The role of deep basement during continent–continent collision: a review

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Abstract: Structural, geophysical and metamorphic studies show that collisional orogeny thickens the crust by a factor of two or more. A large volume of continental material at the base of the orogen is, therefore, subject to eclogite facies conditions. Phase equilibration results in a loss of buoyancy and thermodynamic heating of this crustal root. This dense crustal material may be partially subducted, as in the Alps or the Himalayas, and lost to the system. Alternatively, it may rest isostatically below the Moho until it is partially exhumed during orogenic collapse, as in the Scandinavian Caledonides or the Tonbai-Dabie Mountains. Remnant orogenic roots may exist as seismically reflective mantle and provide a locus for subsequent Wilson Cycle rifting. The rate at which these phase transformations take place may have a profound buffering effect on the amount and duration of orogenic contraction. Isostatically compensated transient 2-dimensional finite element thermal models are presented, which seek to place some limits on these processes. It is interesting to speculate whether more is learnt about the process of orogeny from a single exhumed eclogitic boudin or from mapping nappe complexes.

The topography of mountain belts developed during continent–continent collision is supported by a deep crustal root. Deep seismic reflection and refraction profiles across Cenozoic (e.g. Hirn *et al.* 1984*a*, *b*; Valasek *et al.* 1991; Zhao *et al.* 1993) and ancient orogens (e.g. Mathews & Hirn 1984; Hynes & Snyder 1995) have demonstrated that throughout the Phanerozoic these roots have extended to depths of 70 km or more. Exhumed ultra-high pressure (UHP) metamorphic terranes developed within such orogens (Smith 1984; Chopin 1987; Wang *et al.* 1989) contain coesite, implying that continental crust has been buried to even greater depths, exceeding 90 km. This contribution will argue that the phase assemblages of these roots, particularly the portion below 40 km, has a profound effect upon orogenic evolution. The transformation of lithologies within this root to the eclogite facies during convergence reduces buoyancy and buffers topography. Such phase changes may also exert controls on the subduction and delamination of continental crust and affect the P–T–t path of the orogeny. Crustal roots of mantle density occurring beneath the seismic Moho may play a role in cyclic re-opening of sutured oceans. Retrogression of such large volumes of rock result in a gain in buoyancy that contributes to the collapse process. Dewey *et al.* (1993) calculated the magnitude of the buoyancy effects during both burial and exhumation using a one dimensional airy isostatic model. Ryan & Dewey (1997) have modelled the long-term thermal weakening of the lithosphere caused by eclogitized crustal material beneath the seismic Moho. This study integrates these two approaches to model the topographical and thermal effects of eclogite facies transformations during continental collision. This approach allows the robustness of the earlier topographical models to be tested.

Coherent eclogite facies terranes

Some collapsed orogens contain regions of tens of thousands of square kilometres of continental lithologies, which were subjected to eclogite facies conditions during continental collision (Fig. 1). The eclogite facies assemblages in these terranes are typically preserved in mafic boudins within a lower grade, usually amphibolite facies, matrix showing structural evidence for horizontal extension. This is probably because the basaltic rocks tend to form early boundins which resist subsequent fluid penetration and alteration. However, radiometric and petrological studies show that the rocks of the matrix were also subjected to eclogite facies conditions during peak metamorphism (see, for example, Krogh 1977; Dewey *et al.* 1993; Ames *et al.* 1996); the

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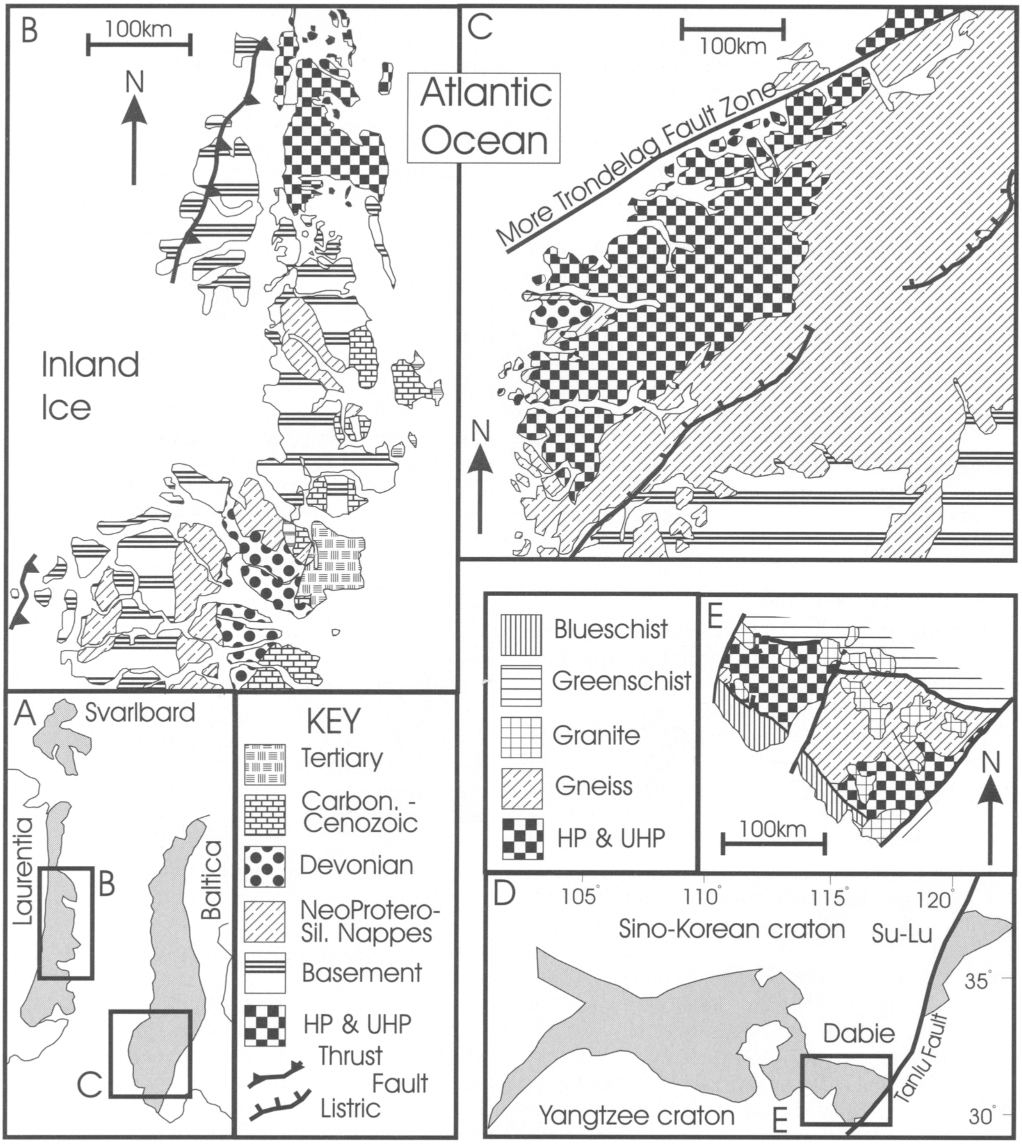

Fig. 1. Outline geological maps of collisional orogens with coherent eclogite facies terranes. (**a**) Location map for Greenland and Baltic plotted using pre-Atlantic opening reconstruction. The Caledonides are shaded. (**b**) Geological map of the Western Gneiss Region, Norway. (**c**) Geological map of the north-east Greenland Caledonides. (**d**) Location map showing the Triassic Dabie–Qinling orogeny. (**e**) Outline geological map of the Dabie Shan region. All geological maps are drawn to the same scale.

current fabric of the terrane being mainly controlled by the process of exhumation which took place during the collapse of the orogen (Andersen & Jamtveit 1990; Chauvet *et al.* 1992; Ames *et al.* 1993; Steltenphol *et al.* 1993; Andersen 1998). A relict high pressure gradient may occur across such terranes which indicates that they have been buried and exhumed wholesale (Krogh 1977) or without apparent substantial disruption of their burial geometry (Cuthbert *et al.* 2000).

These terranes, in which all crustal rocks have undergone HP or UHP metamorphism during prograde metamorphism are here termed coherent eclogite facies terranes (CEFT). They should be distinguished from eclogite bearing terranes formed within subduction zone complexes, which pre-date continent–continent collision (e.g. Eoalpine eclogites of the Sesia zone within the European Alps), are tectonically imbricated with lower grade metamorphic, usually blueschist facies rocks (e.g. the Franciscan mélange, Cloos 1982), and do not affect large regions of older continental basement. Such terranes undergo a much lower bulk density change (±5% for transformations to and from the blueschist facies) than CEFTs (see below) and will always remain positively buoyant with respect to the asthenosphere.

CEFTs were developed during: the Caledonian (Silurian) Baltica–Laurentia collision in the Western Gneiss region (WGR) of Norway (Krogh 1977; Smith 1984) (Fig. 1c) at 420–400 Ma (Griffin & Brueckner 1985) and NE Greenland (Gilotti 1993) at 405 ± 24 Ma (Brueckner *et al.* 1998) (Fig. 1b); the collision of Gondwana with Laurussia during the Variscan orogeny of Europe (Carswell 1990; Le Pichon *et al.* 1997); and in the Dabie-Qinling-Su Lu orogen developed during the Triassic collision of the Sino-Korean and Yangtze cratons (Wang *et al.* 1989; Hacker *et al.* 1998) (Fig. 1e). Such terranes may have developed beneath the Himalayas and Tibet (Sapin & Hirn 1997), and the Alps (Butler 1986). Preservation of Pan-African and Grenvillian continental eclogites suggests that this process may have been active since the Neo-Proterozoic (Sanders *et al.* 1987; Sanders 1989; Bernard-Griffiths *et al.* 1991; Castaing *et al.* 1993; Indares 1993; Möller 1998, 1999). However, the conditions necessary for UHP metamorphism may not have occurred until the Proterozoic–Phanerozoic boundary (Maruyama & Liou 1998). Higher geotherms of perhaps 15–25°C km^{-1} (Maruyama & Liou 1998) and consequent lower lithospheric strengths would make it difficult to form very deep roots during the Archaean or perhaps the Palaeoproterozoic. The positive slope of the granulite to eclogite facies reactions in P–T space mean that a greater thickness of root would be required before such transformations could occur, and lithosphere with a lower strength would be less likely to support the topography associated with such a root.

