

## *Bicentennial Review*

### **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  $7 \times 10^{12} \text{ N m}^{-1}$ , 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.

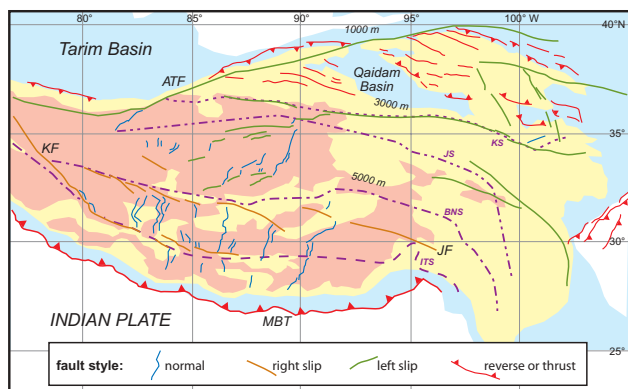
The purpose of this paper is to make the case that collisional plateaux, extending Cordilleran hinterlands, and Mediterranean back-arc basins represent a spectrum of hinterland behaviour with many geological, tectonic, and mechanical features in common. I focus on three examples: the Tibetan plateau, 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**

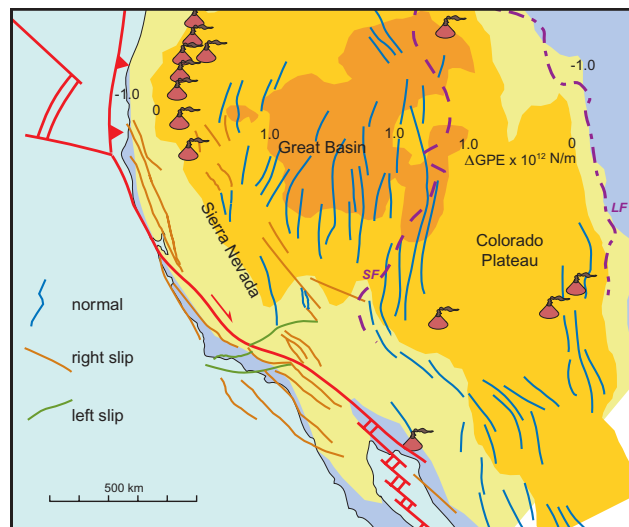
##### *Tectonic evolution*

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



**Fig. 1.** Tectonic map of the Tibetan plateau, based on data from Dewey *et al.* (1988), Molnar & Lyon-Caen (1989), Yin & Harrison (2000) and Tapponnier *et al.* (2001). Suture zones: ITS, Indus–Tsangpo; BNS, Bangong–Nujiang; JS, Jinsha; KS, Kunlun. Active faults: KF, Karakoram; JF, Jiali; ATF, Altyn Tagh.

indentation started, India–Asia convergence has continued at  $c. 50 \text{ km Ma}^{-1}$  (Dewey *et al.* 1988), and global positioning system (GPS) data indicate a present-day convergence rate of  $36\text{--}40 \text{ mm year}^{-1}$  (Paul *et al.* 2001). The crust from which the plateau formed was an amalgam of displaced continental terranes and slices of oceanic crust and sediment that were progressively assembled on the Asian margin by accretion and collision during Late Palaeozoic and Mesozoic time (Dewey *et al.* 1988). These processes are likely to have formed crust of variable thickness and elevation, and there is evidence that parts of the southern margin of Tibet (the Gangdese batholith in the Lhasa terrane) were significantly elevated prior to India–Asia collision (England & Searle 1986), perhaps forming an active convergent margin analogous to the present-day Andes. Scattered remnants of marine sedimentary rocks of Cretaceous age (Yin & Harrison 2000) suggest, however, that by the end of the Mesozoic much of

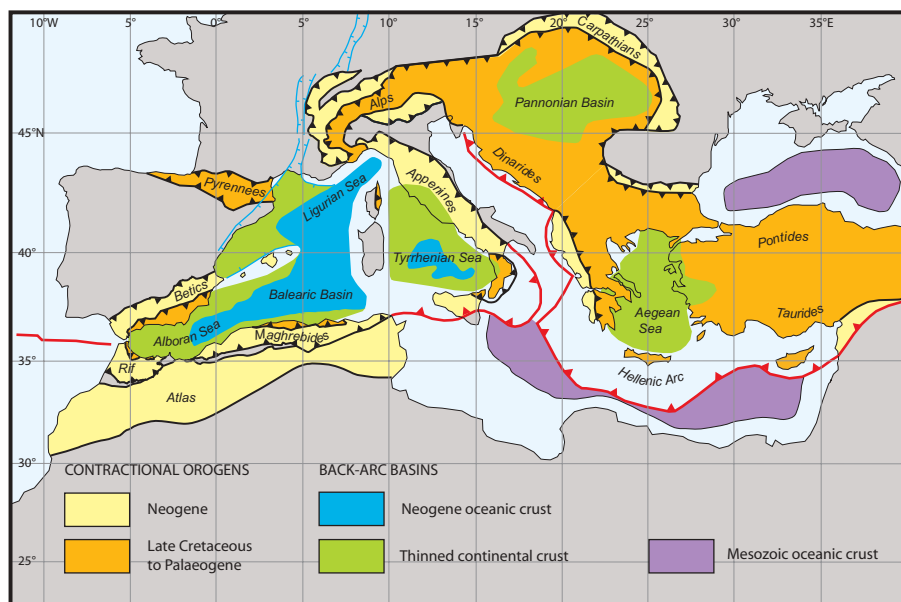


**Fig. 2.** Tectonic map of the Basin and Range Province and its surroundings in the North American Cordillera. Present plate boundaries are shown in red. GPE contours are relative to a standard mid-ocean ridge, after Coblenz & Humphreys (2006). Thrust fronts: SF, Sevier; LF, Laramide.

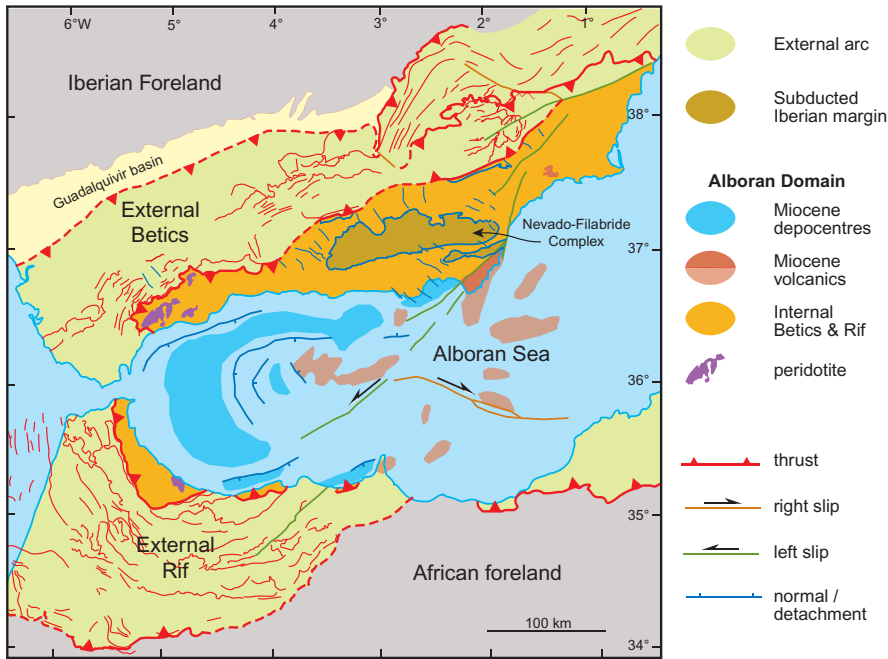
the surface of Tibet lay close to sea level. Given its likely structure and composition, this implies an overall crustal thickness in the range  $25\text{--}35 \text{ km}$ . This crust is likely to have been relatively weak: there is no reason to believe that it would have had a cold, dense, anhydrous, 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  $70 \text{ km}$  in the south to about  $65 \text{ km}$  in the north, where there is a correspondingly greater



**Fig. 3.** Tectonic map of the Mediterranean region, showing Mesozoic–Tertiary contractional orogens and Neogene extensional basins. Africa–Eurasia plate boundary is shown in red; it should be noted that the plate boundary cannot be defined everywhere, because of the widely distributed pattern of deformation.



