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Melting of the continental crust: Some thermal and petrological constraints on anatexis in continental collision zones and other tectonic settings

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Abstract. Useful constraints on the depth-temperature range and degree of melting of lower crustal rocks have been deduced by combining the results from experimental petrology with plausible PTt paths calculated for continental collision zones. H_2O is required to generate melts at the temperatures of crustal orogenesis. Our modeling has shown that residual free water from subsolidus dehydration reactions produces less than 0.5% granitic melt at the H_2O-saturated solidus. Breakdown of hydrous minerals results in fluid-absent (dehydration)- melting. For average geotherms such melting of muscovite and biotite in metapelite will generate at most 25% granitic melt. Because of much higher solidus temperatures, fluid-absent melting of hornblende in metabasalt (amphibolite) will not occur in collision zones without augmented heat input from the mantle. Fluid-absent melting of epidote in metabasalt, and low Ca amphibole and biotite in metavolcanics, can produce up to 10% melt in continental collision zones. Minor melt production from dehydration-melting of chlorite, staurolite, chloritoid, or talc will occur at pressures greater than ~0.6 GPa. The calculated paths envelope small regions of PT space, thus identifying which mineral reactions are important for crustal anatexis and at which PT conditions future experiments should be made. In a dynamically evolving crust during uplift, thermal buffering by melting reactions is a minor component of the heat budget. Extension of thickened continental crust will not promote dehydration-melting without the incursion of mantle heat. Delamination of thickened eclogitic lower crust, or decompression of the asthenosphere in response to lithosphere thinning are the most promising mechanisms to induce extensive lower crustal melting including amphibolite.

Introduction

An ever important question is how much of continental evolution involves remelting of older crustal materials compared to fractionation of hydrous magmas direct from the mantle? While several proposals have been made on geochemical grounds [e.g., *Taylor and McLellan*, 1985], many of the necessary thermal constraints have not been adequately considered. Melting of continental crust can occur in three ways: (1) by supplying H_2O to rocks at temperatures above their H_2O-saturated solidi, (2) by decompression through dehydration-melting reactions, and (3) by increasing the heat supply through the Moho or base of thickened crust.

In recent years, several modeling studies have attempted to simulate crustal anatexis in different tectonic settings. These have ranged from simple continental collision belts [*England and Thompson*, 1984, 1986; *Zen*, 1988; *De Yoreo et al.*, 1989a,b, 1991; *Patiño-Douce et al.*, 1990; *Pinet and Jaupart*, 1987] to melting of oceanic crust in subduction zones [*Peacock et al.*, 1994] to continental extension [*Sonder et al.*, 1987; *Looseveld*, 1989] to massive injection of mantle-derived magma into the continental crust [*Huppert and Sparks*, 1988; *Barton and Hanson*, 1989]. Such studies have varied in their purpose from investigating the ranges of parameters appropriate to achieving high enough temperatures for melting for specific rock types at various aH_2O to attempting to understand the physical and chemical controls on particular intrusive rock settings and migmatites.

From experimental studies we have obtained a good understanding of melting reactions that probably occur in continental crust. We also have information on their temperature intervals, the evolution of melt chemistry at increasing melt fraction and the changes in the compositions of the mineral residues. There are enough data available at least to outline how a entire range of crustal components will melt at different depths (also in thickened orogens) and how this is affected by the absence of fluid or by fluids of different aH_2O. The important modeling study by *Zen* [1988] assumed that substantial melting begins at the wet solidus. This would require a large reservoir of free water and a much larger porosity

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than indicated here and in the models of *Connolly and Thompson* [1989a]. This is unlikely given the instability of fluid in the Earth's crust and, if it is indeed correct, leads to gross overestimation of the amount of melting at low temperatures.

Our purpose here is to use thermal modeling techniques and recent experimental data relevant to continental collision orogens to delimit the anatectic events expected. We will further explore more complex processes in collision orogens as part of a continuum of tectonic processes that effect the whole continental lithosphere. We draw attention here to the fact that we are dealing with thickened continental, rather than oceanic, lithosphere, but which may contain metamorphosed basalt (amphibolite). Partial melting of subducting oceanic crust has recently been considered by *Peacock et al.* [1994] and will not be treated further here.

Within the context of the relatively well constrained case of continental collision orogens, we attempt to explore the following more general problems:

1. To what extent can dehydration reactions deeper in a crustal orogen release fluids that through buoyant upward migration can cause H_2O-saturated melting at shallow crustal depths?
2. How efficient are melting reactions for temperature buffering in evolving orogens?
3. How geologically reasonable is it to maximize the contributory factors in heat transfer models so that increased melt fractions can be obtained?
4. How do tectonic situations that follow distinct paths in PTt space relate to experiments, say, of isobaric determination of the fertility of rocks to melt production?
5. How is the distribution of lithologies in a crustal column responsible for melting in different layers?
6. Which features of magmatism in conventional collision orogens require additional heat from the asthenosphere?
7. What can be deduced about the extraction mechanisms at the melting site from observations on exposed granitoids?

We attempt to show with examples from continental collision orogens that only a few features may be explained within the single-stage event of thickening followed by erosion. In many cases the collision orogens need to undergo later extension and/or delamination of the thickened crustal root to explain some of the magmatic features.

Crustal Melting in Orogenic Settings

A variety of lines of evidence from geochemistry and phase petrology [e.g., *Rudnick*, 1992; *Rushmer*, 1993] have indicated that melting of the lower continental crust results in the in situ generation of migmatites and sometimes in production of sufficient melt to induce granitic melt migration. Intrusive and extrusive granitoids are found in quite different orogenic settings and show variety of rock.type, abundance, source depth, intrusion depth, and content of magmatic water.

Peraluminous granites are thought to result from melting of metapelitic sediments [e.g., *White and Chappell*, 1988; *Burnham*, 1979; *Wyllie*, 1977; *Thompson*, 1982, 1988, 1990]. Their characteristics are residual garnet, or cordierite, and sillimanite, with crystallized muscovite and biotite.

Intrusive granitoids are not very abundant in eroded continental collision belts that typically show prograde medium-pressure (kyanite-sillimanite) metamorphic facies series (e.g., the central European Alps). Such granitoids are more abundant in terrains that have undergone decompression from kyanite to sillimanite stability fields (e.g. the Variscan belt of France). For crust thickened by thrusting and subsequently thinned solely by erosion, thermal modeling has shown that some crustal melting can occur without necessary augmentation of the heat supply from the mantle [*England and Thompson*, 1984, 1986].

In extensional tectonic environments intrusive granitoids are quite abundant when associated with low-pressure (andalusite-sillimanite) metamorphic facies series [e.g., *De Yoreo et al.*, 1991; *Wickham*, 1987a; *Wickham and Oxburgh* , 1987]. Here widespread crustal melting often appears to require massive heat input from cooling of mantle-derived magmas, or at least thinning of the lithosphere and uplift of the asthenosphere [*England and Thompson*, 1984, 1986; *Huppert and Sparks*, 1988; *Powers and Bohlen*, 1985; *Lachenbruch and Sass*, 1978; *Patiño-Douce et al.*, 1990; *Sandiford et al.*, 1987; *Thompson*, 1990; *Wickham and Oxburgh*, 1985].

In Cordilleran terrains, massive batholithic intrusions are characteristic of ocean to continent convergence. Mantle-derived magmas generated during subduction appear to have caused extensive crustal melting and assimilation [e.g., *Hildreth*, 1981; *Miller et al.*, 1988]. Some metaluminous tonalites and trondhjemites, and some low-K dacites, could have amphibolitized basalt as source rocks [*Peacock et al.*, 1994], but the heat source for their melting is not clear.

We will combine petrogenetic grids together with calculated PTt paths for various tectonic settings to understand the broad features of crustal melting, beginning with the case of metapelite melting following continental collision. Our study differs from previous attempts at modeling crustal anatexis in that the melt amount and distribution are regulated by the H_2O released by dehydration reactions. To determine the optimum PT conditions for anatexis, first we consider a metapelite lithology as this is most fertile to lower crustal melt generation. We furthermore constrain the melts to remain in situ.

A Numerical Model of Crustal Anatexis

To assess the possible thermal effects of anatectic melting in collision belts, we have adopted a simple model for crustal metamorphism consisting of two major components: a model petrogenetic grid that determines the conditions for devolatilization and melting; and a thermophysical model that determines the metamorphic thermal budget and the rates of fluid movement.