CEFTs and buoyancy within the orogen

Transformation to eclogite facies assemblages is associated with an increase in density of *c.* 6.7% to 21.1% for a range of crustal lithologies (Table 1). If an orogen develops an eclogitized crustal root then the continental lithosphere will loose buoyancy (Richardson & England 1979; Dewey *et al.* 1993). This effect is illustrated in Figure 2, which shows the increase in mean lithospheric density associated with the development of an eclogitized root. The lithospheric density (ρ_l) is calculated using the following equation (Dewey *et al.* 1993):

$$\rho_l = \frac{\int_0^{C_z} \rho_c(z)\delta z + \int_{C_z}^{L_z} \rho_m(z)\delta Z}{l_z}$$

Table 1. *Model and measured densities for eclogite facies rocks and their precursors*

Rock type	Initial density kg m^{-3}	Granulite facies density kg m^{-3}	Eclogite facies density kg m^{-3}	Percentage density increase	Reference
granodiorite	2760		3100	12.3	Bosquet *et al.* (1997)
		2870		8.0	
andesite	2910		3460	18.9	Bosquet *et al.* (1997)
		3040		13.8	
basalt	2940		3560	21.1	Bosquet *et al.* (1997)
		3270		8.9	
anorthositic gabbro	3055		3260	6.7	Austrheim & Mörk (1988)
anorthosite	2785		3060	9.9	Austrheim & Mörk (1988)
gabbro	3166		3480	9.9	Austrheim & Mörk (1988)
gabbro anorthosite	2870		3190	11.1	Austrheim & Mörk (1988)
mafic mangerite	2980		3210	7.7	Austrheim & Mörk (1988)
pelitic restite	2900		3400	17.2	Hynes & Snyder (1995)

The two values for the percentage change in model density for each rock type calculated by Bosquet *et al.* (1997) represent eclogitization of upper crustal rocks (upper value) and their garnet granulite facies equivalents (lower value). Austrheim & Mörk (1988) measured the change in density in the field at the granulite–eclogite transition. The value for a pelitic restite (Hynes & Snyder 1995) is a model value calculated from a kyanite–garnet containing pelite after 57% melting at 0.8 GPa.

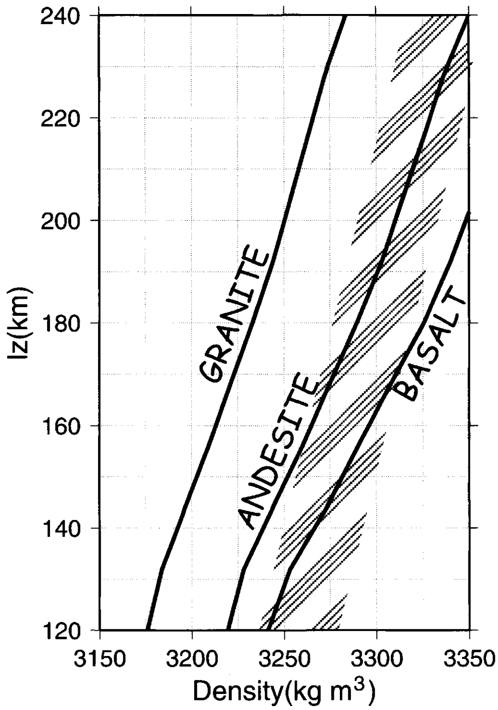

Fig. 2. Plots showing the increase in lithospheric density for a lithosphere of initially 120 km with a continental crust of 40 km. The model assumes that the crust between 30 and 40 km is converted to granulite and that below 40 km to eclogite facies assemblages using the density values of Bosquet *et al.* (1997). The lithosphere is thickened by vertical pure shear by a factor of 2. There is no erosion. The crust is assumed to have a mean temperature of 300°C and the mantle of 950°C. The likely range of asthenospheric densities is patterned.

where ρ_c, ρ_l and ρ_m are the densities of the crust, lithosphere and mantle, respectively; C_z and l_z are the thicknesses of the crust and lithosphere; and z is the depth.

The simple model presented in Figure 2 considers a 40 km thick crust of granodioritic, andesitic or basaltic composition completely transformed to granulite below 30 km and eclogite below 40 km. Homogenous pure shear is assumed, the lithosphere is initially 120 km in thickness and densities are taken from Bosquet *et al.* (1997). The crustal rocks are assumed to have a compressibilities of 10^{-11} Pa^{-1} and $8.0 * 10^{-12}$ Pa^{-1}, and a thermal expansivities of $2.5 * 10^{-5}$ °K^{-1} and $3.0 * 10^{-5}$ °K^{-1} for the crust and mantle, respectively. The likely range of densities of the asthenosphere (3250–3290 kg m^{-3} at 120 km, equivalent to 3284–3324 kg m^{-3} at normal temperature and pressure) is hachured. The mean density of lithosphere with basaltic crust can exceed the likely range for the density of the asthenosphere. A continental lithosphere with andesitic crust might become neutrally buoyant with respect to an asthenosphere of relatively low density when the lithosphere is stretched to a thickness of 190 km (Fig. 2) and consequently the crust to a thickness of 63 km. This means that further thickening will not lead to continued uplift and the orogen would subside below sea level. A lithosphere with a granodioritic crust will always remain positively buoyant with respect to the asthenosphere. These curves assume wholesale instantaneous metamorphic conversion, which for kinetic reasons is unlikely (Austrheim *et al.* 1997), and must represent an extreme end member. They do illustrate, however, that increasing thickening is associated with increasing lithospheric density which will tend to buffer orogenic uplift.

Dewey *et al.* (1993) have modelled the topographical consequences of transforming a significant volume of the root below 40 km to eclogite facies assemblages during convergence, and retrogressing them to amphibolite facies assemblages during exhumation. The prograde reactions allow shortening to continue after the formation of topography of *c.* 3 km because a steady-state is reached where the increase in density of the orogenic root compensates for any further thickening without producing more uplift. If such phase changes do not take place then the orogenic root is unlikely to exceed 60–80 km in thickness (Dewey 1988; Molnar *et al.* 1993) as the resultant gravitational forces, produced by the excess topography, will exceed those driving the convergence.

During exhumation the dense eclogitic root acts as a 'gravitational battery'. The retrogression associated with orogenic collapse and exhumation decreases the density of the root, increasing buoyancy and driving uplift. It is probably this effect which is primarily responsible for the uplift of coherent eclogite terranes (Dewey *et al.* 1993). Austrheim (1990, 1991) makes similar arguments and points out that the loss of feldspar during eclogitization produces a marked loss in strength in the root, which may contribute to the buffering process by promoting lower crustal flow. Le Pichon *et al.* (1997) argue that thermal relaxation beneath the Tibetan plateau has retrogressed eclogite facies rocks formed earlier during the collision of India with Asia. They calculate that the heating over 40 Ma caused by the thickening of the crust and detachment of the mantle root in an orogen without substantial exhumation will transform eclogite into granulite facies assemblages within the orogenic root and the resultant

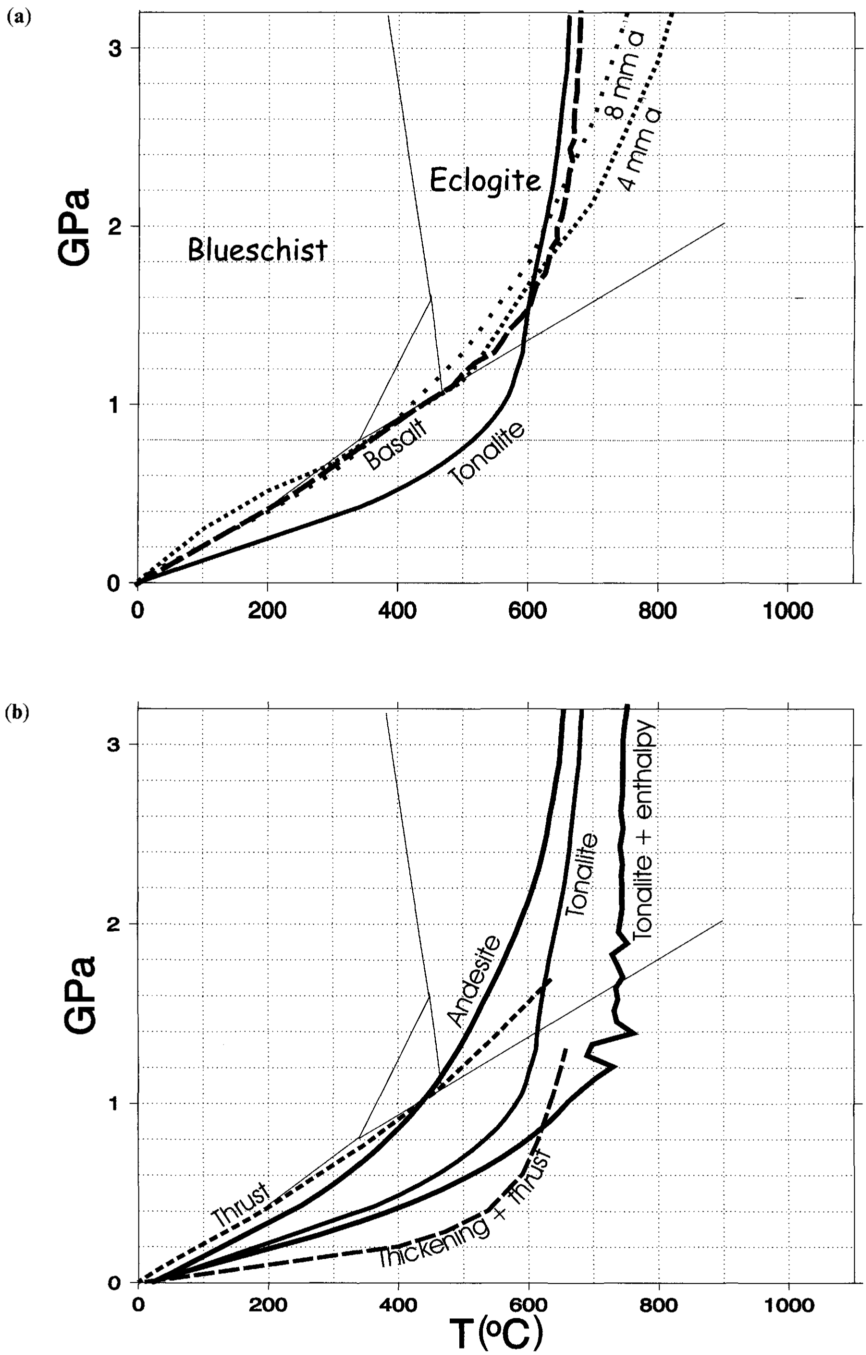

Fig. 3. (**a**) Shows model geotherms for the Alps assuming continental subduction at 4 mm a^{-1} and 8 mm a^{-1} (Bosquet *et al.* 1997) and for a vertical stretch orogen with a crust of granodioritic composition ('granite') and basaltic ('basalt') compositions (this work) after 32 Ma of convergence with phase change to eclogite facies below 40 km. (**b**) Geotherms for different orogenic models. 'Thrust', model where the crust thickens by thrusting and the lithospheric mantle by pure shear (Midgley & Blundell 1997). 'Thickening and thrust', an orogen with vertical thickening and thrusting (Jamieson *et al* 1997). 'Andesite' and 'Tonalite' represent curves for a vertical stretch orogen after 32 Ma of convergence with crusts of andesitic and tonalitic composition respectively with phase change to eclogite facies below 40 km. 'Tonalite + enthalpy', a vertical stretch model after 32 Ma of convergence for a tonalitic crust with phase change that allows for the enthalpy of the reaction *anorthite* ⇌ *garnet* + *kyanite* + *quartz*.

buoyancy gain will account for 2.5 km of the topography of Tibet. Le Pichon *et al.* (1997) also propose that the conversion of Variscan eclogite facies rocks in the Massif Centrale of France to granulite facies during collapse indicates that an equivalent process took place beneath a Tibetan style plateau in central Europe during the late Carboniferous.

