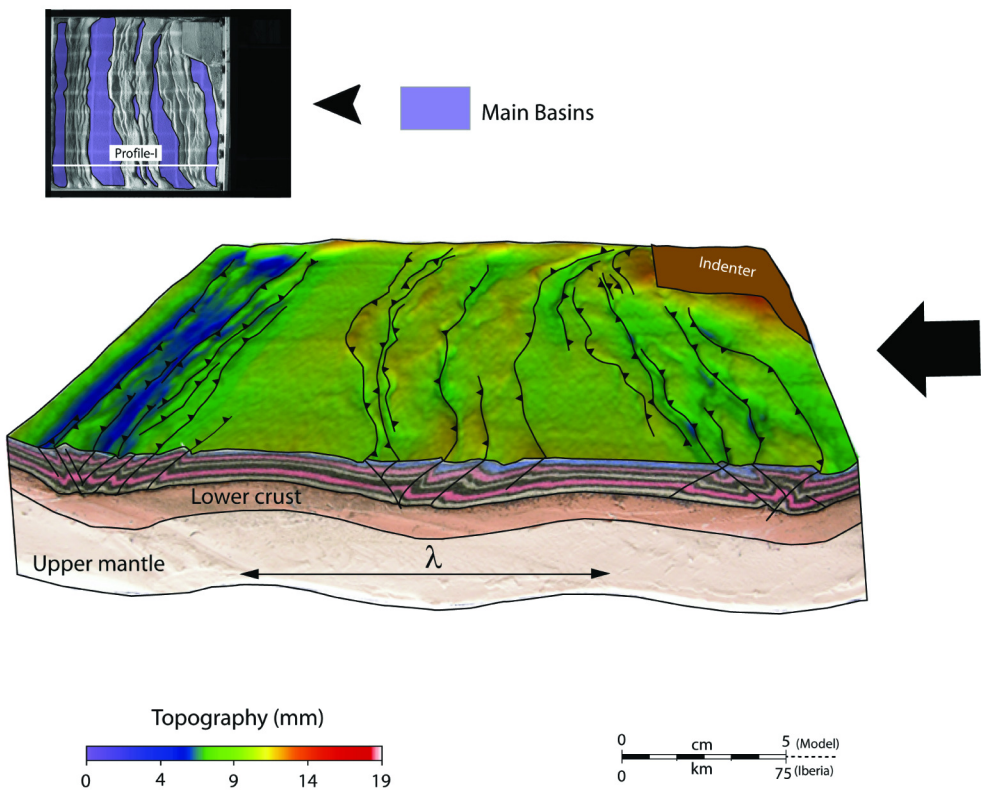
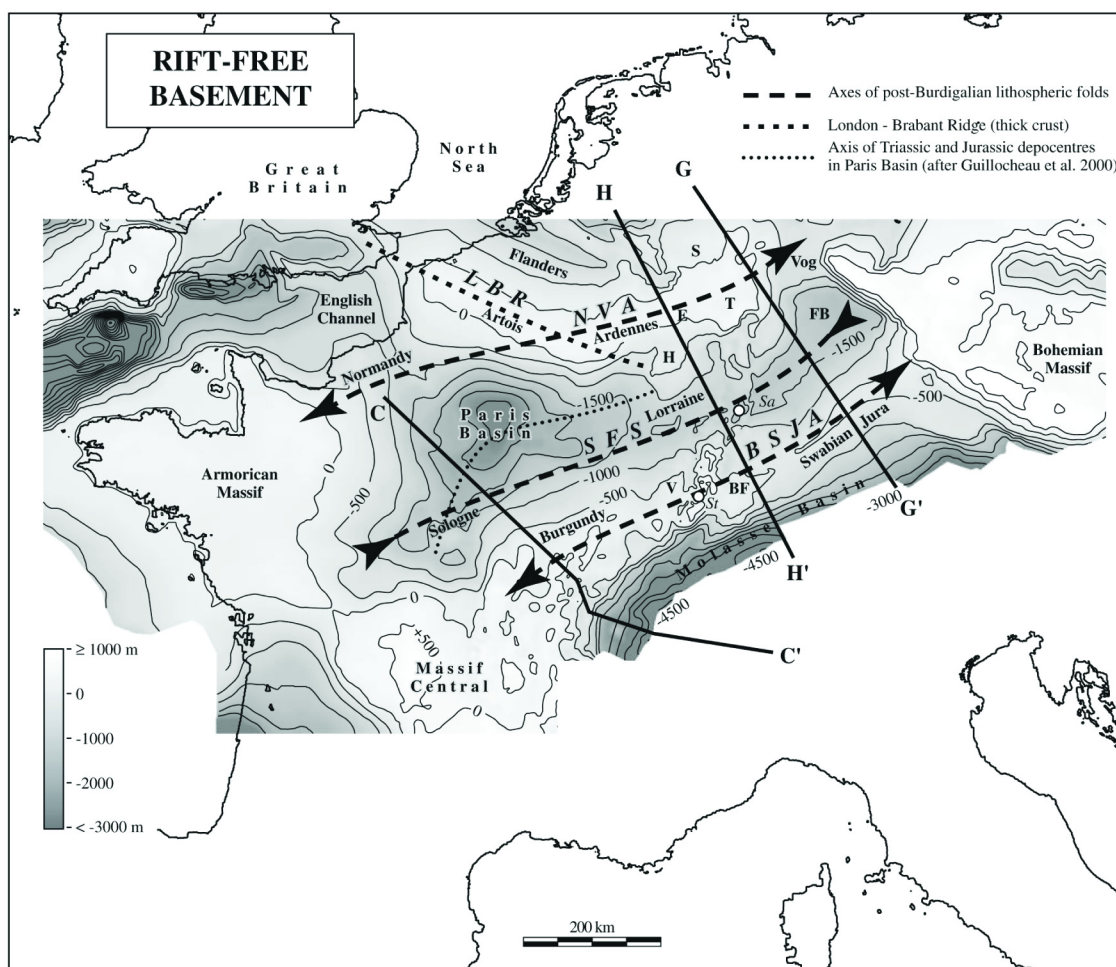
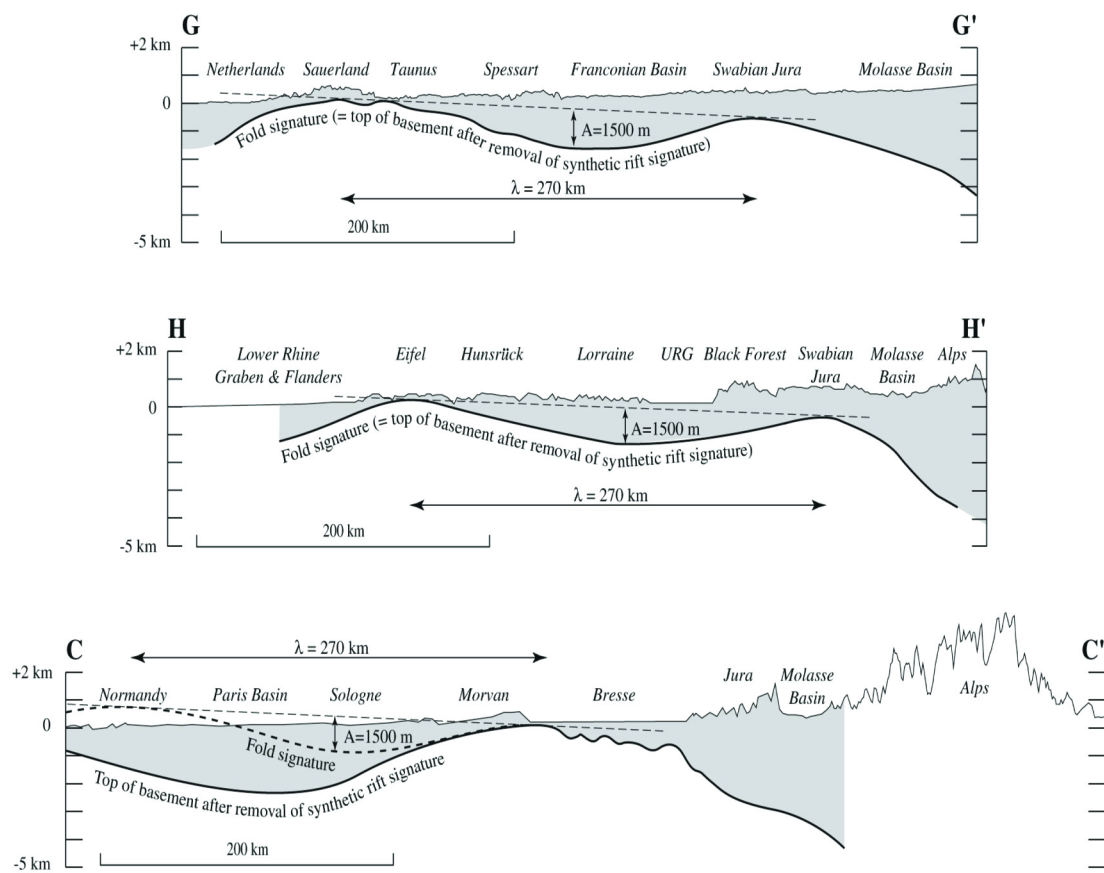


