TECTONICS, VOL. 28, TC4016, doi:10.1029/2009TC002472, 2009

Cenozoic uplift of Europe

Eugenio Carminati,[1,2] Marco Cuffaro,[1,3] and Carlo Doglioni[1,2]

Received 15 February 2009; revised 4 June 2009; accepted 1 July 2009; published 29 August 2009.

[1] Following the diachronous Cretaceous–Neogene onset of seafloor spreading in the northern Atlantic and Arctic oceans, the European passive margin and continental Europe underwent a generalized uplift. This long-wavelength Cenozoic uplift has been attributed either to mantle plumes and/or to the far-field compression in the Alps. We suggest an alternative mechanism or concause, where the asthenosphere depleted at the Mid-Atlantic Ridge was shifted relatively eastward beneath the continent because of the net rotation of the lithosphere. The partial melting at the oceanic ridge leaves the asthenosphere lighter. When migrating beneath the European continental lithosphere, the substitution of the older, denser mantle with the depleted, lighter asthenosphere should have generated an isostatic rebound and associated uplift of about 300–600 m. Dynamic topography exerted by the mantle flowing eastward could have enhanced the uplift process. **Citation:** Carminati, E., M. Cuffaro, and C. Doglioni (2009), Cenozoic uplift of Europe, *Tectonics*, *28*, TC4016, doi:10.1029/2009TC002472.

1. Introduction

[2] Europe recorded Mesozoic diffuse subsidence associated with the Tethyan-Atlantic rifting, followed by later uplift (mainly starting in the Paleocene; Figures 1 and 2) responsible for the emersion of the European passive margin and of basin inversion of the European interior resulting in their elevation up to few hundreds meters above sea level. Why does continental Europe show uplifted sediments at an elevation that cannot be explained by the highest sea level rise? Why are marine sediments of the Mesozoic–early Cenozoic Paris Basin now a few hundreds meters above sea level? Are mantle plumes (e.g., Iceland) responsible for part or the overall emersion of the continent? Are "Alpine" inversion structures also responsible for this uplift [e.g., *Ziegler*, 1987; *Hillis et al.*, 2008]? Or is there another possible or coexisting explanation, as suggested by *Nielsen et al.* [2007], who related the intraplate deformation to the dynamics of the north Atlantic rift, discarding a plume origin for the uplift of Europe?

[1]Dipartimento di Scienze della Terra, Università La Sapienza, Rome, Italy.
[2]Also at IGAG, CNR, Rome, Italy.
[3]Now at MOX, Dipartimento di Matematica, Politecnico di Milano, Milan, Italy.

0278-7407/09/2009TC002472$12.00

[3] The plate-scale uplift or subsidence of continents is one of the most prominent phenomena in geodynamics [e.g., *Bond*, 1979; *Mitrovica et al.*, 1989; *Sengör*, 2003]. Short-term (<0.5 Ma) vertical movements can be easily explained by glacial isostatic adjustment [e.g., *Lambeck*, 1980], but long-term movements (>1 Ma) may have plurime origins, such as subsidence in foreland basins [*Doglioni*, 1994], development of dynamic topography [*Gurnis*, 1990; *Lithgow-Bertelloni and Gurnis*, 1997], thermal perturbations associated to mantle plumes or rifting processes [*Cazenave et al.*, 1989; *Burov and Guillou-Frottier*, 2005], erosion [*Medvedev et al.*, 2008], changes in the far-field stress associated with deformation phases at subduction, collision and rift zones [e.g., *Parmentier*, 1987; *Dèzes et al.*, 2004; *Nielsen et al.*, 2007], etc. In addition to uplift along orogens, clearly associated with subduction and collision zones such as the Alps, there may be uplift associated with inversion of intraplate structures previously generated by rifting processes. However, inversion structures form along narrow elongated belts (e.g., Iberia Chain and Polish trough) and cannot account for the uplift of the whole surrounding continent. Therefore, uplift and subsidence of widespread continents, far from plate boundaries (i.e., unrelated to tectonics) remains of quite obscure origin. The link between deep mantle processes and surface evolution is the topic of a number of ongoing international research projects [e.g., *Cloetingh et al.*, 2007]. For example, the yo-yo motion of the Australian continent has been attributed to the slab mass variations along the western Pacific subduction zones [*Gurnis et al.*, 1998]; Africa stands unusually high for a continent that has not undergone recent shortening, and it has a distinctive hypsometry reflecting broadscale uplift [*Harrison et al.*, 1983]. Erosion alone cannot explain why the whole continent uplifted and it is now in average about 1000 m above sea level. The uplift of Africa has been explained by the presence of a deep and hot raising mantle plume [e.g., *Gurnis et al.*, 2000]. However, this occurrence is debated, and low-velocity anomalies in tomographic images have also been interpreted as related not to positive temperature anomaly, but to compositional variations [*Anderson*, 2000] like larger Fe content [*Trampert et al.*, 2004]. In an alternative scenario, *Doglioni et al.* [2003] proposed that the uplift of Africa is not related to the upwelling of deep lighter mantle, but rather to the horizontal transit of depleted and lighter mantle under the continent. Most of the recent or active African volcanism is located along rift zones (e.g., Benue, Hoggar, Tibesti, Afar). Therefore, assuming like *Gurnis et al.* [2000] an isostatic uplift origin of the entire continental lithosphere, *Doglioni et al.* [2003] suggested a lateral flow of cooler and lighter mantle. In their reconstruction, they noted that most of the mid-oceanic ridges and rift zones in the world (apart along the Pacific Superswell) have an eastern flank a few hundred meters (100–300 m) shallower

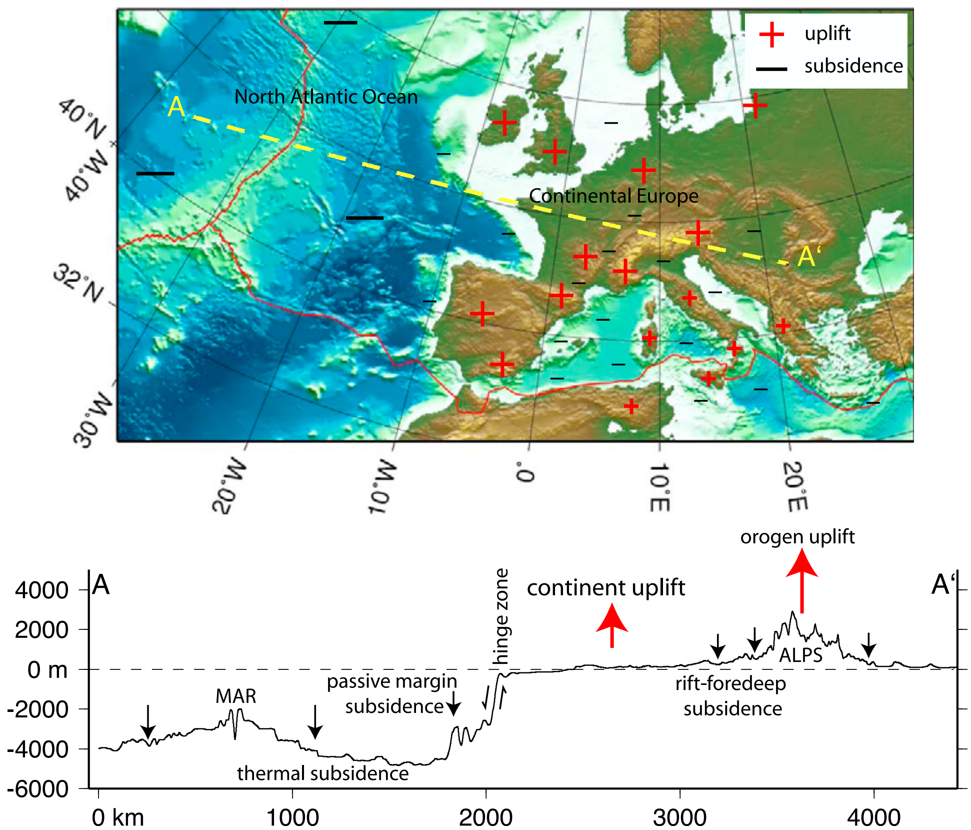

Figure 1. The European continental lithosphere underwent a generalized uplift since the Paleogene except along rift and foredeep zones. The thermal subsidence in the Atlantic Ocean has been contemporaneous with the uplift of continental Europe and still within the same plate. A fault-hinge area separating the two realms is needed along the passive continental margin. Subsidence and uplift signs represent both present and past vertical qualitative trends. MAR, Mid-Atlantic Ridge.

than the western one. This bathymetric asymmetry occurs also in remote oceanic basins not influenced by continental sediment input, excluding differential sediment loading as a potential cause of this asymmetry.

[4] Moreover, continents located to the east of oceanic basins (e.g., Africa) are generally more elevated than their conjugate margins to the west (e.g., South America), if tectonically active areas (such as the Andes) are neglected. This observation has been suggested to be related to (1) the depletion of the asthenosphere at the oceanic ridge and (2) the eastward relative mantle flow induced by the net "westward" rotation of the lithosphere [*Doglioni et al.*, 2005; *Scoppola et al.*, 2006]. In this paper, we test whether the asthenosphere depleted along the Mid-Atlantic Ridge, when shifting "eastward" because of the net rotation of the lithosphere, could have contributed to the Cenozoic uplift of continental Europe. The European vertical motions surely resulted from the diachronous activity of several local- and plate-scale processes and eustacy. However, we suggest that the eastward shift of depleted asthenosphere played a nonnegligible role. It is beyond the scope of the paper to discuss Miocene to recent vertical motions, occurring with time cycles of 10^4–10^5 years, induced and controlled by ice formation and melting, or localized intraplate inversion. We concentrated our study on the poorly constrained uplift history of the European continent, proposing a model that could be applied elsewhere, although each geodynamic setting clearly has variable erosion and sedimentation rates, sediment loading, extensional rates, etc., all factors controlling the vertical motions and the final structure. Our model is only referring to the origin of the overall uplift of the continent, and has not the ambition to solve the previous diachronous rifting episodes, and the intricate details that characterized vertical motions in the different regions of Europe and its margins.

2. Model

[5] Models of oceanic crust generation [*Bonatti et al.*, 1993] predict upwelling and partial melting of a fertile peridotitic mantle (density greater than 3300 kg m^{-3}) below ocean ridges. Extracted basaltic melt (density of 2800 kg m^{-3}) is emplaced at mid-ocean ridges to form the oceanic crust. Partial melting, up to a maximum of 25%, depletes the subridge fertile asthenosphere of high-density phases, such as garnet and produces a Fe-depleted harzburgite. The consequent decrease of the Fe/Mg ratio in depleted harzburgites lowers their density by about 2% (down to 3230 kg m^{-3}), relative to the undepleted mantle

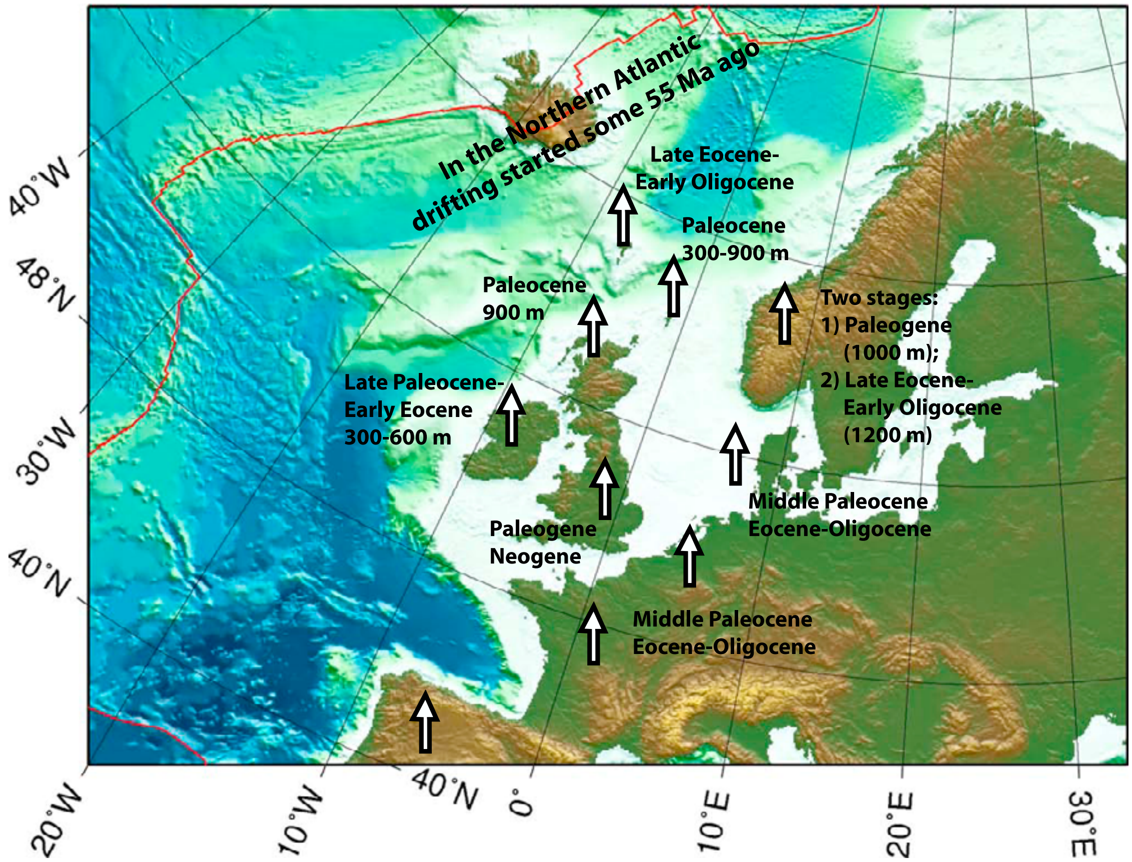

Figure 2. Age and location of uplift occurring in northwestern Europe during the Paleogene and the early Neogene. For sources of information, see text.