Simplified Petrogenetic Grid

Because of crustal heterogeneity there is little to be gained in examining large-scale features by employing a detailed model for crustal dehydration and melting. Instead,

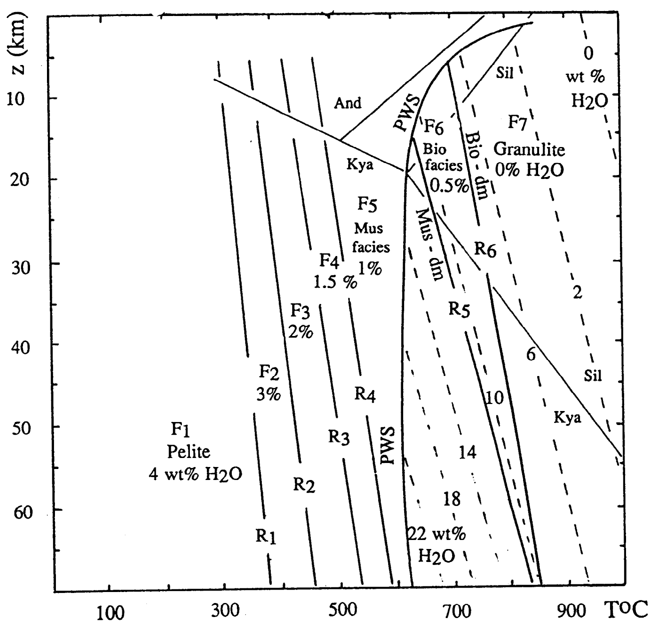

Figure 1a. The simplified petrogenetic grid for metapelite, extended from that used by *Connolly and Thompson* [1989a] to discuss subsolidus dehydration. The reactions R_1 to R_4 lie at temperatures below the H_2O-saturated solidus, whereas R_5 and R_6 occur above it. The water contents of the various facies assemblages (F_1 to F_7) are shown in weight percent, as are those of granitic melts above the wet (H_2O-saturated) solidus. The H_2O-saturated pelite solidus (PWS) is from *Thompson* [1982], dehydration-melting of a muscovite granite (Mus-dm) is from *Huang and Wyllie* [1981], and dehydration-melting of biotite in metapelite (Bio-dm) is extrapolated from *Le Breton and Thompson* [1988] and *Vielzeuf and Holloway* [1988]. The two fluid-absent mica melting curves most likely intersect near 1.5 GPa (~ 55 km [*Thompson*, 1990, Figure 6]). Two version of Al_2SiO_5 (And, Kya, Sil) phase relations are shown [*Richardson et al.*, 1969; *Holdaway*, 1971].

the schematic petrogenetic grid shown in Figure 1 has been designed to broadly reflect the behavior anticipated for pelitic lithologies. The grid consists of six discontinuous dehydration reactions, of which reactions R_1-R_4 represent subsolidus greenschist to amphibolite facies dehydrations and the two higher-temperature reactions, R_5 and R_6, represent muscovite and biotite dehydration-melting (fluid-absent) because they occur above the wet-pelite (granite) solidus. The "unmetamorphosed pelite" contains 4 wt % water. Reactions R_1 and R_2 release 1 wt % water by dehydration of the rock, and those at higher temperature release 0.5 wt % water. Melting behavior is approximated by assuming the rock is composed of a fertile "granite-minimum" component (40 wt %) and a refractory component. Thus the amount of melt generated at a given condition depends only upon the amount of water released by dehydration reactions, or present in porosity, and the solubility of water in the melt. Water solubility in granitic melt, indicated by the dashed curves in Figure 1a, is estimated from the solubility of water in albite melts [*Burnham*, 1979] with a small correction for the lower solidus temperature of granitic melts.

Within 200°C of the wet solidus the amount of continuous melting, due to changing water solubility, is minor in comparison to the amount of melting generated by dehydration-melting reactions. This can be seen in Figure 2 which illustrates two isobaric sections (at about 0.8 and 1.7 GPa) showing melt proportion as a function of temperature generated for the petrogenetic grid in Figure 1. The values of ϕ are zero and 1 vol % initial H_2O-filled porosities. This behavior mimics that observed experimentally [e.g., *Gardien et al.*, 1994] in that melting occurs in steps at dehydration -melting reactions rather than continuously between the wet and dry solidi [see also *Vielzeuf et al.*, 1990]. The energetics of dehydration and melting are calculated assuming that (1) the release of 1 g of water by dehydration requires 3.2 to 4.1 kJ, increasing with temperature [*Connolly and Thompson*, 1989a,b], (2) melting of 1 g of the granite minimum component requires 0.5 kJ [*Burnham*, 1979], and (3) that water and the anhydrous melt component have a negligible enthalpy of mixing [*Clemens and Navrotsky*, 1987].

An entire crustal section being composed of pelitic rocks is implausible, but the pelite-melt model employed here errs so as to maximize melt and fluid production. As such, it is useful for establishing upper limits on the importance of these processes.

Thermophysical Model

The thermal perturbation induced by continental collision (thickening) is modeled after the method of *England and Richardson* [1977] and *England and Thompson* [1984, 1986] by assuming an instantaneous doubling of crustal thickness by overthrusting. The initial model crust is 35 km thick with an average density of 2.8

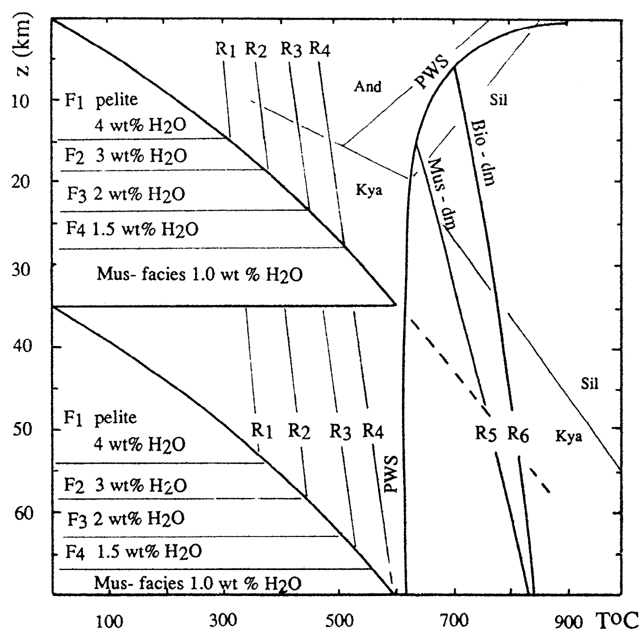

Figure 1b. The facies distribution immediately following 35 km thrusting, together with the perturbed geotherm, relative to the metapelite solidi.

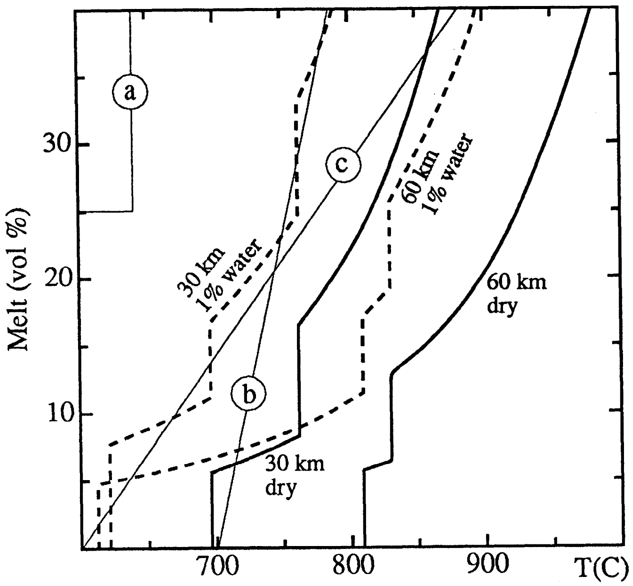

Figure 2. Melt fraction as a function of temperature as determined from the petrogenetic grid (Figure 1) for two isobars (at 30 and 60 km) for 1 wt % free water (thick dashed curves) and no free water (thick solid curves).The thin solid line labeled a is the nonlinear stepwise melting curve of *Zen* [1988] and the lines labeled b and c are the linear melting models of *De Yoreo et al.* [1989b].

x 10^3 kg/m^3 and radioactive heat production (2 x 10^{-6} W/m^3) confined to the upper 15 km The mantle is modeled as a 115-km -hick refractory layer with an average density of 3.5 x 10^3kg/m^3 and a constant heat flux of 3 x 10^{-2} W/m^2 at its lower boundary. Heat capacity of the lithospheric rocks was computed as a function of temperature from the relation given by *England* [1978]; and thermal conductivities were computed by the functions (in W/ m K).

$$K_M = K_{LM} + K_R \quad (1)$$

$$K_C = K_{LC} + K_R \quad (2)$$

where K_{LM} = 418.4 / (31.02 + 0.21 T) for T< 1232 K and K_{LM} = 1.443 for T> 1232 K with K_{LC} = 418.4 / [{(0.6259 + 0.02245w) / 0.005975 - 0.00035854w)} + 0.21 T] and K_R = (-1.151 + 0.002301T), where w is the weight percent water of the rock, the subcripts L and R indicate lattice and radiative components of the conductivity, and the subscripts C and M refer to crust and mantle. Equation (1) is from *Schatz and Simmons* [1972] for a dunite, and equation (2) has been modified from *Schatz and Simmons* to fit the conductivities of quartz-bearing lithologies as a function of metamorphic grade (water content) as summarized by *Clark* [1966]. In comparison to models in which crustal conductivity values are applied to the entire lithosphere [e.g., *Connolly and Thompson,* 1989a,b; *De Yoreo et al.*, 1989, 1991], higher mantle conductivity increases the magnitude, and rate of propagation, of thermal perturbations in the crust resulting from thickening.