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Figure 7a

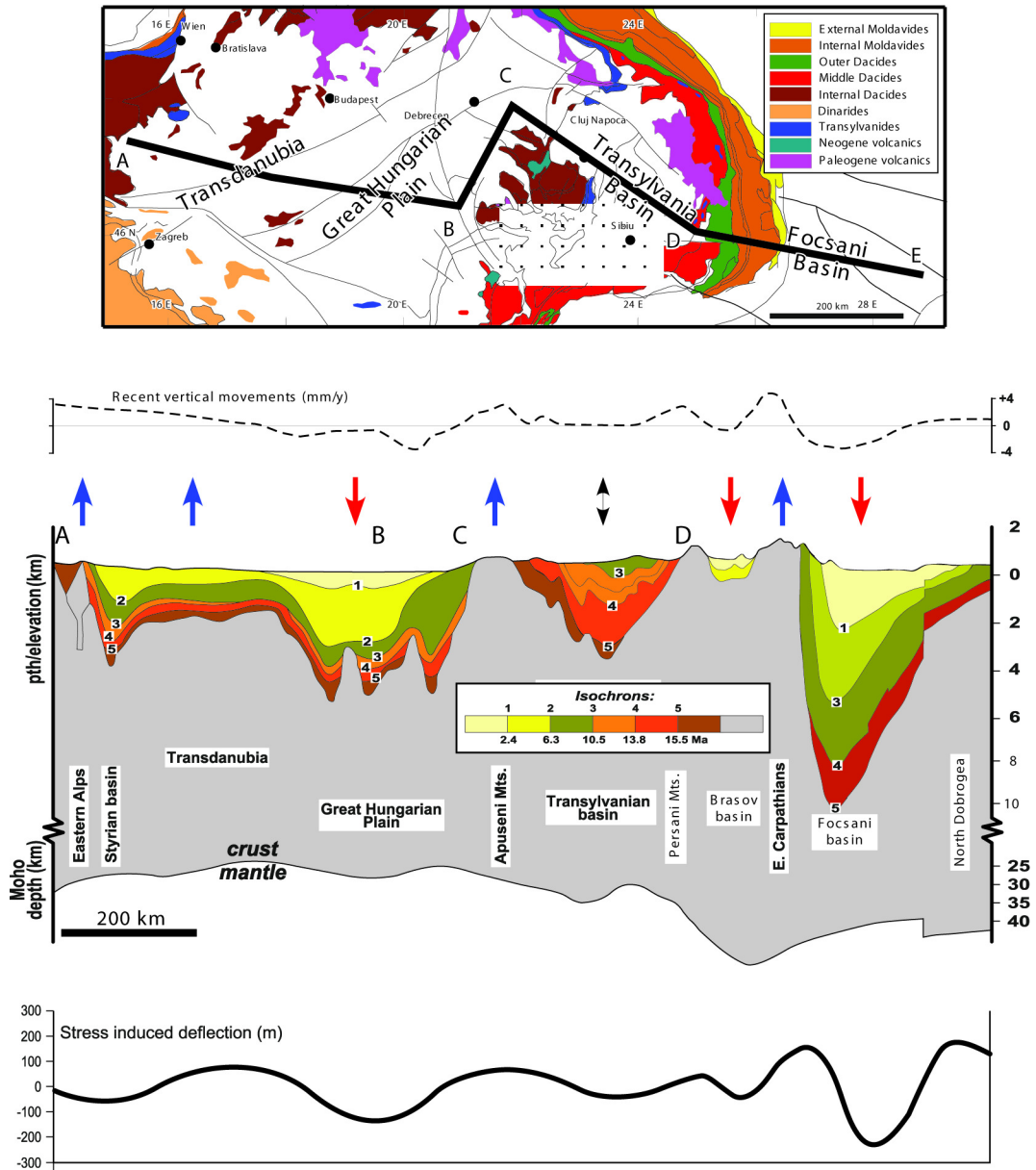
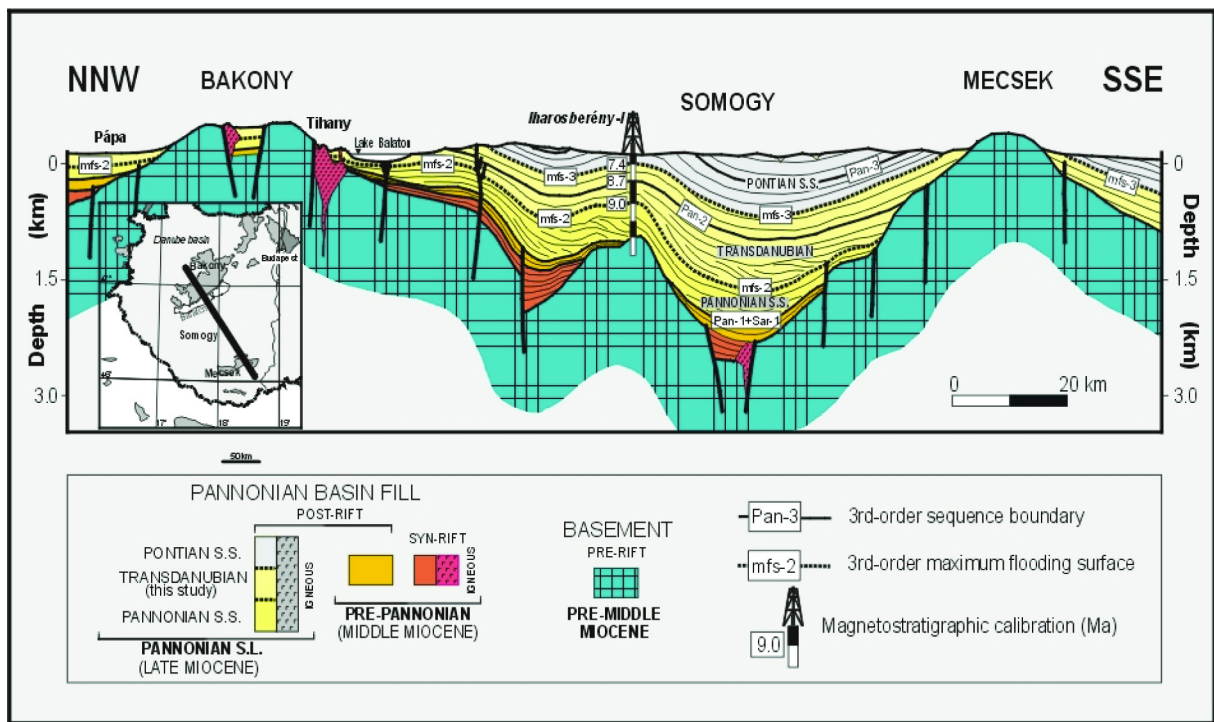
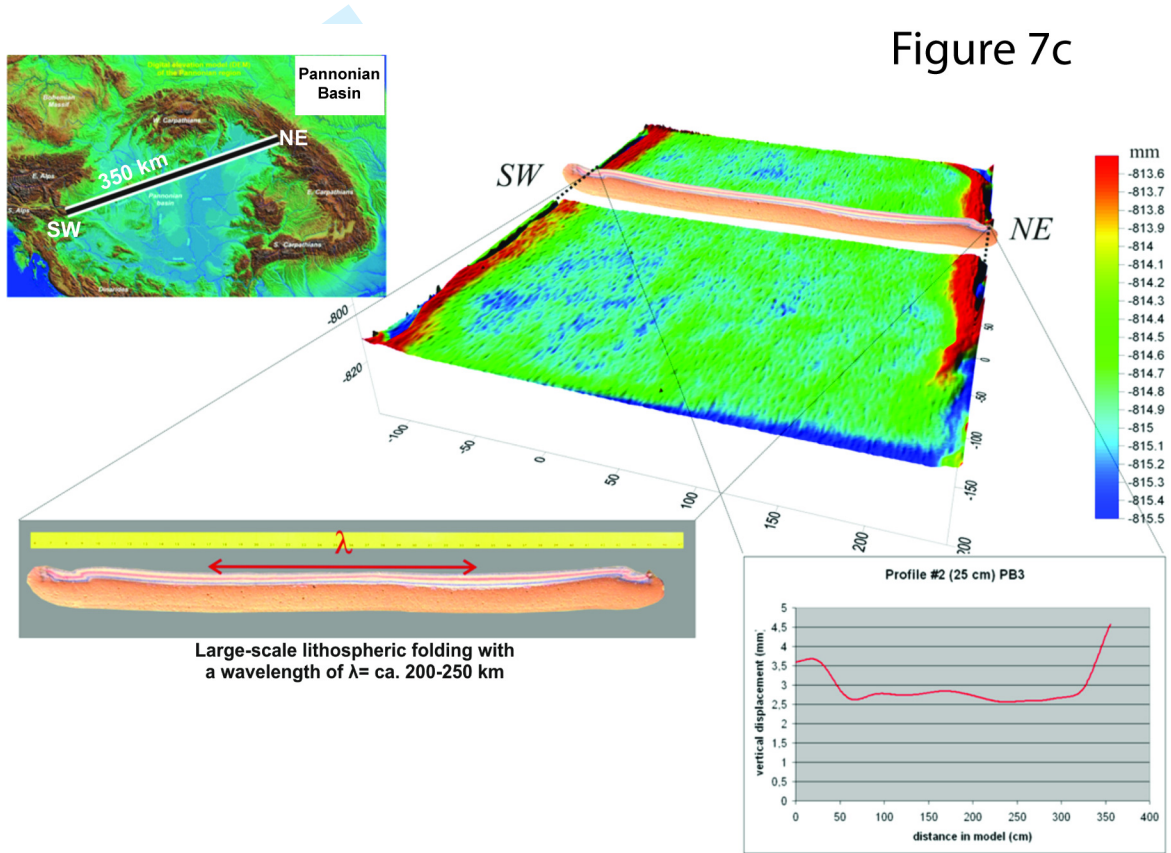


