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From orogenic hinterlands to Mediterranean-style back-arc basins: a comparative analysis

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Abstract: Hinterland plateaux and Mediterranean-style back-arc basins both form behind active subduction zones or collisional megathrusts, and share many characteristics: (1) early crustal thickening to about twice normal continental thickness; (2) thin lithospheric mantle; (3) mixed magmatism including asthenospheric, lithospheric, and crustal melts; (4) late-stage horizontal extension accompanied by vertical shortening. Horizontal extension and vertical shortening are driven by high gravitational potential energy (GPE) contrasts between the hinterland and surrounding lithosphere, which may reach $7 \times 10^{12}$ N m$^{-1}$, equivalent to about 2.5 times the ridge-push force. If extension is rapid relative to the rate of lithospheric cooling, GPE may remain positive even as extension continues, declining to the mid-ocean-ridge value as crustal thickness approaches zero. This suggests that hinterland plateaux could ultimately evolve into oceanic back-arc basins. The rate, direction, and amount of extension, and the rate of vertical shortening, depend on the plate boundary conditions and the GPE of the surrounding lithosphere. Vertical shortening in Tibet is limited by work required to deform the surrounding Asian lithosphere, whereas Mediterranean back-arc basins can extend at the expense of regions of thin continental or oceanic crust.

Continental collision zones are typically marked by a major thrust belt that accommodates a large proportion of the total convergence, behind which lies a hinterland plateau characterized by significant crustal thickening but relatively low strain rates. This plateau may in turn be flanked by thin- or thick-skinned thrust belts on its rear margins. The type example of this configuration is the Tibetan plateau, bounded by the Himalayan thrust belt to the south, and by convergent or transpressive zones to the north and east (Fig. 1). Similar configurations have developed along the southern margin of Asia further west, in the Iranian and Anatolian plateaux, and analogous structures may be developing in the incipient collision zone at the western end of the Sunda arc and in New Guinea. More generally, very similar structures have developed in the Andean and Cordilleran orogens on the western margin of the Americas, where large plateaux have formed through intracontinental shortening behind the major subduction zones along the Pacific rim. Several of these plateaux are currently undergoing some degree of normal faulting at high elevations, and in the US Cordillera this process has led to about 50% vertical shortening in the Basin and Range Province, reducing the crust to about half of its post-orogenic thickness (Fig. 2). A similar process has proceeded much further in the complex orogens of the Mediterranean region, where tightly arcuate thrust belts surround extended hinterlands, several of which have subsided below sea level, and some of which have rifted to form young ocean basins (Fig. 3). Some of these Mediterranean back-arc basins are clearly related to active subduction zones, and may therefore be directly analogous to the back-arc basins of the western Pacific; others, such as the Pannonian basin, the Alboran Sea and the northern Tyrrhenian Sea, have formed in essentially intracontinental settings, and the role of subduction is less clear.

The purpose of this paper is to make the case that collisional plateaux, extending Cordilleran hinterlands, and Mediterranean back-arc basins represent a spectrum of hinterland behaviour with many geological, tectonic, and mechanical features in common. I focus on three examples: the Tibetan plateau, the Basin and Range Province, and the Alboran Domain of the western Mediterranean (Fig. 4). These, and many other examples, share the following characteristics, documented in more detail below: (1) an initial stage of crustal thickening, to as much as twice normal continental thickness; (2) removal of much or all of the lithospheric mantle, either during or after crustal thickening; (3) mixed mode magmatism including asthenospheric, lithospheric, and crustal melts; (4) late-stage vertical shortening and horizontal extension (which may be coeval with continuing plate convergence), reflected by normal faults and conjugate sets of strike-slip faults in the upper crust; (5) core complexes exhuming mid- to lower crustal rocks below extensional detachments.

There are many marked differences between these examples, reflecting their differing geological settings, and differences in the relative importance of the tectonic processes involved. As is commonly the case in the Earth Sciences, identification of fundamental underlying processes requires us to sift through the intricate details of the geology to identify common themes.

The Tibetan plateau

Tectonic evolution

The Tibetan plateau was formed over the last 50 Ma as a result of the indentation of the relatively strong Indian plate into the southern margin of Asia, a process that started after the closure of the Tethys ocean at around 50–54 Ma (Rowley 1996). Since
2000) suggest, however, that by the end of the Mesozoic much of the surface of Tibet lay close to sea level. Given its likely structure and composition, this implies an overall crustal thickness in the range 25–35 km. This crust is likely to have been relatively weak: there is no reason to believe that it would have had a cold, dense, anhydrous, lower crustal layer as is found beneath continental cratons.

Present-day crustal and lithospheric structure

The crustal thickness beneath Tibet is not well defined, but appears to decrease from at least 70 km in the south to about 65 km in the north, where there is a correspondingly greater...
thickness of relatively low-density Palaeozoic sediment in the upper crust (Haines et al. 2003). The mechanism of crustal thickening is debated, but the present-day crustal velocity structure (Haines et al. 2003) is consistent with thickening of non-cratonic crust by a factor of two: hypotheses involving large-scale underthrusting (e.g. Powell & Conaghan 1973) or injection (Zhao & Morgan 1985) of Indian cratonic crust are not supported by the data. Few structures have been identified that are clearly associated with this phase of thickening in southern and central Tibet; Tapponnier et al. (2001) suggested that it occurred largely by reactivation of pre-existing suture zones.

Present-day crustal temperatures are relatively high: the seismic inferred depth to the $\alpha-\beta$ quartz transition suggests a temperature of 700°C at 18 km depth beneath central Tibet (Mechie et al. 2004), and xenolith data suggest temperatures of 800–900°C in the lower crust of northern Tibet (Hacker et al. 2000). These data imply that the lower 70% of the crust is above the wet granite solidus, and this is consistent with the interpretation of bright spots at about 20 km depth in the INDEPTH II seismic reflection line as zones of partial melt (Alsdorf et al. 1998). Seismic velocity and attenuation data suggest that the crust is predominantly felsic in composition, with high geothermal gradients (Galvé et al. 2006). The Phanerzoic history of Tibet suggests that much of the lower crust is made up of metasedimentary and former arc rocks, and these are likely to have been undergoing prograde metamorphism and partial melting since collision. The xenolith data, however, suggest that they may now be anhydrous, with granulite-facies mineral assemblages (Hacker et al. 2000).