The prethickening steady state geotherm for the model was computed by an iterative procedure by imposing the constraint that the initial "metamorphic facies," as defined by the petrogenetic grid (Figure 1), were required to be in equilibrium at the conditions of the geotherm. It is assumed that any water released by the dehydration reactions required to attain this equilibrium has left the crust prior to thickening.

For the specified thermal properties, this procedure leads to an initial configuration (Figure 1) in which the crust has an average water content of 3 wt %, an average geotherm of 18.8 K/km, and an average conductivity of 2.0 W/m K. The mantle geotherm and conductivity average 8.3 K/km, and 4.0 W/m K. After thickening, the crust remains in its thickened state for a period of 20 m.y. (incubation) and then erodes to its original thickness at a constant rate over a period of 100 m.y.

The thermal evolution of the crust is evaluated by solution of the one-dimensional governing equation:

$$C_r \frac{dT}{dt} = \nabla(K\nabla T) + C_f \nabla(q_f T) - T C_r + H + A + u C_r \nabla T. \quad (3)$$

where C_r is rock volumetric heat capacity, K is thermal conductivity, C_f is fluid volumetric heat capacity, q_f is fluid-volume flux, H is reaction enthalpy, A is the rate of radiogenic heat production, and u is the erosion rate. Equation (3) was solved by finite difference approximation using fixed point iteration, Crank-Nicolson weighting, a grid spacing of 462 m, and a variable time step (adjusted so that the maximum change in temperature at any node was less than 2 K per time step). The robustness of numerical solutions was tested by evaluating the energy balance constraint, which was found to be satisfied with errors of less than 1% of the total energy budget. The rate of enthalpy production from dehydration and melting reactions was assumed to be heat flow controlled. In a given time step this rate was iteratively refined so that the temperature at any node undergoing reaction was maintained at the equilibrium temperature for the reaction within a tolerance of 0.01 K.

The melts were considered to be immobile, and water flux rates were computed for a lithostatic pressure gradient and the porosity (ϕ)-permeability (k) relation [*Connolly and Thompson*, 1989a, Equation 2]:

$$k\ (m^2) = 2.3 \times 10^{-14}\ \phi^3 \quad (4)$$

where porosities (ϕ) are determined as the volume of free water present in the rock. The thermal and transport properties of water were treated as constants with the values of density (ρ) = 900 kg m^{-3}, volumetric heat capacity (C_f) = 2.5x 10^6 J/m^3 K, and viscosity (μ) = 1.5x 10^{-5} Pa s. The melt phase was not allowed to migrate, as our purpose was to determine the maximum melt fraction at any depth. Equation (4) is highly uncertain, but by any standard, it yields conservative estimates of the permeability of lower crustal rocks. Accordingly, its use here maximizes the time that free water is present in the lower crust and therefore the importance of wet-solidus melting.

The model parameters have been chosen to reproduce average subcontinental lithospheric thermal structure as described by *England and Thompson* [1984, central model, Figure 3e]. All other aspects of the models (i.e., instantaneous thickening by overthrusting and the assumption, implicit in a one-dimensional model, of no lateral heat flow) led to overestimates of the metamorphic temperatures achieved in collision belt systems. Thus, to the extent that the initial conditions are indeed typical of continental lithosphere, the models maximize the importance of melting.

Results of Modeling for Crustal Melting in Collision Belts

The results of our calculations in one dimension for crust thickened by instantaneous thrusting (configuration of Figure 1b) and following an incubation period of 20 m.y., linear uplift remove the 35 km upper plate in 100 m.y. In Figure 3 it is seen that the PTt paths followed by the crustal rocks intersect the muscovite and biotite dehydration-melting reactions at low angles. Therefore relatively small uncertainties in the model crustal temperatures (~ 80 K) can have major consequences on the predicted melting behavior. As noted earlier, the construction of the model is such that it is more likely to predict excessive temperatures; thus it appears unlikely that biotite melting will be as extensive as shown in Figures 4 and 5 and therefore equally unlikely that extensive plutonism will result from such reactions.

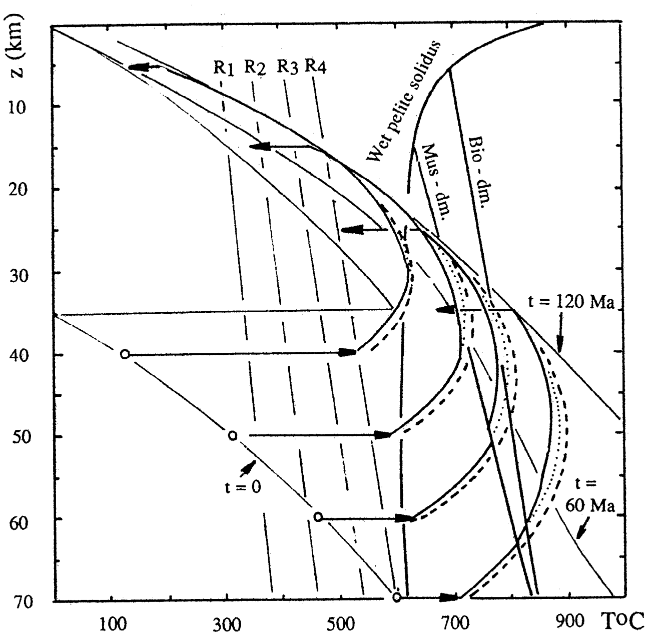

Figure 3. The petrogenetic grid and model PTt paths for the simplified metapelite system in a continental collision environment.The initial geotherm is the central model of *England and Thompson* [1984, Figure 3e].The postthrusting geotherm (t=0) and those at 60 and 120 Ma are shown. Time intervals are not illustrated but can easily be constructed as erosion is assumed to be linear with time. Incorporation of latent heat of melting results in only 20 to 40° C lowering of the PTt paths [see also *Zen*, 1988]. The present model includes the effect of deep dehydration reactions (50 to 70 km in the thickened column). The influence of migrating metamorphic fluid on metapelite melting can be seen in Figure 5.

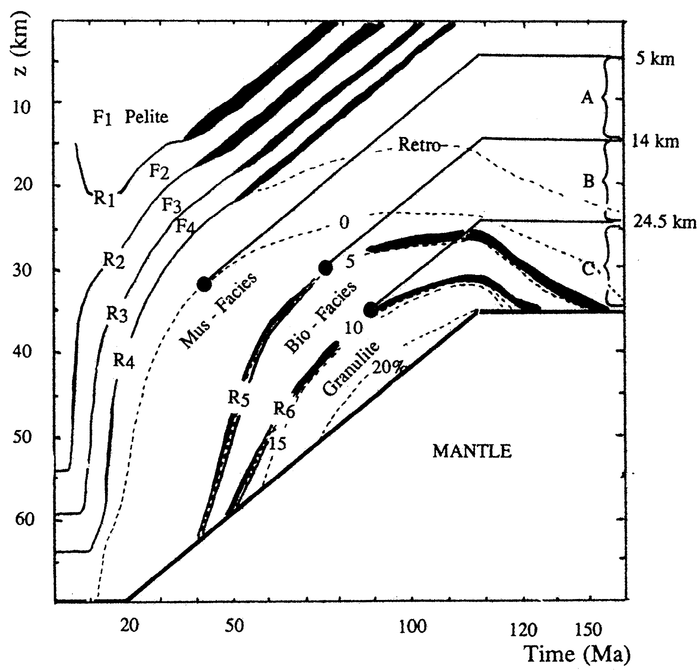

Figure 4. Evolution of the model metamorphic and migmatite facies. The heavy solid regions correspond to the equilibrium facies boundaries in Figure 1. They are smeared because both heat flow and fluid flow control the rate at which reactions take place. Retrograde fronts are shown only for F_4 (dashed curve labeled Retro), the movement downward is due to retrograde hydration effects. The weight percent of in situ melts is shown by the contoured dashed lines at 5 or 10% intervals. The zero contour shows the limit of the H_2O-saturated solidus. The 5% and 15% contours are essentially coincident with the muscovite and biotite dehydration-melting reactions, respectively. The solid circles indicate the maximum height of the melted column. These occur as tangents to the melt contours and thus before the highest active point in the models, because the crust is thinning. The black point at 0 wt % will end up 5 km below the surface, whereas those for muscovite and biotite dehydration-melting will end at 14 and 24.5 km below the surface, respectively. These will consist of the assemblages A, ~ 1 vol % leucogranite + muscovite facies; B, ~ 7 vol % muscovite granite + biotite facies; and C, ~ 18 vol % biotite granite + granulite.

