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Elevated, passive continental margins: Not rift shoulders, but expressions of episodic, post-rift burial and exhumation

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A R T I C L E I N F O

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A B S T R A C T

Many studies of elevated, passive continental margins (EPCMs) assume that their characteristic, large-scale morphology with high-level plateaux and deeply incised valleys has persisted since rifting and crustal separation, and that the absence of post-rift sediments is evidence of non-deposition. The high mountains in West Greenland, however, expose evidence of km-scale, post-rift subsidence, and recent studies showed that typical EPCM morphology with elevated plateaux formed c. 50 Myr after breakup through a process of uplift and dissection of a regional, post-rift erosion surface. Since the West Greenland margin shares all the morphological characteristics of EPCMs, the results from West Greenland lead us to question the common assumption that EPCMs have remained high since the onset of continental separation. We present published evidence of post-rift burial followed by uplift and exhumation from a number of EPCMs and their adjacent basins to support the notion that EPCMs are not permanent highs and that their morphology is unrelated to rifting and continental breakup. Geodynamic models that explain EPCMs as permanent highs since the time of rifting require either no lithospheric mantle extension below extending crust or effective elastic thicknesses >100 km. Such models are, however, not consistent with the subsidence history inferred from actual rifts and their margins. Geodynamic models using low elastic thicknesses and a much more uniform distribution of strain within the lithosphere are more consistent with observations of early post-rift behaviour, but some additional process is needed to uplift the margins later. We suggest that EPCMs represent anticlinal, lithospheric folds formed under compression where an abrupt change in crustal or lithospheric thickness occurs between cratons and rift basins. We propose that EPCMs are expressions of episodes of post-rift burial followed by compression-induced uplift and exhumation; one episode of uplift results in erosion of the region to produce a low-relief surface near the level of the adjacent, opening ocean, and a second (or more) episode(s) raises the plateau to its present elevation, after which the plateau is dissected by fluvial and possibly glacial erosion.

1. Introduction

Many passive continental margins around the world are characterised by elevated plateaux (i.e. large-scale, low-relief, high-level landscapes) at 1 to 2 km or more above sea level (a.s.l.) cut by deeply incised valleys and commonly separated from an adjacent coastal plain by one or more escarpments (Fig. 1). Mesozoic–Cenozoic rift systems parallel to the coast are commonly present offshore with a transition from continental to oceanic crust further offshore. Two observations can easily be made from Fig. 1. First, that landscapes that characterise elevated, passive continental margins (EPCMs) are similar despite different geological settings and despite the time span since breakup (e.g. along the Atlantic from south to north). Second, that other passive continental margins are low-lying, and thus the present elevation of a passive margin cannot be an inevitable consequence of the processes of rifting and breakup. These observations lead us to suggest that the EPCMs were shaped by similar processes that may or may not be activated along them, a long time after breakup.

Studies of the development of EPCMs frequently assume that they have remained high since rifting or continental breakup (Fig. 2). Such studies comprise both geomorphological and geodynamic investigations, and they assume either that the elevated plateau represents a pre- or syn-breakup surface that underwent very little erosion (e.g. Ollier and Pain, 1997), or that it remained high despite significant post-rift denudation (e.g. some of Weissel and Karner's, 1989 models; Gilchrist and Summerfield, 1990; Watts, 2001). Denudational histories following these patterns form the basis for many thermochronological studies (e.g. Gallagher et al., 1998; Brown et al., 2002; Persano et al., 2006; Swift et al., 2008), as well as numerical studies that attempt to show how erosional processes act on rift margins to create the present-day topography (e.g. van der Beek et al., 2002). The assumption that margins remain high long after rifting is accompanied by an interpretation that the absence of a post-rift sedimentary section at many margins is evidence that no section was ever deposited. Little attention is paid to

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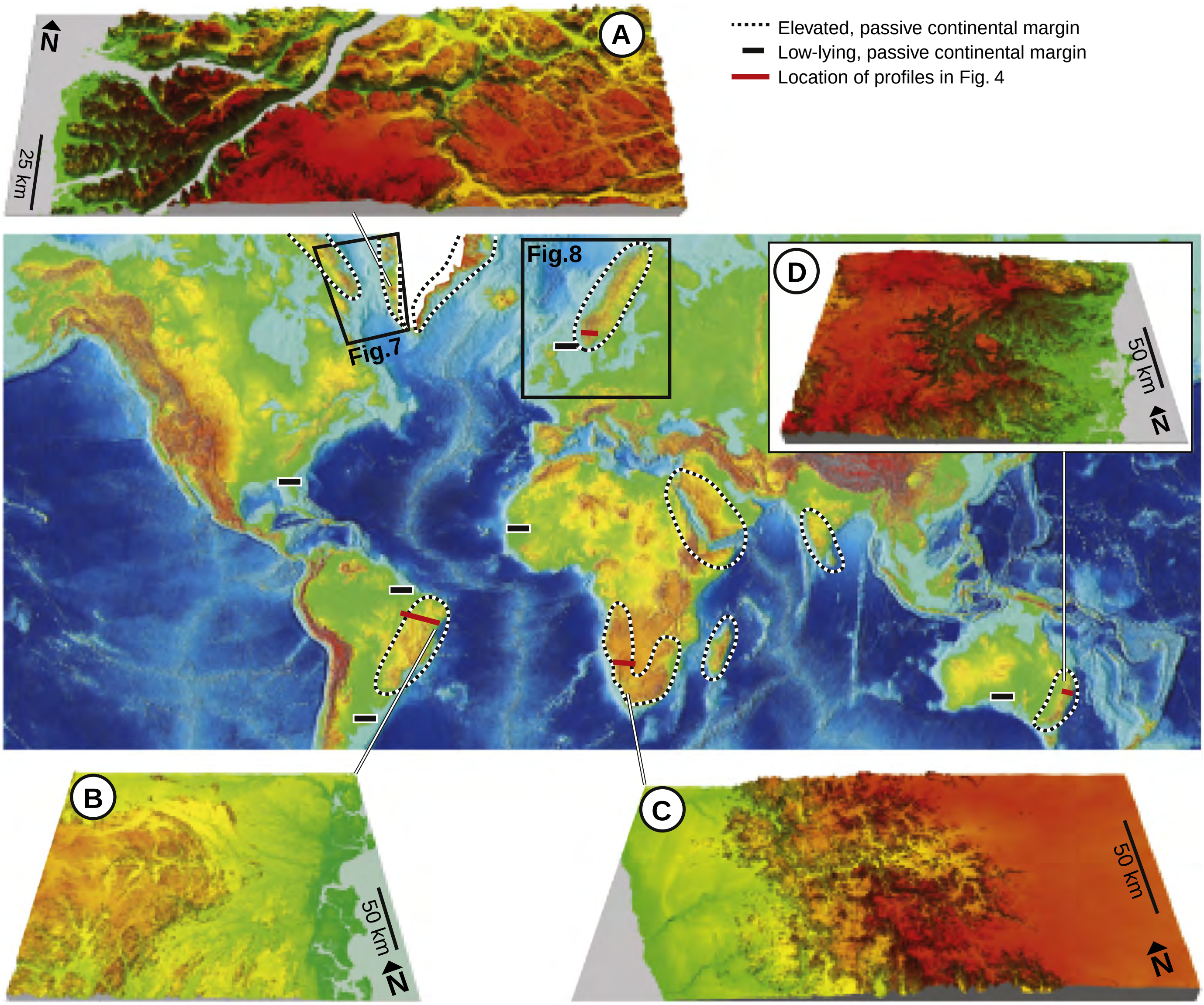

Fig. 1. Topography of the Earth between 72°N and 42°S showing elevated and low-lying passive continental margins. Only Atlantic type margins that can easily be connected to the corresponding spreading centre are indicated and only elevated margins that reach 2 km a.s.l. in more than a single summit (c.f. Bradley, 2008). A–D. Terrain models of EPCM morphology characterised by elevated plateaux: (A) West Greenland (centre coordinates 52°W, 66° 30′N). (B) NE Brazil (centre coordinates 40°W, 13°S). The low-lying areas along the coast reflect the exhumed Early Cretaceous rift where the northern part is the Recôncavo Rift. (C) SW Africa (centre coordinates 18°E, 30°S). (D) SE Australia (centre coordinates 152°E, 31°S). Colour scale: Green (<300 m a.s.l.), yellow (300–500 m), light orange (500–900 m), dark orange (900–1200 m), red (1200–1500 m), brown to black (1500–2000 m): The locations of maps in Figs. 7 and 8 are indicated and the cross-sections shown in Fig. 4 are shown as red lines. Data in A to D from Jarvis et al., 2008. Global relief model from http://www.ngdc.noaa.gov/mgg/global/global.html. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the fact that EPCMs are hundreds of kilometres wide whereas rift shoulders only should be tens of kilometres wide.

In contrast to these ideas, other geodynamic models predict post-rift subsidence of the rift margin. The margin of a rift will normally be uplifted slightly during and shortly after rifting due to flexural rebound and accommodation of fault-block rotation in the rift (e.g. Roberts and Yielding, 1991; Watts, 2001). However, the uplifted margin will normally then subside due to loading of sediments in the nearby rift (Watts et al., 1982; Braun and Beaumont, 1989), erosion of the uplifted area (Roberts and Yielding, 1991) or because stretch in the lithospheric mantle under the rift margin was slightly greater than stretch in the crust (White and McKenzie, 1988). A sedimentary wedge transgresses the subsiding margin, defining the so-called “steer's head” geometry.

