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Subparallel thrust and normal faulting in Albania and the roles of gravitational potential energy and rheology contrasts in mountain belts

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[1] The active tectonics of Albania and surrounding regions, on the eastern margin of the Adriatic Sea, is characterized by subparallel thrust and normal faulting which, we suggest, is likely to be related to gravitational potential energy contrasts between the low-lying Adriatic Sea and the elevated mountainous areas inland. We calculate the magnitude of the force which the mountains and lowlands exert upon each other as a result of this potential energy contrast. It is likely that this force is largely supported by shear stresses on faults, and if so, the average stresses are less than ~20 MPa. Alternatively, if the mountains are supported by stresses in the ductile part of the lithosphere, the stresses are likely to be ~80–240 MPa in magnitude. The mountains of Albania are significantly lower than other ranges, such as the Peruvian Andes, which are thought to be extending in response to potential energy differences, and we discuss the relation between Albania and these other, higher, mountain belts from the perspective of differences in lithosphere rheology. We suggest that the lowlands of western Albania and the Adriatic Sea may have been weakened through time as a result of the deposition of large thicknesses of sediment, which lead to heating of the crystalline basement, a reduction in the potential energy contrast that could be supported by the lowlands, and so normal faulting in the mountains of eastern Albania.

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1. Introduction

[2] In this paper we describe the active deformation in Albania and surrounding regions, on the eastern margin of the Adriatic Sea (Figure 1). The focal mechanisms of earthquakes indicate that ~E-W extension is occurring in eastern Albania. This faulting is spatially distinct from, and almost perpendicular in direction to, the extension in northern and western Greece and throughout the Aegean [*Goldsworthy et al.*, 2002] and is instead part of a pattern of subparallel and spatially separated thrust and normal faulting. Similar patterns of faulting have previously been recognized in other mountain belts, such as the Peruvian Andes, and allow constraints to be placed upon the stresses required to deform the lithosphere [e.g., *Dalmayrac and Molnar*, 1981]. The strength of active faults, and of the lithosphere as a whole, is a subject of much debate [e.g., *Scholz*, 2000; *Jackson et al.*, 2008], and in this paper we seek to use the active deformation in Albania to place some constraints on the rheology of the lithosphere in this area. Other orogenic belts which are thought to be extending in response to contrasts in gravitational potential energy are significantly higher than the mountains of Albania. We suspect that variations in the maximum heights of mountains that can be supported are related to variations in the strengths of their neighboring forelands. This study of Albania is therefore related to a wider debate on the controls of lithosphere rheology, and particularly to what is responsible for the contrasts that are seen between the ancient Precambrian cratons and the younger Phanerozoic orogenic belts [*Jackson et al.*, 2008]. Contrasts in lithosphere rheology and structure control many of the first-order characteristics of continental tectonic geology [e.g., *McKenzie and Priestley*, 2007], and it is now clear that a single generic model for continental lithospheric rheology is inappropriate and does not account for important regional variations [*Jackson et al.*, 2008].

[3] The geology of Albania is described in detail by *Meco and Aliaj* [2000] and *Robertson and Shallo* [2000], and only briefly summarized here. In the area of the "External Albanides" (SW of the dashed line on Figure 1b) the exposed rocks are neritic and pelagic carbonates formed on shallow platforms and in oceanic basins during the Mesozoic and Cenozoic. The oldest rocks exposed at the

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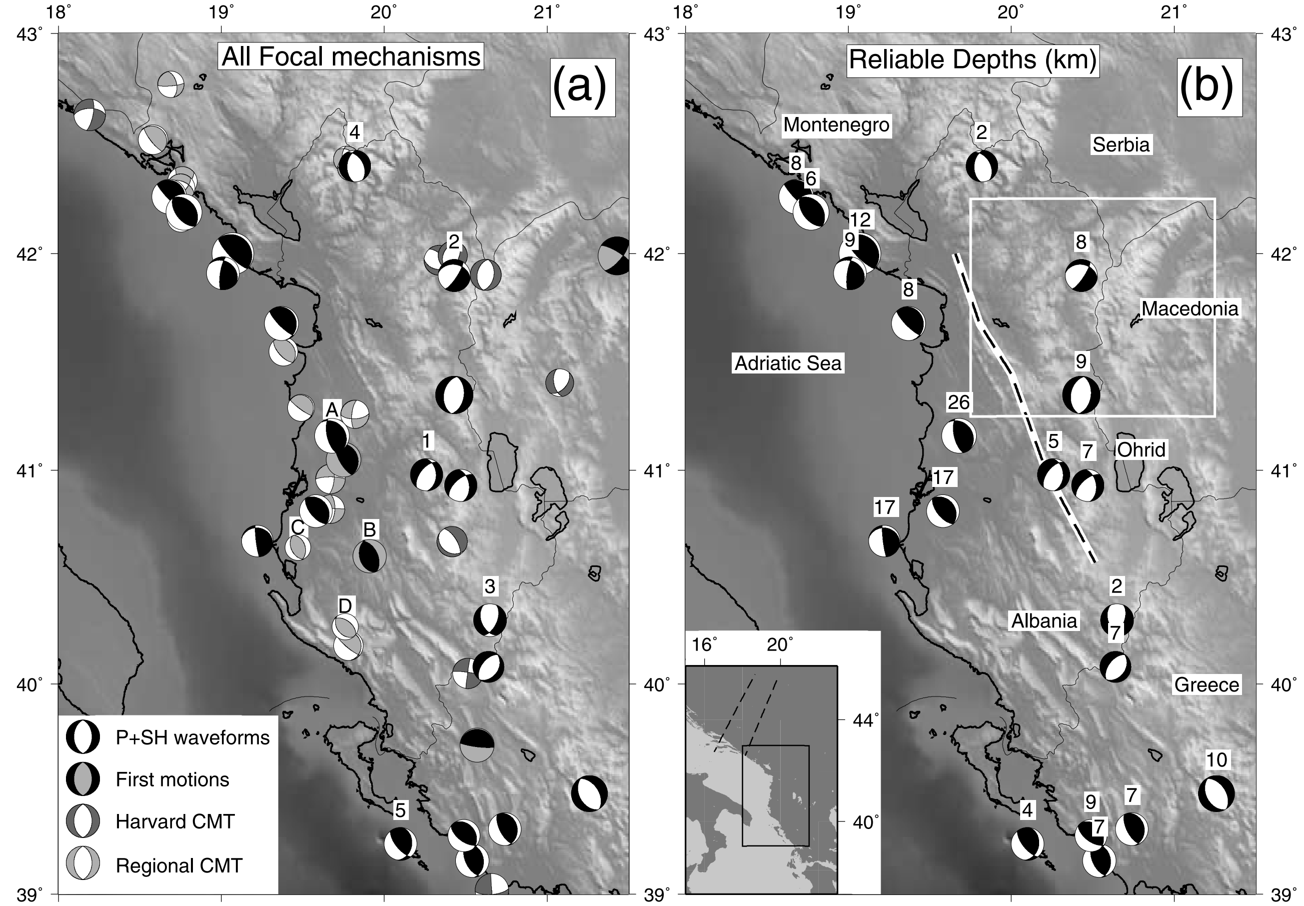

