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Stratigraphic landscape analysis, thermochronology and the episodic development of elevated, passive continental margins

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Apatite fission-track analysis, base level, continental margins, denudation, geomorphology, Great Escarpment, peneplain, stratigraphic landscape analysis, thermochronology, uplift.

Cover illustration
Gåseland, East Greenland (70°N, 28°W), looking north-west. The dominant 1900-m plateau is an erosion surface that truncates Palaeogene basalts which cover the undulating, weathered basement in the foreground. After uplift which started around the Eocene–Oligocene transition, the plateau surface was formed by fluvial incision and slope processes, and ultimately graded to sea level. The landscape was then uplifted again in the Miocene and incised by rivers. The present plateau is smoothed by periglacial processes and the river valleys have been widened and deepened by glacial erosion. (Bonow *et al*. 2014, in press).

Frontispiece: facing page
Planalto da Conquista, Atlantic margin of Brazil (16°S, 42°W, about 200 km from top to bottom). The two low-relief surfaces at *c*. 900 m (red) and 300 m (green) above sea level are erosion surfaces graded to the base level of the adjacent ocean. The higher surface is Palaeogene and was uplifted to its present elevation during the Miocene, after which the lower surface formed by incision below the uplifted, higher surface (Japsen *et al*. 2012b).

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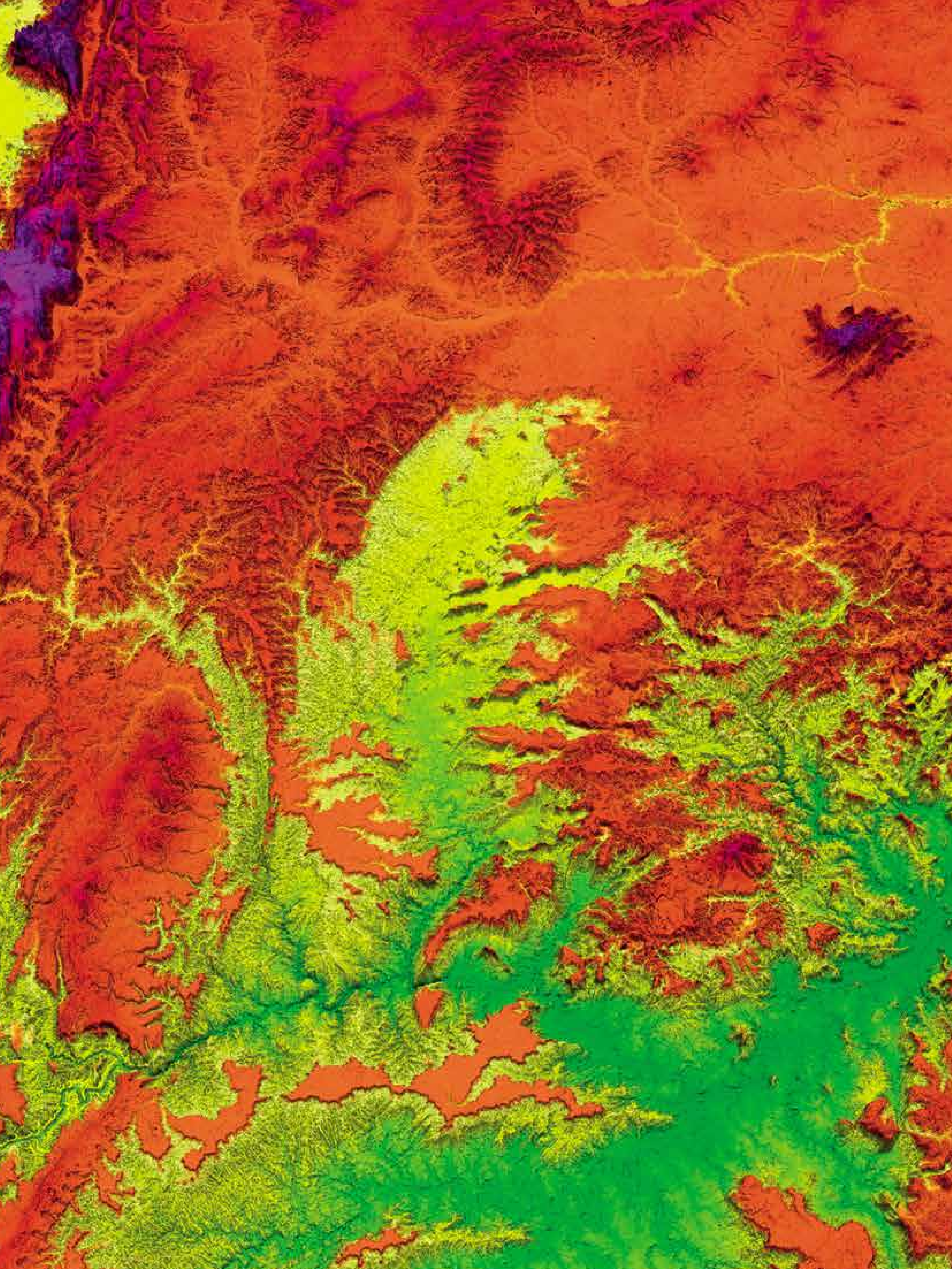

Contents

Abstract

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The continental margin of West Greenland is similar in many respects to other elevated, passive continental margins (EPCMs) around the world. These margins are characterised by extensive regions of low relief at elevations of 1–2 kilometres above sea level sloping gently inland, with a much steeper, oceanward decline, often termed a 'Great Escarpment', terminating at a coastal plain. Recent studies, based on integration of geological, geomorphological and thermochronological evidence, have shown that the high topography of West Greenland was formed by differential uplift and dissection of an Oligo-Miocene peneplain since the late Miocene, many millions of years after continental break-up between Greenland and North America. In contrast, many studies of other EPCMs have proposed a different style of development in which the high plateaux and the steep, oceanward decline are regarded as a direct result of rifting and continental separation. Some studies assume that the elevated regions have remained high since break-up, with the high topography continuously renewed by isostasy. Others identify the elevated plains as remnants of pre-rift landscapes.

Key to understanding the development of the West Greenland margin is a new approach to the study of landforms, stratigraphic landscape analysis, in which the low-relief, high-elevation plateaux at EPCMs are interpreted as uplifted peneplains: low-relief surfaces of large extent, cutting across bedrock of different age and resistance, and originally graded to sea level. Identification of different generations of peneplain (re-exposed and epigene) from regional mapping, combined with geological constraints and thermochronology, allows definition of the evolution leading to the formation of the modern-day topography. This approach is founded particularly on results from the South Swedish Dome, which document former sea levels as base levels for the formation of peneplains. These results support the view that peneplains grade towards base level, and that in the absence of other options (e.g. widespread resistant lithologies), the most likely base level is sea level. This is particularly so at continental margins due to their proximity to the adjacent ocean.

Studies in which EPCMs are interpreted as related to rifting or break-up commonly favour histories involving continuous denudation of margins following rifting, and interpretation of thermochronology data in terms of monotonic cooling histories. However, in several regions, including southern Africa, south-east Australia and eastern Brazil, geological constraints demonstrate that such scenarios are inappropriate, and an episodic development involving post-breakup subsidence and burial followed later by uplift and denudation is more realistic. Such development is also indicated by the presence in sedimentary basins adjacent to many EPCMs of major erosional unconformities within the post-breakup sedimentary section which correlate with onshore denudation episodes.

The nature of the processes responsible is not yet understood, but it seems likely that plate-scale forces are required in order to explain the regional extent of the effects involved. New geodynamic models are required to explain the episodic development of EPCMs, accommodating post-breakup subsidence and burial as well as subsequent uplift and denudation, long after break-up which created the characteristic, modern-day EPCM landscapes.

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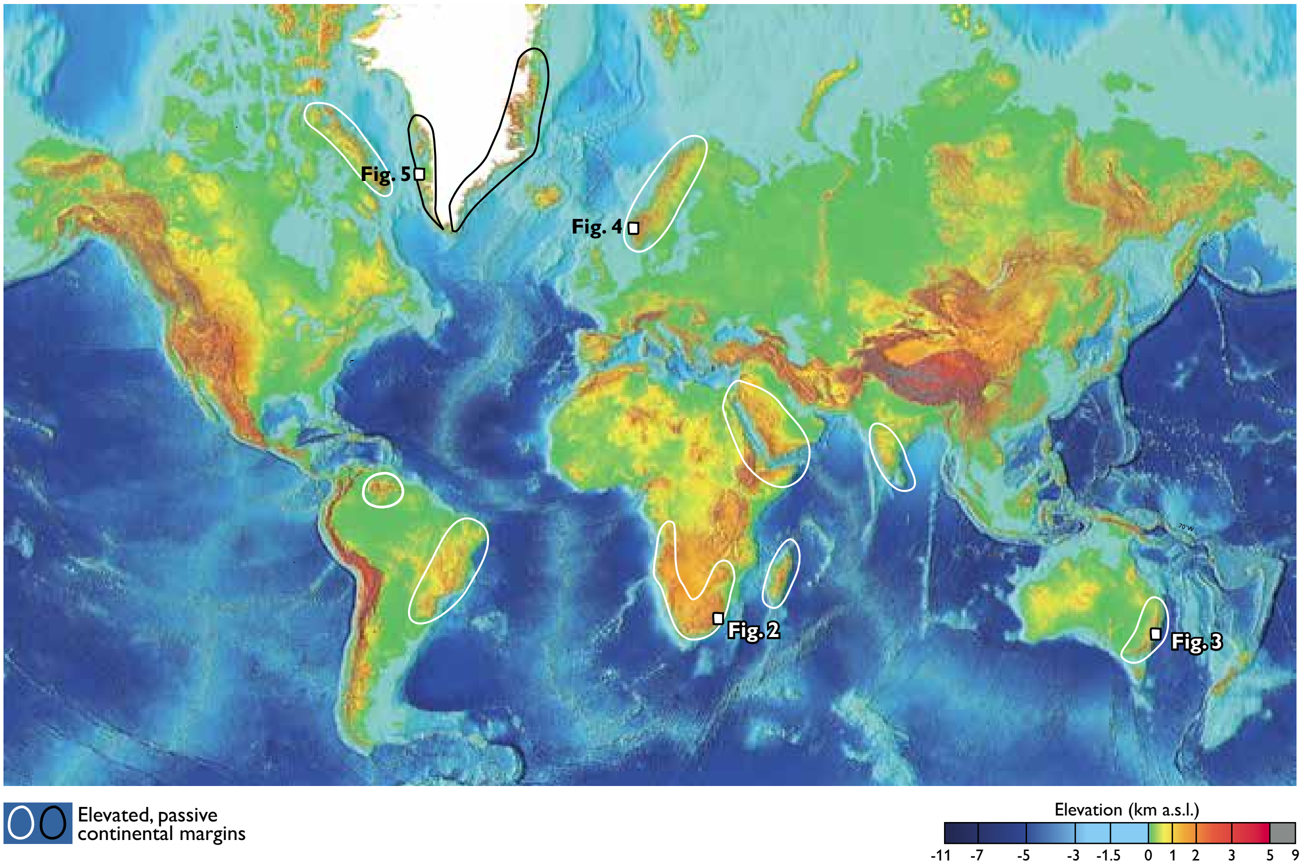


Fig. 1. Topography of the Earth between 80°N and 60°S showing elevated, passive continental margins (EPCMs). Only Atlantic-type margins that can easily be connected to the corresponding spreading centre and only elevated margins that reach 2 km a.s.l. in more than a single summit are indicated (cf. Bradley 2008). Global relief model: *http://www.ngdc.noaa.gov/mgg/image/images/g01929-pos-a0001small.pdf*. Modified from Japsen *et al*. (2012a).

1. Introduction

The topography and relief of the continental margin of West Greenland are similar in many respects to those of other rifted continental margins around the Earth. Margins in locations as diverse as south-east Australia, southern Africa, Brazil, Norway, western India, Saudi Arabia and Yemen (Fig. 1) are characterised by inland regions of low relief up to 1 or 2 km or more above sea level (a.s.l.), declining gently inland and separated from a narrow coastal plain by a region of much steeper oceanward decline (Jessen 1948). The relief of these elevated, passive continental margins (EPCMs) is characterised by high level plains, commonly in stepped sequences, with deeply incised valley systems (Figs 2–5). At many margins, Mesozoic–Cenozoic rift systems parallel to the coast are present offshore with a transition from continental to oceanic crust farther offshore. Post-rift sedimentary sequences offshore commonly dip away from the coast and are truncated below unconformities which correlate with onshore denudation episodes (Fig. 6; Japsen *et al.* 2012a and references therein).

Despite many years of intensive study, the tectonic development of EPCMs is still only poorly understood, particularly in regard to the timing and magnitude of uplift, amounts and timing of denudation ('missing section') and the nature of the underlying processes. Since the 1950s and 1960s there has been much dispute about the origin of the high plains as peneplains graded to former sea levels, and the idea has passed out of favour among many geomorphologists (Phillips 2002). Many studies are based on the assumption that EPCM morphology is directly related to the processes of rifting, continental break-up and separation, and that EPCMs have remained high since prior to break-up, as a result of continuous erosion and isostatic adjustment (e.g. Gilchrist & Summerfield 1991; Gallagher & Brown 1997; Gallagher *et al.* 1998; Bishop 2007). Others (e.g. Ollier & Pain 1997) have identified the elevated plains as remnants of pre-rift landscapes which have remained largely unaffected by later processes.

In contrast, recent studies of the West Greenland margin have shown that the present-day mountains, with plateaux up to 2 km or more a.s.l., represent a differentially uplifted, broken and dissected planation surface graded to sea level during the Oligo–Miocene and uplifted in the late Miocene and Pliocene (Japsen *et al.* 2005, 2006, 2009; Bonow *et al.* 2006a, b). Following Paleocene break-up between Greenland and North America, sea-floor spreading ceased at the end of the Eocene (Chalmers & Pulvertaft 2001), so the uplift of these plateaux postdates rifting and continental separation by many millions of years, and the present-day topography cannot be related to continental break-up. Since the West Greenland margin shares many of the characteristics common to ECPMs worldwide, these results seriously question the common belief that the elevated plateaux with oceanward escarpments characteristic of EPCMs are directly related to continental rifting and separation.

A cornerstone of these studies in West Greenland has been the combination of observations from the onshore geological record with a) an improved methodology for landscape analysis (Stratigraphic Landscape Analysis, SLA, see chapter 3), which defines a relative chronology of denudation, subsidence and uplift, and b) thermochronology incorporating analysis of both basement and overlying sedimentary cover rocks, which provides an absolute timescale and magnitude for key denudational and depositional events (chapter 4). The combination of these methods results in a consistent definition of an episodic style of evolution, involving multiple episodes of both positive and negative vertical movements. Integration of the resulting history with additional information from geological constraints offshore (Chalmers 2000; Japsen *et al.* 2005, 2006; Bonow *et al.* 2007b) has provided a coherent regional framework for deciphering the post-rift tectonic development of the West Greenland EPCM, as explained in chapter 5.

In contrast to this consistent fusion of stratigraphic landscape analysis and thermochronology in West Greenland, many studies of EPCMs in other parts of the world have revealed a conflict between conclusions derived from conventional landscape studies and thermochronology (e.g. Ollier & Pain 1997; Peulvast *et al.* 2008). Any viable model for the development of EPCMs should be consistent with all available constraints. Chapter 6 describes studies of a number of other EPCMs (including southern Africa, south-east Australia and Brazil) where previous studies have led to conflict between landscape studies and thermochronology, and proceeds to illustrate how the different approaches can be integrated to define an episodic style of development similar to that established for the West Greenland margin in chapter 5. Chapter 7 deals with the nature of the underlying processes, and possible mechanisms which might explain the episodic development of EPCMs. Several key issues regarding the development of EPCMs are highlighted and discussed in chapter 8.

This study will particularly focus on two major problems in previous studies of EPCMs. The first is the notion of permanently uplifted margins and the paradigm of steady state (uplift due to isostasy keeps pace with erosion; see section 2.3) which for a long time prevented the use of peneplains as uplift markers in geomorphology. The second is the common adoption in thermochronological studies of monotonic cooling histories. Basic constraints provided by geological observations which impose critical restrictions on viable solutions have commonly been overlooked in both geomorphological and thermochronological studies.

The evidence presented here leads to the conclusion that the high present-day elevation of many passive continental margins is not directly related to processes of continental rifting and break-up, but to processes taking place subsequently. The common assumption that EPCMs are permanently elevated cannot be sustained, and a reassessment of the development of EPCMs is required. The constraints that can be derived from thermochronology and landscape analysis, integrated with the geological record, can provide the basis for such reassessment. A detailed understanding of the development of EPCMs is important not only in terms of understanding the properties of the Earth's lithosphere and mantle but also has major implications for hydrocarbon exploration, e.g. in terms of sediment supply to offshore basins.

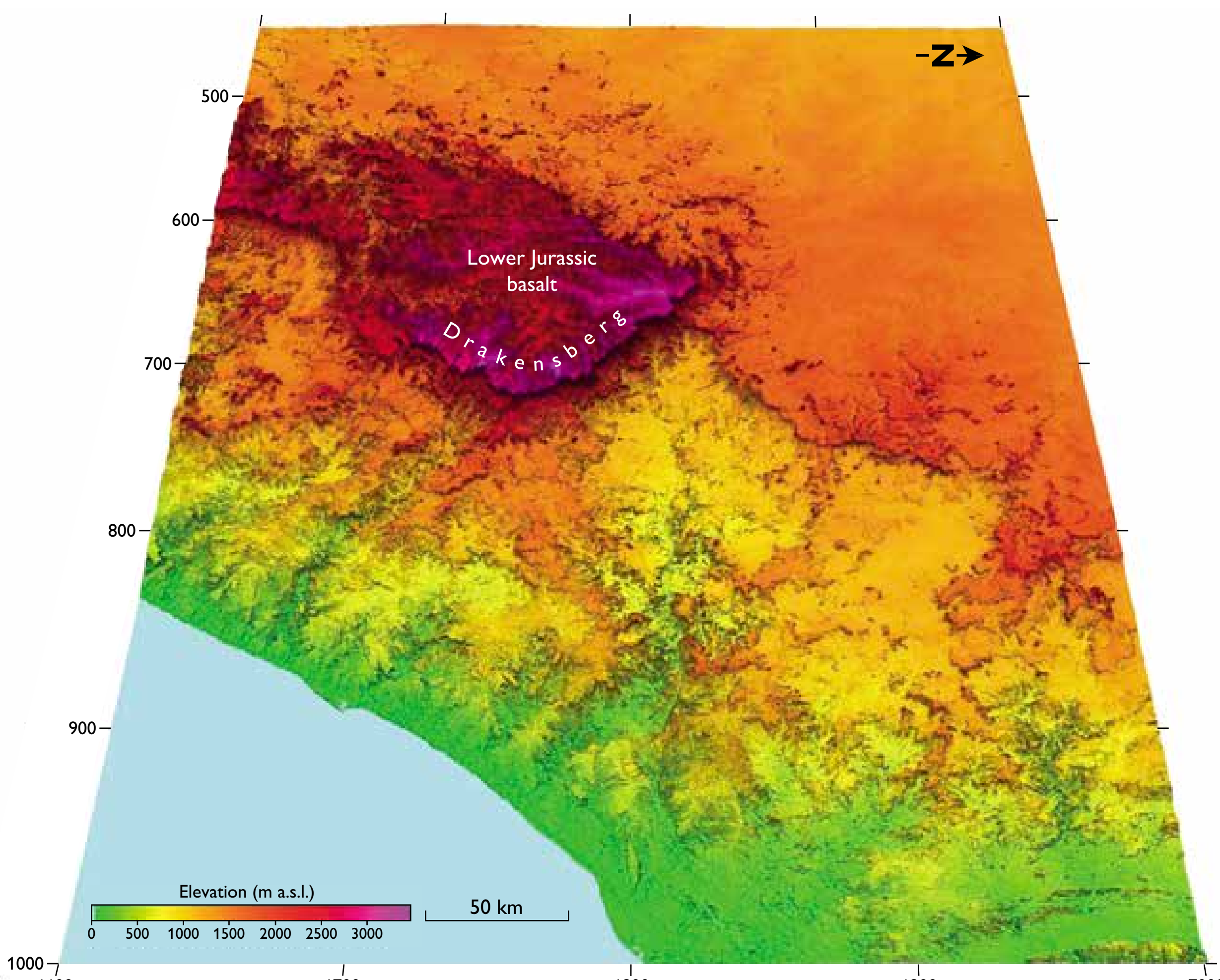


Fig. 2. South Africa, south-east coast, with the Drakensberg Escarpment in Lower Jurassic basalt. Below it are several planation surfaces separated by major steps (cf. King 1972) including the African Surface at about 1800 m a.s.l., a major plain at about 1000 m a.s.l., and a coastal plain. Two valley generations are incised below the 1000 m planation surface. The general picture of the relief at the elevated, passive margin of South Africa is of high-level plains and deeply incised valleys. The map covers most of the KwaZulu-Natal province. 3D-terrain model with vertical shading. UTM coordinates (km) zone 36S. Elevation data from Jarvis *et al*. (2008).

2. Geomorphological concepts related to the evolution of elevated, passive continental margins (EPCMs)

2.1 Peneplains as key data for understanding denudation, subsidence and uplift of EPCMs

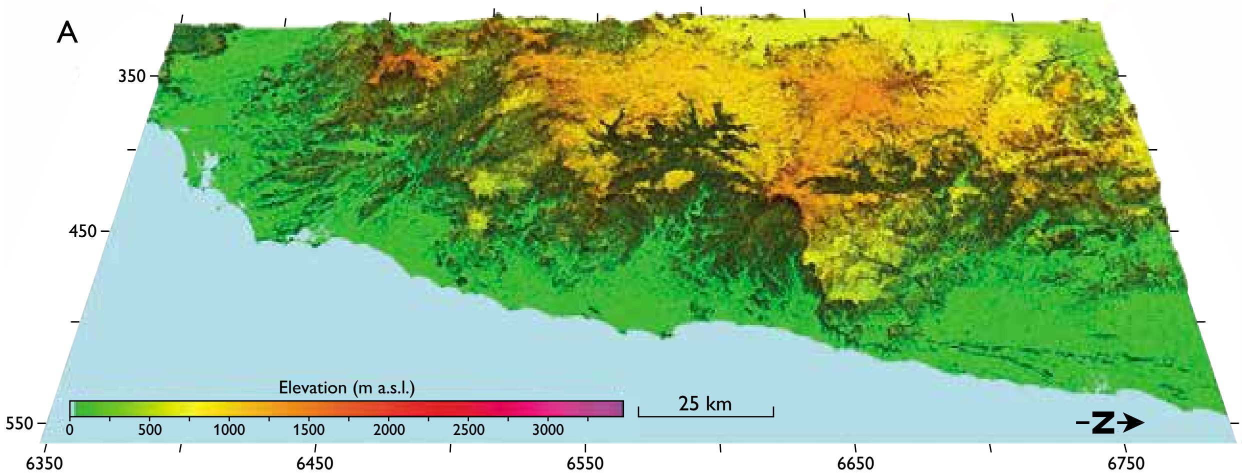
The relief of EPCMs (Figs 2–5) commonly takes the form of high-level plains, separated by more or less pronounced steps. A gentle overall slope characterises their inland continuation, while the oceanward slope is steeper and can take a variety of forms, including distinctly separated steps (Fig. 2), inclined plains (Figs 3, 4) or tilted fault blocks (Fig. 5). In all cases, deeply incised valleys are present. What these margins specifically have in common are the high-level plains and incised valleys, which according to traditional landscape analysis suggest late uplift events, as the plains have been interpreted as having formed close to sea level (Davis 1899; Ahlmann 1919, 1941; Reusch 1901; Johnson 1931; King 1951, 1962, 1967, 1972, 1976; Peulvast 1985;, 1987; Partridge & Maud 1987).