The formation of coherent eclogitic terranes

The crust can be thickened during collisional orogeny by a variety of mechanisms. Modelled geothermal gradients for continental subduction (Bosquet *et al.* 1997), a thickening orogenic wedge (Le Pichon *et al.* 1997), underplating during subduction (Huerta *et al.* 1996), vertical thickening and thrusting (Jamieson *et al.* 1997), crustal overthrusting (Midgley & Blundell 1997) or vertical stretch (this work, see below) are consistent with 'medium temperature' eclogite facies conditions, that is temperatures in the order of 550°C at a pressure of 1 GPa (Carswell 1990), forming in the orogenic root (Fig. 3).

The occurrence of coesite, which forms at depths of >90 km and temperatures around 610–700°C (Fig. 4), giving mean geothermal gradients of *c.* 5°C to 7°C per km, leads most authors to assume that continental crust is subducted. This conclusion is supported by the lack of seismic evidence for modern orogenic roots extending to such depths. However, if rocks at the base of the crustal root are all in the eclogite facies then they will have seismic velocities nearer that of the mantle than the crust and it may not be possible to image them (see Austrheim 1990; Hynes & Snyder 1995; Sapin & Hirn 1997, for

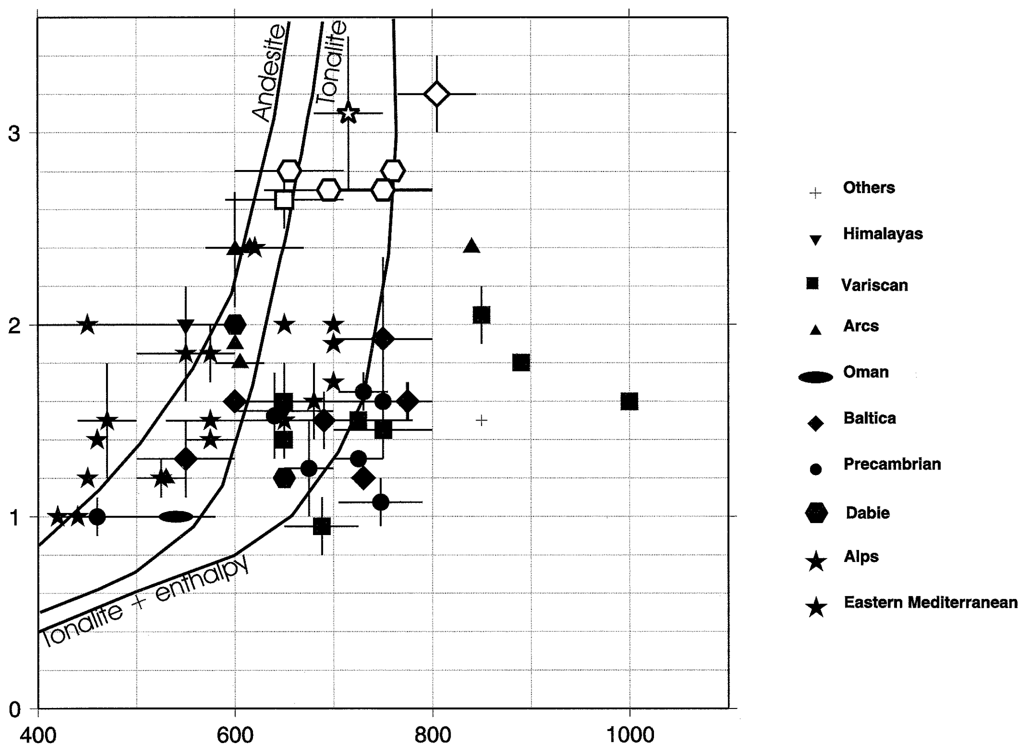

Fig. 4. Plots of pressure temperature estimates for eclogites in collisional orogens. Points for the Alps, Baltica and Dabie (ornaments have hollow centers) at pressures of 2.8 GPa or above represent coesite bearing assemblages. The data plotted is taken from Andersen & Jamtveit (1990), Bernard-Griffiths *et al.* (1993); Boufette & Caron (1991); Camacho *et al.* (1997); Castelli (1991); Chauvet *et al.* (1992); Chopin (1987); Clarke *et al.* (1997); Cliff *et al.* (1998); Cotkin *et al.* (1988); Di Vincenzo *et al.* (1997); El-Din *et al.* (1990); Erdmer *et al.* (1998); Fry & Barnicoat (1987); Gardien (1993); Gomez-Pugnaire & Fernandez-Soler (1987); Guillot *et al.* (1997); Indares (1993); Jamtveit (1987); Kienast *et al.* (1991); Liati & Seidel (1996); Liu & Liou (1995); Mercier *et al.* (1991); Nicollet (1989); O'Brien (1993); Page (1992); Perchuk & Phillippot (1997); Pognante & Spencer (1991); Sanders (1989); Schliestedt (1986); Schmadicke (1991); Wang & Liou (1991, 1993); Wang *et al.* (1989); Waters (1989).

discussion of this problem). If large volumes of continental crust were carried down a subduction zone to a depth of 100 km, buoyancy considerations (see below) require that it had undergone phase changes which made it negatively buoyant. It would also be overlain by the lithospheric mantle of the hanging wall and would be difficult to exhume (see Andersen *et al.* 1991). Most mechanisms appeal to subducted eclogitized crust being accreted onto the hanging wall and rising under its own buoyancy. This suggests that the upper layers of the subducting slab are most likely to be exhumed as they are the most buoyant and structurally highest. Also, such a mechanism is inherently asymmetric and eclogites during collision should be restricted to footwall lithologies. Pre-collision eclogite facies terranes are probably exhumed in

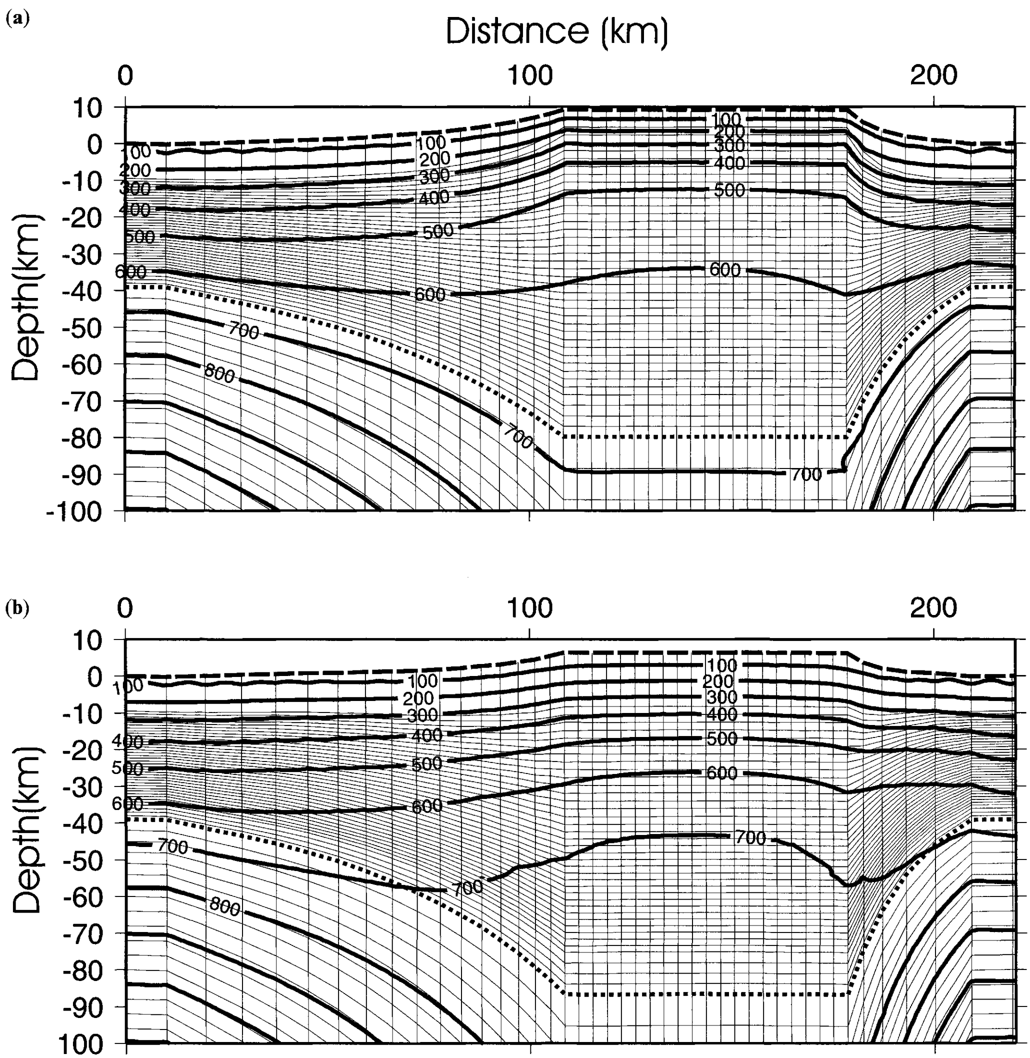

Fig. 5. **(a)** Finite element mesh for a vertical stretch orogen after 32 Ma of convergence at a strain rate of 10^{-15} sec^{-1}. The crust is of a tonalitic composition. No phase transformations take place and a topography of almost 10 km is developed. Thermal contours are drawn using a transient 2-dimensional finite element thermal model. Assumptions are given in Table 2. **(b)** Finite element mesh and transient 2-dimensional finite element thermal model for a lithosphere with the same initial structure as Figure 5a, but which allows for both the volume changes and the enthalpy associated with the *anorthite* $\rightleftharpoons$ *garnet* + *kyanite* + *quartz* reaction. Mass is conserved in each quadrilateral cell. Assumptions are given in Table 2.