Our recent studies of the West Greenland margin showed that the present-day, EPCM morphology, with its characteristic high-level plateaux, is not a remnant of the rifting process but is much younger (Japsen et al., 2005, 2006, 2009; Bonow et al., 2006a, b). Since the West Greenland margin shares all the characteristics of EPCMs as described above, our results from West Greenland lead us to question the assumption that EPCMs have remained high since the onset of continental separation.

The purpose of this paper is to gather observations that document the nature and timing of vertical movements along EPCMs. We present published geological, geomorphological and thermochronological evidence from a number of margins to show that episodic post-rift movements involving km-scale exhumation are common features of EPCMs (as well as other regions—see Holford et al., 2009). We use these observations to assess which of the many explanations for the presence of these prominent, topographic features put forward in the literature are possible and which are impossible. We find that many of the special geodynamic conditions used to account for the present

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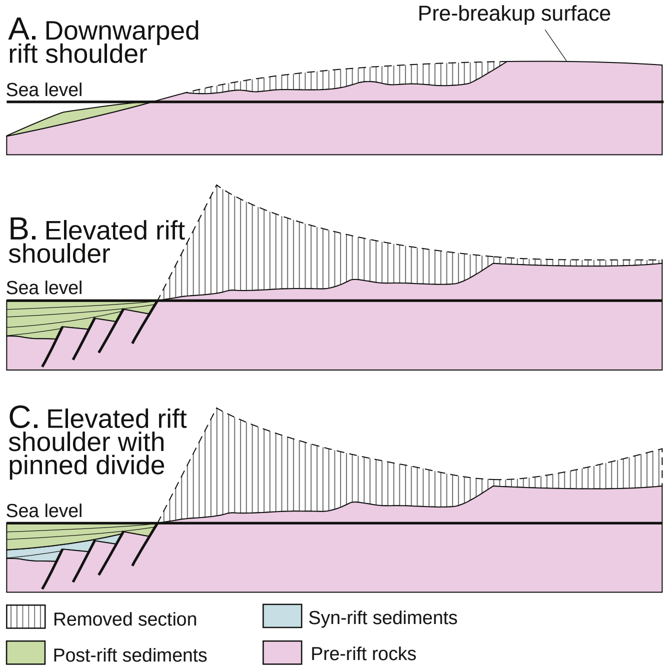

Fig. 2. Geomorphological models that are commonly used in discussions of the development of EPCMs: (A) Down-warped rift shoulder (Ollier and Pain, 1997). (B) Scarp retreat from an elevated rift shoulder (e.g. Gilchrist and Summerfield, 1990). (C) Down-wearing of an elevated rift shoulder with pinned divide (e.g. Brown et al., 2002; Persano et al., 2006). These models are based on the assumption that the margin remained elevated after rifting/breakup and that the plateau surface represents either a pre- or syn-breakup surface or a surface that remained high despite significant erosion. Modified from Ollier and Pain (1997); Gallagher et al. (1998). These models should be compared with that shown in Fig. 6 which explains characteristic features of EPCMs such as those shown in Figs. 1 and 3. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

elevation of rifted margins are based on the assumption that such margins have remained elevated since break-up (or earlier), and are, in fact, incompatible with observations. We conclude that the absence of a post-rift sedimentary section from some margins is best explained in terms of its removal by erosion during post-rift uplift. We find that the uplift events are unrelated to the formation of the margin, but they appear to be related to the presence of the margin and to the state of stress along the margin.

2. Observational constraints on the post-rift development of EPCMs

The best explanation of the development of EPCMs is one that takes into account the features that they all have in common. EPCMs are commonly asymmetric with a slope towards the ocean that is substantially steeper than the slope towards the hinterland. In many cases the higher ground of EPCMs consists of an extensive plateau, or plateaux at different heights, separated by one or more escarpments. The plateaux are typically more than 100 km wide, much larger in some cases, and extend hundreds of kilometres along the margin, cutting across bedrock of different ages and resistances. The plateau surfaces are in turn cut by deeply incised valleys (Fig. 1A–D) (c.f. Ahlmann, 1941; King, 1967; Lidmar-Bergström, 1996; Bonow et al., 2006a, b; Japsen et al., 2009).

The key to understanding the formation of regional, low-relief erosion surfaces is fluvial erosion, because the level to which fluvial erosion cuts (the base level) controls the erosion. Unless a local base level is present, represented for example by a resistant rock surface or an internal drainage basin, the most likely base level is sea level, particularly for locations along continental margins (e.g. during the post-rift development of passive margins). The process of valley widening by river erosion eventually results in a large-scale, low-relief surface; a peneplain. Valley incision below such a surface is evidence of lowering of base level (uplift of the landmass or fall in sea level), with subsequent formation of new valley floors grading to sea level. The height difference between the valley floor and the overlying surface therefore indicates the amount of uplift or fall in base level.

Although pre-rift rocks dominate many EPCMs, there is evidence from a number of EPCMs—which we review in the following—to show that subsidence and burial continued after rifting, and that the post-rift sediments were eroded before or during the uplift of the margin to its present-day altitude. This evidence is, however, frequently overlooked, and the absence of post-rift rocks at high elevation at EPCMs is taken as evidence of non-deposition. The unsupported assumption is then adopted that these margins remained high for many tens of millions of years (e.g. Gilchrist and Summerfield, 1990; Ollier and Pain, 1997; Persano et al., 2006; Swift et al., 2008).

Important insights into the development of EPCMs are available from the rift basins adjacent to the margin where erosional unconformities commonly truncate a tilted, post-rift sequence (Fig. 3) (e.g. Cobbold et al., 2001; Stevenson and McMillan, 2004; Stoker et al., 2005). These truncation events are consistent with uplift of the continents—and of the adjacent shelf—at a time significantly after continental breakup. Such unconformities often represent removal of significant sections of sediment as demonstrated along many margins by e.g. Walford and White (2005); Japsen et al. (2007); Turner et al. (2008); Green and Duddy (2010). These investigations use palaeo-thermal data (vitrinite reflectance, VR, and apatite fission-track, AFT), palaeo-burial data (porosity and sonic velocity) and other methods (cf. Corcoran and Doré, 2005). However, many EPCM studies focus solely on the elevated part of the margin and ignore the information available from the adjacent basins (cf. Fig. 2).

AFT data from many EPCMs have been interpreted within a framework in which the surface inland of the escarpment has remained elevated since the time of rifting or breakup, with post-rift denudation focussed seaward of the escarpment (e.g. Gallagher et al., 1998; Brown et al., 2002; Persano et al., 2006). But in many cases (e.g. Harman et al., 1998; Johnson and Gallagher, 2000; Kohn et al., 2002), such studies actually reveal cooling which post-dates rifting and breakup by many Myr, and also extends much further inland than suggested by the simple models shown in Fig. 2. These observations therefore show that the models in Fig. 2 do not describe the behaviour of real margins.

Japsen et al. (2006) studied the evolution of the West Greenland EPCM using a multi-disciplinary approach, combining the cooling history from apatite fission-track analysis (AFTA) data (Japsen et al., 2005) with the denudation history from landscape analysis (Bonow et al., 2006a, b) and the stratigraphic record (onshore and offshore). Without the AFTA data, the landform analysis would have yielded only a relative event chronology. Without the landform analysis, it would not have been possible to conclude whether the cooling events shown by the AFTA data were due to exhumation or due to changes in the palaeo-temperature gradient.

2.1. West Greenland

Rifting occurred in West Greenland's Nuussuaq Basin (70°N) during mid-Cretaceous and earliest Paleocene times (Chalmers et al., 1999). After the second episode, a km-thick, Palaeogene volcanic sequence with interbedded, marine deposits accumulated during post-rift subsidence (Pedersen et al., 2002). Today, these rocks are exposed on the sides of mountains that reach above 2 km a.s.l., and the marine deposits are at elevations up to 1.2 km a.s.l. (Piasecki et al., 1992). This clearly shows that the present-day elevated topography of the West Greenland margin is not a remnant of the rifting process but developed later. AFTA data (Japsen et al., 2005, 2006) show that post-rift subsidence continued for about 25 million years after cessation of