Figure 7b



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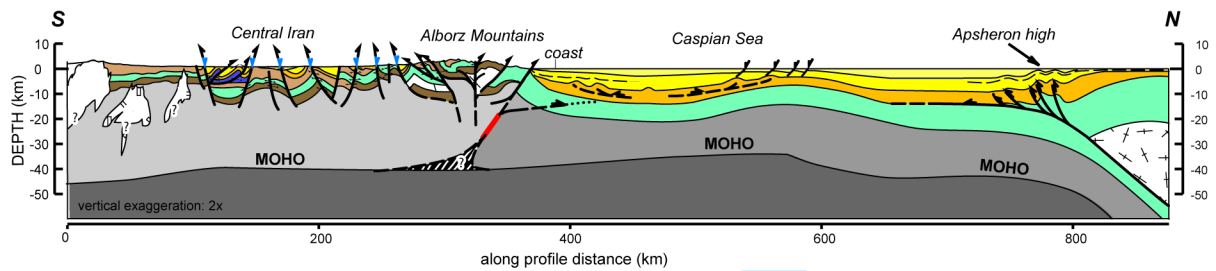
Figure 7c



Large-scale lithospheric folding with a wavelength of $\lambda =$ ca. 200-250 km

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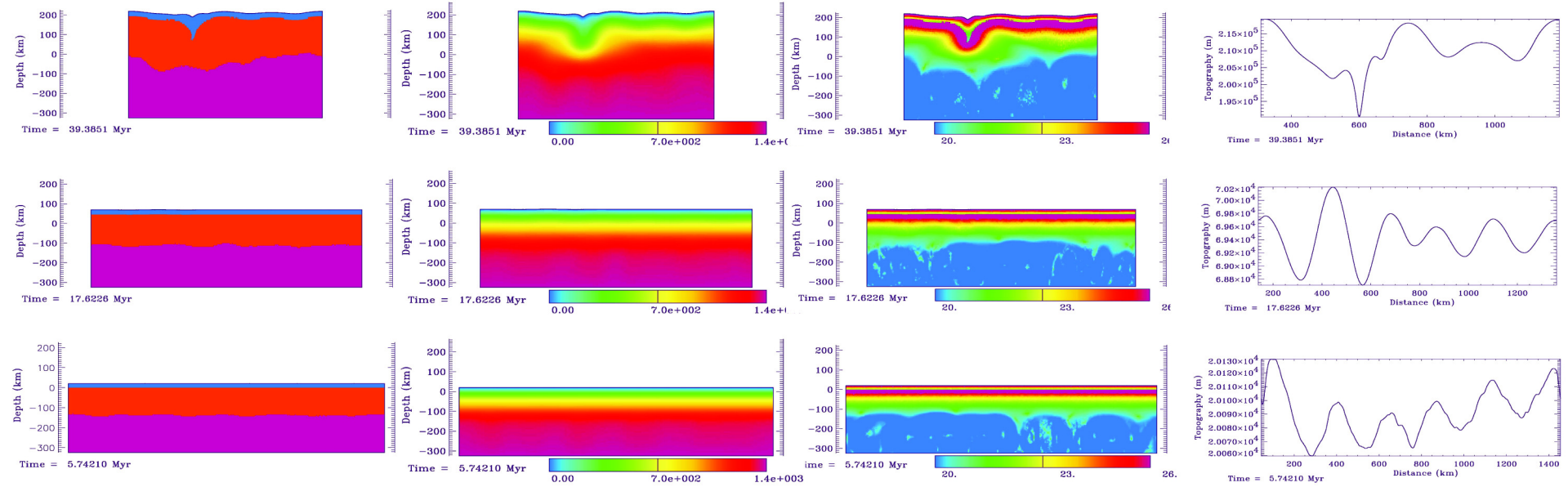
Figure 8



