

Continental geotherm and the evolution of rifted margins

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ABSTRACT

Models of melting accompanying mantle upwelling predict far more melt than is observed at nonvolcanic margins. The discrepancy may be explained if the paradigm of a uniform asthenosphere is incorrect. Work on the velocity structure of the continents has shown that the convecting sublithospheric mantle may have a potential temperature as low as 1200 °C, ~100 °C cooler than that beneath the oceans. The continental geotherm derived from studies on xenoliths brought up by kimberlites is also compatible with a cool sublithospheric mantle except where perturbed by mantle plumes. Upwelling of such cool mantle during rifting leads to little melt production, even for rapid extension rates, explaining the formation of amagmatic margins away from mantle plumes. However, the transition to seafloor spreading and the development of normal-thickness oceanic crust requires the invasion of the amagmatic rift by hotter oceanic asthenosphere and/or plume material. This influx may cause a transient thermal uplift, recorded as a breakup unconformity. Conversely, at volcanic margins, the onset of seafloor spreading is accompanied by the escape of the hot plume puddle along the mid-ocean ridge system away from the volcanic margin, leading to a pulse of rapid subsidence.

Keywords: rifting, breakup, geotherm, melting, mantle, unconformity.

INTRODUCTION

The 6–7 km of basalt (Bown and White, 1994) produced by decompression melting of upwelling mantle with a potential temperature (T_p —the temperature the mantle would have if it rose adiabatically to the surface) of 1300 °C matches the relatively uniform thickness of the oceanic crust (Spudich and Orcutt, 1980; White et al., 1992). Consequently, the T_p of the convecting upper mantle has been assumed to be ~1300 °C (White et al., 1992; Bown and White, 1994), except near mantle plumes (White and McKenzie, 1989). This paradigm appears incompatible with the structure of nonvolcanic rifted margins such as the Iberia Abyssal Plain (IAP in Fig. 1) west of Portugal. During Early Cretaceous rifting over a period estimated as <5 m.y. (Wilson et al., 2001) or 15 ± 5 m.y. (Minshull et al., 2001), mantle rocks were exhumed in a 150-km-wide continent-ocean transition zone (Whitmarsh et al., 2001), implying infinite crustal-stretching factors. If such stretching factors are extrapolated to the entire lithosphere, 4–5 km of melt (Minshull et al., 2001) should have been produced during the upwelling of normal 1300 °C asthenosphere (Fig. 2). However, estimates of the actual igneous thicknesses are in the low end of the range 0–2 km (Minshull et al., 2001).

The Iberia Abyssal Plain is only one, if the best studied (Perez-Gussinye et al., 2001; Whitmarsh et al., 2001), of at least eight nonvolcanic rifted margins that exhibit similar geophysical characteristics (Minshull et al., 2001). At a Jurassic rifted margin (Fig. 1) now exposed in the Alps (Desmurs et al., 2001), rifting leading to mantle exhumation lasted much less than 25 m.y. (Wilson et al., 2001). Models would predict >3 km of magma due to decompression melting of 1300 °C asthenosphere (Fig. 2), but little melting accompanied the extension and mantle exhumation (Desmurs et al., 2001).

As the relevant geologic and geophysical data sets have improved (Whitmarsh et al., 2001), it has become increasingly unlikely that the missing kilometers of igneous rock are present but not observed. The magma is unlikely to have been trapped beneath the less dense conti-

ental crust (Herzberg et al., 1983): where most melt production is expected (Buck et al., 1988), either the crust is thin and brittle (Perez-Gussinye et al., 2001; Reston et al., 1996), facilitating magma transport along fractures, or totally absent (Whitmarsh et al., 2001; Wilson et al., 2001; Desmurs et al., 2001) (Fig. 2). It is unlikely that much magma froze within the highly extended subcontinental mantle, as melt migrates quickly through the lithosphere (McKenzie, 1989); only minor gabbros are observed in the Alpine mantle sections (Desmurs et al., 2001).

