The origin of mountains – implications for the behaviour of Earth's lithosphere

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Mountains are an expression of buoyancy in the outermost part of the Earth, caused by density contrasts in the lithosphere and underlying asthenosphere, as described by the principle of isostasy. The biggest density contrast is between the lower density crust and more dense mantle, and so changes in crustal thickness are the main reason for the creation of the planet's highest and most extensive mountain ranges, in the Andes of South America, and the great ranges of Central Asia, in Tibet and the Himalaya. Crustal thickening is primarily a result of crustal shortening, driven by horizontal forces in the lithosphere. The response of the lithosphere to these forces depends on its thickness and rheology, and when tectonic plates converge, mountain ranges will form where the lithosphere is weak and thin, pushed up by the same forces that drive the plates themselves, though limited by the maximum force that can be transmitted across the plate interface. The smaller density contrast between the lithospheric mantle and underlying less dense asthenosphere will also play a subsidiary role in the elevation of mountain belts, depending on changes in lithospheric thickness. The nature of deformation during mountain building is strongly controlled by the interaction of lithosphere of different strengths.

Keywords: Lithosphere, mountains, plate tectonics, tectonics.

Introduction

HIGH mountains rising up several kilometres form a relatively small fraction of the land area on the planet, yet for those who have travelled amongst the towering snow-capped peaks of the European Alps, or the Rocky Mountains and Andes in the Americas, or the vast ranges of Asia, they are features that can hardly be missed (Figure 1), and discovering their origins has proved to be one of the keys to understanding the way the outer part of the Earth – the strong outer shell called the lithosphere – works.

The fundamental scientific problem is one of dynamics: the nature of the forces that both create and support high mountains, and the response of the lithosphere to these forces. Here, we are not talking about individual peaks, even ones as high as Everest, but rather the general elevation of the Earth over horizontal distances of tens to hundreds of kilometres – this would be the smoother bulge in the Earth's surface one would get if one could somehow bulldozer all the peaks into the surrounding valleys. From this point of view, it is the high plateau of Tibet in Central Asia, or the Altiplano in the Central Andes – extending for 100s to 1000s km at 4 to 5 km above sea level – that are the dominant high bulges in the continents today. All the other mountain ranges are relatively small by comparison, reaching average elevations of only 1 to 2 km, though prominent features of the Earth's surface nonetheless (Figure 2).

The creation of a mountain range leads directly to profound interactions among the solid Earth, biosphere and atmosphere, generating topography and driving rainfall, erosion, sedimentation, chemical rock weathering, CO_2 draw-down, and climate change, as well as giving us in places the dry land we live on. It is therefore not surprising that ever since scientists and philosophers first started thinking about the history of our planet, the formation of large mountain ranges has played a prominent role in their ideas.

Vertical forces beneath mountains

So why are some areas of the continents much higher than everywhere else? It was the work of surveyors in the 18th and 19th centuries, trying to measure the shape of the Earth with the greatest possible accuracy, which provided the first clues. The famous French expedition to Ecuador in 1735, led by Charles-Marie de La Condamine, showed that contrary to expectation, the sideways pull of gravity amongst the high Andean peaks, measured by the deflection of a plumb bob from the vertical, was negligible, despite the vast bulk and mass of these mountains. More than a century later, the great survey of India revealed essentially the same result when it was found that the gravitational deflection of the plumb bob on the plains of northern India, immediately south of the high Himalayan peaks, such as Mt Everest, was only about half that predicted from the mass of these mountains

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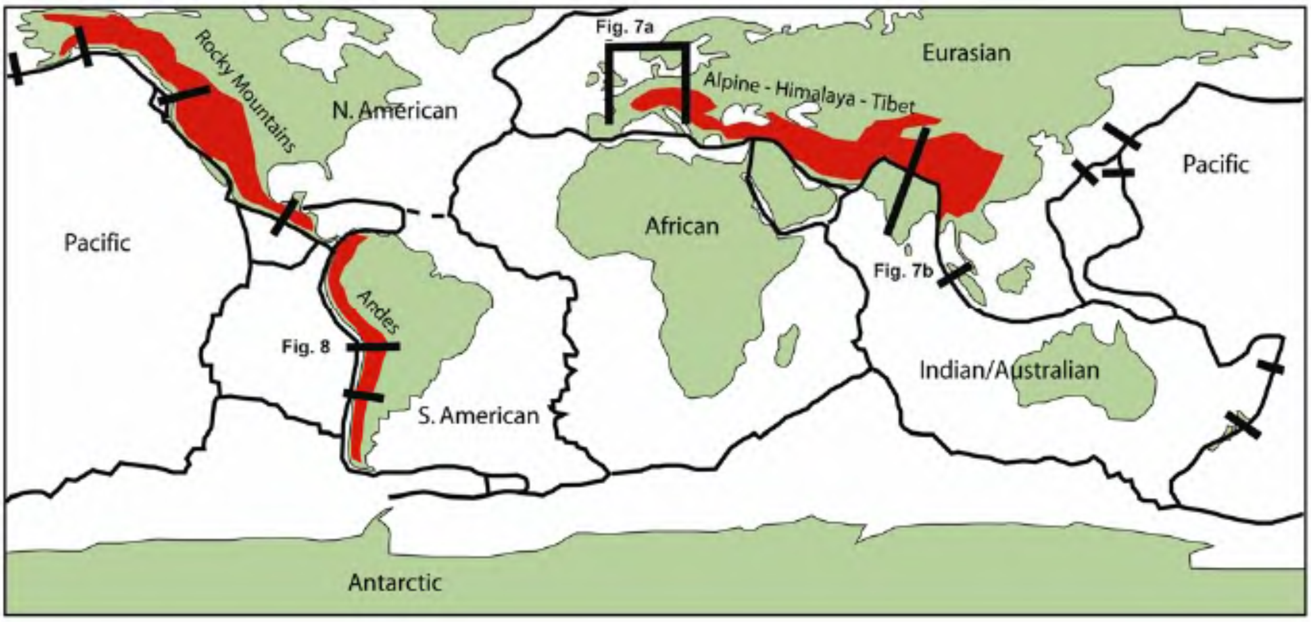

Figure 1. Plate tectonic map of the world showing the main mountain ranges on the planet, along the western margin of the Americas (Rocky Mountains and Andes) and through Europe and Asia (Alpine–Himalaya–Tibet). Also shown are the location of various transects. The topography and bathymetry for transects across subduction zones are plotted in Figure 2; other lithospheric scale transects are shown in Figures 7 *a*, *b* and 8.

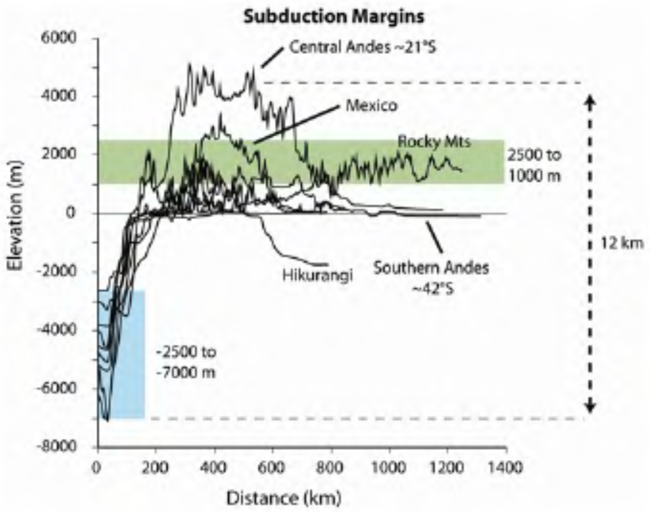

Figure 2. Typical bathymetric and topographic profiles of selected subduction zones with the volcanic arc above sea level. See Figure 1 for location of transects. Trench depths are in the range 2.5–7 km below sea level, whereas the maximum average elevations above sea level are almost all in the range 1–2.5 km. These give rise to maximum elevation contrasts between the sea floor to mountain tops in the range 3.5–8 km. The Peru–Chile subduction zone in northern Chile is a clear exception to this, having both the deepest trench (~ 7 km below sea level) and the highest elevations, reaching over 4 km above sea level, with an elevation contrast ~ 12 km.

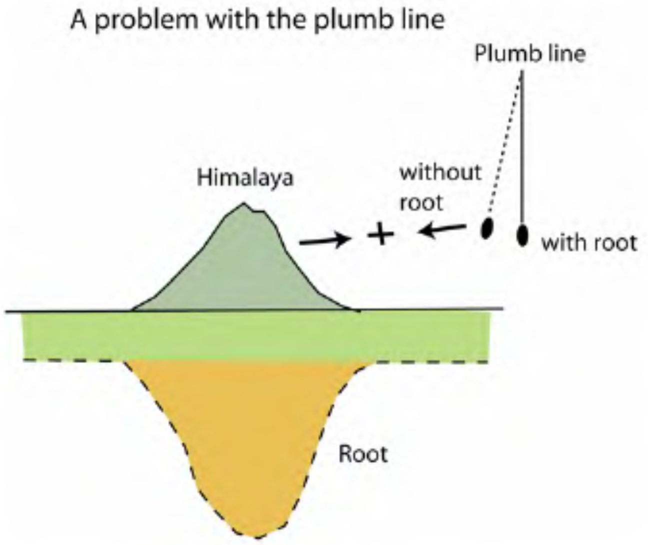

Figure 3. The great survey of India in the nineteenth century used a plumb-line to help fixed the positions of the Indian subcontinent to the greatest precision possible at the time. A triangulation survey and a comparison of astronomical and geodetic latitude showed that the deflection of the plumb-line from the vertical because of the massive gravitational pull of the great ranges of the Himalayas was much smaller than expected – it seemed as though these mountains were almost hollow. In a series of publications in the 1850s, George Airy and Archdeacon Pratt showed that this was because the density of rock deep beneath the mountains was lower than elsewhere, effectively counteracting the pull of the mountains themselves. This was an important clue that mountain ranges are the surface expression of changes in the nature of the Earth's crust at great depth.

(Figure 3). It was George Airy, who, in 1855, suggested an answer to this puzzle that has proved to be one of the most important ideas in geophysics[1].

Airy imagined that the outer part of the Earth – the crust – was effectively floating on its more fluid-like substratum. In doing so, he had made a breath-taking leap in scientific thinking, because the implicit assumption here was that the crust was less dense than the underlying fluid, despite the fact that almost all geologists at the time considered the outer layer of the Earth to be more dense because it was colder and had contracted relative to the interior. It is hard to trace where Airy actually got his idea from, but it may be significant that he was not a geologist, but an astronomer and physicist instead, and so was unbiased by prevailing geological thinking. The concept of a relatively low density crust provided a natural explanation for the surveyors results: just like icebergs floating above the level of the sea, mountains poke their summits high above the general surface of the continents because they have deep roots extending down into the fluid-like substratum, or mantle as it has come to be called (Figure 4 *a*). It was the lesser pull of gravity by these deep and low density roots that was effectively canceling out the gravitational attraction of the overlying mountain peaks, making the mountains seem almost hollow.

Archdeacon Pratt of Calcutta, who had made the first detailed analysis of the Indian surveyors' results, soon showed that there was another way to interpret the results[2]. Instead of thinking of the mountains as being like icebergs, Pratt saw them as more like bread – in the lowlands, the crust was like dense uncooked dough, whereas beneath high mountains, the crust was analogous to light risen bread (Figure 4 *c*).

Whatever the case, Airy and Pratt had established the basis of a physical principle for the building of mountain ranges – the principle of isostasy. This simply states that when the Earth's surface is disturbed by loads such as those associated with large mountain belts, or indeed, the growth and decay of ice sheets, the crust and mantle will adjust and, eventually approach some form of flotational equilibrium. This is no more than Archimedes' principle, worked out over 2000 years ago, where the vertical force that holds a mountain range up above the Earth's surface is the difference in weight between its low density crustal root and the higher density mantle it displaces (Figures 4 *b*, *d* and 5 *a*).