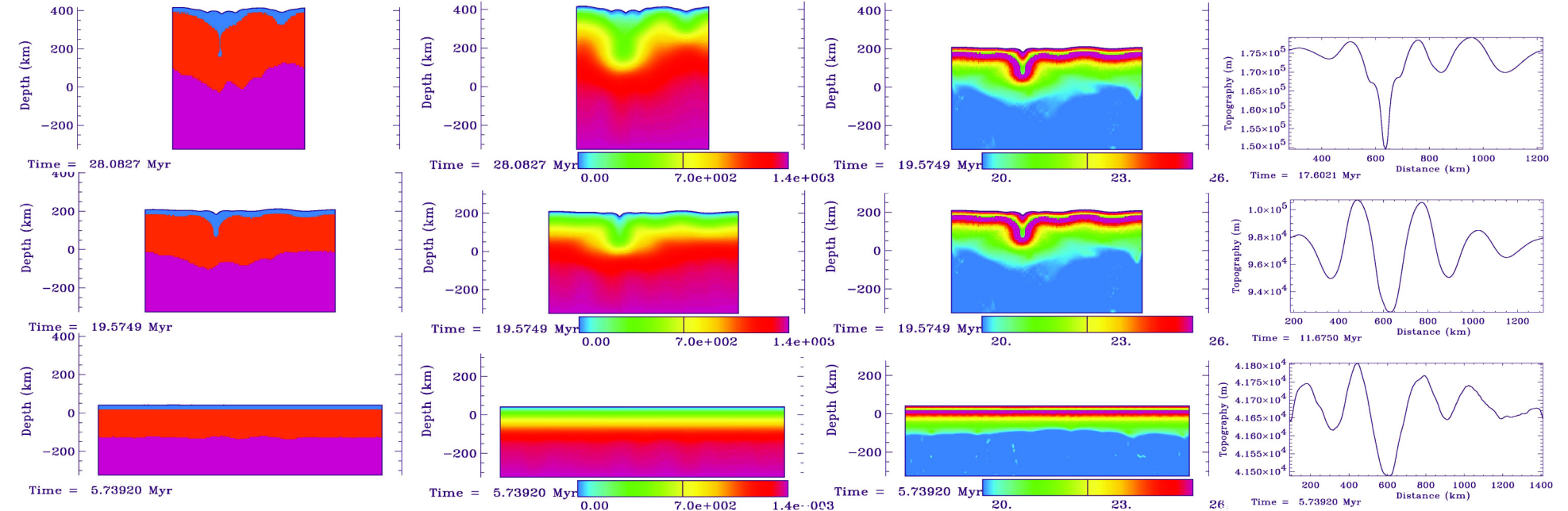
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Figure 9