Seismic velocity data suggest that the lithospheric mantle beneath much of northern Tibet is relatively thin and warm (McNamara et al. 1997), although this has been disputed on the basis of surface-wave analysis (Griot et al. 1998; Tapponnier et al. 2001). Low-density mantle beneath Tibet is also required to satisfy isostatic and gravity data (Jiménez-Munt et al. 2006). The results of the INDEPTH II profile suggest that cold Indian lithosphere may be underthrust horizontally beneath southern Tibet and the Lhasa block, perhaps as far as the Bangong suture (Hauck et al. 1998). This has occurred near the base of the Tibetan crust, implying that the mantle lithosphere beneath this part of Tibet had already been removed when underthrusting started, or was removed during underthrusting. Further north, tomographic inversion of P-wave velocities measured along the INDEPTH III profile indicates significant velocity heterogeneity in the upper mantle down to 300 km, with steeply dipping bodies of seismically fast material beneath central Tibet, and seismically slow material at shallow levels beneath the southern and northern parts of the plateau (Tilmann et al. 2003). The presence of large-scale structure in the upper mantle is also borne out by receiver function analysis, which has identified surfaces of impedance contrast dipping in various directions (Shi et al. 2004). These structures have been interpreted as subducting Indian lithosphere (Tilmann et al. 2003), delaminated lithospheric mantle (Shi et al. 2004), or convective downwellings in the lower part of a thickened thermal boundary layer (Houseman & Molnar 2001). Whatever the detailed interpretation, the tomography suggests circulation of material in the uppermost mantle driven by thermally induced density contrasts.

Evidence that southern Tibet may also have been underlain by anomalously warm mantle (prior to underthrusting by Indian lithosphere) comes from the young magmatic history, which includes K-rich mafic magmatism of mid-Tertiary to late Miocene age (Turner et al. 1996; Williams et al. 2001; Hou et al. 2004). The chemistry of this magmatism suggests a lithospheric contribution, which Turner et al. (1996) interpreted as indicating the convective removal and sinking of lithospheric mantle.

**Present-day tectonics of the Tibetan plateau**

The margins of the Tibetan plateau are defined by contractional or transpresssive zones: the Himalayan thrust belt in the south, which takes up about 40% of the total convergence between India and Asia, the transpressive Altyn Tagh fault in the north, which passes west into the Qilian Shan thrust belt, and a zone of...
folds and reverse faults in the Longmen Shan and the adjacent Szechuan Basin to the east (Fig. 1). A striking feature of these structures is that present-day contractional deformation is almost entirely limited to regions with average elevation less than 4000 m.

Late Miocene to Recent deforming features on the plateau proper are dominated by a linked pattern of strike-slip and normal faults (Armijo et al. 1986; Yin et al. 1999; Taylor et al. 2004), that act together to produce east–west extension, north–south contraction, and vertical shortening. The normal faults trend predominantly north–south, and they are linked by strike-slip faults trending ENE–WSW (left-slip) and ESE–WNW (right-slip), which at least in part act as transfer structures (Fig. 1). This pattern is confirmed by the seismicity: focal mechanisms indicate predominantly strike-slip and normal faulting (Molnar & Lyon-Caen 1989) where the surface elevation of the plateau exceeds 4000 m. These structures absorb a considerable amount of north–south horizontal contraction, a smaller amount of vertical shortening, and significant east–west extension, normal to the overall direction of plate convergence. The amount of vertical shortening is poorly constrained, but the number and geometry of visible graben on the plateau suggests that normal faulting may have caused about 40 km of east–west extension since their inception 10–14 Ma ago. This corresponds to a vertical strain rate of about $1 \times 10^{-16}$ s$^{-1}$. The strain pattern on the plateau therefore lies in the constrictional field, but may not be far from horizontal plane strain.

Although coverage is limited, GPS data across the Tibetan Plateau reveal a very clear and coherent kinematic pattern (Zhang et al. 2004). Northward velocity relative to Asia decreases steadily across the plateau, with the steepest gradients in the south, where c. 15–20 mm year$^{-1}$ of northward motion is accommodated within the 130 km wide Himalayan thrust belt, and in the north, where c. 9 mm year$^{-1}$ of northward motion is accommodated in the much broader Nan Shan and Qilian Shan thrust belts. Northward velocity decreases steadily across the 800 km wide high plateau in central and eastern Tibet, which absorbs 10–14 mm year$^{-1}$ of north–south shortening, giving an average strain rate on an azimuth of 020$^\circ$ of $4.7 \times 10^{-16}$ s$^{-1}$ (Zhang et al. 2004). This decrease in northward velocity is accompanied by an increase in eastward velocity, so that the vectors swing clockwise. Eastward velocity reaches a maximum of around 20 mm year$^{-1}$ in east–central Tibet, but decreases both northwards and eastwards, as relative motion is accommodated in the thrust belts to the north and east of the plateau. Further west there are fewer data, but the measurements suggest a simple pattern of decreasing northward velocity across the plateau without much change in the overall orientation of the velocity vectors. Change in baseline length between Leh in the west and Lhasa 1350 km to the east indicates that the plateau is extending in an east–west direction at $18 \pm 1$ mm year$^{-1}$ (Jade et al. 2004), and Zhang et al. (2004) estimated an overall rate of east–west extension of $22 \pm 3$ mm year$^{-1}$ for the plateau as a whole, giving an average strain rate on an azimuth of 110$^\circ$ of $6.8 \times 10^{-16}$ s$^{-1}$. Assuming conservation of volume, these figures for horizontal strain rates imply a vertical shortening rate of $2.1 \times 10^{-16}$ s$^{-1}$. This is about twice that estimated from the visible structures, but both estimates are subject to large uncertainties. Most of the deformation is being accommodated by distributed strike-slip and normal faulting within the plateau. Both the GPS and recent INSAR data from the western part of the plateau suggest that current slip rates on the Karakoram and Altyn Tagh faults are relatively low, of the order of a few millimetres per year (Wright et al. 2004), in contrast to the much higher estimates from palaeoseismology (e.g. Mériaux et al. 2004). The continuous nature of the velocity gradients revealed by GPS demonstrates that Tibet does not behave in a rigid plate-like fashion (Jade et al. 2004; Zhang et al. 2004).

