Horizontal structures in granulite terrains: A record of mountain building or mountain collapse?

Michael Sandiford
Department of Geology and Geophysics, University of Adelaide
GPO 498, Adelaide 5001, Australia

ABSTRACT

In many high-temperature (>800 °C) granulite terrains, the development of characteristic horizontal structures occurred during the metamorphic culmination and was followed by isobaric cooling. The absolute magnitude of isobaric cooling (commonly >300 °C) implies cooling intervals of the order of the thermal time constant of the continental lithosphere (~100 m.y.). Such prolonged isobaric cooling implies that no significant erosional denudation followed the development of the horizontal structures and thus precludes the prograde deformation being responsible for significant crustal thickening. Rather, the prograde deformation more probably records bulk crustal thinning during extensional collapse of a previously thickened crust possibly triggered by detachment of a thickened thermal boundary layer at the base of the lithosphere.

INTRODUCTION

The prograde deformation history of granulite terrains commonly involves the development of horizontal structures such as recumbent folds and foliations (Coward, 1973; Park, 1981). The development of these structures has been attributed to mountain-building deformations at convergent plate boundaries, the high-temperature metamorphism proceeding partly as a consequence of the induced crustal thickening (Park, 1981; England and Bickle, 1984; Bohlen, 1987; Newton, 1987). The alternative possibility, that the horizontal structures record crustal extension (Sandiford and Powell, 1986a), has received comparatively little attention partly because of the entrenched notion that granulite terrains represent the exhumed mid-levels of very thick crust (60–70 km) rather than the deep levels of normal-thickness crust (25–40 km). The thermal effects of metamorphic cycles involving crustal thickening and erosion are well understood (e.g., England and Thompson, 1984). For plausible rates of erosion, which allow advection of rock at rates faster than the conductive thermal equilibration of the continental lithosphere, crustal thickening-erosion cycles produce characteristic pressure-temperature-strain paths showing significant near-isothermal decompression following deformation. Recent studies suggest that granulite terrains more typically show isobaric cooling following prograde deformation, especially those granulite terrains metamorphosed at maximum temperatures above ~800 °C (e.g., Ellis, 1980; Warren, 1981; Sandiford, 1985a; Bohlen, 1987). In this paper, I argue that the horizontal structures in isobarically cooled granulite terrains are best explained if they developed in response to the extensional collapse of formerly overthickened crust. A fundamental difference between this model and the conventional interpretation is that the deformation giving rise to the horizontal structures is regarded as a consequence rather than a cause of crustal thickening and heating.

The Napier Complex in Enderby Land, Antarctica, is a well-documented, isobarically cooled granulite terrain showing the characteristic association between the high-temperature metamorphism and the development of horizontal structures (Ellis, 1980; Sandiford, 1985a). The timing and extent of isobaric cooling and the effect of prograde deformation on bulk crustal thickness in the Napier Complex are discussed below in order to provide an understanding of the origin of the horizontal structures.

GRANULITE FACIES METAMORPHISM IN THE NAPIER COMPLEX

Background: Metamorphism and Deformation

The Napier Complex underwent high-temperature granulite facies metamorphism and deformation, resulting in the formation of macroscopic recumbent isoclinal folds and regionally pervasive subhorizontal foliations in the mid-Archean (Fig. 1; James and Black, 1981; Sandiford and Wilson, 1984). Granulite facies assemblages include sapphirine-quartz-rutile and osumilite-quartz in metapelites, pigeonite in meta-ironstones, and calcic-mesoperthite in charnockitic gneisses (Ellis et al., 1980; Grew,

Figure 1. Isoclinal folds formed during granulite facies metamorphism in Napier Complex, Antarctica. Cliff face is approximately 250 m high. Unfolding of retrograde dome and basin structures restores these prograde isoclinal folds to initial horizontal attitude (Sandiford and Wilson, 1984).