The margins can show tilted surfaces (e.g. Australia, Fig. 3), where the incised valleys have coalesced to form a Great Escarpment (Ollier 1982). The notion of a Great Escarpment has been used to characterise a variety of margins (e.g. Ollier 1985; Ollier & Marker 1985; Partridge & Maud 1987; Lidmar-Bergström *et al.* 2000; Gunnell *et al.* 2003) although the so-called Great Escarpments greatly differ in shape in different settings (see above). The inclined plains characterising some margins can be interpreted as re-exposed unconformity surfaces (Figs 3, 4; section 3.3) and suggest the former presence of more extensive covers. The high plains and the re-exposed unconformities are important data for revealing histories of denudation, subsidence and uplift (e.g. Lidmar-Bergström 1993; Bonow 2005; Bonow *et al.* 2006a; Lidmar-Bergström *et al.* 2013). Although glaciation produces specific features such as deep fjords and erases summit plains in locations exposed to cirque and valley glaciation, which form an alpine relief (Figs 4, 5; Lidmar-Bergström *et al.* 2000; Mitchell & Montgomery 2006; Bonow *et al.* 2006b; Etzelmüller *et al.* 2007), the high plains and incised valleys are common features at all EPCMs, whether they are glaciated or not.

Low-relief landscapes hundreds of kilometres in extent at both high and low elevations are common features on the Earth's continents (e.g. King 1962, 1967, 1976; Bird 1959; Ollier 1981; Hall 1991; de Brum Ferreira 1991; Lidmar-Bergström 1996; Gunnell 1998; Johansson 1999; Godard *et al.* 2001b; Lageat & Gunnell 2001; Casa-Sainz & Cortés-Gracia 2002; Demoulin 2003; Bonow *et al.* 2003, 2006a). They were labelled peneplains and used at the beginning of the 20th century to constrain base-level changes and uplift (e.g. Davis 1899; Reusch 1901; Ahlmann 1919; Johnson 1931). The following principles were applied: 1) large areas of low relief at high levels forming subhorizontal or tilted planes across different rock types are produced by erosion to a common base level (the sea, in near-shore positions) and have subsequently been uplifted to their present elevations; 2) scattered summit surfaces at higher elevations bevelling different rock types may be remnants of older peneplains; 3) valley incision to common lower levels shows younger generations of erosion graded to new base levels. If the peneplains cannot be related to any covers and have thus been exposed since formation, they are labelled epigene (Twidale 1985).

Since the definition in the late 19th century of a peneplain as the result of fluvial incision and ultimate grading to base level, knowledge about deep weathering processes and resulting bedrock shapes arose in the 1950s (section 3.2). It is now also known that some peneplains have been preserved below cover rocks of different age (section 3.1.2). Therefore a new classification of peneplains is presented in this paper (section 3.2.4.) which embraces low relief surfaces with different shapes and saprolites characteristics. Neither the result of deep weathering nor the importance of former covers have been included in previous discussion of peneplains.

In the prevailing paradigm of steady state and dynamic equilibrium (Inkpen 2005) the existence of peneplains is regarded as elusive (Phillips 2002). On the contrary, here the high plateaux are interpreted as peneplains (see section 3.1), which we define as low-relief erosion surfaces graded to distinct former base levels, irrespective of their detailed development. The elevated peneplains can be labelled transient landscapes (Bishop 2007), as they are under destruction by valley incision.



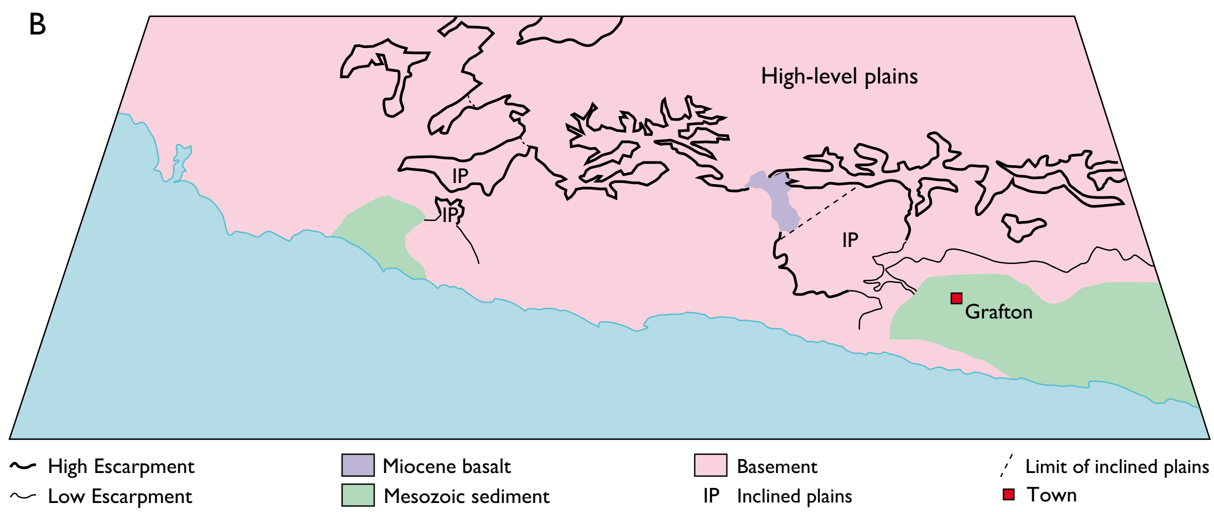


Fig. 3. Eastern Australia. **A**: 3D-terrain model with vertical shading. **B**: Line drawing of the map in **A**. There is a major plateau at 800–1200 m a.s.l. and incised valleys coalescing to the Great Escarpment and a coastal plain. The escarpment is lower where it cuts into inclined plains (**IP**) that are tilted towards the coast. South-west of Grafton, such an inclined plain forms the surface of a triangular basement block, which seems to disappear below Jurassic sedimentary strata (Geology of Australia 1976). This tilted and thus probably sub-Jurassic surface, is cut off by the high plain, here capped by Miocene basalts (Johnson 2004). Scale bar applies only to the foreground of the 3D model. UTM coordinates (km) zone 56S. Elevation data from Jarvis *et al.* (2008).

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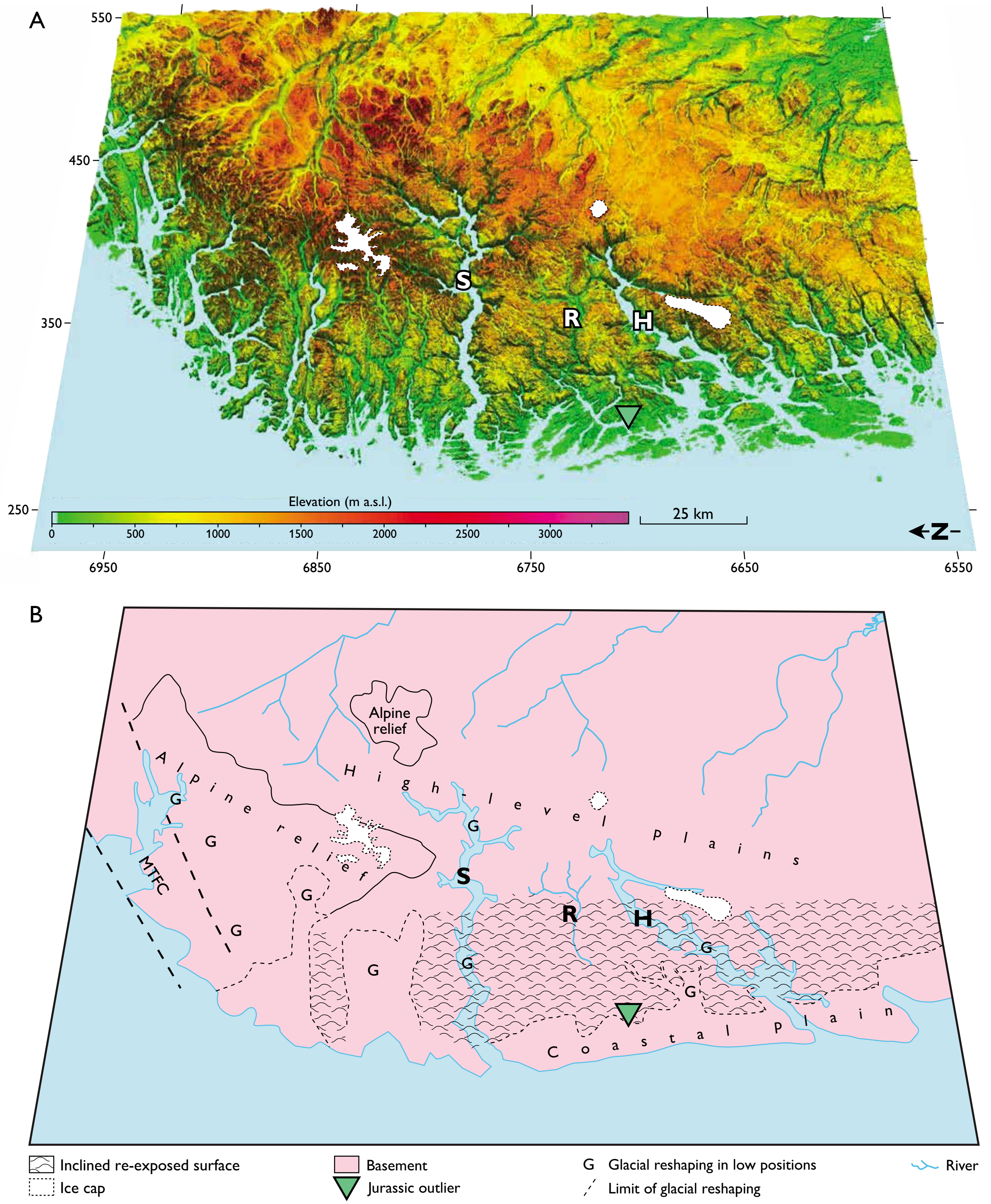
Fig. 4. Western Norway (glaciated). **A**: 3D-terrain model with vertical shading. **B**: Line drawing of the map in **A**. There are extensive, high-level plains and valleys incised below about 800 m a.s.l. and a coastal plain, the strandflat. As in parts of eastern Australia (Fig. 3) a tilted, re-exposed, and probably sub-Mesozoic plain can be traced in the summit relief from the high areas to the coast between the major incised valleys (Lidmar-Bergström *et al.* 2000). Areas with major glacial reshaping occur both in high (alpine relief) and low (**G**) positions. The Raundalen valley (**R**) has escaped major glacial reshaping, while the Sognefjord (**S**) and Hardangerfjord (**H**) have been deepened and widened by major outlet glaciers. As a consequence their inner parts have been fluvially rejuvenated during the interglacials. The Møre-Trøndelag Fault Complex (**MTFC**) is a tectonic zone along which movements occurred in the Mesozoic and maybe also later (Redfield *et al.* 2005). Jurassic outlier in the Bergen area (Fossen *et al.* 1997). Scale bar applies only to the foreground of the 3D model. UTM zone 32N. Elevation data from Jarvis *et al.* (2008).


A
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Elevation (m a.s.l.)
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1000
1500
2000
2500
3000
25 km
N
6950
6850
6750
6650
6550
B
Alpine relief
High-level plains
Alpine relief
G
MTFC
S
R
H
Coastal plain
Inclined re-exposed surface
Ice cap
Basement
Jurassic outlier
G Glacial reshaping in low positions
Limit of glacial reshaping
River

The elevation of epigene peneplains graded to former sea levels can be used as a measure of rock uplift since their formation. This notion on absolute uplift should not be confused with 'surface uplift' defined by England & Molnar (1990) as the change of mean height of an area.

2.2 Historical review of the identification and use of peneplains in landscape analysis

The fundamental importance of a base level for erosion by the fluvial system was recognised at the end of the 19th century and formulated by Davis (1899) in his classic work on 'The geographical cycle'. From observations of different landscapes the following model was constructed. A cycle started with uplift of a flat landscape, which caused incision of valleys and the formation of a youthful landscape. Further development was thought to occur by slope decline to form a landscape with hills and valleys, a mature landscape. The end result was a new peneplain with a few residual hills, called monadnocks. Penck (1924) described peneplains in stepped sequences and many authors (e.g. Reusch 1901; Ahlmann 1919, 1941; Johnson 1931) understood them as markers of former base levels, realising that they could be used to determine uplift events.

Baulig (1928) noted a re-exposed and tilted, mainly sub-Jurassic peneplain along the flanks of the Massif Central in France and a major horizontal summit surface which could be dated to the Oligocene by terrestrial deposits. He interpreted this surface as having formed after an uplift event, as it cuts off the inclined sub-Jurassic peneplain. In addition, incised younger valley generations were identified, confirming further base-level changes. At that time the continents were considered basically stable but it was realised that during certain time intervals the sea had flooded large parts of the continents. For the northern hemisphere the Late Cretaceous was such a period. Surfaces and valleys in stepped sequences were identified in several European massifs and thought to reflect a changing (lowering) sea level from the early Cenozoic and onwards (Baulig 1935). This 'eustatic' theory (not to be confused with the modern concept used in sequence stratigraphy) had an immense influence on geomorphological thinking in Central Europe and particularly in Great Britain. A major theme in geomorphological studies became the identification of as many steps as possible with the aid of topographic profiles. International commissions tried to correlate peneplains across the Atlantic (Lefèvre 1960). The efforts were not successful and the old eustatic theory fell out of favour. As a result, the concept of the peneplain in landscape analysis became widely criticised and Baulig's identification of re-exposed peneplains and their significance for extracting histories of uplift and subsidence was not appreciated at that time.

Although continental drift was not generally accepted in the early part of the 20th century, it was clear to some geologists working in southern Africa (e.g. Du Toit 1937). King (1951, 1956b, 1962, 1967) recognised major low-relief erosion surfaces, which he described as pediplains (see section 3.4) formed by valley incision to former base levels and parallel scarp retreat, as providing important constraints on intermittent uplift after continental break-up and formation of a new coastline. King's model with scarp retreat contrasts to Davis' model with slope decline, but in both models the base level for the erosion surfaces is a marker of uplift. King's ideas on development of EPCMs have had considerable influence and are therefore reviewed here.

King (1956b, 1972, 1983) concluded from analysis of landforms and geology that the landscape of Natal, southern Africa (Fig. 2), had developed through a series of uplifts by warping of the continental margin since the Jurassic, and he dated the pediplains by correlation to offshore or coastal sediments. He recognised small remnants of the pre-break-up Gondwana Surface on some high summits above the Drakensberg Escarpment. The major summit plain below this escarpment, the African Surface, he regarded as Palaeogene with a major erosional phase already in the Late Cretaceous, but finally shaped in the early Miocene. He regarded two major valley generations as evidence of Neogene uplift events. King's (1983) major message was that the present high position of the rims of southern Africa is due to Neogene uplifts after formation of the African Surface close to sea level: "They have nothing to do with the Mesozoic break-up and subsequent continental drift; they refer primarily to Cenozoic tectonic movements operating solely in the vertical sense, both up and down" (King 1983, pp. 203–204). King's view of the importance of Cenozoic tectonics seems to have been forgotten in more recent times, although his work is often acknowledged.

In contrast, Ollier (1982, 1985) described the high-level plains inland of an EPCM as *a single* ancient 'palaeoplain', with its oceanward termination and decline towards the coastal plain defining a 'Great Escarpment' (even though the oceanward slope can look very different in different places, cf. section 2.1). He regarded this pal-

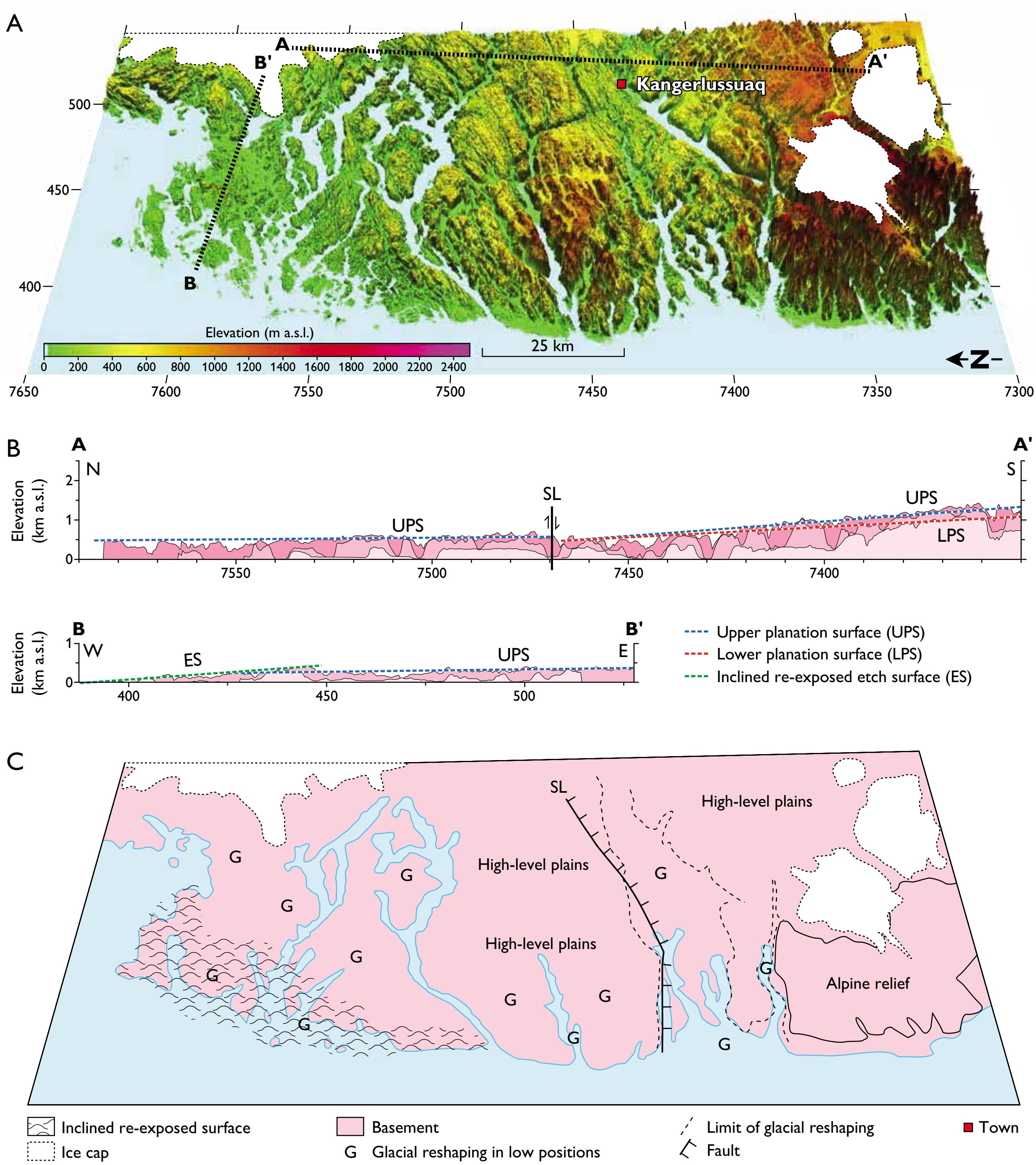


Fig. 5. West Greenland (glaciated). **A:** 3D-terrain model with vertical shading. **B:** Topographical profiles with maximum and minimum elevations within 20 km wide zones. **C**: Line drawing of the map in **A**. There are well-preserved high-level plains inland, deeply incised valleys and a coastal plain (strandflat). The high plain (Upper Planation Surface, **UPS**) is tilted and offset along faults, and a partially developed, lower plain (Lower Planation Surface, **LPS**) is developed in two areas of major uplift. An inclined, Mesozoic etch surface (**ES**) is re-exposed at the coast in the north-west, just south of the Cretaceous–Paleocene Nuussuaq Basin (see Fig. 36). Areas with major glacial reshaping (**G**) occur both in high (alpine relief) and low positions. **SL**: Sisimiut Line offsetting the UPS (see Fig. 34). Scale bar applies only to the foreground of the 3D model. UTM coordinates (km) zone 22N. Elevation data from Jarvis *et al.* (2008). See chapter 5 for uplift history.

aeoplain as an originally high plain of Mesozoic age and discussed how downwarp of this palaeoplain after break-up caused valleys to incise and form a 'Great Escarpment' along the passive margins of eastern Australia and southern Africa. Ollier & Marker (1985) pointed to the general form of high passive margins (what we now call EPCMs) and discussed their development as either uplifted Mesozoic rift shoulders or downwarped Mesozoic surfaces (Ollier & Pain 1997). Ollier and coworkers did not discuss margins in the frame of Neogene uplifts or upwarps, which had been suggested earlier, by e.g. Reusch (1901) for Scandinavia, Craft (1933) for south-east Australia, Ahlmann (1941) for East Greenland and King (1956b) for southern Africa. The conceptual models of marginal development (downwarp or rift shoulder uplift) by Ollier (1985) are of particular importance as they have been used in discussing the thermal and denudation histories of continental margins (as discussed in chapter 6).

2.3 Criticism of landscape analysis and the idea of steady state in geomorphology

Since the 1950s, when a dynamic basis for geomorphology was advocated by Strahler (1952), a controversy has existed within geomorphological circles regarding whether or not landscape generations (stepped peneplains and valley generations) graded to distinct former base levels exist. Researchers such as Chorley (1965) and Summerfield (2000) denied their existence, whereas other geomorphologists of the new dynamic school such as Schumm (1975) and Ahnert (1994) never questioned the old principles of landscape generations. Chorley's (1965) main criticism, however, was aimed at clarifying for geographical geomorphologists that denudation chronology is not a relevant issue "for the geographer with his human theme", although he admitted that landform analysis in combination with stratigraphy could give scientific results (e.g. Wooldridge & Linton 1955). Yet, his criticism of landform analysis resulted in many geomorphologists being reluctant to use peneplains and valley generations as a scientific data set to reveal uplift events.

Criticism of landscape analysis has continued within physical geography through the concept of the 'steady state' (Inkpen 2005), in which the same amount of energy is thought to enter and leave the system without causing any major changes in topography as it is assumed that uplift due to isostasy keeps pace with erosion. This view does not acknowledge the presence of inherited landforms such as high-level peneplains graded to old base levels (cf. Hack 1960), and the information that the landscape elements contribute for deciphering the sequence of events is ignored. Old textbooks based on Davisian geomorphology have been abandoned. Literature on peneplains and incised valleys and what they tell about past events came to be regarded as unscientific within geomorphology. Studies on erosion rates, based on thermochronology of the bedrock and cosmogenic isotopes produced *in situ*, have replaced landform analysis as the major theme in studies of long-term landform evolution and uplift of continental margins (e.g. Brown *et al.* 2000; Cockburn *et al.* 2000; Gleadow & Brown 2000; Summerfield 2000; Persano *et al.* 2002; van der Beek *et al.* 2002).