The same principle applies, but this time in the opposite sense, to the higher density lithospheric root, compared to the lower density asthenosphere it displaces (Figure 6 *b*), though this effect is small compared to that of the buoyancy of the crustal root, because the density contrast between crust and mantle (300–600 kg m^{-3}) is an order of magnitude greater than that between lithospheric and asthenospheric mantle (20–60 kg m^{-3}). Thus, we would expect a rise of 1 km in the surface elevation to be a consequence of either 6–8 km *increase* in crustal thickness, or 50–100 km *decrease* in the thickness of the lithospheric mantle.

Deep structure of mountain ranges

It would be well over a hundred years later, when geophysicists had developed accurate ways to probe the crust and mantle with seismic waves, that it became clear that Airy's idea fitted the data in the world's highest mountain ranges the best (Figure 5 *a*). Thus, beneath the high Bolivian Andes, the crust is 60–75 km thick, compared to around 35 km beneath the lowland plains farther east[3,4] (Figure 5 *b*). Likewise, the crust beneath the Tibetan plateau is up to nearly 80 km thick[5,6], compared to 40 km beneath the plains of northern India. And lesser mountain ranges, like New Zealand's Southern Alps[7] (Figure 6), or the European Alps and Pyrenees, all have their crustal roots, extending to depths of 40–55 km, compared to an average crustal thickness beneath the continents of 30–40 km (refs 8–10) (Figure 5 *a*). An important exception to this are the ranges of the Cordilleras and Rocky Mountains on the western margin of North America, where the average crustal thickness is only ~35 km, the same or thinner than that beneath the lowland plains farther east[11] (see below).

It would be surprising if the density of both the crust and the mantle were completely uniform, given the vastly differing thermal structures, geological histories, as well as the mineralogical changes that must take place during mountain building as a consequence of the high temperatures and pressures at depth. Furthermore, if the denser lithospheric mantle (compared to the underlying asthenosphere) thickens, it will tend to pull down on the overlying crust, creating lower mountains – or even no mountains at all – than one might anticipate from the crustal root alone (Figure 6 *b*). For example, crust up to 50 km thick is found beneath the lowland plains of North America, or beneath the low-lying Baltic region of northern Europe, in regions where the lithosphere is unusually thick[12] (Figures 6 *b* and 7 *a*).

Likewise, a thinner lithosphere will result in higher mountains than would be predicted from the depth of the crustal root alone, buoyed up by the greater thickness of the underlying less dense asthenosphere (Figure 6 *b*). Thus, much of the Canadian Cordillera has average elevations around ~1500 m, yet is underlain by relatively thin crust 30–40 km thick, but the lithosphere here may also be very thin[11,13] (Figure 6 *b*). The buoyancy of the lower density asthenosphere beneath the thin lithosphere of the high parts of the Central Andes is also helping to hold up the mountains here, though the good fit with simple crustal Airy isostasy suggests the effect is relatively small (< 1 km, Figures 5 *b* and 8 *a*).

Isostasy need not only apply at very local scales, over distances of a few kilometres to tens of kilometres. This is because the strength of the lithosphere is capable of spreading the load of mountains laterally over much greater distances (Figure 4 *e*, see below). But on a scale much greater than the thickness of the lithosphere – over 1000s kilometres – the outer part of the Earth must always be close to isostatic equilibrium[14]. And this indeed is what the small regional Free Air gravity anomalies in the continents tell us[14] (Figure 5 *b*).

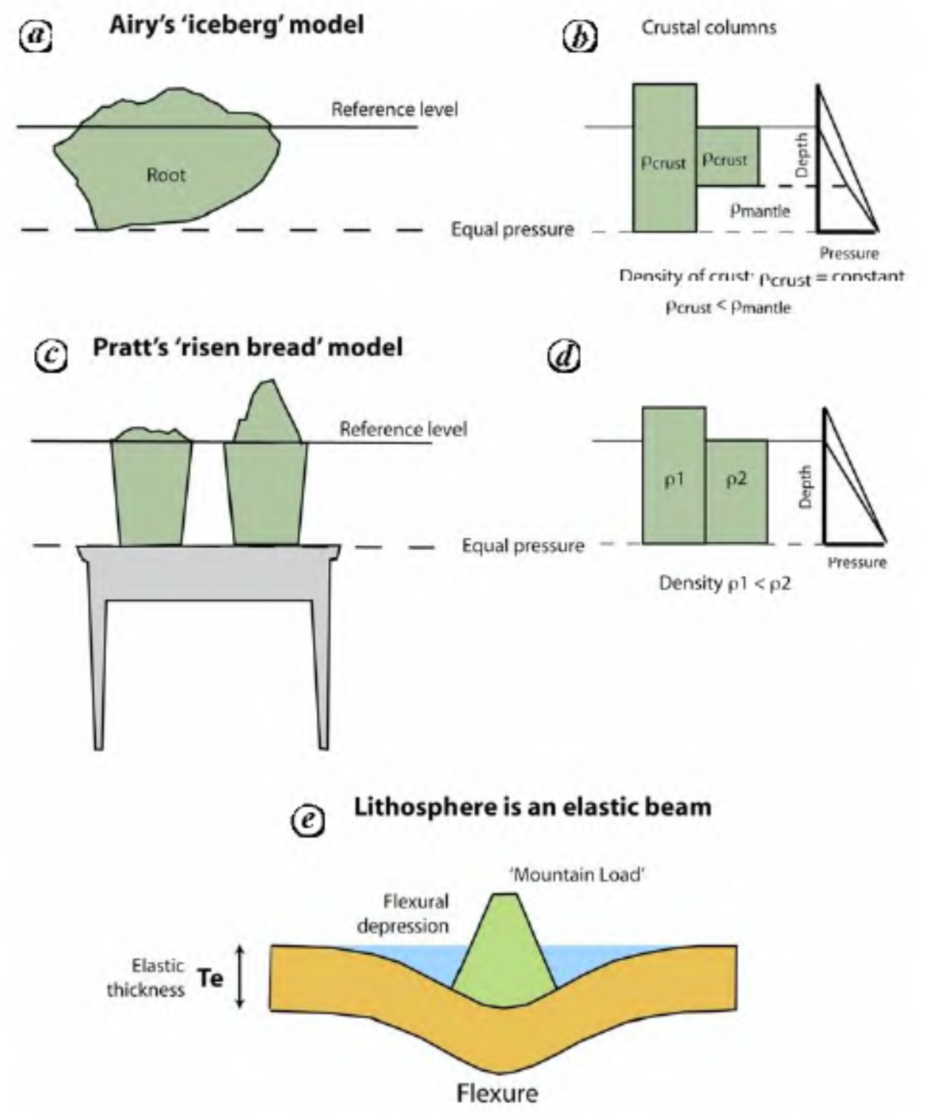

Figure 4. A series of diagrams which illustrate the principle of isostasy. ***a***, George Airy thought of mountains as being like icebergs, forming a low density crust that is floating on a fluid and more dense substratum (now called the mantle). ***b***, As mountain ranges rise higher, so their roots extend deeper into the underlying mantle, and at the depth of the deepest roots the pressure is constant. ***c***, Archdeacon Pratt saw mountains as being more like risen bread, reflecting density changes in the crust itself. ***d***, High mountain ranges are underlain by lower density crust compared to that beneath the lowland plains, again with a level of constant pressure at the base of the crust. The ideas of both Airy and Pratt apply to different parts of the Earth. ***e***, However, the outer part of the Earth, referred to as the lithosphere, has long-term strength, acting like an elastic beam that spreads the load of the mountains, creating pronounced flexural depressions which form important sedimentary basins.

Horizontal forces beneath mountains

Armed with the simple idea of isostasy, we can now see the topography on the continents in terms of density contrasts and marked variations in the thickness of the underlying crust or lithosphere (Figures 5 *a* and 6 *b*). But we still have no explanation for why such variations in thickness exist in the first place.

Lithospheric shortening and crustal thickening

Geologists, ever since they started examining the great mountain ranges of the world, have realized that the rock strata here are intensely folded and cut by reverse faults, with individual displacements up to tens of kilometres. Thus, the most obvious explanation for variations in crustal thickness is one of squeezing. In any vertical section through the crust or lithosphere, if the cross-sectional area is to remain constant, then squeezing it in the horizontal direction will result in thickening in the vertical (Figure 9). Indeed, one can use Airy's model to calculate exactly how much squeezing would be required.

Unfortunately, it is less easy to work out how much squeezing has actually taken place in a mountain belt, because the strata and faults are often not exposed, or have been eroded away, and it is difficult to determine the

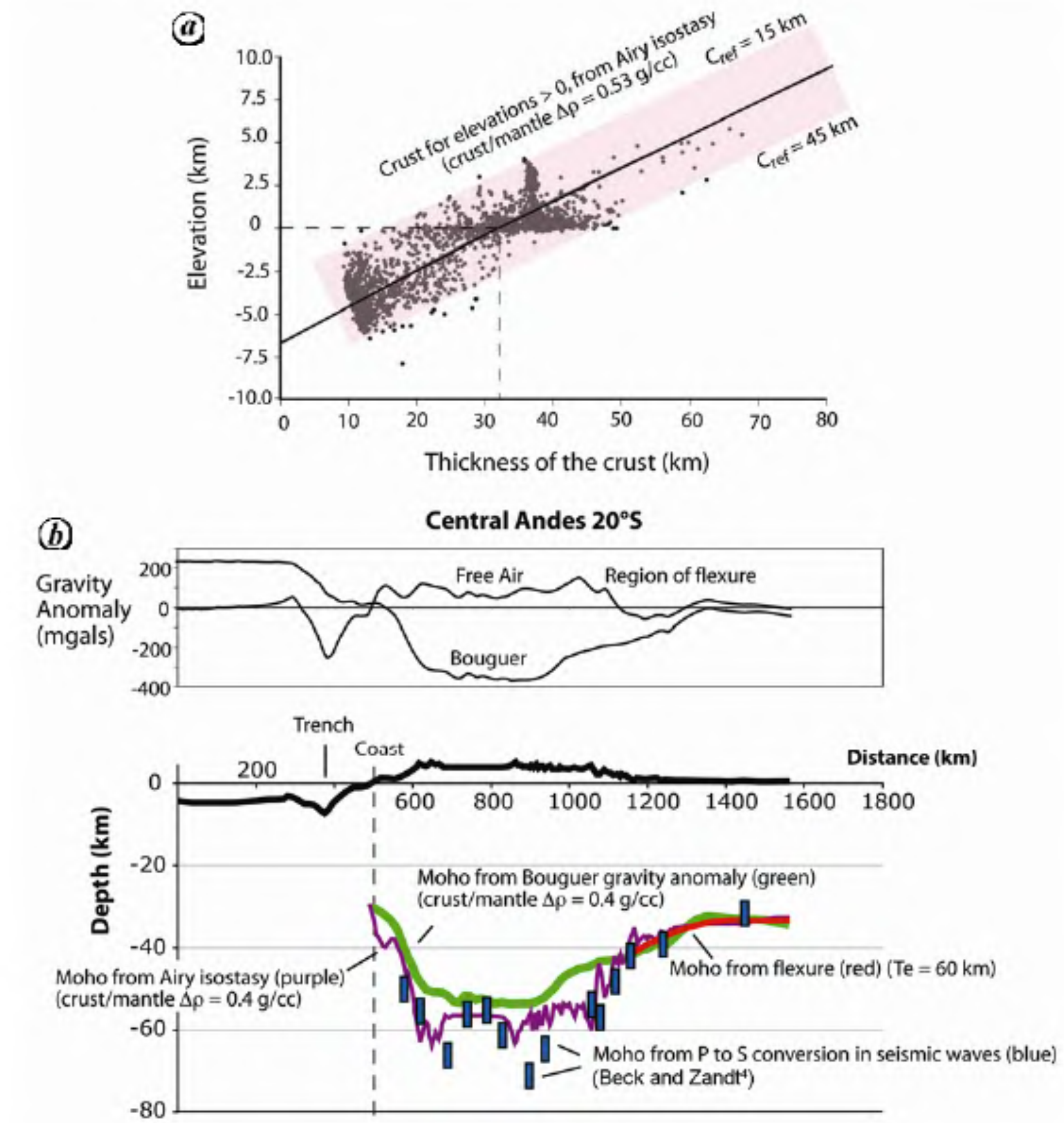

Figure 5. ***a***, Plot of a 5° × 5° average of elevation against crustal thickness[8]. The crustal thickness is based on observed seismic refraction data. The solid line shows a linear fit based on the Airy model, if ρ_{crust} = 2.8 g/cc, $\rho_{\text{sub-oceanic mantle}}$ = 3.18 g/cc, $\rho_{\text{sub-continental mantle}}$ = 3.33 g/cc, ρ_{water} = 1.03 g/cc (note: 1 g/cc = 1000 kg m^{-3}), and zero elevation crustal thickness = 31.2 km. Shaded box shows range for zero elevation crustal thickness between 15 km and 45 km, for the same density model as solid line. ***b***, Cross-section through the Central Andes at 20°S, showing the Free Air and Bouguer gravity anomalies, and various estimates of the crustal structure (see Figure 1 for location). P to S conversions of seismic waves have been used to estimate the depth to the Moho, combined with velocity modelling[4]. Topography together with the Airy model of isostatic compensation shows remarkable agreement with the seismic data. The Free Air and Bouguer gravity anomaly suggests lithospheric flexure on the eastern margin of the Andes, where the 'gravity' Moho is up to 5 km deeper than the 'Airy' Moho, with an elastic thickness of about 60 km. However, in the high parts of the Andes, the gravity Moho is significantly shallower than the 'seismic' and 'Airy' Mohos, suggesting the gravity anomaly also reflects higher density material at depth; presumably this is the high density subducted slab which extends down from the trench farther west.