Most ways to reduce melt production are unsatisfactory. Distributing mantle extension over a wider area than crustal extension can reduce the amount of mantle upwelling (and hence melt production) beneath the rift axis (Keen and Dehler, 1993), but such mechanisms cannot easily explain the paucity of magmatism at the effectively infinite stretching factors implied by the unroofing of 150-km-wide belts of mantle (Minshull et al., 2001). Lithospheric-scale low-angle detachments have not been observed (Reston et al., 1996; Perez-Gussinye et al., 2001; Minshull et al., 2001) and are in conflict with available subsidence and (for active rifts) heat-flow data (Buck et al., 1988). Time-dependent extension rates (accelerating rifting) can decrease melt thickness by as much as 1 km, but become less effective at short rift durations (Fig. 2). Invoking a constant-viscosity mantle (Shen and Forsyth, 1992) may broaden the region of mantle thinning (Minshull et al., 2001), but is in conflict with known flow laws of mantle minerals (Shen and Forsyth, 1992; Karato and Wu, 1993), and the results of the low-resolution numerical experiment of Shen and Forsyth (1992) have not been reproduced by other higher-resolution studies using similar parameters (Sotin and Parmentier, 1989; Cordery and Phipps Morgan, 1993). Finite-element modeling (Bowling and Harry, 2001) of west Iberia extension predicts that mantle extension should focus before crustal separation, leading to the production of 5 km of melt below the rift axis before the crust has thinned to ~4 km.

A remaining possibility is that the paradigm that the convecting asthenosphere has a uniform T_p of 1300 °C is incorrect.

SUBCONTINENTAL GEOTHERM

Estimates of the temperature of the upper mantle come from four sources: heat flow, xenoliths—especially those brought up by kimberlites, the thickness of the mantle transition zone, and seismic velocities. The Kaapvaal cratonic geotherm deduced from mantle xenoliths from kimberlite pipes (Fig. 3) shows a pronounced increase in temperature at ~190 km (Kohler and Brey, 1990; Lenardic and Moresi, 2000), from a T_p of ~1200 °C to one close to 1400 °C. This increase can be explained by the heating of continental lithosphere previously subject to a cooler geotherm (Parmentier and Turcotte, 1974; Lenardic and Moresi, 2000). Other thermobarometers, including those that most faithfully predict the presence or absence of diamond (Pearson et al., 1994), yield similar, if slightly cooler, geotherms. As the very kimberlites that transport these xenoliths to the surface are thought to be generated by the impact of a mantle plume on the base of the lithosphere (Crough et al., 1980), applying Occam's razor would imply that the heating is caused by the impact of the mantle plume and occurs at the very base of the lithosphere (200 km). This scenario implies that the T_p at the base of the lithosphere prior to plume impact was ~1200 °C. Ignoring thermally perturbed xenoliths yields similar cool steady-state geotherms for other kimberlite provinces (Russell and Kopylova, 1999; Fig. 3). These results are consistent with the T_p (~1200 °C) predicted for the

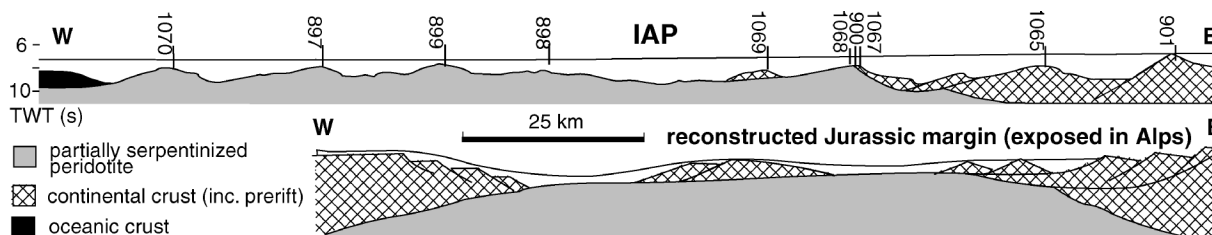


Figure 1. Cross sections across typical nonvolcanic rifted margins (NVRMs). Cross-hachures—rifted crust; white—postrift sequence; gray—mantle; black—oceanic crust. Top: Iberia Abyssal Plain margin (IAP; Whitmarsh et al., 2001) along Ocean Drilling Program drilling transect (numbered lines are Leg 149 and 173 drill sites). TWT—two-way traveltime. Below: reconstruction of margin now exposed in Alps (Froitzheim and Manatschal, 1996). Mantle rocks form top of basement at both these and other NVRMs.

base of cratonic lithosphere from heat flow models (Jaupart et al., 1998), expected values of thermal conductivity and heat production in the mantle, and heat flow across the cratonic Moho (Jaupart and Marschal, 1999).