1980, 1982; Harley, 1985, 1986, 1987; Sandiford, 1985a; Sandiford and Powell, 1986b), and indicate metamorphic temperatures of at least 1000 °C at pressures between 7 and 9 kbar. The terrain is now exposed in oblique profile; pressures increase about 2 kbar southwestward across the terrain. There is, however, no concomitant rise in peak metamorphic temperatures, which were nearly constant over much of the terrain (Sandiford and Powell, 1986b). Consequently, the piezothermic array defined by the highest temperature assemblages is nearly isothermal in the depth range 25–30 km. This contrasts the shallow $\delta P/\delta T$ arrays expected at elevated temperatures from conductive heating during metamorphic cycles involving crustal thickening followed by erosion (England and Richardson, 1977). The systematic preferred orientation of minerals such as sillimanite in metapelites and pyroxenes in mafic granulites within the pervasive foliation indicates that this high-temperature metamorphism was coeval with the development of the horizontal structure (James and Black, 1981; Sandiford and Wilson, 1984).

The retrograde metamorphic history of the Napier Complex is recorded in corona and symplectite textures developed as the products of reactions involving the prograde mineral assemblages, and in shear zones dated as Late Archean/Early Proterozoic (2500–2350 Ma), Late Proterozoic (1100–1000 Ma), and early Paleozoic (500 Ma) (James and Black, 1981; Black et al., 1983). Sillimanite-hypersthene coronas separating prograde sapphirine and quartz, and garnet coronas between prograde pyroxene and plagioclase indicate cooling to temperatures as low as 600–700 °C at constant pressure (Sandiford, 1985a). Proterozoic shear zones preserve garnet-hornblende and kyanite-bearing assemblages equilibrated at ~650 °C and 7–8 kbar (Sandiford, 1985b). These assemblages indicate that, following prograde metamorphism, the Napier Complex granulites remained in the deep crust for some 2000 m.y., during which time they cooled some 300–400 °C with virtually no change in pressure.

Crustal Thickness During Prograde Metamorphism

The prolonged residence of the Napier Complex at depths of 25–30 km precludes any significant erosional denudation of the crust in the 2000 m.y. following the prograde granulite facies metamorphism and associated development of horizontal structures. At this time the Napier Complex could not have been part of a significantly overthick crust; the total crustal thickness was unlikely to be more than 35–40 km, the currently exposed gneisses then being only a few kilometres from the continental Moho (Sandiford, 1985a). The subsequent excavation resulting in the present-day surface exposure of the Napier Complex is related to its incorporation into a Late Proterozoic orogen some 2000 m.y. after the granulite metamorphism (Sandiford, 1985b).

In order to understand the origin of horizontal structures in the Napier Complex, it is necessary to establish how the prograde deformation changed the bulk crustal thickness. Ellis (1987) has elaborated on indirect evidence indicating that the total crustal thickness early in the prograde metamorphic/deformation cycle must have been significantly greater than the 35–40 km inferred for the end of this cycle on the basis of the prolonged isobaric cooling history. This evidence comes from syntectonic felsic intrusives that exhibit a characteristic geochemical signature indicative of derivation from a crustal source at pressures in excess of ~15 kbar (Sheraton and Black, 1983; Ellis, 1987). This observation implies crustal thicknesses in excess of ~50 km during the early stages of granulite facies metamorphism and can only be rationalized with the prolonged isobaric cooling if the prograde deformation resulted in bulk crustal thinning from >50 km to ~35–40 km.

DISCUSSION

The pressure changes following deformation in metamorphic terrains provide an important insight into the nature of the deformation and the relation between the deformation and metamorphism. For deformations resulting in significant crustal thickening and consequent lithospheric heating, the post-deformation reaction textures should reflect substantial erosion-induced decompression. Thermal models for crustal thickening-erosion cycles indicate that decompression will be accompanied initially by only minor changes in temperature (e.g., England and Thompson, 1984). As stated in the preceding sections, many high-temperature granulite terrains show evidence for isobaric cooling following prograde deformation, rather than isothermal decompression. To date, there are few granulite terrains for which the time scale of this isobaric cooling is clearly defined by regional geologic constraints. However, cooling from maximum temperatures in excess of 800–900 °C to ~600 °C is common (e.g., Warren, 1981; Bohlen, 1987), requiring cooling intervals at least of the order of the thermal time constant of the continental lithosphere (~100 m.y.). Such prolonged isobaric cooling is inconsistent with the prograde metamorphism and with deformation having terminated while the granulite terrains were at the mid-levels of substantially overthick crust. Instead of being responsible for crustal thickening, as conventional wisdom would indicate, I suggest that the prograde deformation giving rise to the characteristic horizontal structures in isobarically cooled granulite terrains reflects the extensional collapse of formerly overthick crust. In the Napier Complex, this hypothesis is consistent with geochemical evidence for the existence of thick crust early in the prograde metamorphic cycle.