**Discussion**

The pattern of normal and strike-slip faulting within the plateau strongly suggests that the present-day deformation is dominated by buoyancy forces associated with the high elevation (Molnar & Tappionnier 1978; England & Houseman 1986, 1988, 1991; Molnar & Lyon-Caen 1989; Flesch et al. 2001). This can be conveniently expressed in terms of the difference in gravitational potential energy (GPE) between the plateau and the surrounding Asian lithosphere (Fig. 5a, Table 1), which is about $7 \times 10^{12}$ N m$^{-1}$. GPE is simply the vertical integral of the vertical normal stress, and contrasts in GPE reflect the integrated difference in the vertical loads associated with elevation differences and contrasts in density structure between different columns of lithosphere (Platt & England 1994). GPE contrasts are equivalent to the ability of one lithospheric column to do work on the other, and are directly related to gradients in deviatoric stress. No other physically plausible mechanism exists to explain normal faulting on the plateau. Plate boundary forces associated with India–Asia collision involve horizontal compression, and acting alone should drive crustal thickening. Far-field forces from the subduction zones on the Pacific margin of Asia cannot be driving east–west extension in Tibet, given that eastward flow of material out of the Tibet is being fully absorbed within contractional belts on the eastern margin of the plateau (England & Houseman 1988).

The present high elevation and high GPE of the plateau are not simply the result of crustal thickening. A simple isostatic calculation shows that for reasonable crustal and mantle densities, doubling the thickness of the Asian lithosphere would result in a surface elevation of about 3000 m. This reflects the negative buoyancy of the resulting lithospheric root, which would be around 180 km thick. The present lithospheric structure and surface elevation indicate that most of this lithospheric root has been removed. England & Houseman (1989) suggested that it was removed by convective downwelling at some time after the main phase of thickening beneath the plateau, and Molnar et al. (1993) proposed, based on several separate lines of evidence, that convective removal occurred rapidly beneath the whole plateau at about 8 Ma. Tappionnier et al. (2001), on the other hand, suggested that the lithospheric mantle was removed continuously by delamination during the thickening process. These two hypotheses have very different implications for the mechanical evolution of the plateau, as well as the history of its surface elevation. Recent modelling by Jiménez-Munt & Platt (2006), which takes into account the effects of both internal heating and continuing convergence, suggests that only a relatively recent (<12 Ma) and rapid removal of lithosphere can explain the present topography, the continuing normal faulting on the plateau, and the pattern of eastward flow of material on the eastern margin of the plateau.

These hypotheses can be tested directly only by determination of the palaeo-elevation history of the plateau. This is both difficult and inaccurate, and widely varying conclusions have been reached by the application of techniques such as CLAMP analysis of fossil leaf morphology (Spicer et al. 2003), oxygen isotope analysis of soils (Currie et al. 2005), and $^{14}$C analysis of mammalian teeth (Wang et al. 2006). If the timing of onset of normal faulting can be regarded as a proxy for the timing of significant plateau uplift, then a consensus appears to be
emerging that this may have occurred at around 14 Ma (Coleman & Hodges 1995; Blisniuk et al. 2001; Clark et al. 2005). This timing is broadly consistent with the onset of widespread alkalic magmatism in mid-Miocene time, which may have been triggered by convective removal of lithospheric mantle (Turner et al. 1996; Williams et al. 2001; Chung et al. 2003; Hou et al. 2004).

Deformation of continental lithosphere is likely to be strongly influenced by rheological layering controlled by both temperature and composition. This has led to the suggestion that a weak, partly molten, lower or middle crustal layer has undergone channel flow, extruding eastwards around the margins of Tibet under the topographic load of the plateau (Clark & Royden 2000). Given the weakness of the middle and lower crust beneath Tibet, some degree of differential strain between different crustal layers is likely. Long-range channel flow, however, requires that the necessary pressure gradient be confined between rigid upper and lower layers. The lithospheric mantle is likely to be thin, warm and weak beneath much of central and northern Tibet, and the upper crust is undergoing active deformation. Hence the essential mechanical preconditions for channel flow may not be present. Beaumont et al. (2004), on the other hand, suggested that the Tibetan middle crust has been extruded southwards, emerging in the Himalaya beneath the South Tibetan Detachment as the High Himalayan crystalline unit. The South Tibetan Detachment is a relatively discrete low-angle normal fault system along the northern margin of the Himalaya: it was mainly active in the early Miocene (before 16 Ma, Searle et al. 2003), with a roughly north–south sense of extension. This structure is more likely to be related to the internal mechanics of the Himalayan orogenic wedge (Platt 1986; Royden & Burchfiel 1987) than to the Tibetan plateau, and its activity may in fact predate the final uplift of the plateau.

Conclusions

The Tibetan plateau formed by thickening of Asian continental lithosphere behind the evolving Himalayan thrust belt. Large-scale convective removal of lithospheric mantle beneath Tibet during the mid to late Miocene created the present high plateau, and the resulting high GPE with respect to the surrounding Asian
Table 1. Isostatic and GPE data and calculations

<table>
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<th>Region</th>
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<th>Iberia</th>
<th>Alboran</th>
<th>Atlantic</th>
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<td>1050</td>
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<tr>
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<td>2.22E+12</td>
<td>5.46E+12</td>
<td>2.54E+12</td>
<td>0.79E+12</td>
</tr>
</tbody>
</table>

GPE contrasts between two lithospheric columns A and B are calculated as

$$\Delta GPE = GPE_A - GPE_B = \sum \frac{1}{2} \rho_i g d_z$$

where

- $\rho_i$ is the density in kg m$^{-3}$,
- $g$ is the gravitational acceleration,
- $d_z$ is vertical distance (measured downwards),
- $h$ is an arbitrary depth of isostatic compensation 250 km below sea level,
- $T$ is the temperature at depth $z=250$ km below sea level.

The Basin and Range Province

Tectonic evolution

The extensional Basin and Range Province in the Cordillera of western North America developed during Tertiary time in what could be regarded as the hinterland to both the west-directed accretionary wedges on the western active margin, and the Mesozoic to Early Tertiary east-directed thrust belts along the eastern margin of the Cordillera (Burchfiel & Davis 1972) (Fig. 2). The province is broadest and most clearly defined in the Great Basin, between the elevated regions of the Sierra Nevada and the Colorado plateau, but a comparable region of Tertiary extensional tectonics extends north into the central Canadian Cordillera (Brown & Journee 1987; Parish et al. 1988; Vanderhaeghe et al. 2003), and south into SE California, Arizona and the Mexican state of Sonora (Davis & Listner 1988; Spencer & Reynolds 1991; Henry & Aranda-Gomez 1992).