this manner. However, in western Norway and NE Greenland (Fig. 1), syn-collisional Silurian eclogites form in coherent terranes in continental basement beneath, or within, nappe piles with opposite senses of transport during convergence: to the east in Scandinavia and to the west in Greenland. The increase in pressure in the WGR is systematic towards the west, the supposed root zone of the orogen (Krogh 1977; Cuthbert *et al.* 2000). These geometries are consistent with eclogite facies metamorphism during homogenous thickening. However, mantle peridotites emplaced along basement cover contacts of the UHP terrane of the WGR are perhaps best explained by continental subduction (Cuthbert *et al.* 2000). Also the CEFTs of Greenland and Norway are offset by some 600 km on a post-collision reconstruction (Fig. 1A), a figure that has been reduced by post-collision sinistral strike-slip. This geometry could be attributed to two continental subduction systems of opposite polarity, but would require the existence of an unreported transform separating these terranes. The geology of CEFTs is complex and greatly modified by the exhumation process, however, at least in the case of the Silurian Greenland–Baltica collision there is evidence that supports the contention that CEFTs can develop in orogenic roots, should a suitable geothermal gradient exist.

The likelihood of achieving a sufficiently low geothermal gradient during collision within an orogenic root formed by vertical stretching is investigated below. Figure 5a presents a transient finite element model for an isostatically compensated, asymmetric orogeny, with crustal thickening by vertical stretch and no eclogitic phase transformations within the orogenic root. The advection–conduction heat equation was solved following the method outlined in Ryan & Dewey (1997) and erosion and topography were analysed using the method of Dewey *et al.* (1993). Assumptions made in construction of this figure are given in Table 2. The resultant geothermal gradient at the centre of the orogeny for crusts composed of basalt, andesite and tonalite are plotted on Figures 3a and b along with those obtained by Bosquet *et al.* (1997), assuming continental subduction at 8 mm a^{-1} or 4 mm a^{-1}. The geotherms obtained by Jamieson *et al.* (1997) for thickening and thrusting, and Midgley & Blundell (1997) for an overthrust orogen are plotted on Figure 3b. All geotherms pass through the continental eclogite field, whilst those for continental subduction and vertical stretch geotherms pass through the coesite stability field (Figs 3a, b and 4). The subduction related geotherms (4 and 8 mm a^{-1}) first pass through the blueschist facies and eventually yield higher temperatures at 3 GPa (Fig. 3a).

Table 2. *Assumptions made in the finite element models presented in Figures 5 and 8*

Heat productivity for upper crust	$1.4 * 10^{-6}$ W m^{-3}
Heat productivity for lower crust	$0.7 * 10^{-6}$ W m^{-3}
Heat productivity of mantle	$2.3 * 10^{-8}$ W m^{-3}
Strain rate	10^{-15} sec^{-1}
Crustal thickness	35–40 km
Thermal diffusivity of upper crust	$7.78 * 10^{-7}$ m^2 sec^{-1}
Thermal diffusivity of lithospheric mantle	$7.45 * 10^{-7}$ m^2 sec^{-1}
Thermal diffusivity of asthenosphere	$8.10–11.30 * 10^{-7}$ m^2 sec^{-1}
Time steps	0.5 Ma
Density of crust	2775, 2802 and 2850 kg m^{-3}
Density of eclogitized crust	3080, 3370 and 3560 kg m^{-3}
Density of mantle	3330 kg m^{-3}
Enthalpy of $an \rightleftharpoons gt + ky + qz$	40 KJ $mole^{-1}$
Half-life of conversion to eclogite	5 Ma
Fixed temperature at surface	20°C
Fixed temperature at 240 km	1440°C
Erosion constant	1.0
Heat capacity of crust	$2.5 * 10^6$ J m^{-3}
Heat capacity of mantle	$3.0 * 10^6$ J m^{-3}
Thermal expansion coefficient of the crust	$2.5 * 10^{-5}$ K^{-1}
Thermal expansion coefficient of the	$3.0 * 10^{-5}$ K^{-1}
Compressibility of the crust	10^{-11} Pa^{-1}
Compressibility of the mantle	$8.0 * 10^{-12}$ Pa^{-1}
Percentage melting at threshold temperatures: 675, 700, 740, 780, 820, ≥860°C	15%, 35%, 50%, 60%, 65%, 70%

Their initial lower temperature portion is due to the upper radiogenic crust of the upper crustal wedge not being significantly thickened. The lower temperatures of the vertical stretch model (basalt and tonalite) at 3.0 GPa may be because this geotherm was calculated using a transient, not a steady state solution. Both analyses yield results probably 100–250°C higher at 3.0 GPa than that encountered in subduction zones, such temperatures are typical of 'medium temperature' eclogites formed during continental collision (Carswell 1990). The higher geothermal gradient of the continental (Bosquet *et al.* 1997), as opposed to oceanic subduction, model (e.g. Peacock 1995) is due to the low rates of subduction, the presence of higher concentrations of heat producing elements, and the thickening of the upper crustal material in the hanging wall. The vertical stretch model also produces higher gradients because of progressive thickening of the radiogenic upper crust, plus the relatively low rate of the penetration of the mantle root into the asthenosphere (1.3 mm a^{-1} to 3.2 mm a^{-1} as the model evolved). It is, therefore, possible that coherent eclogite facies terranes can be formed in the hanging wall and the foot wall lithologies of orogens with significant components of vertical thickening.

Role in metamorphism

Recent thermo-mechanical models of orogens (Stüwe 1998; Jamieson *et al.* 1997) show that heating during continent–continent collision due to the thickening of the radiogenic crust is often inadequate to account for the observed metamorphic parageneses. Other sources of heat suggested are frictional heating (Stüwe 1998), advection by igneous intrusions (Dewey & Mange 1999), or accretion of material with relatively high heat productivity to the base of the metamorphic pile (Huerta *et al.* 1996; Jamieson *et al.* 1997). Such models principally investigate the effects of radiogenic heating and cooling by subduction and erosion, and discount the effects of the enthalpy of metamorphic reactions. Eclogite facies phase transformations during convergence affect the crust in the root of an orogen beneath *c.* 40 km, a volume of rock that might exceed that of the non-eclogitized material above. These reactions have negative enthalpy as they tend to reduce the volume for a given mass and, hence, tend to heat the root. The affect of such enthalpy changes on metamorphic heating is modelled (Fig. 5b) by assuming that the root is comprised of anorthite and undergoes the *anorthite* $\rightleftharpoons$ *garnet* + *kyanite* + *quartz* reaction with a half life of 5 Ma when it exceeds depths of 40 km. The reaction is assumed to have an enthalpy of −40 KJ $mole^{-1}$ at NTP, pressure and temperature dependence is taken from Holloway & Wood (1988). The rate of thermodynamic heating is governed by the following equation:

$$\Delta H_{t_{(i+1)}} = \Delta H \cdot (e^{-\lambda \cdot t_i} - e^{-\lambda \cdot t_{(i+1)}})$$

where,

$$\lambda = \ln(2)/t_{1/2}$$

(ΔH), is the total enthalpy; ($\Delta H_{t(i+1)}$), is the aliquot of the enthalpy attributed to time step ($t_{(i+1)}$); and $t_{1/2}$, is the half-life. Other assumptions are as for Figures 3a and 5. Mass is conserved in the finite element grid and the volume of the cells altered appropriately.

The resultant geotherm (tonalite + enthalpy, Figs 3b and 4) passes through the P-T field of most coherent eclogite facies terranes. This phase transformation increases the temperature by some 50°C at 1 GPa, and up to 100°C at 2 GPa, over that of models which do not account for the enthalpy of HP and UHP transformations. The oscillations in the curve at 1.0–1.5 GPa are caused by the instabilities created in the mathematical model used to solve the conduction–advection equation, when the first aliquots of heat of reaction are added to the system. They provide an estimate of the errors inherent in such calculations, which are in the order of ±30°C. Another interesting feature of this model is the near vertical geotherm at pressures between 2 and 3 GPa. This is attributed to the thickening of the cool lithospheric mantle underlying the crustal root, which offsets conductive heating from below making the rocks at 3 GPa colder than they would be in a steady state model. Enthalpic heating in the layers above 2 GPa makes these rocks hotter than the steady state model. The result of these two effects is that there is little temperature change between 2 and 3 GPa. This analysis is not intended to be exhaustive, merely to demonstrate that the enthalpy of metamorphic transformations within a deep crustal root can affect P–T–t gradients and should be incorporated into thermo-mechanical models of collisional orogenies.

The role of fluids

Fluids are essential to many of the prograde and retrograde metamorphic transformations (Austrheim 1990; Pennacchioni 1996; Austrheim *et al.* 1997) involving in the creation and destruction of eclogitic terranes. Evidence from the

Bergen Arcs region of Norway shows that transformations upon the penetration of fluid are fast enough to be associated with seismic activity (Austrheim *et al.* 1996). This suggests that there may be considerable metamorphic overshoot and the availability of fluids, especially in dry basement terranes, may control the formation of eclogite facies assemblages. However, if transformations in the root are limited by lack of fluids, mean crustal density will not increase and excessive topography will develop. Thickening sufficient to develop coesite will produce mean regional elevations of 7 km, rising to 10 km during the collapse phase (figure 7a of Dewey *et al.* 1993). Such topography, which is higher than values recorded today, is unlikely to form as it would be limited by the potential energy gradients it would create (Dewey 1988). Thus the amount of overshoot must either be limited, or the presence of a dry root must inhibit vertical stretch during collision to *c.* 1.7, the value required for a topography of 3 km (Dewey *et al.* 1993), unless the crust is extremely mafic or the asthenosphere of relatively low density. The absence of fluids in rocks at eclogite facies conditions can, therefore, lead to a relatively early cessation of convergence, whilst access to abundant fluids allows continued convergence without developing excessive topography.