Relief is different in different settings and it is evident that there is information in the landscape itself that is of importance for solving tectonic problems (e.g. Gellert 1990; Burbank *et al.* 1999; Sugai & Ohmori 1999; Calvet & Gunnell 2008; Gunnell *et al.* 2009; Westaway 2009). As Bishop (2007, p. 330) states, post-orogenic, non-steady-state landscapes "deserve much more attention than they have received of late". Such landscapes are of interest not only because of the questions they pose about their formation and preservation but also because of the information they can provide on tectonic events through the presence of e.g. epigene and re-exposed peneplains and their cross-cutting relationships.

2.4 Different approaches to geomorphological studies of passive margins

The long-term geomorphological development of passive margins has been studied using a variety of different approaches. In the most common present-day approach (e.g. Bishop & Goldrick 2000; Brown *et al.* 2000; Summerfield 2000; Persano *et al.* 2005), the existence of elevated preserved peneplains graded to ancient common base levels is denied. The topography of EPCMs, described as consisting of a high-level plain with an oceanward Great Escarpment, approximately at the main divide, is interpreted as a steady state phenomenon. In this approach the existence of a high plain, originally formed prior to break-up (Ollier 1982), is accepted, but the question of how this plain was formed is not dis-

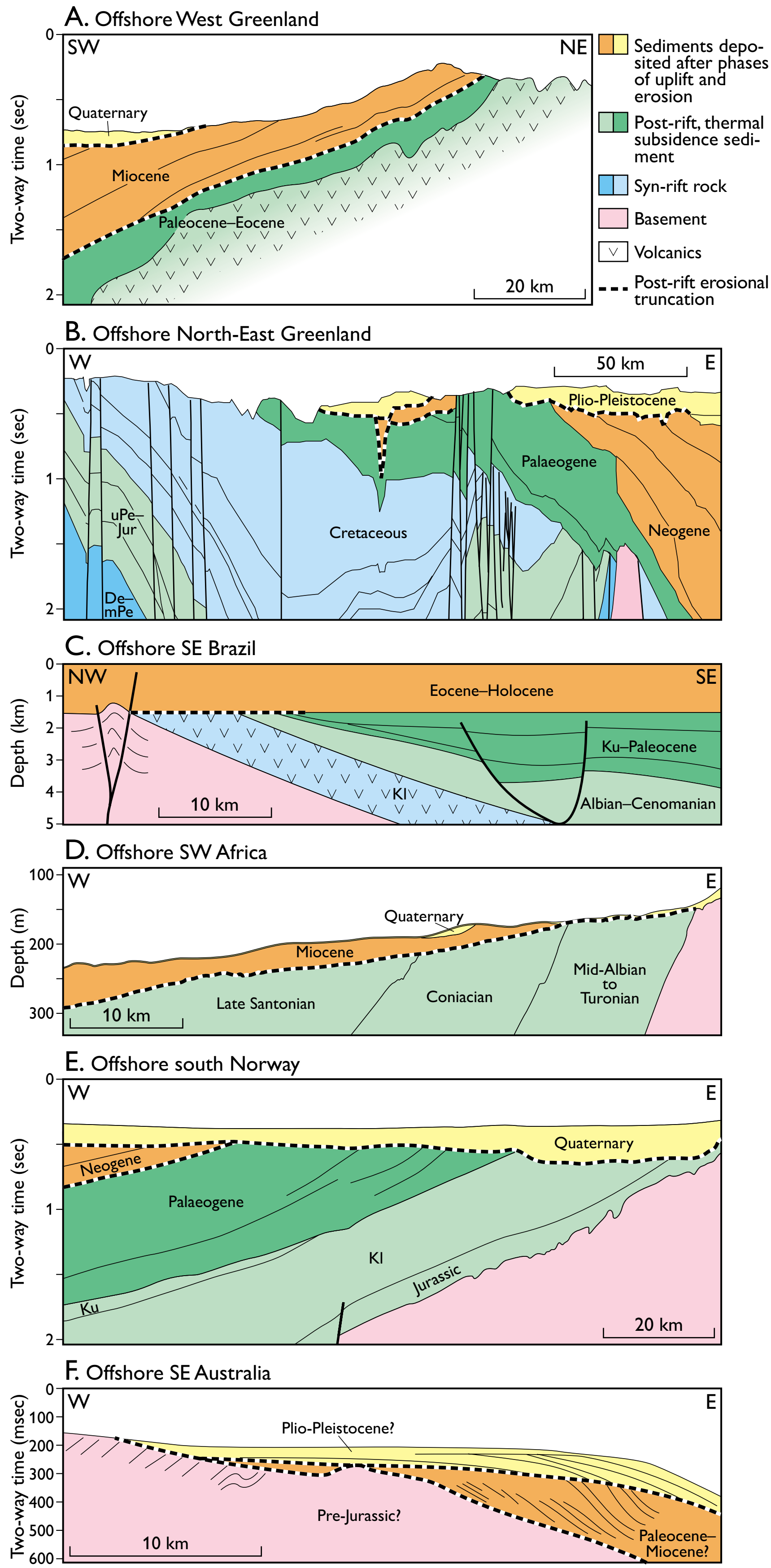


Fig. 6. Post-rift truncation of sedimentary sequences offshore EPCMs indicating post-rift subsidence and uplift of these margins. **A**: Offshore West Greenland (70°30′N); Oligocene and late Neogene truncations along low-angle and high-angle unconformities, respectively (see seismic section in Fig. 42A; redrawn from Chalmers 2000; Japsen *et al.* 2006, 2010 & unpublished AFTA data). **B**: Offshore North-East 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 south-east Brazil (26°30′S); mid-Eocene truncation (redrawn from Cobbold *et al.* 2001). **D**: Offshore south-west Africa (30°S); post-Cretaceous truncation (redrawn from Stevenson &McMillan 2004). Post-rift exhumation along south-west Africa is documented by Walford & 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). The early Neogene exhumation is documented by Japsen *et al.* (2010). **F**: Offshore south-east 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 not all truncations may have been identified on these profiles. **De–mPe**: Devonian – middle Permian. **Kl**: Lower Cretaceous. **Ku**: Upper Cretaceous. **uPe–Jur**: upper Permian–Jurassic. Modified after Japsen *et al.* (2012a).

cussed. The plateau inland of the margin is regarded as having continuously been a plain which has remained elevated since prior to rifting, and no consideration is given to the possibility that the margin may have been buried below a sedimentary cover before development of the present-day relief. The elevated topography is a basic assumption about initial conditions at the time of rifting, and is used to define a framework for calculations of denudation rates in different landscape settings (the high plain, the Great Escarpment, the coastal plain) using low temperature thermochronology.

A different approach is used by many French researchers. They use the palaeolandforms as a source of data from which conclusions on landscape development can be drawn (e.g. Peulvast *et al.* 1996) and regard the landforms *per se* as providing information about past processes (e.g. Godard *et al.* 2001a). The inversion problem connected with this approach was discussed in detail by Gunnell (1998). This field-based approach acknowledges structural and lithological control (Peulvast & Vanney 2001) and the existence of peneplains. When integrating with thermochronology data, approaches differ somewhat between researchers. While Peulvast *et al.* (2008) consider large (km-scale) thicknesses of former sedimentary cover as geologically implausible, Gunnell (1998) regards such kilometre-scale former sedimentary covers as reasonable.

The idea that high-level plains with incised valleys are evidence of late uplift refers to 'landscapes of youth' as described by Davis (1899). It was adopted by Reusch (1901), who referred to the high plateaux of Norway in terms of 'the palaeic surface', which he thought to include two levels. A similar approach was adopted by Ahlmann (1941), who used similar plateaux in east Greenland to infer Neogene uplift. In later studies, a relationship has been acknowledged between high plains and re-exposed relief on tilted peneplains between the incised valleys (Lidmar-Bergström 1982, 1988). Stratigraphic Landscape Analysis (chapter 3) was developed with the South Swedish Dome as a key area, as this dome rises only slightly above surrounding cover rocks, making the relationships between relief and cover more evident. Stratigraphic observations and cross-cutting relationships were used to construct a relative chronology for surface formation and tectonic events, both uplift and subsidence with formation of temporary covers. This method has then been applied to Sweden (Lidmar-Bergström 1995, 1996), other parts of Scandinavia (Lidmar-Bergström 1999; Lidmar-Bergström *et al.* 2000; Lidmar-Bergström & Näslund 2002; Bonow *et al.* 2003; Lidmar-Bergström *et al.* 2013) and West Greenland (Bonow 2005; Bonow *et al.* 2006a). It was integrated with thermochronology and other geological observations in further studies of the West Greenland margin (Japsen *et al.* 2006, 2009). Similar data for constraining late uplift have been presented independently in papers by authors working on uplift in Tibet (importance of former base levels: Schoenbohm *et al.* 2004; Clark *et al.* 2006), Corsica (importance of both former base levels and covers: Kuhlemann *et al.* 2005) and South America (importance of uplift and tilting for valley incision: Schildgen *et al.* 2007).

While thermochronology is used in many approaches, in recent studies of West Greenland (Japsen *et al.* 2006, 2009) it was integrated with stratigraphic landscape analysis and other geological observations to place peneplain formation and subsequent uplift within an absolute timeframe. In these studies, these three types of observations were given equal importance in deriving conclusions in regard to past events, paying particular attention to the possibility of re-burial and the influence of former sedimentary cover (subsequently eroded) in preserving old surfaces. In contrast, some other approaches have resulted in interpretations of thermochronology which are at odds with simple geological constraints, as discussed in chapter 6.

3. Stratigraphic landscape analysis (SLA): a new approach for extracting histories of denudation, subsidence and uplift

This chapter presents a new approach to the analysis of landscape, stratigraphic landscape analysis (SLA; Lidmar-Bergström *et al.* 2013). We highlight observations on the formation of peneplains close to sea level, suggest how different terms can be used to avoid confusion and describe evidence for preservation and destruction of old peneplains (palaeoplains). Subsequent chapters describe how SLA can be combined with low-temperature thermochronology (principally AFTA) to provide quantitative constraints on the timescales and magnitudes of vertical movements involved in these processes.

3.1 Observations and techniques of SLA

This new approach for landscape analysis is based on experience from the Baltic Shield and the South Swedish Dome in particular (Figs 7, 8; Lidmar-Bergström 1988; Lidmar-Bergström *et al.* 2013). Two concepts are important, viz. basement and cover rocks. Basement is defined here as denuded metamorphic and granitic rocks of different orogenies, while cover rocks are undeformed sedimentary or volcanic strata resting directly on basement. We further discriminate between orogeny, a rock-forming process, and mountain building, a process creating topography (Ollier 1981).

3.1.1 The Sub-Cambrian Peneplain

Extensive areas of the Baltic Shield consist of a flat landscape, which has been identified as the Sub-Cambrian Peneplain by the aid of remnants of cover rocks and Cambrian fissure fillings (Högbom 1910; Högbom & Ahlström 1924; Tanner 1938; Mattsson 1962; Rudberg 1960, 1970; Lidmar-Bergström 1996). It extends for 700 km along the east coast of Sweden and 450 km across the central Swedish lowlands (Fig. 7). The Sub-Cambrian Peneplain can be followed as an inclined surface in boreholes and on seismic profiles for more than 400 km and down to over 4000 m below sea level offshore south-eastern Sweden and in the Baltic countries (Kornfält & Larsson 1987; Fredén 1994). The Sub-Cambrian Peneplain continues on land as a re-exposed inclined surface for over *c.*100 km from about sea level in the east to over 300 m a.s.l. in the west and forms the crest of the South Swedish Dome (Fig. 9A; Lidmar-Bergström 1988, 1996). The surface is weathered at its contact with its cover rocks (e.g. Hadding 1929; Lundegårdh *et al.* 1973). At the contact there is shallow weathering to a maximum of 5 m (references in Elvhage & Lidmar-Bergström 1987) and the whole surface is extremely flat (Fig. 9B). Thus the Precambrian basement, which formed during different orogenies, had been denuded to an almost featureless plain before the Cambrian. The peneplain was flooded by shallow seas during the Cambrian and Ordovician, and sea level fluctuated only a few tens of metres (Artyushkov *et al.* 2000). The relationship between the surface and the Cambrian outliers on it shows that the peneplain must have been virtually horizontal and situated close to the sea level before the transgressions.

3.1.2 The South Swedish Dome

The South Swedish Dome emerges from below Cambrian cover rocks in the north (Väner Basin) and east and Mesozoic cover rocks in the south and west and is delimited to the south-west by the Sorgenfrei–Tornquist Zone (Figs 7, 8; Lidmar-Bergström 1988). The Cambrian seas flooded all of southern Sweden (Jaeger 1984), now occupied by the South Swedish Dome. As both Jurassic and Cretaceous cover rocks now rest directly on the basement in areas flanking the present dome in the southwest, it is clear that the Palaeozoic cover had been eroded here by Mesozoic time. Cambrian fissure fillings, nonetheless, confirm the former existence of a lower Palaeozoic sedimentary cover (Martinsson 1968; Samuelsson 1975). Both hilly forms of the fresh basement, with a relative relief up to 200 m, and thick (up to 60 m) remnants of kaolinitic saprolite characterise the sub-Mesozoic peneplains (Fig. 10). Analysis of the cross-cutting relationships to the Sub-Cambrian Peneplain shows that the hilly sub-Mesozoic, mainly sub-Cretaceous, relief around the

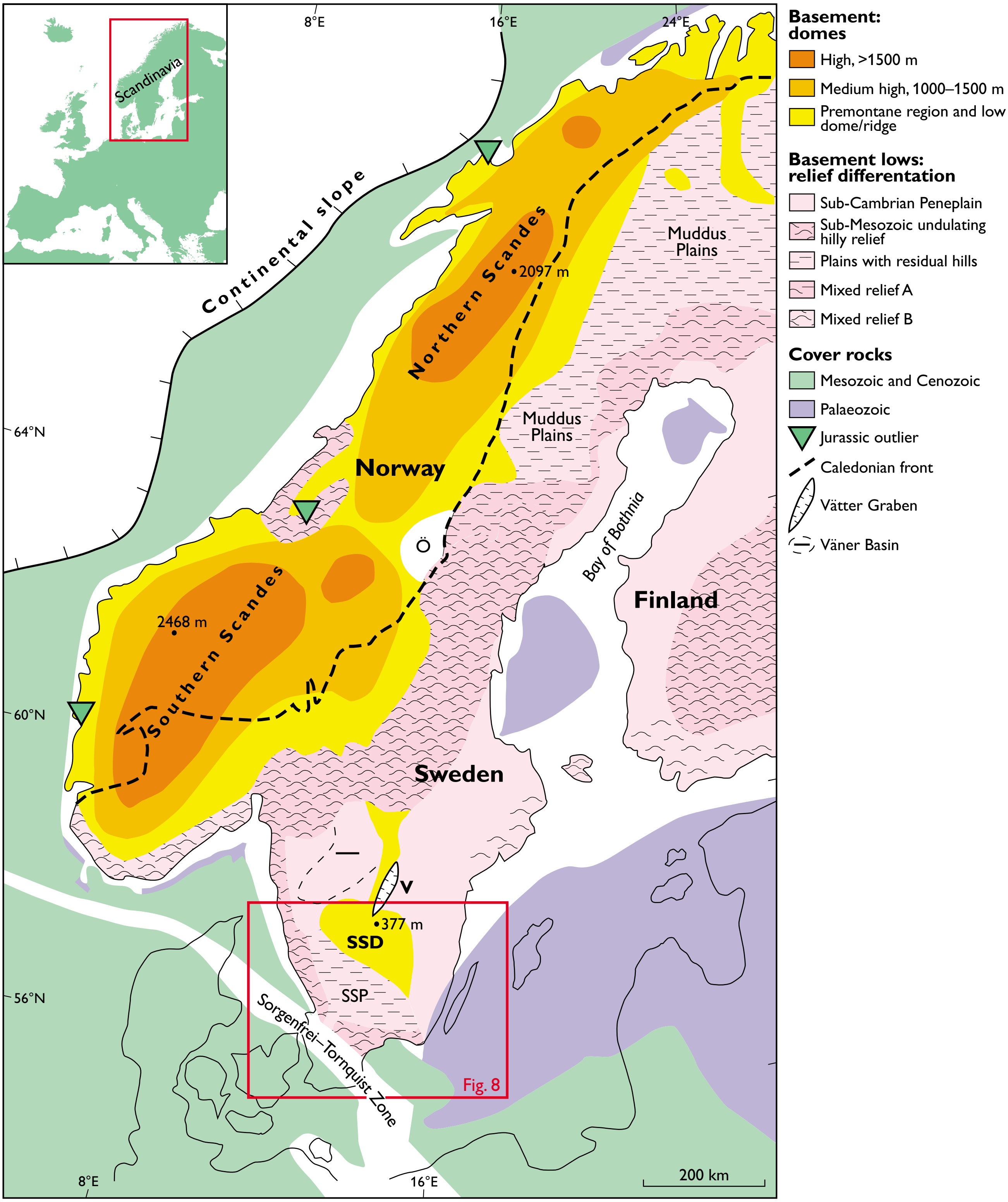


Fig. 7. Topography and relief of basement in Scandinavia in relation to the surrounding cover rocks. Three domes are identified and their highest parts marked on the map: the Northern Scandes, the Southern Scandes and the South Swedish Dome (**SSD**). Three types of low relief are identified on the flanks of the domes and further away: 1) The re-exposed flat Sub-Cambrian Peneplain, 2) the re-exposed hilly sub-Mesozoic peneplain and 3) epigene (never covered) plains with residual hills; i.e. the South Småland Peneplain (**SSP**), the Muddus Plains and the palaeic surfaces of the Southern Scandes (not marked). The re-exposed peneplains are tilted and often cut into different blocks, while the epigene peneplains are near-horizontal. Mixed relief includes stepped plains interfering with hilly relief inside the Bothnian coast in the north (Mixed relief A) and sub-Cambrian summit plateaux interfering with joint aligned valleys in south west Finland/south central Sweden (Mixed relief B). Jurassic outliers on the Norwegian margin are indicated (from south to north Bjorøy, Beitstadfjorden and Andøya; Bøe *et al.* 2010). **V**: Vätter Graben; **Ö**: Östersund low area. Modified from Lidmar-Bergström (1999) and Lidmar-Bergström *et al.* (2013).

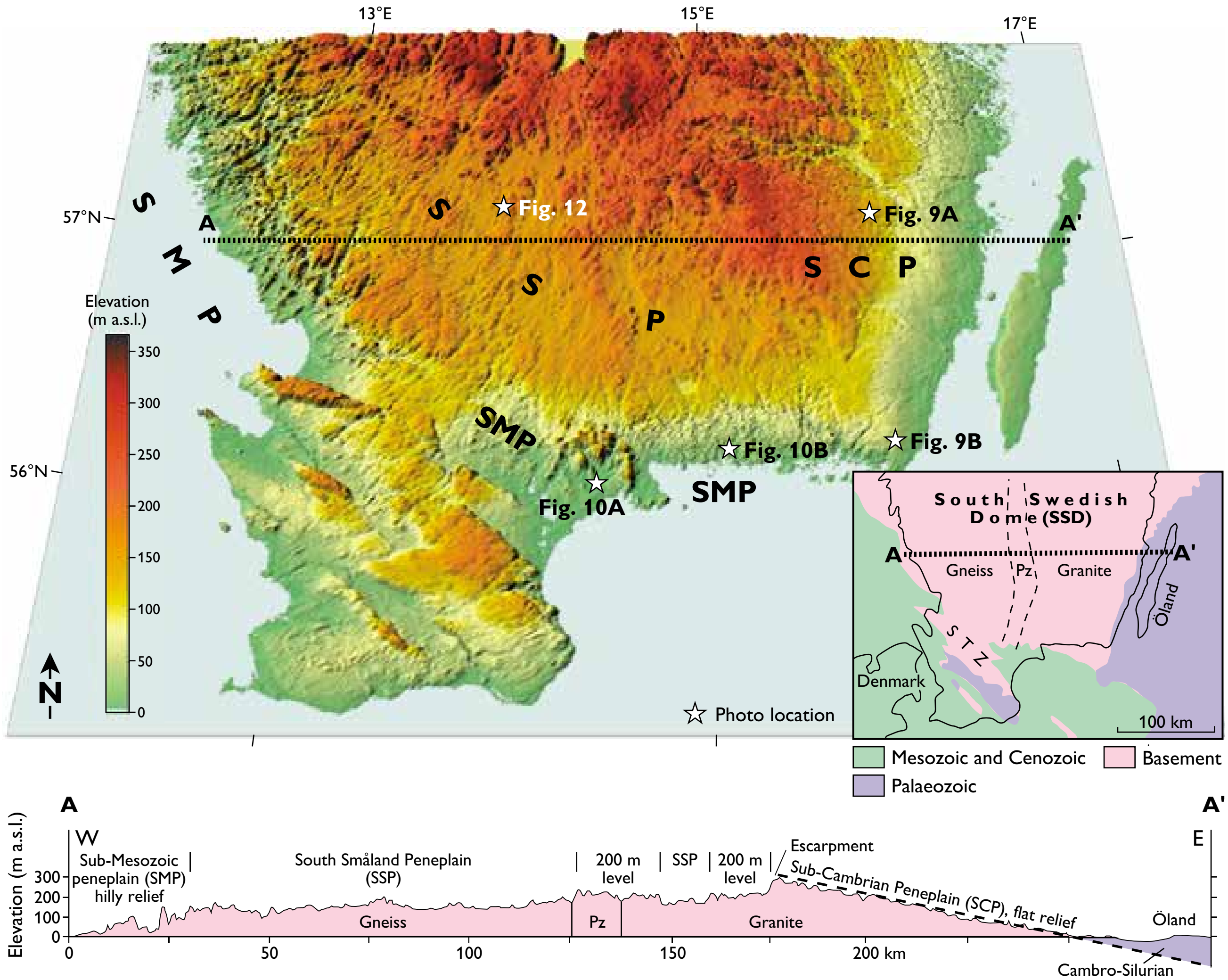


Fig. 8. Elevation of southern Sweden. The South Småland Peneplain (**SSP**), an inselberg plain (Fig. 12) at the south-western flank of the South Swedish Dome (**SSD**), cuts across gneiss in the west and granite/porphyries in the east. Note the tilted cut-off, re-exposed sub-Mesozoic peneplain (**SMP**) (hilly etch surface; Fig. 10) in the west and south and the tilted and uplifted re-exposed, flat Sub-Cambrian Peneplain (**SCP**) (Fig. 9) in the east (profile from Lidmar-Bergström 1988). **PZ**: Protogin Zone. **STZ:** Sorgenfrei–Tornquist Zone. The three different relief types are not related to differences in geology but to denudation history and can be used to reconstruct the tectonic development (Lidmar-Bergström 1993, 1994; Japsen *et al.* 2002).