Distribution of Migmatite Facies and Granulites

The distribution of the hypothetical facies and melt within the crust as a function of time are shown in Figure 4. In the initial stages of the model, isograds migrate upward relative to the base of the crust due to prograde metamorphism. With increasing time, the rate of advance slows and the isograds descend slightly as a result of retrograde effects. For the subsolidus isograds, this descent becomes inhibited as free water is consumed by progressively deeper retrograde reaction fronts (see also Figure 5). Once this occurs, the fossil isograds migrate upward solely in response to erosion.

Because there is always a finite amount of pore water present in the rocks during prograde subsolidus conditions,

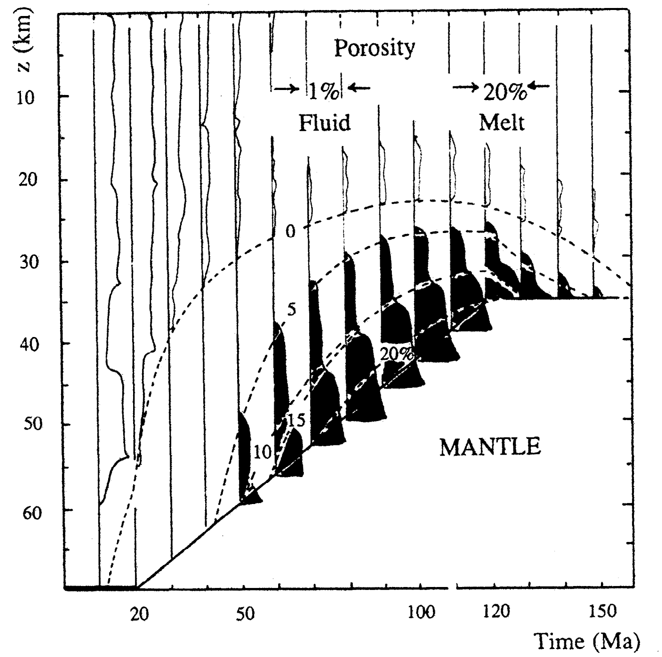

Figure 5. Melt production (open regions) and free H_2O (solid regions) migration in the continental collision orogen as a function of time. These are shown at 10 m.y. intervals, which correspond to 20 wt % melt and 1 wt % H_2O. Fluid saturation occurs at temperatures lower than the H_2O-saturated solidus (the dashed 0% line). The other dashed lines indicate weight percent melt at 5% intervals. The steps on the melt region correspond to the muscovite then biotite dehydration-melting reactions. The system contains melt from 50 m.y. into the history and up to 30 m.y. after erosional uplift has stopped. In these models the melt is not allowed to migrate and does not rise higher than 28 km in the column.

the 0 wt % melt contour of Figures 4 and 5 locates the position of the wet-solidus with time. In this respect, it is noteworthy that although almost the entire lower crustal slab achieves wet-solidus melting temperatures, the amount of wet solidus melt generated is less than 1 wt % (Figures 4 and 5). This reflects both the high solubility of water in the melt and the rapid migration of free water relative to the advance of isotherms through the crust. Consequently, significant anatexis only occurs in association with the muscovite and biotite dehydration-melting reactions, which generate maximum melt fractions around 25 wt %.

Quantities of Melt and Location of Anatectic Layers

The total volume of melt generated from the muscovite and biotite breakdown is about 0.9 and 2.2 km^3 melt/km^2 crust. If this melt were to remain in situ, migmatites would comprise roughly 15 vol % of the lower 20 km of the postmetamorphic crust (Figure 5) but would not be exposed at the surface solely through erosional exhumation.

Migration of Dehydration Fluids and Extent of H_2O-Saturated Melting

A possible scenario for generating leucogranitic melts [see *France-Lanord and LeFort*, 1988; *Scaillet et al.*, 1990] is that water released by dehydration processes, in otherwise refractory rocks, migrates upward into melt-fertile rocks that are above the wet-solidus temperature. Such a mechanism cannot increase the total volume of melt generated, because, in constrast to the melts produced by dehydration-melting reactions, it will result in water-saturated melts. However, it will generate the maximum degree of partial melting, and this may induce mechanical instabilities favoring segregation and plutonism. The efficacy of this process can be estimated by modifying the model so that melting cannot occur in the lower 20 km of the crust. In this case, water released by the muscovite and biotite subsolidus dehydration migrates upward until it reaches the overlying fusible pelitic rocks. The maximum extent of muscovite and biotite dehydration in the numerical model occurs at about 80 Ma (Figure 3), at which point 4.2 x 10^{11} kg H_2O/ km^2 crust have been released. The minimum water solubility of granitic melt at the depth (30 km, ~0.8 GPa) of the overlying pelite is 14 wt %, thus the maximum amount of melt that can be produced is of the order of 1.4 km^3 melt/ km^2 crust.

Heat Budget for Crustal Melting

The above discussion ignores the thermal effects of an change in the distribution of melt enthalpy. The importance of these effects is illustrated in Figure 3, in which computed PTt paths for the model (thick solid lines) are compared with those resulting when the melt enthalpy (dotted lines) and total enthalpy (dashed lines) effects are disregarded. This comparison reveals that the melting heat effect is capable of deflecting PTt paths away from melting reactions and that univariant melting reactions may buffer PTt paths but only on a local scale (1-2 km). However, the overall endothermic enthalpy effect reduces metamorphic temperatures by less than 40 K, and exothermic effects account for temperature increases of less than 10 K. The magnitudes of the heat effects due to dehydration and melting during the erosional history are shown in Figure 6.

It should be kept in mind that Figure 6 shows the total amount of energy involved per unit volume of the crustal column, while the total volume of the column is reduced by half from 20 to 120 m.y.. Thus the total latent heats of melting and crystallization are of similar absolute magnitude. The averaged radioactive heat production for the crustal model is about 0.9 x 10^{-6} W/m^3, so in comparison metamorphic effects are of relatively minor importance.

Thermal Buffering in a Thickened Crust Undergoing Melting

England and Thompson [1984, 1986] in their models for metamorphism and melting in continental collision belts did not explicitly include values for latent heat of melting. They argued that because reaction enthalpies are

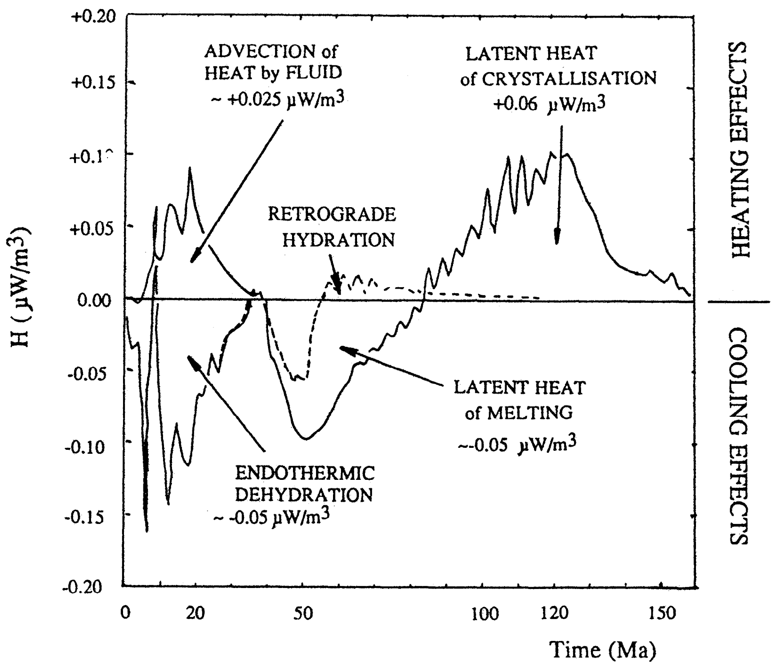

Figure 6. Instantaneous enthalpy effects along the PTt paths for a crustal column initially 70 km thick. All contributions are small compared to radiogenic heat production ($A^* = 0.86\ \mu W/m^{-3}$). The numbers by the arrows for the various endothermic and exothermic effects are averaged values over the time intervals where each effect dominates.

small compared to the overall heat budget for metamorphism, thermal buffering in the lower crust would not be significant during tectonic processes that change crustal thickness. This observation was supported by subsequent models that incorporated latent heat effects [*Zen*, 1988; *De Yoreo et al.*, 1989a,b; *Connolly and Thompson*, 1989a,b; *Thompson and Connolly*, 1990].