($\Delta\rho = -60$ kg m^{-3} [*Oxburgh and Parmentier*, 1977]). Partial melting may also reduce temperatures in the depleted asthenosphere. The density increase eventually associated with temperature decrease can be evaluated considering the relation $d\rho = -\rho\alpha_v dT$, where $\alpha_v = 3 \times 10^{-5}$ K^{-1} is the volumetric coefficient of thermal expansion, $\rho = 3230$ kg m^{-3} is the density and dT is the temperature decrease. Assuming a temperature decrease of 100 K, we obtain $d\rho$ of about 10 kg m^{-3}. Consequently, the density of the depleted asthenosphere, after correction for thermally induced volumetric contraction, should amount to ~3240 kg m^{-3}, far smaller than the density of the fertile asthenosphere. Therefore, a lighter asthenosphere is expected after melting below ridges. A similar process has been postulated for generating the about 1 km uplift of the oceanic seafloor around the Hawaii volcanic chain [*Phipps Morgan et al.*, 1995].

[6] The net rotation of the lithosphere, roughly westerly directed, was proposed by a number of authors [*Rittmann*, 1943; *Le Pichon*, 1968; *Bostrom*, 1971; *Nelson and Temple*, 1972; *Shaw*, 1973]. This so-called westward drift of the lithosphere relative to the mantle is testified to by independent kinematic observations, such as motions of major plates within the hot spot reference frames, using different databases (e.g., geological and geophysical data [*Ricard et al.*, 1991; *Gripp and Gordon*, 2002; *Cuffaro and Doglioni*, 2007] and space geodesy data [*Crespi et al.*, 2007]), and motions of the microplates [*Cuffaro and Jurdy*, 2006]. Classic evidence of the drift was also obtained by measuring plate motions relative to the Antarctica plate, that is considered to be nearly fixed with respect to the mantle [*Le Pichon*, 1968; *Knopoff and Leeds*, 1972], and from geological asymmetries [*Doglioni*, 1993, 1994; *Cruciani et al.*, 2005; *Doglioni et al.*, 2007]. It has been computed with a pole of rotation at about 55.9°S and 69.9°E (ω = 0.4359° Ma^{-1}) assuming, for the hot spots, a deep (at the core-mantle boundary) origin (deep hot spots reference frame [*Gripp and Gordon*, 2002]), and 60.2°S and 83.7°E (ω = 1.4901° Ma^{-1}) assuming an intra-asthenospheric origin of the hot spots (shallow hot spots reference frame [*Cuffaro and Doglioni*, 2007]).

[7] The rotation of the lithosphere determines a relative opposite (eastward) displacement of the underlying mantle, the main decoupling zone being located in the upper asthenosphere, in the low-velocity layer [*Panza*, 1980; *Thybo*, 2006]. The residual mantle depleted below ocean ridges, less dense and containing some melt and fluids [*Scott and Stevenson*, 1989], when displaced to the east relative to the lithosphere, should generate a mass deficit in the eastern limb of the ridge, while the western limb, where this low-density mantle

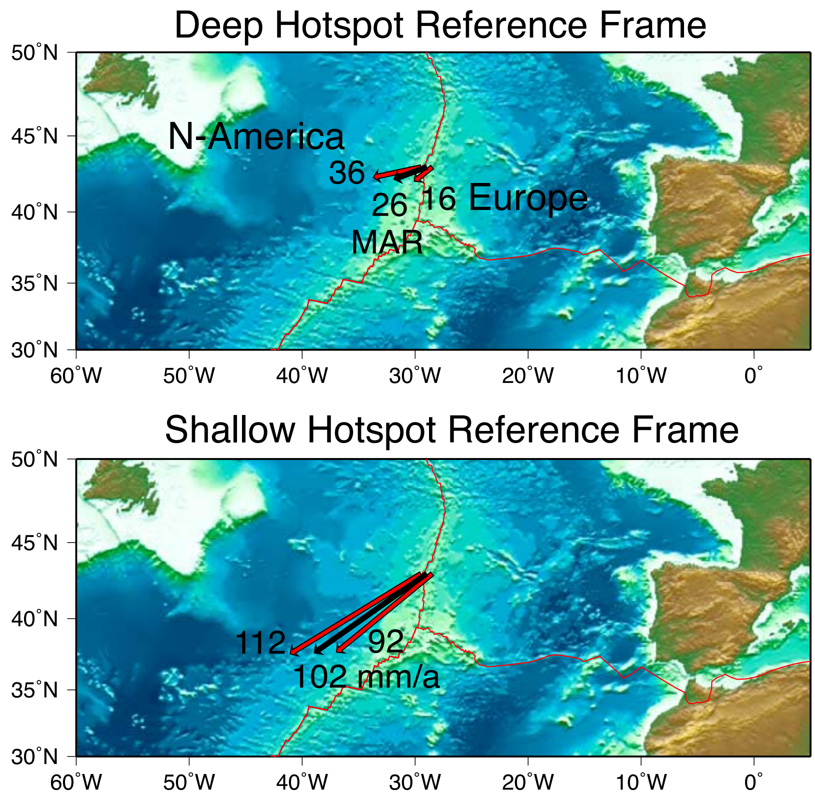

Figure 3. Velocities in mm a^{-1} of the North American plate, the MAR, and the European plate in two different reference frames, i.e., the deep [*Gripp and Gordon*, 2002] and shallow [*Crespi et al.*, 2007; *Cuffaro and Doglioni*, 2007] hot spot reference frames. Note that in both reference frames the MAR is moving westward with Europe, albeit with different velocities. The velocity of Europe should correspond with the counterflow to the "east" of the underlying mantle.

did not propagate, should remain unaffected by this mass deficit [*Doglioni et al.*, 2003, 2005]. As a consequence, the migration of the depleted asthenosphere from west to east (e.g., below the Atlantic) should induce a significant topographic asymmetry because of lower densities in the eastern side (e.g., Africa), therefore partly counteracting the thermal subsidence in the eastern flank of the oceanic ridge.

[8] For analyzing the North American and Eurasian plate motions in the context of a deep and a shallow hot spot reference frame together with the Mid-Atlantic Ridge, we selected points located at a latitude of 43°N, and we computed linear velocities, using plate rotation poles and angular velocities obtained by *Cuffaro and Doglioni* [2007] as shown in Figure 3. The motion of the Mid-Atlantic Ridge is calculated using the expression $V_r = (V_A + V_E)/2$, where V_r is the ridge velocity, V_E is Europe velocity and V_A is North America velocity. Figure 3 displays the calculation results and shows that the Mid-Atlantic Ridge moves to the west with respect to the mantle, regardless of the adopted reference frame (e.g., deep and shallow hot spots, respectively).

[9] The hot spot reference frame is debated for two reasons: (1) it is not firmly understood whether the plume tracks are sourced from deep or shallow (asthenospheric) mantle [*Foulger and Jurdy*, 2007] and (2) hot spots are located both within intraplate settings and along plate boundaries, providing evidence for multiple origins [*Doglioni et al.*, 2005]. Plate boundaries are by definition not stationary and laterally migrating. Therefore they cannot be considered fixed relative to each other like the plumes lying on or close to them. Regardless of these fundamental issues, the Pacific plate has a number of volcanic tracks, such as the Hawaii-Emperor chain, that form continuous trails that (1) allow us to infer the decoupling of the lithosphere from to the underlying mantle and (2) provide a unique independent reference frame. On these bases, we conclude that the Pacific plate is moving westward so fast (>10 cm a^{-1} [*Cuffaro and Doglioni*, 2007]) that the circuit crossing the Pacific, North America, and Europe plates cannot overtake this value, i.e., the sum of the longitudinal component of the Pacific–North America boundary, and the rifting between North America and Europe is not larger than 10 cm a^{-1}. It means that Europe has a westerly component of motion relative to the astenopsheric mantle, as suggested by *Gripp and Gordon* [2002]. *Le Pichon* [1968] also showed a westward shift of Europe relative to Antarctica.

[10] Assuming a deep origin of the Pacific plumes, at the latitude of about 42°N along the Mid-Atlantic Ridge, the

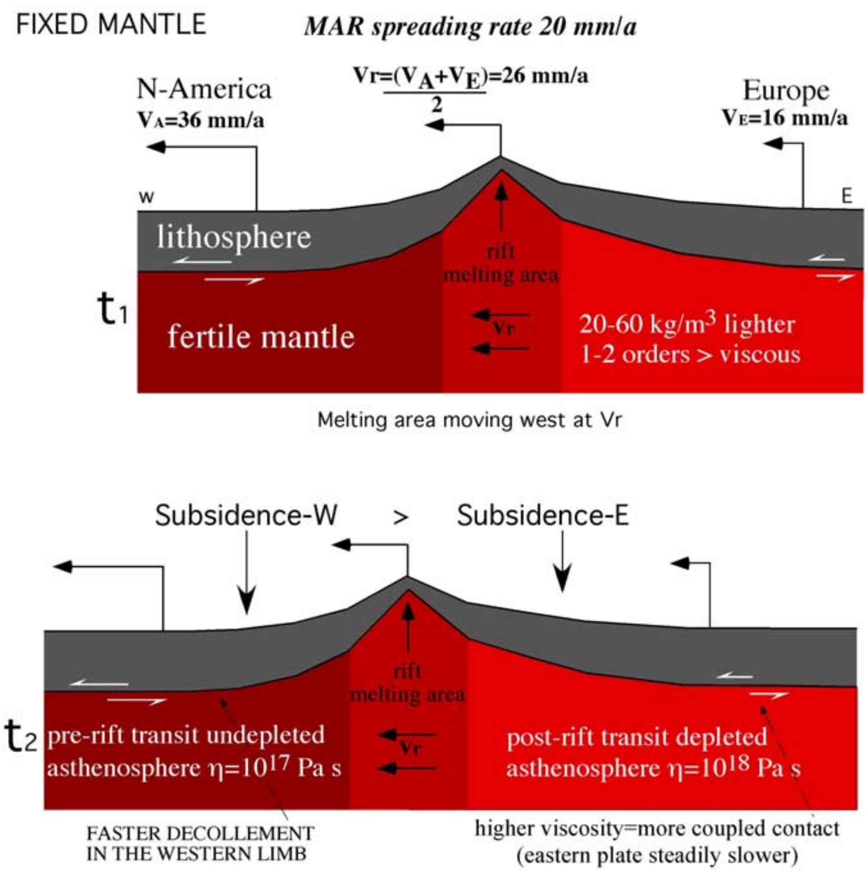

Figure 4. Assuming a fixed mantle, the North American lithosphere moves "west" faster than the European plate because the underlying asthenosphere is less viscous and the decoupling is more efficient. The MAR is also moving westward. The model velocities are the conservative ones of the deep hot spot reference frame. Along the ridge area, the depleted asthenosphere is lighter, determining a lower thermal subsidence in the eastern flank of the rift, which is underlain by an asthenospheric mantle shifting to the east with respect to the lithosphere. Because of the increase in viscosity and decrease in temperature along the rifting area, the asthenosphere below the eastern plate is more viscous, determining a stronger coupling and a steady state lower velocity of the European plate. These kinematics provide a continuous new fertile mantle underneath the oceanic ridge for generating mid-ocean ridge basalt (modified after *Doglioni et al.* [2005]). V_r, velocity of the ridge; V_E, velocity of Europe; V_A, velocity of North America. The melting and depletion area in the mantle is moving with the velocity of the ridge (V_r), affecting more and more fertile mantle to the west, which, migrating below the ridge, becomes depleted.

mantle would shift northeastward beneath the European plate at a speed of 16 km Ma^{-1} (Figure 3), i.e., a drift of 1600 km in 100 Ma. In case of intra-asthenospheric origin of the Pacific plumes, the speed would increase to 92 km Ma^{-1} (Figure 3), a velocity that would shift the underlying mantle at the transition between central and northern Atlantic 9200 km to the northeast. The faster velocity would imply a much faster correlation between the Atlantic rift and the uplift in the adjacent European continent.