More globally, variations in the depths to the 410 and 660 km discontinuities imply that temperatures deep beneath some continents are as much as 100 °C lower than those beneath the oceans (e.g., Gossler and Kind, 1996). Seismic velocity, constrained by heat flow data, provides an estimate of the continental geotherm (Röhm et al., 2000; Goes and van der Lee, 2002), revealing the expected variation in the lithosphere and different lithospheric thicknesses according to lithospheric type. At depths between 100 km (for tectonic provinces) and 200 km (for cratons), the geotherms become parallel to mantle adiabats, indicating the convecting sublithospheric mantle with a T_p in the range 1050–1350 °C (Fig. 4). Although active tectonic provinces have slightly higher average temperatures (~1250 °C) than cratons and stable platforms (both on average ~1200 °C), there is considerable overlap. Given the estimated errors of ± 150 °C (Röhm et al., 2000), even if the seismically derived geotherms systematically underestimate the temperature, many sublithospheric geotherms must occur well below the 1300 °C adiabat. If any errors are more random, it appears likely that away from hotspots the convecting subcontinental mantle has an average T_p closer to 1200 °C.

The relative similarity in temperature structure beneath different types of lithosphere implies that any compositional differences below 200 km have been removed by convective mixing: the idea of a deeper stable tectosphere has also been weakened by the demonstration that the mantle beneath ~200 km is anisotropic, requiring active shearing

(Gung et al., 2003). The resulting anisotropy may reduce the need for cool temperatures slightly but cannot eliminate it entirely, as even the subcontinental slow direction is faster than adjacent oceanic asthenosphere (Gung et al., 2003).

The temperature beneath the lithosphere may be controlled by the proximity of mantle plumes rather than by lithospheric structure (Röhm et al., 2000). The average geotherm for the western United States (Goes and van der Lee, 2002) (asymptotic to the 1300 °C adiabat; Fig. 4) may be influenced by both the Yellowstone and Raton plumes and by the influx of oceanic asthenosphere from the Gulf of California. In contrast, the average geotherm for eastern North America (away from such plumes) is asymptotic to the 1200 °C adiabat (Fig. 4).

In summary, the potential temperature of the convecting mantle beneath the continental lithosphere may in places be considerably cooler than 1300 °C (the T_p of oceanic asthenosphere) and perhaps as low as 1200 °C. The concept of a subcontinental convecting mantle cooler than that beneath the oceans is implicit in some models of mantle convection. Lenardic and Moresi (2000, 2003) used scaling arguments to show that if the continental and oceanic geotherms are coupled and in equilibrium, they must be different due to the thick, buoyant, and heat-producing continental crust. The continents may be underlain by downwelling limbs of cooled, plume-fed oceanic asthenosphere (Lenardic and Moresi, 2000), or beneath the 0–100-km-thick oceanic lithosphere, plume-fed oceanic asthenosphere may form a thin hot layer (Phipps Morgan et al., 1995) not present beneath the thicker continents, except where they overlie a mantle plume.

Figure 2. Plot of melt thickness vs. rift duration (Minshull et al., 2001) at lithospheric stretching factor of 50 (effectively infinite) for potential temperatures (T_p , curves) 1200, 1250, and 1300 °C. If mantle potential temperature was ~1300 °C, between 3 and 5 km of basaltic magma should be generated by decompression melting of upwelling mantle for rift durations of <25 m.y. Up to 1 km less melt thickness is possible at longer rift durations if rifting accelerated (dashed line) (Minshull et al., 2001). However, much less than 2 km of igneous rocks is observed at well-constrained nonvolcanic rifted margins with rift durations (Minshull et al., 2001; Wilson et al., 2001; Desmurs et al., 2001) shown by error bars (IAP—Iberia Abyssal Plain margin). Lithospheric stretching over cooler mantle ($T_p = \sim 1200$ °C) can explain this lack of melt and formation of nonvolcanic margins.