The crust that underlies the province is in part derived from the rifted passive margin that developed on the western side of cratonic North America from mid-Proterozoic to Early Palaeozoic time (the Cordilleran miogeocline), and in part from arc and marginal basin sequences that were progressively accreted to the continental margin during the Late Palaeozoic and Early Mesozoic (Burchfield & Davis 1972; Oldow et al. 1989). In SW Arizona Basin and Range extension transgresses onto former cratonic crust, whereas in the Canadian Cordillera the upper crust of the hinterland is made up entirely of accreted terranes.

Basin and Range crust was initially substantially thickened, initially by accretionary events such as the Late Palaeozoic Antler and Early Mesozoic Sonoma orogenies, and then by the main phase of Cordilleran thrusting, accompanied by granitic magmatism, in Late Jurassic to mid-Cretaceous time (Miller & Gans 1989; Camilleri & Chamberlain 1997; Taylor et al. 2000). Rocks exhumed in the metamorphic core complexes along the axis of the province were metamorphosed at depths of 25–35 km in mid-Cretaceous time (Hodges & Walker 1990, 1992; Lewis et al. 1999; McGrew et al. 2000), and are currently underlain by crust about 30 km thick, suggesting that crustal thicknesses may have reached 60 km or more (Axen et al. 1993). The contribution of the Eocene Laramide orogeny to crustal thickening is less clear: the basement-cored uplifts produced in this event lie well to the east of the province, and there is no geological record of Laramide events within the Basin and Range itself. To the south, the Cordilleran thrust belt trends SW–NE and is truncated by the Mesozoic active margin (Burchfield & Davis 1972). Further south, in SE California and Arizona, crustal thickening may have been primarily magmatic: Mesozoic granites emplaced at up to 30 km depth have been exhumed in detachment footwalls (Anderson et al. 1988; Foster et al. 1990). To the west, clastic sediments from the forearc were underplated beneath the arc at the end of the Mesozoic (Rand, Pelona, and Orocoica schists), and were subsequently exhumed in core complexes close to the later San Andreas Fault (Jacobson et al. 1996, 2000).
The Basin and Range became the locus of widespread broadly calc-alkaline volcanism in mid-Tertiary time, starting during the Eocene in the north, and migrating southward, reaching the southern Great Basin by early Miocene time (Wernicke et al. 1987; Gans et al. 1989; Armstrong & Ward 1991; Axen et al. 1993). A comparable northward migration of magmatism took place from Sonora, where it started in Oligocene time, and reached SE California in the early Miocene. Melts were primarily sourced in lithospheric mantle of Proterozoic age (Hawkesworth et al. 1995). The heat source has been attributed to convective removal of the lower lithosphere (Sonder et al. 1987) or to the foundering of a flat subducted oceanic slab emplaced during the Laramide event (Humphreys 1995). Magmatism for the most part immediately preceded the start of widespread extension in the province (Wernicke et al. 1987; Gans et al. 1989; Axen et al. 1993).

**Present-day crustal and lithospheric structure**

Basin and Range crust is currently 28–35 km thick (Hauser et al. 1987; Jones et al. 1992; McCarthy & Parsons 1994), implying that it has been thinned by a factor of about two since it reached peak thickness in mid-Mesozoic time. Average elevation over the Great Basin is around 1600 m, decreasing southwards into the Colorado extensional corridor. The high heat flow characteristic of the province, the widespread magmatism, and the fairly constant crustal thickness beneath regions showing varying amounts of extension in the surface structure have all been used to suggest that the crust is very weak, and that the lower crust has undergone substantial lateral flow (McCready et al. 1997; Gans 1989; Jones et al. 1992). With a few exceptions, the high-angle normal faults that define the characteristic present-day morphology of the province cannot be traced in seismic lines to depths greater than about 10 km; low-angle normal faults (detachments) can be traced to mid-crustal depths, where they are around horizontal (Allmendinger et al. 1987; Hauser et al. 1987). The contrast between the upper crust, deformed by arrays of brittle normal faults, and the highly ductile middle and lower crust implies the existence of a major zone of differential movement and shear separating the two (Hamilton 1987). This zone, represented by several kilometres thickness of mylonite and high-strain gneiss, has been exhumed in the footwalls of low-angle detachment faults to form metamorphic core complexes (Davis & Lister 1988).

The thickness and structure of the lithospheric mantle beneath the Basin and Range Province is poorly known, although there is general agreement that it is likely to be both thin and warm. Lowry et al. (2000) calculated depths of 35–45 km to the 10^21 Pa s isopose (the rheological base of the lithosphere) from flexural rigidity and heat-flow data, which suggests that most if not all of the original North American lithospheric mantle has been removed. The mechanism by which this occurred is unclear, but the close relationship between the region of thin lithosphere and the Cordilleran thrust belt suggests that a process triggered by the previous history of lithospheric thickening is likely. The lithospheric mantle after Cordilleran intracrustal shortening should have been thicker (>150 km) than in the surrounding regions, favouring some form of convective removal, as proposed by Sonder et al. (1987) and Ranalli et al. (1989). Alternatively, Gemmer & Houseman (2005) have investigated a situation where thick crust overlying normal thickness mantle starts to extend under buoyancy forces, leading to an upward perturbation in the base of the lithosphere. This could then trigger convective upwelling beneath the extending region, and laterally propagating downwelling beneath the extending region, and laterally propagating

**Extensional tectonics in the Basin and Range**

The time at which extension started in the Basin and Range Province is highly uncertain, and extension is likely to have been diachronous. There may have been more than one stage of extension, and more than one cause. Mid- to Late Cretaceous extension has been suggested in several locations on the basis of both structure and cooling ages (Hodges & Walker 1990; Applegate & Hodges 1995; Camilleri & Chamberlain 1997; Wells et al. 2005), but the regional significance of this is as yet unclear. Extension at this stage was coeval with contractional deformation in the adjacent Cordilleran thrust belt, and it is likely that Proterozoic basement of the North American craton was being emplaced directly beneath the areas undergoing extension. This type of syn-convergence extension is most likely to be related to the internal mechanics of the thrust wedge itself (Platt 1986).