**Fig. 4.** Tectonic map of Betic–Rif arc and Alboran Domain. The external arc comprises the Miocene thrust belts in southern Iberia and northern Africa. The Alboran Domain comprises the Internal Betic and Rif orogens, and the floor of the Alboran Sea. The Nevado–Filabride Complex is the subducted leading edge of Iberia, now exhumed from beneath the Alboran Domain.

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 seismically 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 Phanerozoic 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 transpressive 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 deformational 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  $1 \times 10^{-16} \text{ s}^{-1}$ . 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<sup>-1</sup> of northward motion is accommodated within the 130 km wide Himalayan thrust belt, and in the north, where *c.* 9 mm year<sup>-1</sup> 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<sup>-1</sup> of north–south shortening, giving an average strain rate on an azimuth of 020° of  $4.7 \times 10^{-16} \text{ s}^{-1}$  (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<sup>-1</sup> 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  $18 \pm 1 \text{ mm year}^{-1}$  (Jade *et al.* 2004), and Zhang *et al.* (2004) estimated an overall rate of east–west extension of  $22 \pm 3 \text{ mm year}^{-1}$  for the plateau as a whole, giving an average strain rate on an azimuth of 110° of  $6.8 \times 10^{-16} \text{ s}^{-1}$ . Assuming conservation of volume, these figures for horizontal strain rates imply a vertical shortening rate of  $2.1 \times 10^{-16} \text{ s}^{-1}$ . 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 & Tapponnier 1978; England & Houseman 1986, 1988, 1989; 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  $7 \times 10^{12} \text{ N m}^{-1}$ . 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. Tapponnier *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 <sup>14</sup>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



**Fig. 5.** Plots of gravitational potential energy difference ( $\Delta GPE$ ) between hinterlands and their surroundings. Each plot shows the difference in lithostatic pressure ( $\Delta pgz$ ) in MPa (horizontal axis) with depth in km (vertical axis) between a hinterland area and its immediate surroundings. The integral of  $\Delta pgz$  with depth (area under the curve, in white) is the  $\Delta GPE$ : values are shown in units of  $10^{12} \text{ N m}^{-1}$ . It should be noted that some curves show negative contributions (left of zero line) as well as positive. (a) India v. Asia at present day; (b) Basin and Range (B&R) v. Pacific ocean lithosphere at present day; (c) Basin and Range (B&R) v. mid-ocean ridge at time of ridge–trench collision at 26 Ma; (d) Alboran v. Atlantic at present day; (e) Alboran v. Atlantic at start of extension at 24 Ma; (f) Alboran v. Iberia at present day; (g) Alboran v. Iberia at 24 Ma.

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

|                       | Tibet      | Asia   | B&R 0 Ma    | Pacific 0 Ma | B&R 26 Ma   | Pacific 26 Ma | Alboran 0 Ma     | Alboran 24 Ma  | Atlantic         | Iberia         |
|-----------------------|------------|--------|-------------|--------------|-------------|---------------|------------------|----------------|------------------|----------------|
| Elevation             | 4994       | 464    | 1540        | 4019         | 5131        | 2500          | 1003             | 4668           | 5335             | 683            |
| Water depth           |            |        |             | 6000         | 60000       | 6000          | 14500            | 50000          | 6000             | 35000          |
| Crustal thickness     | 70000      | 35000  | 30000       | 51000        | 80000       | 6000          | 60000            | 70000          | 90000            | 125000         |
| Lithosphere thickness | 110000     | 125000 | 40000       | 1050         |             | 1050          | 1050             |                | 1050             |                |
| Water density         |            |        |             | 3000         |             | 3000          | 2750             |                | 3000             |                |
| Crustal density       | 2850       | 2850   | 2750        | 3271         | 3244        | 3000          | 3244             | 2750           | 3271             | 2830           |
| Lithosphere density   | 3244       | 3244   | 3244        | 3198         | 3198        | 3198          | 3198             | 3198           | 3198             | 3244           |
| Asthenosphere density | 3198       | 3198   | 3198        | 3198         | 3198        | 3198          | 3198             | 3198           | 3198             | 3198           |
|                       | Tibet-Asia |        | B&R-Pacific |              | B&R-Pacific |               | Alboran-Atlantic | Alboran-Iberia | Alboran-Atlantic | Alboran-Iberia |
|                       | 7.22E + 12 |        | 0 Ma        |              | 26 Ma       |               | 0 Ma             | 0 Ma           | 19 Ma            | 19 Ma          |
|                       |            |        | 2.22E + 12  |              | 5.46E + 12  |               | 2.54E + 12       | 0.79E + 12     | 6.97E + 12       | 5.06E + 12     |
| ΔGPE                  |            |        |             |              |             |               |                  |                |                  |                |

Elevations and thicknesses in metres; densities in  $\text{kg m}^{-3}$ ;  $\Delta\text{GPE} \times 10^{12} \text{ N m}^{-1}$ . B&R, Basin and Range. Elevations and water depths calculated assuming isostasy relative to a standard mid-ocean ridge (e.g. Pacific 26 Ma; see table). Mantle densities calculated from an assumed temperature of  $3350 \text{ kg m}^{-3}$ , corrected for temperature using coefficient of thermal expansion of  $3.4 \times 10^{-5} \text{ kg m}^{-3} \text{ }^{\circ}\text{C}^{-1}$ , assuming  $T(\text{asthenosphere}) = 1333^{\circ}\text{C}$ ,  $T(\text{Moho}) = 525^{\circ}\text{C}$ . GPE contrasts between two lithospheric columns A and B are calculated as  $\Delta\text{GPE} = \int_{-h}^L [\rho_A(z) - \rho_B(z)] dz$ , where the vertical stress  $\sigma_z = \int_{-h}^L \rho(z) g dz$ ,  $z$  is vertical distance (measured downwards),  $-h$  is the surface elevation of the higher column,  $L$  is an arbitrary depth of isostatic compensation 250 km below sea level,  $\rho(z)$  is the density at depth  $z$ , and  $g$  is gravitational acceleration.

lithosphere resulted in the initiation of normal faulting and eastward extrusion. The plateau is currently underlain by crust that is thick, hot and weak, but the present pattern of deformation reflects the fact that the topographic load of the plateau is sufficient to resist further thickening, and north-south contraction is therefore accommodated by east-west extension. The velocity gradients across the plateau revealed by GPS surveys demonstrate that Tibet does not exhibit rigid plate-like behaviour.

## 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 & Journeay 1987; Parish *et al.* 1988; Vanderhaeghe *et al.* 2003), and south into SE California, Arizona and the Mexican state of Sonora (Davis & Lister 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 (Burchfiel & 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 (Burchfiel & 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 Orocopia 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 (McCreedy *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 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 intracontinental 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

downwellings on the margins that effectively strip away the lithospheric mantle from beneath the extending region. Lithospheric downwellings have been suggested beneath the Sierra Nevada (Liu & Shen 1998; Zandt *et al.* 2004), and may represent the final stages of this process.