The scenario for collapse of thickened crust outlined above can be modeled by considering how lithospheric strength and potential energy evolve as a consequence of crustal and lithospheric thickening in convergent orogens such as continental collisions. Crustal thickening not only greatly increases the lithospheric heat production leading to significant heating on orogenic time scales (e.g., England and Thompson, 1984) but may also induce substantial potential energy contrasts between the deformed zone and surrounding undeformed lithosphere. This is largely because of the difference in the surface elevation between the deformed thickened crust and its surroundings, but also because of Moho

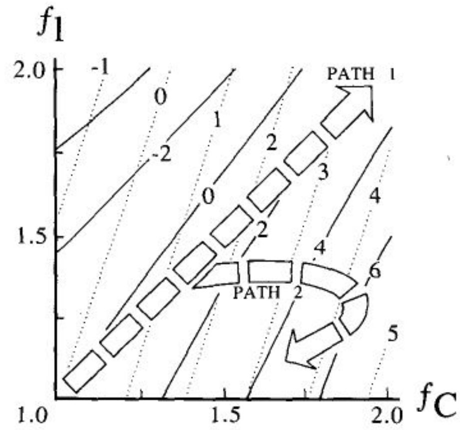

Figure 2. Surface elevation and potential energy of deformed lithosphere evolve as consequence of relative changes in crustal and lithospheric thickness (e.g., Turcotte, 1983), which can be characterized by ratios f_c and f_l of thickness of deformed crust and lithosphere to initial (or undeformed) crust and lithosphere, respectively. Initial lithosphere is assumed to be 100 km thick and crust 35 km thick; both f_c and f_l are modeled in range 1.0 to 2.0, appropriate to continental collisions. Dotted lines show isostatically compensated elevation contrast expected between deformed and initial lithosphere. Thin solid lines show force per unit length of orogen (× 10^{12} N/m) exerted on undeformed lithosphere adjacent to deformed lithosphere (see Appendix 1). Path 1 shows deformation path where thermal boundary layer remains attached to lithosphere during convergent deformation. Path 2 shows deformation path where thermal boundary layer detachment, initiated after onset of deformation, causes lithosphere to thicken at slower rate than crust and leads finally to collapse.

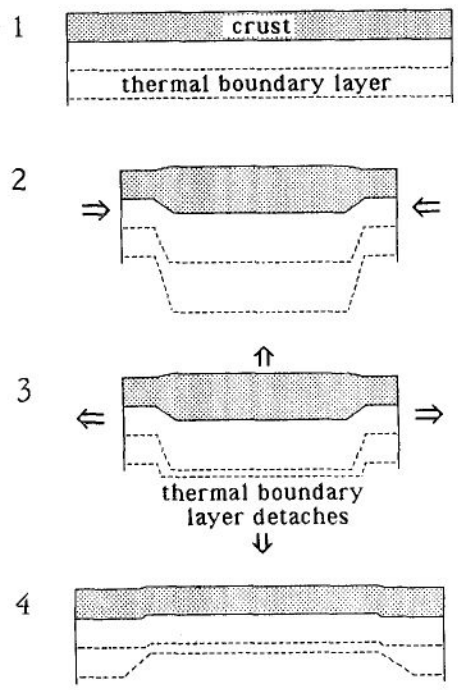

Figure 3. Schematic diagram depicting evolution of convergent orogen where collapse is initiated by detachment of thickened lower thermal boundary layer. Reference continental lithosphere (1) is thickened homogeneously (2). Detachment of thickened lower thermal boundary layer from lithosphere (3) leads to uplift as well as heating, thereby providing necessary increase in potential energy and decrease in lithospheric strength to initiate extensional collapse (4).