a) $t=150$ Ma, $v=1.5$ cm/yr



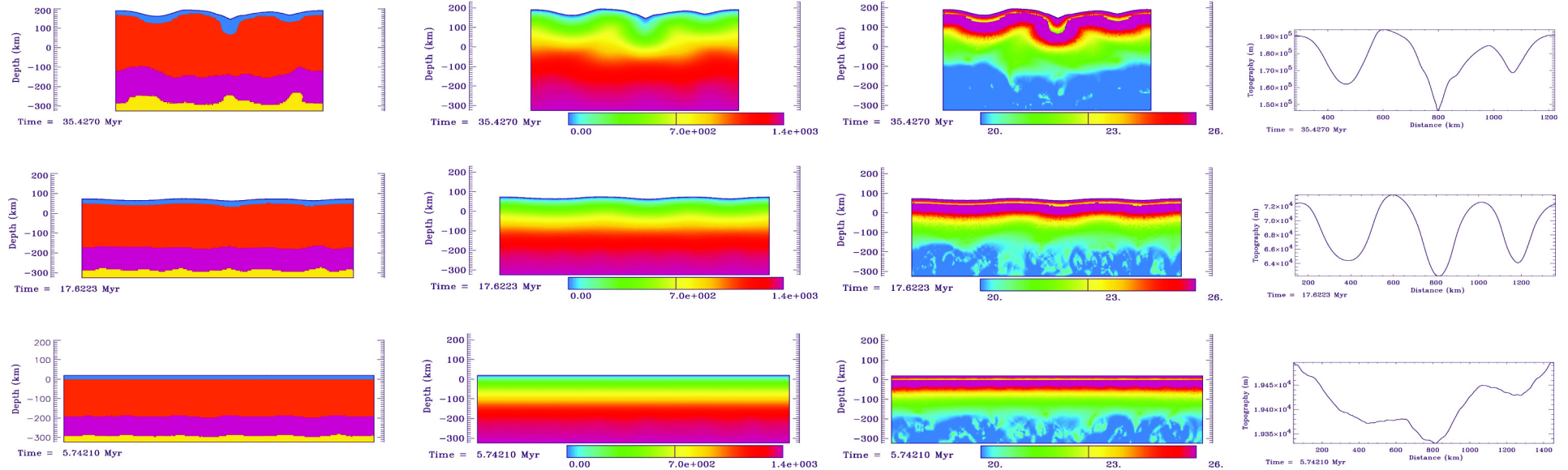
b) $t=150$ Ma, $v=3.0$ cm/yr



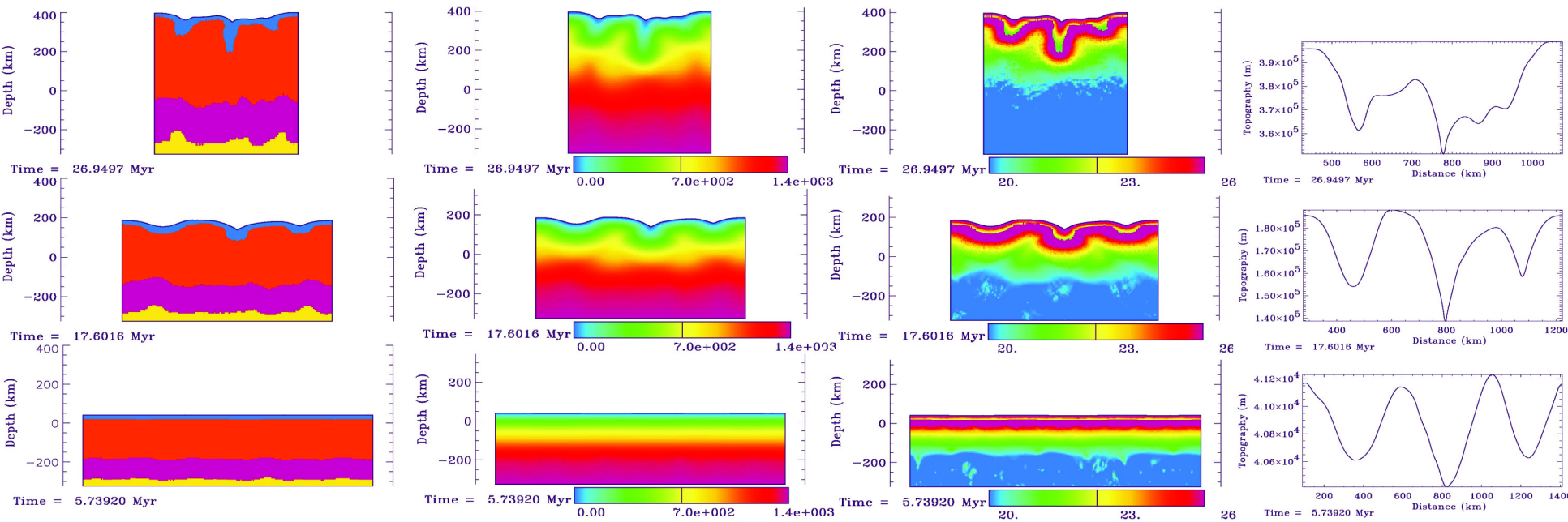