[11] The slower velocity scenario would generate the uplift of western Europe alone, whereas the faster one would be able to explain the uplift of most of Eurasia continental areas. Since the wave would have been diachronous and "easterly" migrating, we should find a corresponding long-wavelength migrating erosional surface. However, this trend should have competed with local effects associated with regional tectonics (rifts, inversion structures, foreland areas, alpine uplift, etc.) rendering it difficult to recognize and separate the rate of such a general long-term and long-wavelength pattern. Further investigation is needed to quantify the different contributions to the vertical movements.

[12] Irrespective of the reference frame, the asthenosphere depleted beneath the Mid-Atlantic Ridge should move beneath the continental lithosphere to the east (i.e., Europe and Africa). This observation does not exclude the possible occurrence of small-scale mantle convection in the asthenosphere. We only assume the relative eastward motion of the mantle as a whole with respect to the lithosphere implicit with the notion of net rotation [e.g., *Le Pichon*, 1968]. However, convective asthenospheric flow is clearly required by mantle

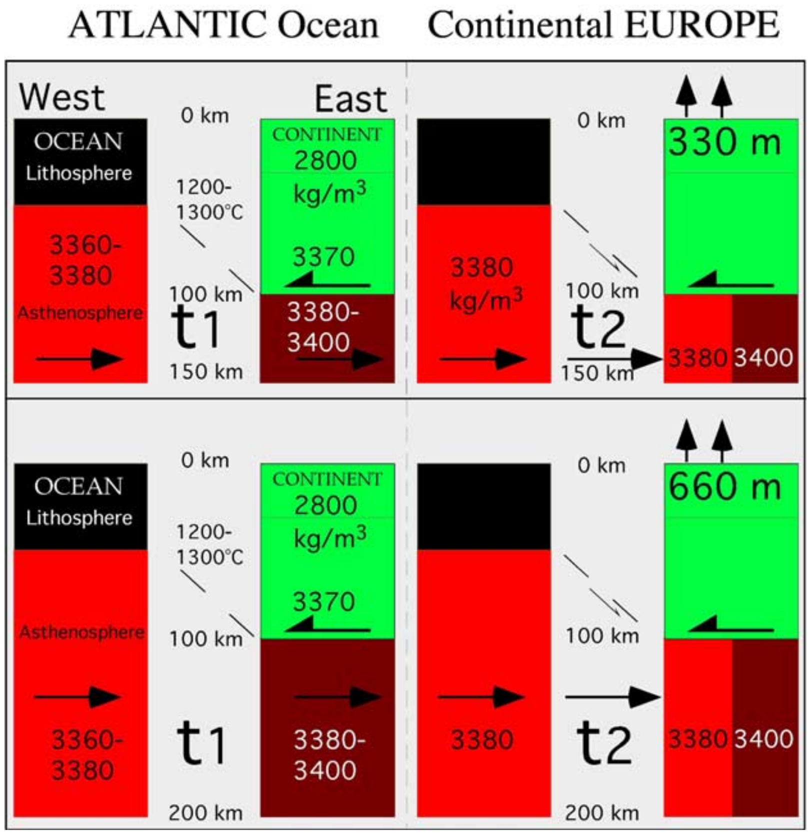

Figure 5. Two different models of density lateral variations assuming (top) a 50 km or (bottom) a 100 km thick asthenosphere beneath Europe. The depleted and lighter asthenophere at the ridge (t_1) migrating below the continent (t_2) would generate an uplift of 330 and 660 m, respectively (modified after *Doglioni et al.* [2003]).

rising along oceanic ridges [*Bonatti et al.*, 2003] and the downward directed corner flow effect along subduction zones [*Turcotte and Schubert*, 2002]. A number of models have also proposed convective rolls in the asthenosphere [e.g., *Korenaga and Jordan*, 2004] as a result of lithosphere-mantle decoupling. These movements internal to the sublithospheric mantle are here neglected for simplicity.

[13] Figure 4 displays the kinematics that can be inferred for the northern Mid-Atlantic Ridge in the conservative deep hot spot reference frame. An asymmetry of seismic velocity anomaly in the Atlantic asthenosphere is evident in mantle tomography [e.g., *Pilidou et al.*, 2005]. This may be interpreted as related to compositional, thermal and density changes, determining a decrease in the rigidity/density ratio that controls the velocity Vs of the *S* waves.

[14] We now assume local isostasy and consider a column composed by a 40 km thick continental crust (density of 2800 kg m^{-3}), a lithosperic mantle (LID) of 60 km (3370 kg m^{-3}) and 50 km (a part of) of asthenosphere (3400 kg m^{-3}; Figure 5). Assumed density values are consistent with the PREM [*Dziewonski and Anderson*, 1981; *Anderson*, 1989], which suggests average density up to 2900 kg m^{-3} in the crust and 3370 kg m^{-3} in the LID, whereas the asthenosphere may have a density as low as 3350 kg m^{-3}. The model of Figure 5 assumes a conservative 20 kg m^{-3} deficit of the asthenosphere at the ridge that is maintained in the upper asthenosphere transiting eastward below the eastern flank of the Atlantic ridge and beneath continental Europe. At a compensation depth arbitrarily chosen at 150 km (i.e., located within the asthenosphere), we would expect a load of about 4842 MPa. The replacement of the original (3400 kg m^{-3}) with lighter (3380 kg m^{-3}) asthenosphere would produce an uplift of Europe of about 330 m, necessary in order to keep constant the load at the assumed compensation depth. This calculation assumes local isostasy, thus neglecting the flexural character of the lithosphere. Increasing the thickness of the asthenosphere down to 200 km and assuming a 200 km compensation depth, would double to 660 m the isostatic continental rebound. Therefore the shift of lighter mantle beneath the European continent with an upper asthenosphere thickness of 50 or 100 km would generate uplift variable between 300 and 600 m. These values are fully compatible with the observed uplift of continental Europe. Note that western Europe (where lithospheric thickness is thinner [e.g., *Artemieva and Mooney*, 2001; *Sandoval et al.*, 2004]) underwent post–Atlantic rift uplift larger than eastern Europe, where the lithosphere is thicker: this could be

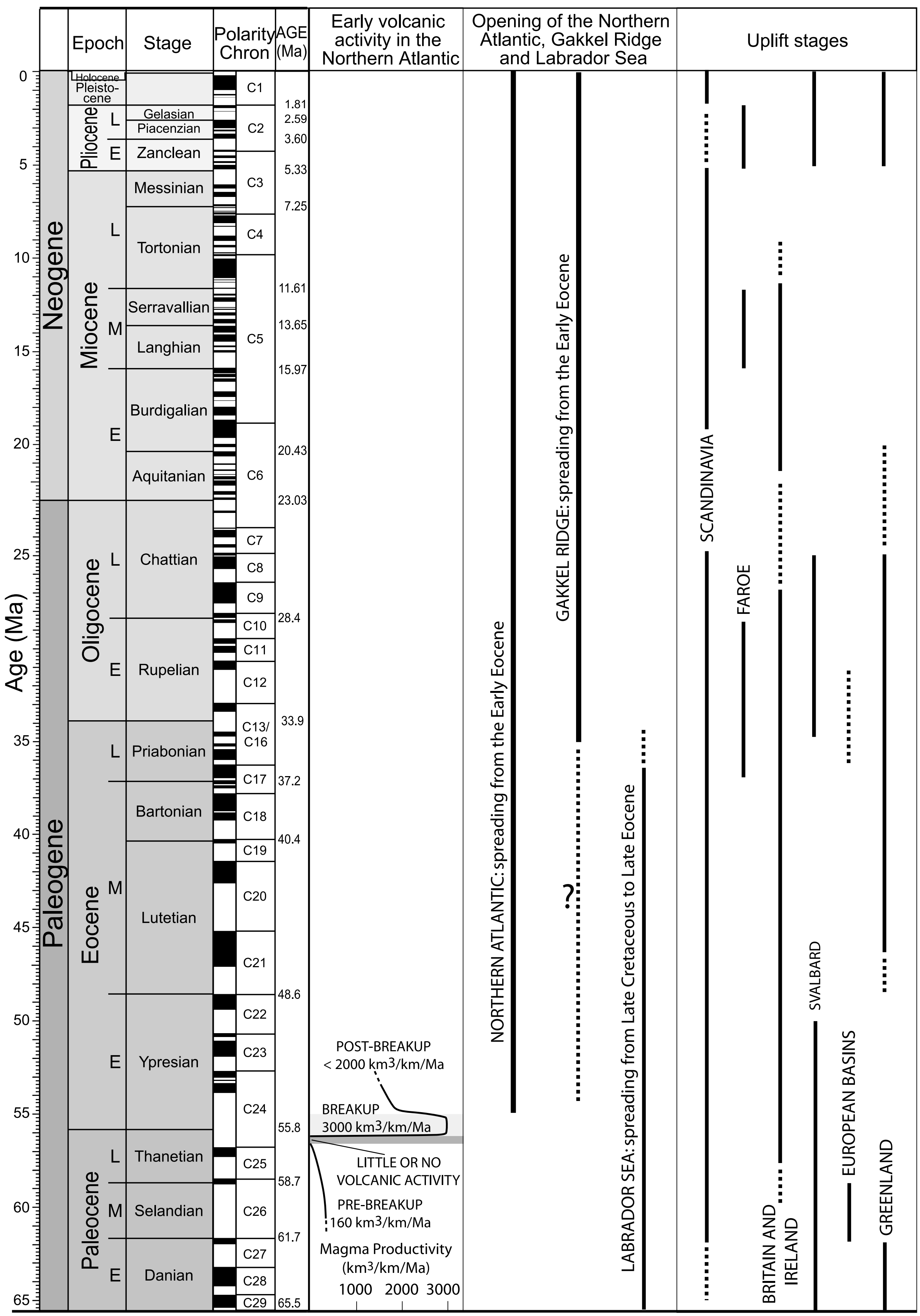
Epoch
Stage
Polarity Chron
AGE (Ma)
Early volcanic activity in the Northern Atlantic
Opening of the Northern Atlantic, Gakkel Ridge and Labrador Sea
Uplift stages
Age (Ma)
Neogene
Paleogene
Holocene
Pleisto-cene
Pliocene
Miocene
Oligocene
Eocene
Paleocene
Gelasian
Piacenzian
Zanclean
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian
Ypresian
Thanetian
Selandian
Danian
1.81
2.59
3.60
5.33
7.25
11.61
13.65
15.97
20.43
23.03
28.4
33.9
37.2
40.4
48.6
55.8
58.7
61.7
65.5
POST-BREAKUP
< 2000 km3/km/Ma
BREAKUP
3000 km3/km/Ma
LITTLE OR NO VOLCANIC ACTIVITY
PRE-BREAKUP
160 km3/km/Ma
Magma Productivity (km3/km/Ma)
1000 2000 3000
NORTHERN ATLANTIC: spreading from the Early Eocene
GAKKEL RIDGE: spreading from the Early Eocene
?
LABRADOR SEA: spreading from Late Cretaceous to Late Eocene
SCANDINAVIA
FAROE
BRITAIN AND IRELAND
SVALBARD
EUROPEAN BASINS
GREENLAND

Figure 6

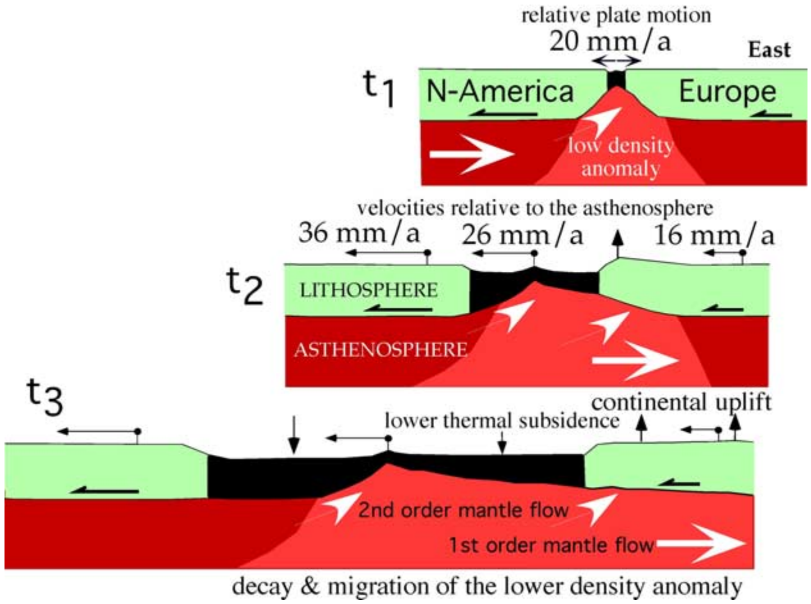

Figure 7. Cartoon of the long-term 300–600 m uplift of Europe generated by the eastward shift of the depleted lighter asthenosphere from the ridge beneath the continental lithosphere and possibly coupled with dynamic topography.

explained by a thicker upper asthenospheric section in western Europe.