Subducting continental crust

Material balance calculations (e.g. Butler 1986; Laubscher 1988; Pfiffner 1992; Marchant & Stampfli 1997) suggests that there is a shortage of deep basement within Cenozoic collisional orogens. This has been attributed to the subduction of continental crust immediately prior to final collision. Although an alternative view is offered by Ménard *et al.* (1991), who suggest that, within the limits of error, convergence was accommodated by thickening of an extended crust and the development of an 50 km orogenic root. Substantial volumes of continental crust can only be subducted if its buoyancy is reduced (McKenzie 1969). Ignoring slab pull and mantle drag forces, it is assumed that the necessary condition for subduction of continental lithosphere is that it should have negative buoyancy with respect to the asthenosphere. Dewey (1988) argues that continental lithosphere becomes neutrally buoyant when the crust (C_z) to overall lithospheric (l_z) thickness ratio is *c.* ≤ 0.16, depending upon the density assumed for the asthenosphere. It may, therefore, be possible to subduct thinned crust, such as a continental margin or lower crust, which has been tectonically stripped of upper crust, as is suggested for the European upper crust in the Alps (Laubscher 1988; Pfiffner 1992).

The mean density of the lithosphere (ρ_l) is given by:

$$\rho_l = \frac{(\rho_c * C_z + \rho_m * (C_z - l_z))}{l_z}$$

where, ρ_c and ρ_m are the mean densities of the crust and mantle, respectively. If ρ_c is increased by eclogite facies metamorphism, then greater thicknesses of crust can be subducted. Figure 6 shows that a mean crustal density of 3000 to 3170 kg m^{-3} is needed for 2/3 of a normal thickness continental crust ($C_z/l_z \simeq 0.2$) to subduct, a value that may not be unreasonable for the Alps (see Laubscher 1988; Pfiffner 1992). Assuming that the crust has a tonalitic composition implies that much of the lower and middle crust of the subducting slab is in eclogite facies. The decrease in strength at the granulite–eclogite transition (Boundy *et al.* 1992) may provide a

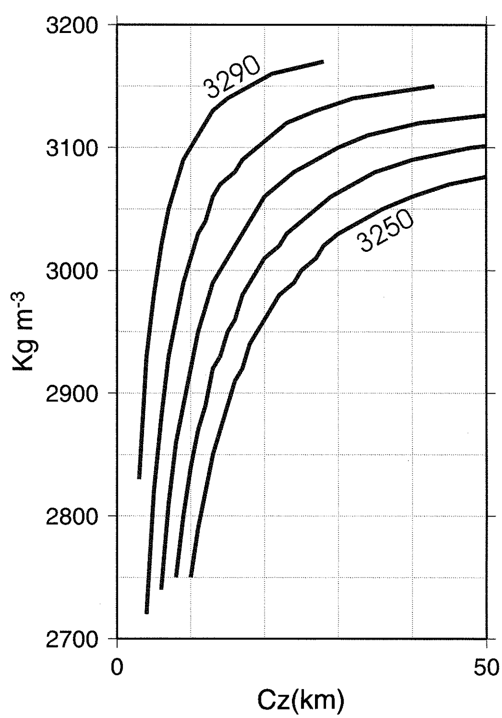

Fig. 6. Curves showing the thickness of crust (Cz) of a given density attached to 80 km of lithospheric mantle that can be subducted. Curves are drawn assuming neutral buoyancy with asthenospheres of density 3250, 3260, 3270, 3280 and 3290 kg m^{-3} equivalent to densities of 3284, 3294, 3304, 3314 and 3324 kg m^{-3} at NTP. The crust is assumed to have a mean temperature of 300°C and the mantle of 950°C.

suitable surface for delamination to take place (Wijbrans *et al.* 1993), allowing the upper crust to rise buoyantly, whilst the eclogitized middle and lower crust are subducted. If the crust is dry and metamorphic fluids cannot penetrate then metamorphic transformation to eclogite facies is unlikely and only the lowermost proportion of the continental crust can be subducted, unless pulled down by an attached oceanic slab (Molnar & Gray 1979). However, this will put the subducting plate into tension and probably result in its failure and the consequent cessation of continental subduction. The thickness of continental crust that can be subducted during the early phases of collision is, at least, partly controlled by basement composition and metamorphic history.

Delamination

Several authors (see Meissner & Mooney 1998; Gao *et al.* 1999 for refs) have pointed out that the average continental crust is more mafic than that of island arcs. The lowermost seismic crustal layer with P wave velocities of greater than 7.2 km sec^{-1} is often missing under orogens. Late tectonic granite composition, including negative Eu anomalies, suggests that the melt was in equilibrium with a dense eclogitic restite phase, which is now not imaged seismically. They, therefore, suggest that much of the basic eclogites or eclogitic restites in the lower crust have delaminated. An appropriate time for this to occur is during orogenesis when the lower crust may be deeply buried, eclogitized and, if basaltic in composition, negatively buoyant. Gao *et al.* (1999) calculated using geochemical constraints that the Dabie–Qinling orogeny has lost between 38 and 74 km of its lower crust beneath Dabie Shan. This is supported by petrophysical and seismic studies (Kern *et al.* 1999) which indicate that large volumes of mafic eclogite do not now exist beneath Dabie Shan, but these results do not exclude the possibility that much of the crust underlying Dabie is an amphibolitized coherent crustal eclogite terrane, such as the WGR.

However, the thermal and isostatic consequences of such delamination must be taken into account. If the lower crust is mechanically attached to the upper plate then its delamination (unless by lateral flow; Meisnner & Mooney 1998) must also remove the lithospheric mantle. A simple isostatic calculation balanced out to the mid-oceanic ridges,

$$h = \frac{(\rho_l - \rho_a)}{(\rho_a)} * l_z - 2.64$$

shows that for a crust of 30–50 km thickness, total instantaneous delamination would produce a topography (h) of some 2–5 km above sea level (Fig. 7). Airy isostasy need only be assumed in this case as the lithosphere would be too weak to have any effective elastic thickness. Also, the exposure of the Moho to asthenospheric temperatures would produce significant volumes of melt. Figure 8 was produced using a two-dimensional transient finite element solution to the conduction–advection equation (see Ryan & Dewey 1997, for the methodology). A 40 km continental crust overlying 90 km of lithospheric mantle was initially allowed to equilibrate for 10 Ma and then all the mantle nodes from 160–320 km horizontally and 40–130 km vertically were to set to 1330°C instantaneously (Fig. 8a). The volumes of melt were then calculated using muscovite dehydration melting models following

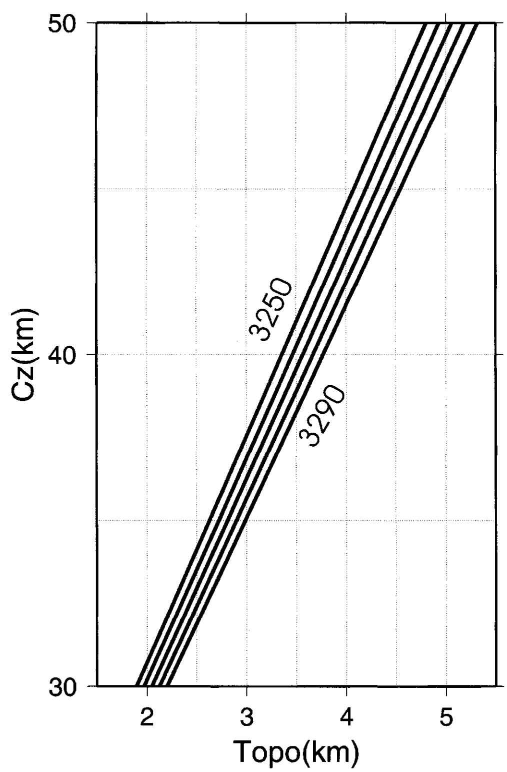

Fig. 7. Model curves showing the topography that would be developed by a crust of tonalitic composition (2810 kg m^{-3}) if the lithospheric mantle were delaminated. Curves are drawn for asthenospheres of density 3250, 3260, 3270, 3280 and 3290 kg m^{-3} equivalent to densities of 3284, 3294, 3304, 3314 and 3324 kg m^{-3} at NTP. The crust is assumed to have a mean temperature of 400°C and the mantle of 1050°C.