The timing of lithospheric removal is also uncertain. Palaeobotanical evidence for Eocene–Oligocene altitudes in the Cordillera as high as, or higher than, at present may suggest that removal had already occurred by late Eocene time (Wolfe *et al.* 1998). Sonder *et al.* (1987) suggested that convective removal of the lower lithospheric mantle immediately preceded the onset of widespread volcanism in the Great Basin during the Oligocene: uplift, extension, and partial melting of the remaining lithospheric mantle and the crust are predictable consequences of this process (Platt & England 1994). The process may have been diachronous: the patterns of converging northward and southward migration of both volcanism and extension in the mid-Tertiary, and then diverging westward and eastward migration in the Late Tertiary to Recent, may reflect complex patterns of laterally migrating downwellings, as discussed above. Livaccari & Perry (1973), however, argued on the basis of isotopic data that North American lithospheric mantle remained in place until the late Miocene. The presence in the Great Basin of pre-late Miocene alkalic basalts with positive  $\epsilon_{\text{Nd}}$  values, characteristic of a lithospheric source, indicates that the lithosphere was at least 50–70 km thick at that time (Livaccari & Perry 1973). This is in fact far less than would be expected as a result of Mesozoic contraction (>150 km), and is consistent with removal of >80 km of lithospheric mantle during the Early Tertiary, followed by Late Tertiary extensional thinning.

An alternative approach has been to explain the evolution of Basin and Range magmatism as a consequence of the foundering of a flat subducted slab of oceanic lithosphere, emplaced beneath western North America during the Laramide contractional event (Best & Christiansen 1991; Humphreys 1995). This process does not explain the present-day lack of North American lithospheric mantle, however; this has to have been removed before, during, or after the emplacement of the flat slab to explain the present-day lithospheric structure, high heat flow, and elevation of the Great Basin.

### *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  $6 \text{ mm year}^{-1}$  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.*

1989), and WNW convergence at  $4.3 \pm 0.5 \text{ mm a}^{-1}$ ) continues at the present day (McClusky *et al.* 2003).

The external Betic–Rif arc is a thin-skinned fold and thrust belt developed during Miocene time from the southern rifted margin of Iberia (García-Hernández *et al.* 1980), the northern rifted margin of Africa (Wildi 1983; Favre *et al.* 1991), and a deep-water clastic basin that lay between the two continental margins (Thurow & Kuhnt 1986). Mass-balance considerations show that the crust that originally underlay this region must have been thin, and may have been in part oceanic (Platt *et al.* 2003a), although there is no direct evidence for the former existence of oceanic crust. The thrust belt formed mainly between about 19 and 6 Ma, as a result of *c.* 250 km westward relative motion of the Alboran Domain between the converging Iberian and African plates (Platt *et al.* 2003a). This westward motion may have been accommodated by rollback of subducting oceanic lithosphere beneath the thrust belt (e.g. Lonergan & White 1997), or by delamination of the underlying lithospheric mantle (e.g. Calvert *et al.* 2000). Contraction in the thrust belt was coeval with large-scale extension in the Alboran Domain. Recent GPS data indicate that SW-directed thrusting may in fact be continuing in the southern Rif, accommodated by continuing extension in parts of the Alboran Domain (Fadil *et al.* 2006).

The Alboran Domain forms the hinterland to the external Betic–Rif arc, and comprises the internal zones of the Betic–Rif orogen and the floor of the Alboran Sea (Fig. 4). It includes the remains of a contractional orogen, formed within the western Neotethys during Eocene time by accretion of micro-continental fragments left stranded during the break-up of Pangaea (Platt *et al.* 2006). Accretion was accompanied by high-pressure–low-temperature (HP–LT) metamorphism (Azañón & Crespo-Blanc 2000). Exhumed lower crust from this orogen has been metamorphosed at pressures of up to 1600 MPa (Tubía *et al.* 1997), suggesting that the crust was up to 60 km thick. During the early Miocene this orogen underwent rapid and dramatic thinning, and much of it has now subsided below sea level, creating the present-day Alboran Sea basin.

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

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 & Visser 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).

#### *Extensional 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 Bousera peridotite massifs (Tubía *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  $9 \text{ mm year}^{-1}$  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-

guide) from rocks affected by HP–LT metamorphism at pressures up to 800 MPa in its footwall (known as Alpujárride). The Málaga–Alpujárride detachment can be traced for about 360 km in the direction of transport (ENE), without much change in structural level in either footwall or hanging wall. Several additional detachments can be defined locally both above and below this main detachment, but a significant amount of extension in the Alpujárride rocks occurred by ductile deformation that was synkinematic with decompressional metamorphism (Tubía *et al.* 1993; Balanyá *et al.* 1997; Argles *et al.* 1999). Ductile shear-sense indicators suggest predominantly ENE-directed shear, though stretching lineations vary locally from north–south through NE–SW to east–west (Tubía *et al.* 1993, 1997; Lonergan & Platt 1995; Platzman & Platt 2004). Palaeomagnetic data suggest that the original slip direction on the Málaga–Alpujárride 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–Filábride 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 (Visser *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–Filábride 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<sup>-1</sup>.

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 °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 time required for conductive heat transfer through 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 & Platt 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 mm year<sup>-1</sup>), however, suggests that the crust is unlikely to return to a normal continental thickness (*c.* 35 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 mm year<sup>-1</sup>, 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 ( $6 \times 10^{-16}$  s<sup>-1</sup>), 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  $5.5 \times 10^{12}$  N m<sup>-1</sup> 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 *c.*  $2.2 \times 10^{12}$  N m<sup>-1</sup>, but is still driving deformation at *c.*  $2.4 \times 10^{-16}$  s<sup>-1</sup>, thinning the 30 km crust by 0.2 mm year<sup>-1</sup>. 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.*  $7 \times 10^{12}$  N m<sup>-1</sup> 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.*  $5.1 \times 10^{12}$  N m<sup>-1</sup> 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-arcs 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.