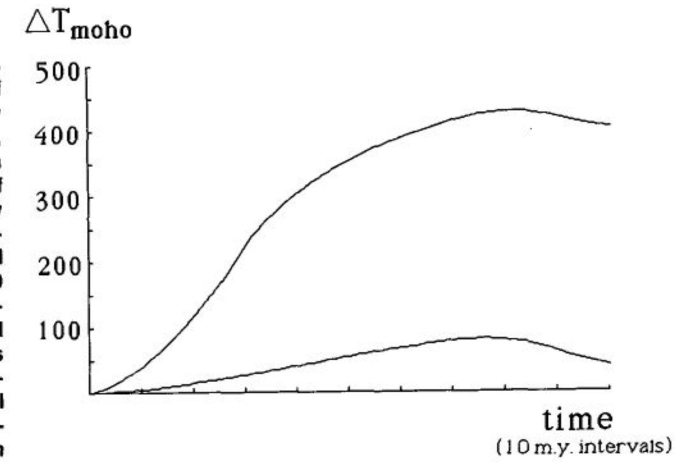

Figure 4. Magnitude of crustal heating during convergent deformations is sensitive to response of lower thermal boundary layer of lithosphere (e.g., Houseman et al., 1981). In this figure thermal state of crust is represented by change in Moho temperatures (ΔT_{Moho}) from initial Moho temperature (500 °C) for deformations involving different thermal boundary layer responses (see Appendix 1 for description of thermal model and deformation parameters). For deformation in which lower thermal boundary layer remains attached to lithosphere (i.e., lithosphere and crust thicken at same rate), crustal heating is minimal with $\Delta T_{Moho} < 100$ °C (illustrated by lower curve) and precludes granulite facies metamorphism. For deformation involving detachment of lower thermal boundary layer, substantial heating of crust occurs. Upper curve models detachment that maintains lithosphere of constant thickness while crust is thickened by factor of two and shows sufficient crustal heating ($\Delta T_{Moho} \sim 400$ °C) for high-temperature granulite facies metamorphism.

topography induced by the isostatic compensation of the thickened crust (e.g., Artyushkov, 1973; Turcotte, 1983). This potential energy contrast may lead to substantial horizontal extensional deviatoric stresses in the deformed lithosphere and corresponding compressional stresses in adjacent undeformed lithosphere (Fig. 2). A criterion for the onset of collapse is that the extensional forces exceed the driving forces for convergence by a critical amount related to the strength of the lithosphere. Collapse may be initiated by a reduction in the driving forces during the terminal stages of collision or by a rapid increase in the potential energy of the deforming zone that may accompany the detachment of the thickened lower thermal boundary layer during convergence (e.g., Houseman et al, 1981; England and Houseman, 1988; Figs. 2 and 3).

Lithospheric strength is strongly temperature-dependent, and the ability of lithosphere to support stresses arising from potential energy contrasts as a consequence of deformation must therefore decrease with the heating induced by the deformation. The ability of a self-heating thickened lithosphere to support the extensional deviatoric stresses arising from isostatically compensated elevated topography has been modeled by England (1987) using existing laboratory-determined rheological data. Using the temperature of the Moho as a guide to the thermal state of the lithosphere, England (1987) suggested that for Moho temperatures in excess of ~750 °C, the lithosphere is likely to be too weak to support the excess topography observed at modern collision zones such as Tibet, where substantial increases in potential energy appear to have accompanied rejuvenated uplift possibly associated with lower thermal boundary layer detachment in the recent geologic past (England and Houseman, 1988). Uncertainties are inherent in the extrapolation of laboratory-determined rheological data to the geological environment (e.g., Patterson, 1987); however, England's results suggest that extensional collapse of thickened crust may be a plausible mechanism for the denudation of old mountain ranges once temperatures in the deep crust are sufficiently high for granulite facies metamorphism. In the ductile levels of the crust, extensional collapse will cause the rotation of preexisting structures into subhorizontal orientations and also will result in the development of horizontal foliations. When collapse has reduced topography to the extent that existing potential energy gradients can be supported by the strength of the lithosphere, deformation will cease and the crust will cool with little further change in thickness.