The limited amount of thermal buffering by melting reactions in these models is a reflection of the case where the crust is changing thickness (here by erosion). Thermal buffering by crustal melting reactions is apparently a more efficient process where magmatic underplating occurs [*Huppert and Sparks*, 1988; *Bergantz*, 1992; *Parsons et al.*, 1992].

The PTt paths for collision belt melting for the central model are shown in Figure 3 to emphasize the effects of the chosen values for latent heat on lowering the temperatures of the paths. Several effects contribute to the thermal budget along any PTt path (Figure 6). The endothermic effects of subsolidus dehydration and latent heat of melting are of similar magnitude to heat advection by fluid and latent heat of crystallization, although occurring at distinct time intervals along the PTt path. *Vielzeuf et al* . [1990, pp. 67-68] considered a zero-dimensional case where a static endothermic reaction with a latent heat of 2 kcal/100 g rock could buffer the rock temperature against an addition of heat which would otherwise be available to raise the temperature by 80°C. The calculations shown in Figure 3 emphasize that this effect is much less for the one-dimensional case, because the latent heat of melting is small compared to the amount of energy that flows through to maintain a steady state in response to thermal gradients.

Segregation of Melts by Residue Compaction and Conditions for Melt Escape

For "pseudogranitic" melts there is considerable range of opinion on the amount of melt production required to induce segregation and separation (see review by *Fountain et al.* [1989]). *Wickham* [1987b] using the experimental results of *van der Molen and Paterson* [1979] concluded that 25-30 vol % melting is required to promote granitic melt segregation, whereas *Arzi* [1978] and *Miller et al.* [1988] concluded that proportions closer to 50 vol % are required. While our calculations show that about 25 vol % melt is generated, it is important to consider how long this melt can remain in situ. The simplest escape mechanism to investigate with the present results is the compaction time (t_h) for collapse of residue at the melting site.

The maximum time required for the anatectic melts to segregate can be roughly estimated from the compaction model of *McKenzie* [1984, Appendix B; 1985], which gives the time to extract two-thirds of the melt from a partially melted layer of thickness h as

$$t_h \sim h\, m/(\varepsilon\, \phi^2). \qquad (5)$$

This relationship shows that the compaction time (t_h) is inversely proportional to the square of the initial melt fraction. Taking viscosity (μ) of wet granitic melts to be of the order of 10^3 to 10^4 Pa s [*Shaw*, 1972; *Wickham*, 1987b], $\varepsilon = 4 \times 10^{-6}$ kg/s^2 [*McKenzie*, 1984], and layer thicknesses of 10.5 km; the time scale necessary to extract the melt formed by muscovite dehydration-melting ($\phi = 8$ vol %) is 13 to 130 m.y., of the same order as the orogenic event. In contrast, the timescale for the melt formed by biotite, following muscovite-dehydration-melting (ϕ = 21 %), is considerably shorter (1.8 to 18 m.y.), so that segregation of these melts is more plausible. Thus only the lower-temperature anatectic melts (generated from muscovite fluid-absent melting, ϕ = 5 to 10%) could remain in situ as migmatite. The higher-temperature melts (resulting from biotite following muscovite fluid-absent melting, ϕ> 15%) could escape by compaction well within the model times. Deformation-enhanced melt extraction could remove melt much faster than the optimum cases illustrated here and diminish the likelihood of migmatites remaining in situ.

Variation of Thermophysical Model Parameters for Anatexis in Continental Collision Belts

For a given set of controlling thermophysical parameters, the resulting PTt paths for all samples define relatively narrow envelopes in PT space. It is useful to examine how systematic changes in the parameters that govern geotherm evolution will modify the PT envelopes in different collision environments.

Exhumation Timescale

With erosion rates of 0.35 km m.y.$^{-1}$ and the simple petrogenetic grid from Figure 1, lower crustal metapelites will undergo some melting, but amphibolites will not (see also discussion regarding Figures 7 to 9, below). Faster erosion rates (e.g., 35 km in 50 m.y., u= 0.70 km m.y.$^{-1}$, [*England and Thompson*, 1984]; and Figures 6, 7, and 8) will not achieve as high temperatures as shown here (Figures, 2 and 8). They can be used, however, to approximate the case for more rapid tectonic exhumation of partially melted lower crust [*England and Thompson*, 1986, Figures 3 and 4].

De Yoreo et al. [1989b, 1991] employed a thermophysical model that leads to considerably higher temperatures than in our models. Specifically, they use very low erosion rates (u= 0.10 km m.y.$^{-1}$), which after crustal thickening cause the geotherm to approach that of a crustal section with twice the normal radioactive heat production; and the thermal conductivity of the mantle is taken to be that of the crust, resulting in extremely high temperatures in the mantle and hence mantle anatexis.

Incubation Period (Elapse Time Before Uplift Starts)

An increase of the incubation period, where the doubled crust remains at the thickened depth before erosion commences (from 20 to 50 m.y. or more) produces much higher temperatures in the lower crust prior to uplift, as shown by the models of *Zen* [1988, pp. 224-226]. For incubation periods longer than about 30 m.y., the geotherm at the beginning of uplift is close to the infinite geotherm (without uplift, see the curves labeled V_∞ in Figure 3 of *England and Thompson* [1984]). England and Thompson argued, [1984 p. 901] following *Richardson and England* [1979] and *Houseman et al.* [1981] that after 20 m.y. the thickened lithosphere might decouple, then drop into the mantle to be replaced by hot asthenosphere. In which case, incubation periods longer than 20 m.y. may not be common.

Depth of Thrusting

The depth at which one-sheet thrusting occurs has relatively minor effects on the temperatures achieved in the lower crust. The effects of thrusting at 20 and 50 km rather than the 35 km used here can be seen by comparing the results obtained by *England and Thompson* [1984; Figures 6c and 8c with Figure 7c]. The location and radioactivity of the underthrust layer can certainly influence the degree of melting [e.g., *Patiño-Douce et al.*, 1990].

Mantle Heat Flux

Downward detachment of thickened lithosphere to be replaced by upwelling of hotter asthenosphere, or massive mantle-derived magma invasion beneath (underplating) or into continental crust (intrusion), can be simplistically modeled by increasing the mantle heat flux. This can be simply explored with a hotter computed geotherm, and can be seen in the results of [*De Yoreo et al.*, 1989b; *Patiño-Douce et al.*, 1990].

An often proposed additional heat source for crustal melting is intrusion of basaltic magma into the lower crust. The efficiency of such a process is thermally limited by conduction cooling time scales for plutons [e.g., *Hanson and Barton*, 1989]. A long-lived increase in mantle heat flux, local or regional scale, in response to global-scale plate motions may ultimately be responsible for events like the Great Basin Cretaceous magmatism and metamorphism [e.g., *Barton and Hanson*, 1989].

Latent Heat of Melting

Zen [1988] and [*De Yoreo et al.*, 1989b] included latent heat of melting in PTt path calculations similar to those of *England and Thompson* [1984, 1986]. They also concluded that endothermic thermal buffering would not lower calculated PTt paths by more than 100 K through endothermic reaction thermal buffering. Similarly, small shifts were found for subsolidus dehydration by *Connolly and Thompson* [1989a] and for dehydration-melting reactions [*Connolly and Thompson* , 1989b; *Thompson and Connolly*, 1990]. The main problem therefore is to determine the amount of melt that can be generated for common crustal rock types for given amounts of H_2O and for likely values of latent heat of melting.

Initial Geotherm

The central geotherm shown in Figures 3 and 7 lies in the amphibolite facies at 35 km depth, whereas hotter initial geotherms (Figure 8) lie at temperatures even higher than mica dehydration-melting reactions in pelites. This would require a metapelite lowermost crust to be permanently partially molten. The initial geotherm also determines the fluid-fertility of the lower crust for subsequent dehydration-melting; in the sense that more dehydrated mineral facies would lie at shallower depths for a hotter geotherm [*Connolly and Thompson* , 1989a, p. 352].