Figure 1. Active faulting in Albania and surrounding regions. The area of coverage of Figures 1a and 1b is shown by the box in the inset of Figure 1b, which also shows the location of the seismic surveys by *Dragasevic and Andric* [1968] as dashed lines. (a) Focal mechanisms in black and white are constrained by P and SH body wave modeling [*Baker et al.*, 1997; *Louvari et al.*, 2001; this study] (see Table 1 and Text S1). Those with quadrants in black (compressional) and grey (extensional) are from observations of first-motion polarities [*McKenzie*, 1972; *Anderson and Jackson*, 1987]; those shown as dark grey and white are from the Harvard CMT catalogue, and pale grey and white mechanisms are the regional CMT solutions of *Pondrelli et al.* [2006]. The earthquakes labeled with letters (above the focal mechanism) are discussed in detail in the text, and those labeled with numbers have mechanisms constrained by waveform modeling, as part of this study (Table 1). (b) The reliable depths (in km) of earthquakes which have been obtained by the modeling of teleseismic P and SH waveforms. The dashed black line shows the approximate boundary between the Internal Albanides (to the northeast) and the External Albanides (to the southwest). The white box shows the area of coverage of Figure 3.

surface are Permo-Triassic evaporites, which have acted as a detachment layer during recent folding and thrusting [*Velaj et al.*, 1999]. Flysch deposition and folding occurred from the Eocene onward, and the surface morphology is characterized by folds that are thought to be cored by the Permo-Triassic evaporites [*Velaj et al.*, 1999], which are also seen to outcrop at the surface as diapiric structures. The rocks of the "Internal Albanides" (NE of the dashed line on Figure 1b) are generally older, and are largely composed of Jurassic ophiolites and Mesozoic carbonates. These rocks are thought to have been accreted to the margin of Eurasia by the end of the Mesozoic. The Adriatic Sea is composed of continental crust and a thick (~15 km [*Patacca et al.*, 2008]) sequence of Mesozoic and Cenozoic sedimentary rocks; the absence of seismicity suggests it is relatively undeforming in this southern part (Figure 1). The Adriatic is thought to have been a "promontory" of the African plate in the Mesozoic and much of the Cenozoic [e.g., *Channell et al.*, 1979] but now appears to be moving independently [e.g., *Anderson and Jackson*, 1987; *Battaglia et al.*, 2004; *D'Agostino et al.*, 2008]. Active shortening occurs on thrust faults in the lowlands and coastal regions of western Albania and active extension, on normal faults subparallel to the thrust faults, occurs in the highlands in the east of the country (Figure 1). In section 2 we describe the active deformation occurring in these regions, and then in section 3 we discuss the likely cause of the spatial separation of

Table 1. Focal Parameters of Earthquakes Modeled in This Study[a]

Event	Year	Julian Day	Time	Latitude	Longitude	M_w	Depth (km)	Strike (deg)	Dip (deg)	Rake (deg)
1	1997	136	0700:48	40.98	20.26	5.3	5	218	57	−65
2	1998	273	2342:54	41.90	20.43	5.2	8	280	32	−27
3	2004	328	0226:16	40.30	20.65	5.4	2	23	55	−59
4	2005	191	1310:12	42.40	19.82	5.3	2	184	49	−67
5	2007	180	1809:15	39.24	20.10	5.4	4	10	32	134

[a]See Text S1, section 1, in the auxiliary material for details. Event number refers to the label for each earthquake on Figure 1a.

thrust and normal faulting and its implications for the mechanical properties of the lithosphere.

2. Active Deformation

[4] Figure 1 shows the earthquakes which have occurred in Albania and surrounding regions during the modern instrumental period (the oldest event shown occurred in 1963). We have used waveform analysis to constrain the source parameters of the five earthquakes labeled with numbers in Figure 1a (Table 1), by inverting long-period P and SH waveforms for the strike, dip, rake, seismic moment, centroid depth, and source time function, using the MT5 version [*Zwick et al.*, 1995] of the algorithm of *McCaffrey and Abers* [1988] and *McCaffrey et al.* [1991]. A detailed description of the method is given by *Nabelek* [1984] and *Taymaz et al.* [1991], and the procedure is now too routine to justify a full explanation here. The results of the waveform inversions are shown in detail in Text S1, section 1, in the auxiliary material.[1] Tests of the type described by *Taymaz et al.* [1991] show that errors in fitting the observed and synthetic seismograms, and uncertainty in the velocity structure, yield errors of about 4 km or less for the centroid depths. Sections 2.1 and 2.2 will describe some details of the active deformation, beginning with the thrust faulting, and following with the normal faults.

2.1. Thrust Faulting

[5] Thrust faulting occurs in the lowlands of western Albania and the Adriatic Sea, and is part of a continuous band which extends from the Alps southward, down the entire eastern margin of the Adriatic sea, to connect with the Hellenic subduction zone south of ~39°N. The earthquakes are mostly low-angle thrusting events, although some events have occurred on steeper-dipping planes (Figure 1). The waveforms of some of the earthquakes on low-angle fault planes have been modeled, and the reliable centroid depths are between 17 km and 26 km in central Albania, although some as shallow as 6 km have occurred along the coasts of northern Albania and Montenegro (Figure 1b). *Louvari et al.* [2001] determined a centroid depth of 30 km for the deepest event in the area (event A on Figure 1a), although they used some narrow-band long-period seismograms when performing their waveform inversions, which generally have poor depth resolution. Comparing observed and synthetic vertical component broadband seismograms for this event (Text S1, section 2) confirms that the centroid depth was actually ~26 km.

[6] Three earthquakes (marked B, C, D on Figure 1a) show relatively high-angle reverse faulting, at depths which are probably shallow (<10 km) but not well constrained. The focal mechanism marked B on Figure 1a represents an earthquake which occurred in 1969, for which surface faulting of an ambiguous nature was reported by *Sulstarova* [1980] and *Aliaj* [1982]. This earthquake occurred near an elongate mountain range which follows the surface trace of two closely spaced anticline axes [*War Office*, 1943], and the strike of the nodal planes is the same as that of the anticline axes. The focal mechanism marked C on Figure 1a also has relatively high-angled nodal planes, and this earthquake occurred where an anticline of Pliocene sandstones and marls, with the same strike as the earthquake nodal planes, emerges from the surrounding low-lying Quaternary sediments (Figure 2), although the depth of this earthquake is not reliably known. The geology in the area surrounding the third high-angle reverse faulting earthquake in Albania (marked D on Figure 1a) is less clear on the geological map, and the earthquake cannot be reliably linked to an individual geological structure.