## References

- ALLMENDINGER, R.W., HAUGE, T.A., HAUSER, E.C., POTTER, C.J. & OLIVER, J. 1987. Tectonic heredity and the layered lower crust in the Basin and Range Province, western United States. In: COWARD, M.P., DEWEY, J.F. & HANCOCK, P.L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 223–246.
- ALSDORF, D., BROWN, L., NELSON, K.D., MAKOVSKY, Y., KLEMPERER, S. & ZHAO, W. 1998. Crustal deformation of the Lhasa terrane, Tibet plateau from Project INDEPTH deep seismic reflection profiles. *Tectonics*, **17**, 501–519.
- ANDERSON, H. & JACKSON, J. 1987. The deep seismicity of the Tyrrhenian Sea. *Geophysical Journal*, **91**, 613–637.
- ANDERSON, J.L., BARTH, A.P. & YOUNG, E.D. 1988. Mid-crustal Cretaceous roots of Cordilleran metamorphic core complexes. *Geology*, **16**, 366–369.
- APPLEGATE, J.D.R. & HODGES, K.V. 1995. Mesozoic and Cenozoic extension recorded by metamorphic rocks in the Funeral Mountains, California. *Geological Society of America Bulletin*, **107**, 1063–1076.
- ARGLES, T.W., PLATT, J.P. & WATERS, D.J. 1999. Attenuation and excision of a crustal section during extensional exhumation: the Carratraca Massif, Betic Cordillera, Southern Spain. *Journal of the Geological Society, London*, **156**, 149–162.
- ARMJO, R., TAPPONNIER, T., MERCIER, J.L. & HAN, T.-L. 1986. Quaternary extension in Southern Tibet: field observations and tectonic implications. *Journal of Geophysical Research*, **91**, 13802–13872.
- ARMSTRONG, R.L. & WARD, P. 1991. Evolving geographic patterns of Cenozoic magmatism in the North American Cordillera: the temporal and spatial association of magmatism and metamorphic core complexes. *Journal of Geophysical Research*, **96**, 13201–13224.
- AXEN, G.J., TAYLOR, W.J. & BARTLEY, J.M. 1993. Space–time patterns and tectonic controls of Tertiary extension and magmatism in the Great Basin of the western United States. *Geological Society of America Bulletin*, **105**, 56–76.
- AZANÓN, J.-M. & CRESPO-BLANC, A. 2000. Exhumation during a continental collision inferred from the tectonometamorphic evolution of the Alpujarride complex in the central Betics (Alboran Domain, SE Spain). *Tectonics*, **19**, 549–565.
- BALANYÁ, J.C., GARCÍA-DUEÑAS, V., AZANÓN, J.M. & SÁNCHEZ-GÓMEZ, M. 1997. Alternating contractional and extensional events in the Alpujarride nappes of the Alboran Domain (Betics, Gibraltar arc). *Tectonics*, **16**, 226–238.
- BANDA, E., GALLART, J., GARCÍA-DUEÑAS, V., DAÑOBEITA, J.J. & MAKIS, J. 1993. Lateral variation of the crust in the Iberian peninsula: new evidence from the Betic Cordillera. *Tectonophysics*, **221**, 53–66.
- BEAUMONT, C., JAMIESON, R. A., NGUYEN, M. G. & MEDVEDEV, S. 2004. Crustal channel flows: 1. Numerical models with applications to the tectonics of the Himalayan–Tibetan orogen. *Journal of Geophysical Research*, **109**, B06406, doi:10.1029/2003JB002809.
- BENNETT, R. A., WERNICKE, B. P., NIEMI, N. A. & FRIEDRICH, A. M. 2003. Contemporary strain rates in the northern Basin and Range province from GPS data. *Tectonics*, **22**, 1008, doi:10.1029/2001TC001355.
- BEST, M.G. & CHRISTIANSEN, E.H. 1991. Limited extension during peak Tertiary volcanism, Great Basin of Nevada and Utah. *Journal of Geophysical Research*, **96**, 13509–13528.
- BLANCO, M.J. & SPAKMAN, W. 1993. The P-wave velocity structure of the mantle below the Iberian Peninsula: evidence for subducted lithosphere below southern Spain. *Tectonophysics*, **221**, 13–34.
- BLISNIUK, P.M., HACKER, B.R. & GLODNY, J. ET AL. 2001. Normal faulting in central Tibet since at least 13.5 Myr ago. *Nature*, **412**, 628–632.
- BOUYBAOUËNE, M., MICHARD, A. & GOFFÉ, B. 1998. High-pressure granulites on top of the Beni Bousera peridotites, Rif belt, Morocco: a record of an ancient thickened crust in the Alboran domain. *Bulletin de la Société Géologique de France*, **169**, 153–162.
- BROWN, R.L. & JOURNEAY, J.M. 1987. Tectonic denudation of the Shuswap metamorphic terrane of southeastern British Columbia. *Geology*, **15**, 142–146.
- BURCHFIELD, B.C. & DAVIS, G.A. 1972. Structural framework and evolution of the southern part of the Cordilleran orogen, western United States. *American Journal of Science*, **272**, 97–118.
- CALVERT, A., SANDVOL, E. & SEBER, D. ET AL. 2000. Geodynamic evolution of the lithosphere and upper mantle beneath the Alboran region of the western Mediterranean: constraints from travel time tomography. *Journal of Geophysical Research*, **105**, 10871–10898.
- CAMILLERI, P.A. & CHAMBERLAIN, K.R. 1997. Mesozoic tectonics and metamorphism in the Pequo Mountains and Wood Hills region, northeast Nevada: implications for the architecture and evolution of the Sevier orogen. *Geological Society of America Bulletin*, **109**, 74–94.
- CARBONELL, R., SALLARÉS, V., POUS, J., DAÑOBEITA, J.J., QUERALT, P., LEDO, J.J. & GARCÍA-DUEÑAS, V. 1998. A multidisciplinary geophysical study in the Betic chain (southern Iberian Peninsula). *Tectonophysics*, **288**, 137–152.
- CARMINATI, E., WORTEL, M.J.R., SPAKMAN, W. & SABADINI, R. 1998. The role of slab detachment processes in the opening of the western–central Mediterranean basins: some geological and geophysical evidence. *Earth and Planetary Science Letters*, **160**, 651–665.
- CARTER, T.J., KOHN, B.P., FOSTER, D.A. & GLEADOW, A.J.W. 2004. How the Harcuvar Mountains metamorphic core complex became cool: evidence from apatite (U–Th)/He thermochronometry. *Geology*, **32**, 985–988.
- CHUNG, S.-L., LIU, D. & JI, J. ET AL. 2003. Adakites from continental collision zones: melting of thickened lower crust beneath southern Tibet. *Geology*, **31**, 1021–1024.
- CLARK, M.K. & ROYDEN, L.H. 2000. Topographic ooze: building the eastern margin of Tibet by lower crustal flow. *Geology*, **28**, 703–706.
- CLARK, M.K., HOUSE, M.A., ROYDEN, L.H., WHIPPLE, K.X., BURCHFIELD, B.C.,