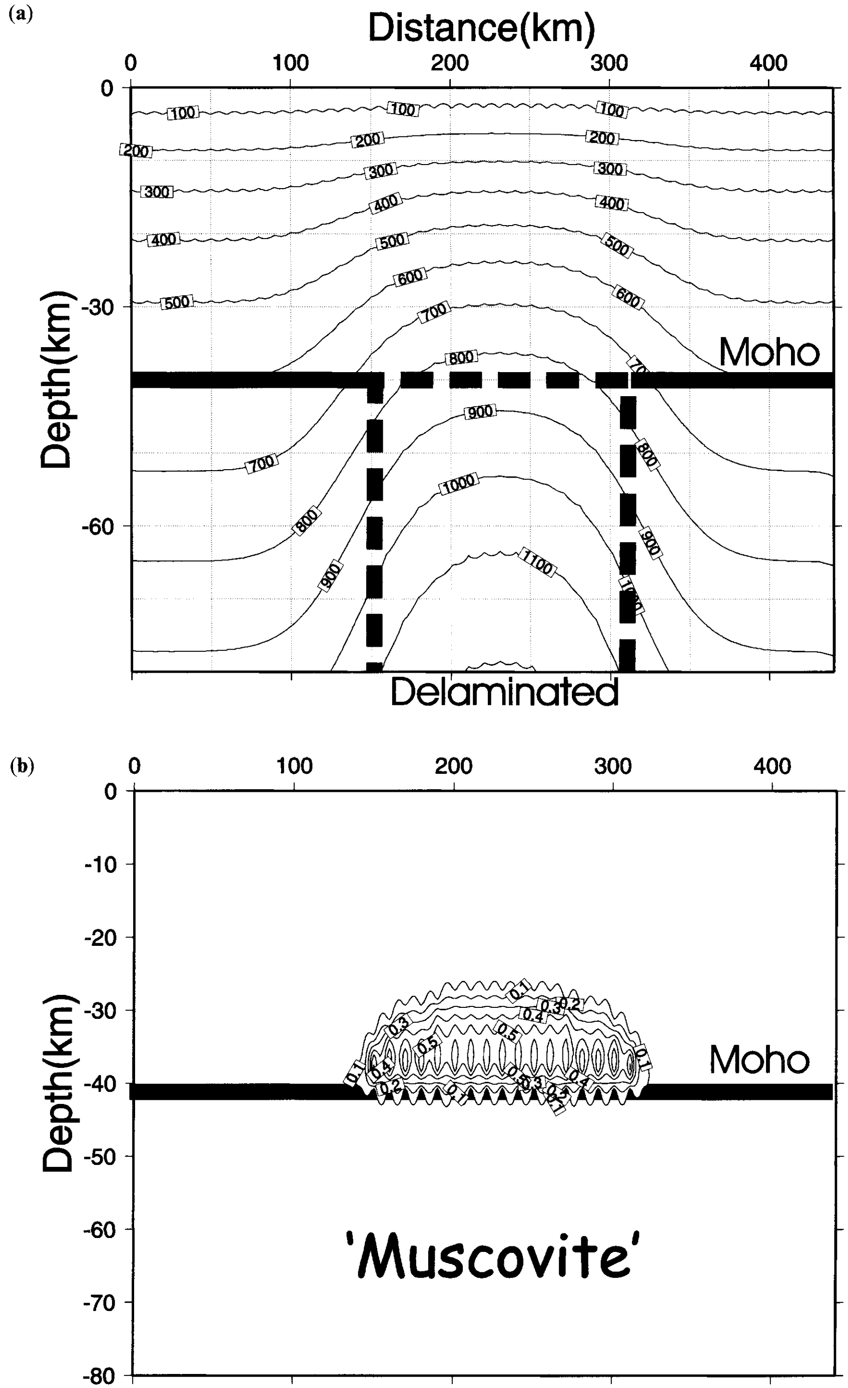

Fig. 8. (**a**) Transient 2-dimensional finite element thermal model 15 Ma after delamination of the lithospheric mantle between 160 and 320 km. The crust is 40 km in thickness. (**b**) Contour plot for the proportion of total melt produced 35 Ma after delamination of the lithospheric mantle between 160 and 320 km. The model assumes muscovite dehydration melting after Zen (1995). Melt is produced up to 35 Ma after delamination. Assumptions are given in Table 2.

the method of Zen (1995). A detailed discussion of this approach is given elsewhere (Ryan & Soper in press). The resultant melt volumes are contoured (Fig. 8b). This analysis suggests that between 20 and 50% of the lower 10–13 km of the lower crust will melt, producing extensive volcanism and plutonism for up to 35 Ma after delamination. This may well have been the case in parts of the Variscides of Europe where some 50% of the crust is either granite or migmatite. This conclusion is supported by the presence of very high temperatures and relatively low pressure eclogites (1000°C, 1.6 GPa, Fig. 4) in the Gfohl Nappe (Medaris *et al.* 1998). However, xenolith evidence suggests that not all of the eclogitized lower crust has delaminated (Mengel 1992). The transition from eclogite to granulite facies in the Massif Centrale suggests thermal relaxation of a thickened crust rather than delamination (Le Pichon *et al.* 1997). Similarly, the formation of early Cretaceous gneiss domes and granitic plutons in Dabie Shan (Wang *et al.* 1989; Faure *et al.* 1998) may be related to delamination triggered by subduction along the margin of the Yangtze Block. The absence of extensive migmatites and granites suggest that this process did not take place in the WGR or NE Greenland.

The nature of the continental crust may control this process. In the Variscides, there is evidence of wide spread mafic vulcanism associated with the break-up of the Gondwana margin in the early Palaeozoic (Floyd *et al.* 2000). Volcanic margins with extensive underplating would have become sites where delamination could take place as eclogitization of significant volumes of basalt would produce an orogenic root that was negatively buoyant with respect to the lithosphere (Fig. 2).

Role in orogenic cyclicity

Ryan & Dewey (1997) suggest that if a substantial volume of eclogite facies rocks of crustal composition remain in isostatic equilibrium with the mantle and are not exhumed during orogenic collapse, they may provide a mechanism for the cyclic re-opening of closed oceans. Such material is likely to have a bulk density of *c.* 3330 kg m^{-3} if it comprises eclogite facies rocks of 50% basalt and 50% granodioritic composition (Table 1), and consequently a P wave velocity in the order of 8 km sec^{-1}. Such material would be imaged seismically as resting beneath the Moho (Griffin & O'Reilly 1987; Hynes & Snyder 1995) and may account for upper mantle reflectivity (Austrheim 1990). These crustal lithologies have a different rheology and a higher heat productivity from that of the mantle they replace. Initially, their lower strength would account for much of the relative weakness of the collapsed orogen compared to its foreland. However, after 300 Ma the radiogenic heat produced by these crustal rocks residing in the mantle amplifies this relative weakness by a factor of 2 to 3 times (Ryan & Dewey 1997) making the old orogen a target for future rifting.

Conclusions

Metamorphic transformations in an orogenic root below 40 km can alter the density of the crust such that it can approach, or even exceed, that of the mantle. This greatly reduces buoyancy during convergence and can, if the correct metamorphic conditions exist, allow vertical stretches to exceed 2.0 without producing excessive topography. The formation of coherent eclogite facies terranes, in which all rocks have undergone HP or UHP metamorphism during convergence may be caused by crustal subduction or by burial within an orogenic root. The broadly coeval CEFTs in the Caledonides of Norway and East Greenland, which lie in nappe stacks of opposite vergence, were probably formed in an orogenic root. Numerical models show that conditions for HP and UHP metamorphism can exist in such a root beneath 40 km and that the associated enthalpy will assist heating within the orogenic pile. Formation of eclogite facies assemblages may assist continental subduction and loss of significant amounts of footwall basement during the early stages of collision. Lithospheres of similar composition and structure may respond differently during convergence depending on the availability of suitable metamorphic fluids.

Eclogitized crustal roots may influence late orogenic history. Whole scale eclogitization of a mafic lower crust, perhaps formed by underplating during earlier rifting, may lead to delamination. Orogenic collapse will then be associated with rapid uplift and high degrees or partial melting and fragmentation of the CEFT. A more gradual retrogression of a less mafic root will tend to buffer subsidence during collapse and assist in exhuming CEFTs. Some continental eclogites, with densities nearer that of the mantle than the crust, may remain beneath the Moho after collapse and cooling of the orogen. They replace mantle rheologies and heat productivities with those of the crust. This leads to a long term weakening of ancient orogenic lithosphere with respect to that of the foreland and may provide an explanation for the Wilson Cycle reopening of sutured oceans.

Research into the controls that crustal roots exert upon the evolution of an orogen is at an early stage. However, existing models suggest that phase transformations within such roots have a significant effect and should be taken into account when modelling collisional orogens. This contribution argues that the existence of a low geothermal gradient within an HP–UHP terrane is not sufficient to prove the mechanism of crustal subduction. Rather, field relationships must be used to establish whether such terranes develop by vertical stretch or crustal subduction, or both. One of the problems in testing such models in ancient orogens is that the process of exhumation destroys much of the collisional fabric within the deep basement to the orogeny. HP and UHP assemblages often only comprise a small percentage of supposed orogenic roots. The coexistence of protolith, HP and UHP assemblages and extensive retrograde assemblages indicate that metamorphic equilibrium was never established. The model presented here requires that at least a significant volume of deep crust underwent transformation to HP and UHP assemblages during collision. Whether or not this happened may depend as much upon the availability of fluids as on P and T conditions. It may be possible to test this model in modern orogens where remotely sensed data exists for that portion of the root still in place. Careful study of exhumed HP and UHP boudins may tell us a great deal about the early history of an orogeny.

References

AMES, L., TILTON, G. R., ZHOU, G. 1993. Timing of collision of the Sino-Korean and Yangtse cratons: U–Pb zircon dating of coesite-bearing eclogites. *Geology*, **21**, 339–342.

AMES, L., ZHOU, G. & XIONG, B. 1996. Geochronology and isotopic character of ultrahigh-pressure metamorphism with implications for collision of the Sino-Korean and Yangtze cratons, central China. *Tectonics*, **15**, 472–489.

ANDERSEN, T. B. 1998. Extensional tectonics in the Caledonides of southern Norway, an overview. *Tectonophysics*, **285**, 333–351.

ANDERSEN, T. B. & JAMTVEIT, B. 1990. Uplift of deep crust during orogenic extensional collapse: a model based on field studies in the Sogn-Sunnfjord region of western Norway. *Tectonics*, **9**, 1097–1111.

ANDERSEN, T. B., JAMTVEIT, B., DEWEY, J. F. & SWENSSON, E. 1991. Subduction and education of continental crust: major mechanisms during continent–continent collision and orogenic extensional collapse, a model based upon the south Norwegian Caledonides. *Terra Nova*, **3**, 303–310.

AUSTRHEIM, H. 1990. The granulite–eclogite facies transition: a comparison of experimental work and a natural occurrence in the Bergen Arcs, western Norway. *Lithos*, **25**, 163–169.

AUSTRHEIM, H. 1991. Eclogite formation and dynamics of crustal roots under continental collision zones. *Terra Nova*, **3**, 492–499.

AUSTRHEIM, H. & MÖRK, M. B. E. 1988. The lower continental crust of the Caledonian mountain chain: evidence from former deep crustal sections in Norway. *Norges Geologisk Undersökelse Special Publication*, **3**, 102–133.

AUSTRHEIM, H., ERAMBERT, M. & BOUNDY, T. M. 1996. Garnets recording deep crustal earthquakes. *Earth and Planetary Science Letters*, **39**, 223–238.

AUSTRHEIM, H., ERAMBERT, M., & ENGVIK, A. K. 1997. Processing of crust in the root of the Caledonian continental collision zone: the role of eclogitization. *Tectonophysics*, **273**, 129–153.

BERNARD-GRIFFITHS, J., PEUCAT, J.-J. & MENOT, R.-P. 1991. Isotopic (Rb–Sr, U–Pb and Sm–Nd) and trace element geochemistry of eclogites from the pan-African Belt: a case study of REE fractionation during high-grade metamorphism. *Lithos*, **27**, 43–57.