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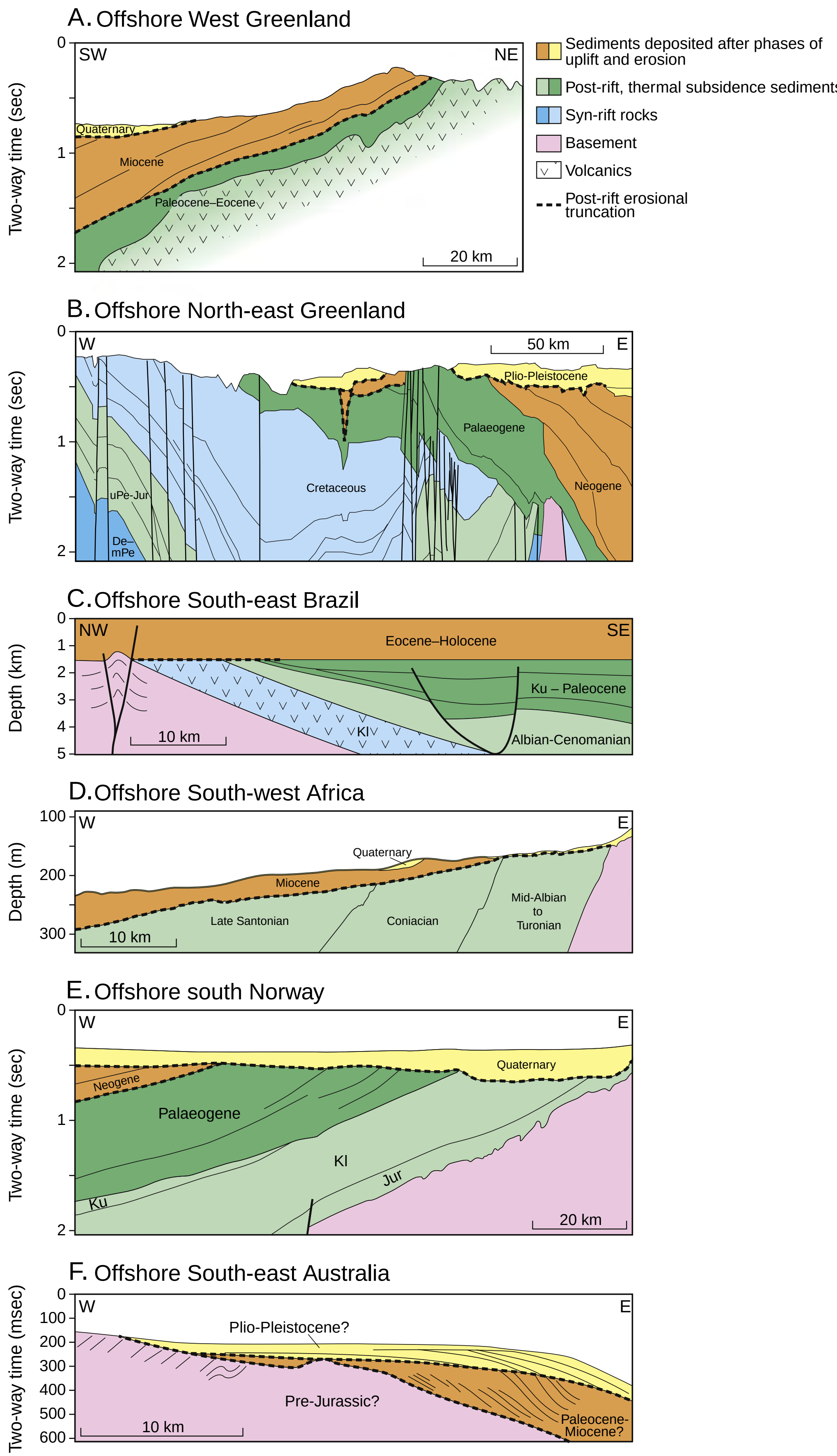
A. Offshore West Greenland
Two-way time (sec)
SW
NE
Quaternary
Miocene
Paleocene–Eocene
20 km
Sediments deposited after phases of uplift and erosion
Post-rift, thermal subsidence sediments
Syn-rift rocks
Basement
Volcanics
Post-rift erosional truncation
B. Offshore North-east Greenland
W
E
50 km
Plio-Pleistocene
Palaeogene
Cretaceous
Neogene
uPe-Jur
De–mPe
C. Offshore South-east Brazil
Depth (km)
NW
SE
Eocene–Holocene
Ku – Paleocene
Kl
Albian-Cenomanian
10 km
D. Offshore South-west Africa
Depth (m)
W
E
Quaternary
Miocene
Late Santonian
Coniacian
Mid-Albian to Turonian
10 km
E. Offshore south Norway
Two-way time (sec)
W
E
Quaternary
Neogene
Palaeogene
Kl
Jur
Ku
20 km
F. Offshore South-east Australia
Two-way time (msec)
W
E
Plio-Pleistocene?
Pre-Jurassic?
Paleocene-Miocene?
10 km

rifting until the end of the Eocene, the time at which sea-floor spreading ceased between Canada and Greenland (Chalmers and Pulvertaft, 2001).

Maximum burial of the post-rift sequence was followed by uplift and erosion in which up to 1 km of the post-rift sequence was removed, resulting in formation of a regional, low-relief erosion surface (a peneplain) that was graded to sea level (Japsen et al., 2005, 2006; Bonow et al., 2006a, b). This erosion surface cuts across middle-Eocene basalts, which it thus post-dates. Once formed, the surface was offset by reactivated faults and uplifted to present-day altitudes of up to 2 km (Fig. 1A) during two episodes in the Neogene.

AFTA results from samples over a wide area onshore, both within the Nuussuaq Basin and its adjacent basement terrain, document a regional cooling episode that began at the Eocene–Oligocene transition. This area is similar in extent to the regional planation surface and we therefore consider it likely that both the formation of the post-middle Eocene planation surface and the end-Eocene cooling event are evidence of the same episode of denudation. Broad and uplifted palaeo-valleys incised below the surface demonstrate that its uplift took place in two phases, and AFTA data date these two phases of uplift to the last 10 Ma. The peneplain was thus formed during the Oligo-Miocene. Offshore, a shelf-wide, low-angle, Oligocene erosional unconformity and a composite, angular, late Neogene unconformity formed at the same time as the uplift onshore (Fig. 3A) (Japsen et al., 2006, 2010). To summarise, the West Greenland margin developed through 25 Myr of initial post-rift subsidence followed by two main phases of uplift; one (starting at 35 Ma) that resulted in formation of the peneplain and a correlative unconformity offshore, and one (starting a 10 Ma) that lifted the peneplain to its present elevation. Redfield (2010) questioned numerous aspects of this interpretation, but offered no alternative view of the timing of events in West Greenland to those summarised here. Green et al. (2011) highlighted that Redfield (2010) explicitly accepted the basic sequence of events involved in our synthesis, including post-rift subsidence and burial, post-39 Ma uplift and erosion, and Neogene uplift and dissection of the resulting planation surface.

2.2. East Greenland

Breakup of the North Atlantic began at the Palaeocene–Eocene transition (c. 55 Ma; e.g. Tegner et al., 2008) and the geological record at Kangerlussuaq and Blosseville Kyst in East Greenland (68–70°N) documents that the area underwent short-lived uplift immediately prior to breakup (Dam et al., 1998) followed by km-scale subsidence during the eruption of basalts with no evidence for crustal upwarping at this time (Brooks, 1985; Larsen and Tegner, 2006). There are marine incursions in some of the earliest basalts (e.g. Nielsen et al., 1981) and in the uppermost of the plateau lavas, the early Eocene Igertivâ Formation (c. 48 Ma; Soper and Costa, 1976; Tegner et al., 1998). These Palaeogene plateau lavas make up Gunbjørn Fjeld which is the highest summit in Greenland (3.7 km a.s.l.), and the present topography is thus the result of post-rift uplift.

Ahlmann (1941) described landscapes in central East Greenland (72–76°N) as alpine along the outer coast and as remnants of a high plateau that become more coherent inland. He concluded that "*the plateau and summit areas are the remains of what has once been a more or less uniform high plateau rising towards the west*". Ahlmann (1941) developed the concept of an "*initial topography*" from which the present-day landscape has developed by dissection. He argued that "*the ultimate formation of the initial topography took place in post-basalt time*" since the high-level plateau cuts across both Palaeogene basalts and Precambrian basement rocks. This fundamental observation has later been ignored; e.g. by Swift et al. (2008), who concluded that the main features of the present landscape formed prior to breakup. Ahlmann (1941) considered that the landscape evidence favoured a "*Late Tertiary*" timing for the formation of the present relief, implying that the topography significantly post-dates North Atlantic breakup when most of the basalts were extruded.

We suggest that the post-basalt, plateau surface formed by fluvial erosion towards the base level of the adjacent sea during the opening of the North Atlantic. The '*initial topography*' thus formed in two steps: first by erosion to form a peneplain near sea level and second by uplift to form a high plateau. The present topography has developed by incision of this high plateau, initially by rivers and later by glaciers below the uplifted peneplain. Offshore seismic profiles show erosional truncation of the Palaeogene and older sequences (Fig. 3B) (Hamann et al., 2005), an observation that supports the hypothesis that East Greenland was affected by Neogene uplift. Marine sediments that were deposited at the Paleocene–Eocene transition are now at high elevations onshore East Greenland (e.g. 400 m a.s.l., c. 74°N; Nøhr-Hansen, pers.comm., 2011), and this is yet another indication of significant post-rift subsidence followed by uplift of this margin.

AFT studies In East Greenland show that Cenozoic maximum palaeotemperatures occurred in the mid- and late Cenozoic (Thomson et al., 1999; Johnson and Gallagher, 2000). In particular, results from Clavering Ø (74°N) show that Carboniferous sandstone overlain by Palaeogene basalts preserved on the summits there were heated to peak temperatures of c. 70°C prior to cooling in the late Cenozoic. As there is no Palaeogene intrusive activity reported from this area, a major cause of the post-basalt heating must be due to burial below rocks that have been removed, presumably by uplift and erosion in the Neogene (Japsen et al., 2010).

Medvedev et al. (2008) showed that 1.1 km of uplift in the Scoresby Sund area (70°N) could be accounted for by flexed isostatic uplift caused by the incision of the fjords below the summit surface, combined with loading from deposition of sediments offshore. These authors also suggested that the elevation of Mesozoic marine sediments at 1.2 km a.s.l. confirmed this conclusion. This cannot, however, explain the entire uplift in this region of East Greenland because it does not account for the initial uplift and erosion necessary to form the low-relief, high-level plateau. In addition, Medvedev et al. (2008) did not consider the possibility that the Mesozoic marine sediments had been buried below a kilometre-thick cover (e.g. Kelly et al., 1998), and thus that the actual, vertical movement of these deposits since maximum burial is likely to have been several kilometres rather than just one. Furthermore, the amount of uplift due to flexed isostacy (1.1 km) is only half the elevation of the plateau in many areas (2 km and more according to Ahlmann, 1941).

