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Mantle-lithospheric deformation and crustal metamorphism with some speculations on the thermal and mechanical significance of the Tauern Event, Eastern Alps

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Abstract

Changes in the thickness of the mantle lithosphere beneath convergent orogens through processes such as the pervasive thickening of the lithosphere and the convective thinning of the thermal boundary layer dictate the heat flow through the base of the crust and thus may profoundly influence the thermal evolution of crustal metamorphic terrains. Calculations based on one-dimensional thermal-energy balances show that when mantle lithosphere is substantially thickened and remains intact through the orogenic cycle, then low-temperature facies series, including blueschist and eclogite facies, may be preserved throughout the crust. In contrast, crust thickened above attenuated mantle lithosphere will develop much higher-temperature facies series. Potential-energy arguments suggest that changes in the thickness in the mantle lithosphere induce changes in the elevation and the potential energy stored within the lithosphere. Therefore, the response to mantle-lithospheric deformation observable in crustal metamorphic terrains should not only be recorded in the thermal regime but also in changes in the incremental strain history in as much as it reflects the force balance operating within the orogen. For example, mantle-lithospheric thinning beneath thickened crust may cause an increase in crustal temperatures synchronous with the termination of convergent strain, or, for large reductions in mantle-lithospheric thickness, with the onset of extensional deformation. Such histories are recorded in a number of young metamorphic terrains, for example the Eastern Alps which may provide an important thermal and isostatic record of the mantle-lithospheric response to convergent deformation. Here, we show how application of these ideas may lead to new insights into some outstanding problems concerning the thermal and mechanical evolution of the Eastern Alps during the Tertiary Tauern Metamorphic Event.

1. Introduction

It is now widely recognized that the response of the mantle lithosphere to deformation exerts a profound influence on the thermal and mechanical evolution of orogenic belts (McKenzie, 1978; Bird, 1978; Houseman et al., 1981; England and Houseman, 1988, 1989; Molnar and Lyon-Caen, 1988; Sandiford and Powell, 1990; Zhou and Sandiford, 1992). Consequently, the thermal-energy budget of crustal metamorphic terrains formed in orogenic belts is likely to reflect at least in part the response of the underlying mantle lithosphere (Houseman et al., 1981; Sandiford and Powell,

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1990). Considerable progress has been made towards understanding the thermal perturbations in orogens by solving the heat equation subject to boundary conditions incorporating the deformation and denudation history of the orogen (e.g., Oxburgh and Turcotte, 1974; Bickle et al., 1975; England and Richardson, 1978; England and Thompson, 1984). While these models have been used to great effect in explaining the thermal perturbations responsible for many "Barrovian" metamorphic terrains, they generally have done so by only considering the contribution of the crustal heat sources and neglecting the role of the mantle lithosphere. For example, many studies of the thermal evolution of orogenic belts have utilized models based on a lower-boundary condition of constant heat flux at a level appropriate to the base of the crust at the beginning of deformation (see discussion in England and Thompson, 1984; Sandiford and Dymoke, 1991). As shown by Sandiford and Dymoke (1991) this somewhat ad hoc treatment imposes a unique thickness evolution for the mantle part of the lithosphere once the crustal deformation is prescribed (Fig. 1) and thus excludes a range of other possibilities for mantle-lithospheric deformation (see Sect. 2). The aim of this paper is to show how the explicit consideration of the mantle-lithospheric deformation may provide important constraints on both the thermal and mechanical evolution of orogenic belts. In doing so we will use the Eastern Alps as a case study, focusing on some hitherto enigmatic and controversial aspects of the Tauern Event. In our discussion we will also question the universal validity of some (entrenched) beliefs that have arisen amongst metamorphic geologists as a consequence of accepting, albeit unwittingly, the conclusions of the thermal models which impose a *constant depth–constant heat flux* lower-boundary condition.

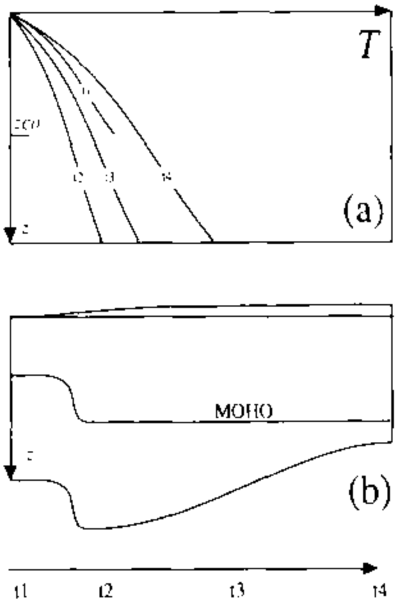

Fig. 1. Schematic illustration of the thermal and the implied thickness evolution of the mantle lithosphere in homogeneously thickened continental crust in models assuming a constant heat flux bottom boundary condition. (a) Temperature–time diagram; *t1* signifies the initial temperature distribution prior to crustal thickening (the geotherm is only shown to the depth of the Moho, z_{co}); *t2* is the temperature distribution just after thickening; *t3* is at a time during the heating process; *t4* is near the new thermal equilibrium geotherm if erosion is neglected. (b) Implied thickness evolution of crust and mantle lithosphere during this process. The bottom line is the lithosphere–asthenosphere boundary, the top line is the surface elevation (exaggerated) above the initial reference elevation (straight line). Note that at time *t2* the crust is thickened without thickening of the underlying lithosphere because the temperature gradient at the Moho is kept constant and the bottom of the lithosphere is defined by an isotherm. Subsequently, the mantle part of the lithosphere must thin continuously because the heating crust dictates a continuously increasing *T*-profile in the mantle lithosphere. This also implies a constant rise in surface elevation. In studies that have attempted to explain the succession of Eo-Alpine and Tauern metamorphic events with such a model, *t2* equates to approximately 65–85 Ma in age and *t4* to approximately 25–35 Ma in age.

In order to provide the basis for the quantitative treatment of the thermal and isostatic consequences of mantle-lithospheric deformation, we begin with a summary of the geological evidence pertaining to the behaviour of the mantle lithosphere during orogenesis. In the development of the quantitative treatment of the thermal and isostatic consequences of mantle-lithospheric deformation we closely follow the approach developed by Sandiford and Powell (1990) and, more recently, Zhou and Sandiford (1992), wherein the interested reader can find more detailed accounts.

2. Constraints on mantle-lithospheric deformation in continental orogens

Interest in the behaviour of the mantle lithosphere during continental orogenesis stems back at least to the late 1970's (McKenzie, 1978; Bird, 1978, 1979a,b), following in part from the realization that convective instability beneath the oceanic lithosphere may provide an explanation for the observed breakdown in the age–depth–heat-flow relation in old oceanic lithosphere (Parsons and McKenzie, 1978). McKenzie (1978) argued that the close proximity of essentially coplanar thrusting and normal faulting in northwestern Greece and Albania may be maintained by the sinking of a lithospheric mantle blob initiated by a convective instability within thickened mantle lithosphere beneath the zone of active thrusting, while Bird (1978, 1979b) suggested that detachment of the whole mantle lithosphere as a discrete entity may have occurred beneath the Himalaya and the Colorado Plateau. In perhaps the most important theoretical development to date, Houseman et al. (1981) described numerical models designed to evaluate the response of a thermally stabilized mantle lithosphere thickened in a zone of continental convergence and showed, for an appropriate parameter range, that after some finite thickening the mantle lithosphere may undergo convective thinning at rates that are rapid compared with the time-scales appropriate to the evolution of continental convergence zones. The implications of this work (Houseman et al., 1981) have only recently been widely appreciated, largely following the realization that lithospheric detachment may provide the mechanism for dramatically increasing the potential energy stored in the overlying orogen (England and Houseman, 1989; Molnar and Lyon-Caen, 1988; Sandiford and Powell, 1990). The model has obvious appeal for the Tibetan Plateau, where the recent onset of extensional deformation reflects a dramatic increase in potential energy, evidently attendant with rapid uplift, as would be expected to result from the convective thinning of a previously thickened mantle lithosphere (England and Houseman, 1988, 1989; Molnar, 1989). It has also been invoked for the Basin and Range Province (Sonder et al., 1987; Gans et al., 1989).