displacements on every single fault. Nonetheless, measurements of the amount of shortening that has taken place in the Tibet or Andes – in the range of hundreds to thousands of kilometres[15–17] – strongly support the conclusion that this shortening is the main mechanism of crustal thickening. Shortening could thicken the mantle part of the lithosphere too, and this may be the explanation for the thick lithospheric root beneath New Zealand's Southern Alps[18] (Figure 6).

In the context of plate tectonics, we would anticipate mountain building wherever continental plates converge, because here the crust is likely to be shortened and thickened. Local compression and thickening of the crust may also occur in continental plate-boundary zones dominated by strike-slip deformation, related to local changes in the orientation of the faults – for example, the Transverse Ranges in the San Andreas Fault System[19] (Figure 10 *c*). However, crustal thickening is not an inevitable

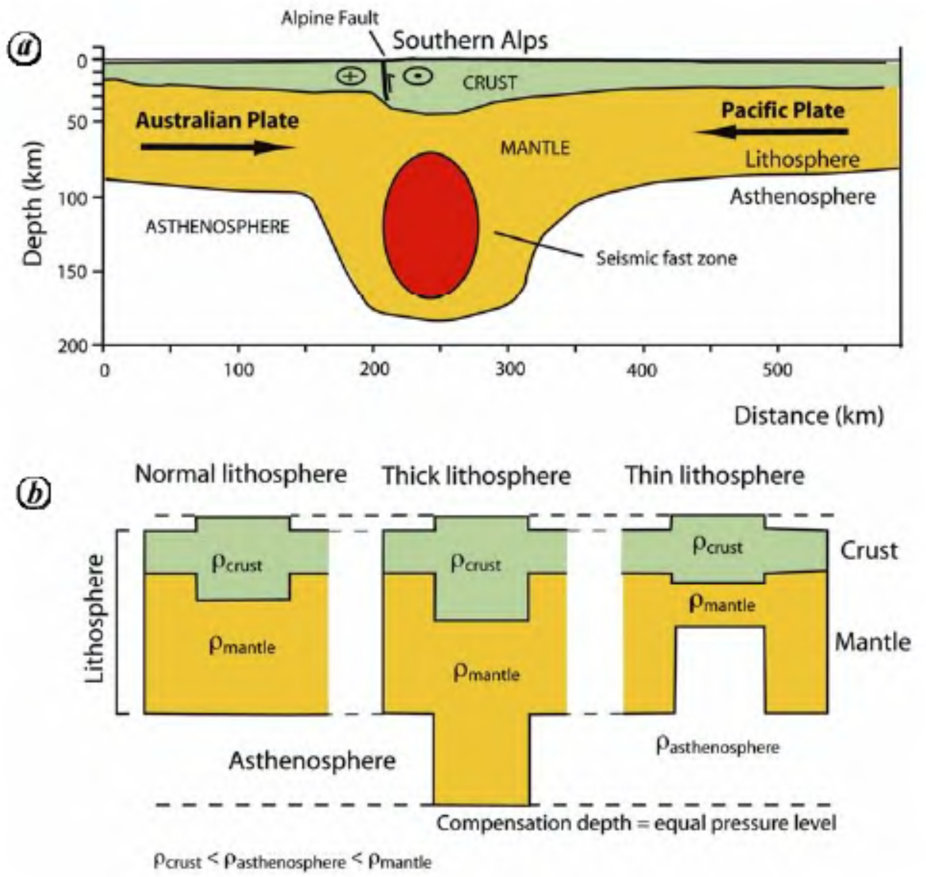

Figure 6. ***a***, Lithospheric scale cross-section through New Zealand's Southern Alps after ref. 18, showing evidence for homogeneous crustal and lithospheric shortening as a result of plate convergence. Here, about 70 km of lithospheric shortening has been accommodated in the last 25 Ma, together with nearly 500 km of dextral strike-slip on the Alpine Fault[75]. Shortening is clearly reflected in both the crustal and lithospheric structure: crustal thickness nearly doubles from about 20 offshore to about 40 km beneath the Southern Alps, and a zone of seismically fast mantle beneath the thick crust suggests the lithosphere has also nearly doubled in thickness[18]. Lithospheric and crustal thickening has resulted in a narrow mountain range, about 50 km wide and reaching average elevations of about 1.5 km above sea level. ***b***, Cartoons illustrating the effect on the elevation of mountains caused by changes in lithospheric thickness. The lithospheric mantle is more dense than the underlying asthenosphere mantle, simply because the lithosphere is colder. Therefore, a thick lithospheric mantle root will tend to reduce the buoyancy effect of the crustal root, compared to normal lithosphere, holding down the mountains. In contrast, the asthenosphere beneath a thinner lithosphere will buoy up the mountains, even if the crustal root is small. These effects are small compared to that of the buoyancy of the crustal root, because the density contrast between crust and mantle (300–600 kg m^{-3}) is an order of magnitude greater than that between lithospheric and asthenospheric mantle (20–60 kg m^{-3}), changing the elevation of the mountains by no more than ~1500 m, and usually only a few hundred metres.

consequence of convergence, or even crustal shortening. Extension – ultimately leading to back-arc spreading – can develop in subduction zones where the roll-back of the subducted slab exceeds the rate of overriding of the overlying plate[20]. And in some cases of continental collision, the crust could also be pushed or extruded sideways along strike-slip faults, without any crustal thickening – or with even thinning – giving rise to what has been termed 'continental extrusion' tectonics.

Thus, it has been suggested that part of the northward convergence of India with Asia has been accommodated by the eastward extrusion of Tibet towards China along major strike-slip faults such as the dextral Karakoram Fault and the sinistral Altyn Tagh Fault[21], together with some extension on north–south normal faults[22], although the very thick crust in Tibet shows that extrusion must at most only absorb a small fraction of the total convergence[23]. Similarly, the westward motion of Turkey towards the Aegean has helped accommodate the northward convergence of Africa with Europe.

Finally, other more local processes, such as sedimentation, and the addition of new material to the crust through mantle melting, magmatic underplating and volcanism, may also locally help to thicken the crust, especially beneath volcanic arcs[24,25]; whereas erosion will act to thin the crust (Figure 9 *a*). Thinning of the lithosphere beneath mountains is more difficult to explain, but could be a consequence of gravitational instabilities in the thickened

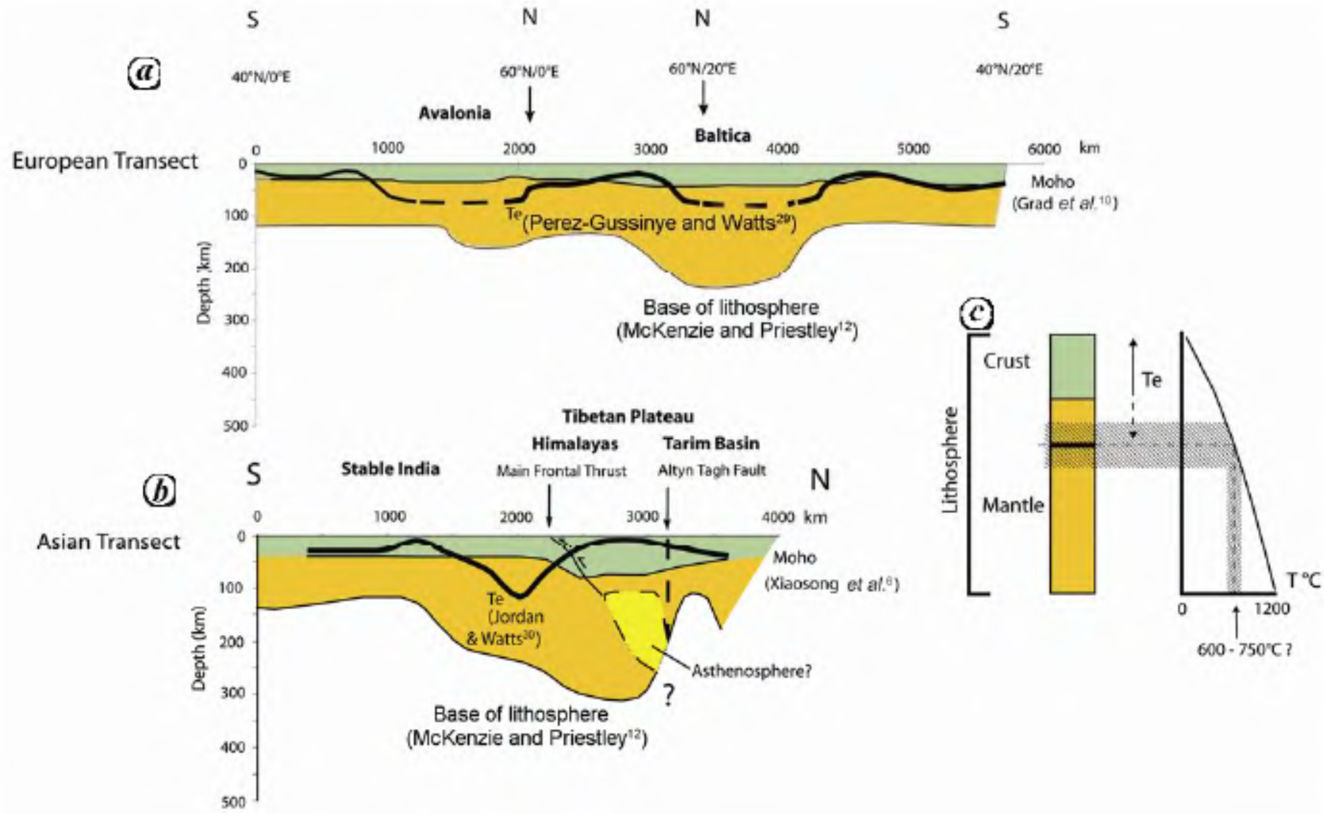

Figure 7. Lithospheric scale transects through northern Europe (***a***) and Central Asia (***b***), based on seismic and gravity measurements, showing lithospheric and crustal structure as well as the effective elastic thickness – note the general marked positive correlation between the elastic thickness and lithospheric thickness in the low-lying and stable continental regions (see Figure 1 for locations). Beneath the Himalayas and Tibet, the thick lithosphere may reflect both underthrusting of the northern Indian plate and homogeneous lithospheric shortening and thickening. ***c***, The elastic thickness fits well with continental lithospheric geotherms, if the limit of elastic behaviour is at mantle temperatures in the range 600–750°C, coinciding with the maximum temperature for earthquakes in the oceanic mantle (see text). In particular, where the lithosphere is thick, the elastic thickness is greater than the crustal thickness, and the elastic 'beam' extends into the mantle.

lithospheric mantle, especially if the lithosphere is weak[26,27] (see further on).