3. Model Predictions Versus Geological Observations

[15] Following the diachronous Cretaceous (Iberia–Grand Banks, Bay of Biscay, Labrador Sea), Paleogene and Neogene (Knipovich ridge) onset of seafloor spreading in the northern Atlantic and Arctic oceans [*Vogt et al.*, 1979; *Torsvik et al.*, 2001], the European passive margin showed a generalized uplift (Figures 2 and 6) [*Japsen and Chalmers*, 2000]. Although in literature such an uplift is normally related to the activity of mantle diapirism related to the Iceland hot spot [e.g., *Rohrman and van der Beek*, 1996], we show that an alternative explanation may be provided by the eastward motion of asthenospheric mantle depleted by partial melting beneath the mid-ocean ridge, as discussed in section 2. In this section, the observations available on the vertical motions of the various parts of the European margin will be summarized. The rough assumption of local isostasy, the rather unconstrained density change of the depleted asthenosphere and the lateral variations of crust, mantle and asthenosphere thicknesses lead us to consider in the following discussion only the order of magnitude of the predicted uplift. In particular, it cannot be excluded that a significant contribution to the long-wavelength topography of the European continent came from dynamic topography, not included in the calculations, related to the interaction between the eastward flowing mantle and the continental keel. Dynamic topography is generally associated with dominant upward mantle flow [e.g., *Gurnis*, 1990]. The relative eastward mantle flow implicit with the net lithospheric rotation could contribute with a first-order flow, and smaller scale (second-order) flows could theoretically be able to generate dynamic topography (Figure 7).

[16] A further complication is due to the superposition on the eastward flow of depleted asthenosphere of (potentially) more than one process (e.g., thermal subsidence, active tectonics, glacial isostasy) generating either concordant or opposing vertical motions. The timescale at which these various processes operate is very variable.

[17] Finally, sediment loading and erosion are known to enhance subsidence and uplift [e.g., *Montgomery*, 1994; *Pelletier*, 2004; *Champagnac et al.*, 2007]. *Stern et al.* [2005] suggested that erosion in mountain ranges can account for 25% of regional uplift in temperate regions, whereas up to 50% of uplift can be attributed to glacial erosion in polar climates. It has been calculated that glacial erosion could have generated an uplift of ca 1.1 km in east Greenland [*Medvedev et al.*, 2008], and the deposition of the eroded material on the adjacent continental platform may be associated with subsidence of the order of 500 m. The same authors suggested that a similar mechanism may be responsible for (at least) part of the uplift along the remaining coasts of the

Figure 6. Chronology of the main tectonic and volcanic events along the northern Atlantic, Gakkel, and Labrador Sea ridges. The time distribution of uplift events occurring in Greenland, the Färoe Islands, Scandinavia, Svalbard, and the British Isles is also shown. The estimated melt production rates for the northern Atlantic Paleogene–early Eocene volcanism are after *Storey et al.* [2007]. Spreading and uplift ages are discussed in the text. The International Commission on Stratigraphy (available at http://www.stratigraphy.org) timescale is adopted. L, late; M, middle; E, early.

North Atlantic. Since these two processes (erosion and sedimentation) are laterally largely variable, their inclusion in the calculations would be highly speculative. It is, however, stressed that such processes may influence the rates but not the pattern (distribution of uplift and subsidence in a region) of vertical motions.

[18] Moreover, the uncertainties in the kinematic reconstructions both for the present and the past (e.g., shallow versus deep hot spot reference frames) are reflected in the prediction of the timing of continental uplift, which is in turn controlled by the relative eastward velocity of the asthenospheric mantle with respect to the lithosphere. The depleted mantle should move eastward relative to the lithosphere with rates between 16 and 92 km Ma^{-1}. As a consequence, assuming a passive margin some 250 km wide, the depleted asthenosphere will take between 15 and 3 Ma (after breakup) to cover the distance. It has been shown by *Doglioni et al.* [2003] that asymmetries are observed also across continental rifts. Rifts characterized by considerable volcanicity could be associated with mantle upwelling and partial melting with the production of a depleted asthenosphere. In such cases, uplift may even predate the breakup because of the migration of this prebreakup depleted asthenosphere.

[19] For all these reasons, the comparison between model predictions and geological observables (in terms of where, how much and when uplift occurred) will necessarily be qualitative. Particular attention will be given to the timing of the beginning of uplift, with respect to the onset of seafloor spreading, in order to assess whether uplift was compatible with the proposed mechanism.

[20] Large areas of the western European continental margin (Scandinavia, Britain, Färoe Platform, and Ireland) and of the inverted basins of the European interior (e.g., Danish Basin, North Sea Basin, and Paris Basin [e.g., *de Graciansky and Jacquin*, 2003] and the Iberia plate [e.g., *Casas-Sainz and de Vicente*, 2009]) were uplifted since the end of the Paleogene and through the Neogene, except ongoing rift zones, which were possibly uplifted at a later stage (e.g., late Miocene–Pliocene). The Paleogene uplift was roughly coincident with the onset of formation of oceanic crust along the Mid-Atlantic Ridge at these latitudes.

[21] It is here stressed that the Paleogene uplift, slightly predating the 55 Ma opening of the northern Atlantic at the latitudes of the above listed regions, may have been induced by the asthenosphere depletion associated with the volcanism in the final rifting stages, rather than by the partial melting beneath an active oceanic ridge. *Storey et al.* [2007] showed that between 61 and 57 Ma (middle–late Paleocene), prebreakup magmatism (Figure 6) occurred in the north Atlantic region (Greenland, Färoe Islands and western British Isles) with an average melt production of 160 km^3 per kilometer of rift per million years. After a short period of little or no volcanism, the average melt production rate increased at 56.1 Ma by more than an order of magnitude over previous levels with an average production of more than 3000 km^3 per kilometer of rift per million years (breakup phase) and diminished at about 55 Ma to less than 2000 km^3 per kilometer of rift per million years (postbreakup phase).

[22] This simple scenario is complicated by the fact that, coeval with the opening of the northern Atlantic, other oceanic basins (in particular the Labrador Sea) were developing and that the opening of the northern Atlantic was diachronous (older, circa 55 Ma, in the southern part and younger in the northern Gakkel Ridge, starting some 35 Ma ago). The opening of the Labrador Sea likely affected the landmasses to the east of the basin (included Greenland), as it will be more widely discussed below. The younger opening of the Gakkel Ridge is instead associated with a post–35 Ma uplift in the European continental margin at high latitudes, as it will be discussed for the Svalbard region.

3.1. Scandinavia

[23] In Scandinavia, significant uplift (of the order of 1000–2000 m in the interior regions, decreasing to ~500 m along the coast) took place since the Paleogene [*Peulvast*, 1978; *Jensen and Schmidt*, 1992; *Riis*, 1996; *Jordt et al.*, 1995; *Clausen et al.*, 1998, 2000; *Lidmar-Bergström et al.*, 2000]. Such estimates are based on structural data, apatite fission track data and geomorphological analyses [e.g., *Rohrman et al.*, 1995; *Redfield et al.*, 2005]. The recent most literature quite agrees in recognizing different phases of uplift for southwestern Norway: in the Paleocene, during the Eocene–Oligocene transition, in the late Miocene–early Pliocene and possibly in the late Pliocene–Pleistocene time [*Redfield et al.*, 2005, and references therein].

[24] Such vertical motions were accompanied, at places, by inversion of previous normal faults [*Dorè and Lundin*, 1996; *Mosar et al.*, 2002], local doming, and broad basin inversion [*Lundin and Dorè*, 2002] during two stages (late Eocene–Oligocene and Miocene). Thus, rifting-related structures locally played an important role in the postrift inversion and in the uplift of Norway, producing minor perturbations to the thermal subsidence on the Norwegian Shelf and in the North Sea [*Nielsen et al.*, 2002].

[25] The Paleogene uplift has been explained in several ways: transient dynamic uplift from the Iceland plume [e.g., *Nadin et al.*, 1995], permanent uplift related to underplating induced by the plume [e.g., *Jones et al.*, 2002], and intraplate stress [e.g., *Cloetingh et al.*, 1990, 2008]. Many models have been proposed for the Neogene uplift of Scandinavia: isostatic rebound from glacial erosion after removal of the ice load, intraplate stress from plate reorganization, edge-driven mantle convection, or mantle diapirism [*Olesen et al.*, 2002; *Cloetingh et al.*, 2005; *Redfield et al.*, 2005] (see also the discussion given by *Ebbing and Olesen* [2005]). In our opinion, a significant part of the uplift of Scandinavia was induced by the eastward flow of depleted asthenosphere, although local tectonics-related uplift clearly occurred.

[26] During the late Eocene, the gradual uplift of the Fennoscandian Shield was contemporaneous with subsidence in the North Sea, in the Norwegian-Danish Basin and in the Central Basin [*Jordt et al.*, 1995]. *Ziegler* [1988] suggested that the subsidence pattern of the North Sea can be explained by the decay of the thermal anomaly that devel-

oped during Triassic–late Paleocene rifting. The contrasting vertical motions in the Fennoscandian Shield and in the North Sea may be explained by the interference between uplift induced by the eastward shift of depleted asthenosphere and thermal subsidence in the strongly stretched North Sea. Uplift prevailed in the mainland and subsidence in the Norwegian-Danish Basin and in the Central Basin. It is likely that this pattern of vertical motions was emphasized by erosion in the uplifting regions and sedimentation of eroded material in the subsiding basins, as previously discussed.

3.2. Färoes, Britain, Scotland, Western North Sea, and Ireland

[27] During the late Eocene–early Oligocene, a phase of domal uplift of the Färoe Platform was associated with compressional tectonics and caused postdepositional tilting of Eocene strata along the southern margin of the platform [*Andersen et al.*, 2000; *Lundin and Dorè*, 2002]. *Andersen et al.* [2000] also recognized a mid-Miocene phase of compressional tectonics, evidenced on seismic transects by gentle anticlines and associated reverse faults and hypothesized a Pliocene uplift of the Färoe Islands as suggested by a progradational wedge of sediments deposited on the eastern Färoe Platform. Three-dimensional seismic reflection data from the Färoe-Shetland Basin allowed *Champion et al.* [2008] to recognize a phase of transient uplift close to the Paleocene-Eocene boundary (circa 56 Ma) of over 490 m, culminating and decaying within 3 Ma.

[28] *Japsen* [1997] showed that there were two distinct uplift episodes in the western North Sea and in eastern Britain: one in the Paleogene and one in the Neogene. The Paleogene phase (post–circa 60 Ma) mainly affected present onshore Britain and brought to an exhumation of about 1000 m. The Neogene phase affected both onshore areas and the western North Sea and was associated with a similar exhumation.

[29] *Hall and Bishop* [2002] and *Mudge and Jones* [2004] provided stratigraphic evidence for a mid-Paleocene uplift event in the Scotland-Shetland and northwestern North Sea region. *Hall and Bishop* [2002] also recognized a tectonic pulse on a lesser scale starting in late Oligocene time and continuing into late Neogene time. *Nadin et al.* [1997], by means of forward and reverse 2-D modeling of synrift and postrift stratigraphy, calculated a Paleocene uplift in the order of 375–525 m in the northwestern North Sea region and 900 m in the Shetland Basin. *MacLennan and Jones* [2006] suggested that uplift in Britain accompanying continental breakup was extremely rapid and may have taken place in just a few tens of thousands of years.