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Figure 10

a) t=300 Ma, v=1.5 cm/yr

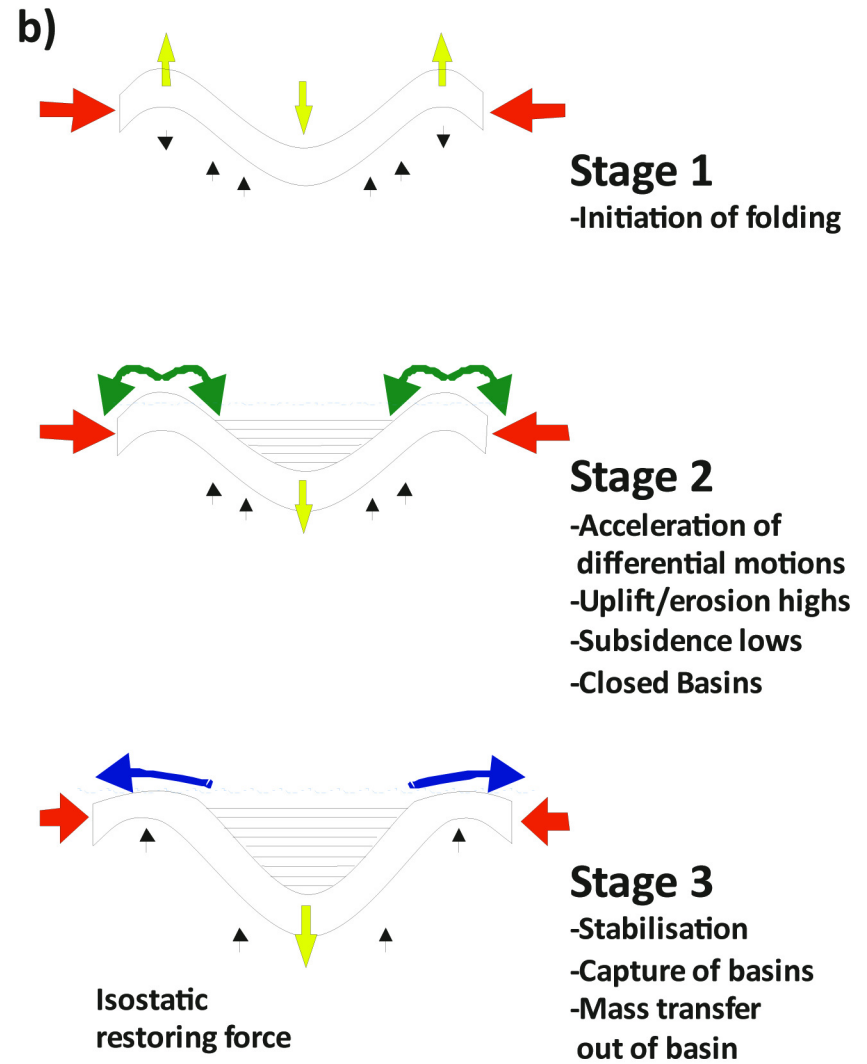
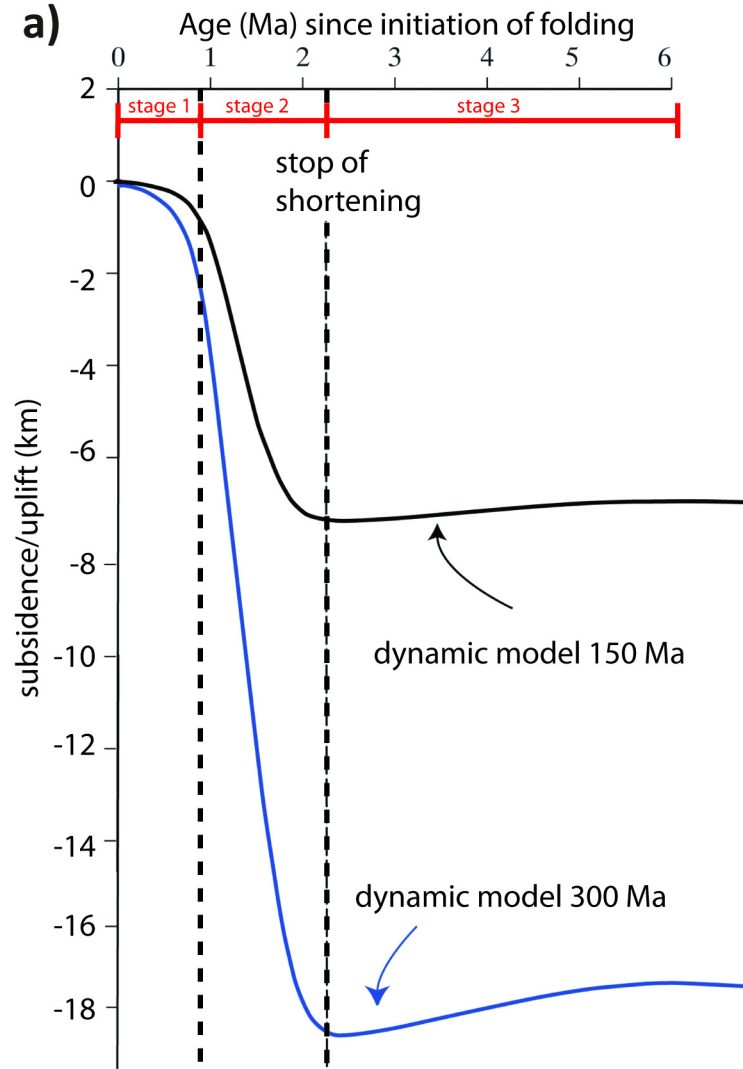