Dome was cut by erosion to a new base level after uplift, tilting and removal of the Palaeozoic cover. This characteristic hilly relief continues northwards along the west coast of Sweden and continues along the south-eastern slope of the southern Scandes (Fig. 7; Lidmar-Bergström *et al.* 2013). The preservation of this surface can only be explained by protection beneath a long-lasting Late Cretaceous and younger cover directly on basement (Elvhage & Lidmar-Bergström 1987; Lidmar-Bergström 1995). The possibility of such a cover has been discussed in thermochronology studies of the area (Cederbom *et al.* 2000; Huigen & Andriessen 2004). The hilly relief is present on the southern and western flanks of the South Swedish Dome only up to about 125 m a.s.l., where it is cut off by a horizontal surface, the South Småland Peneplain (Figs 7, 8, 11). This surface is characterised by plains with scattered residual hills (Fig. 12) and remnants of gravelly saprolites up to over 10 m thick (Lidmar-Bergström *et al.* 1997).

Fig. 9. **A**: View over the inclined Sub-Cambrian Peneplain from Aboda Klint, 30 km west of remaining cover rocks. The Sub-Cambrian Peneplain is an impressive regional feature. **B**: Sub-Cambrian Peneplain close to its Cambrian cover rocks, where it is everywhere extremely flat (Fågelmara). Photo locations shown in Fig. 8. Photo: M. Rowberry, Academy of Sciences of the Czech Republic

3.1.3 Cross-cutting relationships and their implications

The cross-cutting relationships described above lead to the following insights. Regional unconformities with remnants of subaerial weathering provide evidence of peneplain formation in the geological past. Saprolites and relief (flat, hilly, etc.) are different at the contact with different cover rocks. The crosscutting relationships between peneplains of different tilts and with different relief and saprolite give information about the relative denudation chronology. A re-exposed peneplain in basement rocks can be identified at its contact with the cover rocks. Its characteristic topography, saprolite remnants (Lidmar-Bergström *et al.* 1997; Bonow 2005) and inclination can be followed from below its cover. Where the land surface changes character (relief and saprolites) and inclination, this indicates that a former continuation of the re-exposed surface has been erased by later erosion (Fig. 11). The landforms at the south-western border of the South Swedish Dome (Fig. 11) show how such relationships can be used for extracting the geomorphological development in three stages:

1. The Sub-Cambrian Peneplain was re-exposed to denudation along the south-western part of the South Swedish Dome in the Mesozoic. Thus a new surface formed and became graded to sea level by the fluvial system. Due to a warm and humid climate an irregular relief with thick kaolinitic saprolites along fracture zones formed (Fig. 10). This deeply weathered new peneplain was successively tilted, partly stripped of its saprolite and subsequently covered by sediments during the late Mesozoic.

2. After a transition from subsidence to uplift, the South Swedish Dome rose above sea level and a younger, epigene, post-Cretaceous, sub-horizontal surface (the South Småland Peneplain) was formed by erosion cutting across both basement of different lithologies (Stephens *et al.* 1994) and cover rocks,

Fig. 10. The Precambrian basement in south Sweden is weathered along zones of variable width to a kaolinitic saprolite, up to 60 m thick, below or in close connection to preserved Upper Cretaceous strata. **A**: In a quarry on the island of Ivön, partial evacuation of the saprolite has exposed the steep weathering front. The irregular relief with steep-sided hills has formed by deep weathering in the Mesozoic and subsequent stripping of the saprolite. Note person for scale (red circle). **B**: Typical steep-sided hill of the sub-Cretaceous hilly relief at Dalhejaberg with kaolinitic saprolite remnants in fractures. Note the complete contrast to the landscape of the extremely flat Sub-Cambrian Peneplain. Photo location shown in Fig. 8.

guided by a new distinct base level (Fig. 11B). This base level was unrelated to resistant rock types. As the Cenozoic record in Denmark (Rasmussen *et al.* 2008) shows that the area was close to the sea in the west, sea level provides the only feasible base level. Erosion resulting in formation of this younger surface has thus removed both the cover and basement rocks above its present level. The new plain is sub-horizontal, with only a few residual hills, and exhibits gravelly saprolites. Such saprolites started to form in the Miocene (Migoń & Lidmar-Bergström 2001) providing support for the argument that the plain formed during the Neogene.

3. Uplift/lowering of base level (here only about 100 m) has caused re-exposure of the sub-Cretaceous hilly relief with associated clayey saprolites at lower elevations to the south and west of the dome. The lowering of the base level is of a late age, as the hilly relief and often also the associated kaolinitic clayey saprolites remain on the tilted sub-Cretaceous surface. The present elevation of the sub-horizontal South Småland Peneplain indicates uplift relative to the present-day sea level.

3.1.4 Peneplains as unconformities in other areas and their crosscutting relationships

Peneplains as unconformities in the stratigraphic record, such as the sub-Cambrian and sub-Mesozoic surfaces of the Baltic Shield, are of common occurrence in many regions of the Earth. For instance in Germany the peneplanation of the Variscan Mountains was terminated with the formation of 'der Permische Rumpf' (the Permian base levelled plain), on which the upper Permian and Mesozoic sediments were deposited (Gellert 1958). The old peneplain is re-exposed where the cover sediments have been removed by erosion. After uplift, tilting and re-exposure of the surface, younger, more horizontal peneplains have formed and are covered for instance by Palaeogene sedimentary strata (Baulig 1928; Büdel 1977; Demoulin 2003). In Canada sub-Ordovician flat surfaces are an important component of the present relief (Ambrose 1964; Lidmar-Bergström & Jansson 2005).

3.1.5 Sea level, a major base level

An extensive, low-relief erosion surface which cuts across rocks of different resistance provides evidence that its formation was governed by a particular base level for an extended time. The above examples of unconformities developed over large areas and subsequently covered by marine deposits demonstrate that the sea level acted as base level for the formation of extensive peneplains. It is not possible to argue that such low-relief unconformities which have been covered by marine strata, such as the sub-Cambrian flat peneplain or the sub-Cretaceous hilly peneplain, were formed at high elevations (Fig. 11), as they were both progressively transgressed by the sea.

The near-horizontal South Småland Peneplain on the south-western flank of the South Swedish Dome cuts off a re-exposed, inclined, sub-Cretaceous surface. This latter surface would not have retained its characteristic relief and saprolites without a protective cover. Thus the South Småland Peneplain cannot have formed at its present elevation but is part of a larger surface that originally formed across both basement (the present South Småland Peneplain) and Cretaceous cover rocks (protecting the hilly relief) guided by a common base level (Fig. 11). The South Småland Peneplain must have been uplifted in relation to the present base level. If continuously exposed, the relief at low elevations would have been subjected to the same weathering and erosion as at higher elevations and, in accordance with the prevailing climate, only gravelly saprolites would have formed. Instead, hilly relief and kaolinisation at low elevations are observed in the west and south, where the relief disappears below a Mesozoic cover. Where the relief disappears below a Cambrian cover in eastern Sweden, the bedrock surface is flat. A low base level would have caused erosion of the re-exposed surfaces and erased them.

Where a study area is known to have been close to the sea during formation of a peneplain, as in south Sweden (section 3.1.3) or in the case of the post-rift development of margins adjacent to opening oceans, the most likely base level is sea level (see Japsen *et al*. 2009). Thus the importance of sea level as a general base level, as advocated in the early studies (section 2.1; e.g. Davis 1899; King 1962), remains valid.

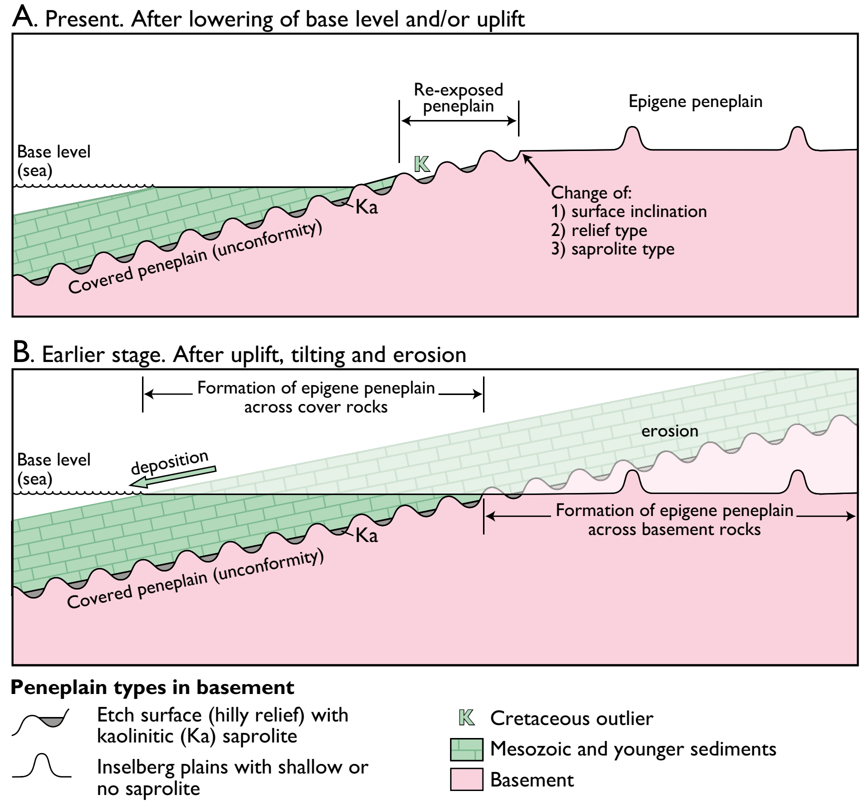
3.1.6 Other implications of re-exposed peneplains

The once-horizontal, Sub-Cambrian Peneplain of the South Swedish Dome has been uplifted, tilted and partly re-exposed long after its formation. It has been incised by valleys and is now being destroyed (Fig. 8). This information leads to the conclusion that uplift leads to incision. Peneplains must also have been close to horizontal when they were being formed, as tilted plains start to be dissected at a slope of 0.4–0.5% (Rudberg 1970; Spönemann 1979) and thus cannot survive for long after tilting. Thus, they are transient features (Bishop 2007). Re-exposed peneplains commonly occur as tilted surfaces at low elevations, but at higher elevations such surfaces are overprinted by younger peneplains (Fig. 11; Garzon *et al*. 1982; Lidmar-Bergström 1988, 1996; Jutras & Schroeder 1999; Jutras & Prichonnet 2004; Peulvast & Claudino-Sales 2004; Bonow *et al*. 2006a; Peulvast *et al*. 2008). Well-preserved re-exposed peneplains are evidence of long-lasting covers. Re-exposed peneplains define both periods of uplift/exposure and periods of subsidence/covering.

3.1.7 Peneplain formation – discussion

Erosion by coastal waves can be responsible for the final modification of low-relief erosion surfaces during major transgressions (e.g. King 1963), but this cannot be the

Fig. 11. Cross-cutting relationships between peneplains. **A**: The relationship between a tilted, re-exposed peneplain and a subhorizontal epigene peneplain, each with its own characteristic relief and saprolite, determines their relative ages. The subhorizontal peneplain, cutting off the tilted peneplain, is the youngest. This case is shown at the western border of the South Swedish Dome, where a Mesozoic peneplain of etch character is cut off by a Cenozoic peneplain with scattered inselbergs (Lidmar-Bergström 1988). **B**: The landform record can be used to extract landscape history in a tectonic/eustatic context. In an earlier stage, before lowering of the base level and/or uplift, the epigene peneplain formed across both basement and cover rocks by grading to a former base level. Formation of the epigene peneplain has required erosion first of the former cover and second of basement rocks. This stage must have been preceded by uplift and tilting (see sections 3.1.3, 3.1.5).



cause of denudation of hundreds of metres or kilometres of rocks across areas of regional extent (Adams 1975). The re-exposed peneplains also reveal that their final shape depends on the environmental conditions during their formation. The south Sweden example shows this well with the different forms and saprolites of the three distinct surfaces: the sub-Cambrian flat peneplain with shallow (maximum 5 m) kaolinitic saprolites, the hilly sub-Mesozoic etch surfaces with thick (60 m) kaolinitic saprolites, and the South Småland Peneplain with scattered residual hills and gravelly saprolites of moderate thickness (up to about 10 m). Weathering processes in combination with slope processes and surface wash tend to flatten hilly relief in arid to semiarid climates in both hot and cold environments (section 3.2.2), but cannot provide a specific base level for a plain of regional extent.

The work of Brozovic *et al.* (1997) has been used in recent papers to argue for peneplain formation at high elevations (Egholm *et al.* 2009; Steer *et al.* 2012). We note that Brozovic *et al.* (1997) speculated (their wording) about a landscape between 4000 and 5000 m in the north-western Himalaya with undissected plateaux, partially dissected plateaux or deeply dissected mountainous regions with comparatively gentle slopes (25 degrees), which *might* (our italics) be interpreted as an erosion surface. However, the relief described by Brozovic *et al.* (1997) does not mimic what we describe as a peneplain, which, by definition, cannot have a relief of 1000 m. Brozovic *et al.* (1997) further speculate that this region at elevations between 4000 and 5000 m is affected by both glacial scour and vigorous freeze-thaw action. This is probable, but the area is not a regional erosion surface or peneplain.

Mitchell & Montgomery (2006) showed that the *distinct peak accordance* with a summit plane cannot *a priori* be regarded as remnants of a former peneplain. They used the 'glacial- buzzsaw' mechanism, proposed by Meigs & Sauber (2000) to show that mountains in the Cascade Range, western USA, have been cut off at a certain elevation averaging 380 m and maximally 600 m higher than the equilibrium-line altitude (ELA), visualised as an inclined plane defined as the floors of 373 cirques in areas exposed to abundant precipitation. On the other hand, the peneplains described here from southern Sweden are not dissected summit surfaces but coherent plains that require another mechanism than the 'glacial buzzsaw' for their formation. Mitchell & Montgomery (2006) state that the 'glacial-buzzsaw' mechanism includes three

Fig. 12. Plain with residual hill on the South Småland Peneplain, the end result of Cenozoic landscape formation. Photo location shown in Fig. 8.

distinctive characteristics in relation to the Quaternary ELA: 1) a line through the peaks is never located more than 600 m above the surface defined from the bases of the cirques; 2) the amount of topography declines above the ELA and 3) the valley sides above the ELA are close to threshold steepness. Thus there is no plain at all above the ELA but a valley landscape with steep slopes with a relative relief of up to 600 m (Fig. 13A).

In the original paper on the 'glacial buzzsaw', Meigs & Sauber (2000) state that the mean height of topography and the mean height of the ELA are coupled. However, this does not mean that there are low relief surfaces similar to peneplains. In the western areas studied by Meigs & Sauber (2000), for example, the land lies in an altitude band between 300 and 2200 m with the highest summit at 6050 m. Thus there is relative relief of at least 3850 m. Despite this, Egholm *et al.* (2009) suggested that the equilibrium-line altitude (ELA) that forms the base level for enhanced glacial erosion of glaciers and cirques can lead to the formation of flattish surfaces, which grow together and form extensive low-relief surfaces at a distinct level. This is not in accordance with geomorphological studies, which instead show that glacial erosion amplifies the relief (Sugden 1978; Lidmar-Bergström 1997; Johansson *et al.* 2001b; Kessler *et al.* 2008) or dissects earlier formed surfaces (Oskin and Burbank 2005; Etzelmüller 2007). It is certainly not valid as a mechanism for the formation of low-relief surfaces of regional extent, and a flattish surface of minor extent is not a peneplain. Egholm *et al.*'s (2009) hypothesis cannot be applied to all EPCMs, since many of them (e.g. in eastern Australia, southern Africa and eastern Brazil) have not been significantly modified by glaciation since Palaeozoic times.

Steer *et al.* (2012, p. 1) studied low-relief surfaces at different elevations in the Sognefjord catchment area of Norway. They argued that these surfaces were shaped by glacial erosion at ELA during different times and wrote: "These surfaces have been attributed to glacial headward erosion in Alpine settings". However, the notion of an alpine setting refers to mountainous areas with glacial forms such as cirques and glacial valleys and not to peneplains. Steer *et al.* (2012) appear to suggest that low-relief surfaces can be formed by glacial headward erosion by cirque retreat as they refer to Oskin & Burbank (2005), but the process they describe simply amplifies the relief (see above), and it has not been shown anywhere that glacial valleys coalesce to produce low-relief erosion surfaces. This type of erosion cannot form the type of regional peneplains that characterise EPCMs all over the world, glaciated or not. In addition, the main distribution of high-level peneplains in southern Norway does not occur on its western and north-western side with its alpine relief (Fig. 4), which developed where precipitation was highest and the frequency of cirques largest. Instead, the best developed high-level plains with residual massifs occur in the east, least affected by glacial erosion (Sollid & Sørbel 1994; Kleman *et al.* 2008). Further evidence that glacial erosion acts to dissect plateaux rather than create them in western Norway was recently published by Hall *et al.* (2013).

It should also be noted that the regional and tilted Upper Planation Surface (UPS) in West Greenland, which covers a region more than 300 km long and 145 km wide between the coast and the Greenland ice sheet and reaches heights over 1500 m a.s.l. in the south-west and 200-500 m a.s.l. in the north and east (Bonow *et al.*

2006a), cannot be correlated with any ELA (see section 5.3 for further discussion on this topic).

Steer *et al*. (2012) also referred to Hales & Roering (2009) who described frost processes at high elevation in New Zealand that prevent peaks rising above 2300 m. Hales & Roering (2009) did not, however, describe the formation of peneplains at high elevation but instead suggested that frost processes cause rock falls, in turn causing the summits to be at the same elevation despite continuous uplift.

High plains, in general, are coherent features. They are best preserved in non-glaciated areas where they cannot have been shaped by glacial processes. Peneplains at high elevations can still be recognised even on glaciated margins (Fig. 13; Bonow *et al*. 2003, 2006a, b), despite glacial reshaping. No credible mechanism has as yet been suggested to produce regional peneplains at high elevations, unless controlled by a structural/lithological base level, but their formation close to sea level is demonstrated by both re-exposed peneplains and their cross-cutting relationships to younger peneplains in positions close to former seas, as has been described in section 3.1.5 and preceding sections.

In conclusion: Low-relief erosion surfaces extending over hundreds of kilometres, identified as re-exposed peneplains, provide evidence of peneplain formation close to sea level and are also important indicators of former covers. Their crosscutting relationships to higher surfaces also confirm the sea as base level for these surfaces that are now at higher elevations.

3.2 Terminology

3.2.1 Landscape, relief and topography

Both the words 'landscape' and 'relief' are used to describe landscapes with slopes of different character, landforms characteristic of different processes in different environments and regions with different relative heights. In SLA we use 'relief' to mean the relative variation in height of the landscape, and the word 'topography' to mean the absolute heights of the landscape above sea level.

3.2.2 Peneplanation or pediplanation

The development of low-relief erosion surfaces graded to distinct base levels by fluvial incision and slope decline/retreat, is fundamental in the landscape models of both Davis (1899) and King (1962; 1967). The main difference between them is how they defined slope behaviour. While Davis based his model on slope decline, King regarded parallel scarp retreat along incised valleys as the fundamental process (Fig. 14). King labelled the resulting low-relief erosion surface a 'pediplain'. The exact behaviour of slope development is now known to depend on details in lithology, structure and fracture systems in combination with different climates (Moon & Selby 1983; Gunnell 1998), and deep weathering plays an important role in the development of slopes between different surfaces (Peulvast 1987; Gunnell 1998). An example of the formation of a minor scarp is where the South Småland Peneplain of south Sweden cuts back into the uplifted Sub-Cambrian Peneplain (Fig. 8; Lidmar-Bergström 1988; Olvmo *et al*. 2005). Resistant rocks can give rise to prominent escarpments such as the Drakensberg Escarpment (Fig. 2; Gunnell 1998; Fjellanger & Etzelmüller 2003), because they cause backward erosion to slow down (Fig. 14). River gaps can form across resistant rocks by superimposition of drainage. The related surface will continue to form behind the obstacle subsequent to breakthrough, and surface development will be more rapid in the less resistant rock. The nature of the slope between two surfaces, either an escarpment or gradual slope, is a function of rock resistance and climate.

In conclusion: 'Peneplain' should be used as the general term for extensive, low-relief erosion surfaces, hilly or flat, graded to a distinct base level. For the purpose of using peneplains as markers of uplift the exact nature of slope development is not critical. The focus must instead be on the level of the erosional base for the peneplain.