2.3. Brazil

The Brazilian EPCM is some 1200 km wide (Fig. 4A). There is abundant evidence that the present topography developed well after

Fig. 3. Post-rift truncation of sedimentary sequences offshore of EPCMs indicating post-rift uplift of these margins. (A) Offshore West Greenland (70° 30′N); Oligocene and late Neogene truncations along low-angle and high-angle unconformities, respectively (redrawn from Chalmers, 2000; Japsen et al., 2006, 2010 and unpublished AFT data). (B) Offshore NE Greenland (78°N); early and late Neogene truncations (rifting continued into the Palaeogene in the central part of the profile) (redrawn from Hamann et al., 2005). (C) Offshore SE Brazil (26° 30′S); mid-Eocene truncation (redrawn from Cobbold et al., 2001). (D) Offshore SW Africa (30°S); post-Cretaceous truncation (redrawn from Stevenson and McMillan, 2004). Post-rift exhumation along SW Africa is documented by Walford and White (2005). (E) Offshore South Norway (59°N); early and late Neogene truncations along low-angle and high-angle unconformities, respectively (the Jurassic sequence may include a thin syn-rift section). Data courtesy of Statoil, seismic interpretation by L.N. Jensen, Statoil, and C. Andersen, GEUS. The early Neogene exhumation is documented by Japsen et al. (2010). (F) Offshore SE Australia (34°S); Truncations are above pre-rift rocks whereas syn-rift sediments are missing (base-Cenozoic and Pliocene dating according to Davies, 1975). Note that low-angle unconformities may be especially difficult to identify as erosional truncations and consequently that all truncations may not have been identified on these profiles. De–mPe: Devonian–Middle Permian, Jur: Jurassic, Kl: Lower Cretaceous, Ku: Upper Cretaceous, uPe–Jur: Upper Permian–Jurassic. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

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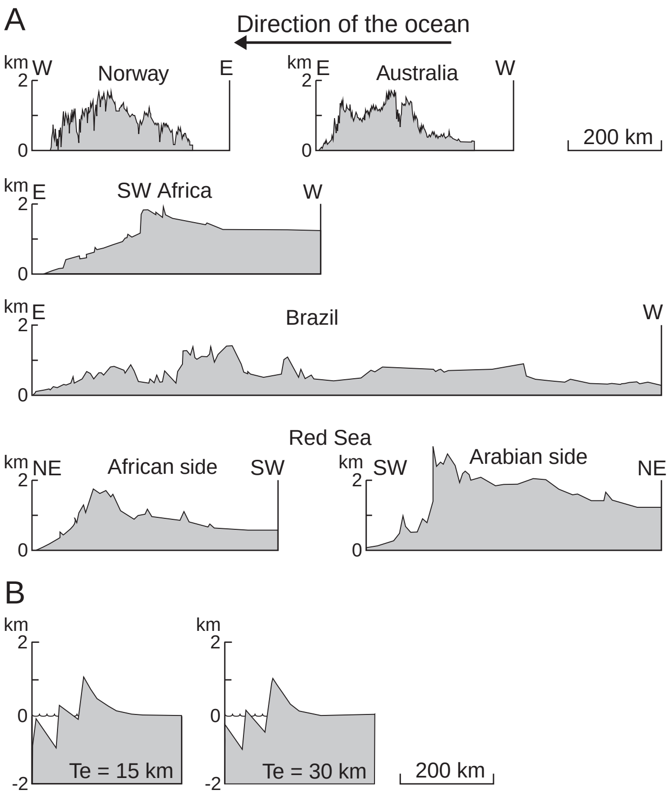

Fig. 4. EPCM cross-sections, present day and just after rifting. (A) Cross-sections of five different EPCMs all at the same horizontal and vertical scale. The sections are oriented so that the nearest ocean is to the left and the continental hinterland to the right. Locations shown in Fig. 1. (B) Flexed rift margins with uniform stretching in the crust and lithospheric mantle and with elastic thicknesses of 15 and 30 km immediately after rifting, shown at the same scale as the cross-sections in A. The amount of uplift is somewhat smaller and the width of the uplifted area is substantially narrower than the real examples (that all represent post-rift situations). As the margins become re-attached to the rift once extension (rifting) ceases, cooling of the extended lithospheric mantle under the adjacent rift will cause the margins to subside along with the rift.

rifting ceased and that erosion removed some of the post-rift sediments. For example, a post-rift section of less than 400 m occurs in the Early Cretaceous Recôncavo Rift, onshore NE Brazil (12–13°S), but VR data from a deep borehole show that an additional post-rift section of c. 1.75 km must have been present prior to exhumation (Magnavita et al., 1994). Further evidence comes from the Araripe Basin (7°N), c. 300 km from the Atlantic margin, which formed during the widespread, Early Cretaceous rifting that eventually led to breakup between South America and Africa at the Aptian–Albian boundary (Torsvik et al., 2009). Marine incursions (e.g. dinocysts and radiolarians) within Albian limestones occur at elevations up to 800 m a.s.l. on the present-day Araripe Plateau (Magnavita et al., 1994; Morais Neto et al., 2009). Such marine incursions occur in sedimentary deposits of mid-Cretaceous age across much of the interior of Brazil (Arai, 2000; Arai et al., 2007), and Arai (2000) argued that this mid-Cretaceous transgression connected together the Pacific and Atlantic oceans. The present elevation of these marine incursions in post-rift sediments across the interior of Brazil thus testifies to post-rift subsidence followed by significant uplift.

Peulvast et al. (2008) studied the belt of plateaux in NE Brazil (including the Araripe Plateau) with relatively constant elevations between 1000 and 1200 m a.s.l. and argued that they are remnants of one, continuous surface, similar to the surface shown in Fig. 1B. They dated the surface by the youngest rocks (Cenomanian continental sandstone in the Araripe Basin) exposed below it and argued that the surface formed as a low-lying, Late Cretaceous rift flank, and that its present elevation is due to post-Cenomanian uplift that thus significantly post-dates breakup in the South Atlantic.

However, Peulvast et al. (2008) assumed that subsidence ended with deposition of the Cenomanian sandstone and did not consider the possibility that additional section was deposited above this surface and subsequently removed. AFT data from the region indicate that rocks now at the surface have cooled by at least 50°C since 80 Ma (Campanian) (Harman et al., 1998), implying km-scale deposition above the Cenomanian sandstone and a corresponding amount of exhumation since the Late Cretaceous. This constraint on the timing of exhumation and the presence of the Palaeogene Serra do Martins Formation covering many of the plateau remnants in the region (Morais Neto et al., 2008), indicates that the plateau surface formed during the Palaeogene as suggested by e.g. King (1967) and Bigarella (1975). This hypothesis is also consistent with the observation that the syn-rift Parana basalts and their overlying post-rift sedimentary sequence in the Santos Basin, offshore SE Brazil, are cut off by an Eocene unconformity (Fig. 3C) indicating a major phase of erosion, most easily explained by a phase of uplift and exhumation of the present-day Serra do Mar which achieved their final shape after Neogene uplift according to Cobbold et al. (2001).

2.4. Southern Africa

There are EPCMs on both the east and west sides of southern Africa where there are mountain ranges near the coast with steep slopes on the ocean side. These slopes are eroding as escarpments in both EPCMs. Southern Africa is, however, atypical because the hinterland between the EPCMs is not close to sea level (compare cross-sections in Fig. 4A), implying that additional forces such as upwelling deep in the mantle (e.g. Al-Hajri et al., 2009) may be present.

Rocks at outcrop as far as 300 km inland on major low-relief surfaces in the highlands of SW Africa have AFT ages which significantly post-date Early Cretaceous breakup (e.g. Gallagher et al., 1998). As most of the analysed samples are stratigraphically much older than the AFT ages, these results demonstrate cooling from paleotemperatures around 100°C or above (corresponding to removal of a km-thick cover) across the region in post-Early Cretaceous times. This is not possible if the present surface is pre-breakup, unless it was reburied by several kilometres of sediment after breakup. More likely, the major planation surfaces that now define the elevated topography across southern Africa formed by erosion that included removal of substantial thicknesses of sedimentary cover, long after rifting and breakup (Fig. 1C). King (1967) used correlation with sediments in coastal areas or offshore to suggest that the major surface above the escarpment, the African surface, formed by erosion to near sea level during the Late Cretaceous to early Miocene. He identified two incised valley generations in the African surface that give evidence for uplift during the Neogene.

The presence inland of Cenozoic marine strata (Pickford, 1998), such as Eocene marine deposits 400 m a.s.l. at Need's Camp, South Africa (Partridge and Maud, 1987), documents that much of the present elevation of Southern Africa is due to post-rift uplift. The presence of tilted and truncated Upper Cretaceous sediments off the coast of SW Africa further supports this hypothesis (Fig. 3D) (Stevenson and McMillan, 2004). The morphology of Southern Africa thus clearly formed post-breakup (see also Burke and Gunnell, 2008, and Al-Hajri et al., 2009). This observation is in strong contrast to the frequent assumption that its elevation is an inherited feature from the time of breakup or even earlier; e.g. Gilchrist and Summerfield (1990) who based models of the margin of southern Africa on the speculation that the elevated topography has persisted since rifting of the eastern margin at c. 150 Ma.