[30] According to *Jones et al.* [2001], the Porcupine Basin, offshore Ireland, was affected by a Jurassic rifting, followed by Cretaceous and Cenozoic postrift subsidence. However, on sedimentological bases, these authors identified 300–600 m uplift at the Paleocene-Eocene boundary. *Ware and Turner* [2002] analyzing sediment porosities in the east Irish Sea basin suggested that exhumation was mostly <1500 m and laterally highly variable. They interpreted such exhumation variations as the effects of Neogene basin inversion.

[31] *Allen et al.* [2002] evaluated the thermal and denudational history of Ireland using an apatite fission track data set and concluded that denudation rates were moderately high in Triassic time, falling to low values in Cretaceous time, and increasing substantially in the Tertiary. *Allen et al.* [2002] concluded that the cumulative amount of denudation during Tertiary time was generally between 1 and 2 km but without clear regional trends.

[32] In summary, a widespread phase of transient uplift of the order of several hundred meters was recognized close to the Paleocene-Eocene boundary (circa 56 Ma). *Champion et al.* [2008] suggested that the amplitude and duration of this transient effect are best explained by a mantle convective phenomenon. The synchronicity between this uplift phase and the beginning of oceanization in the segment of the Atlantic adjacent to these regions is consistent with its origin because of the transit of lighter asthenosphere depleted at the Mid-Atlantic Ridge, although local uplift motions are surely related to local compressional tectonics. Nevertheless, coeval subsidence due to either thermal cooling or active rifting may have locally prevented or obscured the record of the inferred uplift generated by the mantle-depleted anomaly.

3.3. Svalbard

[33] In Svalbard, the Paleogene Central Basin shows elevations between 0.6 and 1.1 km. *Blythe and Kleinspehn* [1998] reconstructed the regional thermal history (constrained by apatite and zircon fission track thermochronology and regional geological data) suggesting initial uplift of sediment source areas recorded by 70–50 Ma cooling signature and recognizing a later phase of uplift between 35 and 25 Ma, consistent with deposition patterns on and offshore. During the Paleogene the broad Barents Sea Shelf was uplifted and subjected to erosion without any significant internal deformation [*Cavanagh et al.*, 2006].

[34] The Late Cretaceous uplift in the Svalbard area was contemporaneous with spreading in the Makarov Basin and in the Labrador Sea (Late Cretaceous–late Eocene [*Srivastava and Roest*, 1999]). During this period Svalbard and Greenland were still in contact with each other. Although this early stage of uplift was synchronous to the opening of the Labrador Sea, it is not possible to relate it directly or entirely to the eastward flow of depleted and lighter asthenosphere, developing under the Labrador Sea ridge. During the Late Cretaceous–Paleogene, transpressional deformation occurred along the Senja Fracture Zone, a major fault zone located south of the Svalbard archipelago [*Ziegler*, 1988]. Latest Cretaceous–Paleocene uplift in the Svalbard Platform may be therefore related to transpressional tectonics. However, a contribution of the transit of lighter asthenosphere depleted at the Labrador Sea ridge under the Svalbard region cannot be excluded a priori.

[35] A later uplift phase occurred between 35 and 25 Ma. *Blythe and Kleinspehn* [1998] suggested that it was the result of rift-related uplift and erosion as the Atlantic spreading ridge propagated to the north some 35 Ma ago and started to separate Greenland from the Svalbard with very slow spreading along the Gakkel Ridge [*Vogt et al.*, 1979]. However, the age of the rifting in this part of the northern Atlantic Ocean

was recently set back to 53.3 Ma [*Engen et al.*, 2008]. Is the Oligocene uplift stage the result of the migration under this region of asthenosphere depleted below the Gakkel Ridge? A post–5 Ma stage of uplift and erosion in Svalbard was attributed to intense glacial denudation, consistently with the evidence of the first ice cover at 5.5 Ma.

3.4. Basins in the European Continental Interior

[36] During the Late Cretaceous and middle Paleocene, many Paleozoic and Mesozoic basins (e.g., Paris Basin, Iberian Basin, Danish Basin, Lower Saxony Basin, southern North Sea Basin) of the European plate interior were characterized by polyphase inversion and/or generalized uplift. *Cloetingh and van Wees* [2005] and *Nielsen et al.* [2005] showed that in the Late Cretaceous, basins were characterized by uplift along narrow zones, by reshear of extensional faults with reverse motion, crustal shortening and development of asymmetric marginal troughs. The middle Paleocene phase was characterized by dome-like uplift of wider areas, associated only with mild fault movements, and formation of more distal and shallower marginal troughs. However, in some areas (e.g., the Polish Basin [*Krzywiec*, 2006]) the late Paleocene uplift has been interpreted to be dominated by inversion, while dome-like uplift was less significant.

[37] This gentle broadscale warping and uplift of the European lithosphere caused the development of a regional erosional unconformity [*Dèzes et al.*, 2004]. These motions were plate wide and simultaneous. A further erosional phase occurred in the southern North Sea around the Eocene-Oligocene boundary [*Nielsen et al.*, 2005; *de Jager*, 2007]. Using apatite fission track data, *Japsen et al.* [2007] recognized a mid-Cenozoic exhumation phase (between 30 and 20 Ma) in the eastern North Sea Basin, associated with an erosion of about 800 to 1100 m. The same authors recognized later uplift phases of late Neogene (between 10 and 5 Ma) and the early Pliocene (at circa 4 Ma) times.

[38] *Nielsen et al.* [2005] interpreted the middle Paleocene phase as a domal, secondary inversion following inevitably the Cretaceous convergence-related inversion after the relaxation of the in-plane tectonic stress. They suggested that this stress change may have followed the elevation of the North Atlantic lithosphere by the Iceland plume [*White and Lovell*, 1997] or may have been triggered by the drop in the north-south convergence rate between Africa and Europe [*Rosenbaum et al.*, 2002].

[39] On the basis of the Paleocene inversion of some intra-European basins (e.g., the Mid-Polish Trough [*Krzywiec*, 2006]) and on the upthrusting of basement blocks in the Bohemian Massif and in southern Sweden [e.g., *Ziegler*, 1988; *Ziegler and Dèzes*, 2007], Paleocene uplift was interpreted as the result of compressional tectonics associated with continued crustal shortening controlled by the buildup of collision-related intraplate compressional stresses [*Ziegler*, 1987; *Ziegler et al.*, 1998; *Dèzes et al.*, 2004]. In any case, intraplate inversion and consequent uplift was concentrated along preexisting rifting-related "weak" zones. In addition to this scenario, we suggest that the middle Paleocene broadscale upwarping of the European plate interior could be, at least in part, related to the eastward propagation of asthenosphere depleted by the prebreakup magmatism occurring in the North Atlantic region (Greenland, Färoe Islands and western British Isles) during the Paleogene (Figure 6).

3.5. Uplift on the Western Side of the Atlantic: The Case of Greenland

[40] According to our model, the eastern North American margins should be unaffected by uplift related to the spreading of the northern Atlantic Ocean. However, several lines of evidence suggest that Greenland was characterized by a complex history of uplift in Cenozoic times [e.g., *Japsen and Chalmers*, 2000]. *Dam et al.* [1998] provide evidence for rapid uplift in the early Paleogene, shortly before the onset of Paleogene volcanism, both in west and east Greenland, contemporaneous to subsidence of several kilometers in the Nuussuaq Basin, central west Greenland [*Bonow et al.*, 2006a]. Although, according to their opinion, the timing of the uplift is not well constrained by borehole data, *Chalmers* [2000] concludes that an uplift stage occurred in southwestern Greenland after the early Eocene. During approximately the last 30 Ma, denudation was ~2 km inland and more than 3 km along the coast of southeastern Greenland, as revealed by fission track analyses [*Hansen*, 2000]. *Bonow et al.* [2006a], using apatite fission track and vitrinite reflectance maturity data recognized an event of uplift and erosion starting between 40 and 30 Ma in the Nuussuaq and Disko areas, central west Greenland. According to the same authors, a further uplift event started between 11 and 10 Ma and caused valley incision [*Bonow et al.*, 2006b].

[41] Finally, *Chalmers* [2000] provided evidence for a very late uplift stage (as late as the onset of glaciation in west Greenland), possibly of Plio-Pleistocene age. Paleogene uplift of Greenland was related in literature to the passage of the Iceland plume [e.g., *Lawver and Muller*, 1994; *Clift et al.*, 1998]. However, the passage of the hot spot across southern Greenland was dated at 45–40 Ma (i.e., much later) by *Lawver and Muller* [1994]. As a working hypothesis, we suggest that the early Paleogene uplift of western and eastern Greenland was induced by the eastward flow of depleted mantle associated with the spreading of the Labrador Sea (Late Cretaceous–late Eocene [*Srivastava and Roest*, 1999]). The contemporaneous subsidence in the Nuussuaq Basin is interpreted as the result of cooling following the last (Maastrichtian-Danian) stages of the rifting process affecting the basin [*Bonow et al.*, 2006b].

[42] However, in central and southern east Greenland, the rate of uplift seems to have accelerated in the Neogene [*Johnson and Gallagher*, 2000; *Mathiesen et al.*, 2000] and, according to *Chalmers* [2000], the post–early Eocene uplift of southern west Greenland took place substantially later than the cessation of magmatism in the early Eocene and the abrupt slowing or cessation of seafloor spreading in the Labrador Sea between chrons 20 and 13 (late Eocene). *Lawver and Muller* [1994] proposed the passage, underneath the southern east Greenland margin, of the Icelandic hot spot from a northwesterly direction between 40 and 35 Ma. This model is based on a global plate kinematic model and is consistent, according to the authors, with the synchronous uplift of this part of Greenland and with the occurrence of

39.8 Ma and 44.6 Ma syenite intrusions in the region. However, such a transit is not revealed directly by the thermal history reconstructed from fission track data from Greenland as discussed by *Hansen* [2000]. If the transit of the hot spot will be confirmed, the post–early Eocene uplift of southern Greenland is best explained by the hot spot activity. In polar areas, the submarine erosion operated by glaciers on the continental margin can generate uplift [*Stern et al.*, 2005; *Medvedev et al.*, 2008]. *Medvedev et al.* [2008] inferred that part of the uplift of east Greenland can be attributed to this process.

[43] Summarizing, the complex uplift of Greenland may be the result of the combination of four diachronous processes: spreading in the Labrador Sea and related eastward flow of depleted and lighter asthenosphere under Greenland (Paleocene stage); passage of the Iceland hot spot (post–early Eocene stage); erosion; ice cycles (Plio-Pleistocene stage). This complex pattern of uplift was further complicated by thermal subsidence in areas affected by Cretaceous–early Paleocene rifting events.

4. Miocene to Recent Vertical Motions

[44] The present work does not fully address and discuss the causes of Miocene to recent vertical motions in the European area. However, a few considerations deserve attention. A part of these motions were undoubtedly controlled by glacial isostatic adjustments associated with ice cycles. Others (late phases of uplift of the Scandinavian shield, differential uplift of the Bohemian, Armorican and Central massifs, accelerated subsidence in the North Sea–north German basin, continuing tectonics in the Rhine Rift System) have been interpreted as the result of the reactivation of preexisting crustal discontinuities and lithospheric folding controlled by a stress field (similar to the present), which came into being during the Miocene and further reinforced in the Pliocene [*Ziegler and Dèzes*, 2007]. Such a stress field has been modeled as the result of compressional forces related to the combination of the Atlantic ridge push and Alpine collision, with only a minor role for body forces induced by topography [*Grunthal and Stromeyer*, 1992; *Gölke and Coblentz*, 1996]. However, more recent numerical models [e.g., *Goes et al.*, 2000] predict, assuming the same combination of processes, stress directions deviating by 20–30° from the measured directions [*Müller et al.*, 1992]. This led *Goes et al.* [2000] to propose that Miocene–present tectonics and vertical motions of northwestern Europe are controlled by a combination of stresses associated with the forces acting on plate boundaries (Atlantic ridge push and collision in the Alps) and to regional mantle processes (buoyancy of anomalous mantle). The upper mantle under west and central Europe exhibits significantly lower velocities [e.g., *Bijwaard et al.*, 1998], interpreted by *Goes et al.* [2000] as being due to the presence of relatively hot asthenosphere. We interpret this observation as the evidence for the eastward flow of hotter and lighter asthenosphere depleted under the Mid-Atlantic Ridge, rather than the presence of a mantle plume. We suggest that these observations are promising for the exportation to the Miocene–present of the model proposed for the early and middle Cenozoic uplift of Europe.