3.2.3 Pediplains (rock-cut plains), pediments, etched surfaces (hilly relief), inselberg plains and etchplains

The term 'pediplain' is also used as a descriptive term, in the sense of a rock-cut plain (Büdel 1970). Flat, minor rock-cut surfaces with shallow or no saprolites are often termed pediments (e.g. Dohrenwend 1994; Vincent & Sadah 1996). They have low slope angles, <6°–11°, ac-

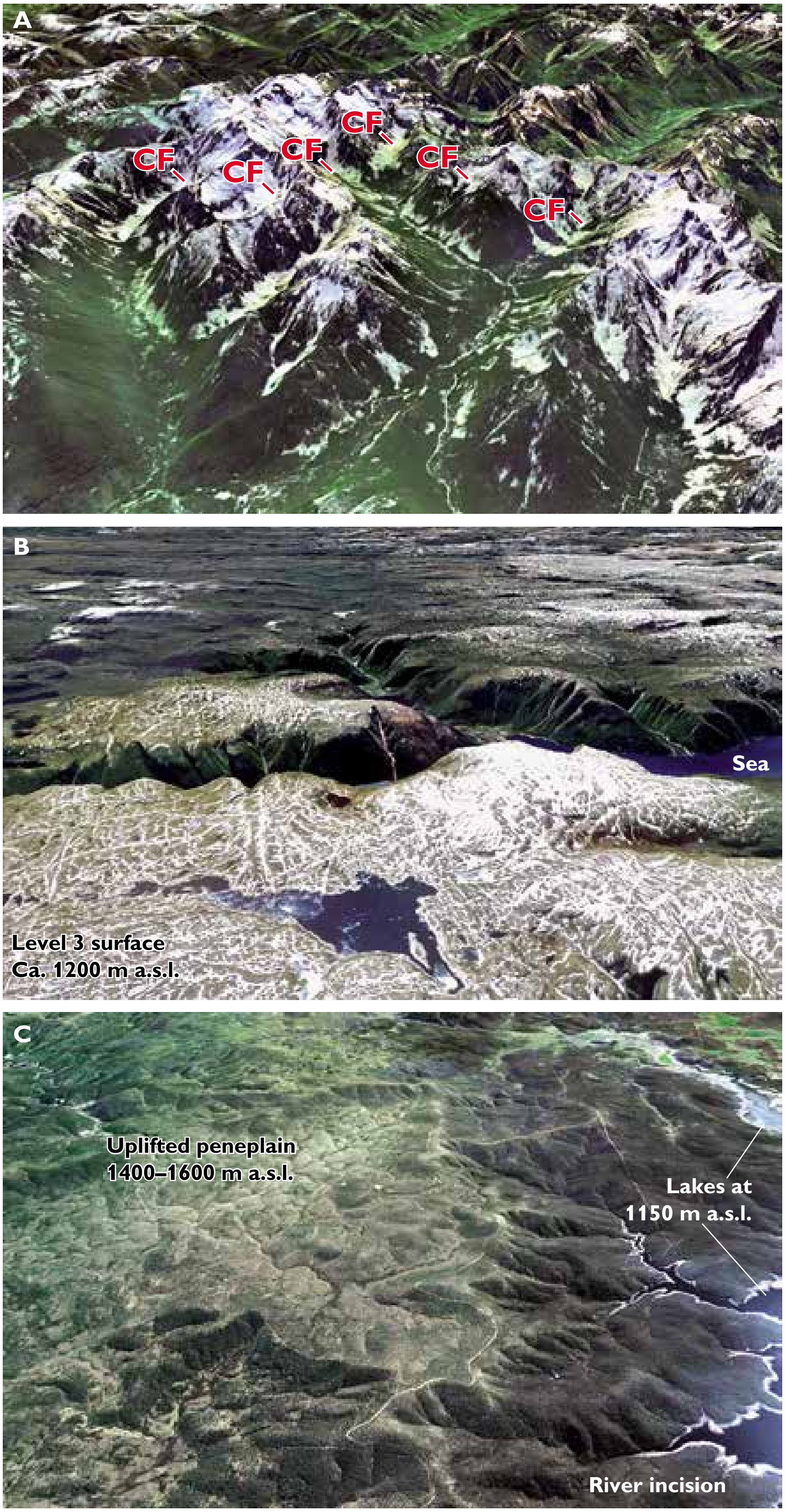


Fig. 13. Comparison of a 'glacial-buzzsaw' landscape with landscapes typical of elevated, passive continental margins (EPCMs). 3D pictures of different mountainous landscapes taken from Google Earth (no vertical exaggeration). All three pictures were taken from the same virtual height of c. 6.8 km. **A**: Cascade Mountains, USA. Typical mountain range affected by the 'glacial buzzsaw' with cirque floors (**CF**) that constitute a tilted planar zone at the equilibrium-line altitude (ELA; Mitchell & Montgomery 2006). Above the cirque floors is an alpine landscape with a relief between 380 and 600 m, but there is no plain despite the claims of Egholm *et al*. (2009). This type of alpine landscape contrasts dramatically with the high-level EPCM peneplains shown in **B** and **C**. **B**: Norwegian EPCM. Hardangervidda, looking approximately south. An elevated peneplain (*c*. 1200 m a.s.l.) is Level 3 in the classification of Lidmar-Bergström *et al*. (2000). In contrast to the 'glacial- buzzsaw' situation in **A**, the valleys to the right (west) have been incised below the peneplain. Glacial erosion has deepened their floors to near or below present-day sea level, because the Pleistocene ELA in Norway was below sea level. **C**: Eastern Australian EPCM. Plateau in Kosciusko National Park, looking slightly east of north. Unglaciated EPCM with an uplifted peneplain to the left (west); 1400–1600 m a.s.l. Fluvial valleys (now flooded by a hydroelectric lake) are eroding into the peneplain from the right (east).

cording to Dohrenwend (1994) or <6.5° according to Bonow *et al.* (2003). The higher figure seems to denote the slope in direct contact with residual hills. Pediments can grow large, forming *pediplains* by scarp retreat in arid climates around residual hills, exposed after erosion of an originally deep saprolite and resulting in elimination of all but a few hills (Thomas 1974; Demangeot 1976; Lidmar-Bergström 1982, 1995) or as pedimentation along the sides of valleys and eventual removal of intervening ridges (Ahnert 1998). Pedimented surfaces evolve in arid and semiarid zones by slope retreat and surface wash (Büdel 1970; Mensching 1970; Young 1972). The Sub-Cambrian Peneplain in Fennoscandia is extremely flat, only a little weathered (see above) and with very few residual hills. It was formed at a time without any vegetation cover, when continuous surface wash therefore contributed to its final shaping as a pediplain (Lidmar-Bergström 1988). Both shallow weathering and surface wash are ingredients in pediplain formation.

On the other hand, field observations in the semihumid tropics have shown that differential weathering by long periods of deep weathering with formation of deep saprolites (up to over 100 m) alternating with periods of stripping of the saprolite, followed by renewed deep weathering (Thomas 1965, 1966) result in a highly irregular weathering front. After final exposure of the weathering front this gives rise to more or less closely spaced hills, depending on fracture density. In Fennoscandia and Greenland, Lidmar-Bergström (1982, 1989, 1995, 1999) and Bonow (2005) have shown that *hilly relief* (irregular bedrock surfaces with hills up to 200 m and more) was caused by periods of kaolinitic weathering (etching) and stripping in subtropical climates in the Mesozoic (Fig. 10). Such surfaces are only preserved where they have been protected until recently by Mesozoic–Palaeogene covers (Fig. 11). Re-exposed hilly relief is often reinforced by glacial erosion in the formerly glaciated areas (Johansson *et al.* 2001a; Migoń & Lidmar-Bergström 2001; Bonow 2005), while it can be partly hidden below thick saprolites in tropical and subtropical areas.

The end product of relief formation in Greenland and Scandinavia during the Cenozoic is plains with residual hills with thin, gravelly, saprolites (Lidmar-Bergström 1982; Elvhage & Lidmar-Bergström 1987 1995; Lidmar-Bergström *et al.* 1997; Bonow *et al.* 2006a, b; Lidmar-Bergström *et al.* 2007). In contrast to the Sub-Cambrian Peneplain, which is almost devoid of residual hills, and to the hilly sub-Mesozoic etched surfaces, the Cenozoic plains are characterised by isolated residual hills (inselbergs; Fig. 12) and are termed *inselberg plains* (Rudberg 1960, 1988; Ebert *et al.* 2012). The observed cross-cutting relationships of peneplains of different shape suggest that the inselberg plains in the tropics are produced in a similar way, rather than by deep weathering to an even, horizontal weathering front as is often suggested (e.g. Moore *et al.* 2009). The irregular weathering guided by joints and fractures instead seems to have originally caused hilly relief (see above) from which the present rock-cut plains with only a few inselbergs have developed (Lidmar-Bergström 1995).

Etch planation is not regarded as a general process for the formation of low-relief surfaces unrelated to base level as suggested by Thomas (1994). Instead we define an *etch plain*, as a plain underlain by a deep saprolite but which also formed in relation to a distinct base level.

In conclusion: Pediplains, inselberg plains, etched surfaces (hilly relief) and etch plains are the result of different climatic conditions during the final stage of peneplain development. However, the formation of the peneplain as such, which involves the kilometre-scale erosion of rock during valley incision and valley widening by running water and slope processes, is guided by a stable base level, often the sea (Davis 1899).

3.2.4 Classification of peneplains

The notion 'planation surface' objectively describes a flat landscape form (Adams 1975). Not all peneplains are planation surfaces. A hilly relief surface caused by irregular deep weathering and belonging to an identifiable plain governed by a base level is a peneplain. There is a plethora of different terms in common use and much confusion surrounds the terminology (Ebert 2009). A classification is listed in Table 1. A peneplain is a low-relief erosional surface graded to a distinct base level. It can be either underlain by a deep saprolite (etch plain) or cut across fresh bedrock, almost without inselbergs (pediplain) or with scattered inselbergs (inselberg plain). Etched bedrock surfaces can be more or less stripped giving rise to a hilly relief, particularly common in formerly glaciated basement terrain where stripping has been efficient. Empirical figures for relative relief are available from glaciated Fennoscandia (Lidmar-Bergström 1995); pediplains <20 m, inselberg plains *c.* 20–200 m (plain between scattered hills <20 m) and hilly relief *c.* 20–200 m.

Table 1. Classification of peneplains§ in basement terrain

Planation surface	Pediplain Inselberg plain Etchplain	Relative relief <20 m† Relative relief *c.* 20–200 m	Rock-cut, flat surface Surface with scattered residual hills Flat surface underlain by deep saprolite
Hilly relief	Etched hilly relief	Relative relief *c.* 20–200 m after total stripping in glaciated terrain	Irregular bedrock surface with closely spaced hills caused by differential deep weathering and stripping of saprolite

§ Peneplains: low-relief erosional surfaces graded to a distinct base level.
† Figures for relative relief from Rudberg (1960).

3.2.5 High-level benchlands (stepped peneplains), incised valleys and the Great Escarpment

There is a clear difference between *pedimented surfaces* with low slope angles (section 3.2.2) and *steeper slopes* in the landscape. Many areas with high-level plains and residual mountain massifs are characterised by benchlands of pediment character flanking the individual massifs with intervening steeper slopes. Besides along some continental margins as e.g. in southern Africa (Fig. 2), such benchlands characterise e.g. Kenya (Ahnert 1982), Spain (Casa-Sainz & Cortès-Gracia 2002) and southern Norway (Bonow *et al.* 2003). The benches sometimes occur along major river valleys and may indicate former base levels for the fluvial systems and related peneplains (Bonow *et al.* 2003). The valleys may have coalesced to partial peneplains which, after step-wise lowering of base level, now fringe the massifs as benchlands.

There is a major difference between high-level peneplains/benchlands with their gentle relief down to a certain base level plain and valleys deeply incised from this lowest plain. The zone with incised valleys along EPCMs is often named the Great Escarpment. Uplifted peneplains with benchlands are called *palaeic* surfaces in Norway (Reusch 1901) and *relict* surfaces in southeastern Tibet (Schoenbohm *et al.* 2004; Clark *et al.* 2006). The once horizontal, Sub-Cambrian Peneplain of the South Swedish Dome in south Sweden is uplifted, tilted, re-exposed and incised by valleys (section 3.1.6), which shows that uplift causes incision. The main base for the lowest high planation surface, from which deep valleys are incised, is regarded as a major marker for calculations of amounts of late uplift (Bonow *et al.* 2006a, b). The cross-cutting relationships of such a surface to a tilted re-exposed peneplain can attest this conclusion (section 3.1.3).

Major knickpoints in river profiles (Fig. 14) might be markers of former base levels (Schoenbom *et al.* 2004; Crosby & Whipple 2006; Bridgland & Westaway 2008). Steep channel profiles incised along the front of the Tibetan Plateau have led to the conclusion that such steep profiles can reflect active uplift of the plateau (Kirby *et al.* 2003). In glaciated areas only knickpoints reflecting incision in major peneplains can be used for deciphering uplift events, while knickpoints in incised rivers might just reflect uneven glacial erosion. Knickpoints can also form in tributary valleys due to glacial deepening of major valleys (Kleman & Stroeven 1997; Bonow *et al.* 2003) and are not usable as uplift markers in such settings.

3.3 Palaeoplains, preservation, destruction and age

When a peneplain is buried beneath a sedimentary or volcanic cover, or loses its direct contact with its base level due to uplift, it is a palaeosurface (Bonow *et al.* 2006a) and can be labelled a *palaeoplain*. Cover rocks can preserve palaeoplains in the form of unconformities for long periods of time. Epigene peneplains that have lost contact with their general base level continue to grade to their own uplifted base level and saprolites can grow (e.g. Carmo & Vasconcelos 2004). Alternatively, the high surfaces can be fossilised by extreme aridity (Migoń & Goudie 2001) or below cold-based ice frozen to the ground (Sugden 1968; Dyke 1993; Kleman 1994). Areas that have experienced continued extreme aridity such as Namibia and the Dry Valleys of Antarctica (Cockburn *et al.* 1999; Summerfield *et al.* 1999) might contain very old features, which make it possible to reveal a history comparatively far back in time. Different types of duricrusts, such as ferricrete, silcrete and calcrete, can form and will then make weathered surfaces resistant to destruction.

Once a peneplain is formed and uplifted (a palaeoplain), it will survive much longer in resistant rock than in more easily eroded rocks, where a new generation of valleys will first be formed (Gunnell 1998; Bonow *et al.* 2003, 2006a; Fjellanger & Sørbel 2007). Narrow

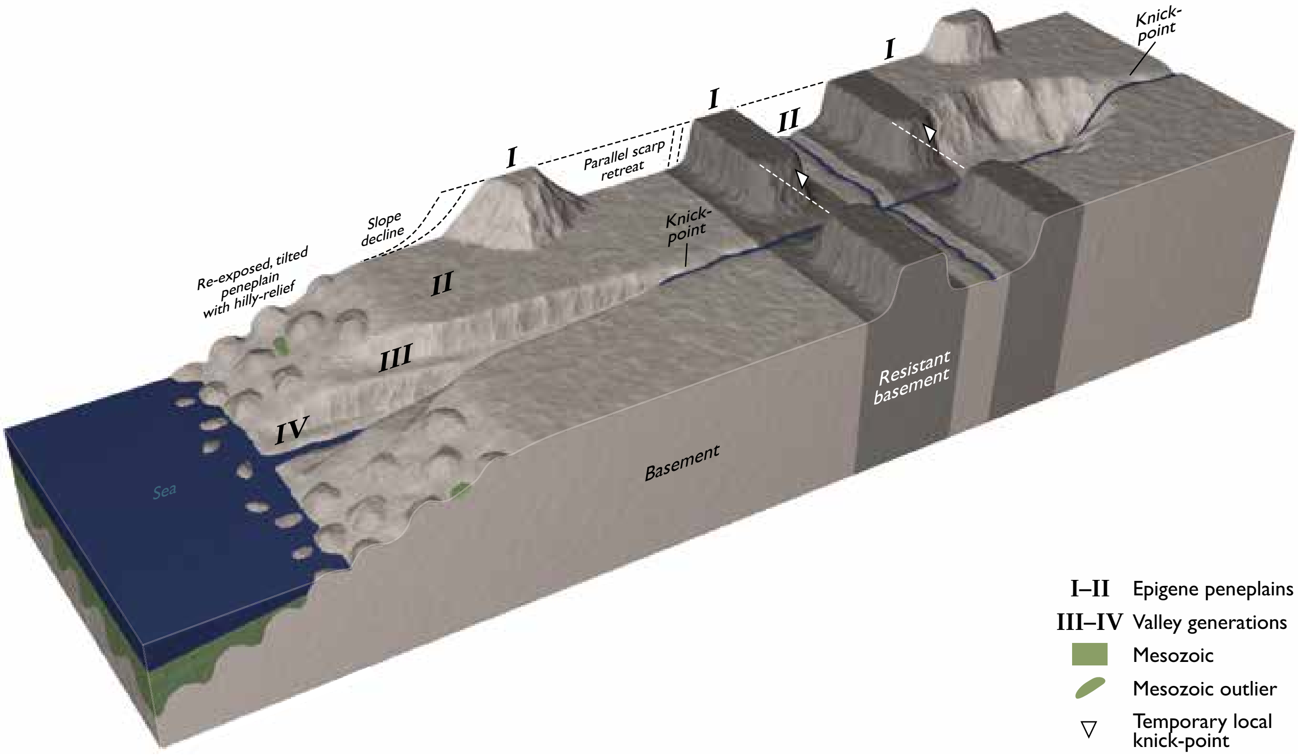


Fig. 14. Generalised model of relief development based on landscape analysis of southern Norway and southern Sweden. Major surface remnants (**I**) on both resistant and non-resistant rocks as a continuation of a common plane are indicative of a dissected higher peneplain. Slope decline according to Davis (1899) and parallel scarp retreat according to King (1962, 1967) is illustrated. The mode of valley incision to a new base level is now known to be dependent on climate and rock type. Superposition of fluvial valleys from the higher base-level plain produces river gaps through resistant rocks and continued surface formation behind the obstacle. Resistant rocks can give rise to temporary escarpments and residual hills. Note that at low elevations valleys (**III** and **IV**) might be incised in re-exposed relief. The near-horizontal peneplain **II** cuts off a re-exposed and inclined sub-Mesozoic surface with hilly relief. This shows that peneplain II formed after uplift following deposition of the Mesozoic sediments, e.g. during the Cenozoic as in the case of the South Swedish Dome where remnants of the Upper Cretaceous cover are preserved on the hilly relief at the flanks below the South Småland Peneplain (see Fig. 11).

valleys are the first sign of destruction. The dissection of the south-dipping basement plain of the recently uplifted Shillong plateau in India is a good example (Fig. 15; cf. Fig. 3; Biswas & Grasemann 2005; Biswas *et al.* 2007). Valleys can only incise due to a lowered base level and elevated surfaces are destroyed successively by fluvial incision and subsequent valley widening (King 1967; Ahnert 1982, 1998; Gunnell 1998; Lidmar-Bergström *et al.* 2000, 2007; Bonow *et al.* 2003, 2007a). Stroeven *et al.* (2009) noted the effectiveness of the fluvial system for dissection of the north-east Tibetan plateau.

High-level peneplains are often assigned very old ages even far away from the contact between the basement and its cover rock (e.g. Twidale 2007). It is argued that they must be older than the cover but without checking if it really is the same surface as at the cover contact. We have noted that such peneplains in Fennoscandia and Greenland always cut off the adjacent re-exposed surface and are thus younger (Fig. 11). They are not continuations of any re-exposed surface, but new surfaces. Thus the old surface has been destroyed and a new surface has formed (sections 3.1.2, 3.1.3).

Experience in Greenland (Japsen *et al.* 2006) suggests that epigene surfaces older than Miocene are unlikely to be preserved unless they have been carved in extremely resistant rocks or preserved in a landscape of high aridity (cf. section 8.1). Re-exposed surfaces can be very old and they are easy to date approximately.

In conclusion: A palaeoplain only exists for a certain time, which varies with different climatologic, tectonic and lithological settings. Palaeoplains (uplifted epigene or re-exposed surfaces) are therefore transient features (cf. Bishop 2007).

3.3.1 Effects of glaciation

Glacial erosion can aid in re-exposing palaeosurfaces as it easily erodes cover rocks, particularly those that are not well consolidated. Glacial erosion easily evacuates saprolites, and therefore hilly etch surfaces are accentuated in formerly glaciated terrain, e.g. the sub-Mesozoic relief in south Sweden (Olvmo *et al.* 1999). How much saprolite remains in formerly glaciated terrain depends on the degree of glacial erosion (Hall & Sugden 1987; Lidmar-Bergström 1997). On the other hand regional flat peneplains remain flat after glacial erosion even in regions with so-called scoured bedrock (Johansson *et al.* 2001b). Scoured refers to the occurrence of thin patchy drift and the grinding of the bedrock surface (Kleman *et al.* 2008) but not to deep erosion of hard, fresh rock. In general, glacial erosion reinforces preglacial relief by preferentially eroding in valley floors and valley sides (Ljungner 1949; Sugden 1978; Rudberg 1992; Fredin 2002, Bonow *et al.* 2003) rather than on summits, which were protected from erosion by the ice cover due to cold-based conditions over long periods of time (Sugden 1978; Kleman 1994; Kleman & Stroeven 1997; Kleman & Glasser 2007; Kleman *et al.* 2008). The deepest glacial erosion is found along major valleys that have hosted outlet glaciers from large ice sheets, where glacial erosion to over 1000 m below sea level has been documented, for example in Sognefjord, Norway (see Fig. 4; Nesje & Whillans 1994). Knickpoints are destroyed but remnants of earlier valley bottoms may occur as local valley benches (Bonow *et al.* 2003). There is also a significant difference in the effect of glacial erosion over basement terrain and across sedimentary rocks as shown by the substantial erosion along the Norwegian coast (Rise *et al.* 2005). In mountainous terrain the erosion of valley sides and valley floors by valley glaciers makes the pattern of preglacial slope and knickpoint development difficult to reconstruct (cf. Bonow *et al.* 2003). Cirques destroy palaeosurfaces to form an 'alpine relief' (Figs 4, 5), where

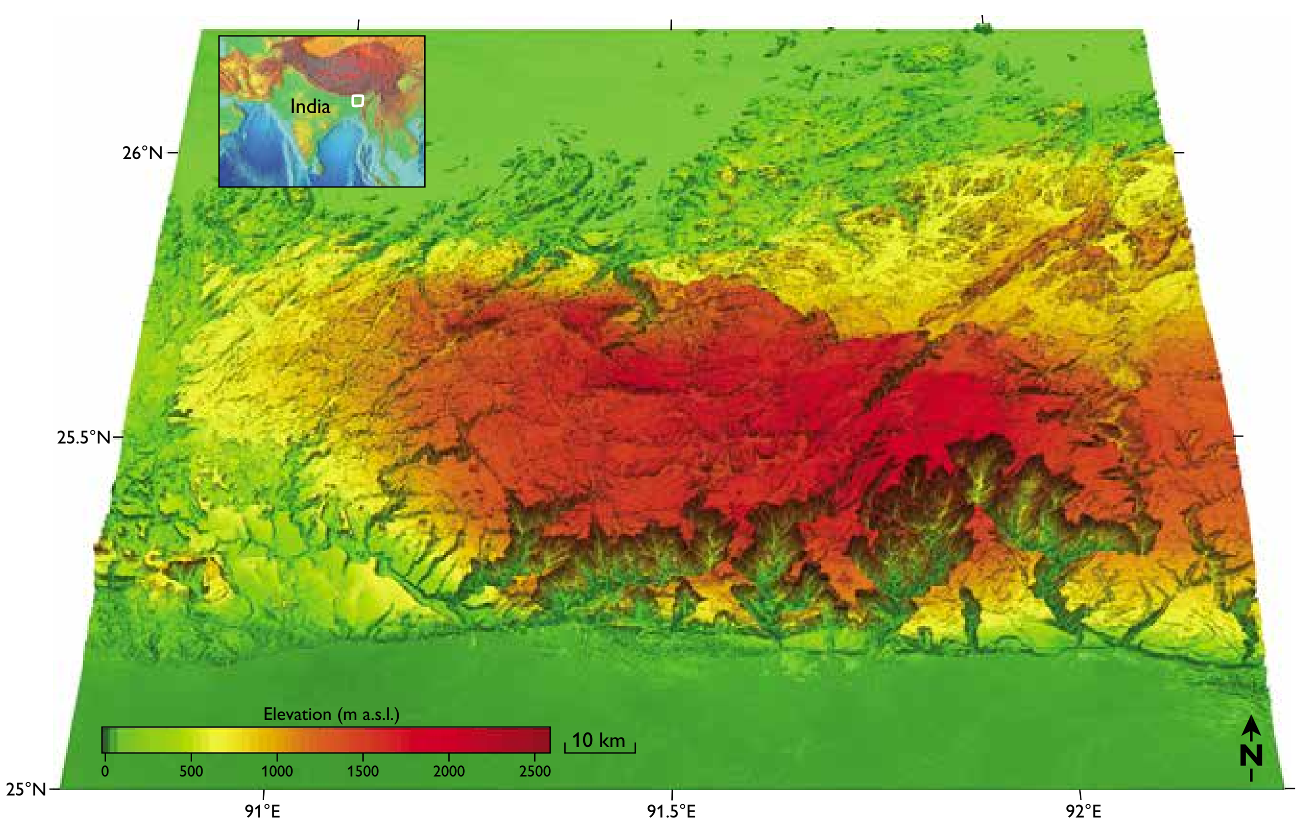


Fig. 15. Shillong plateau, India. The Precambrian basement was exposed in the Late Cretaceous and subsequently buried and covered by thick Cenozoic deposits. The gently south-dipping basement plain is re-exposed from these strata after Pliocene uplift (Biswas *et al.* 2007). The southward-facing, inclined peneplain is currently being dissected by deep valleys down to the new base level, similar to the development of the east Australian escarpment and coastal plain (Fig. 3). The succeeding valley-widening process, ultimately resulting in a new base-level-governed peneplain, depends on lithology and weathering conditions. 3D-terrain model with vertical shading. Elevation data from Jarvis *et al.* (2008).

elevations are sufficiently high and the climate is suitable (Etzelmüller *et al.* 2007).