2.5. SE Australia

Samples of basement and outliers of Permian sandstone from the highest elevations on the dissected plateau of the eastern highlands of Australia (34°S; Fig. 1D), some 150 km inland from an Early Cretaceous rift, yield AFT ages of 100 Ma or less (O'Sullivan et al., 1995; Persano

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et al., 2006). The basement there was thus exhumed to near-surface levels in Permian times and then reheated to palaeotemperatures around 100°C as a result of Early Permian to mid-Cretaceous burial (Green et al., 2007). Kilometre-thick Permian to Early Cretaceous, sedimentary deposits must thus have covered the basement rocks and Early Permian outliers (that now form the elevated plateau surface) at the initiation of cooling that was somehow related to Late Cretaceous breakup along the margin (Gaina et al., 1998). AFT ages less than 100 Ma extend well inland of the coastal escarpment (Kohn et al., 2002), far from the pattern suggested by the simple models illustrated in Fig. 2 which predict post-rift erosion focussed predominantly seaward of the escarpment.

Further south, in the East Victoria Highlands, the presence of outliers of Eocene mudstones preserved below Oligocene basalts indicates that the upland plateau formed after breakup (Holdgate et al., 2008). Holdgate et al. (2008) inferred that these outliers, and similar Eocene sediment now buried offshore, accumulated in a coherent valley system that was incised into a widespread, low-relief landscape near sea level, and that both the sediments onshore and the low-relief plains reached their present-day elevation of almost 2 km a.s.l. in post-Eocene times.

The stratigraphy offshore eastern Australia is poorly known, but Davies (1975) interpreted two pronounced unconformities to represent phases of post-rift, erosional truncations (Fig. 3F). This interpretation gives further support to the notion of significant post-rift uplift of this margin in contrast with the assumption of many researchers that it is a remnant of a rift shoulder that has only been slightly modified since breakup (e.g. Ollier and Pain, 1997; Persano et al., 2006).

2.6. The Red Sea

The Red Sea, where sea-floor spreading began at 5 Ma, has the youngest passive margins on the Earth today (Bradley, 2008). Onset of rifting and flooding by marine waters occurred in the late Oligocene in the southern Red Sea (Crossley et al., 1992), but continued rifting established marine conditions throughout the system by the early Miocene. Episodic isolation in the mid-Miocene led to evaporite deposition in some basins, but marine conditions were re-established in the Pliocene with carbonate build-ups in shallow-water areas. Crossley et al. (1992) interpreted clastic sediment textures to indicate that marginal escarpments, which first developed during the onset of rifting, were strongly uplifted in the Pliocene-to-Recent period. Daradich et al. (2003) also presented evidence that the uplift along the Red Sea—and in particular the tilting of the 1000 km wide Arabian plate—is a young feature (since 12 Ma).

The present topography across the Red Sea (Fig. 4A) is asymmetrical and appears to consist of two components. The high mountains along both margins may contain a remnant of rift shoulders, but the geological evidence, described above, suggests that at least some of the uplift of these mountains was post-rift. The second component is the hinterland of the EPCMs, particularly on the Arabian side, that is at an altitude of around a kilometre and there may be an element of deep mantle upwelling here, similar to what has been described by Al-Hajri et al. (2009). The mantle upwelling may well be related to the presence of the Afar plume/hotspot in the southern Red Sea.

2.7. Scandinavia

The Norwegian mountains along the NE Atlantic margin form a classic EPCM with all of the typical features: a steeper gradient towards the ocean than towards the hinterland, elevated plateaux and post-rift unconformities offshore that truncate sedimentary sequences (Fig. 3E) (e.g. Jensen and Schmidt, 1992; Martinsen et al. 1999; Lidmar-Bergström et al., 2000, 2007; Løseth and Henriksen, 2005; Japsen et al., 2007; Gabrielsen et al. 2010a). The absence of post-Devonian rocks in these mountains means, however, that there are no direct, geological indicators that can be used to evaluate the post-rift development of this EPCM, in contrast to other margins discussed here. There has, consequently, been considerable debate about for how long high mountains in Norway have existed (e.g. Lidmar-Bergström et al., 2000; Lidmar-Bergström and Bonow, 2009; Chalmers et al., 2010; Nielsen et al., 2009a,b, 2010a,b; Gabrielsen et al., 2010a,b).

A number of indirect indicators, however, illustrate the development of the Norwegian margin. Osmundsen et al. (2010), for example, documented major, normal faulting that affected the continental crust of western Scandinavia after rifting and breakup, and prior to late Cenozoic glaciations, and argued that these faults were important during the post-rift uplift of Scandinavia. Osmundsen et al. (2010) considered the Scandinavian topography to be a rejuvenated rift flank, but did not recognise that the AFT ages younger than 100 Ma that they reported from near Jurassic rift basins along the coast of north Norway (69°N), mean that these basins and their margins must have been exhumed from below a kilometre-thick cover.

The presence of hemipelagic, deep-marine deposits of Eocene age near basement exposed in southern Scandinavia (e.g. Heilmann-Clausen et al., 1985) indicates that much of the present onshore areas were covered by a deep sea after breakup in the North Atlantic. Reworked Eocene dinocysts and clasts of Eocene muds interbedded with Miocene deltaic sediments in Denmark indicate the presence of marine Eocene deposits in the Scandinavian hinterland (Rasmussen et al., 2010). Knox et al. (2010) presented a map of early Eocene (latest Ypresian) palaeogeography around southern Norway where only the highest, present-day mountains emerge above the sea. Knox et al. (2010) used microfossil evidence to argue that there was a marine connection between western Siberia via a Baltic Seaway across southern Finland, southern Sweden, southern Norway and the North Sea.

While not being conclusive, the geological evidence is thus consistent with formation of the mountains in southern Norway after Early Eocene breakup of the North Atlantic and after subsidence and burial during later Eocene times. Løseth and Henriksen (2005) used offshore data from the mid-Norwegian margin to suggest that the Norwegian mountains formed as compressional domes during the Miocene. Japsen et al. (2010) summarised results from a number of studies to argue that the mountains achieved their present elevation during at least three phases of post-Eocene uplift (Michelsen et al. 1998; Martinsen et al., 1999; Faleide et al., 2002; Rasmussen, 2005; Japsen et al., 2007; Gabrielsen et al., 2010a; Rasmussen et al., 2010). A first phase that began at the Eocene–Oligocene transition is indicated by the onset of progradation of clastic wedges away from southern Norway. A second phase in the early Neogene is indicated by exhumation of the basins adjacent to the presently exposed basement areas and by early Miocene coarse-grained braided fluvial systems south of Scandinavia. A final phase in the late Neogene is indicated by widespread exhumation, an intra-Pliocene unconformity and subsequent tilting of the Neogene succession in the eastern North Sea.

3. Development of rifted margins predicted from theory of continental stretching

There have been many attempts to explain the presence of EPCMs. A common feature of most, if not all, of these attempts is that they assume that an EPCM has remained elevated (more or less in its present form) since the time of rifting; see Fig. 2 (e.g. Weissel and Karner, 1989; Gilchrist and Summerfield, 1990; Chéry et al., 1992; ten Brink and Stern, 1992; Watts, 2001). Models consequently attempt to account for an EPCM as a feature that remains uplifted during the post-rift thermal subsidence of the adjacent basin to the present day.

Weissel and Karner (1989) published perhaps the most influential of these papers. These authors rightly state that "*Uplift ceases when extension ends in the case of dynamically supported rift flanks, and whether the topography can be maintained after extension is debatable.*

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In the case of thermally supported rift flanks, uplift diminishes over time at a rate commensurate with thermal equilibrium of the extended lithosphere" (Weissel and Karner, 1989, p. 13919). They then make the following (unsupported) statement that is the cornerstone of a thesis that other authors have accepted uncritically in subsequent years: "*Elevated topography observed along many passive margins of the Gondwana continents . . . suggests, however, that uplifted rift flanks are maintained permanently*". Weissel and Karner (1989) then constructed two models of extension of the lithosphere that can produce the anticipated permanent uplift (their figs. 7 and 8) and a third model (their fig. 9) in which extension of a rifting lithosphere is uniform with depth through the whole lithosphere. There are no problems with Weissel and Karner's (1989) modelling but, in our opinion, there are serious problems with how these models have been interpreted, particularly by other authors.