- ZHANG, X. & TANG, W. 2005. Late Cenozoic uplift of southeastern Tibet. *Geology*, **33**, 525–528.
- COBLENTZ, D. & HUMPHREYS, E.D. 2006. North American dynamics and western U.S. tectonics. *Journal of Geophysical Research*, in press.
- COLEMAN, M.E. & HODGES, K.V. 1995. Evidence for Tibetan plateau uplift before 14 Myr ago from a new minimum estimate for east–west extension. *Nature*, **474**, 49–52.
- COMAS, M.C., PLATT, J.P., SOTO, J.I. & WATTS, A.B. 1999. The origin and tectonic history of the Alboran Basin: insights from Leg 161 results. In: ZAHN, R., COMAS, M.C. & KLAUS, A. (eds) *Proceedings of the Ocean Drilling Project, Scientific Results, 161*. Ocean Drilling Program, College Station, TX, 555–579.
- CRESPO-BLANC, A. 1995. Interference pattern of extensional fault systems: a case study of the Miocene rifting of the Alboran basement (north of Sierra Nevada, Betic Chain). *Journal of Structural Geology*, **17**, 1559–1570.
- CURRIE, B.S., ROWLEY, D.B. & TABOR, N.J. 2005. Middle Miocene paleoaltimetry of southern Tibet: implications for the role of mantle thickening and delamination in the Himalayan orogen. *Geology*, **33**, 181–184.
- DAVIS, G.A. & LISTER, G.S. 1988. Detachment faulting in continental extension; perspectives from the southwestern U.S. Cordillera. In: CLARK, S.P. JR, BURCHFIELD, B.C. & SUPPE, J. (eds) *Processes in Continental Lithospheric Deformation*. Geological Society of America, Special Papers, **218**, 133–159.
- DEWEY, J.F., SHACKLETON, R.M., CHENGFA, C. & YIYIN, S. 1988. The tectonic evolution of the Tibetan plateau. *Philosophical Transactions of the Royal Society of London*, **A327**, 379–413.
- DEWEY, J.F., HELMAN, M.L., TURCO, E., HUTTON, D.W.H. & KNOTT, S.D. 1989. Kinematics of the Western Mediterranean. In: COWARD, M.P., DIETRICH, D. & PARK, R.G. (eds) *Alpine Tectonics*. Geological Society, London, Special Publications, **45**, 265–284.
- DOKKA, R.K., MAHAFFIE, M.J. & SNOKE, A.W. 1986. Thermochronologic evidence of major tectonic denudation associated with detachment faulting, northern Ruby Mountains, East Humboldt Range, Nevada. *Tectonics*, **5**, 995–1006.
- DUGGEN, S., HOERNLE, K., VAN DEN BOGAARD, P. & HARRIS, C. 2004. Magmatic evolution of the Alboran region: the role of subduction in forming the western Mediterranean and causing the Messinian Salinity Crisis. *Earth and Planetary Science Letters*, **218**, 91–108.
- ENGLAND, P.C. & HOUSEMAN, G.A. 1986. Finite strain calculations of continental deformation. 2. Comparison with the India–Asia collision zone. *Journal of Geophysical Research*, **91**, 3664–3676.
- ENGLAND, P.C. & HOUSEMAN, G.A. 1988. The mechanics of the Tibetan plateau. *Philosophical Transactions of the Royal Society of London, Series A*, **326**, 301–320.
- ENGLAND, P.C. & HOUSEMAN, G.A. 1989. Extension during continental convergence, with application to the Tibetan plateau. *Journal of Geophysical Research*, **94**, 17561–17579.
- ENGLAND, P.C. & SEARLE, M.P. 1986. The Cretaceous–Tertiary deformation of the Lhasa block and its implications for the crustal thickening of Tibet. *Tectonics*, **5**, 1–14.
- FACENNA, C., BECKER, T.W., LUCENTE, P., JOLIVET, L. & ROSSETTI, F. 2001. History of subduction and back-arc extension in the Central Mediterranean. *Geophysical Journal International*, **145**, 809–820.
- FADIL, A., VERNANT, P., MCCLUSKY, S., et al. 2006. Active tectonics of the western Mediterranean: geodetic evidence for rollback of a delaminated subcontinental lithospheric slab beneath the Rif Mountains, Morocco. *Geology*, **34**, 529–532, doi:10.1130/G22291.1.
- FAULDS, J.E., HENRY, C.D. & HINZ, N.H. 2005. Kinematics of the northern Walker Lane: an incipient transform fault along the Pacific–North American plate boundary. *Geology*, **33**, 505–508.
- FAVRE, P., STAMFILLI, G. & WILDI, W. 1991. Jurassic sedimentary record and tectonic evolution of the northwestern corner of Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **87**, 53–73.
- FEINBERG, H., SADDIQI, O. & MICHARD, A. 1996. New constraints on the bending of the Gibraltar arc from palaeomagnetism of the Ronda peridotites (Betic Cordillera, Spain). In: MORRIS, A. & TARLING, D.H. (eds) *Palaeomagnetism and Tectonics of the Mediterranean Region*. Geological Society, London, Special Publications, **105**, 43–52.
- FLESCH, L.M., HAINES, A.J. & HOLT, W.E. 2001. Dynamics of the India–Eurasia collision zone. *Journal of Geophysical Research*, **106**, 16435–16460.
- FOSTER, D.A., HARRISON, T.M., MILLER, C.F. & HOWARD, K.A. 1990. The  $^{40}\text{Ar}/^{39}\text{Ar}$  thermochronology of the eastern Mojave desert, California, and adjacent western Arizona, with implications for the evolution of metamorphic core complexes. *Journal of Geophysical Research*, **95**, 20005–20024.
- GALVÉ, A., JIANG, M. & HIRN, A. ET AL. 2006. Explosion seismic P and S velocity and attenuation constraints on the lower crust of the North–Central Tibetan Plateau, and comparison with the Tethyan Himalayas: implications on composition, mineralogy, temperature, and tectonic evolution. *Tectonophysics*, **412**, 141–157.
- GANS, P.B. 1989. An open-system, two layer crustal stretching model for the eastern Great Basin. *Tectonics*, **6**, 1–12.
- GANS, P.B., MAHOOD, G.A. & SCHERMER, E. 1989. *Syn-extensional magmatism in the Basin and Range province; a case study from the eastern Great Basin*. Geological Society of America, Special Papers, **233**.
- GARCÍA-HERNÁNDEZ, M., LOPEZ-GARRIDO, A.C., RIVAS, P., SANZ DE GALDEANO, C. & VERA, J.A. 1980. Mesozoic palaeogeographic evolution of the external zones of the Betic Cordillera. *Geologie en Mijnbouw*, **59**, 155–168.
- GEMMER, L. & HOUSEMAN, G. A. 2005. The effects of gravitational instability on the tectonic evolution of continental orogens. *Abstracts and Programs, American Geophysical Union, Spring Meeting 2005*, abstract T51A-03.
- GHOSH, A., HOLT, W.E., FLESCH, L.M. & HAINES, J.H. 2006. Gravitational potential energy of the Tibetan Plateau and the forces driving the Indian plate. *Geology*, **34**, 321–324.
- GRIMSON, N.L. & CHEN, W.P. 1986. The Azores–Gibraltar plate boundary: focal mechanisms, depths of earthquakes, and their tectonic implications. *Journal of Geophysical Research*, **91**, 2029–2047.
- GRIOT, D.-A., MONTAGNER, J.-P. & TAPPONNIER, P. 1998. Phase velocity structure from Rayleigh and Love waves in Tibet and its neighboring regions. *Journal of Geophysical Research*, **103**, 21215–21232.
- GUTSCHER, M.-A., MALOD, J., REHAULT, J.-P., CONTRUCCI, I., KLINGHOEFER, F., MENDES-VICTOR, L. & SPAKMAN, W. 2002. Evidence for active subduction beneath Gibraltar. *Geology*, **30**, 1071–1075.
- HACKER, B.R., GNOS, E. & RATSCHBACHER, L. ET AL. 2000. Hot and dry deep crustal xenoliths from Tibet. *Science*, **287**, 2463–2466.
- HAINES, S. S., KLEMPERER, S. L., BROWN, L., et al. 2003. INDEPTH III seismic data: from surface observations to deep crustal processes in Tibet. *Tectonics*, **22**, 1001, doi:10.1029/2001TC001305.
- HAMILTON, W. 1987. Crustal extension in the Basin and Range province, southwestern United States. In: COWARD, M.P., DEWEY, J.F. & HANCOCK, P.L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 155–176.
- HAUCK, M.L., NELSON, K.D., BROWN, L.D., ZHAO, W. & ROSS, A.R. 1998. Crustal structure of the Himalayan orogen at c. 90° east longitude from Project INDEPTH deep reflection profiles. *Tectonics*, **17**, 481–500.
- HAUSER, E., POTTER, C. & HAUGE, T.A. ET AL. 1987. Crustal structure of eastern Nevada from COCORP deep seismic reflection data. *Geological Society of America Bulletin*, **99**, 833–844.
- HAWKESWORTH, C., TURNER, S., GALLAGHER, K., HUNTER, A., BRADSHAW, T. & ROGERS, N. 1995. Calc-alkaline magmatism, lithospheric thinning and extension in the Basin and Range. *Journal of Geophysical Research*, **100**, 10271–10286.
- HENRY, C. & ARANDA-GÓMEZ, J.J. 1992. The real southern Basin and Range: mid to late Cenozoic extension in Mexico. *Geology*, **20**, 701–705.
- HODGES, K.V. & WALKER, J.D. 1990. Petrologic constraints on the unroofing history of the Funeral Mountain metamorphic core complex, California. *Journal of Geophysical Research*, **95**, 8437–8446.
- HODGES, K.V. & WALKER, J.D. 1992. Thermal evolution of a portion of the Sevier hinterland: the northern Ruby Mountains–East Humboldt Range and Wood Hills, northeastern Nevada. *Tectonics*, **11**, 154–164.
- HOISCH, T.D., HEIZLER, M.T. & ZARTMAN, R.E. 1997. Timing of detachment faulting in the Bullfrog Hills and Bare Mountain area, southwest Nevada: inferences from  $^{40}\text{Ar}/^{39}\text{Ar}$ , K–Ar, U–Pb, and fission track thermochronology. *Journal of Geophysical Research*, **102**, 2815–2833.
- HOU, Z.-Q., GAO, Y.-F., QU, X.-M., RUI, Z.-Y. & MO, X.-X. 2004. Origin of adakitic intrusives generated during mid-Miocene east–west extension in the southern Tibet. *Earth and Planetary Science Letters*, **220**, 139–155.
- HOUSEMAN, G.A. & MOLNAR, P. 2001. Mechanisms of lithospheric rejuvenation associated with continental orogeny. In: MILLER, J.A., HOLDSWORTH, R.E., BUICK, I.S. & HAND, M. (eds) *Continental Reactivation and Reworking*. Geological Society, London, Special Publications, **184**, 13–38.
- HUMPHREYS, E.D. 1995. Post-Laramide removal of the Farallon slab, western United States. *Geology*, **23**, 987–990.
- JABALOY, A., GALINDO-ZALDÍVAR, J. & GONZÁLEZ-LODEIRO, F. 1993. The Alpujarride–Nevado–Filabride extensional shear zone, Betic Cordillera, SE Spain. *Journal of Structural Geology*, **15**, 555–570.
- JACOBSON, C.E., OYARZABAL, F.R. & HAXEL, G.B. 1996. Subduction and exhumation of the Pelona–Orocopia–Rand schists, southern California. *Geology*, **24**, 481–576.
- JACOBSON, C.E., BARTH, A.P. & GROVE, M. 2000. Late Cretaceous protolith age and provenance of the Pelona and Orocopia Schists, southern California: implications for evolution of the Cordilleran margin. *Geology*, **28**, 219–223.
- JADE, S., BHATT, B.C., YANG, Z., et al. 2004. GPS measurements from the Ladakh Himalaya, India: preliminary tests of plate-like or continuous deformation in Tibet. *Geological Society of America Bulletin*, **116**, 1385–1391; doi:10.1130/B25357.1.
- JIMÉNEZ-MUNT, I. & PLATT, J.P. 2006. Influence of mantle dynamics on the topographic evolution of the Tibetan Plateau: Results from numerical modelling. *Tectonics*, **25**, TC6002; doi:10.1029/2006TC001963/.