[7] The thrusting earthquakes in western Albania and the Adriatic Sea can therefore be subdivided into two types. The majority have occurred on low-angle fault planes, and extend to depths of up to 26 km. A smaller number of earthquakes have occurred on higher-angle planes, and appear to be spatially correlated with anticlines visible at the surface, suggesting that the anticlines are currently active and underlain by (possibly blind) reverse faults. The geology and geomorphology of the anticlines associated with these steeper reverse faults are reminiscent of the better known Zagros mountains of southwest Iran, where large thicknesses of salt are also present [e.g., *Talebian and Jackson*, 2004; *Ramsey et al.*, 2008].

2.2. Normal Faulting

[8] The mountainous areas of eastern Albania and the adjoining regions of Greece and the Republic of Macedonia are characterized by active normal faulting (Figure 1a). The earthquakes have shallow centroid depths (<10 km, Figure 1b), and nodal planes which vary in strike between NE and NW. A number of intramontane basins have the morphology characteristic of normal fault-bounded basins (such as at Lake Ohrid, Figure 1b [e.g., *Aliaj et al.*, 2001]), suggesting that the faulting in this belt reaches the surface as it is known to do at its southern end in NW Greece [*Goldsworthy et al.*, 2002]. In places the normal faulting has had a visible effect on the large-scale drainage network, an example of which is shown in Figure 3.

3. Role of Gravitation Potential Energy Differences

3.1. Calculating Potential Energy Differences

[9] Figure 4 shows how the style of faulting varies with surface elevation within the area shown on Figure 1 north of 40°N, clearly indicating that thrust faulting is concentrated

[1]Auxiliary materials are available in the HTML. doi:10.1029/2008JB005931.

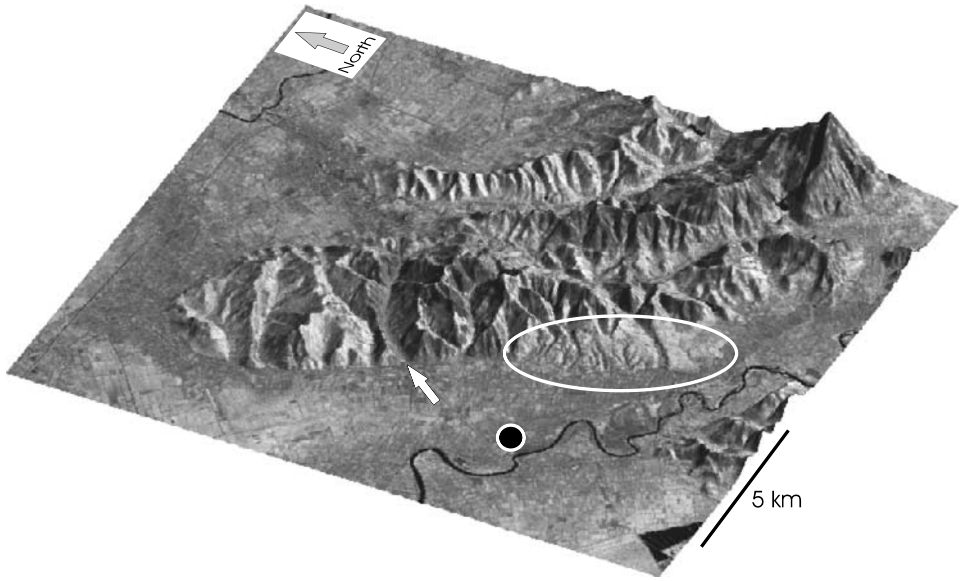

Figure 2. Perspective image (Landsat imagery overlain on SRTM topographic data) looking NE at the location of the earthquake marked C on Figure 1a. The high topography in the center of the image rises ~200 m above the level of the surrounding plains. The black circle shows the location of the epicenter of the earthquake (M_w 4.2) determined by *Pondrelli et al.* [2006], which may be in error by as much as 10 km. The white oval shows an area where incision is clearly visible, suggesting active uplift. The valley indicated by the white arrow crosses the entire width of the high topography, although the central portions are above the level of the alluvial plain surrounding the anticline and may represent an old river course which has been beheaded by uplift. The watershed lies toward the NE side of the anticline, which may suggest the underlying fault dips to the SW.

in low-lying areas while normal faulting occurs at high elevations. A number of mechanisms have been proposed to explain the presence of normal faulting within orogenic belts, such as slab roll-back, gravitationally driven deformation around the margin of an arcuate mountain range [e.g., *Copley and McKenzie*, 2007], and "traction reversal" above a rigid underthrusting and subducting layer [e.g., *Willett*, 1999]. However, the absence of active subduction beneath Albania, the linearity of the mountain range, and the clear relation between focal mechanism style and elevation leads us to suggest that the faulting is caused by differences in gravitational potential energy between the lowlands and the mountains [e.g., *Dalmayrac and Molnar*, 1981; *England and Houseman*, 1988]. As a mountain belt is formed, work is done against gravity when thickening the crust, which means that mountain ranges have higher gravitational potential energy than lowland areas [e.g., *Artyushkov*, 1973]. In such a situation, crustal shortening tends to concentrate in the lowlands on the margins of the mountain belt, and the high mountains will have a tendency to extend, as is seen in Albania. The difference in gravitational potential energy between two isostatically compensated lithospheric columns of unit area is equivalent to the vertical integral of the difference in vertical normal stress between the two areas, i.e., $\Delta\text{GPE} = \int \Delta\sigma_{zz}\, dz$, where the integral is between the surface of the mountains and the depth of isostatic compensation, σ_{zz} is the vertical normal stress ($\partial\sigma_{zz}/\partial z = \rho g$), and $\Delta\sigma_{zz}$ refers to the difference in σ_{zz} at a given depth between the two lithospheric columns being considered [e.g., *Molnar and Lyon-Caen*, 1988]. Therefore, if the surface elevation and deep structure of the mountain belt and lowland are known, it is possible to estimate the contrast in gravitational potential energy between them, and so the force which the lowlands and mountains exert upon each other as a result of the lateral differences in crustal thickness. For the purposes of the following discussion it will be assumed that the topography in Albania and surrounding regions can be approximated as two-dimensional, and we will consider topography along a cross section oriented perpendicular to the main geological structures and the nodal planes of earthquake focal mechanisms. Thrust faulting occurs in regions where the surface elevation is close to sea level (Figure 4), and the mean elevation of the mountainous region of eastern Albania is ~1 km. The Moho depth in the region is not well known. The surface wave analysis of *Marone et al.* [2003] shows Moho depths in excess of 40 km in the mountainous regions, and as low as 30 km in the Adriatic sea, although the horizontal resolution of surface wave studies means that these values are likely to be affected by the crustal thickness in the surrounding regions, in addition to the areas with which we are concerned. Northwest of the area discussed here, *Dragasevic and Andric* [1968] conducted active source seismic investigations along two profiles which extended from the Adriatic Sea across the mountains of the former Yugoslavia (dashed lines on the inset in Figure 1b). The greatest depths they estimated for the Moho were ~50 km, below the mountains, and the Moho depth was around 40 km on the coast of the Adriatic Sea. In a similar area, *van der Meijde et al.* [2003] used receiver functions to estimate Moho depths of 41 and 47 km at two sites on the margin of the Adriatic sea. Therefore, the crustal