While the discussion above implies that the vertical component of the strain field in the mantle lithosphere and in the crust may vary considerably during orogeny, the precise constraints on how this "decoupling" occurs, are poorly understood, in part because of the difficulties involved in modelling convective processes in the interior of the Earth, and in part because until recently there has been little understanding of the types of observations needed to constrain the response of the mantle lithosphere beneath orogens. If the mechanism suggested by Houseman et al. (1981) is appropriate, then it seems likely that at any given time the thickness of the mantle lithosphere may vary considerably in a lateral direction. For example, Molnar (1989) has suggested that in the Himalaya–Tibet orogen convective instability in the mantle lithosphere leading to substantial thickening of the lithosphere beneath the Karakorum is accompanied by simultaneous thinning of the mantle lithosphere beneath the northern part of the Tibetan Plateau. Similar spatial and temporal variations in convective instability have been suggested for the Basin and Range Province (Gans et al., 1989). Thus, one possibility with interesting thermal ramifications that arises from the mechanism suggested by Houseman et al. (1981) is that the onset of convective instability in the lower lithosphere may lead to significant lateral variations in mantle-lithospheric thickness, with thickening of the lithosphere beneath the part (or parts) of an orogen where convective instability is initiated inducing the excision of the mantle lithosphere at variable rates beneath other parts of the orogen.

Finally, the discussion so far is prejudiced in as much as all models pertaining to mantle-lithospheric deformation cited above assume that the mantle lithosphere is thermally stabilized and therefore more dense than the subjacent convective asthenosphere. Indeed, it is exactly this unstable density stratification that allows the possibility for convective instability. While this is likely for the lower thermal boundary layer of the lithosphere, it may not necessarily be the case for the mechanical boundary layer where metasomatism by the accumulation of small melt fractions derived from the subjacent convective asthenosphere may significantly alter the density structure (e.g., McKenzie, 1989). The importance of this is profound; a chemically stabilized lithosphere will have little potential for convective thinning, and may behave in a way that differs considerably from the lithosphere modelled by Houseman et al. (1981), with convergent deformation potentially generating a thickened mantle-lithospheric root which persists well beyond the active life of the convergent zone depending on how rapidly heat is supplied to the base of the lithosphere by the convective motion in the subjacent asthenosphere.

3. Thermal consequences of lithospheric deformation

In order to quantify the thermal consequences of crustal deformation many previous workers have used models which assume a constant depth–constant heat flux boundary condition at the bottom of the crust to solve the thermal-energy balance (e.g., England and Thompson, 1984). However, these models imply that the mantle lithosphere responds uniformly during the orogenic process. Following the discussion in the previous section, it seems more appropriate that formulations of thermal-energy balances for the evolution of orogenic belts should allow that the vertical strain of the mantle lithosphere may be strongly decoupled from that of the overlying crust, and the practical question arises as to how to treat the basal boundary condition. Two alternative strategies allow consideration of such possibilities: (1) a *variable depth–constant heat flux* lower-boundary condition, and (2) a *variable depth–constant temperature* lower-boundary condition.

The *variable depth–constant heat flux* description of the lower boundary assumes that heat supply from beneath the lithosphere remains constant through the orogenic process and has been employed by England and Thompson (1984) who, in their defense, argued "we know of no evidence that such a change (in heat flux) from the mantle should occur, so we have not investigated it". However, models appropriate to convection within the Earth clearly show that the convective heat flux at the base of the lithosphere depends upon the location with respect to the convective motion (McKenzie et al., 1974; Davies, 1988), and since the argument for convective instability of a thickened lithosphere implies a coupling between lithospheric deformation and convective motion, there must be considerable lateral variations in the heat supplied to the base of the lithosphere within orogenic belts if convective instabilities are allowed. Specifically, the convective heat flux should be enhanced in regions where mantle lithosphere is actively thinned compared with regions where the convective instability and downwelling are initiated. Moreover, the use of a *variable depth–constant heat flux* lower-boundary condition again only implicitly addresses the question of mantle-lithospheric deformation since the base of the lithosphere is defined in nature by an isotherm (presumably through the temperature-dependent rheology of mantle peridotite). Here it is our primary objective to explicitly address the influence of variable mantle-lithospheric deformation geometries and we prefer therefore to employ a *variable depth–constant temperature* lower-boundary condition. It is important to realise, however, that in treating the base of the lithosphere as an isotherm, T_l = 1280°C, the heat flux through the mantle lithosphere must also vary through the orogenic cycle in a way which is unlikely to maintain constant heat flow at any particular depth, including the base of the lithosphere. We will assume that at any time the topographic gradients at the crust–mantle boundary (the Moho) and at the bottom of the lithosphere are small so that heat transfer occurs only in the vertical dimension and the thermal effects of the deformation can be approximated by the one-dimensional diffusion–advection equation. While this assumption may well be invalid for real orogenic belts, it suffices here for the purposes of illustrating the first-order thermal consequences of different mantle-lithospheric deformation geometries.

The appropriate choice of thermal parameters to model the thermal evolution of continental orogens is not well-constrained because of our poor understanding of the distribution of conductivity and heat sources within the continental lithosphere. In the steady-state, where crustal thickness tends to ~ 35 km, the surface heat flow, q_s, is typically around 65 mW m^{-2} and while there is a large natural variation, this value can be used to provide an approximate constraint on "acceptable" parameter ranges for heat source distributions and conductivities (Sandiford and Dymoke, 1991). We note that Menard and Molnar (1991) suggest relatively thin thicknesses for the Alpine crust prior to collision of the eastern Alps and, with foresight to our application below, we evaluate the parameter space for a range of comparably thin crustal thicknesses. Since it is generally recognized that deep crustal metamorphic rocks have low heat production rates and mantle rocks have negligible heat production, we consider only heat source distributions which are confined to the crust. In order to keep the analytical treatment simple we only consider crustal heat source distributions which are constant within the crust or depend analytically on depth. Following Lachenbruch (1970) we employ an inverse exponential dependence that can be characterized by the length scale, h_r, for the depth-dependent description. Then, the heat production, $H_{(z)}$, at any depth, z, within the crust is related to the surface heat production, H_s, by:

$$H_{(z)} = H_s \exp(-z/h_r) \tag{1}$$

In order to explore the consequences of various mantle-lithospheric thicknesses on the thermal evolution of the lithosphere, we will assume deformations which, on the scale of the lithosphere, are homogeneous. We therefore introduce a dimensionless form of the length-scale defined here as $h'_r = h_r / z_c$, (where z_c is total crustal thickness) which remains invariant through the deformation.