Horizontal force balance

The process of shortening the crust or lithosphere must be driven by horizontal forces. We can use the isostatic balance to estimate what these horizontal forces might be (Figures 4 *b*, *d* and 11 *a*). Imagine a small cube of rock buried deep in the lithosphere, with faces that are orientated both horizontally and vertically (Figure 11 *b*). The force on the horizontal faces is the weight of rock bearing down vertically. Isostasy tells us what we would expect this weight to be in neighbouring columns of rock, at any level down to the base of the deepest mountain root (Figure 11). Any difference between this force and the horizontal force on the vertical faces is the real measure of the force driving mountain building and also the strength of the crust or lithosphere – a more general way of expressing this is in terms of forces per unit area (i.e. stresses or pressures measured in pascals where $1\ \mathrm{Pa} = 1\ \mathrm{kg\,m^{-1}\,s^{-2}}$) on the faces of the cube (Figure 11).

The mountains will grow, maintaining their tendency to approach isostatic equilibrium at the same time, until their weight, together with their strength, balance the available horizontal push to hold them up[28] (Figure 11). More precisely, in this case, if their strength is small compared to the forces holding them up, then a mountain range will approach its maximum height when the total available horizontal push essentially balances the buoyancy force, which is the depth integrated vertical force (weight) per unit area in our column of rock ($F_T = F_B$ in Figure 11). Measured this way, average lithospheric horizontal stresses that drive mountain building tend to be < 150 MPa. At this point, the mountain range can only continue to grow by becoming wider rather than higher. If at some later date, the horizontal force decreases, the mountains will begin to collapse and extend, shrinking until their height again balances the new state of forces in the lithosphere.

Continental extrusion will occur where the lateral support of the mountain range, along its length, is insufficient, and so rather than the crust being pushed up, against the force of gravity, the crust moves sideways, in

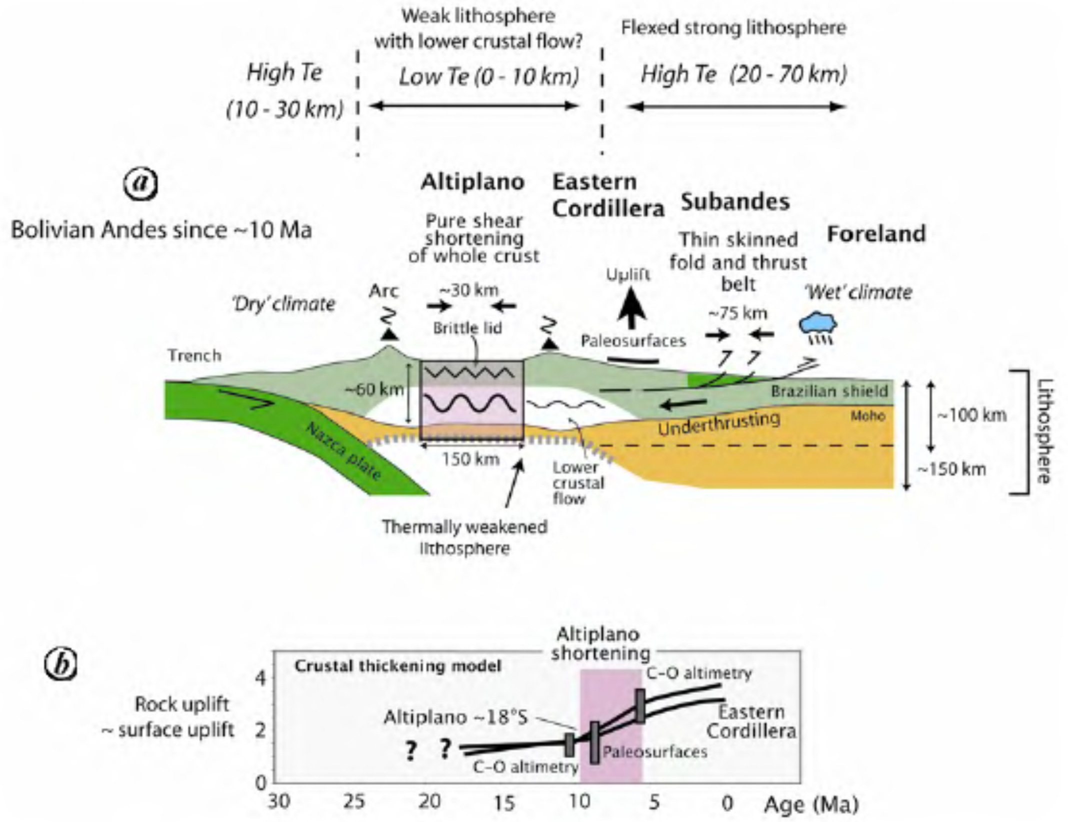

Figure 8. Lithospheric cross-section through the subducting Andean margin at 20°S, in Chile and Bolivia (see Figure 1 for location, after refs 24, 25, 37), showing the main tectonic features. Here, the Nazca plate is subducting beneath the western margin of South America, with a region of thick crust and high elevations extending several hundred kilometres farther east (see also topographic profile in Figure 2). ***a***, Since ~10 Ma, there has been more than 100 km of shortening in this region, with marked vertical partitioning of the deformation, so that at any point on the surface the amount of shortening varies significantly with depth. This has been mainly accommodated by thin-skinned shortening (~75 km) on the eastern margin, where the 'strong' lithosphere of the Brazilian Shield is underthrusting the Subandes, acting as a sort of 'piston', pushing its way into the ductile lower crust farther west, beneath the high Altiplano. The flexural rigidity of this lithosphere (Figure 5, elastic thickness ~60 km) both helps to hold up the Andean mountains themselves, as well creating a flexural foreland basin, now filled with sediment eroded from the mountains themselves. Seismic and geochemical evidence (summarized in ref. 25) suggests that the mantle part of the lithosphere beneath the high Altiplano is thinner compared to beneath the foreland, farther east (~100 km compared to ~150 km thick), possibly related to delamination around ~25 Ma, and the lithosphere is weak with significant lower crustal flow, with a very low elastic thickness (T_e is 0–10 km). Crustal shortening of the whole thickness of the crust beneath the Altiplano has accommodated about 30 km of additional shortening. Note that Neogene deformation has been focused on the wetter side of the Andes. ***b***, All this deformation is driving uplift of the high Andes. Estimates of palaeo-elevation from the geomorphology of uplift peneplains, as well as stable isotope measurements of carbonates in paleosols[44,65,66], suggest 1–3 km of rock uplift (~ surface uplift, given the small amount of erosion) of the high Andes since ~10 Ma, which can be explained in terms of the crustal shortening and thickening here[25].

effect collapsing laterally under the force of gravity towards regions of lower elevation[23].

Elastic behaviour of the lithosphere

Earthquakes clearly show that the outermost part of the Earth behaves elastically and has some finite strength: when forces in the Earth exceed this strength, the rocks rupture, generating earthquakes – it is only this way that significant horizontal deformation of the Earth's surface can occur to push up a mountain range. Thus, the elastic properties of the lithosphere have important implications for the formation of mountains.

If large enough, this elastic strength would also tend to spread the load of mountains laterally (Figure 4 *e*). In fact, clear evidence for this can be found at the edges of all major mountain belts, where the weight of the mountains has pushed down the surrounding lowlands several kilometres, creating a foreland basin up to a few hundred kilometres wide that has filled with sediment mainly derived from the mountains themselves (Figures 4 *e* and 12 *a*). In effect, the lithosphere here is behaving like an elastic beam, bending or flexing under a vertical load and the elevations are lower and the depth to the base of the crust is deeper than that suggested by simple local isostasy[14] (Figure 12). The corollary of this is that the

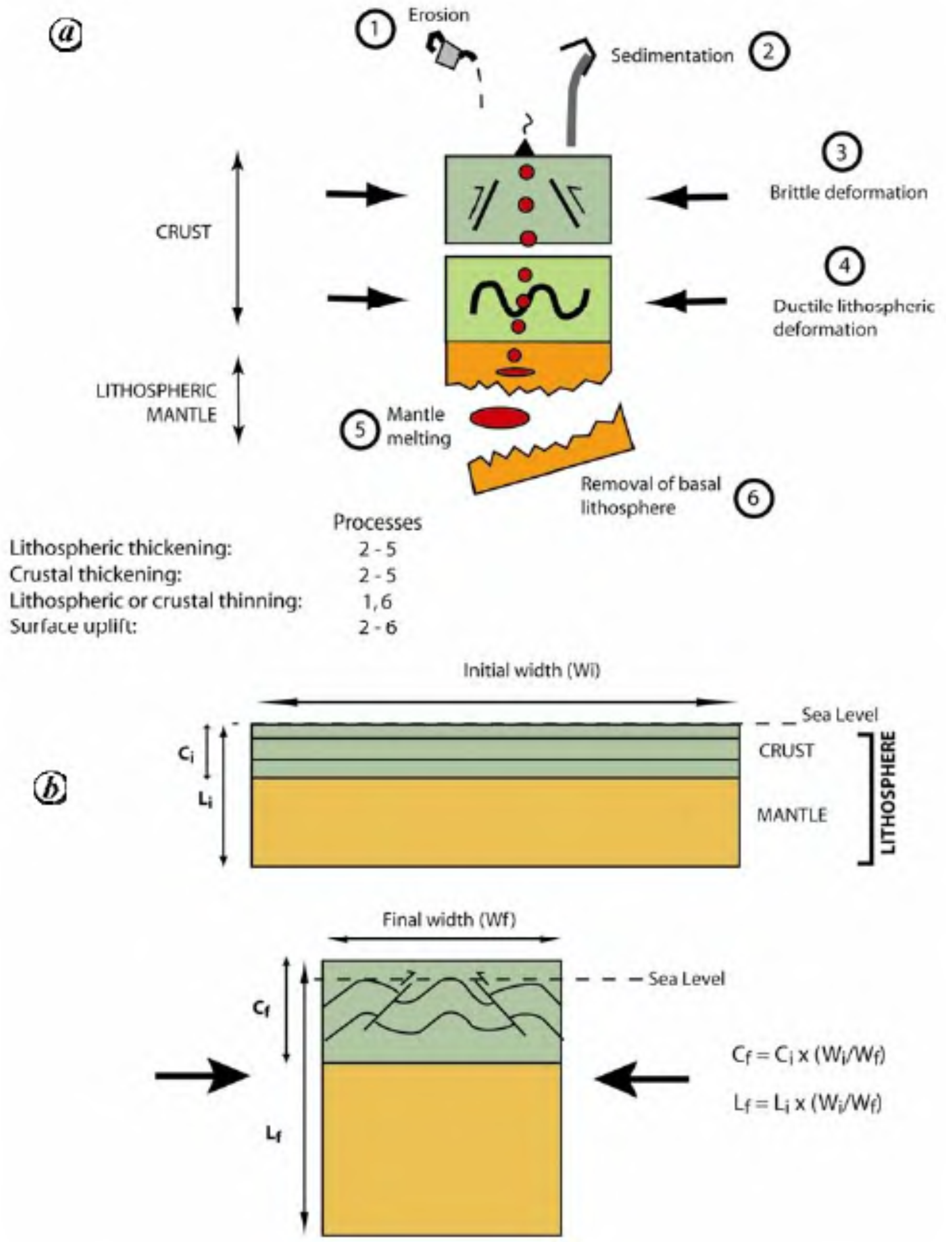

Figure 9. Processes to thicken or thin the crust and lithosphere. ***a***, Crustal thickening is a consequence of both shortening in the brittle upper crust, accommodated by reverse faulting and folding, and ductile flow in the lower crust and mantle lithosphere. Surface processes, such as sedimentation or erosion, can add to or remove the top part of the brittle crust. At depth, mantle melting can thicken the crust by adding new melt to the crust, whereas removal of the basal part of the lithosphere can thin the lithosphere. All these processes except erosion will drive surface uplift. ***b***, If no material enters or leaves the plane of the section, the area of any vertical section of crust will be preserved during crustal and lithospheric shortening. In the simplest case, crustal thickening and surface uplift can be directly related to horizontal crustal shortening.

strength of the beam is helping to hold up the mountains themselves, so that the foothills of the mountain range will be slightly higher and the base of the crust shallower beneath them than simple local isostatic models would suggest.