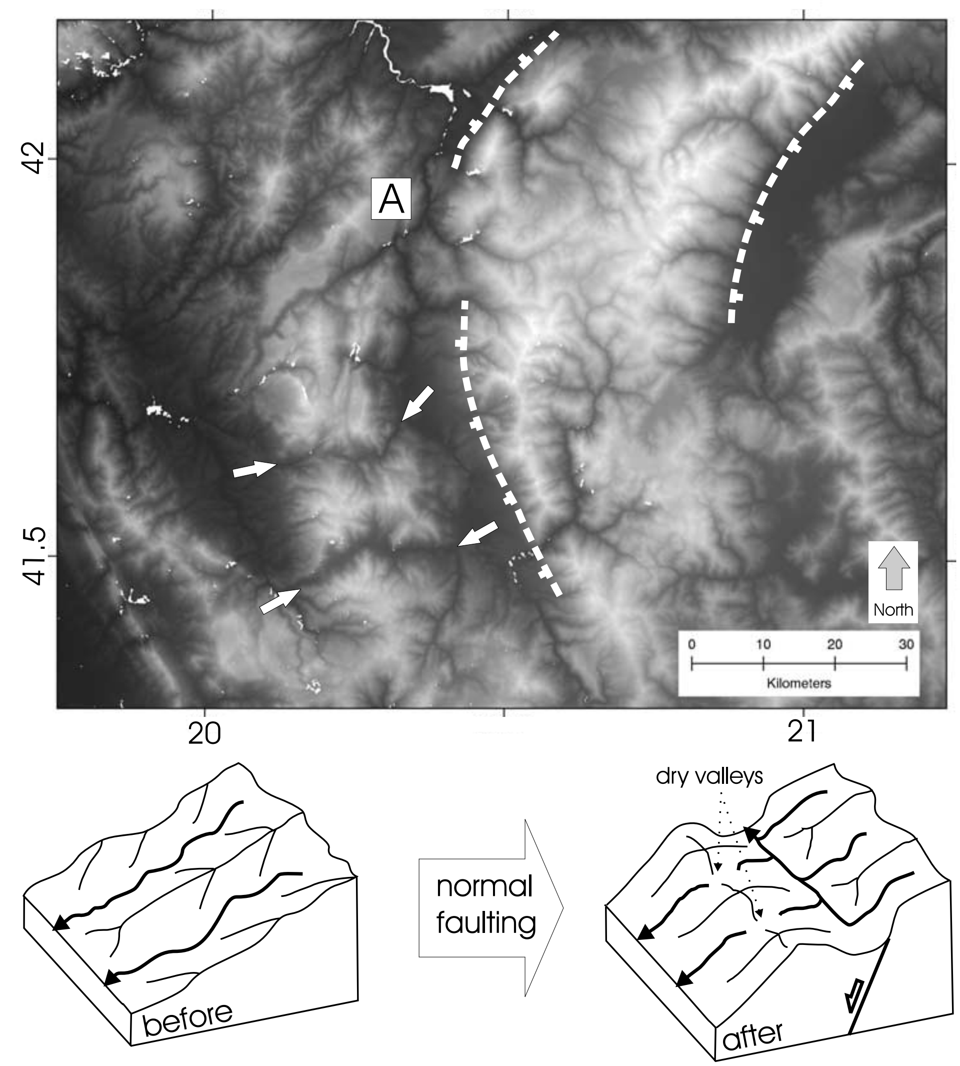

Figure 3. Topography within the white box shown on Figure 1b (pale shades represent higher elevations). The dashed white lines show the locations of large linear topographic features which have the geomorphological characteristics of active normal faults, with ticks on the downthrown side. The white arrows show the locations of two valleys which cross the high topography to the west of the southernmost normal fault. The drainage divide in these valleys is within the high topography, and so rivers flow down them to the major valleys to the east and west, and the valleys are dry at their watersheds. The hanging wall basin of the southernmost normal fault is drained by the river flowing along the valley marked A. It is probable that streams in the two valleys which cross the high topography flowed west before the normal faulting began and that motion on the southernmost normal fault, and the associated subsidence in its hanging wall, caused the drainage direction to be reversed in the eastern parts of the valleys, creating the situation we see today (as shown in the cartoons).

thicknesses are taken to be 40 km on the margins of the Adriatic sea and 50 km beneath the mountains of eastern Albania. Isostatic compensation is assumed to take place at the base of the thickest crust, and the densities of the crust and mantle are taken to be 2800 and 3300 kg m^{-3}. The vertical normal stress can then be calculated for lithospheric columns beneath the mountains and the lowlands, and by integrating over depth the difference between these stresses, the force exerted upon each other by the mountain belt and the adjacent lowlands is calculated to be $\sim 1.2 \times 10^{12}$ N per unit length along strike.

3.2. Geotherms and Rheology

[10] Having estimated the magnitude of the force resulting from the gravitational potential energy contrast, it is now worth considering which part of the lithosphere is likely to be supporting these forces. Our first step will be to calculate a geotherm for the Adriatic lithosphere. There is no evidence for recent tectonic activity, except around the margins of the Adriatic, so we have constructed a geotherm by assuming that the temperature profile is in steady state and have only considered vertical diffusion [e.g., *McKenzie et al.*, 2005]. *Lavenia* [1967] measured the heat flow at the

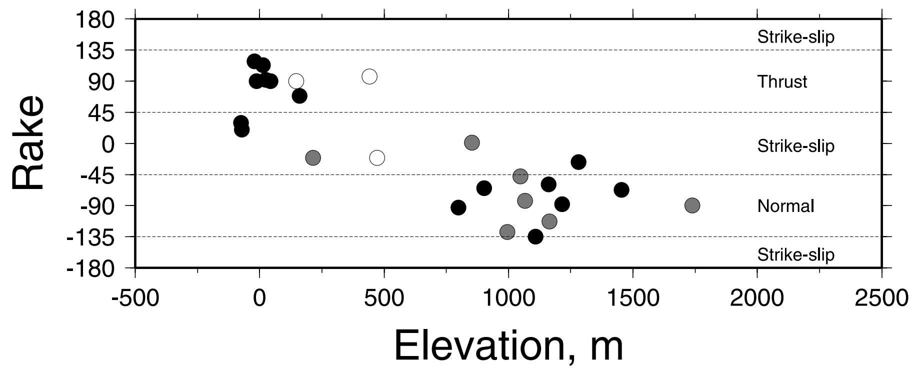

Figure 4. The relationship between surface elevation and style of faulting in the area shown on Figure 1 north of 40°N. The elevation is taken from topography smoothed with a Gaussian filter with a full width of 50 km in order to remove small-scale topographic fluctuations. The black circles represent waveform-modeled earthquakes, the grey circles are CMT solutions, and the white circles are from earthquakes with mechanisms constrained by first-motion polarities. The rake of only one nodal plane of each earthquake is plotted, but the pattern does not change appreciably whichever plane is plotted, as most of the earthquakes are close to pure dip slip or strike slip.