For the *potential-temperature* arguments developed below we have found it useful to parameterize the surface heat source term in terms of the reference surface heat flow, q_{s0} (Sandiford and Dymoke, 1991):

$$H_s = \frac{e^{1/hr'}(T_l k - q_{s0} z_{l0})}{z_{c0} h'_r [h'_r z_{c0}(e_r^{1/hr} - 1) - z_{c0} + z_{l0}(1 - e^{1/hr'})]} \tag{2}$$

where z_{c0} and z_{l0} are the thicknesses of the crust and the lithosphere in the reference condition (that is, prior to any deformation). We assume a range of conductivities for the continental lithosphere of $k = 2$ to 3 W m^{-1} K^{-1} (for discussion see England and Thompson, 1984) and assume that conductivity is both temperature- and depth-independent. While this approach is certainly not defensible given our present state of knowledge concerning both the temperature and mineralogical dependence of thermal conductivity, it does follow common practice and has the virtue of allowing a simple analytical treatment so that other controlling parameters can be explored (such as the thickness of the mantle lithosphere). We assume a reference lithosphere, of thickness of $z_{l0} = 126$ km, which is thermally stabilized in the sense of McKenzie and Bickle (1988) extending to the isotherm $T_l = 1280$°C.

As described by Sandiford (1989) and Sandiford and Powell (1990), deformation in orogenic belts where the vertical component of the strain field may be decoupled between the crust and mantle by processes such as convective thinning of the mantle lithosphere is readily portrayed in f_c–f_l space where f_c and f_l represent the thickening factors on the scales of the crust and lithosphere, respectively. The utility of this parametrization is that the thermal and isostatic consequences of various lithospheric deformation geometries can be readily evaluated (Sandiford and Powell, 1990).

We define the *potential thermal structure* as the thermal structure that would be attained at thermal equilibration for an arbitrary deformation that can be represented by a set of f_c–f_l coordinates. In terms of the surface heat production, H_s, the potential temperature at any given depth, T_p, is given by:

$$T_p = T_s + \frac{H_s h_r^2}{k} - \frac{H_s h_r}{k}\left(h_r e^{z/hr} + z e^{f_c z c_0/hr}\right) + z\left(\frac{T_l - T_c}{f_l z_{l0} - f_c z_{c0}}\right) \tag{3}$$

where z is the depth in the crust at which potential temperatures are calculated, and T_c the temperature at the Moho, which is given by:

$$T_c = \frac{T_l f_c z_{c0} e^{f_c z_{c0}/h_r}}{f_l z_{l0}} + \left\{ H_s h_r f_c z_{c0} \left(h_r e^{f_c z_{c0}/h_r} - h_r - f_c z_{c0} \right) + H_s h_r f_l z_{l0} \left(h_r + z_c - h_r e^{f_c z_{c0}/h_r} \right) \right\} \times \left\{ k f_l z_{l0} e^{1/h_r} \right\}^{-1} \quad (4)$$

where f_c and f_l are given by the ratios of the deformed crust and lithospheric thicknesses, z_c and z_l, to the reference crust and lithospheric thickness, z_{c0} and z_{l0}, respectively:

$$f_c = \frac{z_c}{z_{c0}} \quad (5)$$

$$f_l = \frac{z_l}{z_{l0}} \quad (6)$$

In order to track the *potential* thermal evolution of individual rocks as they are incorporated into, and exhumed from, zones of thickened crust, it is useful to use a normalization factor, p, that relates the depth to the total crustal thickness, so that:

$$z = z_c p \quad (7)$$

By expressing depths in terms of constant p (< 1.0), the potential temperatures can be shown for specific material points within the crust assuming homogeneous deformation on the scale of the crust. Thus, the potential Moho temperature is set by substituting Eqs. (5) and (7) into Eqs. (2) and (3) and setting $p = 1$. Equivalent (and simpler) expressions may be derived for the assumption of constant heat production throughout the crust or no heat production (Zhou and Sandiford, 1992). The typical form of potential temperatures at various crustal levels ($p = 1.0$ and $p = 0.5$) are shown in Figs. 2a and 2b and Figs. 2c and 2d, respectively. Theses figures are drawn for constant heat production (i.e., $h'_r \rightarrow \infty$) in the crust for two different initial crustal thicknesses that span the range believed to be appropriate to the Eastern Alps (e.g., Menard and Molnar, 1991). Note that, although the absolute magnitude of the Moho temperature variation is dependent on the initial thermal parameters, the qualitative shape of the contours is similar. The influence of the mantle lithosphere is readily appreciated with reference to Fig. 2 for any given crustal thickness (i.e., a line of constant f_c). Potential temperatures decrease with increasing f_l because of the cooling influence of the mantle lithosphere. Because of the potential sensitivity of the results to the assumptions of the thermal parameters, we have explored the results for a large range of thermal parameters in Fig. 3. It may be seen that the profound influence of the mantle-lithospheric

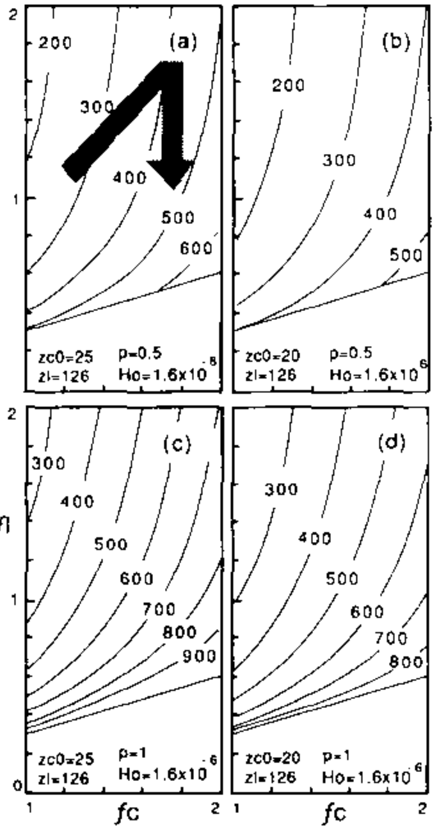

Fig. 2. The $f_c - f_l$ plane contoured for potential temperature in °C and at half crustal depth (a and b) and at the Moho (c and d), for a reference lithosphere of 126 km thickness with two different initial crustal thicknesses of 25 km (a and c) and 20 km (b and d). Heat production, H_0, is constant in the crust and zero in the mantle part of the lithosphere. The shaded arrow in (a) is a deformation path discussed in Fig. 6 and below

thickness is evident for a wide range of plausible reference thermal parameters for the crust (Fig. 3).

It is important to realise that the *potential-temperature* arguments discussed above and shown in Figs. 2 and 3 only pertain to thermal equilibrium. Rapid changes in mantle-lithospheric or crustal thickness at rates appropriate to the construction and demolition of orogenic belts, through processes such as deformation and erosion, will obviously not evolve under conditions of thermal equilibrium. Recognising this, the utility of the *potential-temperature* arguments is that they provide a limiting bound on the range of thermal states that are attainable as the lithospheric-scale conductive response to the deformation, and therefore provide an simple analytical tool for exploring the plausible limits to the response to the lithospheric-scale deformation.