As discussed in section 3.1.7, a 'glacial-buzzsaw' effect can reduce mountain heights in areas exposed to high precipitation and can produce summits of equal heights (Mitchell & Montgomery 2006), and the Equilibrium Line Altitude (ELA) can form a base level for intensified glacial erosion in valleys (Brozovic *et al.* 1997). On the Nuussuaq peninsula and the island of Disko in central West Greenland, the Upper Planation Surface (UPS) is tilted to the west and well preserved below 1000 m a.s.l. (see chapter 5), while it is more or less destroyed above this level close to the coast (Bonow *et al.* 2006b). The UPS is again identifiable 100 km eastwards, where it is a coherent surface descending from 2000 m a.s.l. in the west to 500 m a.s.l. in the east (Bonow *et al.* 2006b). In the intervening region, glacial erosion has limited the summit heights to about 1500 m a.s.l., possibly by the 'glacial-buzzsaw' mechanism (Mitchell & Montgomery 2006).

In conclusion: Peneplains in stepped sequences can also be identified in formerly glaciated areas, while preglacial valley steps are destroyed. In areas close to a coast, high peneplains might be dissected by cirques and valley glaciers.

3.4 Combining SLA and thermochronology

Development of peneplains clearly requires erosion of large amounts of rock. Analysis of peneplains, re-exposed and epigene, and their cross-cutting relationships, provides basic information on the relative chronology of their formation, subsidence, burial, uplift, tilting and formation of new peneplains. The most recent uplift relative to sea level is revealed by re-exposure of formerly covered peneplains and also by valley incision below the lowest of uplifted peneplains. SLA is based on integration of information on the character of surfaces (forms and saprolites), the relationships between them, and geological constraints. Special emphasis is placed on identifying sedimentary remnants which define re-exposed peneplains. Where applicable, the effects of glacial reshaping should also be taken into account.

SLA is independent of thermochronology so the two disciplines are discussed separately here. The relative chronology of events provided by SLA can be placed within an absolute frame of reference using low-temperature thermochronology. Integration of the two approaches allows determination not only of the timing of erosional episodes but also quantifies amounts of removed section, including both basement and sedimentary cover rocks. This is discussed in the next chapter.

4. Low-temperature thermochronology

4.1 Apatite fission-track methods

4.1.1 Historical background

Naturally occurring fission tracks are radiation-damage trails produced by the spontaneous fission of ^{238}U atoms, in which a uranium atom splits into two fragments. These fragments are stripped of electrons and positively charged, so they repel each other through the lattice and create a linear damage zone consisting of displaced atoms. These damage zones are highly chemically reactive, and can be selectively dissolved and thereby enlarged by a simple etching treatment. Thus fission tracks produced by spontaneous fission of individual U atoms within an apatite crystal can be revealed where they intersect a surface (Fig. 16).

The number of tracks per unit area of a polished and etched grain surface is controlled by uranium content and time, through standard decay laws, and track length, through geometrical considerations (Galbraith 2005). Therefore, by counting the number of tracks and measuring the uranium content, a 'fission-track age' can be calculated which, in the absence of other factors, should indicate the time over which tracks have accumulated. The uranium content is measured by irradiating a grain mount with thermal neutrons, which produces induced fission of ^{235}U atoms in the apatite grains. Some of the resulting fission fragments are emitted from the grain surfaces and are recorded as 'induced tracks' in a muscovite external detector after etching. The ratio of spontaneous tracks (daughter product) to induced tracks (parent) is converted to a numerical age using age standards (e.g. Hurford & Green 1983; Green 1985). Reviews of the basics of fission-track dating are provided by e.g. Fleischer *et al.* (1975), Wagner & Van den Haute (1992) and Galbraith (2005).

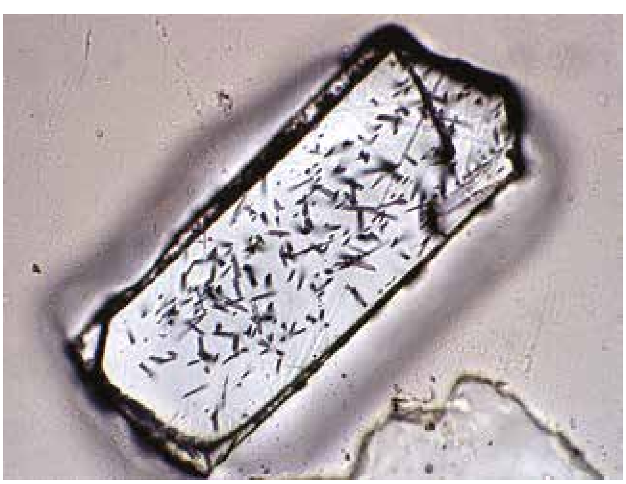


Fig. 16. Spontaneous fission tracks in an apatite grain, revealed after etching for 20 seconds in dilute nitric acid. The track openings are 1–2 µm in width. Individual tracks range up to a maximum of *c.* 15 µm in length.

Early applications of apatite fission-track dating to accessory apatite grains extracted from granitic rocks suggested that fission-track ages could be reset at relatively low temperatures around 100°C over geological timescales (e.g. Wagner & Reimer 1972). This observation was supported by early laboratory annealing studies (Wagner 1968; Naeser & Faul 1969), and subsequently confirmed by direct measurement of fission-track ages in subsurface samples (Naeser & Forbes 1976).

Integration of fission-track ages with 'confined track' length measurements (Fig. 17), first reported by Bhandari *et al.* (1971), led to a deeper understanding of the method. Confined tracks are totally enclosed within the body of the crystal and are revealed when the etchant penetrates cracks or other tracks and intersects tracks below the surface of the grain. For this reason, the entire length of the track can be etched and measured, in contrast to the tracks used in age determination, which are truncated by intersection with the surface.

Early measurements showed that whereas new tracks produced by induced fission of ^{235}U were characterised by mean confined track lengths around 16 µm, spontaneous tracks are always shorter. Even in volcanic rocks which have only experienced very low temperatures after initial post-eruption cooling, mean confined track lengths are typically around 14 to 15 µm (Gleadow *et al.* 1986a). For both types of track, the distribution of track lengths showed a standard deviation of *c.* 1µm, due to the variation in mass and energies of the fission fragments. Green (1980) showed that the track shortening by 1–2 µm in these volcanic apatites can be explained in terms of thermal annealing of these tracks at low temperatures (<50°C) over geological timescales, highlighting the sensitivity of the technique. Donelick *et al.* (1990) later showed that a small degree of initial track-length reduc-

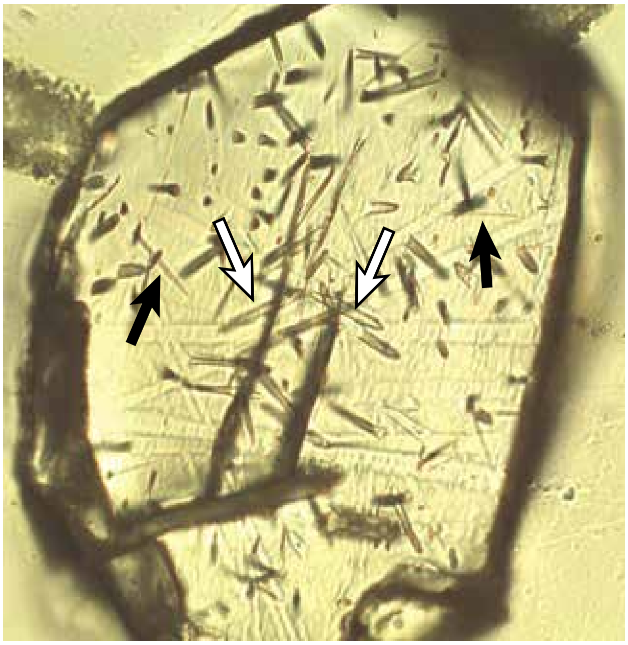


Fig. 17. Confined fission tracks in apatite, revealed by focusing below the grain surface. Confined tracks can be revealed by intersections with other tracks (Track-IN-Track or TINTs; black arrows) or with cracks or fractures (Tracks-IN-CLEavage or TinCLEs; white arrows).

tion proceeds quickly even at room temperature for fission tracks produced by induced fission, further emphasising that annealing proceeds at finite rates even at low temperatures.

Results from boreholes in the Mesozoic–Cenozoic Otway Basin of south-east Australia (Gleadow & Duddy 1981) provided the first rigorous constraints on the thermal stability of tracks in geological conditions, revealing the progressive reduction in fission-track age with increasing depth and temperature, and showing that this was complemented by a corresponding decrease in mean confined track length (Fig. 18). Laboratory studies (Green *et al.* 1985, 1986; Green 1988), together with detailed mathematical analysis (Laslett *et al.* 1982; Galbraith & Laslett 1988, 1990), established that the progressive reduction in track length causes the reduction in fission-track age, by reducing the proportion of

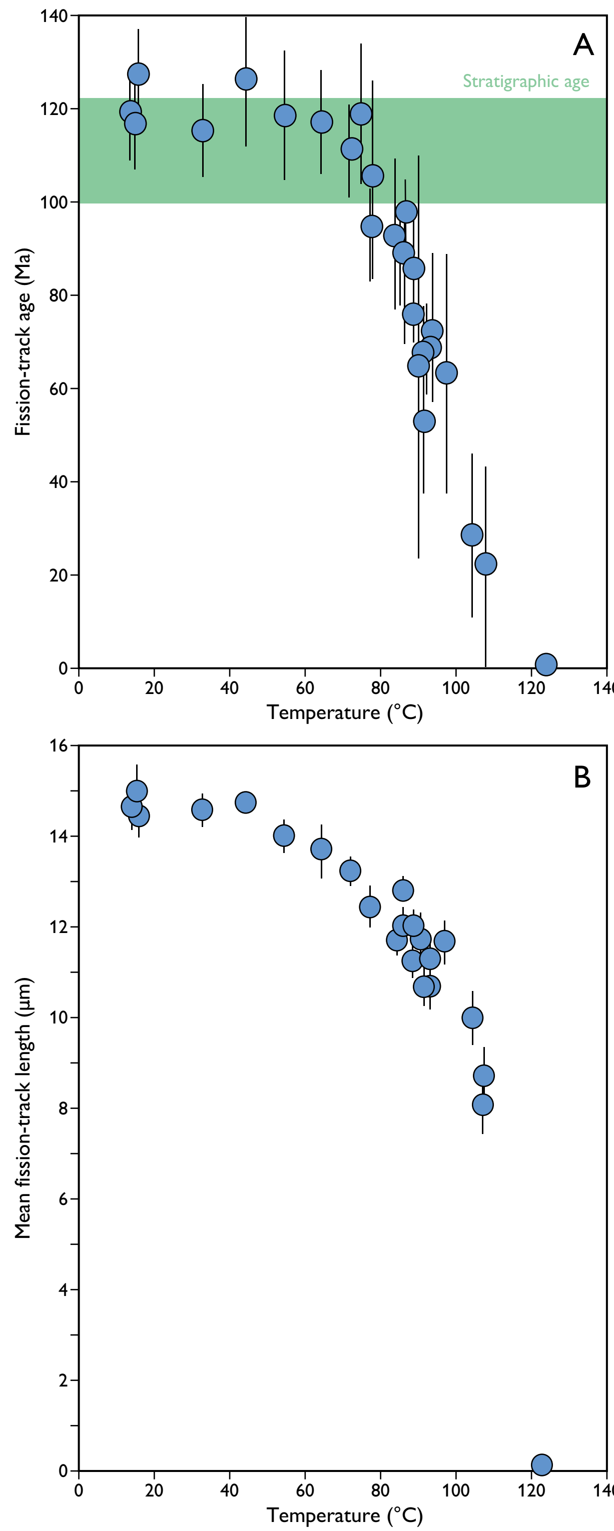


Fig. 18. Fission-track ages and mean confined track length in apatites extracted from samples of Early Cretaceous volcanogenic sandstones in a number of wells located in the Otway Basin, SE Australia. The sedimentary sections intersected in these wells are characterised by simple histories involving continuous burial, such that all samples are now at their maximum postdepositional temperatures. The decrease of both fission-track age and mean track length with down-hole temperature therefore provides a direct expression of the thermal sensitivity of fission tracks in these apatites over geological timescales. Modified from Green & Duddy (2013).

tracks that can intersect a polished grain surface. The reduction in track length is, in turn, a manifestation of the reduction in the degree of damage within the track region as displaced atoms return to their original lattice sites by thermally activated diffusion. Recognition of track-length reduction as the controlling process has underpinned all subsequent efforts to quantitatively predict apatite fission-track parameters and extract thermal history information from such data.

4.1.2 Thermal response of fission tracks in apatite

Early laboratory studies of the thermal sensitivity of fission tracks in apatite (e.g. Wagner 1968; Naeser & Faul 1969) were based on measuring the reduction of fission-track age resulting from various heat treatments. The large uncertainties in measured ages or densities, combined with possible unrecognised effects due to compositional differences (as discussed later) introduces considerable uncertainty into such measurements, allowing a range of interpretations. With the advent of confined track-length measurements, the greater precision and reproducibility of such measurements for representing the degree of annealing allowed significant refinement in the quantitative understanding of the kinetics of fission-track annealing in apatite.

Laslett *et al.* (1987) showed that the variation of mean track length with temperature and time in laboratory annealing studies could be described by a 'fanning Arrhenius plot' model (adopted from earlier studies), in which contours of equal track-length reduction form straight lines in a plot of time against inverse absolute temperature, the slopes of which (reflecting an 'activation energy') increase as the degree of annealing increases. Laslett *et al.* (1987) favoured a model in which all contours of equal annealing converge to a point at infinite temperature ($1/T_o = 0$), because it allowed a simplified mathematical treatment. A variety of alternative annealing models have subsequently been published (Crowley *et al.* 1991; Laslett & Galbraith 1996; Ketcham *et al.* 1999). All of these represent refinements or alternative forms of the basic fanning Arrhenius plot model proposed by Laslett *et al.* (1987), e.g. involving finite values of $1/T_o$, or curved contours, and the principles described in the following sections also apply to these models.

The improved definition of the kinetics of fission-track annealing provided by using mean confined track length as the fundamental parameter (Laslett *et al.* 1987), combined with a detailed understanding of the way in

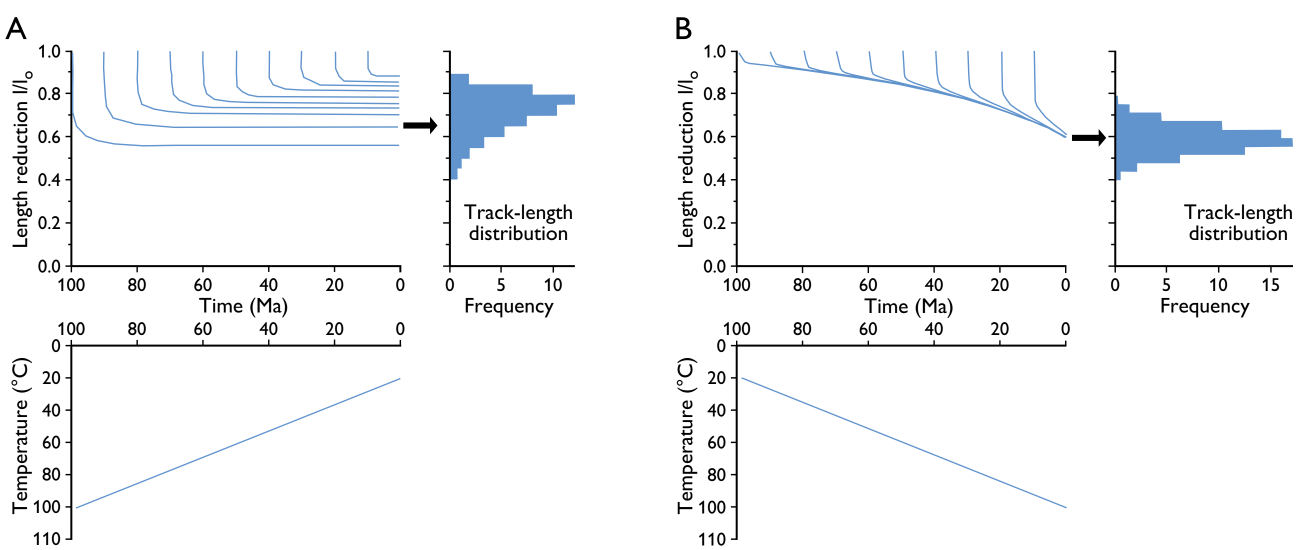


Fig. 19. Predicted track-length reduction vs. time for tracks formed at different times within histories involving continuous cooling (**A**) and heating (**B**), together with final predicted track-length distributions (right). The trajectories and final track-length distributions are calculated following the methods outlined by Green *et al.* (1989). Whereas in the cooling history, tracks formed at different times end up at the present day with different lengths, in the heating case almost all tracks are shortened to the same degree, reflecting the dominance of temperature over time in the kinetics of annealing. Track length reduction, l/l_0, where l is measured mean track length and l_0 is the initial (unannealed) mean track length.

which reduction in track length is manifested in the fission-track age (Green 1988), provided the basis for the quantitative modelling of the response of fission tracks in apatite to different styles of thermal history in geological conditions (Green *et al.* 1989). A key step in this process is the transition from isothermal annealing models to histories in which temperature varies with time. Duddy *et al.* (1988) provided a way forward by adopting the principle of 'equivalent time' (originally postulated by Goswami *et al.* 1984), by which the rate of annealing of a track at any given time only depends on the length to which the track has already been reduced and the prevailing temperature, and not on the history of how the track reached that length. Duddy *et al.* (1988) verified this assumption in a series of variable temperature-annealing experiments.

A key outcome of this work is the recognition of the dominance of temperature over time in fission-track annealing. In thermal histories involving continuous cooling, tracks produced at different times throughout the history are shortened at a rate which is initially high but is quickly reduced to a minimal level as the temperature falls through time. Tracks produced at different times experience declining temperatures and are therefore shortened to different degrees (Fig. 19A). In contrast, for histories involving continuous heating, because temperature dominates over time in the kinetics, the lengths of tracks produced at different times are progressively reduced through time, and all tracks are shortened by more or less the same amount, except for those produced within the last few per cent of the history (Fig. 19B) although it should be appreciated that tracks are produced with a finite spread of lengths, as discussed below.

These contrasting styles of behaviour have serious implications for the response of the apatite fission-track system in thermochronology. Figure 20 illustrates the development of apatite fission-track parameters through a history involving heating followed by cooling and subsequent residence at low temperature. During the heating phase all tracks formed at different times are progressively reduced in length, and at the palaeothermal maximum all but a small proportion of tracks that have formed during the heating phase are shortened to the same degree. Tracks formed after the onset of cooling remain longer, reflecting the lower temperature. At the present day, two populations of tracks are present, one with short lengths representing tracks formed during the heating phase, and another comprised of longer tracks formed after the onset of cooling (Fig. 20). The population of shorter tracks will contribute a reduced component to the fission-track age, compared to the time interval over which tracks have been retained, while the contribution to the fission-track age from the longer population will be much closer to the time interval after the onset of cooling. The final measured fission-track age will represent the summed contributions of both components.

A sample which reached a maximum palaeotemperature sufficient to reduce track lengths to zero prior to cooling will only contain a single population of tracks, formed after cooling to a temperature at which tracks are retained (analogous to the population of longer tracks in shallower samples). The fission-track age in such a sample will be determined by the time since the sample began to retain tracks, but the precise value will depend on the degree of length reduction of tracks formed during the cooling history. Figure 21 illustrates how this results in a characteristic variation of apatite fission-track parameters with depth in a sedimentary section that has been heated and then cooled.

4.1.3 Variation in annealing kinetics between different apatite species

Studies of apatite fission-track parameters in subsurface samples from the Otway Basin, south-east Australia (Gleadow & Duddy 1981; Green *et al.* 1985, 1986), showed that the chlorine content of the apatite grains exerts a systematic influence on annealing rates (Fig. 22). This was subsequently confirmed in laboratory studies by Carlson *et al.* (1999) and Barbarand *et al.* (2003; Fig. 23). Although both these studies tended to downplay the influence of chlorine in favour of other factors, the importance of differential annealing within individual samples related to wt% Cl has been demonstrated in a number of studies (e.g. Argent *et al.* 2002; Crowhurst *et al.* 2002; Green *et al.* 2002; Green 2005; Green & Duddy 2013). In contrast, evidence for systematic differences in annealing rates in geological samples due to any element other than Cl has yet to be demonstrated. A correlation between etch pit diameters and annealing rates has been used in a number of studies to allow for differential annealing between different apatite species (Ketcham *et al.* 1999), although Green *et al.* (2005) showed that annealing rates correlate much more strongly with wt% Cl than with etch pit size (see also Green & Duddy 2013).

In geological conditions, differential annealing effects within individual samples are maximised in rocks which have been heated into the critical temperature range (typically 90–120°C). Over this range, the most sensitive (i.e. low-Cl) apatites are totally annealed while more resistant apatites (high-Cl) are unaffected (cf. Fig. 22). In such cases, the systematic dispersion in fission-track age, correlating with wt% Cl, can provide added precision to a thermal history solution (Green *et al.* 2002).

An example of an apparently anomalous observation that can be simply explained in terms of differences in wt% Cl is the difference in fission-track age between adjacent gneissic and charnockitic terrains in India, which Gunnell (2000) attributes to differing resistance to erosion. This is more likely to be due to a difference in annealing rates in apatites from the two rock types, with charnockitic apatites likely to be richer in Cl, and hence giving older ages. In such cases, measurement of Cl content by electron microprobe can easily resolve such effects. Lorencak *et al.* (2004) provide graphic evidence of the impact of small-scale variations in apatite Cl content within a single rock body on measured apatite fission-track ages due to the influence of wt% Cl on annealing rates.