- JIMÉNEZ-MUNT, I., FERNÁNDEZ, M., VERGÉS, J. & PLATT, J.P. 2006. Lithosphere structure underneath the Tibetan Plateau inferred from elevation, gravity and geoid anomalies. *Journal of Asian Earth Sciences*, **26**, 121–171.
- JOHNSON, C.J. & HARBURY, N. 1997. The role of extension in the Miocene denudation of the Nevado–Filabride Complex, Betic Cordillera (SE Spain). *Tectonics*, **16**, 189–204.
- JOLIVET, L. & FACENNA, C. 2000. Mediterranean extension and the Africa–Eurasia collision. *Tectonics*, **19**, 1095–1106.
- JONES, C.H., WERNICKE, B.P., FARMER, G.L., WALKER, J.D., COLEMAN, D.S., MCKENNA, L.W. & PERRY, F.V. 1992. Variations across and along a major continental rift: an interdisciplinary study of the Basin and Range Province, western USA. *Tectonophysics*, **213**, 57–96.
- JONES, C.H., SONDER, L.J. & UNRUH, J.R. 1998. Lithospheric gravitational potential energy and past orogenesis: implications for conditions of initial Basin and Range and Laramide deformation. *Geology*, **26**, 639–642.
- LEE, J. 1995. Rapid uplift and rotation of mylonitic rocks from beneath a detachment fault: insights from potassium feldspar  $^{40}\text{Ar}/^{39}\text{Ar}$  thermochronology, northern Snake Range, Nevada. *Tectonics*, **14**, 54–77.
- LEWIS, C.J., WERNICKE, B.P., SILVERSTONE, J. & BARTLEY, J. 1999. Deep burial of the footwall of the northern Snake Range décollement, Nevada. *Geological Society of America Bulletin*, **111**, 39–51.
- LIVACCARI, R.F. & PERRY, F.V. 1993. Isotopic evidence for the preservation of Cordilleran lithospheric mantle during the Sevier–Laramide orogeny, western United States. *Geology*, **21**, 719–723.
- LIU, M. & SHEN, Y. 1998. Sierra Nevada uplift: a ductile link to mantle upwelling under the Basin and Range province. *Geology*, **26**, 299–302.
- LONERGAN, L. & PLATT, J.P. 1995. The Malaguide–Alpujarride boundary: a major extensional contact in the Internal Zone of the eastern Betic Cordillera, SE Spain. *Journal of Structural Geology*, **17**, 1655–1671.
- LONERGAN, L. & WHITE, N. 1997. Origin of the Betic–Rif mountain belt. *Tectonics*, **16**, 504–522.
- LOWRY, A.R., RIBE, N.M. & SMITH, R.B. 2000. Dynamic elevation of the Cordillera, western United States. *Journal of Geophysical Research*, **105**, 23371–23390.
- MARTÍNEZ-MARTÍNEZ, J. M., SOTO, J. I. & BALANYÁ, J. C. 2002. Orthogonal folding of extensional detachments: structure and origin of the Sierra Nevada elongated dome (Betics, SE Spain). *Tectonics*, **21**, 1012, doi:10.1029/2001TC001283.
- MAUFFRET, A., FRIZON DE LAMOTTE, D., LALLEMANT, S., GORINI, C. & MAILLARD, A. 2004. East–west opening of the Algerian basin (Western Mediterranean). *Terra Nova*, **16**, 257–264, doi:10.1111/j.1365-3121.2004.00559.x.
- MCCARTHY, J. & PARSONS, T. 1994. Insight into the kinematic Cenozoic evolution of the Basin and Range–Colorado transition from coincident seismic refraction and reflection data. *Geological Society of America Bulletin*, **106**, 747–759.
- MCCLUCKY, S., REILINGER, R., MAHMOUD, S., BEN SARI, D. & TEALEB, A. 2003. GPS constraints on Africa (Nubia) and Arabia plate motions. *Geophysical Journal International*, **155**, 126–138.
- MCCREADY, T., SNOKE, A.W., WRIGHT, J.E. & HOWARD, K.E. 1997. Mid-crustal flow during Tertiary extension in the Ruby Mountains core complex, Nevada. *Geological Society of America Bulletin*, **109**, 1576–1594.
- MCGREW, A.J., PETERS, M.T. & WRIGHT, J.E. 2000. Thermobarometric constraints on the tectonothermal evolution of the East Humboldt Range metamorphic core complex, Nevada. *Geological Society of America Bulletin*, **112**, 45–60.
- MCMANARA, D., WALTER, W., OWENS, T. & AMMON, C. 1997. Upper mantle velocity structure beneath the Tibetan Plateau from Pn travel time tomography. *Journal of Geophysical Research*, **102**, 493–505.
- MECHIE, J., SOBOLEV, S.V. & RATSCHBACHER, L. ET AL. 2004. Precise temperature estimation in the Tibetan crust from seismic detection of the  $\alpha$ – $\beta$  quartz transition. *Geology*, **32**, 601–604.
- MÉRIAUX, A.-S., RYERSON, F.J., TAPPONNIER, P., ET AL. 2004. Rapid slip along the central Altyn Tagh Fault: morphochronologic evidence from Cherchen He and Sulamu Tagh. *Journal of Geophysical Research*, **109**, B06401, doi:10.1029/2003JB002558.
- MEULENKAMP, J.E., WORTEL, M.J.R., VAN WAMEL, W.A., SPAKMAN, W. & HOOGENDOORN STRATING, E. 1988. On the Hellenic subduction zone and the geodynamic evolution of Crete since the late Middle Miocene. *Tectonophysics*, **146**, 203–215.
- MILLER, E.L. & GANS, P.B. 1989. Cretaceous crustal structure and metamorphism in the hinterland of the Sevier thrust belt, western U.S. Cordillera. *Geology*, **17**, 59–62.
- MILLER, E.L., GANS, P.B. & GARING, J. 1983. The Snake Range décollement: an exhumed mid-Tertiary ductile–brittle transition. *Tectonics*, **2**, 239–263.
- MOLNAR, P. & HOUSEMAN, G.A. 2004. The effects of buoyant crust on the gravitational instability of thickened mantle lithosphere at zones of intracontinental convergence. *Geophysical Journal International*, **158**, 1134–1150.