The arguments and calculations presented in Figs. 2 and 3 and the previous section, highlight the fact that mantle-lithospheric deformation may profoundly influence the thermal structure of the overlying orogen. The thermal response reflects the fact that the mantle lithosphere controls the heat flux into the crust evidenced by the inverse correlation between f_l and potential temperature (for a given f_c). Indeed, the effective refrigeration provided by mantle-lithospheric thickening may be much greater than suggested by the potential-temperature arguments because the thickened lithosphere may not thermally equilibrate on the orogenic time-scale. Thickening of the mantle lithosphere during convergence not only

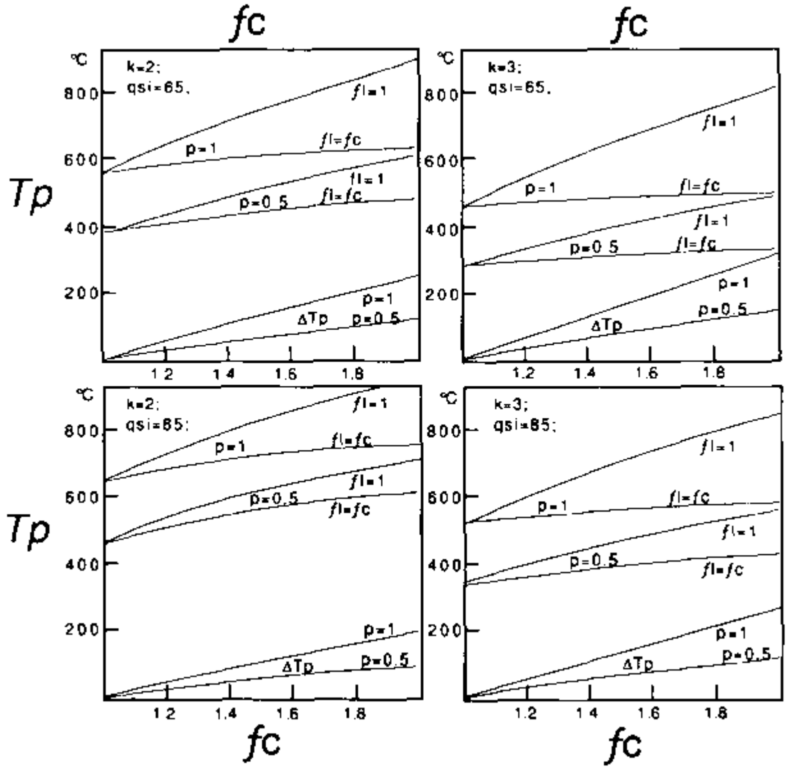

Fig. 3. The potential temperature at the Moho ($p = 1$) and at mid-crustal levels ($p = 0.5$) as a function of a large range of thermal parameters appropriate to the crust. In particular, potential temperatures are shown for thermal conductivities $k = 2$ and $k = 3$ W m^{-1} K^{-1} and two different surface heat flows with $q_{si} = 65$ mW m^{-2} and $q_{si} = 85$ mW m^{-2}. On each diagram, potential temperatures are shown for lithosphere thicknesses corresponding to $f_l = 1$ and $f_l = fc$. ΔTp is the difference between the potential temperature for $f_l = 1$ and $f_l = f_c$.

allows the possibility of the formation of very low-temperature metamorphic facies series at all levels within the crust, but also these low-temperature assemblages may be preserved as long as the thickened mantle lithosphere persists. As shown by Sandiford and Dymoke (1991) low-temperature facies series, *including blueschist and eclogite facies*, may be preserved throughout the crust if thickened mantle lithosphere remains intact through the orogenic cycle. In contrast, thickened crust above attenuated mantle lithosphere (through processes such as convective thinning of the thermal boundary layer) will develop much higher-temperature facies series. For such conditions (i.e., $f_c \gg f_l$), potential Moho temperatures may exceed melting temperatures raising the possibility that metamorphic temperatures in the shallower crust are significantly boosted through the advection of heat in crustal and mantle-lithospheric melts (Sandiford and Powell, 1991).

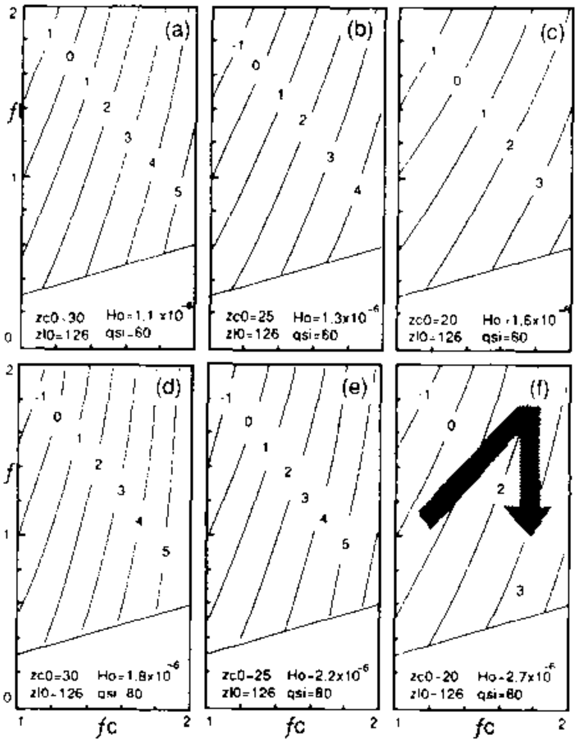

Fig. 4. Surface elevation relative to a reference lithosphere in kilometres mapped onto the $f_c - f_l$ plane for three different initial crustal thicknesses and two different initial surface heat flows. The heat production is assumed to be constant throughout the crust and is scaled to match the assumed values for the surface heat flow. The shaded arrow in (f) is the same deformation path as in (a) and will be discussed in Fig. 6 and below.

4. Isostatic consequences of lithospheric deformation

Just as the potential thermal response to deformation can be mapped onto the f_c–f_l plane, so can the isostatic response as reflected in changes of the surface elevation (e.g., Sandiford and Powell, 1990; Zhou and Sandiford, 1992; Fig. 4) or the potential-energy changes (Fig. 5). Such mappings can be combined with the thermal predictions from above and can be used as an integrated tool to interpret structural and metamorphic field data. Again, it is emphasized that the potential thermal structure used for these calculations, provides only an upper limiting bound to the magnitude of the effect shown in these mappings. Using the relationships of Zhou and Sandiford (1992) we have evaluated the elevation changes for a range of initial thicknesses and thermal parameters. The precise elevation changes accompanying deformation are clearly dependent on the thermal state of the lithosphere through the coefficient of thermal expansion. Therefore, there is some dependence of calculated elevation on the assumed thermal structure (Zhou and Sandiford, 1992), but this is not normally significant (Fig. 4). Since isostasy only specifies a vertical stress balance, changes in the thickness of the crust and mantle lithosphere will also lead to changes in the gravitational potential energy, with variations in lithospheric potential energy giving rise to mechanically important extensional forces (horizontal buoyancy forces). These extensional forces arise only from the lateral density structure and they are mapped relative to the reference lithospheric column for the thermal structure assumed above (Fig. 5).