Post-convergence extension appears to have started first in the core complexes of the Canadian Cordillera in Eocene time (56–48 Ma; Vanderhaeghe et al. 2003), where it followed the end of contractional deformation in the thrust belt by not more than a few million years. There is also evidence for extension in late Eocene time (48–41 Ma) in the northern Great Basin (Mueller &
Snoke 1993; Lee 1995; Wells et al. 2000; Rahl et al. 2002). The main phase of activity on the low-angle detachments that bound the metamorphic core complexes along the axis of the province took place in late Eocene to Late Oligocene time (36–23 Ma) throughout the Great Basin, following the onset of calc-alkaline volcanism (Dokka et al. 1986; Gans et al. 1989; Mueller & Snoke 1993; Wells et al. 2000). Further south, in Arizona, SE California, and Mexico, detachment faulting started later (26–15 Ma) (Davis & Lister 1988; Foster et al. 1990; Henry & Aranda-Gomez 1992; Wong & Gans 2003; Carter et al. 2004), and as late as middle Miocene in Death Valley (Hoisch et al. 1997). During the Late Tertiary, extensional deformation migrated outwards in the Great Basin, both west towards the Sierra Nevada and east towards the Colorado plateau (Gans et al. 1989). The total extension is estimated at around 100%; over a 35 Ma period this corresponds to an average rate of extension of \(6 \times 10^{-16} \text{ s}^{-1}\), although at times it may have been substantially greater than this.

The direction of extension in the Basin and Range Province varies both spatially and temporally. Slip on the detachment faults of the Great Basin in Eocene to Oligocene time was mainly WNW- or ESE-directed (Miller et al. 1983; Wust 1986; McCready et al. 1997; Wells et al. 2000; Rahl et al. 2002). The north–south trend of the later high-angle normal faults that dominate the present morphology suggests a swing to east–west extension in the Miocene, and the current direction of extension across most of the Great Basin is c. WNW–ESE (290° azimuth) (Thatcher et al. 1999; Bennett et al. 2003). Late Miocene to Recent extension directions in the Death Valley area and the western Great Basin, however, swing to the NW–SE (Unruh et al. 2003; Faulds et al. 2005), reflecting increasing influence of the Pacific–North American transform boundary. Further south, in the SE California, Arizona, and Sonora, extension directions have been uniformly SW–NE-directed since the start of extension (Wust 1986; Davis & Lister 1988; Spencer & Reynolds 1991; Henry & Aranda-Gomez 1992; Wong & Gans 2003). The current rate of extension is close to the limit of measurement by GPS or seismology, and it is difficult to separate out the effects of lithospheric thinning from transform-related deformation, but it seems likely that thinning-related strain amounts to around 6 mm year\(^{-1}\) horizontal extension across the 800 km extending region in the northern Great Basin (Thatcher et al. 1999), equivalent to a strain rate of \(2.4 \times 10^{-16} \text{ s}^{-1}\).

**Discussion**

The remarkable feature of Basin and Range kinematics is that with the exception of post-12 Ma deformation in the westernmost Great Basin, where transform-related deformation has become important, extension directions show little relation to either relative or absolute plate motions. They do, however, show a simple relation to the trend of the continental margin, and to gradients in GPE (Fig. 2). Present-day GPE (Jones et al. 1998; Sonder & Jones 1999; Coblenz & Humphreys 2006) has been reduced by crustal thinning in the Basin and Range Province (Fig. 5b), and present-day extension rates are low, so it is difficult to test this hypothesis rigorously. If we assume the existence during Late Eocene to early Miocene time of an elevated crustal welt along the axis of the North American Cordillera from Canada to Mexico, with crustal thickness up to 60 km and thin (c. 20 km) lithospheric mantle beneath it, then the predominant gradients in GPE in the past (Fig. 5c, Table 1) should have been normal to the continental margin. Available data on extension directions throughout the province are consistent with this. The regional stress field may have been influenced by forces associated with the evolving convergent plate-boundary during the Tertiary, but active subduction zones may be associated with both contractional and extensional deformation in their hinterlands, and we lack any physical basis for predicting this behaviour. The observational evidence suggests that Basin and Range extension was not primarily driven by plate-boundary forces.

**Conclusions**

The Basin and Range extensional province has formed along the locus of maximum crustal thickening in the Cordilleran orogen, and shows evidence for removal of much of the underlying lithospheric mantle. This may have started in late Eocene to Oligocene time, and be continuing today on the margins of the province. The resulting gradients in GPE, relative to Pacific Ocean lithosphere, are the most likely driving forces for mid-Tertiary to Recent extension, which has been predominantly directed normal to the continental margin and to the trend of the Cordillera. A combination of spatially variable extension and lower crustal flow has reduced crustal thicknesses throughout most of the province to some 30 km: the present-day surface elevation reflects the low density of the immediately subjacent mantle. Reduced GPE contrasts have resulted in a substantial slowing of the rate of extension to the present.

**The Alboran Domain of the western Mediterranean**

**Tectonic evolution**

The contractional orogens and extensional basins of the Mediterranean region (Fig. 3) have formed since Late Mesozoic time in a tectonic environment dominated by the northward relative motion of Africa with respect to Eurasia (Dewey et al. 1989). During the Late Tertiary, a distinctive pattern has evolved of arcuate thrust belts or accretionary wedges surrounding extensional basins floored by thinned continental or new oceanic crust (Jolivet & Facenna 2000). Some of these arc–basin systems are associated with clear-cut subduction zones: the Aegean–Hellenic and Calabrian–Tyrrhenian systems are both underlain by seismically active and tomographically imaged zones along which Neotethyan oceanic lithosphere of the eastern Mediterranean is being subducted (Anderson & Jackson 1987; Meulenkamp et al. 1988; Facenna et al. 2001). In the Carpathian–Pannonian, Apennine–northern Tyrrhenian, and Betic–Rif–Alboran systems the evidence for the subduction of oceanic lithosphere is fragmentary, disputed or absent; the thrust belts were largely formed by shortening the old passive continental margins created during the early Mesozoic break-up of Pangaea. Tectonic analysis is complicated by the existence of several sub-plates (e.g. Iberia and Adria) and numerous micro-continental fragments, the kinematics of which are poorly known.