In order to generate lower crustal temperatures in excess of 750–800 °C during continental collisions, it is necessary for the thickened lower thermal boundary layer to detach from the lithosphere during or following convergent deformation (Houseman et al., 1981; Figs. 3 and 4). The isostatic consequence of thermal boundary layer detachment is to enhance the surface topography by as much as several kilometres, causing a substantial increase in the potential energy of the deformed lithosphere (path 2 in Fig. 2). Detachment of a thermal boundary layer beneath a zone of thickened crust formed during convergent deformation therefore provides the mechanism for generating the high-temperature granulite facies metamorphism as well as providing the necessary energy required to trigger the onset of extensional collapse of the thickened crust.

APPENDIX 1.

The horizontal force arising from deformation-induced gradients in surface elevation, crustal thickness, and lithospheric thickness is given by

$$F = \int_0^{z_a} P(L_d)\,dz - \int_0^{z_a} P(L_u)\,dz, \quad (1)$$

where F is the force (per unit length) in a reference or undeformed lithosphere (L_u) bounding a zone of deformed lithosphere (L_d), P is the lithostatic pressure, and z_a is the depth at which the density structure is uniform beneath both the deformed and undeformed lithosphere.

The thermal evolution of deforming lithosphere is given by

$$\partial T/\partial t = \kappa \cdot \partial^2 T/\partial z^2 - v \cdot \partial T/\partial z + H/(\rho \cdot c_p), \quad (2)$$

where T is temperature, v is vertical velocity, t is time, z is depth, κ is diffusivity, H is heat production, ρ is density, and c_p is heat capacity. Thermal parameters are modeled to give an initial surface heat flow of 63 $mW \cdot m^{-2}$ and a mantle heat flux of 30 $mW \cdot m^{-2}$ and an initial Moho temperature of 500 °C. An initial crust 35 km thick and an exponentially decreasing heat production are assumed. The initial lithospheric thickness is 100 km. Deformation producing double crustal thickening occurs from time 0 to 30 Ma. From time 30 to 60 Ma, there is no deformation. From time 60 to 90 Ma, extensional collapse restores the original crustal thickness. Equation 2 has been solved by using a Crank-Nicolson finite difference algorithm with time-stepping intervals of 1 m.y.

REFERENCES CITED

Artyushkov, E.V., 1973, Stresses in the lithosphere caused by crustal thickness inhomogeneities: Journal of Geophysical Research, v. 78, p. 7675-7708.

Black, L.P., James, P.R., and Harley, S.L., 1983, Geochronology and structure of the Early Archaean rocks at Fyfe Hills, Antarctica: Precambrian Research, v. 21, p. 197-222.

Bohlen, S., 1987, Pressure-temperature-time paths and a tectonic model for the evolution of granulites: Journal of Geology, v. 5, p. 617-632.

Coward, M.P., 1973, Flat lying structures within the Lewisian basement gneiss complex of NW Scotland: Geologists' Association Proceedings, v. 85, p. 459-572.

Ellis, D.J., 1980, Osumilite-sapphirine-quartz granulites from Enderby Land: P-T conditions of metamorphism, implications for garnet-cordierite equilibria and the evolution of the deep crust: Contributions to Mineralogy and Petrology, v. 74, p. 201-210.

—— 1987, Origin and evolution of granulites in normal and thickened crusts: Geology, v. 15, p. 167-170.

Ellis, D.J., Sheraton, J.W., England, R.N., and Dalwitz, W.B., 1980, Osumilite-sapphirine-quartz granulites from Enderby Land, Antarctica, mineral assemblages and reactions: Contributions to Mineralogy and Petrology, v. 72, p. 353-367.

England, P.C., 1987, Diffuse continental deformation: Length scales, rates and metamorphic evolution: Royal Society of London Philosophical Transactions, ser. A, v. 321, p. 3-22.

England, P.C., and Bickle, M., 1984, Continental thermal and tectonic regimes during the Archaean: Journal of Geology, v. 92, p. 123-143.

England, P.C., and Houseman, G.A., 1988, The mechanics of the Tibetan Plateau: Royal Society of London Philosophical Transactions, ser. A, v. 326, p. 301-320.