surface to be ~45–50 mW m^{-2}, and we fixed the lithospheric thickness to be 100 km (a likely value for unthickened, noncratonic continental lithosphere). The base of the lithosphere is constrained to have the isentropic temperature at that depth (with a potential temperature of 1315°C), and the surface is taken to be at 0°C. For the thermal conductivity of the mantle lithosphere we used the expressions of *McKenzie et al.* [2005] for the temperature-dependent thermal conductivity of olivine. The thermal conductivity of the ~15 km thick sedimentary pile was estimated by calculating the percentage pore space at a given depth using the relation $\phi = \phi_0 \exp(-z/\lambda)$, where ϕ_0 is the porosity at the surface and λ is the porosity scale height. We used values for ϕ_0 and λ of 0.5 and 2500 m for sandstones and 0.4 and 3200 m for carbonates. We used a simplified sedimentary section, having a sandstone composition for the top half and a limestone composition for the lower half. We then used the expression of *Budiansky* [1970] to calculate the conductivity of the solid-water mixture (where pore-free sandstone and limestone were taken to have conductivities of 3.7 and 3.2 W m^{-1} K^{-1}, respectively). The choice of lithologies and their depth extents have a limited effect on the shape of the resulting geotherm. The crystalline lower crust is taken to have a thermal conductivity of 2.5 W m^{-1} K^{-1} [e.g., *Jaupart and Mareschal*, 1999; *McKenzie et al.*, 2005]. The distribution of radiogenic heat production within the crust can then be adjusted to reproduce the observed surface heat flux for the input lithospheric and crustal thicknesses. A number of different distributions of heat production can match the observed surface heat flux, but the general forms of the geotherms produced are similar to that shown as the black line in Figure 5, which corresponds to a heat production of 4 μW m^{-3} for the recent flysch deposits at the top of the sedimentary column, and a value of 1.5 μW m^{-3} for the remainder of the crust. The deepest earthquake which has been recorded in the area (at 26 km, earthquake A on Figure 1a) occurred at a depth which corresponds to a temperature of ~600°C within the crust. However, this temperature should only be treated as a rough estimate, as the earthquake occurred in western Albania, in an area of crustal thickening, where the temperature structure will be altered by the competing influences of cold material being advected downward relative to the surface, and of increased radiogenic heat production because of the thickened crust. *Jackson et al.* [2004] argued that crustal rocks are only likely to be seismically active at such high temperatures if they are anhydrous; a characteristic of many Precambrian lower crustal granulites. If the Adriatic Sea is indeed relatively strong because its basement is anhydrous, that may explain why it has remained relatively undeformed as mountains have been formed in its surroundings. The geotherm has mantle temperatures at the Moho of above 600°C, so mantle earthquakes would not be expected to occur [*McKenzie et al.*, 2005], and none have been recorded. *Kruse and Royden* [1994] used the geometry of recent sediments to study flexure in the northern and central Adriatic Sea and estimated values of the effective elastic thickness which varied between 10 and 15 km. *D'Agostino and McKenzie* [1999] used profiles through gravity data to estimate similar values of 5–12 km for the effective elastic thickness. However, it should be noted that these estimates are both from fitting modeled geometry and gravity along one-dimensional profiles and in some cases, when estimating the elastic thickness using one-dimensional profiles, the upper limit on the range of estimates can be relatively poorly resolved [e.g., *Jackson et al.*, 2008]. *Maggi et al.* [2000] studied other parts of the Alpine-Himalayan belt and found that the elastic thickness was similar to, but slightly lower than, the seismogenic thickness, which they interpreted to mean that the long-term strength of the continental lithosphere was concentrated in the seismogenic layer. An alternative explanation for the same observations could be that the seismogenic layer is weak in the long term while the underlying aseismic layer holds long-term strength; which they suggest is unlikely and unnecessarily complicated. Recent and relatively unconsolidated sediments would not be expected to be able to support significant loads over geological time, and earthquakes do not appear to be generated within the sediments of the Adriatic Sea (the

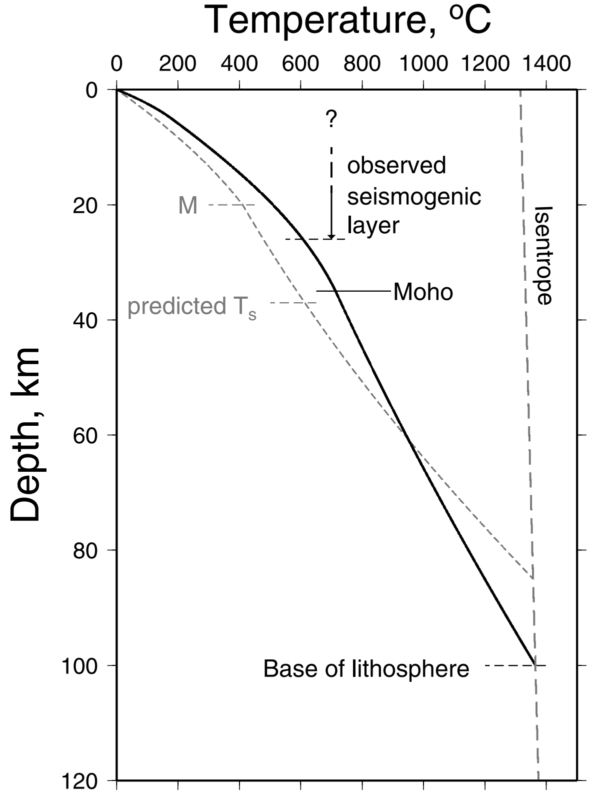

Figure 5. The black line shows a geotherm for the Adriatic Sea, calculated from the thermal parameters described in the text (section 3.2). The grey dashed line shows a geotherm calculated using the same parameters, but assuming the ~15 km thick layer of sediments had not been deposited (as discussed in section 4.5). Labeled on the grey line are the Moho (M), and the depth to the 600° isotherm, and therefore the predicted deepest limit of seismicity in the mantle in this case (predicted T_s) [*McKenzie et al.*, 2005].

only thrusting earthquake in Albania that may have produced coseismic surface faulting is earthquake B on Figure 1a, which occurred where anticlines of Mesozoic carbonates reach the surface). *Velaj et al.* [1999] suggested that the thrust faults visible on land in western Albania are mostly decollements formed within the thick sequence of Permo-Triassic evaporites, so any faults present in the upper part of the sedimentary sequence offshore are likely to be capable of supporting only minor shear stresses ($\leq$1 MPa [e.g., *Carter et al.*, 1982]) and are probably aseismic. The simplest interpretation of the seismic and elastic thickness estimates in Albania is therefore that the seismogenic basement (below the thick covering of weak sediments) is of a similar thickness to the estimates of effective elastic thickness, and is therefore likely to be the part of the lithosphere supporting long-term stresses.