The Betic–Rif–Alboran system is in some ways the most dramatic of these arc–back-arc basin pairs, as it is developed right in the collision zone between Africa and Iberia (Fig. 4). The system comprises the west-facing external Betic–Rif arc, which surrounds an extensional hinterland known as the Alboran Domain (Fig. 4). In the early Mesozoic the area experienced sinistrally oblique extension between Africa and Iberia, which opened up a narrow seaway connecting the Atlantic to the Neotethys ocean (Dewey et al. 1989; Roest & Srivastava 1992). During the Late Cretaceous and Tertiary, Africa moved between 200 and 500 km north with respect to Iberia. The convergence direction changed to NW in the Late Tertiary (Dewey et al.
The Alboran Sea is underlain by crust currently between about 12 and 20 km thick (Torné et al. 2000). Metamorphic rocks recovered from Ocean Drilling Program (ODP) Site 976 in the western Alboran Sea are comparable with those exposed onshore in the internal Betic–Rif orogen (Platt et al. 1998; Soto & Platt 1999), so that these regions are grouped together as the Alboran Domain. In the east much of the Alboran Sea is underlain by Miocene calc-alkaline volcanic rocks (Comas et al. 1999), and the crust thins eastwards into Miocene oceanic crust of the Balearic Basin (Mauffret et al. 2004). Peripheral areas of the Alboran Domain were emplaced onto the Iberian and African margins during the thrusting that formed the external Betic–Rif arc (Fig. 4). These emergent areas are underlain by crust up to 39 km thick (Banda et al. 1993), but only the top 10 km or so consists of rocks belong to the Alboran Domain. The remainder presumably comprises basement and cover rocks of the two continental margins (Watts et al. 1993; Carbonell et al. 1998), emplaced beneath the Alboran Domain during formation of the external thrust belt.

Interpretation of the upper mantle structure beneath the Alboran region is highly contentious. It is clear that the lithosphere is thin beneath much of the Alboran Sea. Torné et al. (2000) estimated a thickness of 45 km from heat-flow data, increasing to c. 100 km under the West Alboran Basin, and Calvert et al. (2000) estimated a thickness of c. 60 km under the central Alboran Sea from P-wave velocity data. Seismic tomography suggests the existence of one or more large bodies of cold mantle at depths between 150 and 600 km (Blanco & Spakman 1993; Calvert et al. 2000): these have been interpreted as the remains of a detached subducting slab of Tethyan oceanic lithosphere (Carminati et al. 1998), a convective downwelling beneath a region of previously thickened continental lithosphere (Platt & Vissers 1989), or an actively subducting slab of Atlantic oceanic lithosphere dipping east beneath the Alboran Sea (Gutscher et al. 2002). Isolated earthquakes at around 600 km depth presumably occurred within this body (Grimison & Chen 1986); shallow seismicity in a steep zone down to 150 km beneath the south coast of Spain may indicate a zone of present-day mantle downflow (Calvert et al. 2000).

Interpretation of the mantle structure in terms of an actively subducting slab dipping east beneath the arc, as proposed by Gutscher et al. (2002), faces three difficulties. First, there is no support from seismicity, geodesy or surface structural data for continuing east–west convergence (Platt & Houseman 2003; Fadil et al. 2006). Second, a continuous slab more than 600 km long, as proposed, requires more than twice the amount of convergence in the Betic–Rif arc that is estimated from the structural data (Platt et al. 2003a). Third, a slab of this length should be generating a magmatic arc where it reaches 125 km depth, near the Straits of Gibraltar. No such arc is observed: the volcanic rocks in the Alboran Domain are mainly Miocene in age, and lie east of the downdip termination (at 600 km depth) of the proposed slab.

The thin lithosphere beneath the Alboran Sea and the presence of cold mantle at depth suggest large-scale convective circulation of material, and the present mantle structure is likely to reflect that process (Calvert et al. 2000). The current pattern of crustal motion in the Rif (Fadil et al. 2006), and the presence of young (2–10 Ma) alkali basalts and potassic lavas interpreted as lithospheric melts (Turner et al. 1999; Duggen et al. 2004) to the north and south of the Alboran Sea in southern Spain and Morocco, may indicate that the original downwelling beneath the Alboran Sea has propagated laterally outward, as suggested by Houseman & Molnar (2001) and Platt & Houseman (2003).

**Present-day crustal and lithospheric structure of the Alboran Domain**

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**Extensions tectonics in the Alboran Domain**

A phase of rapid extension started close to the beginning of the Miocene (24 Ma), and continued until about 18 Ma (Platt et al. 2003b): this resulted in the exhumation of rocks from various depths down to the base of the crust and locally the underlying upper mantle, now exposed as the Ronda and Beni Bouensa peridotite massifs (Tubia et al. 1993; Bouybaouène et al. 1998; Argles et al. 1999; Platt et al. 2003c). The orogenic crust was stretched and thinned from as much as 50–60 km to as little as 5 km in places: rates of exhumation reached 9 mm year$^{-1}$ in the most deep-seated rocks. Exhumation and decompression were accompanied locally by heating, producing a distinctive low-pressure–high-temperature metamorphic overprint on more deeply exhumed rocks (Platt et al. 1998, 2003c; Soto & Platt 1999; Negro et al. 2006). Much of the Alboran Domain had subsided below sea level by the end of this phase of extension, and parts of the west Alboran basin have accumulated as much as 8 km of Miocene–Recent sediment (Comas et al. 1999).

The early Miocene phase of extension produced a range of structures, including a major detachment that separates largely unmetamorphosed rocks in the hanging wall (known as Malá-
Late Miocene sedimentary basins (Vissers et al. 2003) created several west- to WSW-directed transport (Jabaloy et al. 1993, 1997; Lonergan & Platt 1995; Platzman & Platt 2004). Palaeomagnetic data suggest that the original slip direction on the Maláguide–Alpujarra detachment was NNE (Feinberg et al. 1996; Platzman & Platt 2004). Brittle normal faults formed towards the end of the exhumation history show displacement predominantly towards the north (Crespo-Blanc 1995).

Extension and westward motion of the Alboran Domain was accommodated by coeval contraction in the Iberian and African continental margins and the basin that lay between them, forming the external Betic–Rif arc in the process. The lithospheric mantle beneath the external arc was presumably subducted or delaminated. Direct evidence for subduction comes from a distal part of the Iberian margin (the Nevado–Fila-bride Complex, Fig. 4), which was subducted beneath the Alboran Domain in the early Miocene, reaching eclogite facies at around 17 Ma (Platt et al. 2006). This rock body was then rapidly exhumed during the mid- to late Miocene beneath a second major detachment, with west- to WSW-directed transport (Jabaloy et al. 1993; Martínez-Martínez et al. 2002). Brittle normal faults associated with the final stages of extension slip mainly NE–SW, and created several Late Miocene sedimentary basins (Vissers et al. 1995). The detachment can be traced in the transport direction for >160 km, and the footwall rocks show a significant increase in the maximum pressure of metamorphism from around 1200 MPa in the east to >1600 MPa in the west (Platt et al. 2006), suggesting that the detachment footwall had a SW component of original dip. Ar–Ar ages suggest that cooling started immediately following high-P metamorphism (Platt et al. 2006), and fission-track data and the arrival of Nevado–Fila-bride sediment in adjacent basins show that exhumation was largely complete by 12 Ma in the east and 10 Ma in the west (Johnson & Harbury 1997). The overall rate of exhumation of the eclogitic rocks was therefore around 8 mm year$^{-1}$.