- MOLNAR, P. & LYON-CAEN, H. 1989. Fault plane solutions of earthquakes and active tectonics of the Tibetan Plateau and its margins. *Geophysical Journal International*, **99**, 123–153.
- MOLNAR, P. & TAPPONNIER, P. 1978. Active tectonics of Tibet. *Journal of Geophysical Research*, **83**, 5361–5375.
- MOLNAR, P., ENGLAND, P.C. & MARTINOD, J. 1993. Mantle dynamics, uplift of the Tibetan plateau, and the Indian monsoon. *Reviews of Geophysics*, **31**, 357–396.
- MOREL, J.L. & MEGHRAOUI, M. 1996. Goringe–Alboran–Tell tectonic zone: a transpression system along the Africa–Eurasia plate boundary. *Geology*, **24**, 755–758.
- MUELLER, K.J. & SNOKE, A.W. 1993. Progressive overprinting of normal fault systems and their role in Tertiary exhumation of the East Humboldt–Wood Hills metamorphic complex, northeast Nevada. *Tectonics*, **12**, 361–371.
- NEGRO, F., BEYSSAC, O., GOFFÉ, B., SADDIQI, O. & BOUYBAOUËNE, M. L. 2006. Thermal structure of the Alboran Domain in the Rif (northern Morocco) and the Western Betics (southern Spain). Constraints from Raman spectroscopy of carbonaceous material. *Journal of Metamorphic Geology*, **24**, 309–327, doi:10.1111/j.1525-1314.2006.00639.x.
- OLDOW, J.S., BALLY, A.W., AVÉ LALLEMANT, H.G. & LEEMAN, W.P. 1989. Phanerozoic evolution of the North American Cordillera; United States and Canada. In: BALLY, A.W. & PALMER, A.W. (eds) *The Geology of North America—an Overview*. Geological Society of America, Boulder, CO, 139–232.
- PARISH, R.R., CARR, S.D. & PARKINSON, D.L. 1988. Eocene extensional tectonics and geochronology of the southern Omineca Belt, British Columbia and Washington. *Tectonics*, **7**, 181–212.
- PAUL, J., BÜRGMANN, R. & GAUR, V.K. ET AL. 2001. The motion and active deformation of India. *Geophysical Research Letters*, **28**, 647–650.
- PLATT, J.P. 1986. Dynamics of orogenic wedges and the uplift of high-pressure metamorphic rocks. *Geological Society of America Bulletin*, **97**, 1037–1053.
- PLATT, J.P. & ENGLAND, P.C. 1994. Convective removal of lithosphere beneath mountain belts: thermal and mechanical consequences. *American Journal of Science*, **294**, 307–336.
- PLATT, J.P. & HOUSEMAN, G. 2003. Evidence for active subduction beneath Gibraltar. *Geology*, doi:10.1130/0091-7613.
- PLATT, J.P. & VISSERS, R.L.M. 1989. Extensional collapse of thickened continental lithosphere: a working hypothesis for the Alboran Sea and the Gibraltar arc. *Geology*, **17**, 540–543.
- PLATT, J.P., SOTO, J.-I., WHITEHOUSE, M.J., HURFORD, A.J. & KELLEY, S.P. 1998. Thermal evolution, rate of exhumation, and tectonic significance of metamorphic rocks from the floor of the Alboran extensional basin, western Mediterranean. *Tectonics*, **17**, 671–689.
- PLATT, J.P., ALLERTON, S., KIRKER, A., MANDEVILLE, C., MAYFIELD, A., PLATZMAN, E.S. & RIMI, A. 2003a. The ultimate arc: differential displacement, oroclinal bending, and vertical axis rotation in the External Betic–Rif arc. *Tectonics*, **22**, doi:10.1029/2001TC001321.
- PLATT, J.P., WHITEHOUSE, M.J., KELLEY, S.P., CARTER, A. & HOLICK, L. 2003b. Simultaneous extension and exhumation across a collapsing back-arc basin: implications for the causes of late-orogenic extension. *Geology*, **31**, 251–254.
- PLATT, J.P., ARGLES, T.W., CARTER, A., KELLEY, S.P., WHITEHOUSE, M.J. & LONERGAN, L. 2003c. Exhumation of the Ronda peridotite and its crustal envelope: constraints from thermal modelling a  $PT$ –time array. *Journal of the Geological Society, London*, **160**, 655–677.
- PLATT, J.P., ANCZKIEWICZ, R., SOTO, J.I., KELLEY, S.P. & THIRLWALL, M. 2006. Early Miocene continental subduction and rapid exhumation in the Western Mediterranean. *Geology*, **34**, 981–984.
- PLATZMAN, E. & PLATT, J.P. 2004. Kinematics of a twisted core complex: oblique axis rotation in an extended terrane (Betic Cordillera, southern Spain). *Tectonics*, **23**, doi:10.1029/2003TC001549.
- PLATZMAN, E., PLATT, J.P. & KELLEY, S.P. 2000. Large clockwise rotations in an extensional allochthon, Alboran Domain (southern Spain). *Journal of the Geological Society, London*, **157**, 1187–1197.
- POWELL, C.M. & CONAGHAN, P. 1973. Plate tectonics and the Himalayas. *Earth and Planetary Science Letters*, **20**, 1–12.
- RAHL, J.M., MCGREW, A.J. & FOLAND, K.A. 2002. Transition from contraction to extension in the northeastern Basin and Range: new evidence from the Copper Mountains, Nevada. *Journal of Geology*, **110**, 179–194.
- RANALLI, G., BROWN, R.L. & BOSDACHIN, R. 1989. A geodynamic model for extension in the Shuswap core complex, southeastern Canadian Cordillera. *Canadian Journal of Earth Sciences*, **26**, 1647–1653.
- ROEST, W.R. & SRIVASTAVA, S.P. 1992. Kinematics of the plate boundaries between Eurasia, Iberia and Africa in the North Atlantic from the Late Cretaceous to the present. *Geology*, **19**, 613–616.
- ROWLEY, D.B. 1996. Age of initiation of collision between India and Asia: a review of stratigraphic data. *Earth and Planetary Science Letters*, **145**, 1–13, doi:10.1016/S0012-821X(96)00201-4.
- ROYDEN, L. & BURCHFIELD, B.C. 1987. Thin-skinned extension within the convergent Himalayan region: gravitational collapse of a Miocene topo-