3.3. Shear Stresses on Faults

[11] We have now estimated the magnitude of the forces exerted upon each other by the lowlands of western Albania and the mountains in the east of the country, and suggested that these forces are likely to be supported by the seismogenic layer. The dip of the low-angle nodal planes of the thrust faulting earthquake focal mechanisms (which are presumably the fault planes), and the depth to which the faulting extends in the Adriatic Sea and western Albania, can therefore be used to estimate the average shear stress which can be supported by the faults. We will initially use a simple model in which the mountains are considerably weaker than the lowlands, so that the force resulting from the potential energy contrast is completely supported by the lowlands, and in which the force is entirely accommodated by the shear stress on the faults, i.e., $\tau = \Delta\mathrm{GPE} \tan \theta / T_s$, where τ is the average shear stress on the faults, θ is the dip of the thrust faults, T_s is the seismogenic thickness, and ΔGPE is the gravitational potential energy difference between the two lithospheric columns and, equivalently, the force exerted upon each other by the lowlands and mountains. We assume that the seismogenic layer is 15 km thick (extending down to depths of 26 km, but not including the top 11 km of the sedimentary sequence), which implies that to support the $\sim 1.2 \times 10^{12}$ N estimated above, the average shear stress on the faults should be around 40 MPa if the faults dip at 25° (as suggested by the focal mechanism of earthquake A on Figure 1a, and other nearby earthquakes). Following the approach of *Lamb* [2006], if we resolve and balance the forces on the hanging wall of a thrust fault and take into account the weight of the rock in the hanging wall, then we can estimate the average shear stress on the fault as $\tau = 0.5\,[(F/T_s) - (0.5\rho g T_s)] \sin 2\theta$, where F is the sum of the horizontal forces due to the gravitational potential energy contrast and those due to the weight of rock in adjacent crustal columns. If this second method is used, the estimate for the average shear stress on the thrust faults reduces to ~30 MPa. The analysis presented above assumes that the normal faults in the mountains support no stresses, and that the elevation contrast is entirely supported by the faults in the lowlands. However, if the normal faults support some shear stress, the estimate of the shear stress on the thrust faults will be reduced, because the elevation of the mountains will not only be supported by the forces transmitted across the thrust faults in the lowlands, but also in part by the resistance to deformation of the normal faults in the mountains. If we further assume that the normal faults in the high mountains dip at 45° and extend from the surface to depths of 10 km (Figure 1), and support the same average shear stress as the thrust faults in the lowlands, then we can estimate the average shear stress on the faults to be ~30 MPa for the simple model and ~20 MPa for the more complex model. The results of all the calculations in this section are highly dependent on the values chosen for the seismogenic thickness and the dip of the thrust faults. The effects of changing these parameters can be seen in Figure 6. If significant stresses are also supported below or above the seismogenic layer (in the ductile crust or the overlying sediments), then the force resulting from the potential energy difference will be accommodated by a larger thickness of lithosphere, and the shear stress supported by the thrust faults will be less than the values we have estimated, which should therefore be treated as an upper limit. It therefore seems likely that, for plausible estimates of the dip of the thrust faults and the seismogenic thickness, the average shear stress on the faults is likely to be less than ~20 MPa. Our estimate for the shear stress on the faults in the Albanian lowlands is comparable to the estimate of 7–15 MPa by *Lamb* [2006] for the average

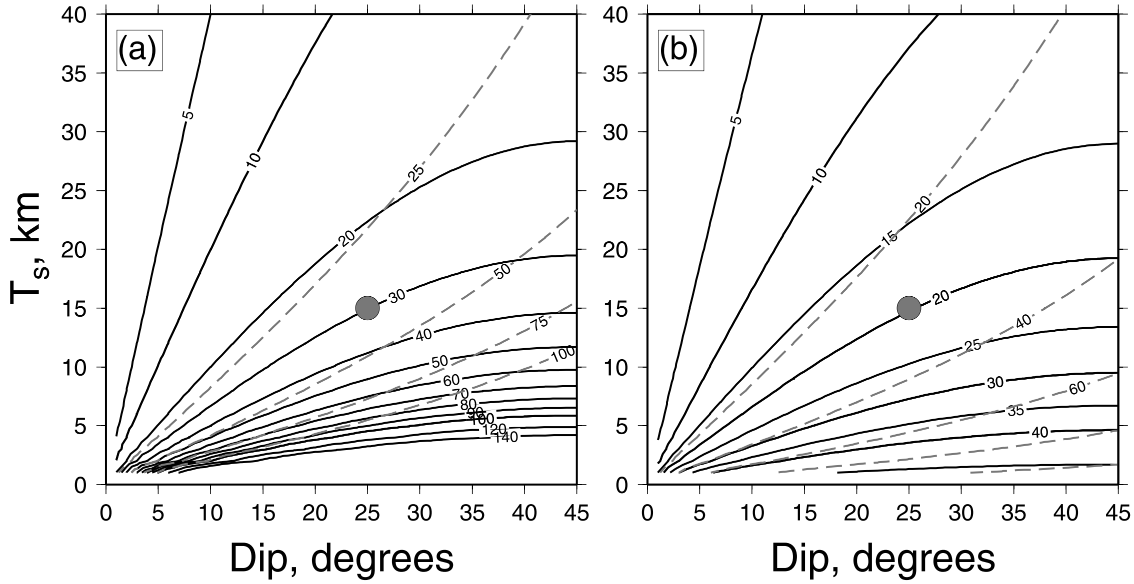

Figure 6. Graphs showing the average shear stress (in MPa) on faults required to support the forces resulting from gravitational potential energy differences, as a function of thrust fault dip and seismogenic thickness (T_s). Black lines are calculated using the method of *Lamb* [2006], which takes into account the weight of rock in the hanging wall of the faults when calculating the average shear stress, and the grey dashed lines are calculated using a more simple model (see text for details). (a) The stresses required to support the forces resulting from an isostatically compensated elevation difference of 1 km if the mountains are weak. (b) The stresses required to support the same elevation difference if the faults in the mountains support the same average stresses as those in the lowlands, so there is a resistance to extension (see section 3). The grey circles show a thrust fault dip of 25° and a seismogenic thickness of 15 km, as discussed in the text.

shear stress on most of the subduction zone thrust faults on the margins of the Pacific Ocean, and is similar to the size of the stress drop commonly observed in earthquakes [e.g., *Scholz*, 2002].

[12] If the logic presented above is wrong, and the mountains are supported entirely by stresses in the ductile part of the lithosphere, below the seismogenic layer, then it is possible to estimate the deviatoric stresses which must be present. If the stresses are supported in an elastic layer then the effective elastic thickness (5–15 km) can be used to estimate that the layer must be able to support deviatoric stresses of 240–80 MPa. If the stresses supporting the mountains are present because of active deformation, rather than long-term strength, then we cannot use the flexural parameters to estimate the thicknesses of the layers involved, but the stresses estimated above can be simply scaled to give those which must be present in a layer of any given thickness.