The elastic beam-like nature of the lithosphere is revealed in the Earth's gravitational field[14]. In simple terms, the gravitational field is reduced in the lowlands where the lithosphere has been pushed down, and increased in the foothills where elastic strength is locally helping to hold up the mountains. Like the geologically important foreland basins, these negative/positive couples in the Free Air gravity anomalies are clear to see on the edges of major mountain belts (Figure 5 *b*), with amplitudes of ± 200 mgal (1 mgal = 10^{-5} m s^{-2}). They can be used quantitatively to estimate the thickness of the effective elastic beam in the lithosphere[14]. But it is important to appreciate that this beam need not actually exist as a single entity, rather it is a description of the general mechanical behaviour of the lithosphere[14].

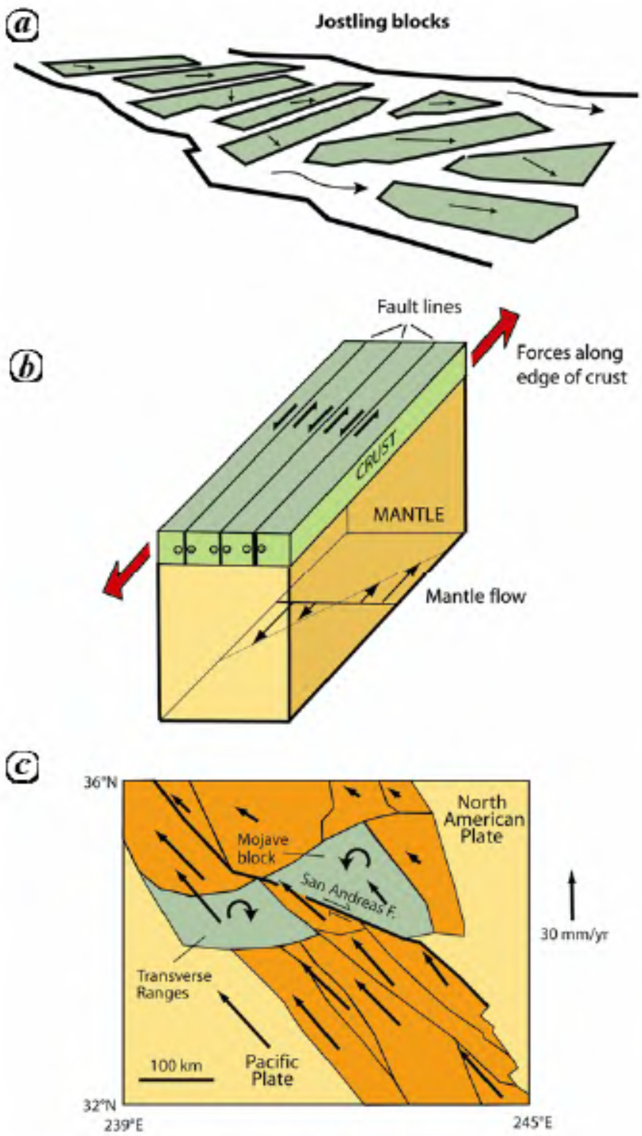

Figure 10. ***a***, In wide zones of continental deformation, the brittle crust is broken up into blocks by faults, forming a mosaic of 'jostling' blocks, which on a large scale give rise to an overall pattern of flow. ***b***, A crucial question is whether the motions of these blocks are mainly controlled by forces exterted by the faults themselves, defining the edges of the blocks, or by forces at the base of the blocks related to some more fundamental pattern of lower lithospheric flow. The relative magnitudes of these edge or basal forces depend on the rheology of the lithosphere, and, in particular, the relative strength of the crust and mantle (see Figure 13). ***c***, Lithospheric block model for the pattern of deformation in the San Andreas Fault Zone in western North America, after ref. 19. The motion of the blocks, shown by arrows relative to North America, is constrained by GPS measurements, with velocities generally between 5 and 40 mm/yr. The blocks are also rotating about a vertical axis at 0–1.2°/Ma, both clockwise (for example, Transverse Ranges) and anticlockwise (for example, Mojave Region), and the overall pattern of motion accommodates the relative motion between the Pacific and North American plates. The Transverse Ranges have been pushed up in a zone at a high angle to the main dextral strike-slip faults.

Elastic thickness and lithospheric strength

The thickness of the elastic beam (elastic thickness T_e) is a measure of the strength of the lithosphere – the thicker the beam, the greater the horizontal forces in the lithosphere needed to break it (Figures 4 *e* and 12 *b*). The main conclusion of both gravity and seismic studies is that both the elastic strength and thickness of the lithosphere vary markedly in the continents[12,14,29,30] (Figure 7).

Those areas that form the low lying – close to sea level – stable parts of the continents, which have not undergone significant geological activity for hundreds or thousands of millions of years, have the highest values of elastic thickness, up to 100 km or more (Figures 5, 7 and 12) and consistent with a strong lithosphere. The thickness of the lithosphere in these regions is also large, in the range 150–350 km, and so the high values of elastic thickness fit well with an upper temperature in the range 600–750°C for brittle failure in oceanic mantle lithosphere[31–33], if the base of the lithosphere approximately follows the 1200°C isotherm (Figure 7 *c*).

In marked contrast, the highly deformed interiors of large mountain belts, such as the Tibetan plateau or Altiplano in the Andes, where we would anticipate the lithosphere to be much weaker, have much smaller elastic thicknesses in the range 0–20 km (refs 30, 34–36) (Figures 7, 8 and 12). The lithosphere beneath the Central Andes is also thinner than that farther east, beneath the foreland[25,37] (Figure 8). However, it is unclear whether or not thin lithosphere lies beneath Tibet[5,12] (Figure 7).

Elastic thickness and thin-skinned deformation

The flexure of the lithosphere has an important implication for mountain building. Put simply, if mountain building is a consequence of shortening and thickening of the crust, then the elastic beam must be pervasively broken by faults. But to break the beam means to lose much of its flexural properties. The large elastic thickness at the margin of a mountain belt, in the Himalaya or eastern foothills of the Andes, suggests that despite the intense faulting here, the elastic beam remains essentially intact.

There is a simple solution to this apparent paradox. Whereas in the interior of a large mountain belt, deformation can extend to any depth in the lithosphere – sometimes referred to as thick-skinned deformation – on the margin, the deformation is only in a thin wedge, overlying the stronger and rigid basement that makes up the elastic beam, referred to as thin-skinned deformation (Figures 7, 8 and 12). This basement is merely the lithosphere of the neighbouring lowland regions that has been pushed – underthrust – beneath the margin of the mountain belt, sliding along a giant gently inclined fault extending for up to several hundred kilometres. In the process, flexing of the elastic beam-like basement

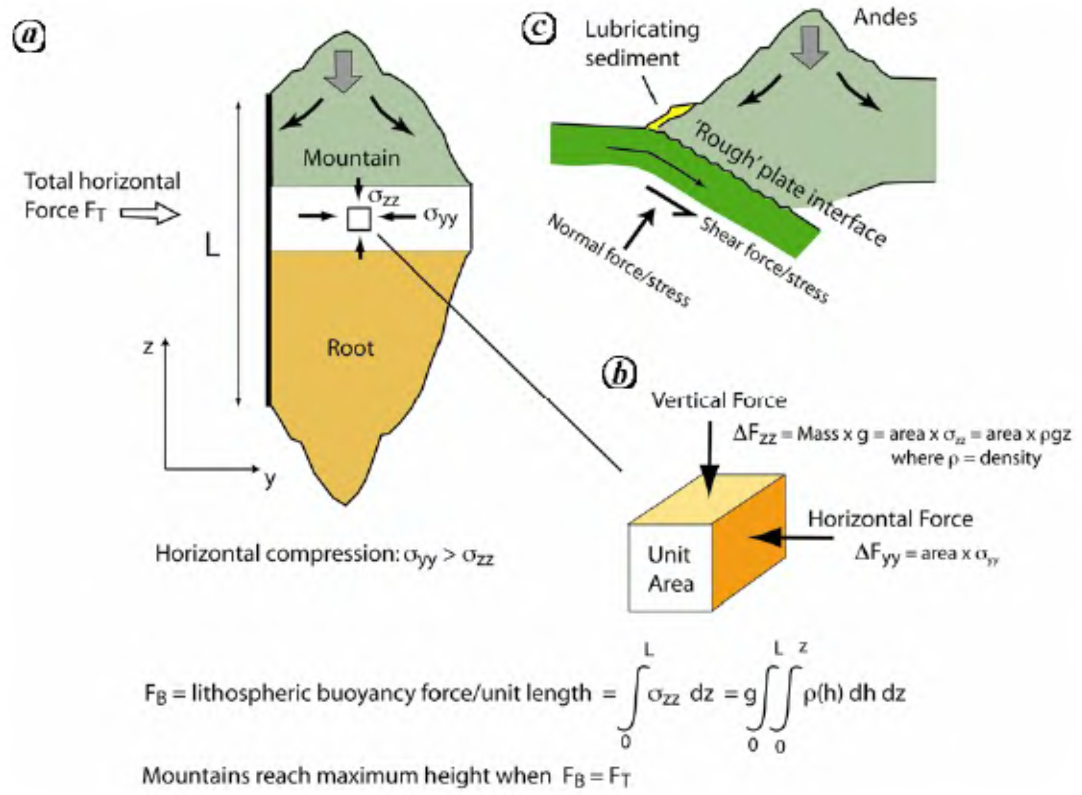

Figure 11. The lithospheric force balance beneath high mountain ranges. ***a***, For the long-term existence of a mountain range, the weight of the crust must be balanced by both the strength and horizontal force in the lithosphere. ***b***, These forces can be analysed in terms of forces (or stresses σ) on the faces of a small cube of rock. The weight of rock exerts a vertical force (or pressure σ_{zz}) on the horizontal faces of the cube, whereas the horizontal force (or stress σ_{yy}) is exerted on vertical faces. The difference between these two forces or stresses is a measure of the strength of the rock. The integrated vertical force (weight) with depth, in a column with unit cross-sectional area, defines the buoyancy force per unit length (F_B). In general, because the strength of the crust in deforming zones is likely to be small compared to the forces, the mountains will reach their maximum height when the buoyancy force F_B equals the total horizontal force F_T. ***c***, In subduction zones, the horizontal force must be transmitted across the subducting plate interface. For this reason, the 'roughness' or coupling of the plate interface will be an important factor in determining the maximum height of mountains behind subduction zone. This coupling may be increased if there is a lack of water-rich trench sediment fill that could act as a lubricant, raising fluid pressures and smoothing out the plate interface.

beneath the Himalayas and plains of northern India has resulted in earthquakes in the 'beam' itself, where normal faulting has accommodated extension in the upper part of the curved beam, and thrust faulting in the lower part – here, the neutral surface is not in the centre of the beam because it is also transmitting the horizontal force to hold up the mountains[38–40] (Figure 12).

As the underthrust basement slides farther beneath the edge of the mountain range, the crust near the tip of the thin overlying wedge is shortening and thickened. On the eastern margin of the Central Andes – in the Subandean zone – and also the southern margin of the Himalayas, this has resulted in prominent fold and thrust belts that have accommodated tens to hundreds kilometres of shortening (Figure 8 *a*). This shortening is a measure of how far the underlying basement has slid, acting as a sort of 'piston' that pushes and squeezes its way into the much weaker lower crust in the more internal parts of the mountain belt (Figure 8 *a*).

Overall, there is a marked vertical partitioning of deformation, so that the thin-skinned deformation on the margin of a mountain belt is accommodated by intense lower lithospheric shortening – together with some upper lithospheric shortening as well – in its weaker interior[41], giving rise to the distinct patterns of Cenozoic shortening in both Central Andes[42–44] (Figure 8 *a*), and Himalayas and Tibet[15,45,46] (Figure 7).