Extension in the Alboran Domain was accompanied by volcanism, which started at around 23 Ma with a swarm of tholeiitic dykes (Platzman et al. 2000), and continued through the Miocene with predominantly andesite to rhyolite compositions. These rocks have been interpreted as related to subduction (Gutscher et al. 2002; Duggen et al. 2004), or as asthenospheric melts related to extension, with variable amounts of crustal assimilation (Turner et al. 1999). Lithosphere-derived alkali basalt and highly potassic volcanic rocks were erupted onshore in the eastern Betics and Rif during the late Miocene (Duggen et al. 2004).

**Discussion**

Three critical features of the extensional tectonic event in the Alboran Domain guide thinking about its causes, and stimulate comparison with the evolution of the Tibetan plateau and the Basin and Range province.

1. Extension was localized in a region that had undergone previous crustal thickening.
2. Extension was accompanied by significant heat transfer into the extending crust. This is most clearly demonstrated by the $P$–$T$ path from ODP Site 976, where temperature rose by around $100^\circ$C during decompression from 1050 to 350 MPa (Soto & Platt 1999). Decompressional $P$–$T$ paths from onshore exposures in the Alboran Domain are approximately isothermal, but thermal modelling suggests that substantial heat input is needed to explain isothermal decompression of rocks at around 800°C to depths as shallow as 15 km (Platt et al. 2003c).
3. Extension was accompanied and followed by mixed-mode magmatism, localized in the area of greatest crustal thinning, and resulted from decompression melting of asthenospheric mantle at shallow depths, with contributions from lithospheric and crustal melts.

These features suggest that high GPE associated with thick crust and thin or absent lithospheric mantle was a factor contributing to extension (Fig. 5). Removal of lithospheric mantle would have caused a significant increase in GPE, triggering extension; and would also have resulted in mantle melting and conductive heat transfer into the crust. It is important to note that a combination of two separate processes is required to produce the observed thermal effects: (1) removal of lithospheric mantle, creating a heat source at depth; (2) vertical shortening and horizontal extension, to reduce the thickness and hence the rate required for conductive heat transfer though the remaining crust and lithosphere mantle (Platt & England 1994). Vertical shortening alone does not result in heating, except transiently in the vicinity of major discontinuities.

The distinctive metamorphic features of the Alboran Domain, which have not been recorded from Tibet or the Basin and Range Province, resulted from the very large amounts and very high rates of extension that it experienced. This was in turn a result of the small scale of the system, and the fact that extension could be accommodated at the expense of lithosphere with thin continental or oceanic crust. Large-scale extension resulted in subsidence, and the clearest evidence for the thermal history is found in rocks recovered from 2000 m below sea level at ODP Site 976.

**Conclusion**

The Alboran Domain formed on the site of an Eocene accretionary orogen that reached a thickness of around 60 km. Removal of much or all of the lithospheric mantle beneath it triggered rapid extension in the early Miocene, accompanied by mixed mode magmatism. Decompressional metamorphism in exhumed deep-seated rocks was accompanied by significant input of heat from the underlying asthenospheric mantle. This phase of extension occurred by NNE-directed slip on extensional detachments, and by ductile stretching and shear in the same sense and direction, approximately parallel to the orientation of the orogen at this time. Extension was accommodated largely by consumption of thin continental or possibly oceanic crust between the converging African and Iberian margins, but also resulted in shortening and thickening of the continental margins themselves. Peripheral regions of the Alboran Domain were emplaced onto the two continental margins, and rocks adjacent to the Iberian margin were subducted beneath the Alboran Domain and were then rapidly exhumed beneath a middle to late Miocene detachment.
Discussion and conclusions

The crucial questions with respect to all three of these extending terranes are as follows.

1. What drives lithospheric thinning, and what controls the rate and direction of extension?

2. When did lithospheric thinning start, how long will it last, and when will it stop?

3. Can we predict the end product in terms of crustal and lithospheric thickness?

The fact that in all three cases extension was initially localized within regions of thickened crust clearly suggests that high GPE associated with the buoyancy of this crust is an important factor. In all three cases, however, there is also evidence for thin or even absent lithospheric mantle. Removal of lithospheric mantle contributes to positive GPE not only by its effect on surface elevation, but also by changing the overall density structure of the lithospheric column with respect to its surroundings. Increasing the thickness of continental lithosphere may not increase overall GPE by much, and may even decrease it (England & Houseman 1989; Platt & England 1994), but if the lithospheric mantle is removed, the GPE contrast can reach values comparable with the largest energy differences that drive plate motions.

GPE contrasts are only one possible driver for lithospheric deformation: others include far-field forces transmitted through plates as a result of their rigidity, forces exerted on plate boundaries, and integrated tractions exerted on the base of plates by the flowing mantle beneath (Ghosh et al. 2006). All of these forces may contribute to the net state of deviatoric stress in an extending region, and not all of them are easily quantifiable. Strain rates will depend on the rheology of both the extending lithosphere and the deforming region that is accommodating the resulting displacements.

In the case of Tibet, we can make some useful estimates of the relative magnitude of the deviatoric stresses. The horizontal north–south deviatoric stress $\tau_{xx}$ associated with India’s northward motion drives north–south shortening in Tibet, but the excess GPE of the plateau prevents thickening in response, and in fact drives vertical thinning, albeit at a significantly lower rate. The vertical deviatoric stress $\tau_{zz}$ is therefore small, and the east–west deviatoric stress $\tau_{yy}$ $\approx -\tau_{xx}$. As a result, the extension direction is horizontal and east–west, and is accommodated primarily by the deformation of adjacent Asian lithosphere to the NE, east and SE of the plateau.

How will the Tibetan plateau evolve in the future? The current episode of vertical shortening may be almost over (Jiménez-Munt 2006): as full-thickness Indian and Asian lithosphere continue to converge, the region underlain by thin lithospheric mantle will be progressively reduced, GPE will be reduced, and vertical shortening will cease. The GPE will nevertheless still be sufficient to drive plane strain deformation within the plateau and horizontal contraction in the surrounding lithosphere, and as long as India continues its northward motion, the plateau will continue to grow outwards.