4. Discussion

[13] The work presented in section 3 raises two obvious points for further discussion. The first is the question of how the situation in Albania compares with other parts of the world, such as the Andes and the Tibetan Plateau, where gravitational potential energy differences are thought to play a major role in the active tectonics [e.g., *Dalmayrac and Molnar*, 1981; *England and Houseman*, 1988]. The second question is how the topography in eastern Albania, which is currently too high to be supported by the adjacent lowlands, was initially produced.

4.1. Support of Forces and Dips of Faults

[14] We first compare our conclusions regarding the active tectonics of Albania with work done on other mountain belts. There are two main factors to consider when discussing the potential energy contrasts that can be supported by the lowlands bordering mountain belts. First, the geometry and properties of the faulting on the margin of the mountain belt need to be considered, and secondly the ability of the lowlands to support the stresses transmitted across these faults is also important. The force that can be supported by thrust faults on the margins of a mountain range will depend on a number of factors. The downdip width of the faulting will determine the area over which stresses are transmitted, and faulting extending deeper will be able to support larger elevation contrasts. If the downdip width of the faulting is controlled by temperature [e.g., *Oleskevich et al.*, 1999; *Wang and Suyehiro*, 1999], then the convergence velocity will play a major role in determining the downdip width of the faulting, as faster convergence will advect cold material further down the fault plane [e.g., *Molnar and England*, 1990], and brittle faulting will extend to deeper depths. If the faults have a "Coulomb" rheology, where the shear stress which can be supported depends on the normal stress, then the deeper parts of the faults will be able to support higher stresses (per unit area of fault plane) than the shallow parts. The dip of the fault plane is

important, because not only will it play a role in determining the thermal structure, but also for a given maximum shear stress on the faults, and assuming the lithosphere remains intact except for the faults being considered, low-angle faults can support considerably higher topographic differences than high-angle faults (Figure 6). However, the assumption in the previous sentence is an important one. No matter what the magnitude of a force that can be transmitted across thrust faults on the margins of a mountain belt, if the resultant stresses cannot be supported by the lowlands adjacent to the mountains, new faults will form, and the maximum potential energy contrast will be limited by the properties of the lithosphere in the lowlands. Variations in the geometry and rate of faulting, and in the characteristics of the lithosphere in the lowlands, are therefore able to produce a large range in the forces which can be supported by the lowlands adjacent to mountain belts, even in the absence of any variation in the strength of the faults themselves.

4.2. Comparison With the Peruvian Andes

[15] *Dalmayrac and Molnar* [1981] discussed the presence of parallel thrust and normal faulting in the Peruvian Andes from the perspective of differences in gravitational potential energy. The most striking difference between the Peruvian Andes and the mountains of Albania is the large difference in elevation contrast between the mountains and adjacent lowlands. In the Andes active extension occurs in areas where the elevation difference between the mountains and the thrust faults in the lowlands is $\sim$4 km, rather than the $\sim$1 km observed in Albania. The lowlands east of the Andes largely consist of the Precambrian Brazilian Shield, and earthquakes have been recorded on the margins of the range, and up to 400 km to the east, extending to depths of around 40 km [*Assumpcao and Suarez*, 1988; *Assumpcao*, 1992; *Emmerson*, 2007], which is approximately the depth of the Moho in the area [e.g., *James and Snoke*, 1994; *Gilbert et al.*, 2006]. These earthquake are likely to represent the breaking of the Brazilian shield in response to the gravitational potential energy contrast with the high Andes. The earthquakes commonly have nodal planes which dip at relatively high angles (e.g., 45°), which is to be expected as maximum shear stresses occur on planes at an angle of 45° to the applied force, so relatively steeply dipping faults will experience more shear stress than low-angle faults. The difference in elevation contrast between Albania and the Andes may therefore be the result of a difference in load-bearing ability in the lowlands adjoining the mountain belts, with the larger seismogenic thickness east of the Andes able to support higher mountains than the relatively thin seismogenic layer in the Adriatic. Alternatively, if the mountains are supported by stresses in the ductile parts of the lithosphere, then the relatively cool, and therefore more viscous, upper mantle in areas of thick lithosphere may explain some of the difference in elevation contrast which can be supported. If either of these are the case, then the difference between Precambrian lithosphere (which is generally thick, with an anhydrous lower crust) and Phanerozoic lithosphere [e.g., *Jackson et al.*, 2008] appears to play a fundamental role in determining the heights of mountain belts which can be created and supported.

4.3. Comparison With the Tibetan Plateau

[16] The situation in Tibet is more complex. Rather than being subparallel, the normal and thrust faulting earthquakes occur on roughly perpendicular planes. It is likely that this is, at least in part, due to an extensional stress parallel to the margin of the mountain belt, resulting from arc-normal gravitationally driven flow at the southern margin of the plateau [e.g., *Copley and McKenzie*, 2007]. The tectonics are therefore not easily discussed in relation to the two-dimensional framework adopted in this paper. However, *Bollinger et al.* [2004] constructed a three-dimensional model to study the state of stress in the Himalaya, and used topography, geodetic information, and seismicity to suggest that the deviatoric stress on the Main Himalayan Thrust was less than 35 MPa, similar to the values we estimate for Albania. It should be noted that the Tarim Basin (to the north of the Tibetan Plateau) is thought to be underlain by Precambrian cratonic material and that the lowlands of northern India, another Precambrian shield, display seismic activity throughout the thickness of the crust, as in Brazil, and a similarly high mountain belt is present.

[17] The Shillong Plateau, in northeastern India, is thought to be a block of crust which has been uplifted on relatively high-angle reverse faults (dip > 45°) bounding the plateau margins [*Bilham and England*, 2001]. This structure may represent a place where the lithosphere has broken on high-angle faults in response to the stresses transmitted across the low-angle thrust faults on the margin of the Tibetan Plateau because, as discussed above, shear stresses will be larger on relatively high-angle planes (up to a dip of 45°) than on lower angled planes. The Shillong Plateau is the only "pop-up" structure in northern India with high topography (>1000 m) and a well-developed geomorphological signature. However, the motion in the Bhuj earthquake of 2001 in NW India was reverse faulting on a plane dipping at $\sim$50° [*Antolik and Dreger*, 1986], and the distribution of aftershocks suggests that the fault ruptured through most or all of the thickness of the crust [*Bodin and Horton*, 2004]. The faulting and geomorphology are more complex in the area of the Bhuj earthquake than at the Shillong Plateau, and Bhuj may therefore represent a less well developed version of the situation seen at Shillong. As we suggested for the earthquakes with relatively high-angle fault planes in the Brazilian shield, Shillong and Bhuj may represent areas where the crust has broken on relatively high-angle planes in response to the potential energy contrast with the adjacent mountain belt. *Rodgers* [1987] described uplifts bordered by high-angle faults which cut the entire crust on the margins of Paleozoic to Cenozoic mountain belts in North America, which may be examples, preserved in the geological record, of places where gravitational potential energy contrasts between mountains and adjacent lowlands have become large enough to break new high-angle faults through the entire crust. Although we have suggested that rupture occurs on high-angle faults in the lowlands surrounding mountain belts because shear stresses are larger on planes with relatively steep dips than on shallow-dipping thrusts, it is notable that many mountain belts (such as Tibet, the central Andes, the Tien Shan, and the Caucasus) have low-angle thrusts on their margins, not

relatively steeply dipping reverse faults. This is likely to be because faulting on high-angle planes builds topography faster then low-angle faulting. Therefore, low-angle faults can accumulate larger amounts of displacement before the topography which they have produced becomes prohibitively large, and the associated increase in potential energy prevents further faulting from occurring. The faults on the margins of orogenic belts, which have large displacements, must therefore have relatively shallow dips.