England, P.C., and Richardson, S.W., 1977, The influence of erosion upon the mineral facies of rocks from different metamorphic terrains: Geological Society of London Journal, v. 134, p. 201-213.

England, P.C., and Thompson, A.B., 1984, Pressure-temperature-time paths of regional metamorphism, 1. Heat transfer during the evolution of regions of thickened crust: Journal of Petrology, v. 25, p. 894-928.

Grew, E.S., 1980, Sapphirine + quartz association from Archaean rocks in Enderby Land, Antarctica: American Mineralogist, v. 65, p. 762-787.

—— 1982, Osumilite in the sapphirine-quartz terrane of Enderby Land, East Antarctica: Implications for osumilite petrogenesis in the granulite facies: American Mineralogist, v. 67, p. 762-787.

Harley, S.L., 1985, Garnet-orthopyroxene bearing granulites from Enderby Land, Antarctica: Metamorphic pressure-temperature-time evolution of the Archaean Napier Complex: Journal of Petrology, v. 26, p. 819-856.

—— 1986, A sapphirine-cordierite-garnet-sillimanite granulite from Enderby Land: Implications for FMAS petrogenetic grids in the granulite facies: Contributions to Mineralogy and Petrology, v. 94, p. 452-460.

—— 1987, A pyroxene-bearing metaironstone and other pyroxene granulites in Enderby Land, Antarctica: Further evidence for very high T (>980 °C) Archaean regional metamorphism in the Napier Complex: Journal of Metamorphic Geology, v. 5, p. 341-356.

Houseman, G., McKenzie, D., and Molnar, P., 1981, Convective instability of a thermal boundary layer and its relevance for the thermal evolution of continental convergent belts: Journal of Geophysical Research, v. 86, p. 6115-6132.

James, P.R., and Black, L.P., 1981, A review of the structural evolution and geochronology of the Archaean Napier Complex of Enderby Land, Australian Antarctic Territory: Geological Society of Australia Special Publication 7, p. 71-83.

Newton, R.C., 1987, Petrologic aspects of Precambrian granulite facies terrains bearing on their origins, *in* Kroner, A., ed., Proterozoic lithospheric evolution: American Geophysical Union Geodynamics Series, v. 17, p. 11-26.

Park, R., 1981, Origin of horizontal structure in Archaean high grade terrains: Geological Society of Australia Special Publication 7, p. 481-490.

Patterson, M., 1987, Problems with extrapolation of laboratory determined rheological data: Tectonophysics, v. 133, p. 33-44.

Sandiford, M., 1985a, Metamorphic evolution of granulites at Fyfe Hills: Implications for Archaean crustal thickness: Journal of Metamorphic Geology, v. 3, p. 155-178.

—— 1985b, The origins of retrograde shear zones in the Napier Complex: Implications for the tectonic evolution of Enderby Land, Antarctica: Journal of Structural Geology, v. 7, p. 477-488.

Sandiford, M., and Powell, R., 1986a, Deep crustal metamorphism during continental extension: Modern and ancient examples: Earth and Planetary Science Letters, v. 79, p. 151-158.

—— 1986b, Pyroxene exsolution in granulites from Fyfe Hills, Enderby Land, Antarctica: Evidence for 1000°C temperatures in Archaean continental crust: American Mineralogist, v. 71, p. 946-954.

Sandiford, M., and Wilson, C.J.L., 1984, The structural evolution of the Fyfe Hills-Khmara Bay region, Enderby Land, Antarctica: Australian Journal of Earth Sciences, v. 31, p. 403-426.

Sheraton, J.W., and Black, L.P., 1983, Geochemistry of Precambrian gneisses: Relevance for the evolution of the east Antarctic Shield: Lithos, v. 16, p. 273-296.

Turcotte, D.L., 1983, Mechanisms of crustal deformation: Geological Society of London Journal, v. 140, p. 701-724.

Warren, R.J., 1981, Metamorphic and tectonic evolution of granulite, Arunta Block, central Australia: Nature, v. 305, p. 300-303.

Manuscript received June 28, 1988
Revised manuscript received December 12, 1988
Manuscript accepted January 24, 1989