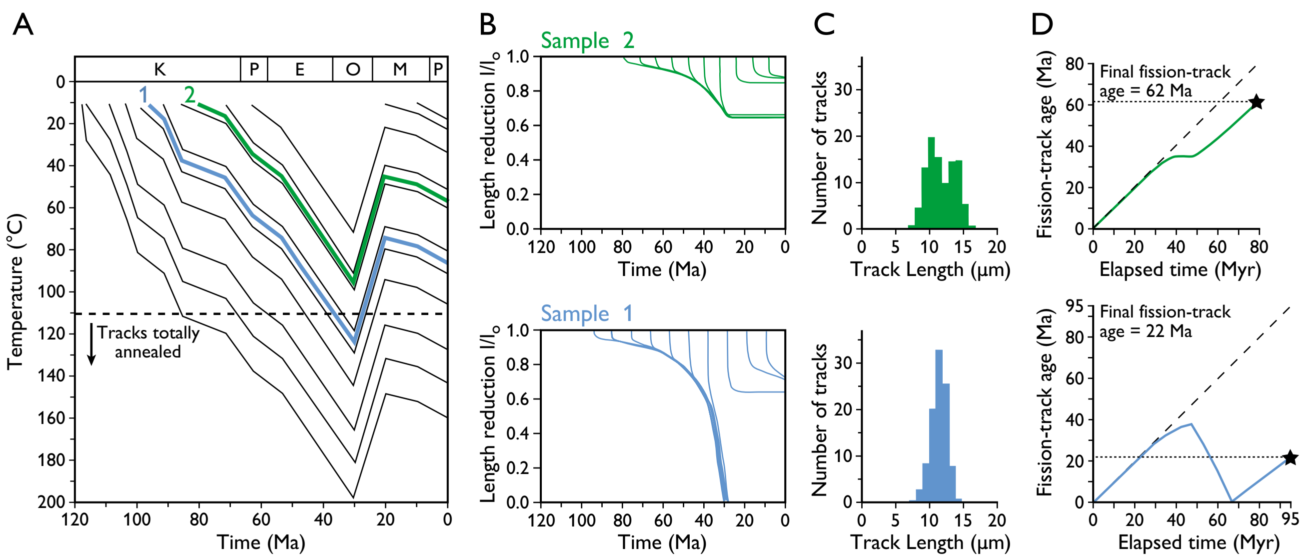


Fig. 20. Thermal history response of fission tracks in apatite under geological conditions. **A**: Notional thermal history for a sedimentary sequence that underwent progressive burial through the Cretaceous to middle Cenozoic, followed by cooling due to uplift and erosion commencing at 30 Ma and completed by 20 Ma, with minor reburial from 20 Ma to the present day. The thermal histories of two samples are highlighted. **B**: Track-length shortening vs. time for tracks produced at different times in samples 1 and 2. As temperature increases, the length of all tracks is progressively reduced, and because temperature dominates over time in the kinetics of annealing, at any time during this phase all but the most recently formed tracks at any given time have the same mean length (although each population of tracks has a finite spread in length). At the point when the maximum temperature is reached and the history changes from heating to cooling, all tracks formed up to that point in time are effectively 'frozen' at the length to which they have been reduced. They do not undergo further shortening because annealing rates are much slower at the reduced temperatures now prevailing, and they do not get longer because the annealing process is irreversible. Those tracks formed after the onset of cooling remain longer because of the lower annealing rates at the prevailing lower temperatures. Sample 2 reached a maximum temperature sufficient to reduce the length of all tracks produced up to that time to zero. At the present day, this sample contains only one track population, formed after the sample cooled to temperatures at which tracks could be retained (*c.* 110°C for typical apatite compositions). **C**: Track-length distributions for samples 1 and 2 resulting from the thermal histories shown in **A**. For sample 1, two populations of tracks are present at the present day; a shorter population representing tracks formed up until the onset of cooling from the palaeo-thermal maximum, and a longer population formed after the onset of cooling. For sample 2, the measured track-length distribution will reflect the thermal history in the post-cooling period only. **D**: Evolution of fission-track age with time resulting from the thermal histories shown in **A**. For sample 1, the final measured fission-track age will represent the summed contribution of the two populations of tracks; the shorter component will contribute a reduced component to the fission-track age compared to the time interval over which tracks have been retained. On the other hand, the contribution to the fission-track age of the longer population will be much closer to the time elapsed since the onset of cooling. For sample 2, the final fission-track age will be determined by the time when the sample began to retain tracks, but moderated by the degree of length reduction of tracks formed during the cooling history. While this is based on a mono-compositional apatite of Durango composition using the Laslett *et al.* (1987) model, the nature of this response is common to all forms of kinetic models, and is a fundamental property of the apatite fission-track system governed by a fanning Arrhenius plot. **Myr**: million years. Modified from Green *et al.* (2002).

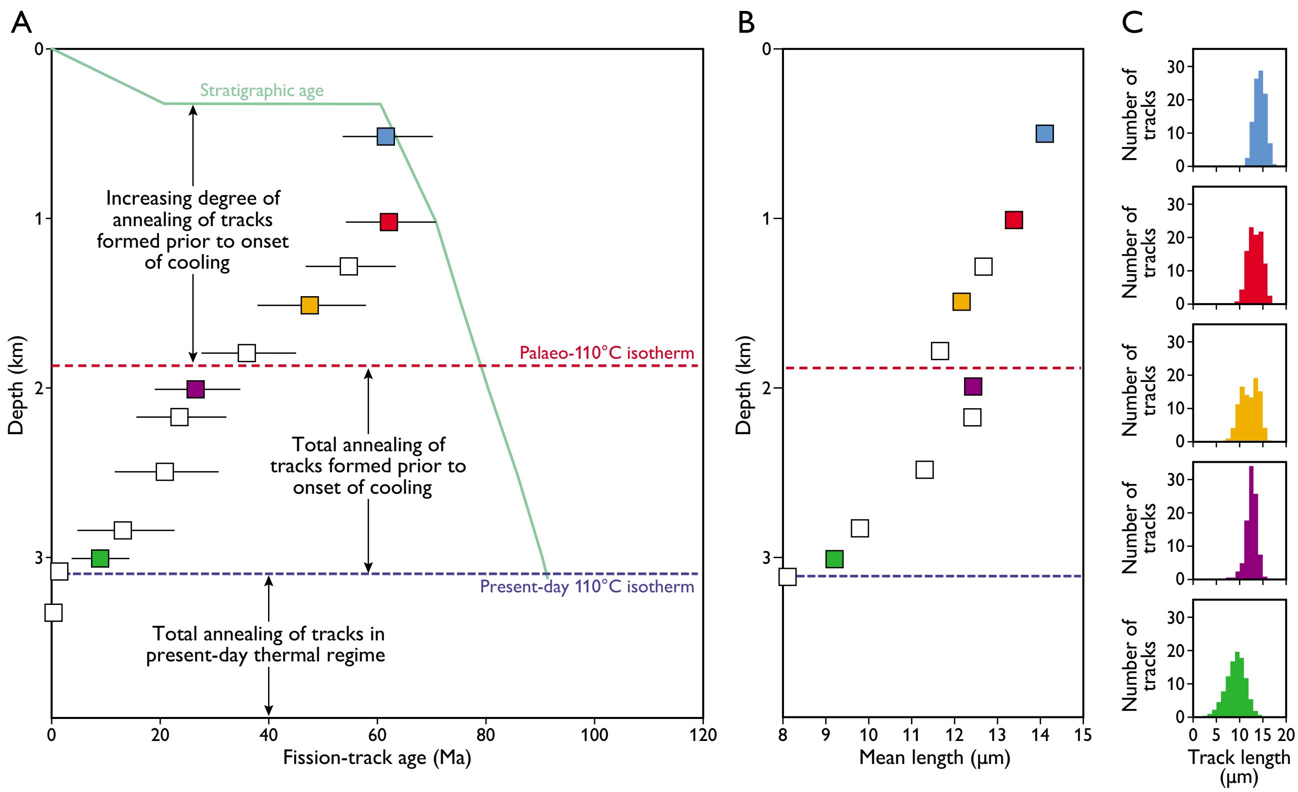


Fig. 21. Predicted variation of fission-track parameters with depth for a vertical rock section with a thermal history of the style shown in Fig. 20. **A**: Fission-track age declines rapidly with increasing depth through the shallower section, as the mean length of the shorter population of tracks (formed up to the onset of cooling) is progressively shortened, and the proportion of these tracks that can reach the polished grain surface and be revealed, decreases. **B**: Similarly, the mean track length reduces due to the decreasing mean length of the shorter population, evident in the track-length distributions (**C**). As the depth (and temperature of *c.* 110°C) corresponding to total annealing of all tracks formed prior to the onset of cooling is approached, the mean track length begins to increase again, as the shorter population of tracks becomes increasingly difficult to reveal and therefore contributes less to the overall mean for the sample, which is increasingly dominated by the longer population of tracks formed after cooling. As the transition from partial to total annealing of tracks formed prior to cooling is crossed at the palaeo-isotherm of *c.* 110°C, the mean track length increases abruptly as the sample is now dominated only by longer tracks formed after the onset of cooling. And the fission-track age reduction shows a characteristic 'break-in-slope', below which only a single component of tracks is present, with parameters controlled by the history after the onset of cooling. With further increase in depth, both the fission-track age and mean track length show progressive reduction to zero at a present-day 110°C, although in detail this temperature is controlled by apatite composition and the timescale of heating/burial. Modified from Green *et al.* (2002).

4.1.4 Extracting thermal history information from apatite fission-track data

Extracting explicit thermal history solutions directly from apatite fission-track data is not possible because of the high degree of redundancy in the data (i.e. many histories result in the same measured age and length parameters, Fig. 24). Instead, the problem is approached by forward modelling the apatite fission-track parameters expected from a range of specified thermal histories and defining the range of conditions which provide predictions that are consistent with the measured data. Within this common philosophy, a range of approaches has been developed, each of which has its advantages and disadvantages. This disparity can often confuse the non-specialist, so below we review the most commonly used approaches. Most published studies represent a variation on one of these basic themes. The basics of data generation in all approaches are similar in most respects.

Geotrack approach (AFTA®): The approach adopted in our own studies, referred to as AFTA (apatite fission-track analysis), is designed primarily for application to sedimentary basins, and takes account of the fact that sedimentary horizons are deposited at the surface then

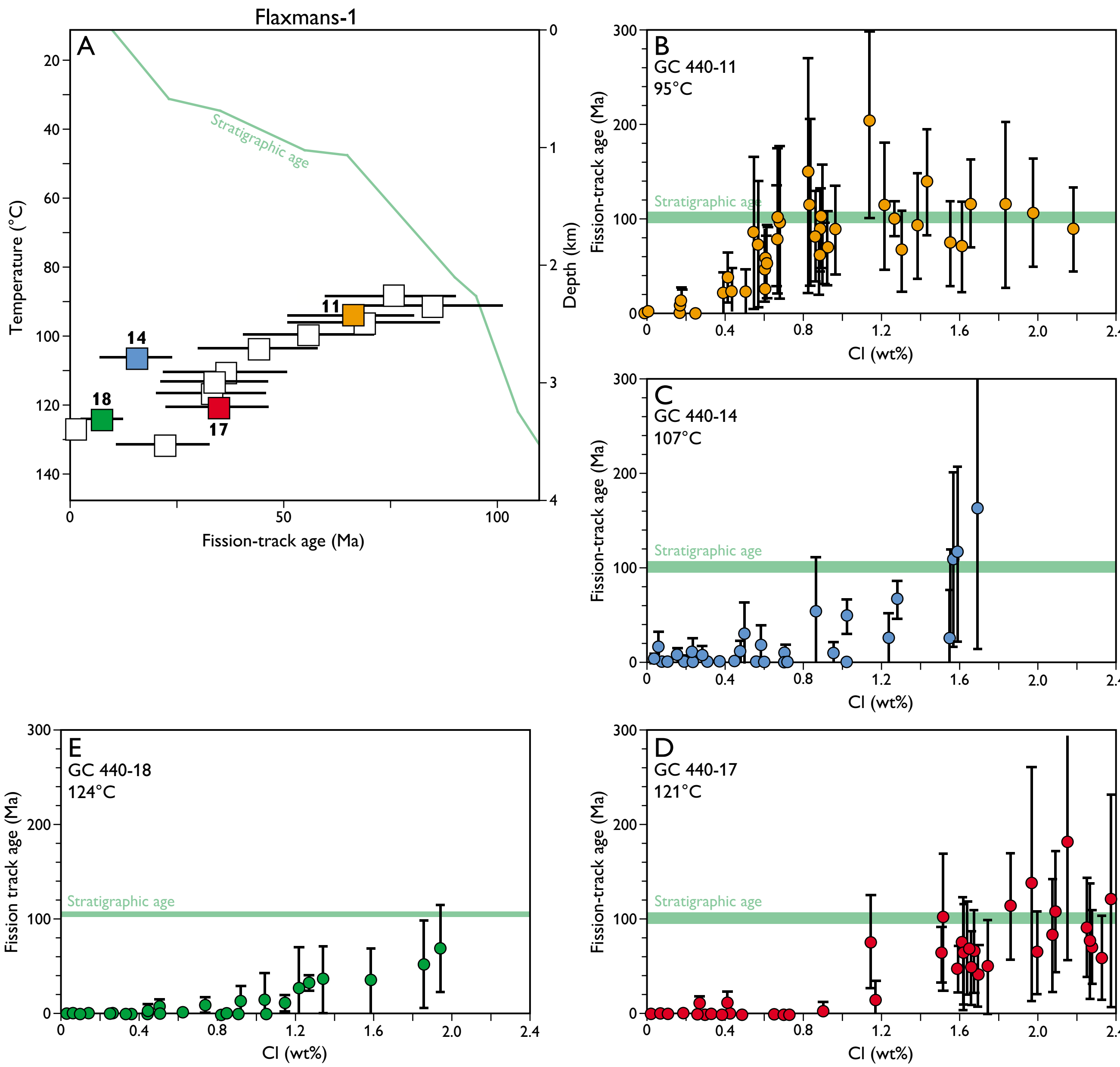


Fig. 22. **A**: Variation of central fission-track age with depth for samples from the Flaxmans-1 well, Otway Basin, SE Australia. **B–E**: The variation of fission-track age with chlorine content for individual apatite grains from four selected samples. In each of these samples the most sensitive (low-Cl) grains are totally annealed (i.e. zero FT age) while more retentive (higher-Cl) grains giving ages up to the depositional age and above. With increasing present-day, down-hole temperature (the maximum post-depositional temperature in these samples), the transition to total annealing shifts to progressively higher Cl contents, demonstrating the systematic influence of chlorine content on annealing. Note that while most of the central fission-track ages define a generally smooth decrease with increasing temperature, the age for sample GC440-14 is off trend. This is due to the absence of grains with Cl >1.6 wt% compared to adjacent samples, which are dominated by more retentive grains and therefore give higher central ages. This major effect of chlorine on apatite fission-track age must be taken into account in order to extract meaningful geological constraints from apatite fission-track data. Modified from Green & Duddy (2013).

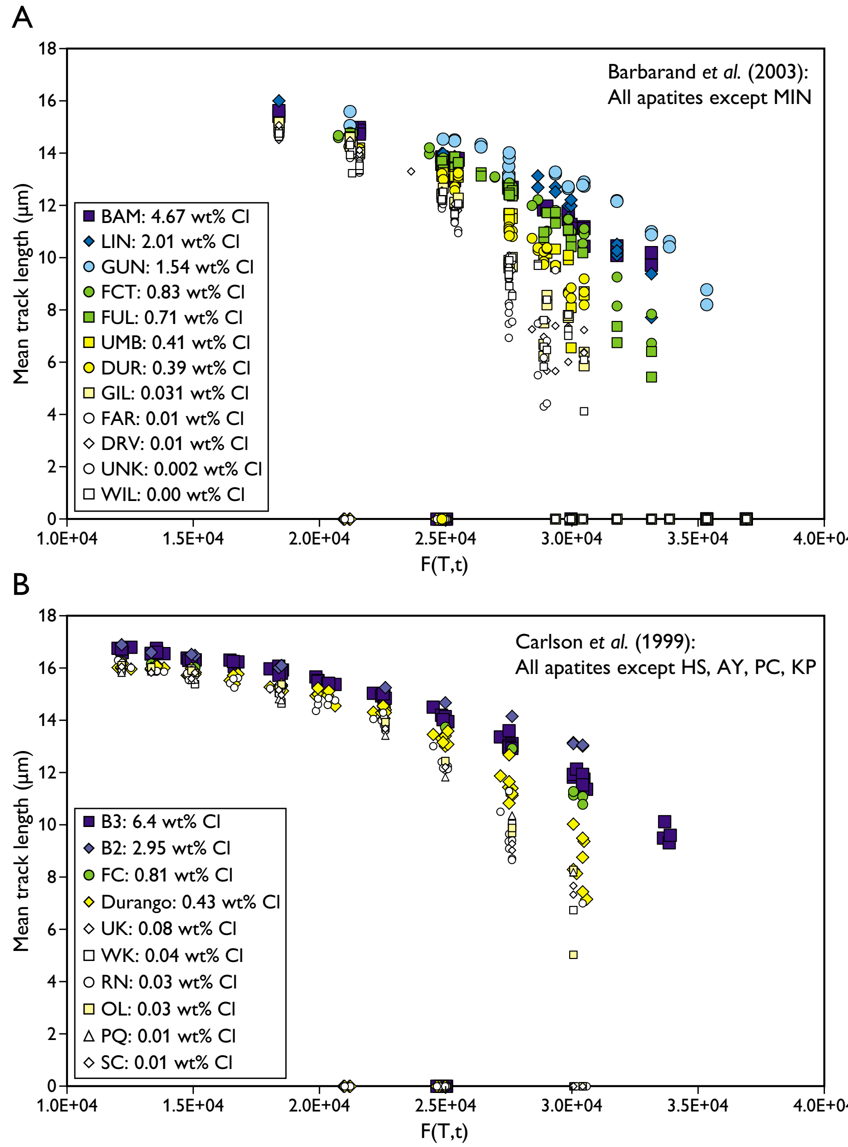


Fig. 23. Mean track lengths from laboratory annealing experiments reported by Carlson *et al.* (1999) and Barbarand *et al.* (2003), plotted against a unifying function of temperature (T) and time (t), which reduces all data to a common scale. This function is of the form $F(T,t) = [\log t - \log t_o]/[(1/T)-(1/T_o)]$, $\log t_o = -10$ and $1/T_o = 0.001$. Apatites of different Cl content are coded to illustrate this variation, with high Cl contents (>1 wt% Cl) shown in blue colours and large symbols, apatites low in Cl (<0.1 wt % Cl) shown in pale colours and smaller symbols, compositions around 0.4 wt% Cl shown in yellow and those around 0.8 wt% Cl in green colours. The Durango apatite (*c.* 0.4 wt% Cl, yellow) and Fish Canyon Tuff apatite (*c.* 0.8 wt% Cl, green circles) are common to both datasets. These results clearly illustrate the first order control on annealing rates exerted by Cl content, with apatites high in chlorine giving longer lengths for any given heat treatment than those low in Cl. While the first-order control from chlorine is clear, other elements produce additional variation, and several apatites from each dataset have been omitted as they are not consistent with the main body of data. However, in data from natural samples, no systematic compositional control on annealing other than chlorine has yet been identified. Modified from Green & Duddy (2013).

buried and heated to some maximum depth and temperature, after which they may be exhumed and cooled. This allows a convenient parameterisation of the history in terms of one or more episodes of heating and cooling.

Because of the high level of redundancy in the data (see above), in order to extract useful information from AFTA data it is important to establish a framework with some fixed independent constraints, without which the range of possible solutions is unworkably large. For this purpose, we construct a 'Default Thermal History' (DTH), which represents that part of the history that can be constrained from geological evidence. The DTH for an outcrop sample of sedimentary rock is defined by two points, deposition and the present day, at which the sample is at the appropriate mean surface temperature. For basement outcrop samples, a similar approach can be adopted using the age of the oldest overlying sedimentary unit (if no cover is present, the intrusion age or youngest metamorphic age can be used). For downhole samples, the DTH is calculated from the burial history derived from the preserved section combined with the present-day thermal gradient.

If the AFTA data can be explained either solely by the DTH or by a combination of the DTH and tracks inherited from sediment source regions (for sedimentary rock samples), no further information can be obtained from the data because the data are dominated by the maximum temperature. But if the data show a higher degree of annealing (in terms of either fission-track age reduction or track-length reduction) than can be accommodated by the DTH then the sample has been hotter during the period covered by the DTH and the data can

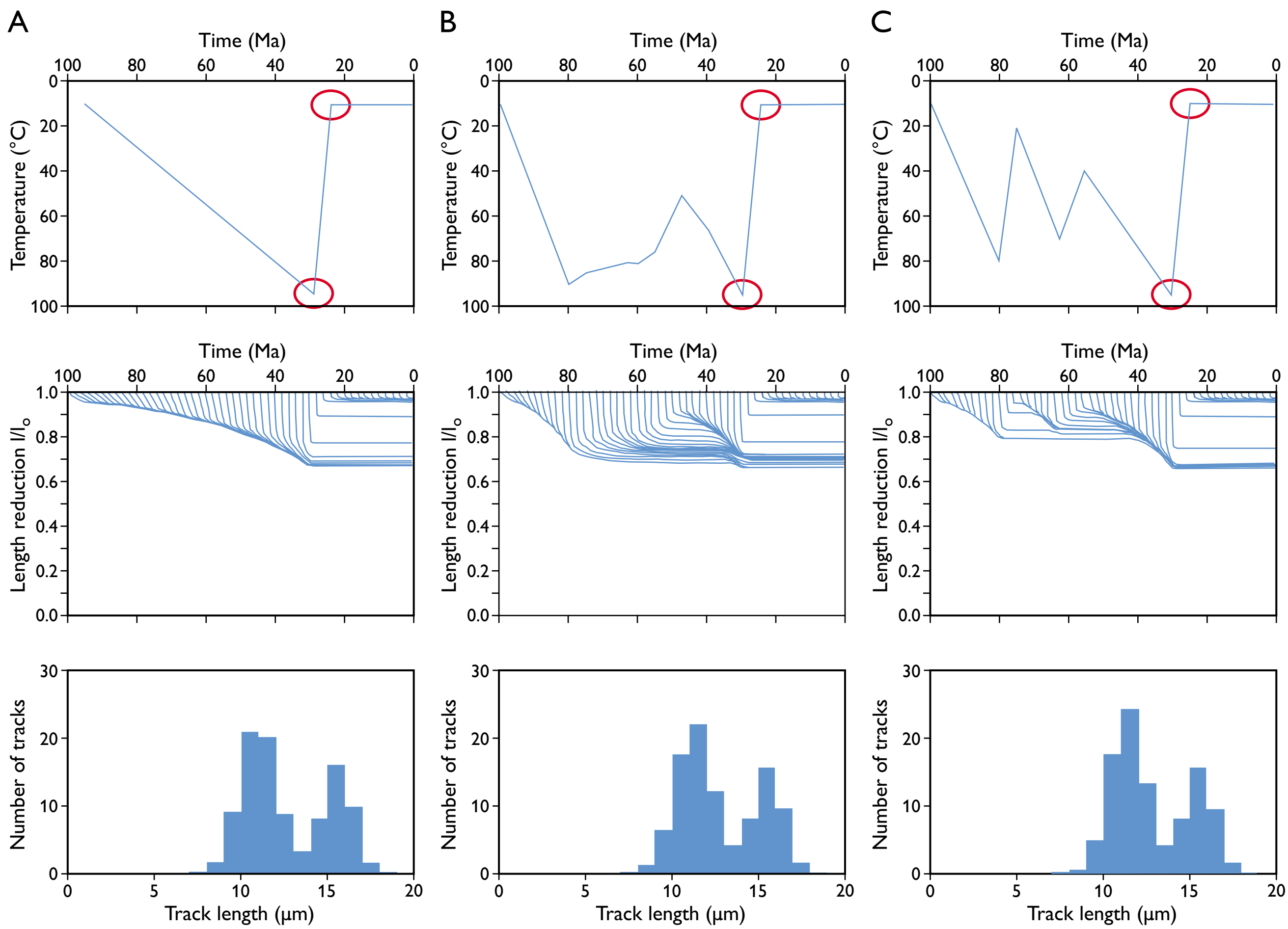


Fig. 24. Shortening trajectories (centre row) for tracks produced at different times through three thermal history scenarios (top row) representing increasing levels of complexity from **A** to **C.** The resulting track-length distributions are also shown (bottom row). Despite the obvious differences in the thermal histories, the resulting track-length distributions are almost indistinguishable because both the time and magnitude of maximum temperatures and the rate of cooling from maximum palaeotemperatures are the same for each history (circled points on the thermal histories). This outcome reflects the fundamental kinetics of the apatite fission-track system such that the data are sensitive only to the magnitude of the maximum temperature and the timing of the onset of cooling (in relation to the overall time over which tracks have been retained), and preserve no information on the prior history (except that temperatures must have been lower than at the palaeo-thermal maximum). Thus, successive heating episodes overprint the effects of earlier episodes, leaving only evidence of the maximum temperature episode and the subsequent history after cooling from the palaeo-thermal maximum. It is therefore not possible to discriminate between these three scenarios from apatite fission-track data. Modified from Green & Duddy (2013).

be used to define the main features of palaeothermal history (i.e. that period when the sample was hotter than it is today). In sedimentary rock samples which have not been heated above *c.* 60°C, the AFTA data may be dominated by tracks formed prior to deposition, and it may not be possible to resolve the effects of post-depositional heating. In such cases only an upper limit to the magnitude of the maximum post-depositional temperature can be obtained (Green & Duddy 2013).