5. Discussion and Conclusions

[45] The plate-scale uplift or subsidence of continents (e.g., Africa, Australia) has been so far generally assumed as triggered by the presence of deep mantle positive thermal anomalies (uplift), or the presence of a deep cold slab (temporary subsidence of Australia [*Gurnis et al.*, 1998]). Although we agree that the generalized epeirogenic movements of continents are associated with isostatic adjustment, in our research we further suggest that (1) the isostatic rebound can be controlled also by horizontal mantle flow rather than vertical mantle movements alone and that (2) the origin of the density anomaly might be controlled not only by thermal anomalies, but also by compositional variations.

[46] Available evidence and age dating indicate a post-Atlantic rift age for the uplift of the European continent. When looking at regional paleogeographic reconstructions [*Ziegler*, 1988], it appears that the European continent as a whole was not uplifted until crustal separation was achieved in the Arctic–North Atlantic domain. Its post–early Eocene subsidence and uplift pattern was very differentiated and did not follow a regionally simple pattern of progressive eastward advancing uplift. This observation would cast doubt on the general model proposed here. However, as already stated, the process could have started when parts of the European continent were still undergoing rifting, such as the North Sea or the Rhine and Rhône grabens, provided that magmatic activity was associated with stretching.

[47] The far-field effect of the Alps on the uplift of northern Europe has been questioned [*Michon and Merle*, 2005]. Furthermore, the supposed mantle plumes are not consistent with the low heat flow values, the paucity of volcanism in Europe (apart sporadic episodes along the intraplate rift zones, e.g., Rhine and Rhône grabens) and the ambiguous evidence from mantle tomography [*Lustrino and Carminati*, 2007].

[48] We interpret the long-term uplift of the continent as related to the underlying transit of a lighter asthenosphere, previously depleted along the Mid-Atlantic Ridge (Figures 7 and 8). In some areas (e.g., North Sea), local vertical motions controlled by tectonics (normal and reverse faulting) and by sedimentary loading and erosion were superimposed on plate-scale motions induced by plate tectonic mechanisms (e.g., postrift thermal relaxation of attenuated lithosphere, rift shoulder uplift, eastward shift of depleted asthenosphere, glacioisostasy). These processes were active on very different timescales. The interference resulted in a complicated pattern of uplifting and subsiding areas, with uplift and subsidence migrating through time. Moreover, if the long-wavelength uplift wave associated with the depletion of the mantle in the Mid-Atlantic Ridge was real, it should have interacted with local tectonic rates (both uplift and subsidence), and acted on a lithosphere with variable thickness and composition. This resulted in regional variable rates of vertical motions, these being the sum of different mechanisms.

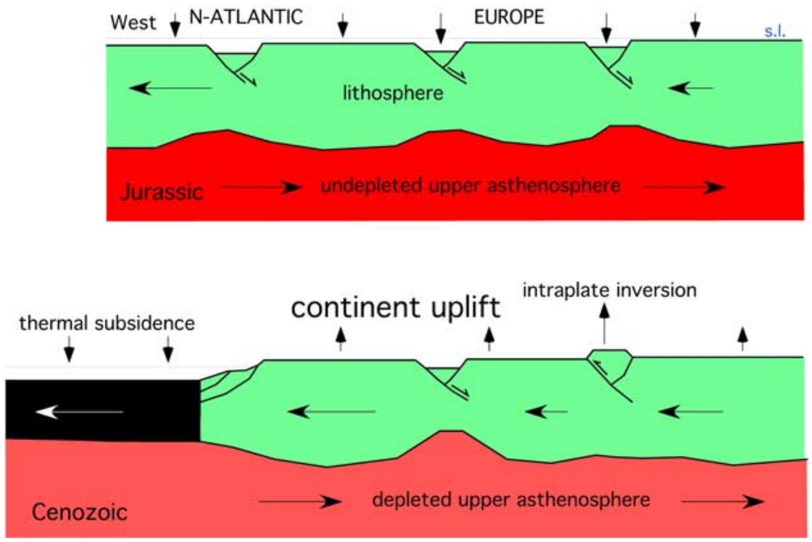

Figure 8. Continental Europe underwent diffuse rifting and subsidence during the Mesozoic and Cenozoic. After the opening of the Atlantic, the depletion of the underlying eastward flowing mantle was possibly contributing to a generalized uplift, while localized areas were inverted or still subsiding.

[49] The generalized continental uplift rate should have been maximal as the low-density anomaly was initially flowing from west to east beneath Europe; subsequently, in a steady state transit, it should have stabilized (Figure 8). However, uplift and subsidence may have occurred as a function of the degree of melting and of the composition of the mantle at the ridge. The eastward relative motion of the mantle could have enhanced the uplift process and a dynamically induced topography cannot be ruled out. Even if the mantle anomaly generated along the Mid-Atlantic Ridge were stationary through time, and the related uplift wave should have hypothetically remained in a steady state, other contemporaneous mechanisms controlling intracontinental vertical movements (e.g., active rifts, thermal relaxation, inversion of structures, deep mantle anomalies, lithospheric foreland flexure) could have laterally varied the wavelength and frequency of uplift (Figure 8).

[50] Along passive margins, the uplift of the continent is contrasting with the thermal subsidence in the oceanic realm. A hinge or a fault zone is necessarily required to separate the two areas. *Jackson et al.* [2005] have shown the occurrence of an active normal fault system along the Angola coast, which could separate the area of oceanic subsidence from that of continental uplift. Similarly, a hinge zone is expected to occur along the western European coast (Figure 1), separating the thermally and loaded subsiding passive continental margin from the uplifting continent. This hinge might be responsible for scattered, isolated "anomalous" seismicity (e.g., 1755, Lisbon? earthquake). In fact, besides the well known seismicity associated with plate boundaries in the Mediterranean and the rifts of the European realm, indeed segments of the Atlantic-European margin are marked by seismicity [*Cloetingh et al.*, 2007].

[51] **Acknowledgments.** The paper benefited from the comments of Daniel Bernoulli and an anonymous referee. Discussions with E. Bonatti, C. Faccenna, F. Innocenti, G. Panza, M. Ligi, F. Riguzzi, and D. Scrocca were very fruitful. Figures 1–3 were made with the Generic Mapping Tools of *Wessel and Smith* [1995]. Research was supported by Sapienza University. The research was supported by Eurocores-CNR (Topoeurope-Topo4D project).

References

Allen, P. A., et al. (2002), The post-Variscan thermal and denudational history of Ireland, in *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration*, edited by A. G. Doré et al., *Geol. Soc. Spec. Publ.*, *196*, 371–399, doi:10.1144/GSL.SP.2002.196.01.20.

Andersen, M. S., T. Nielsen, A. B. Sørensen, L. O. Boldreel, and A. Kuijpers (2000), Cenozoic sediment distribution and tectonic movements in the Faroe region, *Global Planet. Change*, *24*, 239–259, doi:10.1016/S0921-8181(00)00011-4.

Anderson, D. L. (1989), *Theory of the Earth*, 366 pp., Blackwell, Boston, Mass.

Anderson, D. L. (2000), Thermal state of the upper mantle: No role for mantle plumes, *Geophys. Res. Lett.*, *27*(22), 3623–3626, doi:10.1029/2000GL011533.

Artemieva, I. M., and W. D. Mooney (2001), Thermal structure and evolution of Precambrian lithosphere: A global study, *J. Geophys. Res.*, *106*, 16,387–16,414, doi:10.1029/2000JB900439.

Bijwaard, H., W. Spakman, and E. R. Engdahl (1998), Closing the gap between regional and global travel time tomography, *J. Geophys. Res.*, *103*, 30,055–30,078, doi:10.1029/98JB02467.

Blythe, A. E., and K. L. Kleinspehn (1998), Tectonically versus climatically driven Cenozoic exhumation of the Eurasian plate margin Svalbard: Fission track analyses, *Tectonics*, *17*, 621–639, doi:10.1029/98TC01890.

Bonatti, E., M. Seyler, and N. Sushevskaya (1993), A cold suboceanic mantle belt at the Earth's equator, *Science*, *261*, 315–320, doi:10.1126/science.261.5119.315.

Bonatti, E., M. Ligi, D. Brunelli, A. Cipriani, P. Fabretti, V. Ferrante, L. Gasperini, and L. Ottolini (2003), Mantle thermal pulses below the Mid-Atlantic Ridge and temporal variations in the oceanic lithosphere, *Nature*, *423*, 499–505, doi:10.1038/nature01594.

Bond, G. C. (1979), Evidence for some uplifts of large magnitude in continental platforms, *Tectonophysics*, *61*, 285–305, doi:10.1016/0040-1951(79)90302-0.

Bonow, J. M., P. Japsen, K. Lidmar-Bergström, J. A. Chalmers, and A. K. Pedersen (2006a), Cenozoic uplift of Nuussuaq and Disko, west Greenland—Elevated erosion surfaces as uplift markers of a passive margin, *Geomorphology*, *80*, 325–337, doi:10.1016/j.geomorph.2006.03.006.

Bonow, J. M., K. Lidmar-Bergström, and P. Japsen (2006b), Palaeosurfaces in central west Greenland as reference for identification of tectonic movements and estimation of erosion, *Global Planet. Change*, *50*, 161–183, doi:10.1016/j.gloplacha.2005.12.011.

Bostrom, R. C. (1971), Westward displacement of the lithosphere, *Nature*, *234*, 536–538, doi:10.1038/234536a0.

Burov, E., and L. Guillou-Frottier (2005), The plume head-continental lithosphere interaction using a tectonically realistic formulation for the lithosphere, *Geophys. J. Int.*, *161*, 469–490, doi:10.1111/j.1365-246X.2005.02588.x.

Casas-Sainz, A. M., and G. de Vicente (2009), On the tectonic origin of Iberian topography, *Tectonophysics*, doi:10.1016/j.tecto.2009.01.030, in press.

Cavanagh, A. J., R. Di Primio, M. Scheck-Wenderoth, and B. Horsfield (2006), Severity and timing of Cenozoic exhumation in the southwestern Barents Sea, *J. Geol. Soc.*, *163*, 761–774, doi:10.1144/0016-76492005-146.

Cazenave, A., A. Souriau, and K. Dominh (1989), Global coupling of Earth surface topography with hotspots, geoid and mantle heterogeneities, *Nature*, *340*, 54–57, doi:10.1038/340054a0.

Chalmers, J. A. (2000), Offshore evidence for Neogene uplift in central west Greenland, *Global Planet. Change*, *24*, 311–318, doi:10.1016/S0921-8181(00)00015-1.

Champagnac, J. D., P. Molnar, R. S. Anderson, C. Sue, and B. Delacou (2007), Quaternary erosion-induced isostatic rebound in the western Alps, *Geology*, *35*, 195–198, doi:10.1130/G23053A.1.

Champion, M. E. S., N. Y. White, S. M. Jones, and J. P. B. Lovell (2008), Quantifying transient mantle convective uplift: An example from the Faroe-Shetland Basin, *Tectonics*, *27*, TC1002, doi:10.1029/2007TC002106.

Clausen, O. R., M. Huuse, and O. Michelsen (1998), Palaeogene onset of the ''Neogene uplift'': Tectonics and sequence stratigraphy in the Danish North Sea, in *Neogene Uplift and Tectonics round the North Atlantic. International Workshop*, edited by L. Boldreel and P. Japsen, pp. 57–63, Geol. Surv. of Den. and Greenl., Copenhagen.

Clausen, O. R., O. B. Nielsen, M. Huuse, and O. Michelsen (2000), Geological indications for Palaeogene uplift in the eastern North Sea Basin, *Global Planet. Change*, *724*, 175–187, doi:10.1016/S0921-8181(00)00007-2.

Clift, P. D., A. Carter, and A. J. Hurford (1998), The erosional and uplift history of NE Atlantic passive margins: Constraints on a passing plume, *J. Geol. Soc.*, *155*, 787–800, doi:10.1144/gsjgs.155.5.0787.

Cloetingh, S., and J. D. van Wees (2005), Strength reversal in Europe's intraplate lithosphere: Transition from basin inversion to lithospheric folding, *Geology*, *33*, 285–288, doi:10.1130/G21051.1.

Cloetingh, S., F. M. Gradstein, H. Kooi, A. C. Grant, and M. A. Kaminski (1990), Plate reorganization: A cause of rapid late Neogene subsidence and sedimentation around the North Atlantic?, *J. Geol. Soc.*, *147*, 495–506, doi:10.1144/gsjgs.147.3.0495.

Cloetingh, S., P. A. Ziegler, F. Beekman, P. A. M. Andriessen, L. Matenco, G. Bada, D. Garcia-Castellanos, N. Hardebol, P. Dezes, and D. Sokoutis (2005), Lithospheric memory, state of stress andrheology: Neotectonic controls on Europe's intraplate continental topography, *Quat. Sci. Rev.*, *24*, 241–304, doi:10.1016/j.quascirev.2004.06.015.