- graphic front. In: COWARD, M.P., DEWEY, J.F. & HANCOCK, P.L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 611–619.
- SEARLE, M.P., SIMPSON, R.L., LAW, R.D., PARRISH, R.R. & WATERS, D.J. 2003. The structural geometry, metamorphic and magmatic evolution of the Everest massif, High Himalaya of Nepal–South Tibet. *Journal of the Geological Society, London*, **160**, 345–366.
- SHI, D., ZHAO, W. & BROWN, L. ET AL. 2004. Detection of southward intracontinental subduction of Tibetan lithosphere along the Bangong–Nujiang suture by P-to-S converted waves. *Geology*, **32**, 209–212.
- SONDER, L.J. & JONES, C.H. 1999. Western United States extension: how the west was widened. *Annual Review of Earth and Planetary Sciences*, **27**, 417–462.
- SONDER, L.J., ENGLAND, P.C., WERNICKE, B.P. & CHRISTIANSEN, R.L. 1987. A physical model for Cenozoic extension of western North America. In: COWARD, M.P., DEWEY, J.F. & HANCOCK, P.L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 187–201.
- SOTO, J.I. & PLATT, J.P. 1999. Petrological and structural evolution of high-grade metamorphic rocks from the floor of the Alboran Sea basin, Western Mediterranean. *Journal of Petrology*, **40**, 21–60.
- SPENCER, J.E. & REYNOLDS, S.J. 1991. Tectonics of mid-Tertiary extension along a transect through west central Arizona. *Tectonics*, **10**, 1204–1221.
- SPICER, R.A., HARRIS, N.W.B. & WIDDOWSON, M. ET AL. 2003. Constant elevation of Tibet over the past 15 million years. *Nature*, 622–625.
- TAPPONNIER, P., XU, Z., ROGER, F., MEYER, B., ARNAUD, N., WITTLINGER, G. & YANG, J. 2001. Oblique stepwise rise and growth of the Tibet Plateau. *Science*, **294**, 1671–1677, doi:10.1126/science.105978.
- TAYLOR, M., YIN, A., RYERSON, F. J., KAPP, P. & DING, L. 2004. Conjugate strike-slip faulting along the Bangong–Nujiang suture zone accommodates coeval east–west extension and north–south shortening in the interior of the Tibetan plateau. *Tectonics*, **22**, 1044, doi:10.1029/2002TC001361.
- TAYLOR, W.J., BARTLEY, J.M., MARTIN, M.W., GEISSMAN, J.W., WALKER, J.D., ARMSTRONG, P.A. & FRYXELL, J.E. 2000. Relations between hinterland and foreland shortening: Sevier orogeny, central North American Cordillera. *Tectonics*, **19**, 1124–1143.
- THATCHER, W., FOULGER, G.R., JULIAN, B.R., SVARC, J., QUILTY, E. & BAWDEN, G.W. 1999. Present-day deformation across the Basin and Range Province, Western United States. *Science*, **283**, 1714–1718.
- THUROW, J. & KUHN, W. 1986. Mid Cretaceous of the Gibraltar arc area. In: SUMMERHAYES, C.P. & SHACKLETON, N.J. (eds) *North Atlantic Palaeoceanography*. Geological Society, London, Special Publications, **21**, 423–445.
- TILMANN, F., NI, J. & INDEPTH III SEISMIC TEAM 2003. Seismic imaging of the downwelling Indian lithosphere beneath central Tibet. *Science*, **300**, 1424–1427, doi:10.1126/science.1082777.
- TORNÉ, M., FERNÁNDEZ, M., COMAS, M.C. & SOTO, J.I. 2000. Lithospheric structure beneath the Alboran Basin: results from 3D gravity modeling and tectonic relevance. *Journal of Geophysical Research*, **105**, 3209–3228.
- TUBÍA, J.M., NAVARRO-VILA, F. & CUEVAS, J. 1993. The Maláguide–Los Reales Nappe: an example of crustal thinning related to the emplacement of the Ronda peridotites (Betic Cordillera). *Physics of the Earth and Planetary Interiors*, **78**, 343–354.
- TUBÍA, J.M., CUEVAS, J. & GIL-IBARGUCHI, J.-I. 1997. Sequential development of the metamorphic aureole beneath the Ronda peridotites and its bearing on the tectonic evolution of the Betic Cordillera. *Tectonophysics*, **279**, 227–252.
- TURNER, S., ARNAUD, N. & LIU, J. ET AL. 1996. Post-collision, shoshonitic volcanism in the Tibetan plateau: implications for convective thinning of the lithosphere and the source of ocean island basalts. *Journal of Petrology*, **37**, 45–71.
- TURNER, S.P., PLATT, J.P., GEORGE, R.M.M., KELLEY, S.P., PEARSON, D.G. & NOWELL, G.M. 1999. Magmatism associated with orogenic collapse of the Betic–Alboran Domain, S. Spain. *Journal of Petrology*, **40**, 1011–1036.
- UNRUH, J., HUMPHREY, J. & BARRON, A. 2003. Transtensional model for the Sierra Nevada frontal fault system, eastern California. *Geology*, **31**, 327–330.
- VANDERHAEGHE, O., TEYSSIER, C., MCDUGAL, I. & DUNLAP, W.J. 2003. Cooling and exhumation of the Shuswap Metamorphic Core Complex constrained by  $^{40}\text{Ar}/^{39}\text{Ar}$  thermochronology. *Geological Society of America Bulletin*, **115**, 200–216.
- VISSERS, R.L.M., PLATT, J.P. & VAN DER WAL, D. 1995. Late orogenic extension of the Betic Cordillera and the Alboran Domain: a lithospheric view. *Tectonics*, **14**, 786–803.
- WANG, Y., DENG, T. & BIASATTI, D. 2006. Ancient diets indicate significant uplift of southern Tibet after ca. 7 Ma. *Geology*, **34**, 309–312.
- WATTS, A.B., PLATT, J.P. & BUHL, P. 1993. Tectonic evolution of the Alboran Sea basin. *Basin Research*, **5**, 153–177.
- WELLS, M.L., SNEE, L.W. & BLYTHE, A. 2000. Dating of major normal fault systems using thermochronology: an example from the Raft River detachment, Basin and Range, western United States. *Journal of Geophysical Research*, **105**, 16303–16327.
- WELLS, M.L., BEYENE, M.A., SPELL, T.L., KULA, J.L., MILLER, D.M. & ZANETTI, K.A. 2005. The Pinto shear zone; a Laramide synconvergent extensional shear zone in the Mojave Desert region of the southwestern United States. *Journal of Structural Geology*, **27**, 1697–1720.
- WERNICKE, B.P., CHRISTIANSEN, R.L., ENGLAND, P.C. & SONDER, L.J. 1987. Tectonomagmatic evolution of Cenozoic extension in the North American Cordillera. In: COWARD, M.P., DEWEY, J.F. & HANCOCK, P.L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 203–221.
- WILDI, W. 1983. La chaîne tello-rifaine (Algérie, Maroc, Tunisie): structure, stratigraphie et évolution du Trias au Miocène. *Revue de Géologie Dynamique et Géographie Physique*, **24**, 201–297.
- WILLIAMS, H., TURNER, S., KELLEY, S.P. & HARRIS, N. 2001. Age and composition of dikes in southern Tibet: new constraints on the timing of east–west extension and its relationship to postcollisional volcanism. *Geology*, **29**, 399–342.
- WOLFE, J.A., FOREST, C.E. & MOLNAR, P. 1998. Palaeobotanical evidence of Eocene and Oligocene paleoaltitudes in midlatitude western North America. *Geological Society of America Bulletin*, **110**, 664–678.
- WONG, M.S. & GANS, P.B. 2003. Tectonic implications of early Miocene extensional unroofing of the Sierra Mazatán metamorphic core complex, Sonora, Mexico. *Geology*, **31**, 953–956.
- WRIGHT, T.J., PARSONS, B., ENGLAND, P.C. & FIELDING, E.J. 2004. InSAR observations of low slip rates on the major faults of western Tibet. *Science*, **305**, 236–239.
- WUST, S.L. 1986. Regional correlation of extension directions in Cordilleran metamorphic core complexes. *Geology*, **14**, 828–830.
- YIN, A. & HARRISON, M.T. 2000. Geologic evolution of the Himalayan–Tibetan orogen. *Annual Review of Earth and Planetary Sciences*, **28**, 211–280.
- YIN, A., KAPP, P.A. & MURPHY, M.A. ET AL. 1999. Significant late Neogene east–west extension in northern Tibet. *Geology*, **27**, 787–791.
- ZANDT, G., GILBERT, H., OWENS, T.J., DUCEA, M., SALEEBY, J. & JONES, C.H. 2004. Active foundering of a continental arc root beneath the southern Sierra Nevada in California. *Nature*, **431**, 41–46.
- ZHANG, P.-Z., BÜRGMANN, R. & MOLNAR, P. ET AL. 2004. Continuous deformation of the Tibetan Plateau from global positioning system data. *Geology*, **32**, 809–812.
- ZHAO, W. & MORGAN, W.J. 1985. Uplift of Tibetan plateau. *Tectonics*, **4**, 359–369.

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