Modelling of the way Rayleigh–Taylor instabilities grow and propagate in the lithospheric mantle beneath regions of buoyant continental crust suggests that, once started, the process may continue by lateral propagation of the downwellings (Molnar & Houseman 2004). The complex mantle structure beneath Tibet at the present day suggests that this may already be happening, in which case vertical shortening may continue while continental convergence lasts. The current low rate of thinning ($c. \ 0.2 \ \text{mm \ year}^{-1}$), however, suggests that the crust is unlikely to return to a normal continental thickness ($c. \ 35 \ \text{km}$), as it has in the present-day Basin and Range, and the region will certainly not subside below sea level.

How would Tibet evolve should India–Asia convergence cease? This would be the ultimate test of the role of locally generated forces on the tectonics of the plateau. If $\tau_{xx}$ is zero, the situation reduces to plane stress in the east–west vertical plane, with $\tau_{zz} = -\tau_{yy}$. If the resistance to deformation offered by the surrounding Asian lithosphere remained unchanged, $\tau_{yy}$ would have about half its present-day value, and the dominant mode of deformation would be vertical shortening and east–west extension. Depending on the rheology of the Tibetan lithosphere, we might expect the rate of vertical shortening to be as much as half the current rate of east–west extension. This could lead to thinning of the 70 km thick crust by $c. \ 0.5 \ \text{mm \ year}^{-1}$, significantly faster than the current rate. The plateau might therefore evolve towards something like the present-day Basin and Range Province over several tens of million years. Reduction in crustal thickness would reduce the GPE contrast, as would lithospheric cooling, hence it is unlikely that thinning would continue indefinitely.

The situation in the Basin and Range Province is clearly different: the main phase of crustal thinning did not occur in an environment of continental indentation, and the average rate of vertical shortening has been relatively high ($6 \times 10^{-16} \ \text{s}^{-1}$), approximating vertical east–west plane strain. Extension was accommodated primarily at the expense of oceanic lithosphere, either by subduction prior to the evolution of the present transform boundary, or by preventing sea-floor spreading between the Pacific and North American plates since the inception of the San Andreas transform. In either case, the GPE contrast should be measured with respect to young ocean lithosphere (Fig. 5, Table 1), and may have amounted to about $5.5 \times 10^{12} \ \text{N} \ \text{m}^{-1}$ at the start of major extension. The rate of deformation is controlled primarily by the rheology of the Basin and Range lithosphere, as little work is required to drive subduction, or to prevent sea-floor spreading. The present-day GPE contrast has been reduced by thinning and subsidence to $2.2 \times 10^{12} \ \text{N} \ \text{m}^{-1}$, but is still driving deformation at $2.4 \times 10^{-16} \ \text{s}^{-1}$, thinning the 30 km crust by 0.2 mm year$^{-1}$. What then is the future for this region? Cooling and subsidence are likely to significantly decrease GPE over the time scale at which further extension (at current rates) becomes significant. The GPE contrast that drives extension may therefore dissipate before much more extension occurs. The ultimate fate of the province is therefore likely to be thermal subsidence to form a shallow marine basin along the axis of the Cordillera, separating the active margin of North America from the Rocky Mountains and the cratonic interior.

The Alboran Domain is different again. Although this region of extension formed in the middle of a zone of continental convergence, its GPE relative to its surroundings was high enough to prevent internal horizontal shortening. The greatest contrast ($c. \ 7 \times 10^{12} \ \text{N} \ \text{m}^{-1}$ at the start of extension) was with the thin crust to the south and west (Fig. 5, Table 1). As a result, the Alboran Domain extended to the SW, thinned rapidly and subsided below sea level. The rate of deformation was high because of the small scale of the system, and the fact that little work was required to drive shortening in the external thrust belt to the SW.

Positive GPE contrasts of $c. \ 5.1 \times 10^{12} \ \text{N} \ \text{m}^{-1}$ between the Alboran Domain and the continental crust to the north and south meant that north–south plate convergence was accommodated at the expense of the Iberian and African continental margins. The lack of a significant thickness of lithospheric mantle beneath the Alboran Domain meant that GPE remained positive with respect
to surrounding lithosphere even after the Alboran Domain had subsided below sea level. Lithospheric cooling has now reduced the GPE contrasts, so that vertical shortening has largely ceased, and the dominant mode of deformation is strike-slip on conjugate sets of faults, plus minor amounts of reverse faulting and folding, allowing north–south shortening and east–west extension (Morel & Meghraoui 1996; Comas et al. 1999).

The overall conclusions to be reached from the comparison of these three regions are as follows.

(1) Vertical shortening in all three areas is a direct result of two related processes: crustal thickening, and the removal of most or all of the lithospheric mantle beneath the region. Both are required: without lithospheric removal GPE contrasts would not be sufficient to drive vertical shortening. The process by which the lithospheric mantle is removed continues to be debated, but the close relationship in time and space between thickening and lithospheric removal favours some form of gravitational instability that causes relatively dense mantle lithosphere to separate from the thickened crust.

(2) The rate and direction of horizontal extension, and the rate of associated vertical shortening, depend on the potential energy contrast with the surrounding lithosphere, and the magnitude and direction of the plate boundary forces. Tibet extends to the east, at the expense of low-GPE Asian continental lithosphere, and normal to the plate convergence direction. The low rate of vertical shortening reflects the fact that the rate of work being done by India in north–south shortening of Tibet is about the same as that required to cause Tibet to flow out of the way at the expense of the surrounding Asian lithosphere. The Basin and Range province extended westwards at the expense of young oceanic lithosphere to the west, even though this was initially parallel to the direction of plate convergence. The relatively high rate of vertical shortening reflects the fact that little work is required to shorten oceanic lithosphere, which is readily subducted. The Alboran Domain extended to the SW and west, at the expense of thermally mature lithosphere with either oceanic or thin continental crust: this had low GPE and was apparently easily subducted. Even though Africa and Iberia continued to converge, the excess GPE of the Alboran Domain was sufficient to prevent internal shortening; north–south convergence was accommodated by the African and Iberian margins.

(3) Extending hinterlands can continue to thin until they rupture and produce new oceanic lithosphere, as has happened in several extending back-arc in the Mediterranean region. This will only happen, however, if there is a region of sufficiently low GPE to accommodate the extension, such as lithosphere with oceanic or thin continental crust. Subduction, lithospheric delamination, or laterally propagating Rayleigh–Taylor instabilities are probably required to dispose of the lithospheric mantle beneath the region of crustal shortening.

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