b) t=300 Ma, v=3.0 cm/yr



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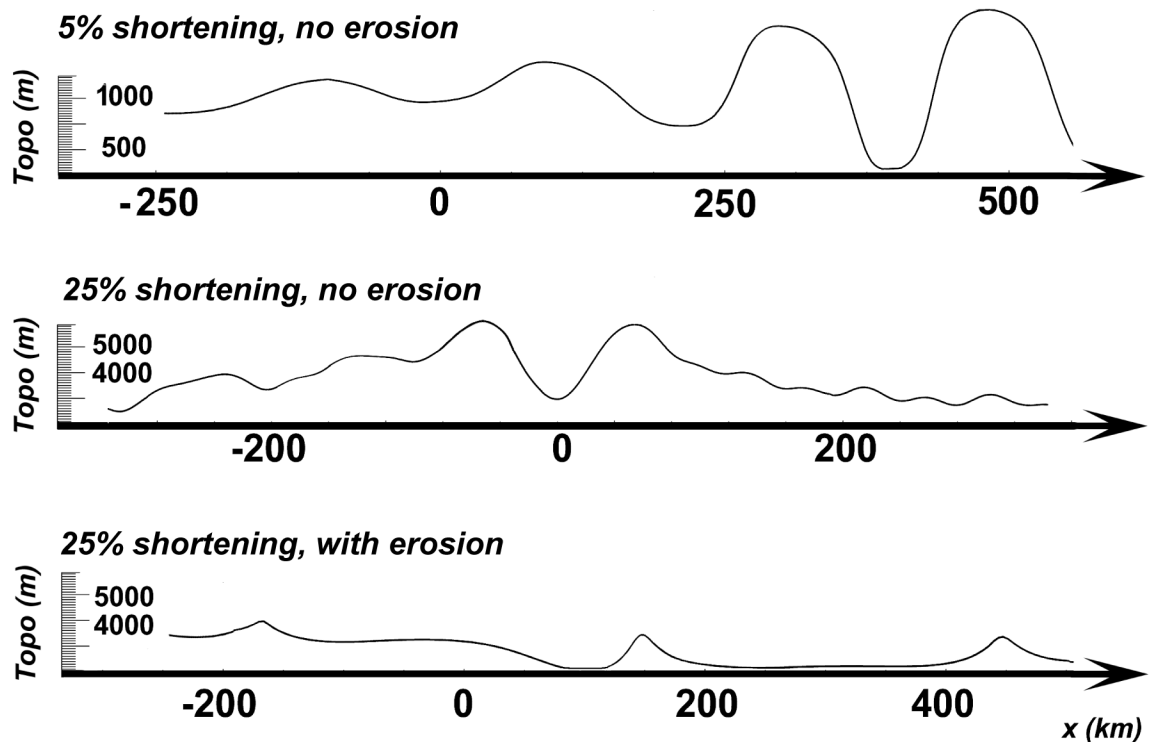
Figure 11



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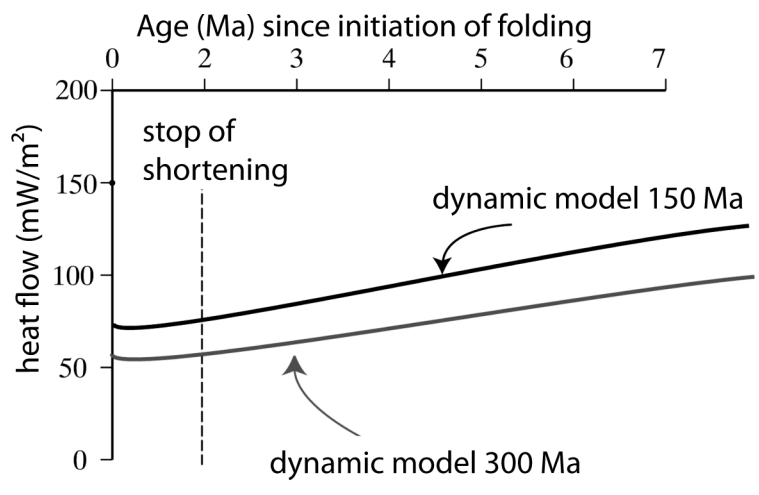
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Figure 12