Cloetingh, S., et al. (2007), TOPO-EUROPE: The geoscience of coupled deep Earth-surface processes, *Global Planet. Change*, *58*, 1–118, doi:10.1016/j.gloplacha.2007.02.008.

Cloetingh, S., F. Beekman, P. A. Ziegler, J. D. van Wees, and D. Sokoutis (2008), Post-rift compressional reactivation potential of passive margins and extensional basins, in *The Nature and Origin of Compression in Passive Margins*, edited by H. Johnson et al., *Geol. Soc. Spec. Publ.*, *306*, 27–70, doi:10.1144/SP306.2.

Crespi, M., M. Cuffaro, C. Doglioni, F. Giannone, and F. Riguzzi (2007), Space geodesy validation of the global lithospheric flow, *Geophys. J. Int.*, *168*, 491–506, doi:10.1111/j.1365-246X.2006.03226.x.

Cruciani, C., E. Carminati, and C. Doglioni (2005), Slab dip vs. lithosphere age: No direct function, *Earth Planet. Sci. Lett.*, *238*, 298–310, doi:10.1016/j.epsl.2005.07.025.

Cuffaro, M., and C. Doglioni (2007), Global kinematics in deep versus shallow hotspot reference frames, in *The Origins of Melting Anomalies: Plates, Plumes, and Planetary Processes*, edited by G. R. Foulger and D. M. Jurdy, *Spec. Pap. Geol. Soc. Am.*, *430*, 359–374.

Cuffaro, M., and D. M. Jurdy (2006), Microplate motions in the hotspot reference frame, *Terra Nova*, *18*, 276–281.

Dam, G., M. Larsen, and J. C. Sørensen (1998), Sedimentary response to mantle plumes: Implications from Paleocene onshore successions, west and east Greenland, *Geology*, *26*, 207–210, doi:10.1130/0091-7613(1998)026<0207:SRTMPI>2.3.CO;2.

de Graciansky, P. C., and T. Jacquin (2003), Evolution des structures et de la paléogéographie au passage Lias-Dogger dans le bassin de Paris d'après les données de la subsurface, *Bull. Soc. Geol. Fr.*, *174*(1), 3–17, doi:10.2113/174.1.3.

de Jager, J. (2007), Geological development, in *Geology of the Netherlands*, edited by T. E. Wong, D. A. J. Batjes, and J. de Jager, pp. 5–26, R. Neth. Acad. of Arts and Sci., Amsterdam.

Dèzes, P., S. M. Schmid, and P. A. Ziegler (2004), Evolution of the European Cenozoic Rift System: Interaction of the Alpine and Pyrenean orogens with their foreland lithosphere, *Tectonophysics*, *389*, 1–33, doi:10.1016/j.tecto.2004.06.011.

Doglioni, C. (1993), Geological evidence for a global tectonic polarity, *J. Geol. Soc.*, *150*, 991–1002, doi:10.1144/gsjgs.150.5.0991.

Doglioni, C. (1994), Foredeeps versus subduction zones, *Geology*, *22*, 271–274, doi:10.1130/0091-7613(1994)022<0271:FVSZ>2.3.CO;2.

Doglioni, C., E. Carminati, and E. Bonatti (2003), Rift asymmetry and continental uplift, *Tectonics*, *22*(3), 1024, doi:10.1029/2002TC001459.

Doglioni, C., D. Green, and F. Mongelli (2005), On the shallow origin of hotspots and the westward drift of the lithosphere, in *Plates, Plumes and Paradigms*, edited by G. R. Foulger et al., *Spec. Pap. Geol. Soc. Am.*, *388*, 735–749.

Doglioni, C., E. Carminati, M. Cuffaro, and D. Scrocca (2007), Subduction kinematics and dynamic constraints, *Earth Sci. Rev.*, *83*, 125–175, doi:10.1016/j.earscirev.2007.04.001.

Dorè, A. G., and E. R. Lundin (1996), Cenozoic compressional structures on the NE Atlantic margin: Nature, origin and potential significance for hydrocarbon exploration, *Petrol. Geosci.*, *2*, 299–311.

Dziewonski, A. M., and D. L. Anderson (1981), Preliminary reference earth model, *Phys. Earth Planet. Inter.*, *25*, 297–356, doi:10.1016/0031-9201(81)90046-7.

Ebbing, J., and O. Olesen (2005), The northern and southern Scandes—Structural differences revealed by an analysis of gravity anomalies, the geoid and regional isostasy, *Tectonophysics*, *411*, 73–87, doi:10.1016/j.tecto.2005.09.002.

Engen, A., J. I. Faleide, and T. K. Dyreng (2008), Opening of the Fram Strait gateway: A review of plate tectonic constraints, *Tectonophysics*, *450*, 51–69, doi:10.1016/j.tecto.2008.01.002.

Foulger, G. R., and D. M. Jurdy (Eds.) (2007), *Plates, Plumes and Planetary Processes*, *Spec. Pap. Geol. Soc. Am.*, *430*, 997 pp.

Goes, S., J. J. P. Loohuis, M. J. R. Wortel, and R. Govers (2000), The effect of plate stresses and shallow mantle temperatures on tectonics of northwestern Europe, *Global Planet. Change*, *27*, 23–38, doi:10.1016/S0921-8181(01)00057-1.

Gölke, M., and D. Coblentz (1996), Origins of the European regional stress field, *Tectonophysics*, *266*, 11–24, doi:10.1016/S0040-1951(96)00180-1.

Gripp, A. E., and R. G. Gordon (2002), Young tracks of hotspots and current plate velocities, *Geophys. J. Int.*, *150*, 321–361, doi:10.1046/j.1365-246X.2002.01627.x.

Grunthal, G., and D. Stromeyer (1992), The recent crustal stress field in central Europe: Trajectories and finite element modeling, *J. Geophys. Res.*, *97*, 11,805–11,820, doi:10.1029/91JB01963.

Gurnis, M. (1990), Bounds on global dynamic topography from Phanerozoic flooding of continental platforms, *Nature*, *344*, 754–756, doi:10.1038/344754a0.

Gurnis, M., R. D. Müller, and L. Moresi (1998), Cretaceous vertical motion of Australia and the Australian-Antarctic discordance, *Science*, *279*, 1499–1504, doi:10.1126/science.279.5356.1499.

Gurnis, M., J. Mitrovica, J. Ritsema, and H. J. van Heijst (2000), Constraining mantle density structure using geological evidence of surface uplift rates: The case of the African superplume, *Geochem. Geophys. Geosyst.*, *1*(7), 1020, doi:10.1029/1999GC000035.

Hall, A., and P. Bishop (2002), Scotland's denudational history: An integrated view of erosion and sedimentation at an uplifted passive margin, in *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration*, edited by A. G. Doré et al., *Geol. Soc. Spec. Publ.*, *196*, 271–290.

Hansen, K. (2000), Tracking thermal history in east Greenland: An overview, *Global Planet. Change*, *24*, 303–309, doi:10.1016/S0921-8181(00)00014-X.

Harrison, C. G. A., K. J. Miskell, G. W. Brass, E. S. Saltzman, and L. Sloan (1983), Continental hypsography, *Tectonics*, *2*, 357–377, doi:10.1029/TC002i004p00357.

Hillis, R. R., S. P. Holford, P. F. Green, A. G. Doré, R. W. Gatliff, M. S. Stoker, K. Thomson, J. P. Turner, J. R. Underhill, and G. A. Williams (2008), Cenozoic exhumation of the southern British Isles, *Geology*, *36*, 371–374, doi:10.1130/G24699A.1.

Jackson, M. P. A., M. R. Hudec, and K. A. Hegarty (2005), The great West African Tertiary coastal uplift: Fact or fiction? A perspective from the Angolan divergent margin, *Tectonics*, *24*, TC6014, doi:10.1029/2005TC001836.

Japsen, P. (1997), Regional Neogene exhumation of Britain and the western North Sea, *J. Geol. Soc.*, *154*, 239–247, doi:10.1144/gsjgs.154.2.0239.

Japsen, P., and J. A. Chalmers (2000), Neogene uplift and tectonics around the North Atlantic: Overview, *Global Planet. Change*, *24*, 165–173, doi:10.1016/S0921-8181(00)00006-0.

Japsen, P., P. F. Green, N. H. Nielsen, E. S. Rasmussen, and T. Bidstrup (2007), Mesozoic-Cenozoic exhumation events in the eastern North Sea Basin: A multi-disciplinary study based on palaeothermal, palaeoburial, stratigraphic and seismic data, *Basin Res.*, *19*, 451–490, doi:10.1111/j.1365-2117.2007.00329.x.

Jensen, L. N., and B. J. Schmidt (1992), Late Tertiary uplift and erosion in the Skagerrack area: Magnitude and consequences, *Nor. Geol. Tidsskr.*, *72*, 275–279.

Johnson, C., and K. Gallagher (2000), A preliminary Mesozoic and Cenozoic denudation history of the north east Greenland onshore margin, *Global Planet. Change*, *24*, 261–274, doi:10.1016/S0921-8181(00)00012-6.

Jones, S. M., N. White, and B. Lovell (2001), Cenozoic and Cretaceous transient uplift in the Porcupine Basin and its relationship to a mantle plume, in *The Petroleum Exploration of Ireland's Offshore Basins*, edited by P. M. Shannon, P. D. W. Haughton, and

D. V. Corcoran, *Geol. Soc. Spec. Publ.*, *188*, 345–360.

Jones, S. M., N. White, B. J. Clarke, E. Rowley, and K. Gallagher (2002), Present and past influence of the Iceland plume on sedimentation, in *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration*, edited by A. G. Doré et al., *Geol. Soc. Spec. Publ.*, *196*, 13–25.

Jordt, H., J. I. Faleide, K. Bjørlykke, and M. T. Ibrahim (1995), Cenozoic sequence stratigraphy of the central and northern North Sea Basin: Tectonic development, sediment distribution and provenance areas, *Mar. Pet. Geol.*, *12*, 845–879, doi:10.1016/0264-8172(95)98852-V.

Knopoff, L., and A. Leeds (1972), Lithospheric momenta and the deceleration of the Earth, *Nature*, *237*, 93–95, doi:10.1038/237093a0.

Korenaga, J., and T. H. Jordan (2004), Physics of multiscale convection in Earth's mantle: Evolution of sublithospheric convention, *J. Geophys. Res.*, *109*, B01405, doi:10.1029/2003JB002464.

Krzywiec, P. (2006), Structural inversion of the Pomeranian and Kuiavian segments of the Mid-Polish Trough—Lateral variations in timing and structural style, *Geol. Q.*, *50*, 151–168.

Lambeck, K. (1980), *The Earth's Variable Rotation: Geophysical Causes and Consequences*, edited by G. K. Batchelor and J. W. Miles, Cambridge Univ. Press, Cambridge, U. K.

Lawver, L. A., and R. D. Muller (1994), Iceland hotspot track, *Geology*, *22*, 311–314, doi:10.1130/0091-7613(1994)022<0311:IHT>2.3.CO;2.

Le Pichon, X. (1968), Sea-floor spreading and continental drift, *J. Geophys. Res.*, *73*, 3661–3697, doi:10.1029/JB073i012p03661.

Lidmar-Bergström, K., C. D. Ollier, and J. C. Sulebak (2000), Landforms and uplift history of southern Norway, *Global Planet. Change*, *24*, 211–231, doi:10.1016/S0921-8181(00)00009-6.

Lithgow-Bertelloni, C., and M. Gurnis (1997), Cenozoic subsidence and uplift of continents from time-varying dynamic topography, *Geology*, *25*, 735–738, doi:10.1130/0091-7613(1997)025<0735:CSAUOC>2.3.CO;2.

Lundin, E., and A. G. Dorè (2002), Mid-Cenozoic post-breakup deformation in the 'passive' margins bordering the Norwegian–Greenland Sea, *Mar. Pet. Geol.*, *19*, 79–93, doi:10.1016/S0264-8172(01)00046-0.

Lustrino, M., and E. Carminati (2007), Phantom plumes in Europe and the circum-Mediterranean region, in *The Origins of Melting Anomalies: Plates, Plumes, and Planetary Processes*, edited by G. R. Foulger and D. M. Jurdy, *Spec. Pap. Geol. Soc. Am.*, *430*, 723–745.

MacLennan, J., and S. M. Jones (2006), Regional uplift, gas hydrate dissociation and the origins of the Paleocene-Eocene Thermal Maximum, *Earth Planet. Sci. Lett.*, *245*, 65–80, doi:10.1016/j.epsl.2006.01.069.