BERNARD-GRIFFITHS, J., PEUCAT, J. J. & OHTA, Y. 1993. Age and nature of protoliths in the Caledonian blueschist- eclogite complex of western Spitsbergen: a combined approach using U–Pb, Sm–Nd and REE whole-rock systems. *Lithos*, **30**, 81–90.

BOSQUET, R., GOFFÉ, B., HENRY, P., LE PICHON, X. & CHOPIN, C. 1997. Kinematic, thermal and petrological model of the central Alps: Lepontine metamorphism in the upper crust and eclogitization of the lower crust. *Tectonophysics*, **273**, 105–127.

BOUFFETTE, J. & CARON, J.-M. 1991. Trajets metamorphiques prograde et retrograde dans des eclogites piemontaises (Ouest du Massif du Rocciavre, Alpes cottiennes) (Metamorphic prograde and retrograde P–T paths in Piemontese ophiolites, western Rocciavre Massif, Cottic Alps) *Comptes Rendus – Academie des Sciences, Serie II*, **312**, 1459–1465.

BOUNDY, T. M., FOUNTAIN, D. M. & AUSTRHEIM, H. 1992. Structural development and petrofabrics of eclogite facies shear zones, Bergen Arcs, western Norway: implications for deep crustal deformational processes. *Journal of Metamorphic Geology*, **10**, 127–146.

BRUECKNER, H. K., GILOTTI, J. A. & NUTMAN, A. P. 1998. Caledonian eclogite-facies metamorphism of early Proterozoic protoliths from the North-East Greenland Eclogite Province. *Contributions to Mineralogy and Petrology*, **130**, 103–120.

BUTLER, R. W. H. 1986. Thrust tectonics, deep structure and crustal subduction in the Alps and Himalayas. *Journal of the Geological Society, London*, **143**, 857–873.

CAMACHO, A., COMPSTON, W., MCCULLOCH, M. & MCDOUGALL, I. 1997. Timing and exhumation of eclogite facies shear zones, Musgrave Block, central Australia. *Journal of Metamorphic Geology*, **15**, 735–751.

CARSWELL, D. 1990. Eclogites and eclogite facies: definitions and classification. *In*: CARSWELL, D. A. (ed.) *Eclogite Facies Rocks*. Blackie, Glasgow, 1–13.

CASTAING, C., TRIBOULET, C., FEYBESSE, J. L. & CHEVREMONT, P. 1993. Tectonometamorphic evolution of Ghana, Togo and Benin in the light of the Pan-African/Brasiliano orogeny. *Tectonophysics*, **218**, 323–342.

CASTELLI, D. 1991. Eclogitic metamorphism in carbonate rocks: the example of impure marbles from the Sesia–Lanzo zone, Italian western Alps. *Journal of Metamorphic Geology*, **9**, 61–77.

CHAUVET, A., KIENAST, J. R., PINARDON, J. L. & BRUNEL, M. 1992. Petrological constraints and PT path of Devonian collapse tectonics within the Scandian mountain belt (Western Gneiss Region, Norway). *Journal of the Geological Society, London*, **149**, 383–400.

CHOPIN, C. 1987. Very-high-pressure metamorphism in the western Alps: implications for subduction of continental crust. *Philosophical Transactions of the Royal Society of London*, **321**, 183–197.

CLARKE, G. L., AITCHISON, J. C. & CLUZEL, D. 1997. Eclogites and blueschists of the Pam Peninsula, NE New Caledonia: a reappraisal. *Journal of Petrology*, **38**, 843–876.

CLIFF, R. A., BARNICOAT, A. C. & INGER, S. 1998. Early Tertiary eclogite facies metamorphism in the Monviso ophiolite. *Journal of Metamorphic Geology*, **16**, 447–455.

CLOOS, M. 1982. Flow mélanges: numerical modeling and geologic constraints on their origin in the Franciscan subduction complex, California. *Geological Society of America Bulletin*, **93**, 330–345.

COTKIN, S. J., VALLEY, J. W. & ESSENE, E. J. 1988. Petrology of a margarite-bearing meta- anorthosite from Seljeneset, Nordfjord, western Norway: implications for the P–T history of the Western Gneiss Region during Caledonian uplift. *Lithos*, **21**, 117–128.

CUTHBERT, S. J., CARSWELL, D. A., KROGH-RAVNA, E. J. & WAIN, A. 2000. Eclogites and eclogites in the Western Gneiss Region, Norwegian Caledonides. *Lithos*, **52**, 165–195.

DEWEY, J. F. 1988. The extensional collapse of orogens. *Tectonics*, **7**, 1123–1139.

DEWEY, J. F. & MANGE, M. 1999. Petrography of Ordovician and Silurian sediments in the western Irish Caledonides: tracers of a short-lived Ordovician continent-arc collision orogeny and the evolution of the Laurentian Appalachian–Caledonian margin. *In*: MAC NIOCAILL, C. & RYAN, P. D. (eds) *Continental Tectonics*. Geological Society, London, Special Publications, **164**, 55–107.

DEWEY, J. F., RYAN, P. D. & ANDERSEN, T. B. 1993. Orogenic uplift and collapse, crustal thickness, fabrics and metamorphic phase changes: the role of eclogites. *In*: PRICHARD, H. M., ALABASTER, T., HARRIS, N. B. W. & NEARY, C. R. (eds) *Magmatic Processes and Plate Tectonics*. Geological Society of London Special Publications, **76**, 325–343.

DI VINCENZO, G., RICCI, C. A., PALMERI, R., TALARICO, F. & ANDRIESSEN, P. A. M. 1997. Petrology and geochronology of eclogites from the Lanterman Range, Antarctica. *Journal of Petrology*, **38**, 1391–1417.

EL-DIN, E. A., COLEMAN, R. G. & LIOU, J. G. 1990. Eclogites and blueschists from northeastern Oman: petrology and P–T evolution. *Journal of Petrology*, **31**, 629–666.

ERDMER, P., GHENT, E. D., ARCHIBALD, D. A. & STOUT, M. Z. 1998. Paleozoic and Mesozoic high-pressure metamorphism at the margin of ancestral North America in central Yukon. *Bulletin of the Geological Society of America*, **110**, 615–629.

FAURE, M., LAIN, W. & SUN, Y. 1998. Doming in the southern foreland of the Dabieshan (Yangtze block, China). *Terra Nova*, **10**, 307–311.

FLOYD, P. A., WINCHESTER, J. A., SESTON, R., KRYZA, R. & CROWLEY, Q. G. 2000. Review of geochemical variation in Lower Palaeozoic metabasites from the NE Bohemian Massif: intracratonic rifting and plume-ridge interaction. *In*: FRANCKE, W., HAAK, V., ONCKEN, O. & TANNER, D. (eds) *Orogenic Processes: Quantification and Modelling in the Variscan Belt*. Geological Society, London, Special Publications, **157**, 155–174.

FRY, N. & BARNICOAT, A. C. 1987. The tectonic implications of high-pressure metamorphism in the western Alps. *Journal of the Geological Society, London*, **144**, 653–659.

GAO, S., ZHANG, B., JIN, Z. & KERN, H. 1999. Lower crustal delamination in the Qinling–Dabie orogenic belt. *Science in China, Series D: Earth Sciences*, **42**, 423–433.

GARDIEN, V. 1993. Les reliques petrologiques de haute a moyenne pression des series du Vivarais oriental (Est du Massif Central francais) (High to medium pressure relics in the eastern Vivarais series, eastern part of the French Massif Central). *Comptes Rendus – Academie des Sciences, Serie II*, **316**, 1247–1254.

GILOTTI, J. A. 1993. Discovery of a medium-temperature eclogite province in the Caledonides of northeast Greenland. *Geology*, **21**, 523–526.

GOMEZ-PUGNAIRE, M. T. & FERNANDEZ-SOLER, J. M. 1987. High-pressure metamorphism in metabasites from the Betic Cordilleras (SE Spain) and its evolution during the Alpine orogeny. *Contributions to Mineralogy and Petrology*, **95**, 231–244.

GRIFFIN, W. L. & BRUECKNER, H. K. 1985. Rb–Sr and Sm–Nd studies of Norwegian eclogites. *Chemical Geology*, **52**, 249–271.

GRIFFIN, W. L. & O'REILLY, S. Y. 1987. Is the continental Moho the crust-mantle boundary? *Geology*, **15**, 241–244.

GUILLOT, S., DE SIGOYER, J., LARDEAUX, J. M. & MASCLE, G. 1997. Eclogitic metasediments from the Tso Morari area (Ladakh, Himalaya): evidence for continental subduction during India–Asia convergence. *Contributions to Mineralogy and Petrology*, **128**, 197–212.

HACKER, B. R., IRELAND, T., WALKER, D., SHUWEN, D., RATSCHBACHER, L. & WEBB, L. 1998. U/Pb zircon ages constrain the architecture of the ultrahigh-pressure Qinling-Dabie Orogen, China. *Earth and Planetary Science Letters*, **161**, 215–230.

HIRN, A., LÉPINE, J. C., JOBERT, G. ET AL. 1984*a*. Crustal structure and variability of the Himalayan border of Tibet. *Nature*, **307**, 23–25.

HIRN, A, NERCESSIAN, A., SAPIN, M. ET AL. 1984*b*. Lhasa block and bordering sutures: a continuation of a 500 km Moho traverse through Tibet. *Nature*, **307**, 25–28.

HOLLOWAY, J. R. & WOOD, B. J. 1988. *Simulating the Earth: experimental geochemistry*. HarperCollins, London.

HUERTA, A. D., ROYDEN, L. H. & HODGES, K. V. 1996. The interdependence of deformational and thermal processes in mountain belts. *Science*, **273**, 637–639.

HYNES, A. & SNYDER, D. B. 1995. Deep-crustal mineral assemblages and potential for crustal rocks below the Moho in the Scottish Caledonides. *Geophysical Journal International*, **123**, 1–17.

INDARES, A. 1993. Eclogitized gabbros from the eastern Grenville Province: textures, metamorphic context, and implications. *Canadian Journal of Earth Sciences*, **30**, 159–173.

JAMIESON, R. A., BEAUMONT, C., FULLSACK, P. & LEE, B. 1997. Barrovian regional metamorphism: where's the heat? *In*: TRELOAR, P. J. & O'BRIEN, P. J. (eds) *What Drives Metamorphism and Metamorphic Reactions?* Geological Society, London, Special Publications, **138**, 23–51.

JAMTVEIT, B. 1987. Metamorphic evolution of the Eiksunddal eclogite complex, western Norway, and some tectonic implications. *Contributions to Mineralogy and Petrology*, **95**, 82–99.