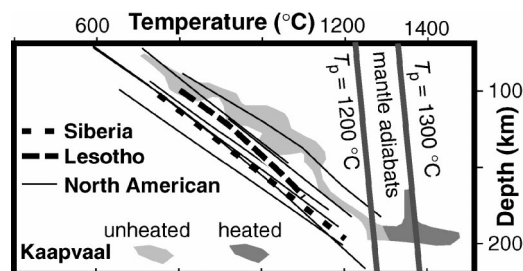
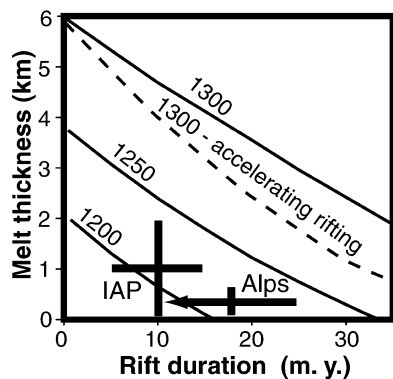


Figure 3. Cratonic geotherm from xenoliths in kimberlite pipes. Lines: steady-state geotherms derived from North American, Lesotho, and Siberian kimberlite pipes (Russel and Kopylova, 1999). Shading: geotherm derived from all Kaapvaal kimberlite xenoliths (Kohler and Brey, 1990). Dark shading: xenoliths from base of Kaapvaal lithosphere heated by mantle plume that led to kimberlite eruption (Crough et al., 1980). Unheated (light shading) and other steady-state geotherms reach 1200 °C adiabat at base of lithosphere, suggesting that away from plumes, convecting sublithospheric mantle has potential temperature (T_p) of ~1200 °C.

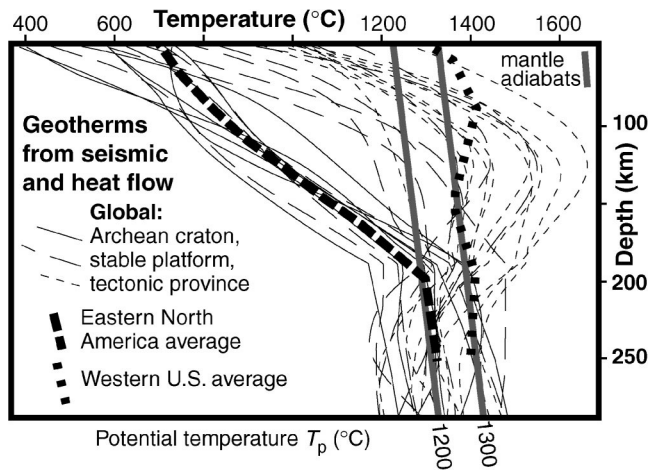


Figure 4. Continental geotherm from P-wave and S-wave velocities, and heat flow. Global curves: Röhm et al. (2000); averaged North America and U.S. curves: Goes and van der Lee (2002). Note large variations in lithospheric structure, but in convecting sublithospheric mantle, geotherms parallel mantle adiabats (gray lines) with potential temperature (T_p) commonly below 1300 °C.

CONSEQUENCES FOR RIFTING AND BREAKUP

Pressure-release melting of a hot ($T_p > 1400$ °C) mantle plume produces the thick sequences of igneous rocks found at volcanic rifted margins (White and McKenzie, 1989; Bown and White, 1995; Hopper et al., 2003). Lowering the T_p of the sublithospheric mantle to below 1300 °C similarly leads to a decrease in the amount of melt generated at effectively infinite stretching factors (Fig. 2). For a base lithospheric T_p of 1200 °C, <2 km of basalt is predicted, even for very rapid stretching (Minshull et al., 2001). As the mantle is too cool to melt sufficiently to produce new oceanic crust, continued stretching over several million years leads to separation of the continental crust followed by the amagmatic unroofing of a broad expanse of mantle (Perez-Gussinye et al., 2001). The little melt produced will be enriched in incompatible elements (Maaloe, 1985), as are those few basalts that have been sampled within the continent-ocean transition west of Iberia (Seyfert and Brunotte, 1996).