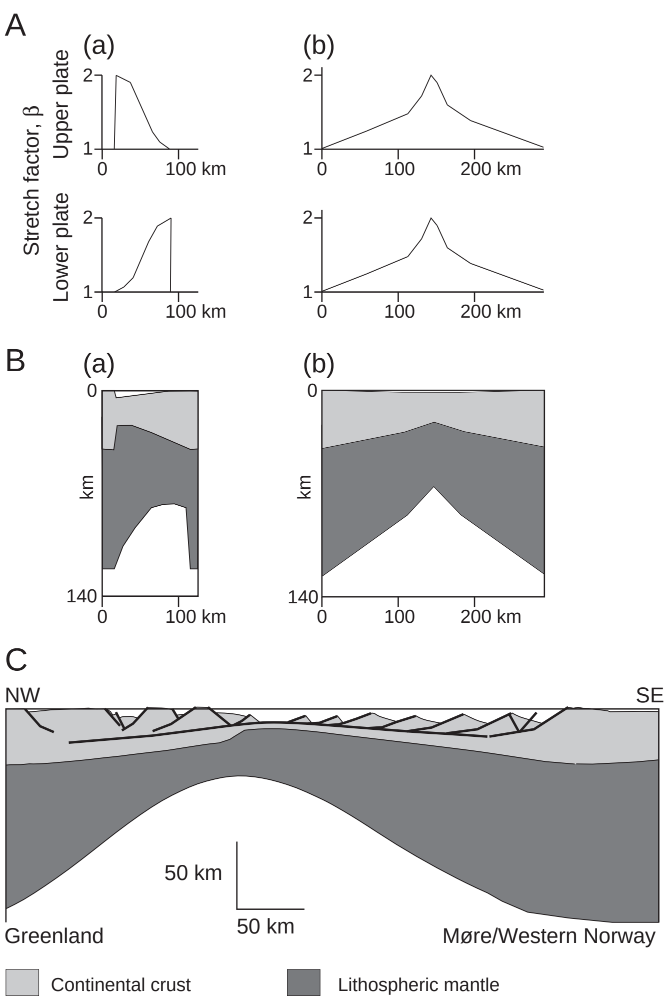

Fig. 5. Distribution of stretching, geodynamic models and a case from the North Atlantic. (A) Relative distribution of stretching in an upper and lower plate (crust and lithospheric mantle) used by Weissel and Karner (1989) to model the syn- and post-rift flexed behaviour of rifts. The amounts of initial foot-wall uplift in all three Weissel and Karner models (their figs. 7, 8 and 9) are very similar to those shown in Fig. 4B. The rift margin remains permanently uplifted if stretching takes place in the very different distributions (asymmetric, simple shear) shown in A(a) (Weissel and Karner, 1989, fig. 7). The rift margin subsides to only minor uplift after 100 Myr (or even negative values; steer's head geometry; White and McKenzie, 1988, fig. 9) if distribution of stretching is uniformly distributed (pure shear) between crust and lithospheric mantle as shown in A(b) and if the elastic thickness is small. Distributions of relative extension in real situations resemble the situation shown in B(b) more than B(a) (see C). Note that the modelled, rift-margin uplifts are much narrower than the real EPCMs whose cross-sections are shown in Fig. 4A. (B) Cross-sections of crust and mantle immediately after rifting produced from the distributions of stretching shown in A, assuming the detachment is at the crust–mantle boundary. Asymmetric distribution (simple shear) of stretching, B(a), leads to permanent uplift of the rift margin, whereas the more realistic symmetric distribution (pure shear), B(b), leads to post-rift subsidence of the rift margins. (C) Interpretation of the situation across the North Atlantic immediately prior to the onset of sea-floor spreading (simplified slightly from Mosar et al., 2002). This interpretation illustrates that the distribution of thinning is locally similar in the crust and in the lithospheric mantle on both sides of the rift, although the rift is asymmetric; cf. B(b). This distribution of stretching corresponds to pure shear. The present-day EPCMs of East Greenland and Norway are respectively just beyond the NW and SE ends of this cross-section. See also e.g. Cloetingh et al. (2005).

Weissel and Karner's (1989) first two models (their figs. 7 and 8) consist of a variation of Wernicke and Burchfiel's (1982) asymmetric, simple shear model where the lithosphere extends independently above and below a detachment (which may be within or at the base of the crust), and where the distribution of stretching above the detachment is very different from that below; see Figs. 5A(a) and 4B(a). The assumptions behind these figures lead inevitably to the formation of a permanently uplifted footwall, because the extension below the detachment is distal relative to the rift margin. The broken upper crust reacts with flexural uplift (Watts, 2001). The material below the detachment is, however, not extended and thinned in this vicinity. It does not, therefore, participate in post-rift cooling, so there is no post-rift subsidence due to cooling of the lithosphere mantle.

Weissel and Karner's (1989) third model (their fig. 9) is based on pure shear where extension of a rifting lithosphere is uniform with depth through the whole lithosphere (Fig. 5A(b) and B(b)). Where the detachment is dipping and there is flexure during the entire rifting and post-rift subsidence period, an initial, small footwall uplift has disappeared by 100 Myr and sediments transgress the footwall. A flat detachment results in a small uplift on both sides of the rift after 100 Myr, but this uplift is very small compared with the sizes of EPCMs. Since this model does not lead to permanent uplift of the rift flanks, it has tended to be dismissed, particularly in the geomorphological and fission-track literature. We find, however, that the geological evidence presented in Section 2 suggests that this model is the most likely of Weissel and Karner's (1989) three models to represent what actually happens at rift margins.

Van der Beek et al. (1994) reviewed the models and parameters needed to produce rift margins that would remain uplifted permanently since the time of rifting and onto which little or no sediment would be deposited. They concluded that the best flexural model for producing permanently-uplifted rift flanks required either of two conditions: (1) that there is lateral offset between crustal extension and mantle extension (simple shear), a conclusion that is essentially identical to that modelled in Weissel and Karner's (1989) figs. 7 and 8; see Fig. 5A(a) and B(a); or (2) that there has to be a substantial difference in effective elastic thickness between the basin and its margins; relatively low elastic thickness in the basin, but up to over 100 km on the rift margin.

Perhaps the best argument that van der Beek et al.'s point (1) does not apply is the countless, successful 1-D models of basin subsidence based on McKenzie's (1978) theory of post-rift subsidence. Those models require for their success the implicit assumption that the extension of the lithospheric mantle underlying a sedimentary basin is similar to that of the crust (pure shear). Models using extension by simple shear (van der Beek et al.'s point 1) are incompatible with the commonly observed post-rift development of steer's-head geometry with post-rift onlap of sediment across the basin margin (Watts et al., 1982; White and McKenzie, 1988; Braun and Beaumont, 1989; Roberts and Yielding, 1991). While White and McKenzie's (1988) model requires extension of the lithospheric mantle under the proximal rift and its margin that is slightly *greater* than extension in the crust, the opposite condition to that assumed by the model's shown in Weissel and Karner's (1989) figs. 7 and 8. Steer's head geometry develops naturally, either by flexural loading of the margins of the basin (Watts et al., 1982; Braun and Beaumont, 1989) or by erosion of the flexurally-uplifted rift flank that then subsides below sea-level by cooling-induced subsidence (Roberts and Yielding, 1991).

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The latter three models are all compatible with extension being symmetric, pure shear.

The evidence that we can find in the literature that is independent of assumptions about permanent uplift of rift margins strongly suggests that there is much less offset between the extension in the crust and the lithospheric mantle than assumed in Weissel and Karner's (1989) figs. 7 and 8 and concluded by van der Beek et al. (1994); see for example Mosar et al. (2002) (Fig. 5C) and Cloetingh et al. (2005). Models in which the distribution of extension is similar in the lithospheric mantle and in the crust show that any syn-rift uplift of the margin should decay to small and even negative values during the subsequent post-rift cooling, unless the effective elastic thickness of the rift margin is very large. We therefore find that there is no *a priori* reason to assume major differences in the distribution of extension between the lithospheric mantle and the crust. Models that assume that such major differences exist have been devised on the assumption that elevated margins remain high after rifting.

With regard to van der Beek et al.'s point (2), some authors (e.g. Chéry et al., 1992; ten Brink and Stern, 1992) have proposed effective elastic thicknesses as great as 115 km to account for some EPCMs. However, Watts' (2001, fig. 8.30) global compilation of effective elastic thicknesses shows that, while such large thicknesses may occur in the centres of ancient shields, effective elastic thicknesses are much lower along rifted continental margins, typically <50 km, including all areas where EPCMs are observed. Additionally, models within rifts that are well-controlled from palaeo-water-depth estimates from boreholes require very low elastic thicknesses, less than 10 km in some cases (e.g. Kusznir and Egan, 1990; Kusznir et al., 1991). White (1999, table 2) lists basins where effective elastic thicknesses less than 10 km have been reported.

Margins with moderate elastic thicknesses will result in initial uplift that is less than and, in particular, much narrower than those observed along present-day EPCMs (compare Fig. 4A and B) and they will still subside with time due to the subsidence of the base of the lithosphere back to its pre-rift depths. We have examined evidence from actual EPCMs in Section **2** and show that many features of these margins cannot be explained if the margins are elevated permanently since breakup. We can find no independent evidence in the literature that they have been.

Isostatic response to valley erosion within uplifted area will cause the uneroded portions of the margin to remain uplifted longer. However, the average height of an eroded margin will still decrease with time. Medvedev et al. (2008) claimed that loading by sediments in the rift may give rise to a peripheral bulge. Watts et al. (1982) and Braun and Beaumont (1989), on the other hand, showed that the load causes additional subsidence of the rift margin, forming steer's head geometry, not a bulge.

We conclude that there is nothing in current theories of the development of rifts and their margins that *require* there to be margins that remain uplifted since the time of rifting. The existence of EPCMs requires explanation, but the explanation has to be consistent with observations such as those listed in Section 2 which demonstrate that EPCMs are not permanent highs but formed post-rift. In Section 4 we outline a *conceptual* model of the development of an EPCM that is consistent with these observations. Any *geodynamic* model to explain EPCMs needs to be consistent with this qualitative model.