The magnitude of the extensional forces developed as a consequence of lithospheric deformation is not only sensitive to the way crust and mantle lithosphere are deformed, but also to the thermal evolution, particularly in the upper mantle. In many previous quantitative discussions (e.g., Turcotte, 1983; Molnar and Lyon-Caen, 1988; England and Houseman, 1988; Sandiford and Powell, 1990), the changes in the buoyancy forces attendant with lithospheric deformation have been calculated assuming a linear geotherm (that is, assuming that the lithosphere contains no heat sources). In contrast, Zhou and Sandiford (1992) have provided a quantitative evaluation of the horizontal buoyancy forces in terms of the potential thermal structure. In reality, the gravitational potential-energy state of the orogen is likely to be bounded by these two different assumptions and with time will progress towards the configuration appropriate to the potential thermal structure. The magnitudes of the changes in potential energy associated with the development of large orogens are reflected in the magnitude of the extensional forces per unit length of orogen (up to $5–10 \times 10^{12}$ N m^{-1}). These are comparable with the magnitude of the forces that drive plate tectonics ($2–3 \times 10^{12}$ N m^{-1} for ridge push and a few times 10^{13} N m^{-1} for slab pull) (Fig. 5). Thinning of the mantle lithosphere may contribute as much as $6–10 \times 10^{12}$ N m^{-1} to the potential energy of the overlying orogen (see Fig. 6, which shows the bounds on changes in the potential energy of such a scenario as recently suggested for the Tibetan Plateau by Turner et al., 1993). Therefore, changes in thickness of the mantle lithosphere must clearly have significant mechanical consequences and such a process may leave a distinctive coupled response in both the thermal and incremental strain history of crustal metamorphic terrains. For example, rapid mantle-lithospheric thinning may mediate the change from compressive ($\sigma_{xx} > \sigma_{zz}$) to extensional ($\sigma_{xx} < \sigma_{zz}$) deformation in the overlying orogen (e.g., England and Houseman, 1989; Sandiford, 1989) and this may be accompanied by heating of the

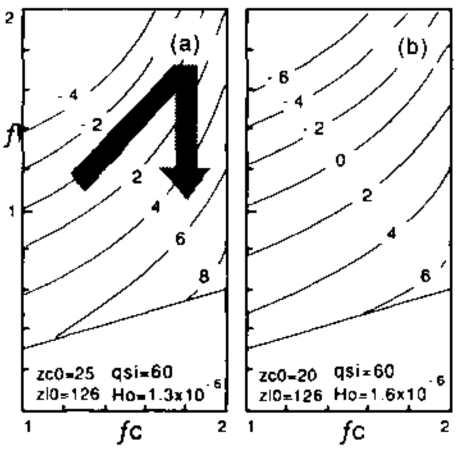

Fig. 5. Gravitational extensional forces (horizontal buoyancy forces) in 10^{12} N m^{-1} mapped onto the f_c–f_l plane for two different initial crustal thicknesses and a surface heat flow of 60 mW m^{-2}. The shaded arrow indicates a deformation path discussed in Fig. 6.

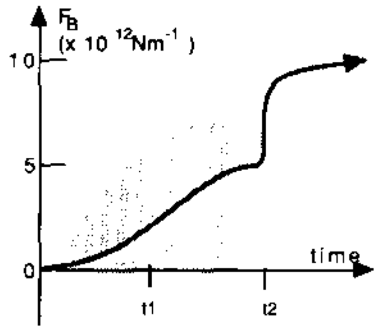

Fig. 6. Gravitational potential-energy changes expressed in horizontal extensional forces, F_B, during an orogeny involving delamination of the mantle lithosphere. The origin of the time axis marks the beginning of convergent deformation, time $t1$ marks the end of deformation, time $t2$ marks the time of delamination of the mantle lithosphere. The shaded regions indicate the large uncertainties that are associated with this process as a consequence of uncertainties in the thermal evolution (Zhou and Sandiford, 1992). The upper bound of the plausible region is provided by the assumption that the orogen evolves in thermal equilibrium with the initial lithosphere (here characterized by a heat production distribution giving a surface heat flow of 65 mW m^{-2}). The lower bound is provided by assuming that the orogen evolves with no thermal reequilibration. In both cases the initial lithosphere is assumed to be in potential-energy and isostatic balance with the mid-ocean ridge system, against which the potential-energy changes may be measured. Assuming that the thermal time constant for the thickened lithosphere is of the order of 100 m.y., the potential-energy evolution for the model outlined here will follow a path similar to the bold line with an increase in F_B of approximately 5×10^{12} N m^{-1} accompanying mantle-lithospheric thinning at time $t2$. The deformation history illustrated here corresponds to the shaded arrows in Figs. 2, 4 and 5.

middle crust to amphibolite facies metamorphic conditions (Fig. 2) as well as topographic development (Fig. 4). In the following section we discuss the Eastern Alps as an example with a unique record for temporally coupled metamorphic and structural processes which may be used to constrain the thickness evolution of both the crust and the mantle part of the lithosphere.

5. Application to the Tauern Event in the Eastern Alps

The discussion in the previous section suggests that the mantle lithosphere has the potential to mediate both the thermal and the mechanical behaviour of the lithosphere during orogenesis. Whilst this realization has lead to the interpretation of a number of present-day geological observations (e.g., Tibet, Basin and Range; Molnar 1989; Bird 1979b), an important question remains as to whether a record of such mantle-lithospheric deformation is preserved in the thermal and kinematic record of crustal metamorphic terrains. Our purpose here is to show that this is not only likely, but that the realization of the importance of the mantle lithosphere may help to resolve some important, and otherwise perplexing, aspects of the evolution of metamorphic terrains. We concentrate our efforts on the Tauern Event in the Eastern Alps, in part because it has provided the impetus for some of the classic studies on the thermal-energy budgets of crustal metamorphism (e.g., Oxburgh and Turcotte, 1974; Bickle et al., 1975; Oxburgh and England, 1980). More recently, significant lateral extension has been recognised to be linked to the thermal events (e.g., Ratschbacher et al., 1989, 1991a,b). We will argue that the consideration of both the kinetic and the thermal constraints point to a significant role of the mantle lithosphere in the evolution of the Eastern Alps. In order to formulate our arguments we begin with an outline of the metamorphic and structural evolution of the Eastern Alps in which we show that some aspects of the spatial and temporal distribution of metamorphic and