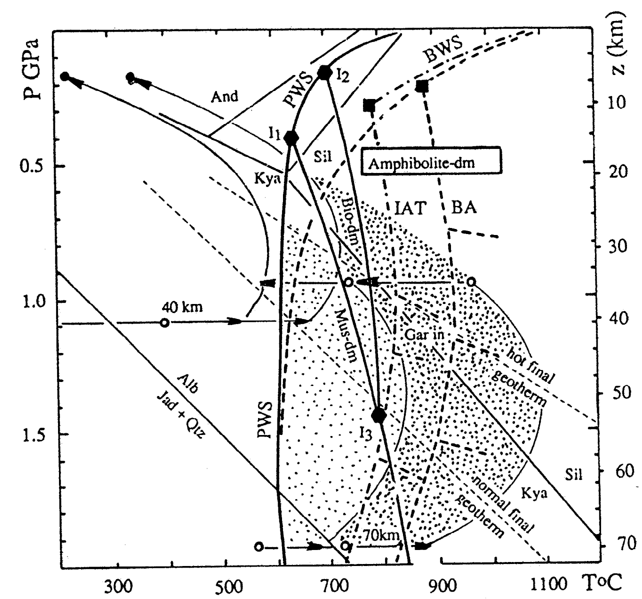

Figure 7. This petrogenetic grid includes reactions observed experimentally in metapelite, where the muscovite- and biotite-dehydration-melting reactions intersect at about 15 kbar [e.g., *Le Breton and Thompson*, 1988]. The PTt envelopes for continental collision resulting from the central geotherm lie entirely in the kyanite field. However, geothermometric-barometric estimates for migmatites mostly lie in sillimanite field. The data of *Rushmer* [1991] for hornblende dehydration-melting are taken to be typical of amphibolites (BA; see also *Wyllie and Wolf* , 1993) and for cummingtonite-biotite (IAT; see also *Conrad et al.* [1988]) is taken to be typical of metavolcanics (the interval labeled Amphibolite-dm). These have been extrapolated to higher pressures following the procedures of *Thompson* [1990] and appear to cross the metapelite dehydration-melting reaction deeper than 55 km. The Al_2SiO_5 (And, Kya, Sil) phase relations are from Figure 1, and albite to jadeite + quartz is from *Huang and Wyllie* [1981]. Other abbreviations:- BWS -basalt wet solidus; Chl-chlorite: Gar- garnet: Alb -albite, Jad- jadeite, Qtz -quartz.

PTt Paths for Continental Collision and Different Initial Geotherms

The quantities of melt produced for a particular rock type at a given depth, and the nature of the residue, are mostly a function of the initial geotherm. This in itself is an indication of the tectonic setting immediately prior to anatexis. These effects can be seen by comparing our results with those using hotter initial geotherms [e.g., *De Yoreo et al.*, 1989b; *Patiño-Douce et al.*, 1990] despite the other differences in the modeling.

Central (Average) Initial Geotherms

We will first consider crustal anatexis in terms of the PTt paths for the central model of *England and Thompson* [1984, Figure 3e; 1986, Figures 2 and 3]. These are shown Figure 7, together with several pelite and amphibole dehydration-melting reaction curves (the data sources are given in the caption). For continental crust of standard thickness (Moho at 30 to 35 km), the central initial geotherm lies at temperatures lower than even the H_2O-saturated solidus for granite melting. Thus melting of non thickened crust requires tectonic perturbation of the geotherm within the crust, or an increase in mantle heat flux. The PTt paths in Figure 7 for samples in the lower slab, illustrated for 40 and 70 km, define an envelope for the occurrence of in situ migmatites. The envelope of PTt paths for the "central" initial geotherm lies entirely in the kyanite stability field. Kyanite migmatites are described [e.g., *Barbey et al.*, 1990] but sillimanite is widespread in pelitic migmatites

Crustal Anatexis in Continental Collision Belts With Hot Initial Geotherms

Higher intitial geotherms result mainly from higher mantle heat flux, supplemented by lower thermal conductivity, higher radiogenic heat production, and longer incubation times (between time of burial and beginning of uplift). Variations in the latter three parameters would be local rather than general effects.

The PTt paths resulting from 35km thrust thickening from an initially hotter geotherm, use the calculations from *England and Thompson* [1984, Figure 3f; 1986, Figure 4]. They are illustrated in Figure 8 by the heavily stippled envelope at higher temperatures, compared to those for the central initial geotherm (Figures 3 and 7). All PTt paths in the lower slab would reflect both the kyanite-

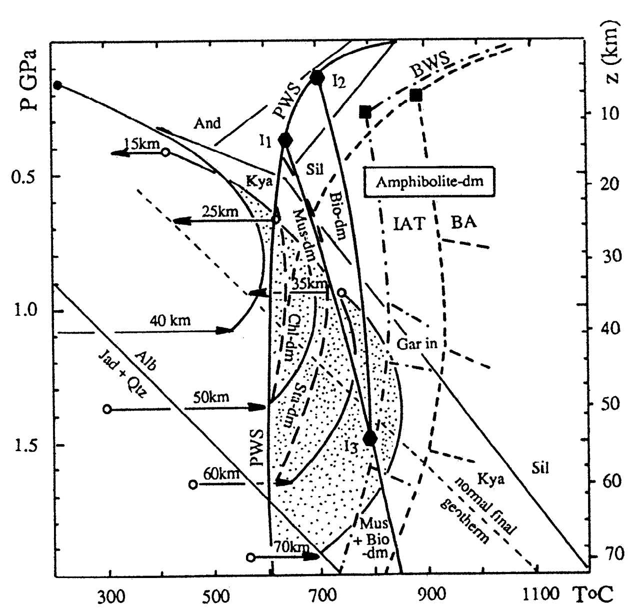

Figure 8. The central and hot PTt envelopes for continental collision, shown by heavy stipple at higher temperatures and light stipple at lower temperatures, respectively. The latter lie well into the sillimanite field, agreeing with many geothermometric-barometric estimates for migmatites (see Figure 9). The principal differences in the PTt envelopes lies deeper than 20 km. Special cases such as decreased thermal conductivity (blanket effect) or increased radiogenic heat production certainly can result in hotter PTt envelopes. In the general case, higher heat from the mantle is the biggest contributor.

sillimanite transition and extensive "decompression dehydration-melting" of pelitic and amphibolitic lithologies.

De Yoreo et al. [1989b] have also modeled crustal anatexis with a hotter initial geotherm and find extensive anatexis. Their results [*De Yoreo et al.*, 1989b, p. 190] do not easily separate the effects of the initial geotherm from the very slow erosion rates in their models (35 km in 350 m.y. compared to 35 km in 100 m.y. as used here). The "hotter" initial geotherm, shown in Figure 8, would require the lower few kilometers of metapelitic continental crust of "normal" thickness (Moho at 30 to 35 km) to be permanently partially molten, even by dehydration-melting. Such an observation is not consistent with all seismic data [e.g., *Mooney and Meissner*, 1992]. This in itself indicates that metapelite lithologies are not common at Moho depth, or the steady state geotherm for continental crust is generally not as high as shown in Figure 8.

The effect of higher initial geotherms on the present results are most easily seen by comparing Figures 3 and 4 of *England and Thompson* [1986] with our calculations that include the various heat effects (Figures 3 and 8, here). Such thermal conditions are not normally achieved in continental collision environments and require an augmented heat supply, such as upwelled asthenosphere or injection of massive amounts of mantle magma into the lower continental crust.

Migmatite Terrains and Decompression-Melting in Continental Collision Belts

For the PTt envelope shown in Figure 7 (central initial geotherm) samples beginning uplift from shallower than about 42 km will not experience temperatures high enough to even cross the H_2O-saturated solidus for granite melting. Samples from between 42 to about 58 km will only undergo melting if free H_2O is available at the melting site. Only samples from deeper than about 58 km will experience mica dehydration (fluid-absent) - melting in metapelitic compositions. All migmatites will be muscovite or kyanite (rather than andalusite- or sillimanite-) bearing, as the PTt paths lie deeper than the invariant point $\{I_1\}$ for muscovite melting. If melts remain in situ as migmatites they will never be found shallower than about 20 km, unless subsequently tectonically exhumed and emplaced at shallower depths [e.g., *Thompson*, 1990].

The common occurrence of kyanite relicts in sillimanite migmatite terranes indicates that such migmatization is related to decompression with maximum depth of burial occurring before T_{max} [e.g., *Jones and Brown*, 1989]. That such behavior is typical for migmatite terrains can be seen from the geothermobarometric(T-P) estimates in Figure 9.