4. A conceptual model for EPCM development

The features that characterise most, if not all EPCMs were described in Section 2c. Here we present a conceptual model (Fig. 6) which may explain how these characteristics of EPCMs develop and which is also consistent with the low elastic thicknesses estimated from well-controlled models within rifts.

Fig. 6A shows a cross-section through a margin some 30–50 Myr after cessation of rifting. Fig. 6B shows the margin after a first phase of post-rift uplift and denudation, at a stage when fluvial erosion has formed a peneplain near sea level. The surface of the peneplain reflects only to a small degree the age and the resistance of the underlying rocks. The uplift represents a reversal of the initial post-rift

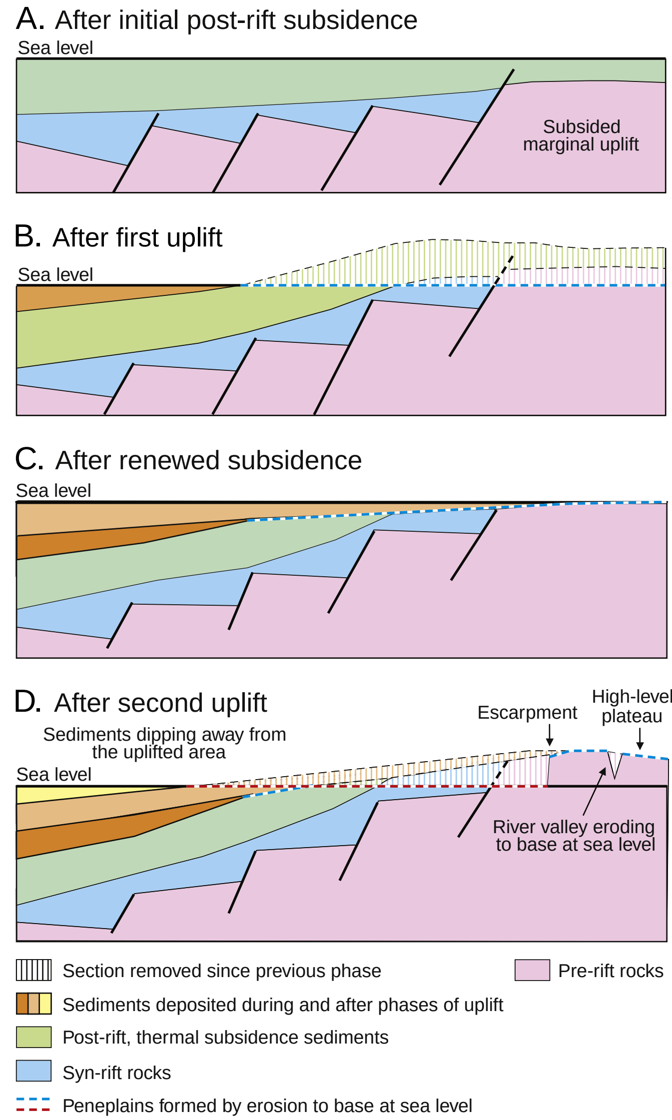

Fig. 6. Cartoon to illustrate the post-rift development of an elevated, passive continental margin based on observations of EPCMs around the world. Compare with Fig. 2. (A) A rift margin in continental crust approx. 30–50 Myr after cessation of rifting: Continental breakup to form oceanic crust is farther to the left. Syn-rift sediments (blue) accumulated in the accommodation space made available by down-faulting and rotation of fault-blocks. After cessation of rifting, cooling of the rift and its margins caused the area to subside and post-rift sediments (green) to fill the available accommodation space and to transgress the former flexurally-uplifted margin. (B) After one phase of uplift of the rift and its margin: Fluvial erosion has formed a peneplain a short distance above sea level. The surface of the peneplain is uniform regardless of the resistance of the underlying rocks (basement, syn- or post-rift sediment. Subsidence continued offshore and the accommodation space filled with sediment (orange). (C) After renewed subsidence: Sediments (beige) cover the erosion surface forming an erosional unconformity in the basin and burying the peneplain across basement rocks to some extent. (D) After a second phase of uplift: River valleys grading to sea level coalesce to form an escarpment that erodes back to form a new peneplain (see Fig. 1C–E). The peneplain that formed after the first phase of uplift is now a high-level planation surface. Sediments that were deposited horizontally in the post-rift section now dip seaward and are truncated by the new erosion surface whereas the distal part of the old erosion surface remains buried. Subsidence still continued offshore and the accommodation space filled with sediment (yellow). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

subsidence and thus leads to exhumation of the basin proximal to the margin. In West Greenland, post-rift subsidence lasted for c. 25 Myr whereas the subsequent formation of the peneplain took c. 20 Myr (Japsen et al., 2006, 2009).

Renewed burial may affect the margin after the first uplift phase (Fig. 6C). Sediments cover the erosion surface leading to reburial of the peneplain to some extent beyond the rift margin and to formation of an erosional unconformity within the basin. This is similar to the situation in offshore West Greenland where Neogene sediments cover the erosional, Oligocene unconformity. A second phase of uplift (Fig. 6D) raises the peneplain that formed after the first phase of uplift, to become a high-level, low-relief surface as in West Greenland c. 50 Myr after rifting (Japsen et al., 2006). The new uplift causes rivers to erode below the now-uplifted peneplain and eventually to the formation of a new peneplain near sea level (compare with Fig. 1A–D). The new erosion surface truncates sediments in the post-rift section that now dip seaward after the uplift whereas the distal part of the old erosion surface remains buried.

This qualitative model explains satisfactorily the erosional truncation of the post-rift sedimentary sequences along passive margins (Fig. 3) and the presence of low-relief surfaces at high elevation on many EPCMs (Fig. 1). In particular, our model suggests that the elevated surfaces are uplifted peneplains that formed as erosion surfaces near sea level. We know of no other published model that explains formation of these extensive surfaces in both tropical and polar regions (Fig. 1).

A consequence of our explanation is that the formation of an EPCM requires at least two episodes of uplift, one to provide the relief to be eroded to form the first peneplain and a second to lift it to its present heights. The model can be extended to EPCMs with stepped topography such as Norway (Lidmar-Bergström et al., 2000). Each low-relief surface is a peneplain that formed near sea-level and requires a separate uplift episode to raise it.

5. Plausible and implausible explanations

There is a wide range of passive continental margins at the present day (Fig. 1); some are low-lying and some are elevated. This simple observation shows that the processes of rifting and breakup do not necessarily lead to the formation of an EPCM. The present elevation of EPCMs is therefore not a consequence of breakup, but of later processes that may or may not be activated along the margins. Elevated plateaux cut by deep valleys appear to be ubiquitous in the large-scale morphology of all EPCMs and appear naturally within the conceptual model presented in the previous chapter. We therefore suggest that EPCMs were shaped by a sequence of events similar to that shown in Fig. 6. It is important to note that the characteristic features of EPCMs are present in margins with clear differences in their geological development prior to and during breakup (e.g. volcanic and non-volcanic) and in the time span available since breakup (e.g. c. 112 Myr vs. 55 Myr for the South and the North Atlantic, respectively) (see Fig. 1).

Numerous models presented in the literature attempt to explain why there are mountains along many passive continental margins. But many of these models are based either on theoretical considerations without sufficient empirical constraints or they are based on the assumption that a single EPCM can be explained by special conditions rather than by conditions that apply to EPCMs world-wide. Most models fail to account for the presence of many features that characterise most, if not all EPCMs as outlined in Section 2. Based on these observations and other facts known from the literature it is possible to conclude that EPCMs are:

- not due to movements that took place during rifting or breakup (Figs. 4, 6).
- not permanent. The West Greenland margin was uplifted several times post-rift and subsided below sea-level again at least once between two uplift events.

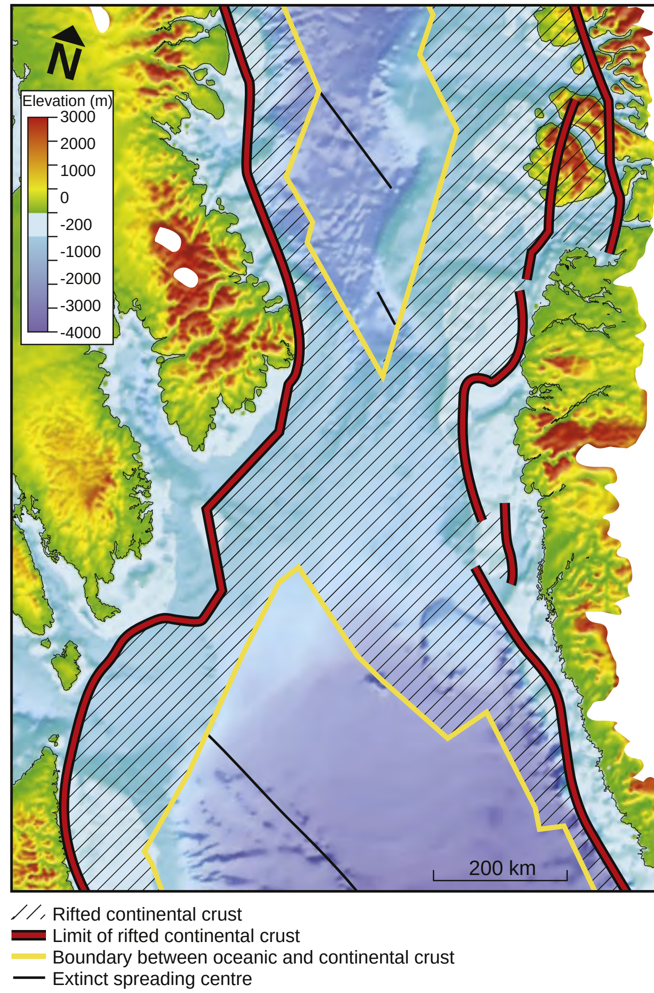

Fig. 7. Map of the area between Canada and Greenland illustrating the close relation between EPCMs and the edges of the cratons as indicated by the limit of the rifted continental crust (Chalmers and Pulvertaft, 2001). The West Greenland EPCM was formed by late Neogene uplift, c. 50 Myr after breakup west of Greenland (Japsen et al., 2006) and the east–west symmetry suggests that this was also the case with the Canadian EPCM. The presence of Eocene marine sediments several hundred metres above sea level on the Canadian margin (MacLean and Falconer, 1979) supports this conjecture. Location of map shown in Fig. 1. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