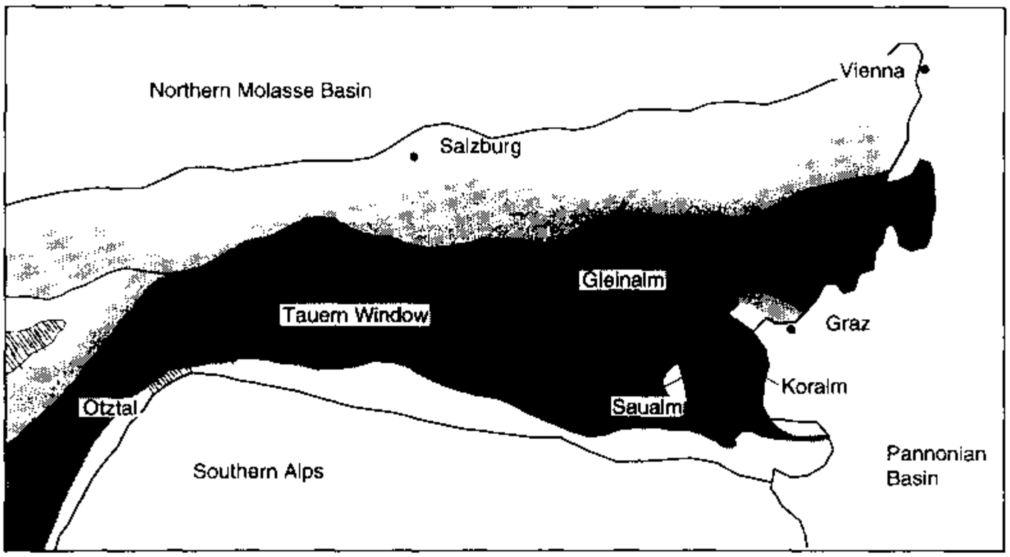

Fig. 7. A sketch map of the Eastern Alps with localities discussed in the text and a schematic indication of the distribution of the Alpine metamorphic grade (modified after Frank, 1987, and Krohe, 1987). The Tauern Window is shown in the centre of the diagram. All other units not specified such as Southern Alps, Northern Molasse Basin or Pannonian Basin are part of the Austro-Alpine nappe complex. Light shading indicates Eo-Alpine sub-greenschist facies grade metamorphism. Dark shading indicates greenschist facies Eo-Alpine metamorphism. Black indicates Eo-Alpine amphibolite facies metamorphism. The dashed pattern indicates areas affected by mid-Tertiary Barrovian metamorphism

structural events are difficult to reconcile with interpretations that consider thickness changes of the crust alone.

5.1. Background to Eastern Alpine metamorphism

Orogeny in the Eastern Alps occurred as a result of the thrusting of Mesozoic sediments and the Austro-Alpine nappes, over pre-Alpine granitic basement (e.g., Dewey et al., 1973) before 65 Ma (e.g., Cliff et al., 1985). The contact between the basement and cover nappes is exposed in tectonic windows, one of which is the Tauern Window, where early (ca. 65 Ma: Hawkesworth, 1976; Cliff et al., 1985) blueschist–eclogite (Eo-Alpine) assemblages are overprinted, often pervasively, by later Barrovian assemblages of the Tertiary Tauern Metamorphic Event (Cliff et al., 1985; Frank et al., 1987a,b). While the Tauern Event is spatially and temporally well-confined to the region in and around the Tauern Window at a time of 25–35 Ma (Fig. 7), the Eo-Alpine event affected the Austro-Alpine rocks at variable grade and at different times with absolute ages ranging from 65 Ma to > 100 Ma. Nevertheless, the time gap between "Eo-Alpine" and Tauern events has been used to formulate some of the classic conductive heating models used to explain Barrovian type terrains (England and Richardson, 1978; England and Thompson, 1984). In the summary below we focus particularly on the temporal and spatial relationships of these two events. For summaries of the absolute ages and regional evidence for the two events we refer the reader to published summaries and references therein (amongst others: Frank et al., 1987a,b; Genser and Neubauer, 1989; Ratschbacher et al., 1991a,b).

The Tauern Window. P–T–t paths of Penninic rocks and their cover sequences exposed in the Tauern Window comprise an initial heating phase and burial to pressures of 8–10 kbar during the Eo-Alpine (Selverstone et al., 1984; Droop, 1985; Selverstone, 1985; Selverstone and Spear, 1985), followed by decompression (Holland and Richardson, 1979; Selverstone and Spear, 1985; Droop, 1985; Cliff et al., 1985) at increasingly rapid rates throughout the Tertiary. The thermal peak was reached during the Tauern Event about 40 Ma after the early high-pressure peak (Cliff et al., 1985). In the southeast Tauern Window, rapid approximately isothermal decompression immediately post-dated the Tauern Event, giving exhumation rates up to 5 mm a^{-1} between 25 and 16 Ma ago (Cliff et al., 1985). Rocks on the present-day surface did not cool through 380–400°C until they were within 5 km of the surface (Selverstone and Spear, 1985; Selverstone, 1988). In the eastern Tauern Window top-to-the-east extensional transport has been documented to be associated with this event (Genser and Neubauer, 1989). In the western Tauern Window, Barrovian mineral growth during the Tauern Event occurred during a top-to-the-west-directed shallowly aligned extensional deformation resulting in a pervasive mineral lineation and tectonic fabrics, but no major folding (Selverstone, 1988). It was also shown that metamorphic pressures indicate a greater maximum overburden thickness than can be estimated from present-day lateral equivalents of the overburden-forming lithological units (Selverstone, 1985, 1988).

The 40-Ma hiatus separating the Tauern and the Eo-Alpine events was a period of continued deformation and imbrication of the Eastern Alps, but it is important to note that there is a substantial time lag between the two thermal events and that the rapid topography development, exhumation and the onset of lateral extension occurred only in association with the younger Tauern Event.

The Austro-Alpine. Metamorphic studies in the Austro-Alpine nappe pile west and east of the Tauern Window have shown the the following.

(1) Many of the eclogites in the Austro-Alpine west and east of the Tauern Window formed from continental rather than oceanic protoliths and are of Eo-Alpine age (e.g., Miller et al., 1980; Miller and Frank, 1983; Thöni and Jagoutz, 1992). This indicates that the absolute exposure level of many Austro-Alpine rocks (measured in kbar) must be similar to that of the Penninic rocks in the Tauern Window, despite their higher structural position (measured in stratigraphic depth).

(2) Coalification studies on coals from Tertiary intracratonic basins indicate that a very

low-grade overprint related to the Tauern Event can be recognized in the Austro-Alpine with the grade of the overprint increasing with proximity to the Tauern Window (Sachsenhofer, 1991). This indicates that the Tauern Event may be geographically laterally confined and not related to the depth of exhumation of the Penninic rocks.

(3) The temperature–time paths established for parts of the Austro-Alpine indicate that both high-P metamorphism and significant cooling occurred in the Cretaceous (Thöni and Jagoutz, 1992; Neubauer et al., 1995-this issue). Retrograde assemblages in the greenschist and amphibolite facies formed soon after the high-P event and a mid-Tertiary overprint is absent. The relationship of this cooling history to the exhumation history is much less well-established. In the Gleinalm Complex, Neubauer et al. (1995-this issue) suggest that Cretaceous cooling can be correlated with exhumation on the basis of structural arguments. On the other hand, cooling rates calculated for the Koralm Complex suggest that cooling was extremely rapid (Ehlers et al., 1994; Stüwe and Powell, 1994) and cannot be related to exhumation. The direct correlation of cooling and exhumation may therefore be unwarranted and it is possible that large parts of the Austro-Alpine were exhumed much later, for example at the time of exhumation of the Tauern Window. Cretaceous Gosau basins preserved in parts of the Austro-Alpine show that this interpretation cannot be valid for the whole of the Austro-Alpine nappe pile, but it is consistent with sedimentological evidence from the Molasse basins. For example, England (1981) showed that the sediment volumes of post-mid-Tertiary basins surrounding the Alps correspond largely to the erosion volumes predicted by geobarometry implying no significant exhumation of the Eastern Alps prior to the mid-Tertiary.