4.4. Elevation of the Mountains of Albania

[18] An obvious question to now ask is why Albania is in active extension, despite being such a low mountain range, when there are several mountain ranges that are considerably higher which are not experiencing extension? The answer to this may lie in the relatively small seismogenic thickness of the Albanian lowlands which, we show later, may be related to the thick sequence of recent sediments at the surface. Alternatively, if the $\sim 25^{\circ}$ dip suggested by earthquake focal mechanisms is representative of the dips of the faults on the margins of the range, they may be steeper than those on the margins of other relatively low mountain ranges, such as the Talesh, on the western margin of the South Caspian Sea, where earthquake focal mechanisms suggest the faults dip at less than 15° [e.g., *Priestley et al.*, 1994; *Jackson et al.*, 2002]. It is also possible that the relatively slow rates of convergence in Albania (<5 mm a^{-1} [*Battaglia et al.*, 2004]) will result in a downdip width of faulting which is less than in other mountain belts where the deformation is faster (as discussed above). As Figure 6 shows, these effects would reduce the height of the mountain belt that could be supported for a given maximum shear stress on the faults (provided, in the case of the dip and downdip width of the thrust faults, that the lithosphere in the lowlands can support the forces transmitted across the faults, as described above). Throughout this discussion we have assumed that mountain belts are supported by the shear stresses on faults. If potential energy contrasts are supported elsewhere within the lithosphere, then the composition, mineralogy, hydration state, and temperature of the deeper parts of the lithosphere will also play a role in determining the height of mountain belts.

4.5. Origin of Extension in Albania

[19] A simple conceptual model of mountain belt formation involves the range increasing in height until the difference in gravitational potential energy between it and the adjacent lowlands is as large as can be supported by stresses within the lithosphere. At this limit the mountain belt will grow perpendicular to strike, strike-slip faulting will replace thrust faulting in the high mountains, and a wide flat-topped plateau will form [e.g., *England and McKenzie*, 1982]. The extension at high elevations in Albania suggests that the simple model needs modification in this area. *England and Houseman* [1988] discuss the mechanisms by which a high mountain belt can become gravitationally unstable and begin to extend and thin. They require there to be a change in the balance between the horizontal compressive stress applied to the boundary of the mountain belt and the stresses within the mountain belt which result from elevation contrasts, and suggested three ways in which the balance could be altered; either the stresses supportable by the neighboring lowlands could change, the boundary conditions acting on the edge of the mountain belt could change, or the potential energy of the mountain belt could increase. *Meco and Aliaj* [2000] suggest that the majority of the intermontane basins in eastern Albania formed in the upper Miocene to Pliocene as a result of extensional tectonics, which places a rough estimate on the age of initiation of the normal faulting, and so the change in stress state, which is similar in timing to a phase of increased exhumation at $\sim$4–6 Ma documented by *Muceku et al.* [2008] in northern Albania. With the information we presently have it is not possible to suggest why the balance of stresses in the Albanian region has been altered. It is possible that part of the mantle lithosphere may have been removed, as has been suggested for other mountain belts (e.g., Tibet [*England and Houseman*, 1988], the Betics [*Vissers et al.*, 1995], and the Andes [*Garzione et al.*, 2006]), but it is equally possible that changes in the motions of the bounding plates [*D'Agostino et al.*, 2008] could have altered the stress balance, or that the lithosphere of western Albania and the Adriatic Sea could have become weaker. One possibility is that the lithosphere beneath the lowlands may have been weakened by the deposition of a large thickness of sediments within and on the margins of the Adriatic Sea. As a simple illustration of this point, the grey dashed line on Figure 5 shows a steady state geotherm calculated in exactly the same way as that shown by the black line, with the sole exception that in this model there is no $\sim$15 km thick layer of sediments. Temperatures in the crust are lower, and the upper part of the mantle is at temperatures less than 600°, and so potentially seismogenic [*McKenzie et al.*, 2005]. The seismogenic thickness would therefore be larger than in the geotherm shown as the black line and, if the mountains are supported by shear stresses on faults, the lowlands would have been capable of supporting greater forces, and so higher mountains. If the mountains are supported by stresses in the ductile parts of the lithosphere, then cooler temperatures in the lower crust and upper mantle would result in more load-bearing ability. It is therefore possible that deposition of a large thickness of sediment caused temperatures in the uppermost crystalline basement to increase, which over time would have weakened the basement, and reduced the size of the topographic contrast that could be supported by the lowlands. The characteristic timescale for thermal diffusion can be estimated as $t = l^2/\pi^2\kappa$, where l is the thickness of the layer involved, and κ is the thermal diffusivity. For a layer thickness of 40 km (the crust and uppermost mantle), and a thermal diffusivity of 10^{-6}, the characteristic timescale is around 5 Ma. The sedimentary sequence in western Albania and the Adriatic was deposited throughout the Mesozoic and Tertiary and, although crude, the timescale we have estimated shows that the effects of thermal blanketing by sediment deposition at the surface, and radiogenic heat production within the sediments, would have had time to affect the entire crust and the upper mantle. However, this suggested mechanism for weakening the lowlands of western Albania and the Adriatic Sea, leading to normal faulting in the mountains of

eastern Albania, is speculative, and at present there is no information available which would allow us to confidently suggest the reason for the change in the stress balance between the mountains of eastern Albania and the adjacent lowlands.

5. Conclusions

[20] We have described the active deformation currently occurring in Albania and surrounding areas. The subparallel thrust and normal faulting is likely to be a result of gravitational potential energy differences between the lowlands of western Albania and the mountains in the east of the country. We have calculated the force which the mountains and lowlands exert upon each other as a result of the potential energy contrast. If the mountains are supported by shear stresses on the faults in the area, as we believe likely, then we can estimate that the average stresses are likely to be less than ~20 MPa. If, however, the mountains are supported by stresses in the ductile part of the lithosphere, then these stresses are likely to be on the order of 80–240 MPa. The deposition of large thicknesses of sediment in western Albania and the Adriatic Sea throughout the Mesozoic and Cenozoic may have weakened the crystalline lower crust and upper mantle by increasing the temperature, and therefore reduced the potential energy contrast supportable by the lowlands, and lead to the normal faulting currently occurring in the mountains of eastern Albania.

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