Mathiesen, A., T. Bidstrup, and F. G. Christiansen (2000), Denudation and uplift history of the Jameson Land Basin, east Greenland—Constrained from previous maturity and new apatite fission track data, *Global Planet. Change*, *24*, 275–301, doi:10.1016/S0921-8181(00)00013-8.

Medvedev, S., E. H. Hartz, and Y. Y. Podladchikov (2008), Vertical motions of the fjord regions of central east Greenland: Impact of glacial erosion, deposition, and isostasy, *Geology*, *36*, 539–542, doi:10.1130/G24638A.1.

Michon, L., and O. Merle (2005), Discussion on "Evolution of the European Cenozoic Rift System: Interaction of the Alpine and Pyrenean orogens with their foreland lithosphere" by P. Dèzes, S. M. Schmid and P. A. Ziegler, *Tectonophysics 389* (2004) 1–33, *Tectonophysics*, *401*, 251–256, doi:10.1016/j.tecto.2005.01.006.

Mitrovica, J. X., C. Beaumont, and G. T. Jarvis (1989), Tilting of continental interiors by the dynamical effects of subduction, *Tectonics*, *8*, 1079–1094, doi:10.1029/TC008i005p01079.

Montgomery, D. R. (1994), Valley incision and the uplift of mountain peaks, *J. Geophys. Res.*, *99*, 13,913–13,921, doi:10.1029/94JB00122.

Mosar, J., G. Lewis, and T. H. Torsvik (2002), North Atlantic sea-floor spreading rates: Implications for the Tertiary development of inversion structures of the Norwegian-Greenland Sea, *J. Geol. Soc.*, *159*, 503–515, doi:10.1144/0016-764901-093.

Mudge, D. C., and S. M. Jones (2004), Palaeocene uplift and subsidence events in the Scotland-Shetland and North Sea region and their relationship to the Iceland Plume, *J. Geol. Soc.*, *161*, 381–386.

Müller, B., M. L. Zoback, K. Fuchs, L. Mastin, S. Gregersen, N. Pavoni, O. Stephansson, and C. Ljunggren (1992), Regional patterns of tectonic stress in Europe, *J. Geophys. Res.*, *97*, 11,783–11,803, doi:10.1029/91JB01096.

Nadin, P. A., N. J. Kusznir, and J. Toth (1995), Transient regional uplift in the early Tertiary of the northern North Sea and the development of the Iceland Plume, *J. Geol. Soc.*, *152*, 953–958, doi:10.1144/GSL.JGS.1995.152.01.12.

Nadin, P. A., N. J. Kusznir, and M. J. Cheadle (1997), Early Tertiary plume uplift of the North Sea and Faeroe-Shetland basins, *Earth Planet. Sci. Lett.*, *148*, 109–127, doi:10.1016/S0012-821X(97)00035-6.

Nelson, T. H., and P. G. Temple (1972), Mainstream mantle convection: A geologic analysis of plate motion, *Am. Assoc. Pet. Geol. Bull.*, *56*, 226–246.

Nielsen, S. B., G. E. Paulsen, D. L. Hansen, L. Gemmer, O. R. Clausen, B. H. Jacobsen, N. Balling, M. Huuse, and K. Gallagher (2002), Paleocene initiation of Cenozoic uplift in Norway, in *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration*, edited by A. G. Doré et al., *Geol. Soc. Spec. Publ.*, *196*, 45–65, doi:10.1144/GSL.SP.2002.196.01.04.

Nielsen, S. B., E. Thomsen, D. L. Hansen, and O. R. Clausen (2005), Plate-wide stress relaxation explains European Palaeocene basin inversions, *Nature*, *435*, 195–198, doi:10.1038/nature03599.

Nielsen, S. B., R. Stephenson, and E. Thomsen (2007), Dynamics of mid-Palaeocene North Atlantic rifting linked with European intra-plate deformations, *Nature*, *450*, 1071–1074, doi:10.1038/nature06379.

Olesen, O., E. Lundin, Ø. Nordgulen, P. T. Osmundsen, J. R. Skilbrei, M. A. Smethurst, A. Solli, T. Bugge, and C. Fichler (2002), Bridging the gap between the onshore and offshore geology in Nordland, northern Norway, *Nor. Geol. Tidsskr.*, *82*, 243–262.

Oxburgh, E. R., and E. M. Parmentier (1977), Compositional and density stratification in oceanic lithosphere—Causes and consequences, *J. Geol. Soc.*, *133*, 343–355, doi:10.1144/gsjgs.133.4.0343.

Panza, G. F. (1980), Evolution of the Earth's lithosphere, in *Mechanisms of Continental Drift and Plate Tectonics*, edited by P. A. Davies and S. K. Runcorn, pp. 75–87, Academic, New York.

Parmentier, E. M. (1987), Dynamic topography in rift zones: Implications for lithospheric heating, *Philos. Trans. R. Soc. London, Ser. A*, *321*(1557), 23–25, doi:10.1098/rsta.1987.0002.

Pelletier, J. D. (2004), Estimate of three-dimensional flexural-isostatic response to unloading: Rock uplift due to late Cenozoic glacial erosion in the western United States, *Geology*, *32*, 161–164, doi:10.1130/G20059.1.

Peulvast, J. P. (1978), Le bourrelet Scandinave et les Caledonides: Un essai de reconstitution des modalites de la morphogenese en Norvege, *Geogr. Phys. Quat.*, *32*, 295–320.

Phipps Morgan, J. P., W. J. Morgan, and E. Price (1995), Hotspot melting generates both hotspot volcanism and a hotspot swell?, *J. Geophys. Res.*, *100*, 8045–8062, doi:10.1029/94JB02889.

Pilidou, S., K. Priestley, E. Debayle, and O. Gudmundsson (2005), Rayleigh wave tomography in the North Atlantic: High resolution images of the Iceland, Azores and Eifel mantle plumes, *Lithos*, *79*, 453–474, doi:10.1016/j.lithos.2004.09.012.

Redfield, T. F., P. T. Osmundsen, and B. W. H. Hendriks (2005), The role of fault reactivation and growth in the uplift of western Fennoscandia, *J. Geol. Soc.*, *162*, 1013–1030, doi:10.1144/0016-764904-149.

Ricard, Y., C. Doglioni, and R. Sabadini (1991), Differential rotation between lithosphere and mantle: A consequence of lateral viscosity variations, *J. Geophys. Res.*, *96*, 8407–8415, doi:10.1029/91JB00204.

Riis, F. (1996), Quantification of Cenozoic vertical movements of Scandinavia by correlation of morphological surfaces with offshore data, *Global Planet. Change*, *12*, 331–357, doi:10.1016/0921-8181(95)00027-5.

Rittmann, A. (1943), Zur Thermodynamik der Orogenese, *Geol. Rundsch.*, *33*, 485–498, doi:10.1007/BF02198643.

Rohrman, M., and P. van der Beek (1996), Cenozoic postrift domal uplift of North Atlantic margins: An asthenospheric diapirism model, *Geology*, *24*, 901–904, doi:10.1130/0091-7613(1996)024<0901:CPDUON>2.3.CO;2.

Rohrman, M., P. van der Beek, P. Andriessen, and S. Cloetingh (1995), Meso-Cenozoic morphotectonic evolution of southern Norway: Neogene domal uplift inferred from apatite fission track thermochronology, *Tectonics*, *14*, 704–718, doi:10.1029/95TC00088.

Rosenbaum, G., G. S. Lister, and C. Duboz (2002), Relative motions of Africa, Iberia and Europe during Alpine orogeny, *Tectonophysics*, *359*, 117–129, doi:10.1016/S0040-1951(02)00442-0.

Sandoval, S., E. Kissling, J. Ansorge, and the SVEKALAPKO STWG (2004), High-resolution body wave tomography beneath the SVEKALAPKO array: II. Anomalous upper mantle structure beneath central Baltic Shield, *Geophys. J. Int.*, *157*, 200–214, doi:10.1111/j.1365-246X.2004.02131.x.

Scoppola, B., D. Boccaletti, M. Bevis, E. Carminati, and C. Doglioni (2006), The westward drift of the lithosphere: A rotational drag?, *Geol. Soc. Am. Bull.*, *118*, 199–209, doi:10.1130/B25734.1.

Scott, D. R., and D. J. Stevenson (1989), A self-consistent model of melting, magma migration and buoyancy-driven circulation beneath mid-ocean ridges, *J. Geophys. Res.*, *94*, 2973–2988, doi:10.1029/JB094iB03p02973.

Sengör, A. M. C. (2003), *The Large-Wavelength Deformations of the Lithosphere: Materials for a History of the Evolution of Thought From the Earliest Times to Plate Tectonics*, *Mem. Geol. Soc. Am.*, *196*, 347 pp.

Shaw, H. R. (1973), Mantle convection and volcanic periodicity in the Pacific: Evidence from Hawaii, *Geol. Soc. Am. Bull.*, *84*, 1505–1526, doi:10.1130/0016-7606(1973)84<1505:MCAVPI>2.0.CO;2.

Srivastava, S. P., and W. R. Roest (1999), Extent of oceanic crust in the Labrador Sea, *Mar. Pet. Geol.*, *16*, 65–84, doi:10.1016/S0264-8172(98)00041-5.

Stern, T. A., A. K. Baxter, and P. J. Barrett (2005), Isostatic rebound due to glacial erosion within the Transantarctic Mountains, *Geology*, *33*, 221–224, doi:10.1130/G21068.1.

Storey, M., R. A. Duncan, and C. Tegner (2007), Timing and duration of volcanism in the North Atlantic Igneous Province: Implications for geodynamics and links to the Iceland hotspot, *Chem. Geol.*, *241*, 264–281, doi:10.1016/j.chemgeo.2007.01.016.

Thybo, H. (2006), The heterogeneous upper mantle low velocity zone, *Tectonophysics*, *416*, 53–79, doi:10.1016/j.tecto.2005.11.021.

Torsvik, T. H., R. Van der Voo, J. G. Meert, J. Mosar, and H. J. Walderhaug (2001), Reconstructions of the continents around the North Atlantic at about the 60th parallel, *Earth Planet. Sci. Lett.*, *187*, 55–69, doi:10.1016/S0012-821X(01)00284-9.

Trampert, J., F. Deschamps, J. Resovsky, and D. Yuen (2004), Probabilistic tomography maps chemical heterogeneities throughout the lower mantle, *Science*, *306*, 853–856, doi:10.1126/science.1101996.

Turcotte, D. L., and G. Schubert (2002), *Geodynamics*, 456 pp., Cambridge Univ. Press, New York.

Vogt, P. R., P. T. Taylor, L. C. Kovacs, and G. L. Johnson (1979), Detailed aeromagnetic investigations of the Arctic Basin, *J. Geophys. Res.*, *84*, 1071–1089, doi:10.1029/JB084iB03p01071.

Ware, P. D., and J. P. Turner (2002), Sonic velocity analysis of the Tertiary denudation of the Irish Sea basin, in *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration*, edited by A. G. Doré et al., *Geol. Soc. Spec. Publ.*, *196*, 355–370, doi:10.1144/GSL.SP.2002.196.01.19.

Wessel, P., and W. H. F. Smith (1995), New version of Generic Mapping Tools released, *Eos Trans. AGU*, *76*(33), 329, doi:10.1029/95EO00198.

White, N. J., and B. Lovell (1997), Measuring the pulse of a plume with the sedimentary record, *Nature*, *387*, 888–891, doi:10.1038/43151.

Ziegler, P. A. (1987), Late Cretaceous and Cenozoic intra-plate compressional deformations in the Alpine foreland—A geodynamic model, *Tectonophysics*, *137*, 389–420, doi:10.1016/0040-1951(87)90330-1.

Ziegler, P. A. (1988), *Evolution of the Arctic—North Atlantic and the Western Tethys*, *AAPG Mem.*, *43*, 198 pp.

Ziegler, P. A., and P. Dèzes (2007), Cenozoic uplift of Variscan Massifs in the Alpine foreland: Timing and controlling mechanisms, *Global Planet. Change*, *58*, 237–269, doi:10.1016/j.gloplacha.2006.12.004.

Ziegler, P. A., J. D. van Wees, and S. Cloetingh (1998), Mechanical controls on collision-related compressional intraplate deformation, *Tectonophysics*, *300*, 103–129, doi:10.1016/S0040-1951(98)00236-4.

E. Carminati and C. Doglioni, Dipartimento di Scienze della Terra, Università La Sapienza, Piazzale Aldo Moro 5, I-00185 Roma, Italy.

M. Cuffaro, MOX, Dipartimento di Matematica, Politecnico di Milano, I-20133 Milano, Italy.