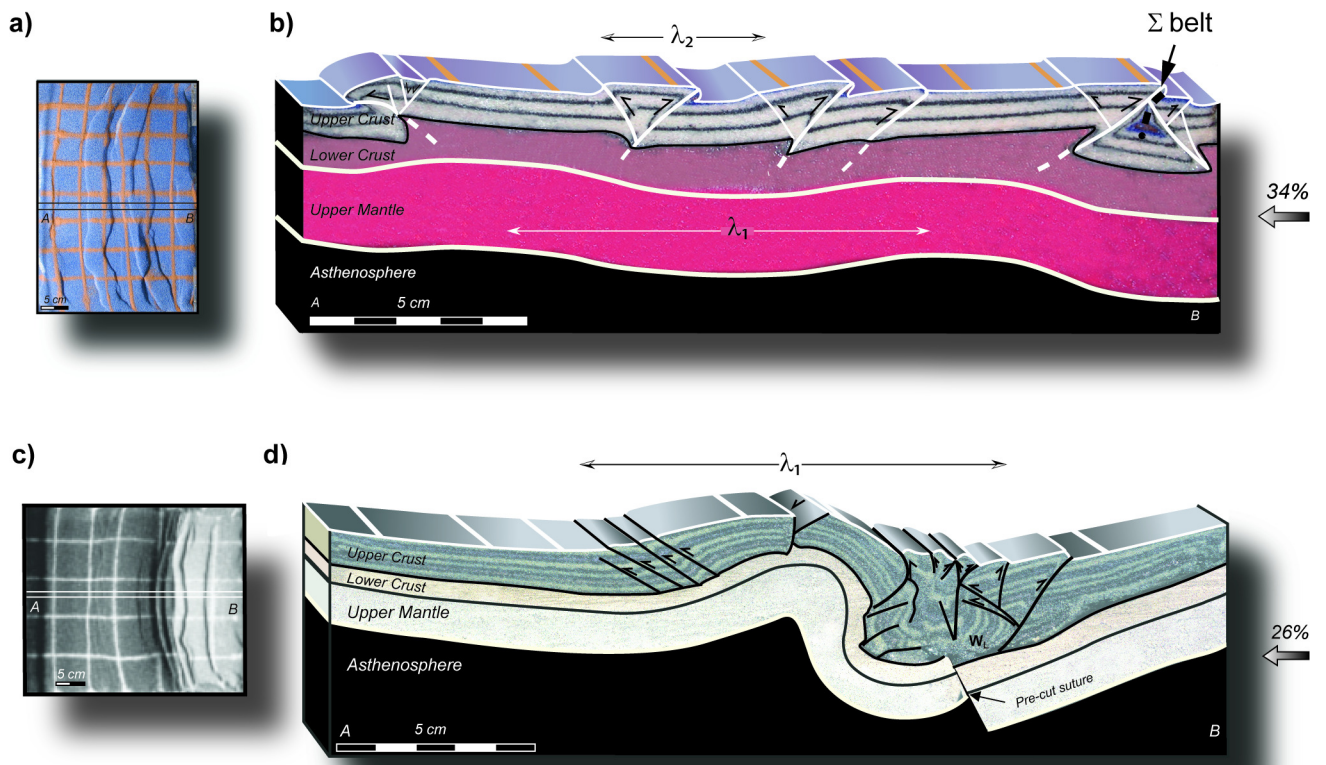
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Figure 13



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Figure 14



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Figure 15

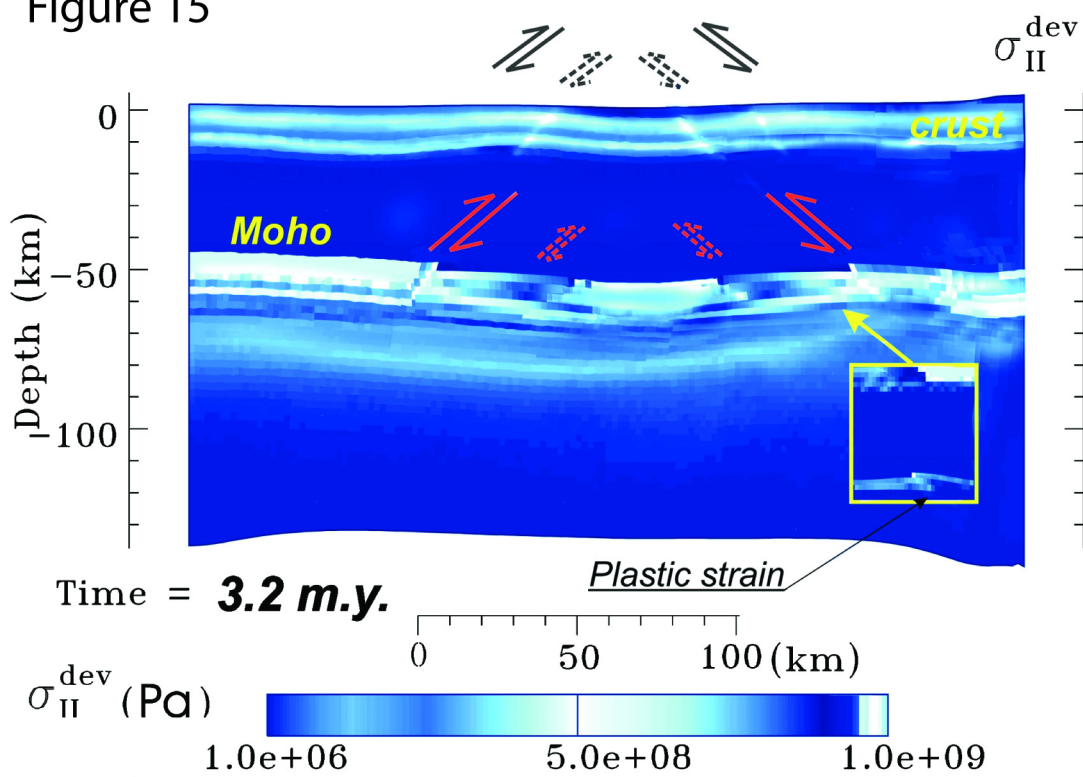


Figure 16