The 'flow' of the lithosphere

It is clear that where the lithosphere is deforming, it is cut by faults at the surface. Earthquakes provide a historical snap shot of where at least some of this brittle deformation in the continents is occurring deeper in the lithosphere – generally at depths shallower than about 40 km, and mainly in the top ~ 15 km of the crust[47,48] (Figure 13), forming a mosaic of jostling rigid blocks, translating and rotating (Figure 10 *a*, *c*). So, on a scale much bigger than the blocks – 100s kilometre scale – the motion of the crust will result in a large-scale pattern of flow[49].

But what controls this overall motion of the blocks? In particular, is it mainly the resistance on faults, forming the sides of the blocks, or is it deeper in the lithosphere – this boils down to where the main long-term strength of the lithosphere resides, and whether it is mainly in the crust or mantle[38,39,49] (Figure 10 *b*). If it is the resistance on faults, then the dynamics of lithospheric deformation are difficult to constrain, requiring an understanding of the forces involved in the interaction of many rigid blocks. However, observations about deeper levels in the lithosphere suggest that in some cases, particularly in the interiors of large mountain ranges like the Central Andes or Tibet, there may be an overall simplicity to the large-scale flow.

Experiments on rocks at high temperatures and pressures have long suggested that sufficiently deep in the Earth, rocks should flow rather than break along faults in a brittle manner[50], and so the large-scale pattern of flow observed at the surface really is a fluid flow at depth. Thus, if the long-term strength of the deforming lithosphere actually resides in the deeper ductile parts of the lithosphere – in the mantle – rather than the overlying brittle crust, it may be the forces at the base of the crustal blocks, where the rock starts to become ductile, that govern their overall motion[12,38,39,51] (Figure 13). In this case,

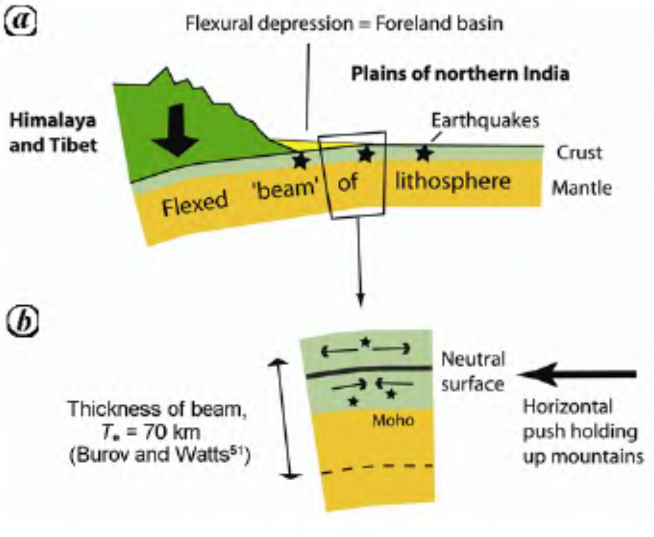

Figure 12. The effect of flexure of a lithospheric plate on the margin of a mountain belt. ***a***, In northern India, the Indian lithosphere, where the crust is ~40 km thick, is being underthrust beneath the Himalayas (see also Figure 7). In the process, it is behaving as an elastic beam, with an effective thickness of about 70 km (ref. 51), pushed down and flexed by the weight of the mountains themselves. This way, a depression several hundred kilometres wide has been created beneath the plains of northern India, now filled with sediment eroded from the mountains themselves. ***b***, Flexing of the lithosphere has resulted in extensional or compressional rupture and earthquakes in the top and bottom parts of the curved beam[38]. However, the transition between these two types of failure – the neutral surface – is not in the centre of the beam because the beam is also under horizontal compression, providing the horizontal force to hold up the mountains[39,40].

the faulted crust is merely a thin layer of brittle material floating on top, carried along by the underlying flow[49,52,53] (Figure 10 *b*).

Thus, where the effects of the deeper ductile flow are dominant, we should expect the physics of fluid flow to describe lithospheric deformation – particularly at a scale much bigger than the individual fault blocks (over 100s km) – and the force of gravity may play a large role in driving this flow. And for mountains to exist at all, the lithosphere must be undergoing some kind of ductile flow at depth. This suggests that if we want to really understand mountain building, rather than just looking at the end result of all the faulting in the crust, we should be studying how they are moving. The global positioning system (GPS) is now so accurate, that coherent images of moving mountains can be obtained in just a few years by repeatedly tracking the positions of a network of survey markers, revealing displacements ranging from millimetres to tens of millimetres each year (Figures 10 *c* and 14 *b*).

An analysis of the flow in the great ranges of Asia[54], or the Andes of South America[55] shows that the measured movements over 100s kilometres do have many of the characteristics of a fluid, flowing under the influence of gravity – and where the mountains are highest they are presently being squeezed and thickened the least rapidly, and where they are lowest they are being squeezed the most rapidly, and in a direction normal to their general trend, with an average flow viscosity of 10^{21}–10^{22} Pa s (refs 55, 56) (Figure 14 *c*). Thus, over distances of 100s kilometres, and timescales as short as a few years, the deforming continental lithosphere in these places really does seem to have fluid-like characteristics, with an average viscosity that is about 10 to 100 times greater than that of the underlying asthenosphere, but 10 to 100 times weaker than strong plate-like lithosphere.

Thin crustal blocks, thick lithospheric blocks and lower crustal flow

In general, a truly fluid view of the continents cannot be the whole story, and the behaviour of the lithosphere is likely to be controlled by both the strength of the brittle top layer of the crust (and possibly mantle) and the ductile flow in the deeper parts of the lithosphere, and the relative contributions will vary in the Earth, depending on the rheology of the crust and mantle (whether it is wet or dry, for example), the amount of deformation, thermal structure and thickness of the lithosphere[50] (Figure 13). Thus, it has been argued that in some places the resistance to slip on faults in the top layer of brittle crust may comprise most of the overall strength of the lithosphere, and the underlying bottom layer of the crust and mantle flow is too weak to drive the motion of the fault blocks because these regions are too ‘wet’ and ‘hot’[38,57] (Figure 13 *b*). Thus,

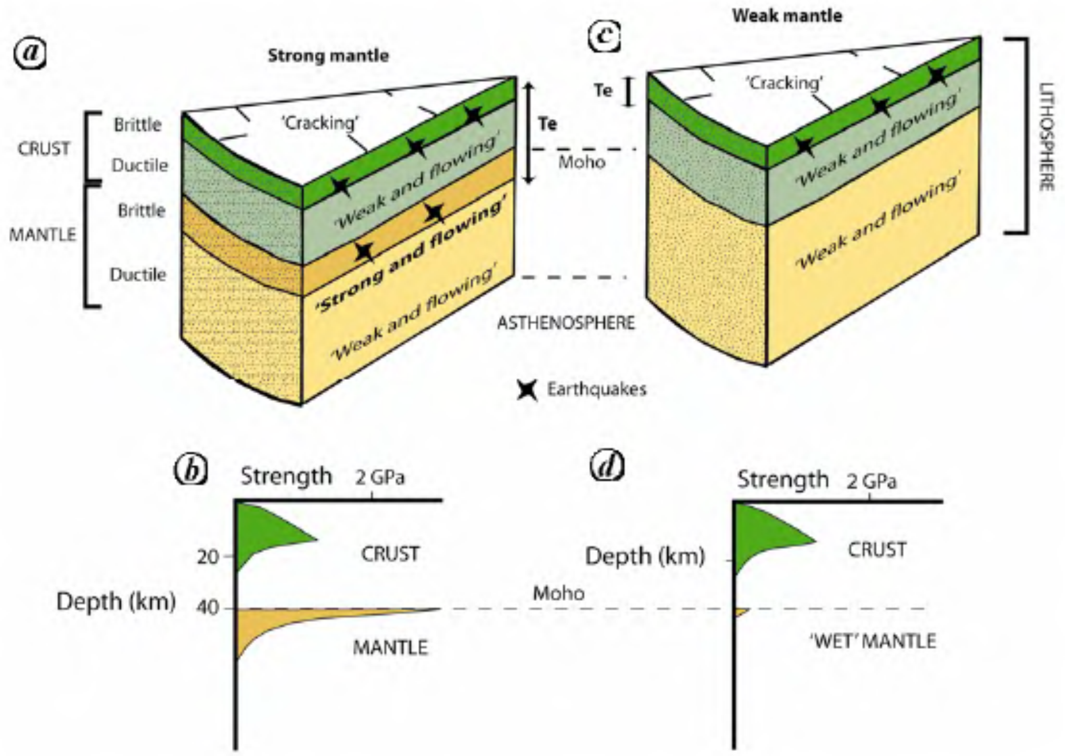

Figure 13. Possible strength profiles and rheologies for the continental lithosphere. The actual rheology well depend on both the composition of the crust and the thermal history of the lithosphere. In (***a***) and (***b***), the strength of the lithosphere lies in both the crust and mantle, with a strong brittle crust and uppermost mantle – here earthquakes are likely to occur. Elsewhere, in the lower crust and mantle, the rocks are ductile and much weaker, flowing during continental deformation. In (***c***) and (***d***), the strength of the lithosphere lies in the brittle crust, and below this the rocks are weak and ductile. The plots in (***b***) and (***d***) show plausible strength profiles with depth, for a wet or dry mantle rheology (modified from ref. 51).

the behaviour of the deforming lithosphere is block-like, but the blocks are relatively small and 'thin', comprising only the brittle crust.

In some cases, the whole thickness of the lithosphere may be too strong to deform, forming 'thick' lithospheric blocks, with deformation focused in weak zones at their margins – in effect, this is plate tectonics on a scale of 10s to 100s km, and much smaller than the major tectonic plates that extend for many thousands of kilometres. For example, the patterns of active faulting in western North America, Alaska or parts of the Alpine–Himalayan belt have been interpreted as a number of translating and rotating lithospheric blocks[19,57,58] (Figure 10 *c*). However, it is difficult to see how this style of deformation could result in regional crustal thickening, resulting in a large-scale mountain belt, although local block interaction could create local uplift.

In any case, the strength of the lithosphere will change during mountain building itself, both as a consequence of the thermal evolution of the deforming lithosphere and changes in thickness of the crustal and mantle components (with their different rheologies/strengths). For example, if the lower crust is weak enough, perhaps because it has become hot through radiogenic heating in a thickened crust, it may flow laterally, effectively forced horizontally in a channel by the high topographically induced pressures beneath the high parts of a mountain belt, tending to even out lateral variations in crustal thickness[46,59]. Searle *et al.*[16] have proposed that such heating was shallow enough to have triggered middle crustal flow and ductile extrusion of high-grade metamorphic rocks from beneath the Tibetan plateau southward towards the Greater and Lesser Himalaya.

A weak lower crust effectively decouples the brittle uppermost crust from the underlying lower lithospheric deformation, acting as a sort of spirit level for the high plateau[46,60] (see further on). In this case, the lateral support required to hold up the high plateau presumably comes from the increased strength of the lithosphere at the margins of the region of weak lower crust – reflected, for example, in the higher values of elastic thickness (T_e) at the margins of the Altiplano in the Central Andes, compared to that in the Altiplano itself[36] (Figure 8 *a*). Evidence for crustal extension on normal faults in Tibet and some parts of the high Andes[22,61] suggests that locally the margins, especially across the length of the mountain ranges, may not be strong enough to hold up the plateau.