Samples that have undergone "decompression-melting" will recross their appropriate solidi and crystallize long before erosional uplift is complete. The chemography of the melting reactions (fluid-absent and fluid-present) requires that water be released on solidification except in the cases where the paths cross the reactions at identical reaction stoichiometries. This migmatitic water can hydrate adjacent restite, or migrate upward to cause

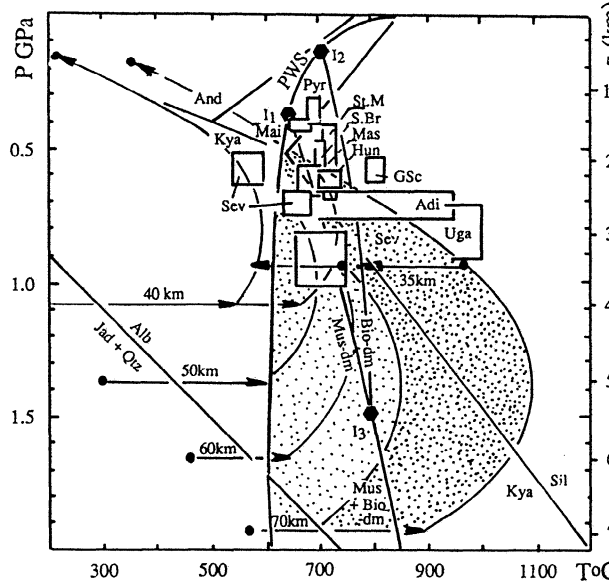

Figure 9. Geothermo-barometric estimates of migmatite zones compared to some contact aureoles and granulites. Most lie in the sillimanite stability field. Data sources are for migmatites: Mai, Maine [*De Yoreo et al.* [1989a,b]; Pyr, Pyrenees [*Wickham* , 1987]; St.M, St.Malo [*Brown*, 1979]; S.Br., S.Brittany [*Jones and Brown*, 1989]; Mas, central Massachussetts [*Tracy*, 1978]; Sev, Sevier [*Patiño-Douce et al.*, 1990]; for some contact aureoles: Hun, Huntly; GSc, Glen Scaddle [*Ashworth and Chinner*, 1978]; Adi, Adirondacks [*Powers and Bohlen*, 1985]; granulites: Uga, Uganda [*Sandiford et al.*, 1987]. Apart from Sevier Hinterland all P-T estimates lie in the sillimanite field; Maine and Pyrenees lie in andalusite field for the *Richardson et al.* [1969] Al_2SiO_5 triple point. Only Sevier Hinterland samples could have undergone anatexis within the PT envelope for average geotherms. Many of the migmatite terains show final equilibration in the depth range 10 to 20 km at higher temperatures than the PT envelope for hot geotherms governed by erosional uplift. Tectonic exhumation (by extensional uplift?) seems to be required to explain these low-P high-T conditions [see *Loosveld*, 1989; *De Yoreo et al.*, 1991].

retrogression, or in some cases melting, elsewhere in the metamorphic pile. If the water-undersaturated melt behaves as a closed system then the migmatites will contain muscovite stoichiometrically replacing Al_2SiO_5. If the water-undersaturated melt behaves as a closed system where less H_2O is consumed by remaking mica than is actually available in the melt, then this migmatitic water can induce partial retrogression in nearby restite.

The large PT region of the low-pressure (andalusite-sillimanite) metamorphic facies series will not be characterized by migmatites without the effect of an additional heat source [*De Yoreo et al.*, 1991]. Many migmatite terrains show final equilibration in the depth range 10 to 20 km at temperatures higher than the PT envelope for hot geotherms governed only by erosional uplift (Figure 9). Tectonic exhumation (extensional uplift) seems to be be required to explain these PT conditions.

Crustal Anatexis in Heterogeneous Crust

At the beginning, we stated our objective was to determine the optimum conditions for melt production in collision orogens. We now consider which aspects of real crust will change the amount of melt expected.

Additional Hydrous Minerals in Metapelites (e.g., Chlorite+ Staurolite in Kyanite Migmatites)

Ashworth [1975] has reported staurolite with sillimanite+K feldspar in muscovite-free migmatites, and *Thompson et al.* [1985, 1986] report staurolite+chlorite in kyanite migmatites. Considerations of subsolidus petrogenetic grids for metapelites led *Koons and Thompson* [1985, Figure 6] to suggest that dehydration-melting of staurolite and even chlorite could be locally important. Examples of dehydration-melting reactions involving Sta+Chl are included in Figure 7 extrapolated from the recent subsolidus petrogenetic grid by *Spear and Cheney* [1989].

Because of chlorite's high water content, its dehydration-melting could produce noticeable amounts of melt, over a moderate pressure range (~ 0.6 to 1.4 GPa) at temperatures just above the H_2O-saturated solidus (Figure 7). Likewise, staurolite dehydration-melting contributes to melt production before the micas react, over quite a large pressure range (~ 0.4 to 1.7 GPa). Thus prominent facies regions for chlorite migmatites would be quite restricted in their depth-time distribution compared to staurolite and mica facies The details can be deduced by constructing a diagram similar to Figure 4 but based on Figure 7 rather than Figure 3. For typical modes of chlorite and staurolite in average metapelites, up to 10 vol % melts could be generated for the indicated depth ranges.

Other Petrogenetic Grids

Real petrogenetic grids are obviously much more complex than shown here in our simplified case (Figure 1). However, reactions with slopes with different dP/dT than illustrated here spend less time coincident with the evolving geotherms. In any case, only those parts of petrogenetic grids that lie within P-T envelopes are of interest. More complex grids can be investigated in the field through studies of dehydration-melting reactions [e.g., *Waters*, 1990; *Tait and Harley*, 1988], but it is necessary to separate out effects of natural varitions in aH_2O from those of temperature, in trying to assess regional metamorphic heat supply.

Dehydration Melting of Amphibolites and Pyroclastics

Dehydration-melting of amphibole in basaltic amphibolites [*Rushmer*, 1991; *Beard and Lofgren*, 1991; *Rapp et al.*, 1991; *Wyllie and Wolf*, 1993] requires higher temperatures than the dehydration melting of biotite in metasediments [*Le Breton and Thompson*, 1988]. More magmatically evolved basalts metamorphosed in the amphibolite facies, or metavolcanics containing low Ca-amphibole and biotite, could undergo dehydration melting at similar temperatures to biotite in metasediments [*Conrad et al.*, 1988]. Field observations in central Massachussetts by *Hollocher* [1991] clearly suggest that biotite and cummingtonite-bearing metavolcanics undergo dehydration-melting close to that of muscovite in metapelites and long before hornblende in amphibolites.

In the petrogenetic grid of Figure 7 the amphibole dehydration-melting curves (IAT, BA) are shown crossing the dehydration-melting of muscovite + biotite deeper than 60 km. If this is true, then it is conceivable that at the lowermost part of thickened continental crust (60 to 70 km), anatexis could begin in metavolcanic, and perhaps even amphibolite, lithologies before in metasedimentary compositions. New experimental investigations in this PT range on metavolcanics are clearly needed.

Melting of a Layered Crust

A layered crust (e.g., Figure 10) will exhibit quite different general dehydration and melting behaviour to that discussed so far. While amphibolites and gneisses may not melt along the PTt paths for the central model, it is conceivable that their dehydration might induce melting in pelites. Such reactions would produce different mineralogies that should be observable in the field.

It will be noted in Figures 7, 8, and 10 that at less than 1.0 GPa pressure, the H_2O-saturated solidus for pelitic metasediments (PWS) occurs at temperatures lower than

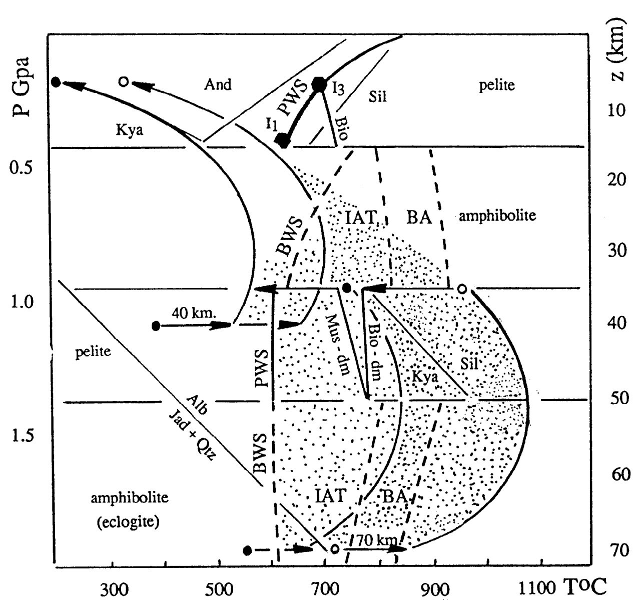

Figure 10. Melting reactions and PTt paths for a layered continental crust, arbitrarily consisting of 15 km of pelite above 20 km of amphibolite. No melting occurs in the upper 15 km of pelite. Upper slab amphibolites can melt if H_2O is supplied from beneath or for unusually hot geotherms. The buried pelite layer (between 35 and 50 km after thrusting) is very fertile to melting. The buried amphibolite layer (50 to 70 km after thrusting) is fertile to melting of metavolcanics for central geotherms and amphibolite melting for hot geotherms. Experimental studies are required of the stability of other hydrous minerals (chlorite, staurolite, epidote) near to the H_2O-saturated solidi of both rock types at lower slab pressures.

the H_2O-saturated solidus for basaltic-amphibolite (BA). This means that "boiling" of pelitic migmatites at their H_2O-saturated solidus could not induce H_2O-saturated melting in nearby amphibolites at normal crustal depths.

