Chapter 9

Continental Deformation

The structural geometry of both convergent and extensional continental orogens at the outcrop scale is very (some would say - horribly) complex. More than anything else, this complexity reflects the very strong mechanical anisotropy of crustal rocks; that is, the structural geometry resulting from the deformation is largely a consequence of the inherited structure and is only weakly coupled to the nature of the forces driving the deformation. Indeed this complexity begs the question as to whether it is possible, or even useful to attempt, to evaluate the parameters governing the geodynamic evolution of continental orogenic belts. However, some of the most impressive large scale features of orogenic belts are much more regular than their internal structural geometry. For example, the topography of orogenic belts, while very fragmented (fractal) on small scales, is very regular at the scale of the orogenic belt. Indeed, just as an understanding on the control on topographic variation in the ocean basins provides fundamental insights, understanding the controls on topography provides very important insights into the mechanics of continental orogens.

We begin by examining the controls and some consequences of topography and potential enegrgy using simple calculations based on the assumption of local isostatic equilibrum. This assumption is only likely to be valid for thermally mature orogenic systems which have started to develop plataeus (e.g., Tibet) or, in extension, wide basins, in which the deformation of the lithsophere is induced by forces applied as *end loads*. The margins of orogenic belts involving wedge shaped thrust belts are certainly not in isostatic equilbrium and so the topographic variation in such circumstances need to be evaluated using a differnet set of *boundary conditions*. To tackle the mechanics of the frontal parts of mountain belts, and accretionar wedges, we need to investigate the dynamics of *critical wedges*, where the driving forces are imparted to the deforming crust as *basal tractions* along some kind of master thrust or decollement.

9.1 Deformation of the lithosphere subject to an end load

Any consideration of the behaviour of the continental lithsophere during deformation needs to recognise the contrasting influence of the crust and the mantle lithosphere in mediating both the thermal and isostatic response:

- *The crust* represents the buoyant part of the lithosphere as well as the part where lithospheric heat production is concentrated. Thickening of the crust therefore increases the buoyancy and potential energy of the lithospheric column, as well as steepening the geotherm through the increase in the heat production in the thickened column.

Note that any increase in the geotherm accompanying thermal equilibration of the thickened lithosphere may take considerable time following deformation (i.e, 50 Ma).

- *The mantle lithosphere* represents the dense negatively buoyant part of the lithosphere, which is relatively devoid of heat producing elements. Thickening of the mantle lithosphere therefore reduces the buoyancy and potential energy of the isostatically equilibrated column as well suprressing the geotherm in the overlying crust by reducing the heat flow into the base of the crust.

Because of this contrasting influence it is useful to consider the effects of the crust and mantle lithosphere, independently. This can be achieved by describing the deformation of the lithosphere by the ratio of the changes in thickness of the crust, f_c, and mantle lithosphere, f_m, where these are defined at any stage of the deformation by:

In terms of these parameters the change in thickness of the total lithosphere f_l as used by Sandiford & Powell, 1990 etc. is given by $f_l = f_m + \psi\,(f_c - f_m)]$ where $\psi = z_{c0}/z_{l0}$.

$$f_c = \frac{z_c}{z_{c0}}$$

$$f_m = \frac{z_m}{z_{m0}}$$

where z_c and z_m are the deformed thickness of the crust and mantle lithosphere, respectively, and z_{c0} and z_{m0} are the initial thickness of crust and mantle lithosphere prior to deformation.

Airy Isostasy and crustal thickening

Assuming Airy isostasy then crustal thickening results in both an increase in surface elevation, c_h, and the the development of the crustal root, c_r.

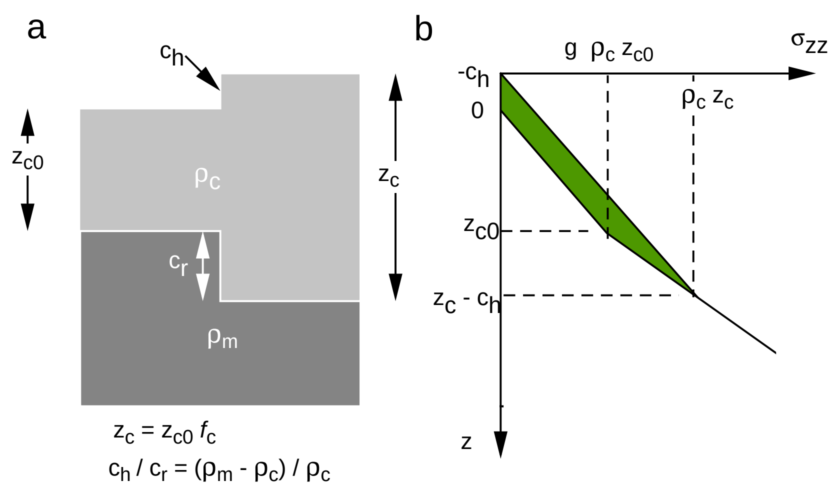

Figure 9.1: Isostatic effects of crustal thickening.

For a crust of constant density, ρ_c, overlying mantle of density ρ_m, the ratio of the change in surface elevation c_h to the thickness of the crustal root c_r is given by

$$\frac{c_h}{c_r} = \frac{\rho_m - \rho_c}{\rho_c}$$

The change in crustal thickness is given by the sum of $c_h + c_r$:

$$f_c = \frac{c_h + c_r + z_{c0}}{z_c 0}$$

giving

$$c_r = f_c\, z_{c0} - c_h - z_{c0}$$

Therefore

$$c_h = z_{c0}\,(1 - f_c)\left(\frac{\rho_c - \rho_m}{\rho_c}\right)$$

Note that the change in surface elevation is linear in the change in crustal thickness; that is, it is linear in $f_c\, z_{c0}$. The change in potential energy U_c for this scenario is given by:

$$U_c = \frac{g\,\rho_c}{2}\left((z_{c0}\,f_c)^2 - z_{c0}^2 - 2\,c_r^2\right) - \frac{g\,\rho_m}{2}c_r^2$$

Note that potential energy changes with the square of the crustal thickening.

The effects of crustal thinning (i.e., $f_c < 1$) are exactly opposite crustal thickening, that is it results in subsidence and a reduction in potential energy (see Chapter 10).

Airy Isostasy and the mantle lithosphere

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See Sandiford & Powell (1990) *EPSL*

9.2 Deformation within the lithosphere due to basal tractions