The role of lithospheric detachment

Whatever the mechanism for uplift of the Earth's surface, it must involve a change in the vertical density structure

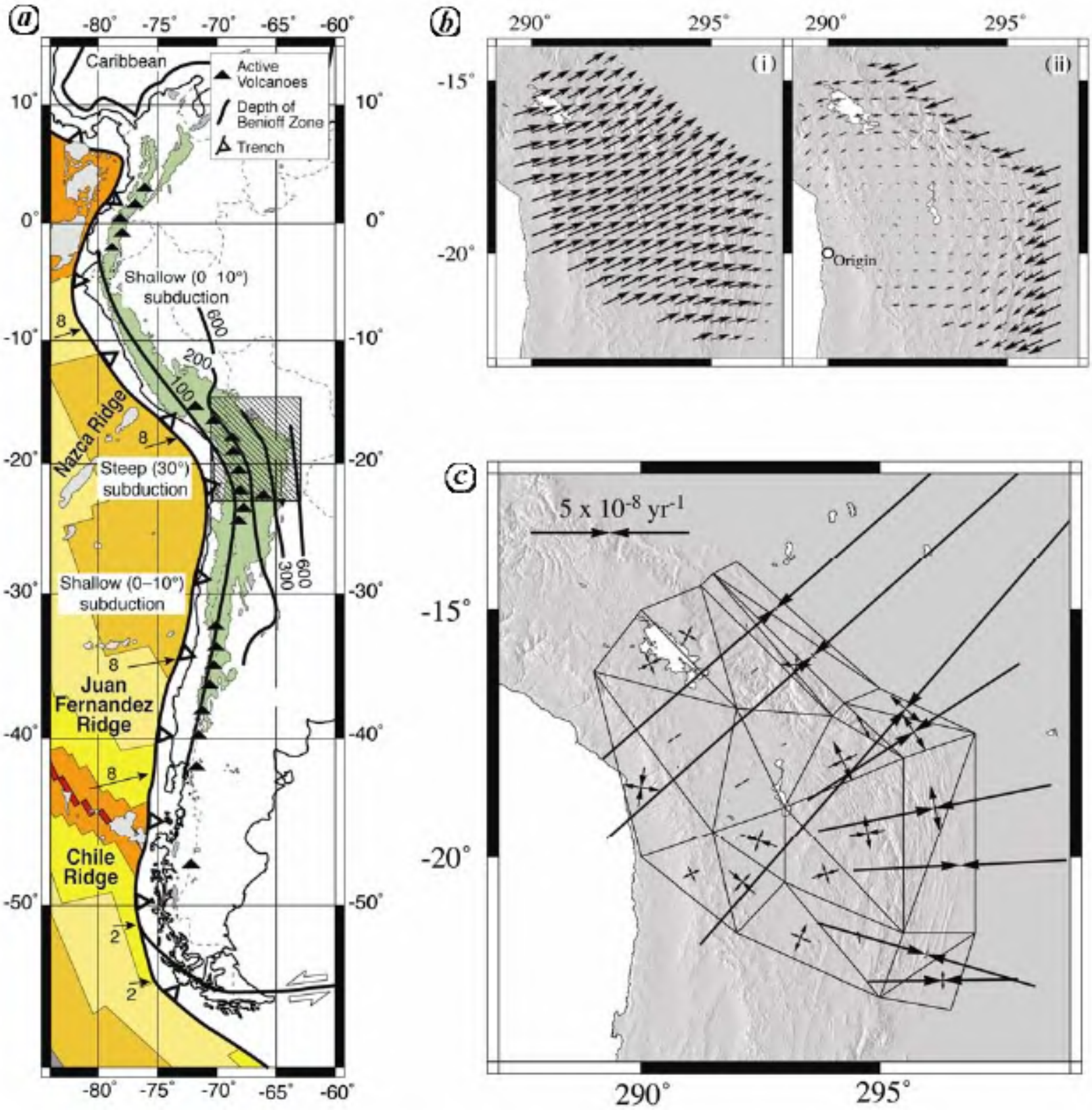

Figure 14. The horizontal surface motion of the crust in mountain ranges like the Andes can be measured over periods of a few years using the global positioning system (GPS) and the pattern of active faulting, revealing a pattern of large-scale flow described by a velocity field. ***a***, Map showing the plate tectonic setting and main features of the Andes, along the western margin of South America. Shaded box shows location of maps in (***a***) and (***b***). In (***b***), a long-term velocity field (over 10,000s years) for the Andes of Bolivia and northern Chile, in two frames of reference: (i) relative to stable South America, (ii) relative to an origin on western margin of South America, but not rotating with respect to South America. Velocity field has been constructed from a combination of GPS measurements and estimates of neotectonic fault displacements[55]. Here, in the hinge of the Bolivian orocline, velocities ~ 15 mm/yr on the western margin are essentially parallel to the plate convergence vector, but farther east they tend to swing round to more nearly orthogonal to the trend of the mountains, with a marked reduction in rate that reflects the intense crustal shortening on the eastern margin in the Subandes. ***c***, Map showing the pattern of crustal strain in the Andes of Bolivia and Chile, based on the velocity field in (***b***) (after ref. 55). This clearly shows the highest strain rates on the eastern margin, with the direction of shortening more nearly orthogonal to the range front and trend of the mountains. The overall pattern of crustal strain rate is similar to that expected if the deforming zone was behaving on a large scale like a flowing fluid, strongly influenced by the force of gravity[55].

of the Earth, so that isostasy – either local or regional – is maintained. Because the Moho is the main density contrast in the outer part of the Earth, we have focused so far on the effects of crustal shortening and thickening. In a fluid view of continental deformation, the GPS constrained flow of the strongest parts of the deforming continental lithosphere, mainly in the mantle, transmits the force to thicken the crust and push up mountains: the timescale for uplift is the same as the timescale for crustal thickening, over tens of millions of years.

There are other potential density contrasts in the lithosphere – more akin to Pratt's, rather than Airy's original idea for mountains – that could play a role in isostasy and the uplift of mountain belts. For example, mineralogical changes in the lower crust during crustal thickening, such as the transformation of basalt to high density eclogite – when mafic crust reaches depths greater than 40 km – could counteract a significant part of the uplift due to crustal thickening, in addition to the effects of thickening of the higher density mantle part of the lithosphere (Figures 10 and 15). In effect, the eclogized lower crust and thickened mantle are acting as an anchor, holding down the more buoyant thickened crust (Figure 15 *c*). This immediately suggests that if this anchor could be somehow be removed, the mountains would bob up, resulting in uplift of the mountains independent of crustal shortening and thickening[26] (Figure 15 *c*).

Theoretical and numerical modelling suggests that during lithospheric shortening, there is tendency for this underlying root of high density eclogized lower crust and thickened mantle lithosphere to start growing at a faster rate than the overall shortening, forming a blob that becomes more and more gravitationally unstable, until it eventually detaches and sinks into the surrounding asthenosphere[20,26] (Figure 16 *c*). This process, referred to as lithospheric detachment, will result in large and rapid surface elevation changes up to a few kilometres, on a timescale of a few million years or less and much faster than that due to crustal thickening alone.

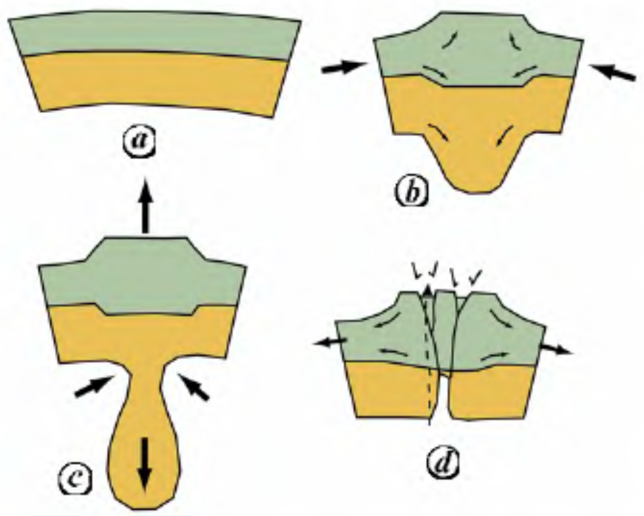

Figure 15. Theoretical models of mountain building suggest that if the underlying mantle in the lithosphere is thickened too, it will form a cold dense root that acts as an anchor holding the mountains down, illustrated in (***a***)–(***b***). If the mantle root becomes too thick (***c***), it may become gravitationally unstable, detaching itself and sinking into the underlying asthenosphere. In response, the mountains bob up. In addition, (***d***) volcanic activity may be triggered by the inflow of hot asthenosphere. Ultimately the mountains will collapse (***d***) if they rise higher than can be supported by the horizontal forces in the adjacent lithosphere. All these processes have been suggested for the uplift of the Tibetan plateau[27] and Altiplano in the Central Andes[66]. However, there is evidence that crustal thickening is a better explanation for uplift in both these regions, and lithospheric detachment, if it occurs at all, plays only a secondary role.

There are a number of presumed consequences of this detachment, discussed by England and Houseman[62], and Molnar *et al.*[27], such as the likelihood of the onset of extension when the gravitational potential energy in the uplifted lithosphere exceeds that in the surrounding regions, and also mantle and possibly crustal melting, triggered by the inflow of hot asthenosphere to replace the detached lithosphere (Figure 15 *d*).

Lithospheric detachment beneath the Altiplano and Tibet?

This mechanism was first proposed to explain rapid uplift, extension and volcanism at about 8 Ma, resulting in a thin lithosphere beneath the Tibetan Plateau[27]. Subsequently, it has been shown that this region has been high for most of the Neogene, and possibly before that[63,64], and recent seismological studies suggest that the lithosphere here may be unusually thick, extending to depths of 200 km (ref. 12) (Figure 7 *b*). In this case, catastrophic lithospheric detachment is unlikely to be an important process in at least the Neogene evolution of this region. Attention has now therefore turned to the second highest mountain belt, in the Central Andes, which offers another opportunity to evaluate the relative roles of both crustal thickening and lithospheric detachment as mechanisms for creating their present high ~ 4 km surface elevation.

Ghosh *et al.*[65] and Garzione *et al.*[66] used oxygen and carbon isotope data on carbonate paleosols in the high Bolivian Altiplano to suggest that the Altiplano had risen about 2.5 km in a brief period between about 10 Ma and 6 Ma. They inferred that crustal shortening was too slow in this region to explain such rapid uplift, and proposed instead catastrophic detachment of eclogized lower crust and mantle lithosphere. However, there is evidence that crustal shortening and an Airy-type thickening on the appropriate timescale can account for all or most of this uplift, and lithospheric detachment as a major mechanism of uplift is not required[24,25,44] (Figure 8, S. H. Lamb, manuscript in preparation, 2010).

Nonetheless, the lithosphere beneath the Altiplano in the Central Andes does indeed appear to be unusually thin (Figure 8). However, if this thinning occurred prior to significant lithospheric and crustal thickening, it would not cause large-scale surface uplift > 1 km. Thus, the magmatic evolution of the Altiplano strongly suggests that thinning of the lithosphere occurred here relatively early on in Andean evolution, at ~ 25 Ma, associated with steepening of the subducted plate and the re-establishment of the volcanic arc[25]. At this time, mafic–felsic volcanism abruptly flared up right across the width of the high Andes, with evidence for only a modest uplift < 1 km (ref. 25) (Figure 8 *b*).