KERN, H., POPP, T., JIN, S., GAO, S. & JIN, Z. 1999. Petrophysical studies on rocks from the Dabie ultrahigh-pressure (UHP) metamorphic belt, Central China: Implications for the composition and delamination of the lower crust. *Tectonophysics*, **301**, 191–215.

KIENAST, J. R., LOMBARDO, B., BIINO, G. & PINARDON, J. L. 1991. Petrology of very-high-pressure eclogitic rocks from the Brossasco-Isasca Complex, Dora-Maira Massif, Italian western Alps. *Journal of Metamorphic Geology*, **9**, 19–34.

KROGH, E. J. 1977. Evidence for a Precambrian continental collision in western Norway. *Nature*, **267**, 17–19.

LAUBSCHER, H. 1988. Material balance in Alpine orogeny. *Geological Society of America Bulletin*, **100**, 1313–1328.

LE PICHON, X., HENRY, P. & GOFFÉ, B. 1997. Uplift of Tibet: from eclogites to granulites – implications for the Andean Plateau and the Variscan Belt. *Tectonophysics*, **273**, 57–76.

LIATI, A. & SEIDEL, E. 1996. Metamorphic evolution and geochemistry of kyanite eclogites in central Rhodope, northern Greece. *Contributions to Mineralogy and Petrology*, **123**, 293–307.

LIU, J. & LIOU, J. G. 1995. Kyanite-anthophyllite schist and southwest extension of the Dabie Mountains ultrahigh- to high-pressure belt. *Island Arc*, **4**, 334–346.

MARCHANT, R. H. & STAMPFLI, G. M. 1997. Subduction of the continental crust in the Western Alps. *Tectonophysics*, **269**, 217–235.

MARUYAMA, S. & LIOU, J. G. 1998. Initiation of ultrahigh pressure metamorphism and its significance on the Proterozoic–Phanerozoic boundary. *Island Arc*, **7**, 6–35.

MATHEWS, D. & HIRN, A. 1984. Crustal thickening in the Himalayas and the Caledonides. *Nature*, **308**, 497–498.

MCKENZIE, D. P. 1969. Speculations on the consequences and causes of plate motions. *Geophysical Journal of the Royal Astronomical Society*, **18**, 1–32.

MEDARIS, L. G., MISAR, Z., FOURNELLE, J. H., GHENT, E. D. & JELINEK, E. 1998. Prograde eclogite in the Gfohl Nappe, Czech Republic: new evidence on Variscan high-pressure metamorphism. *Journal of Metamorphic Geology*, **16**, 563–576.

MEISSNER, R. & MOONEY, W. 1998. Weakness of the lower continental crust: a condition for delamination, uplift and escape. *Tectonophysics*, **296**, 47–60.

MENARD, G., MOLNAR, P. & PLATT, J. P. 1991. Budget of crustal shortening and subduction of continental crust in the Alps. *Tectonics*, **10**, 231–244.

MENGEL, K. 1992. Evidence from xenoliths for the composition of the lithosphere. *In*: BLUNDELL, D., FREEMAN, R. & MUELLER, S. (eds) *The European Geotraverse*. Cambridge University Press, Cambridge, 91–102.

MERCIER, L., LARDEAUX, J.-M. & DAVY, P. 1991. On the tectonic significance of retrograde P-T-t paths in eclogites of the French Massif Central. *Tectonics*, **10**, 131–140.

MIDGLEY, J. P. & BLUNDELL, D. J. 1997. Deep seismic structure and thermo-mechanical modelling of continental collision zones. *Tectonophysics*, **273**, 155–167.

MÖLLER, C. 1999. Sapphirine in SW Sweden: A record of Sveconorwegian (-Grenvillian) late-orogenic tectonic exhumation. *Journal of Metamorphic Geology*, **17**, 127–141.

MÖLLER, C. 1998. Decompressed eclogites in the Sveconorwegian (-Grenvillian) orogen of SW Sweden: petrology and tectonic implications. *Journal of Metamorphic Geology*, **16**, 641–656.

MOLNAR, P., ENGLAND, P. & MARTINOD, J. 1993. Mantle dynamics, uplift of the Tibetan plateau, and the Indian monsoon. *Review of Geophysics*, **31**, 357–396.

MOLNAR, P. & GRAY, D. 1979. Subduction of continental lithosphere: some constraints and uncertainties. *Geology*, **7**, 58–62.

NICOLLET, C. 1989. L'eclogite de Faratsiho (Madagascar): un cas exceptionnel de metamorphisme de haute-P–basse-T au Proterozoique Superieur (The Faratsiho eclogite, Madagascar: a record of late Proterozoic high-pressure–low-temperature metamorphism). *Precambrian Research*, **45**, 343–352.

O'BRIEN, P. J. 1993. Partially retrograded eclogites of the Munchberg Massif, Germany: records of a multi-stage Variscan uplift history in the Bohemian Massif. *Journal of Metamorphic Geology*, **11**, 241–260.

PAGE, L. M. 1992. Pressure-temperature-time constraints for the Seve Nappe of the Singis-Tjuoltajaure area central Norrbotten Caledonides, Sweden: implications for early Caledonian marginal Baltica. *Geodinamica Acta*, **5**, 3–16.

PEACOCK, S. M. 1995. Ultrahigh-pressure metamorphic rocks and the thermal evolution of continent collision belts. *Island Arc*, **4**, 376–383.

PENNACCHIONI, G. 1996. Progressive eclogitization under fluid-present conditions of pre-Alpine mafic granulites in the Austroalpine Mt Emilius Klippe (Italian Western Alps). *Journal of Structural Geology*, **18**, 549–561.

PERCHUK, A. & PHILIPPOT, P. 1997. Rapid cooling and exhumation of eclogitic rocks from the Great Caucasus, Russia. *Journal of Metamorphic Geology*, **15**, 299–310.

PFIFFNER, A. 1992. The Alpine Orogeny. *In*: BLUNDELL, D., FREEMAN, R. & MUELLER, S. (eds) *The European Geotraverse*. Cambridge University Press, Cambridge, 180–190.

POGNANTE, U. & SPENCER, D. A. 1991. First report of eclogites from the Himalayan belt, Kaghan Valley (northern Pakistan). *European Journal of Mineralogy*, **3**, 613–618.

RICHARDSON, S. W. & ENGLAND, P. C. 1979. Metamorphic consequences of crustal eclogite production in overthrust zones. *Earth and Planetary Science Letters*, **42**, 183–190.

RYAN, P. D. & DEWEY, J. F. 1997. Continental eclogites and the Wilson Cycle. *Journal of the Geological Society, London*, **154**, 437–442.

RYAN, P. D. & SOPER, N. J. 2001. Modeling anatexis in intra-cratonic rift basins: an example from the Neoproterozoic of the Scottish Highlands. *Geological Magazine*, in press.

SANDERS, I. S. 1989. Phase relations and P-T conditions for eclogite-facies rocks at Glenelg, northwest Scotland. *In*: DALY, J. S., CLIFF, R. A. & YARDLEY, B. W. D. (eds) *Evolution of Metamorphic Belts*. Geological Society, London, Special Publication, **43**, 513–517.

SANDERS, I. S., DALY, J. S., DAVIES, G. R. 1987. Late Proterozoic high-pressure granulite facies metamorphism in the north-east Ox inlier, northwest Ireland. *Journal of Metamorphic Geology*, **5**, 69–85.

SAPIN, M. & HIRN, A. 1997. Seismic structure and evidence for eclogitization during the Himalayan convergence. *Tectonophysics*, **273**, 1–16.

SCHLIESTEDT, M. 1986. Eclogite–blueschist relationships as evidenced by mineral equilibria in the high-pressure metabasic rocks of Sifnos (Cycladic Islands), Greece. *Journal of Petrology*, **27**, 1437–1459.

SCHMADICKE, E. 1991. Quartz pseudomorphs after coesite in eclogites from the Saxonian Erzgebirge. *European Journal of Mineralogy*, **3**, 231–238.

SMITH, D. C. 1984. Coesite in clinopyroxene in the Caledonides and its implications for geodynamics. *Nature*, **310**, 641–644.

STELTENPOHL, M. G., CYMERMAN, Z., KROGH, E. J.; Kunk, M. J. 1993. Exhumation of eclogitized continental basement during Variscan lithospheric delamination and gravitational collapse, Sudety Mountains, Poland. *Geology*, **21**, 1111–1114.

STÜWE, K. 1998. Heat sources of Cretaceous metamorphism in the eastern Alps – a discussion. *Tectonophysics*, **287**, 251–269.

VALASEK, P., MUELLER, S., FREI, W. & HOLLINGER, K. 1991. Results of NFP20 seismic reflection profiling along the Alpine sector of the European Geotraverse (EGT). *Geophysics Journal International*, **105**, 85–102.

WANG, X. & LIOU, J. G. 1991. Regional ultrahigh-pressure coesite-bearing eclogitic terrane in central China: evidence from country rocks, gneiss, marble, and metapelite. *Geology*, **19**, 933–936.

WANG, X. & LIOU, J. G. 1993. Ultra-high-pressure metamorphism of carbonate rocks in the Dabie Mountains, central China. *Journal of Metamorphic Geology*, **11**, 575–588.

WANG, X., LIOU, J. G. & MAO, H. K. 1989. Coesite-bearing eclogite from the Dabie Mountains in central China. *Geology*, **17**, 1085–1088.

WATERS, C. N. 1989. The metamorphic evolution of the Schistes lustres ophiolite, Cap Corse, Corsica. *In*: DALY, J. S., CLIFF, R. A. & YARDLEY, B. W. D. (eds) *Evolution of Metamorphic Belts*. Geological Society, London, Special Publications, **43**, 557–562.

WIJBRANS, J. R., VAN WEES, J. D., STEPHENSON, R. A. & CLOETINGH, S. A. P. L. 1993. Pressure–temperature–time evolution of the high-pressure metamorphic complex of Sifnos, Greece. *Geology*, **21**, 443–446.

ZEN, E. A. 1995. Crustal magma generation and low-pressure high-temperature regional metamorphism in an extensional environment: possible application to the Lachlan Belt, Australia. *American Journal of Science*, **295**, 851–874.

ZHAO, W. J., NELSON, D. K. & THE INDEPTH TEAM. 1993. Deep seismic reflection evidence for continental underthrusting beneath southern Tibet. *Nature*, **366**, 557–559.