In the light of these observations and their possible interpretation some specific aspects of the metamorphic evolution of the Eastern Alps may require a new synthesis. These are: (1) the significance of the overprinting of early high-P Eo-Alpine assemblages by Tertiary Barrovian assemblages during a phase of extension and thinning of previously thickened crust; (2) the causes of rapid isothermal decompression, in particular during and following the Tauern thermal event; (3) the apparent absence of appreciable erosion during the ~ 40 Ma long hiatus between the Eo-Alpine and Tauern Event, at least within the Tauern Window but possibly also in the Austro-Alpine; and (4) the relatively restricted occurrence of the Tauern Event to the Tauern Window and the absence of corresponding Tertiary overprint in the Austro-Alpine. In the next section we will discuss these aspects in the light of the discussion in the first part of this paper. It is our aim to demonstrate that this scenario is difficult to reconcile with some existing interpretations and that these conflicts may be circumvented with models which are formulated in terms of both crustal and mantle processes.

5.2. *Thermal and mechanical constraints on the Tauern Event*

One of the intriguing aspects of the geological evolution summarized above is that the Eo-Alpine high-pressure assemblages are widely distributed in the Austro-Alpine crystalline units (e.g., the Koralm—Thöni and Jagoutz, 1992; Ehlers et al., 1994; the Ötztal—Miller and Frank, 1983) but that it is only in the Tauern Window that the Tertiary overprint is recorded on Eo-Alpine high-pressure assemblages. Bickle et al. (1975) argued that the Tauern Event represents the conductive thermal response to Eo-Alpine nappe stacking. In this interpretation, in terms of crustal processes, the 40-Ma hiatus between the Eo-Alpine and Tauern metamorphic events is seen as the conductive response to the crustal thickening, while the lack of appreciable erosion has been interpreted as the consequence of relative subdued topography developed on the dense "eclogitized" nappe pile prior to the Tauern Event (Richardson and England, 1979; England and Holland, 1979). This interpretation has been widely accepted and has formed the basis for the thermal models of England and Richardson (1978) and England and Thompson (1984).

An outstanding problem with this conventional interpretation is that it provides no obvious reason why the Tauern Event heating is associated with the extensional deformation which occurred during the cooling path from this event (e.g., Selverstone, 1988; Genser and Neubauer, 1989) and it does not explain the absence of a Tertiary overprint in the Austro-Alpine west and east of the window. The conductive heating model implies a continuous heating path between Eo-Alpine and the mid-Tertiary and therefore implies a continuous thinning of the mantle lithosphere during this hiatus (Fig. 1). In view of the f_c–f_l arguments presented above we would expect such a thermal response to be accompanied by a continuous topographic evolution and therefore probably also erosion. Substantial topography should have been developed above the Tauern Window in the interval between Eo-Alpine and the Tauern Event and no rapid increase of the denudation in the mid-Tertiary should be observed. This is clearly inconsistent with the observations on the topographic, sedimentological and metamorphic evolution of the range (England, 1981; Cliff et al., 1985; Droop, 1985; Stüwe and Sandiford, 1994).

Recognizing the need for a dynamic interpretation, Ratschbacher et al. (1989, 1991a,b) have suggested that the Tauern deformation reflects lateral crustal extension (and tectonic escape) in front of a rigid, north-moving indenter. They regard the Tauern Window as simply resulting from the exhumation of deeper, and therefore hotter, crustal levels. Their model has found wide acceptance in the recent literature but is difficult to reconcile with the mechanical consequences of indentation since it is difficult to understand how indentation can produce bulk crustal thinning on regionally extensive shallow foliations as documented by Selverstone (1988), Genser and Neubauer (1989) and Wallis et al. (1993). Rather, indentation is likely to lead only to compression or strike-slip faulting, depending on the boundary conditions applying to the orogen (e.g., England and Houseman, 1989). Recent mechanical studies (e.g., Houseman and England, 1986; England and Houseman, 1986; Vilotte et al., 1986) have shown that bulk crustal thinning of previously thickened crust as indicated by decompression associated with flat-lying foliations is mechanically possible only if: (1) other causes for local variations in gravitational potential-energy increase can be invoked; or (2) dramatic rheological changes occur in the actively deforming zone; or (3) the forces supporting the thickened crust are reduced (Houseman and England, 1993). Indeed, even in case of reduction of the driving forces, the potential for extension is dependent on the potential-energy contrast which may not be significant if previous indentation caused homogeneous thickening (Fig. 5).

A further problem with the conductive heating models (Bickle et al., 1975; England and Richardson, 1978; England and Thompson, 1984) relates to the absence of a Tertiary overprint in the Austro-Alpine rocks west and east of the Tauern Window. This has been interpreted as the result of their earlier exhumation from depth (e.g., Frank, 1987; Neubauer et al., 1995-this issue), but we have shown above that a number of problems are associated with this interpretation. For example, it involves a complicated scenario where rocks both to the east and west of the Tauern Window were exhumed, whilst rocks within the window remained at depth. It is also inconsistent with the coalification studies that show a continuous increase in grade for the Tertiary Tauern Event with proximity to the Tauern Window and it is not required in the light of recent data on the temperature–time evolution for at least some parts of the Austro-Alpine (Ehlers et al., 1994) (see above). On the other hand, it has to be noted that the existence of Gosau basins testifies to Cretaceous exhumation of at least some parts of the Austro-Alpine.

5.3. *The case for the mantle lithosphere*

The discussion outlined in the previous section shows that there are a number of observations with respect to the temporal and spatial distribution of the depth changes of rocks in the Austro-Alpine and the Penninic rocks in the Tauern Window that are difficult to reconcile within conductive heating models. Here, it is suggested that the Tauern Event may rather be seen as a temporally and spatially confined event unrelated to Eo-Alpine metamorphism and occurring as a consequence of localised mantle-lithospheric thinning beneath the Tauern Window. This process has been discussed previously and *has* been suggested as a possible trigger of the lateral extrusion by Ratschbacher et al. (1991), but we believe that the importance of this process has not been sufficiently appreciated in terms of its thermal and mechanical consequences. As we have shown, the coupled thermal and isostatic response of such a process in the mantle lithosphere may be sufficient to induce significant changes in thermal regime and mechanical state of the overlying crust (Fig. 6). Details of a possible thickness evolution of the mantle lithosphere for the Eastern Alps will be discussed in the next section.

Thickening of the crust during the Cretaceous deformation is also likely to also have thickened the mantle part of the lithosphere. It is now clear that this deformation did not occur synchronously in all parts of the Eastern Alps, but we summarise it here as "Eo-Alpine", encouraged by the similarity of the supporting evidence in a number of regions: (1) the absence of evidence for significant topography development at this time; (2) the low-temperature high-pressure assemblages formed at this time; and (3) the present excess thickness of much of the Alpine crust and mantle lithosphere (Spakman, 1988, 1989; Aric et al., 1989). Observations from the eastern part of the Austro-Alpine support this interpretation: the sedimentary basins of the Gosau subsided at the same time as the formation of small metamorphic domes like the Gleinalm (Neubauer et al., 1995-this issue). The parallel existence of several of such structures on the scale of tens of kilometres cannot be the result of lithospheric-scale processes and indicates therefore a general elevation-neutral isostatic equilibrium. This can only have existed if the mantle lithosphere underneath that part of the Eastern Alps was substantially thickened. It is therefore indeed not surprising that high-pressure low-temperature facies rocks which were not subsequently heated, are common in many parts of the Austro-Alpine. Such a path is illustrated by the first arrow in Fig. 8 and should be read in conjunction with Figs. 2, 4 and 5.