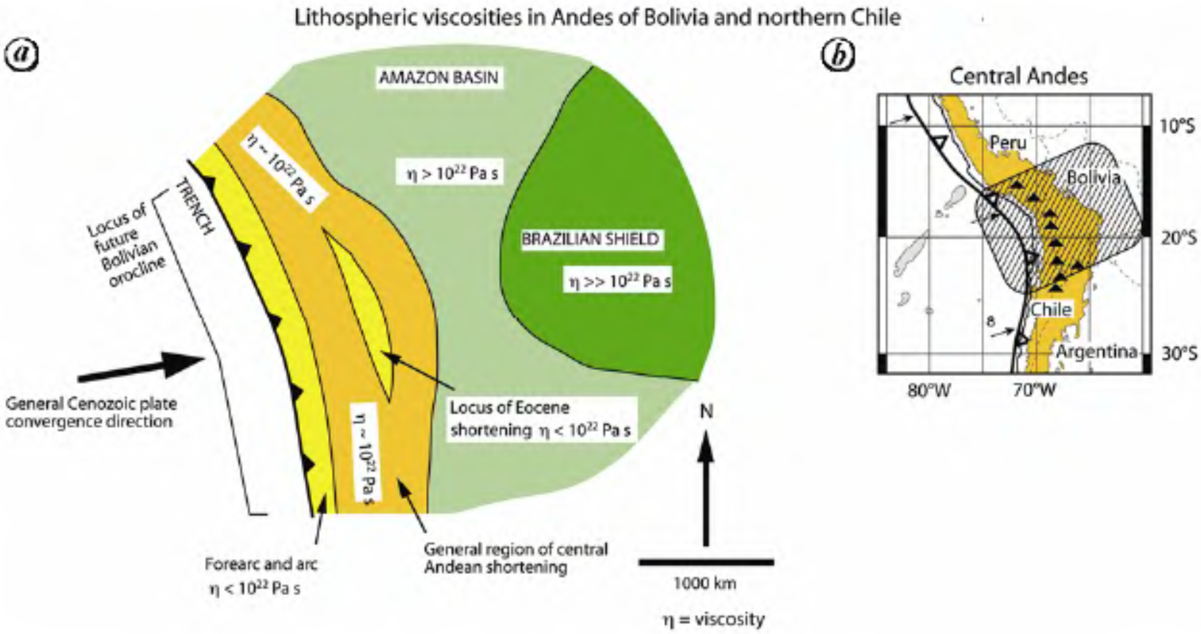

Figure 16. *a*, The postulated distribution of bulk lithospheric viscosity in the vicinity of the Bolivian Andes, as it might have been in the early Cenozoic before significant shortening in the Bolivian Andes commenced (region today in the obliquely shaded box in (*b*), after ref. 76). The map is based on considerations of the dynamical controls of the locus and style of Cenozoic deformation in this region, including the history of tectonic rotation (after ref. 76). Lower lithospheric viscosities in the west ($\sim 10^{22}$ Pa s) may have been responsible for focusing Andean deformation into an elongate zone on the western margin of South America. Viscosities are assumed to have been particularly low ($< 10^{22}$ Pa s) beneath the active volcanic arc at this time, and in a narrow zone farther east, where Eocene behind-arc Andean deformation was initiated. Viscosities in the east, beneath the foreland and Brazilian Shield, are assumed to be significantly higher ($> 10^{22}$ Pa s), resulting in an effective rigid eastern margin of the deforming zone, which has been progressively underthrust beneath the Bolivian Andes in the last 10 Ma (see Figure 8 *a*).

Though plausible, lithospheric detachment as a mechanism for uplift of large mountain belts like the Andes or the great ranges of Central Asia seems at most to play a second order role in comparison to crustal thickening. Perhaps a long history of mantle melting in these regions tends to reduce the density of the underlying mantle part of the lithosphere, inhibiting the development of gravitational instabilities during lithospheric thickening[12]. Alternatively, the mantle root may be too weak to develop long-term instabilities, but is continually detaching during lithospheric thickening, so that discrete large-scale uplift events (> 1 km) never occur[25]. However, this mechanism may still be significant in some convergent plate boundary zones (e.g. New Zealand and western North America), driving rapid uplift of several hundred metres during periods of a million years or less[7].

Why are some mountain ranges higher and wider than others?

Mountains will form wherever the continental lithosphere is sufficiently weak that it cannot support the available horizontal compressive forces in the lithosphere without deforming. In this case, one might anticipate that the initial locus of mountain building will always be at subducting plate margins. Here, the process of subduction carries water down into the mantle, and then releases it into the overlying lithosphere, thereby weakening it[11,13]. Shear stresses set up in the mantle wedge between the subducted and overriding plates can provide the forces to drag portions of the mantle lithosphere away, allowing hot asthenosphere to flow in, thereby thermally weakening the lithosphere further. Crustal thickening can also weaken the lithosphere, because it changes the thermal structure. In particular, the mantle is at greater depths, and will thermally evolve to higher temperatures, and the thickness of heat producing radiogenic crust increases.

Squeezing in the weakened lithosphere thickens the crust, raising the surface of the Earth as required by the principle of isostasy. How wide and high the mountain range becomes depends on both the time over which the horizontal forces have been acting, the magnitude of those forces, and the size of the region of relatively weak lithosphere. In addition, as the crust thickens, its weight increases, resisting further squeezing. Eventually it reaches the maximum elevation that can be supported by the available horizontal forces (Figure 11), and so deformation is focused into neighbouring and lower regions.

This way, mountain belts tend to get wider as deformation progresses, though the development of a high plateau, such as the Altiplano in the Andes or the high plateau in Tibet, seems to require a very weak crustal layer, capable of flowing laterally and leveling out the surface[24,46] (Figure 8 *a*). As this deformation impinges on stronger lithosphere of the mountain flanks, the mode of deformation will switch from thick-skinned to thin-skinned, where the strong lithosphere is pushed beneath the rising mountains, as is happening today in the Himalayas and eastern margin of the Central Andes (Figures 7, 8, 12 and 16).

The Andes, Himalayas and Tibet have had about 50 million years of mountain building to get to their present size, though significant uplift has occurred in the last 10 Ma (refs 42, 44, 64, 66). Other more youthful mountain ranges, like New Zealand's Southern Alps are mere children in comparison, reaching the relatively narrow width of a few tens of kilometres in the last five million years (Figure 9). But what is striking is that the general average elevation of mountain ranges around the world is in the range 1–2 km, whereas the Central Andes and Tibet are much higher, with average elevations between 4 and 5.5 km above sea level[67] (Figure 2). These heights must reflect the local horizontal forces available for mountain building[67].

The origin of driving forces for mountain building

Ultimately the forces that drive mountain building are the same as those that drive plate tectonics, and are gravitational forces, mainly resulting from the negative buoyancy of subducted slabs or the buoyancy of spreading ridges, though buoyancy forces related to mantle plumes, or other density contrasts in the lithosphere could play a role. These forces have to be transmitted across plate boundaries. So, for example, at subducting plate margins, the force needed to push up mountains is ultimately limited by the force that can be transmitted across the plate interface along the subduction megathrust[67] (Figure 11).

Many factors can influence this force. Lamb and Davis[68] suggested that the degree of lubrication, controlled by the presence of thick water saturated trench fill, may be responsible for the marked variations in height along the Andean margin, with no or very little lubrication in the vicinity of the Central Andes where the mountains are highest[67,68] (Figure 11, see below). However, the normal stresses on the plate interface may be important too, so that the rate of overriding of the South American plate[20] or the tendency for roll-back of the subducting Nazca plate[69], which both determine the normal stresses on the plate interface, could play an even bigger role.

The Indian–Asian collision zone, in the Himalayas and Tibet, has, of course, the highest and widest mountain ranges of all. The buoyancy of continental lithosphere requires almost all the relative motion between India and Asia to be absorbed by deformation, with a large amount of squeezing and crustal thickening. Thus, zones of continental collision are in some senses more efficient sites of mountain building; in a subduction zone like the Andean margin, more than 80% of the relative plate motion is absorbed by subduction, as the oceanic plate slips into the mantle, and less than 20% is actually taken up by lithospheric shortening in the overlying plate. However, this does necessarily mean that the forces available for mountain building are significantly less.

Thus, the megathrust in the subduction zone along the Central Andean margin, in Peru and northern Chile, is one of the strongest faults known on the planet, with an average shear stress of ~ 40 MPa (ref. 67), providing the necessary support to hold up the Central Andes where they are rising up from the depths of an oceanic trench at about 7 km below sea level. In addition, the stresses on the major faults in the foothills of the Andes are not much less than those in the foothills of the Himalayas, with both in the range 10–20 MPa (refs 54, 67).

Climate and mountain building

Mountains exert a major influence on weather, deflecting air masses and controlling regional and possibly global patterns of rainfall[70]. In addition, mountain belts are the main places of surface erosion, driven by rivers or glaciers. The eroded rock reacts with water and carbon dioxide in the atmosphere to form carbonate minerals, effectively sucking CO_2 out of the atmosphere. CO_2 is an important greenhouse gas, and so mountain building can affect the planet's climate as well. In fact, the long-term cooling of the climate since about 50 million years is associated with a net draw down of CO_2 from the atmosphere and coincides with erosion and uplift of the great ranges of Asia and Andes[64,71].

The eroded rock (and carbonate) is eventually carried to sea where it accumulates offshore as vast submarine fans. Thus, erosion of the Himalayas has created an apron of sediment in the Bay of Bengal, several kilometres thick and extending into the Indian Ocean for hundreds of kilometres. Since the late Miocene, much of the eroded detritus from the Andes has been diverted from the Pacific side of South America and transported by the transcontinental Amazon River, to be deposited in the Atlantic. However, it is important to remember that because there is a large crustal root beneath the mountains, it requires the removal of a layer of rock 6–8 km thick to lower the general elevation of the mountains by only 1 km – in other words, it requires tens of kilometres of erosion to reduce elevations back to sea level, and remove the mountain range all together. This way, mountain building forms an important part of the rock cycle near the surface of the Earth. Eventually, when there is a change in the motion of the tectonic plates, the eroded detritus may be pushed up to form new mountains.

The erosion of mountain belts also has profound effects on the creation of the mountains themselves. Thus, if erosion is sufficiently rapid, such as in New Zealand's Southern Alps, where erosion rates are as high as 10 mm/yr (equivalent to 10 km per Myr), the growth of the mountains is limited because most of the uplifted rock is eroded and deposited offshore. This way, rocks from deep levels in the crust can be rapidly exhumed at the surface in only a few million years of geological time – rather than being pushed down to form deep crustal roots – and the rapid rock uplift raises isotherms, thereby thermally weakening the crust and focusing deformation[46,60,72]. In addition, erosion modifies the overall topography of the mountain belt at its edges, often resulting in local steep topographic gradients. Thus, climate and erosion can influence the overall polarity and asymmetry of deformation across a mountain range, helping to explain why deformation in New Zealand's Southern Alps, Central Andes, or Tibet and Himalayas, is predominantly focused on the wetter side[72] (Figure 8 *a*).

In the Himalayas and eastern margin of the Andes, some of the eroded rock is deposited in the adjacent foreland basin, and is eventually incorporated into the toe of the mountain belt as deformation propagates towards the continental interior. Along the western margin of South America (Figure 14 *a*), some of the eroded material from the Andes is transported into the deep sea trench in the subduction zone, and is eventually subducted. Lamb and Davis[68] suggested that this process carries water deep into the subduction zone, lubricating the plate interface and ultimately limiting the forces available for mountain building.

Unlike the southern and northern parts of the Andean margin, the Atacama region in northern Chile is so dry that virtually no sediment is transported into the trench. It is plausible that these special climatic conditions have played an important role in the creation of the Central Andes themselves, because here, the plate interface is deprived of potential water-rich sediment lubrication, and so is rougher with lower fluid pressures, transmitting significantly more stress to push up and support the higher mountains of the Central Andes, compared to farther north and south[67,68] (Figure 8 *a*).

Final thoughts

The previous discussion highlights the central role that the origin of mountain belts has played in our understanding of the way the Earth works, and particularly the behaviour of the lithosphere. The basic conclusions can be found in even the earliest observations of mountain belts, from their apparent lack of mass – suggested by the early gravity measurements – to the abundant geological field evidence that we now have for the intensity of their deformation.

An intriguing final insight of all this is that the central highly deformed parts of mountain belts, by being such weak and mobile parts of the Earth, may be the places where the strong cratonic cores of the continents were first formed, comprising what are today the most stable parts of the dry land we live on[12]. This is because the process of mountain building, by squeezing both the crust and mantle parts of the lithosphere, creates a thick lithosphere. Over time, as geotherms relax and the crust heats up as a result of the increased radiogenic heat generation in the thickened crust, granulite grade metamorphism will occur, eventually dehydrating and further strengthening the crust[73].

If, at some later stage, the crust in this thick lithosphere is eroded back down to its original thickness of around 30–40 km, as isostasy would predict, the land surface will return to around sea level, but with the deep crustal levels of granulite grade metamorphic basement now exposed at the surface[12,73]. So, as has been long suspected by geologists[74], mountain building, although occurring in only a small fraction of the surface area of the continents at any one time, might have shaped most of the Earth's continental crust.

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