The layered model in Figure 10 (doubling of 15 km pelite underlain by 20 km amphibolite) shows that the buried pelite layer could melt and release hydrous melt into the overlying amphibolite of the upper slab. Subsolidus pelite dehydration might also induce H_2O-saturated amphibolite (basalt) melting. New experimental results are needed on the stability of hydrous minerals (such as epidote) in basaltic compositions, before the melting processes in amphibolitic crust can be quantified.

Crustal Anatexis in Other Tectonic Settings

We have concluded that for the central PTt envelopes following continental collision, melting of metapelites is to be expected but not extensive and for amphibolites is insignificant. Does it automatically follow that if amphibolites melt, then the orogenic processes involved cannot simply be continental collision? To investigate this question we will examine what can be deduced for more complex collision orogens and how important tectonic histories are that immediately precede or follow the continental collision.

Other Modes of Crustal Thickening and Thinning

It was shown by *England and Thompson* [1984, 1986], *Zen* [1988], *De Yoreo et al.*, [1989b, 1991] and *Patiño-Douce et al.* [1990] that homogeneous crustal thickening does not attain as high temperatures as thrusts. The gradual build up of thrust layers [*Gillet et al.*, 1985], or homogeneous thickening over long time periods does not change thermal effects much.

Extensional Environments and Andalusite/Sillimanite Facies Series

According to *Gaudemer et al* . [1988, p. 50], "orogenic belts of the world seem to belong to one of two classes; narrow ones with little (syn to) post-orogenic extension and plutonism and broad ones with widespread extension and plutonic activity". *Gaudemer et al.* [1988, p. 50] append "where there is no extension there are few or no granites." They consider that the differences are partly due to lateral heat-transfer which efficiently cools narrow orogenic belts. In addition, erosion attacks mostly the steep outer parts of mountain belts, thus removing the excess crust faster for narrow than for broad belts.

Extensional thinning following thickening by continental collision can cause rocks in the lower crust to achieve high temperatures during uplift only if a hot Moho temperature is rapidly uplifted [*England*, 1987; *Sandiford and Powell*, 1986; *Sonder et al.*, 1987]. The temperatures achieved during extension are not as high as those attained during the much slower process of erosional thinning. This results because tectonic exhumation transports rocks faster to the zero-temperature condition of the surface than the slower process of erosion [*England*, 1987; *Thompson and Ridley*, 1987].

It is not known to what extent continental crust can be thinned by plastic extension. If the maximum amount of thinning is to half-previous thickness [*McKenzie*, 1978], then thinning of a 35 km crust to 17.5 km could induce fluid-absent melting of muscovite by decompression and a transition from sillimanite to andalusite. Prograde andalusite to sillimanite facies metamorphism apparently requires augmented heat supply to the lower crust [e.g., *De Yoreo et al.*, 1991]. This can be achieved by massive basalt intrusion into an extending lower crust [e.g., *Huppert and Sparks*, 1988] or a long-lived increase in mantle heat flux [e.g., *Sandiford and Powell*, 1986].

We will shortly examine two sequential tectonic histories: (1) extensional thinning following thrust thickening [*Keen*, 1987; *England*, 1987; *Thompson and Ridley*, 1987] and (2) collision of continental crust thinned (by extension) immediately proir to collision [*Thompson*, 1989].

Sequential Tectonic Histories

PTt paths for continental thickening followed by erosional or extensional thinning will cross the solidi for fluid-saturated melting of metapelites, and perhaps of amphibolites, early in their histories at temperatures determined by the local fluid composition. *Thompson* [1989] has suggested that crust thinned by extension then involved in collision within 20 m.y. afterward will achieve temperatures high enough for tonalite melting (about $850^{o}C$ at 1 GPa). This can easily be modeled by juxstaposing two blocks of different thickness, the thinner block having greater basal (mantle) heat flux. These aspects must be investigated further through two-dimensional modeling. Thermomechanical modeling of later extension, during or after convergence, is important to understanding the emplacement of granitoid intrusions in belts that have undergone collision but where the magma emplacement occurs late in the history.

Concluding Remarks

The lower continental crust is not normally partially molten. Only tectonic perturbation of average geotherms will lead to anatexis. Lower continental crust in collision-thickened orogens can undergo some partial melting in fertile rocks of the underthrust slab during uplift, without an additional mantle heat source. These concluding remarks refer to the questions raised at the beginning of this paper;

1. The amount of free H_2O likely in the lower crust will produce negligible amounts of melt at the H_2O-saturated solidus. Metapelites will undergo dehydration-melting of muscovite and in deeper rocks of biotite. These lithologies are the most fertile to melt production, leading to up to 25 vol % in our models. In thermally perturbed lower crust some melting (<10 %) of biotite gneisses and metavolcanics is to be expected, but melting is not likely of hornblende in amphibolites for the central envelope of PTt paths.

Melt fractions of up to 25% are produced in our models using average thermophysical models for fertile continental crust. These melts were not allowed to segregate in the models and thus remained in situ as migmatite. Such melts always crystallize deeper than 20 km and thus will never be exposed by erosional uplift alone. The migmatitic H_2O becomes bound in new hydrous minerals on recrossing the solidi, so that boiling and retrogression of adjacent rocks is negligible.

2. Consideration of latent heat of melting lowers the temperature of crustal PTt paths, but only by 30 to 40 K, and then for periods less than 5 m.y. Thus thermal buffering by melting reactions is not a significant process in collision orogens [see also *Zen*, 1988]. Only a few percent less melt is produced compared to the simple case without including fusion enthalpy.

3. Hotter initial geotherms, long incubation periods, slow erosion rates, low crustal thermal conductivity, and high crustal radiogenic heat production in collision belts (as illustrated by *Zen* [1988], and *De Yoreo et al.* [1989b, 1991]) can induce more widespread lower crustal melting in metapelites, in gneisses, and even in amphibolites.

4. Despite the recent experimental studies that have determined melt fraction with increasing temperature along an isobar, new studies need to be conducted, beginning at the lowest temperature dehydration-melting reactions and within the PT envelopes revealed by thermal modeling studies.

5. By considering the most fertile lithology to crustal melting (metapelite), we have optimised the amount of in situ melt for the case of dehydration-melting. Our first attempt to model anatexis of lithologically layered crust shows greatly diminised melt fertility because of the refractory nature of hornblende amphibolites in the lower crust for central geotherms.

6. While hot geotherms can lead to melting of even amphibolitic lower crustal lithologies, the actual mechanisms by which augmented heat supply from the mantle can be achieved need specific geological demonstration from convergent plate boundaries. Single-event models (such as collision followed by uplift as discussed here, or thinning of asthenosphere by augmented mantle heat supply) appear to be oversimplified explanations for crustal anatexis. Sequential tectonic histories (e.g., collision following continental extension) and careful consideration of thermomechanical interactions (such as during oblique convergence) will certainly be the focus of many future studies.

Finally, the calculated compaction times for the simplest mode of melt extraction are of the order of 13 to 130 m.y. for melt fractions of about 8% (muscovite melting) according to our modeling. These compaction times are significantly longer than for melt fractions of about 21% (biotite melting), which are of the order of 1.8 to 18 m.y. *Harris and Inger* [1992] and *Inger and Harris* [1993] have deduced that melt fractions of about 12 vol % from a muscovite-bearing source were responsible for the Himalayan leucogranites. Their observations clearly indicate how future modeling can investigate the time-volume details of melt extraction and perhaps help to determine the efficiency of deformation in melt segregation.

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