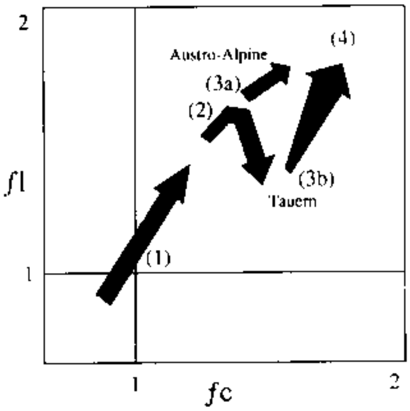

Fig. 8. Sketch of the suggested deformation path for the lithosphere in the Eastern Alps. (1) Prior to continent-continent collision at greater than about 65 Ma. The lithosphere is of normal thickness ($f_l = 1$) and the crust may have been comparatively thin ($f_c < 1$). Homogeneous thickening of the whole lithosphere until the Late Cretaceous is indicated by the path to (2). Then, the region of the Tauern and that further east may have commenced separating paths: dramatic thinning of the lower lithosphere led to increase in elevation, extensional forces and crustal temperatures in the area of the Tauern Window (path to (3b)): minor continued crustal thickening may have occurred in the Austro-Alpine (3a). The last stage of the evolution may have been characterised by repeated rapid thickening of the mantle lithosphere to attain its present thickness (Spakman, 1988, 1989) and is shown by the path to (4).

Such a deformation is unlikely to produce significant surface topography and is therefore consistent with the absence of significant sediment volumes of Eo-Alpine age. We suggest that this event may have been later followed by local and transient thinning of the mantle lithosphere in the region of the Tauern Window but not elsewhere. Such a process may have been responsible for rapid heating in that area during the Tauern Event independently of the events outside the Tauern Window. The process would also invoke a sudden increase in the surface elevation of this region and a dramatic increase of potential energy of this region (Figs. 5, 6). This potential-energy increase may be of the same order as the driving force and may be the cause for the onset of lateral extension. Rapid repeated thickening of the mantle lithosphere must have occurred during continued convergence to attain the present state of thickness (e.g., Spakman 1988, 1989; Aric et al., 1989; Fig. 8).

Within this model it is possible to explain the different thermal histories of the Tauern Window and the Austro-Alpine without the need of invoking substantial differential uplift in various parts of the Alpine chain in the Late Cretaceous. As such other events of the mid-Tertiary, including the onset of sedimentation in the Molasse basins, the onset of lateral extension and the maximum uplift in the Tauern Window may be better explained as isostatic consequences of thinning of the mantle lithosphere in the Tertiary. Our model provides a mechanism for lateral extrusion of the Eastern Alps without invoking dramatic changes in the driving forces. Ratschbacher et al. (1991a) speculated on the possibility of this process as a sudden trigger of the extension process, but we believe that a number of observations including the spatial distribution of the Tauern Event, indicate that such a process is a likely pre-requisite for extrusion to commence.

This model raises the question as to the strict geographical confinement of the thermal effects to the region of the Tauern Window. It is possible that lateral extension compressed the isograds and it may also be possible that thermal conductivity contrasts between Penninic rocks and the Austro-Alpine may have contributed to the effect (Jaupart and Provost, 1985). However, the question remains unresolved. Finally, the model suggested above has a number of important implications for the interpretation of the evolution of the Eastern Alps which should allow testing of the hypothesis. For example, the Eo-Alpine and Tauern metamorphic events are viewed as two fundamentally different processes. Consequently, heating and cooling rates for the two events are independent of each other and it is at least possible that the thermal history *between* the two events was characterised by a cooling phase (Fig. 9).

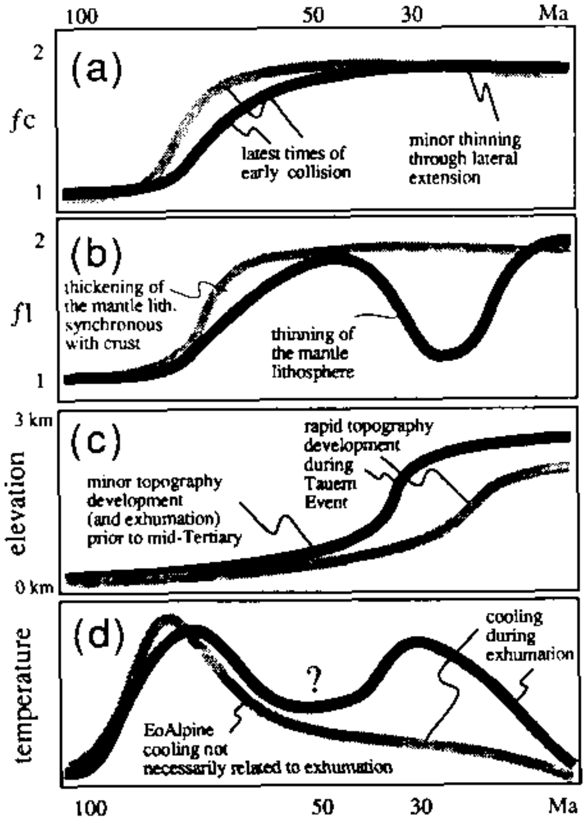

Fig. 9. Possible evolution of surface elevation, crustal thickness, mantle lithosphere thickness and metamorphic grade in the Eastern Alps. In all four diagrams the light-shaded path refers to the Austro-Alpine east of the Tauern Window and the dark path to the Tauern Window itself. Paths on (a) and (b) correspond to those of Fig. 8. Paths on (d) correspond to the expected temperature evolution. It can be seen that our model would allow for a drop in metamorphic grade between Eo-Alpine and Tauern metamorphism. The application of a model after England and Thompson (1984) would require a continuous path connecting the early and late metamorphic peak through slowly rising temperatures.

6. Conclusion

Thickness changes of the mantle part of the lithosphere during orogenesis potentially have a significant influence on the thermal and mechanical evolution of the overlying orogen. If thickening of the crust is accompanied by thickening of the mantle lithosphere, the changes of the thermal state and potential-energy structure of the orogen will be only limited. However, if thickening of the crust is accompanied by attenuation of the underlying mantle lithosphere, then dramatic increases in crustal temperatures, elevation and horizontal buoyancy forces may occur. In particular, the potential-energy increase that may accompany the rapid attenuation of the mantle lithosphere may be comparable to the magnitude of the driving forces of continental deformation so that the transition from convergent to extensional deformation may take place during the process. It is suggested that the Tertiary Tauern Metamorphic Event in the Eastern Alps may provide an important mid- to deep-crustal record of such a process. There, Tertiary heating was accompanied by the onset of both uplift and lateral extension of the range. Conductive heating models of the crust that have been previously applied to explain the temporal sequence of Eo-Alpine and Tauern metamorphic events imply thickness evolutions of the subcrustal lithosphere that are not easily reconciled with these observations and, in particular, with the wide preservation of Eo-Alpine high-pressure low-temperature assemblages.

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