- not due to ridge push alone because a rifted margin can also subside during the spreading of the adjacent ocean as we have shown in this paper. The West Greenland EPCM developed tens of millions of years after cessation of sea-floor spreading in Labrador Sea and Baffin Bay.
- not the erosional remains of a former orogen. The mountains around the North Atlantic extend well beyond the Palaeozoic, Caledonian orogen and contain rocks formed from the Archaean to the Paleogene. Many EPCMs contain the eroded remains of post-orogenic rift basins now at substantial altitudes.
- not caused by glacial processes or solely by isostatic uplift in response to glacial erosion. The presence of elevated plateaux in both tropical and polar regions shows that formation of these landscapes is independent of climatic conditions (Fig. 1A–D). Some component of uplift must, however, be due to the flexed isostatic response to removal of material after initial tectonic uplift.
- not in general supported by crustal roots. The Norwegian EPCM is underlain by crust of the same or even less thickness than that under the adjacent lowlands (Fig. 8) (Tesauro et al., 2008, Stratford et al., 2009).

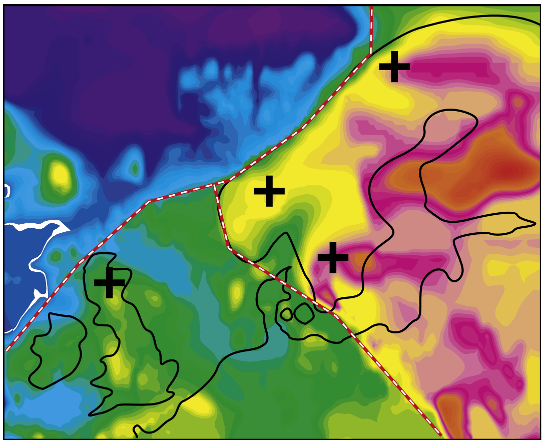

Fig. 8. Map of crustal thickness under north-west Europe. There is no crustal root under the Norwegian mountains, although thin crust under the Oslo Graben gives the false appearance of one under southern Norway. Note the sharp contrast in crustal thickness between the north-west European EPCMs and the rifted crust bordering the Atlantic Ocean to the NW and between the elevated area in southern Sweden and the thinner continental crust to its west. Map after Tesauro et al. (2008). Colour scale: Purple (< 10 km), light blue (c. 10–25 km), yellow (c. 25–40 km), pink (c. 40–50 km), dark orange (c. 50– > 60 km). Centres of Neogene uplift after Japsen and Chalmers (2000). Location of map shown in Fig. 1. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

- not supported by magmatic underplating. P-wave velocities in the lower crust under the Norwegian EPCM are slower (and the lower crust is thus probably less dense) than under adjacent low-lying Sweden (Tesauro et al., 2008).
- not lifted by edge-driven convection because this mechanism causes downwelling at the lithospheric discontinuity (King and Anderson, 1998).
- not caused by conditions that are unique to only one or a few EPCMs, since they are common to passive margins of all continents (Fig. 1). Processes related only to e.g. the Iceland Plume, the African Superplume, glaciation around the North Atlantic or to movements along the Andes cannot be invoked alone to explain the general aspects of EPCMs.

EPCMs are, however, located above thick crust/lithosphere that is closely juxtaposed to thinner crust/lithosphere. The presence of mountains close to where continental crust starts to thin towards oceanic crust, illustrates the relation between EPCMs and the edges of cratons; e.g. on Baffin Island (eastern Canada) and in western Greenland (Fig. 7), in Scandinavia, Scotland and Ireland (Fig. 8). Similarly, the dominant topographic feature in southern Scandinavia, the South Swedish Dome (<400 m a.s.l.; Lidmar-Bergström, 1996), is situated on thick crust close to the thinner crust of west Europe (Fig. 8).

Passive continental margins clearly formed as a result of extension. Despite this, the World Stress Map (www.world-stress-map.org) shows that, where data exist, all EPCMs are today under compression. Cloetingh and Burov (2010 and references therein) showed that the lithosphere within continents folds under compression to form massifs and intervening basins, but they did not extend their analysis to the margins of continents. We suggest that EPCMs may be similar anticlinal, lithospheric folds caused by compression applied a long time after rifting. Similarly, Leroy et al. (2004) proposed that deformation and uplift along passive margins were related to horizontal compression. Løseth and Henriksen (2005) suggested this to be the case for the Norwegian margin, and based on the results of e.g. Davy and Cobbold (1991) and Burg and Podladchikov (1999), they argued that lithospheric folding is a basic response to large-scale continental shortening and an efficient mountain building process (see also Doré et al., 2008). Our observations and our conceptual model suggest that the heights of EPCMs are large compared to massifs within cratons, and that this additional height may be due to processes operating where lateral contrasts in thickness of the crust or lithosphere make the margins of the cratons unstable (e.g. Praeg et al., 2005; Japsen et al., 2006; Osmundsen et al., 2010). Whether a passive margin is low-lying or elevated will consequently depend on the geometry of the edge of the crust or lithosphere and on the stress exerted on it in the immediate geological past (typically, within the Neogene).

Application of compressive stress to the edge of a craton may thus lead to uplift, the consequent formation of extensive erosion surfaces and their subsequent uplift as well as to truncation of the post-rift sequence along the margin. That such vertical movements are related to changes in stress is also supported by the observation that a phase of uplift and erosion that affected margins around the North Atlantic at the Eocene–Oligocene transition (c. 35 Ma), correlates in time with a major plate reorganisation there (cf. Gaina et al., 2009; Green and Duddy, (2010); Japsen et al., 2010). Holford et al. (2009, 2010) also suggested that episodes of regional exhumation of the British Isles and of southern Australia were caused by far-field, compressional stresses. Geodynamic modelling of lithospheric folding along EPCMs and documentation of the changes in plate movements that are needed to produce the necessary forces for the episodic, vertical movements along these margins are important subjects for future studies.

6. Conclusions

Elevated plateaux with deeply incised valleys are common features of EPCMs around the world. The West Greenland margin is a typical EPCM, and the geological record shows that it is not a remnant of the rifting process but the result of much later uplift which partially removed a thick, post-rift section. Evidence from other EPCMs suggests that such post-rift subsidence and later uplift is a general feature of EPCMs. The deposition of post-rift sediments inland of a rifted margin agrees with models of margin development that assume that extension in the crust and in the mantle is locally similar and that effective elastic thicknesses are low or moderate. The wide range between low-lying and elevated, passive continental margins indicates that the present elevation of a passive margin is not a consequence of the processes of rifting and breakup, but of much later processes that may or may not be activated along them.

Much geomorphological literature is, however, based on the assumption that the present EPCM morphology has remained essentially unchanged since the time of rifting and that absence of post-rift sediments is evidence of non-deposition. Many geophysical studies attempt merely to adjust model parameters to the assumption that EPCMs are permanently elevated after rifting. These assumptions lead to the conclusion that EPCMs have to form as a result of Wernicke and Burchfiel's (1982) asymmetric, simple-shear extension in which extension of the lithospheric mantle takes place very distally compared to extension of the crust. Other models require that the effective elastic thickness of EPCMs must be greater than 100 km, in contrast to published estimates independent of assumptions of permanent uplift (e.g. White 1999; Watts, 2001). More realistic models of rifting using low values of elastic thickness controlled by well data predict initial flexural uplift of rift flanks followed by subsidence and deposition of a thick post-rift sequence overlying both the rift and its margins. The recognition that the elastic thickness of EPCMs and their adjacent basins are of a similar, low to moderate magnitude implies that the margins of cratons are much more mobile than assumed in many studies.

We have demonstrated that the morphology of EPCMs is unrelated to the processes taking place during rifting and continental separation and that EPCMs have not remained elevated since the time of rifting, as is assumed in many studies. We suggest instead that EPCMs are anticlinal, lithospheric folds formed under compression similar to those discussed by Cloetingh and Burov (2010). If so, the vertical

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movements are related to compressive stress operating in the lower crust and/or mantle and are produced where there is an abrupt, lateral change in crustal or lithospheric thickness at such locations. We suggest that EPCMs are produced by repeated episodes of compression-induced uplift, and that the elevated plateaux observed along many margins are expressions of these episodic movements. An initial episode of uplift is needed to erode the region to produce the low-relief surface near the level of the adjacent, opening ocean. A second episode is then needed to raise the margin to its present elevation, after which the plateau is dissected and progressively denuded to produce the characteristic EPCM landscapes of e.g. West and East Greenland, SE Australia, southern Africa and eastern Brazil.

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