In our model, true seafloor spreading only begins at a nonvolcanic rifted margin when hotter suboceanic asthenosphere penetrates laterally beneath the rifting region. This influx follows the connection of the rift to the ridge network and is controlled in part by the shape of the base of the lithosphere, itself dependent on the three-dimensional evolution of the rifting and spreading system (one possibility, not meant to represent any particular margins, is shown simplified in Fig. 5). The rate of asthenospheric influx beneath the rift is a complex function of the temperature (hence viscosity and density) contrast between oceanic asthenosphere and cooler subcontinental mantle, the thickness of the asthenospheric layer, the proximity of mantle plumes, and the kinematics and geometry of rifting. At the Iberia Abyssal Plain, the J-anomaly ridge marks the onset of seafloor spreading and is consistent with the northward influx and subsequent upwelling of hot asthenosphere beneath a previously cool rift, leading to increased melting (Fig. 2). Farther north, the J-anomaly and ridge are absent, implying that breakup west of the Galicia Bank occurred later and that the spread of 1300 °C asthenosphere beneath the thinned lithosphere was more gradual. Initial oceanic crustal thickness there (Bown and White, 1995) is <4 km, perhaps reflecting the rate at which the warm asthenosphere could invade the rift rather than just cooling by the adjoining continental lithosphere.

Reduced sublithospheric temperatures are compatible with the subsidence history of rift basins (McKenzie, 1978) and rifted margins:

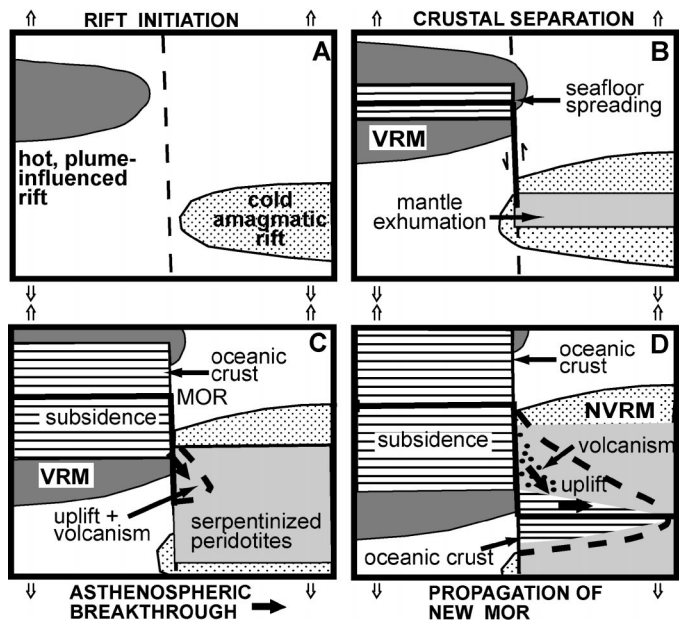


Figure 5. Plan-view sketches of possible evolution of adjacent volcanic rifted margin (VRM) and nonvolcanic rifted margin (NVRM) during breakup. A: VRM and NVRM start as plume-influenced and cool amagmatic rifts, respectively. B: Crustal separation leads to seafloor spreading and formation of oceanic crust (stripes) at VRM, but to mantle unroofing (light shading) at amagmatic rift. C: Hotter plume (or oceanic) asthenosphere invades amagmatic rift, initiating melting. D: Propagation of new mid-ocean ridge (MOR), accompanied by thermal uplift of NVRM, reflecting rate at which hotter asthenosphere penetrates (dashed line) beneath amagmatic rift. Escape of hot plume material from beneath VRM accelerates subsidence there.

the effect on synrift and postrift subsidence is less than that due to uncertainties in other parameters (e.g., lithospheric thickness or thermal-expansion coefficient). However, the change in sublithospheric temperature that occurs when hot asthenosphere invades the rift axis should have a resolvable effect on subsidence. The influx of hot asthenosphere from the global ridge system should not only initiate melting and the formation of oceanic crust, but also cause simultaneous transient thermal uplift of the margin, providing an explanation for the breakup unconformity found at many margins (Braun and Beaumont, 1989; Embry and Dixon, 1990). Most estimates of the required uplift (perhaps 200 m; Embry and Dixon, 1990) imply only modest rates of hot asthenosphere influx. Conversely, at volcanic margins, the transition to seafloor spreading should provide a pathway for hot plume material to escape laterally and be replaced by cooler mantle. At such margins, breakup should be marked by a pulse of accelerated subsidence that could be confused with a pulse of extension. Such an event occurred over much of the northwest European margin at the Paleocene-Eocene boundary, when seafloor spreading began in the Norwegian-Greenland Sea (Hopper et al., 2003; Joy, 1992). The reduced influence of the plume on northwest Europe after this time has also been inferred from the pattern of clastic deposition in the region (White and Lovell, 1997).

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