By modelling the expected fission-track age and track-length distribution and their variation with wt% Cl resulting from a range of possible thermal histories, we can define the range of values of maximum palaeotemperature and the onset of cooling resulting in predictions which match the measured data within 95% confidence limits (Figs 25–27). The approach is based on likelihood theory similar to that described by Gallagher (1995) but it also includes the influence of composition. A multi-compositional annealing model is used which takes specific account of the influence of wt% Cl. This model consists of a series of parallel equations, each taking the form of a linear fanning Arrhenius plot with non-zero

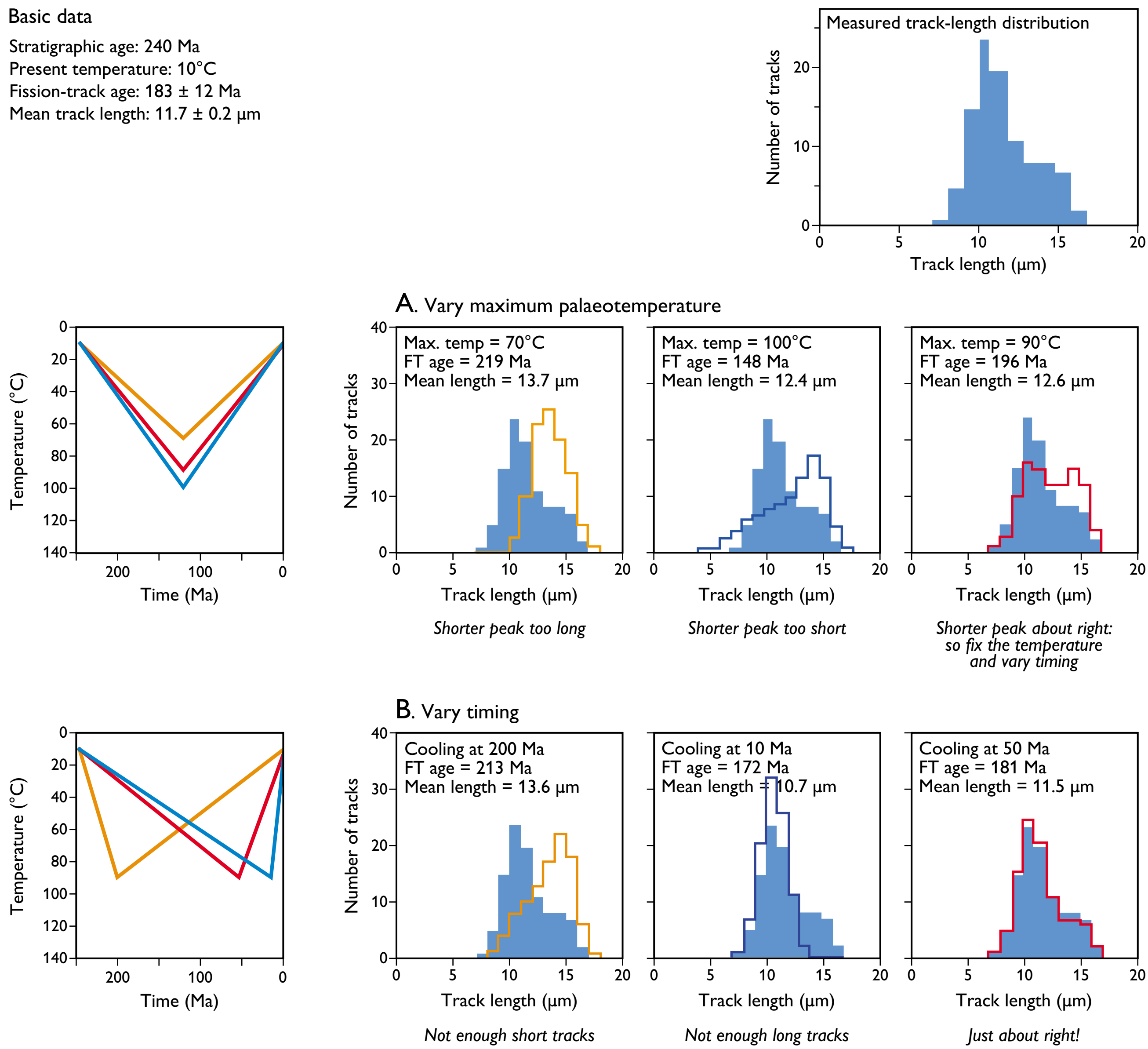


Fig. 25. Principles of AFTA interpretation illustrated for a monocompositional apatite, showing how a thermal history solution can be extracted from measured AFTA parameters (fission-track age, mean track length and track-length distribution). For samples of sedimentary rock it is necessary to know the stratigraphic age and present temperature of the sample. By predicting the AFTA parameters for various thermal history scenarios, we can define the best-fit thermal history. As a first step, assume that cooling from the maximum palaeotemperature occurred at the midpoint of the history (120 Ma in this case). **A**: By varying the maximum temperature and comparing measured and predicted parameters, we find a good match with the shorter population of tracks in the measured track-length distribution at a maximum palaeotemperature of 90°C. But the predicted track-length distribution contains too many long tracks. **B**: By fixing the maximum temperature at 90°C and varying the timing of cooling, a good match between the predicted and measured track-length distributions, as well as the fission-track age, is achieved with cooling commencing at 50 Ma. Note that no attempt is made to define the whole thermal history because the history prior to the onset of cooling is overprinted by the thermal maximum. Note also that by itself, the measured fission-track age of 183 ± 12 Ma provides no direct indication of the time of cooling, which only comes from kinetic modelling of the details of the track-length distribution together with the fission-track age. Predictions based on a mono-compositional apatite of Durango composition using the Laslett *et al.* (1987) model. Modified from Green *et al.* (2002) and Green & Duddy (2013).

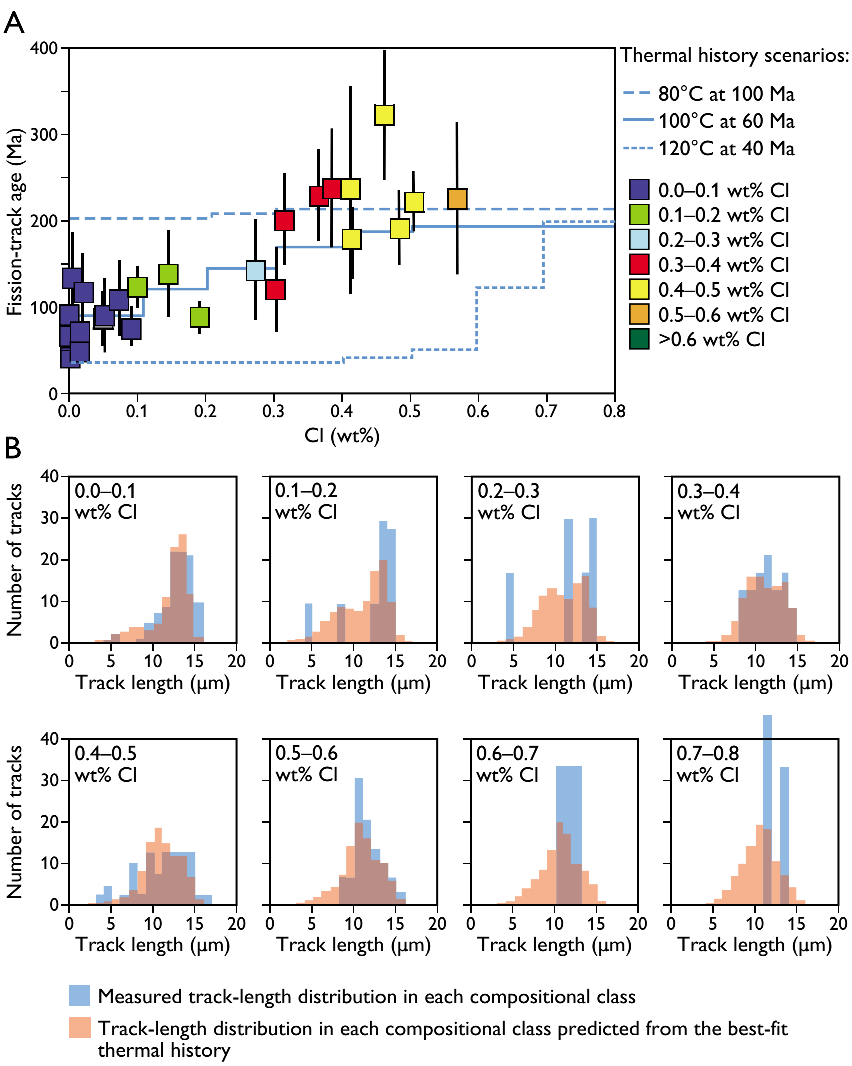


Fig. 26. Apatite fission-track interpretation methodology for a multi-compositional apatite sample. The same basic information and interpretation strategy as described in Fig. 25 is used for samples which contain apatite of different compositions, but supplemented by the wt% chlorine of each apatite grain (measured by electron microprobe) in which a fission-track age or track length is measured. Fission-track ages and track lengths are grouped according to the chlorine content into 0.1 wt% Cl intervals, and a multi-compositional annealing model is used which takes specific account of the influence of wt% Cl on annealing rates. The matching procedure is the same as for a sample of a single composition, but now involves the simultaneous matching of fission-track age and the details of the track-length distribution in all compositional groups present in the sample. In this example, 8 groups are present containing between 0.0 and 0.8 wt% Cl. The best-fit match to the data is achieved for cooling from a maximum temperature of 100°C beginning at 60 Ma (data in a Permian sandstone from north-eastern England). Modified from Green *et al.* (2002) and Green & Duddy (2013).

intercept, with constants which vary systematically with wt% Cl.

No attempt is made to define the whole thermal history because, as shown in Figs 19, 20, 24, the history prior to the onset of cooling is overprinted by the thermal maximum. Instead we focus on determining those aspects of the thermal history that directly control the measured AFTA parameters, viz. the maximum palaeotemperature and the time at which cooling from the palaeothermal maximum began (Fig. 20). Additional episodes of heating and cooling following the onset of cooling from the palaeothermal maximum can often be resolved, utilising the shortening of tracks formed after the initial cooling phase, provided that the magnitude and timing of the palaeothermal maximum and subsequent peak are sufficiently separated in temperature and time. In rare cases, three discrete episodes can be resolved in data from a single sample (e.g. Turner *et al.* 2008). The principles involved are illustrated in Fig. 27.

While the episodic history of heating and cooling adopted in our approach is designed specifically for application to sedimentary basins, we believe that it is also relevant to basement terrains (Green & Duddy 2010; Japsen *et al.* 2010; Lidmar-Bergström *et al.* 2013). Such an approach is essential in basement regions where sedimentary outliers and re-exposed peneplains occur, revealing earlier cycles of exhumation, burial and re-exhumation (e.g. Green & Duddy 2006, 2007). Examples of these situations are discussed in chapters 6 and 8.

In routine application, AFTA data are commonly integrated with data from other palaeothermal indicators such as vitrinite reflectance (VR), and/or indicators of burial such as sonic velocity (e.g. Japsen *et al.* 2007a). Such data provide an independent check on the interpretation of the AFTA data, and ensure that the resulting thermal histories and the information on denudation derived from them are not affected by artefacts of any individual technique. Final interpretations can also be integrated with information from SLA (chapter 3), such as the presence of re-exposed peneplains and incised valleys.

'HeFTy' and related approaches: The majority of published apatite fission-track studies over the last 10 years or so rely on extracting thermal history information from the 'HeFTy' software package (Ketcham 2005; Ehlers *et al.* 2005) which is a development from the earlier 'AFTSolve' software (Ketcham *et al.* 2000). Further details of analytical approach are provided by Ketcham *et al.* (2007, 2009). In this approach, repeated forward modelling of data through various thermal history scenarios within specified limits results in definition of a range of viable histories for which predictions provide 'acceptable' fits to the measured data, including a more restricted range of solutions providing 'good' fits. The ranges of good and acceptable fits provide an assessment of the degree of uncertainty in the history. This approach typically allows for within-sample variation in annealing kinetics using etch pit size as the kinetic parameter, although wt% Cl may be used (Ketcham *et al.* 1999). The variation in kinetics with wt% Cl embodied in this model is very similar to that in the Geotrack model (above), and the two approaches give similar results when applied with similarly specified thermal history scenarios.

Either monotonic cooling or episodic heating and cooling can be employed, but in most published studies too much flexibility is built into the thermal history structure, resulting in wide ranges of allowed fits which reflect the redundancy issues discussed earlier (Fig. 24), rather than providing meaningful constraints on the thermal history. This can be avoided to some extent by specifying thermal histories in a series of simple linear segments involving heating to a maximum temperature followed by cooling, although this is rarely done. Temperature–time constraints can be imposed on the range of allowed fits, as boxes in temperature–time space through which the histories must pass. Such constraints derived from geological evidence (e.g. from overlying sedimentary units) can narrow the range of likely histories, and in some ways can be thought of as analogous to the Default Thermal History approach outlined above. However, in practice details of such constraints, together with the underlying justification, are rarely provided. This is problematical, since these constraints can exert a critical control on the nature of the resulting thermal histories.

The HeFTy approach provides solutions which are defined up to 200°C (with corresponding uncertainty limits), but this is purely an abstraction resulting from the assumed style of history, and solutions in this temperature range have no real meaning. Note also that as this approach attempts to constrain the entire thermal history below 200°C, considerable computing time and effort is wasted because of the issue of redundancy highlighted in Fig. 24.

'Monte-Trax' and related approaches: Gallagher (1995) showed how forward modelling could be combined with a 'genetic algorithm' to constrain the range of thermal histories that is consistent with the measured data and to converge on the most likely history. This approach, embodied in the 'MonteTrax' software package, has commonly been used in basement terrains and crystalline rocks to constrain the complete thermal history below *c.* 110°C assuming monotonic cooling (although this is not essential). While this approach has been superseded by more sophisticated approaches (see below), it remains of interest because it underlies several key studies of denudation histories of the southern Africa and Brazil continental margins discussed in chapter 6, which are still the subject of considerable discussion.

The kinetic model of Laslett *et al.* (1987) is commonly employed in this approach, on the assumption that the Durango apatite (Young *et al.* 1969), for which this model was developed, is representative of common apatites. In many studies using this approach, apatite fission-track data have been used in isolation as the only constraint on thermal and denudation histories. Adoption of the Laslett *et al.* (1987) model as the keystone of this approach introduces some problems, particularly with the low temperature (<60°C) aspect of thermal history solutions, which commonly show evidence of major cooling within the last 10 to 20 million years. This is commonly regarded as spurious, and attributed to inaccuracy in predictions of the model at low temperatures (e.g. Hendriks & Andriessen 2002). However, a large contribution to this effect may arise from the fact that the Durango apatite is not representative of apatite grains commonly analysed from basement rocks, which typically contain <0.1 wt% Cl. In contrast, the Durango apatite contains *c.* 0.4 wt% Cl, and is thus more resistant to annealing than typical low-Cl basement-derived apatites (Ketcham *et al.* 1999; Barbarand *et al.* 2003), which therefore show a greater degree of annealing compared to the predictions of a model based on Durango apatite. Green (2004) suggested that this can readily explain the 'anomalous

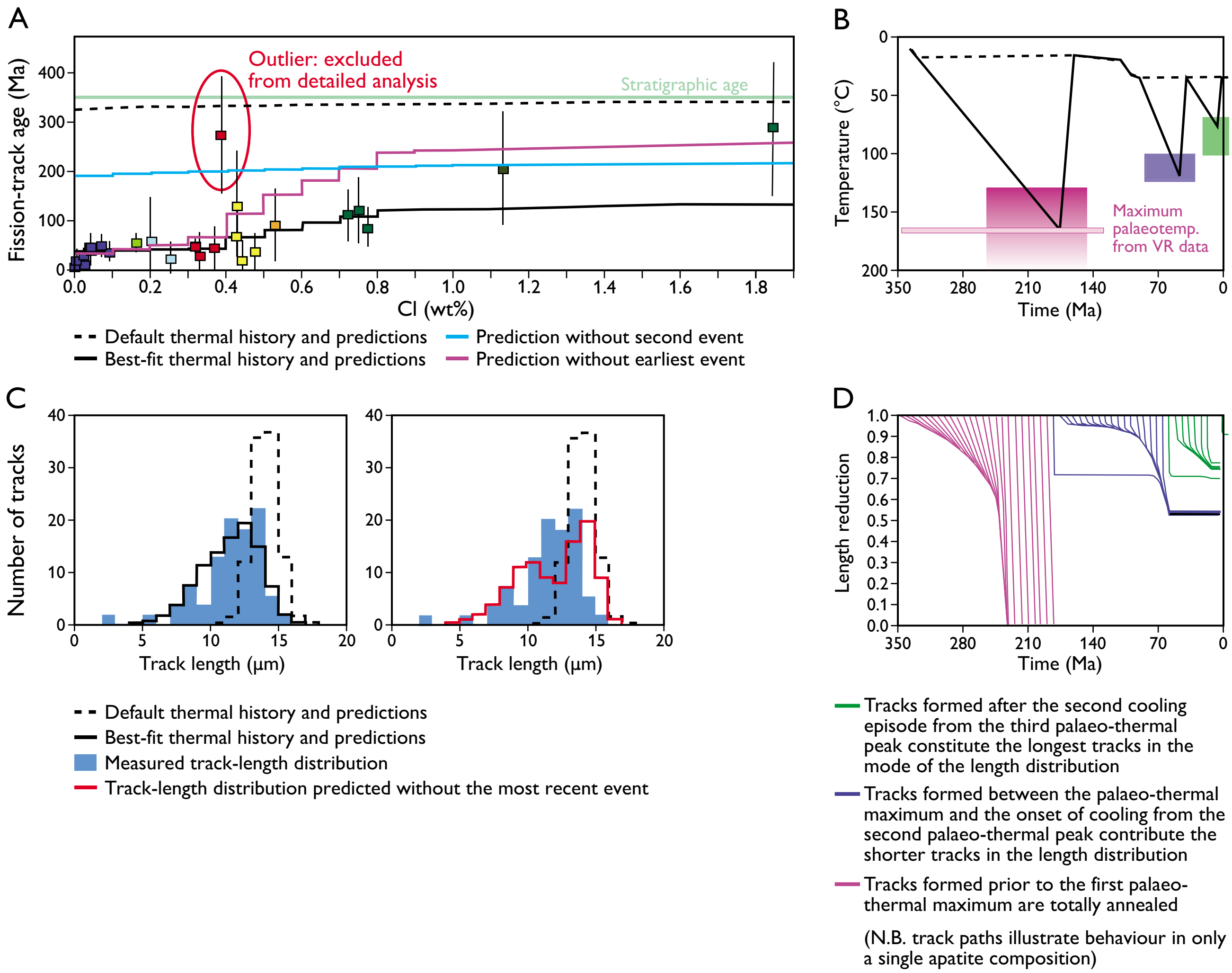


Fig. 27. Definition of three events from AFTA data in one sample. This example, based on AFTA data from the Dodo Canyon K-03 well in the Mackenzie Valley, Northwest Territories, Canada (after Green & Duddy 2013), illustrates how multiple events can be resolved from AFTA data, where the events are sufficiently separated in temperature and time. **A**: The variation of fission-track age with wt% Cl over a wide range of chlorine contents in this sample provides definition of the two earlier events (colour scale as in Fig. 26), while the most recent event is defined from track-length data (**C** and **D**). **B**: Three palaeo-thermal episodes are required to explain all aspects of the AFTA data in this sample. The best-fit thermal history results in predictions which provide an excellent fit to both the variation of fission-track age with wt% Cl (**A**) and also the measured track-length distribution (**C**). Omitting either of the two earlier events results in a failure to fit the trend of age vs. wt% Cl (**A**), while omitting the most recent event produces a poor fit to the longer end of the track-length distribution (**C**). **D**: Predicted track-shortening trajectories for the best-fit history (**B**) illustrate the development of AFTA parameters through this history. All tracks are erased in the earliest episode, which leads to resetting of fission-track ages across the range of wt% Cl. A population of shorter tracks is then produced at the second palaeo-thermal peak (note that the sedimentary section deposited after the initial cooling shows that re-heating is required). Finally a third palaeo-thermal episode produces another population of tracks shortened to a lesser degree (constituting the longer part of the main mode of the distribution). A few longer tracks are produced after this final episode, but contribute only a small proportion of tracks. **VR** (vitrinite reflectance) data provide independent verification of the magnitude of the earliest episode. Modified from Green & Duddy (2013).

Miocene cooling' commonly seen in 'Monte-Trax' solutions.

Unfortunately, despite the advent of kinetic models which explicitly include the influence of chlorine content on annealing rates (e.g. Ketcham *et al.* 1999), these have not been be incorporated into the Monte-Trax software. In an attempt to account for these problems, Gunnell *et al.* (2003) renormalised the Laslett *et al.* (1987) model. But in addition to modifying the low temperature behaviour, this also made the model more sensitive at high temperatures, and results obtained using this modified model are therefore unreliable.

A number of variations of this approach have been developed (e.g. Corrigan 1991; Lutz and Omar 1991; Willett 1997), some of which implement the Laslett *et al.* (1987) annealing model, while others have used alternative models (e.g. Crowley *et al.* 1991). However, extrapolation of these alternative models to geological timescales is not as accurate as that of the Laslett *et al.* (1987) model, and these models have not been widely adopted.

More recently Gallagher (2012) described a new approach for inverse modelling of thermal history from apatite fission-track data using a Bayesian transdimensional Markov Chain Monte Carlo approach, allowing the use of multiple samples and different styles of thermal history scenario, as well as combination of data from multiple techniques. Cogné *et al.* (2012) demonstrated the use of this approach in samples from the Brazil continental margin.

Fission-track stratigraphy: The qualitative use of apatite fission-track age patterns with depth or elevation to identify offsets between crustal blocks was pioneered in the Transantarctic Mountains by Fitzgerald & Gleadow (1987). This approach has been adopted more recently in other regions, e.g. the Pyrenees (Fitzgerald *et al.* 1999) and Scandinavia (Redfield *et al.* 2005). While this approach is largely free of any problems associated with the detailed thermal response, it fails to make full use of the data to provide quantitative thermal history constraints, and is susceptible to the influence of changes in apatite composition through the section (Lorencak *et al.* 2004). As with most of the other approaches described above, apatite fission-track data are usually used in isolation in this approach.

4.1.5 Monotonic cooling vs. episodic heating and cooling

As discussed above, thermochronology studies of basement terrains are commonly carried out within a framework involving monotonic cooling from above *c.* 110°C to surface temperatures. In such studies, sedimentary outliers are often ignored (or avoided), and relief differentiation as evidence of former cover (now removed) in basement terrains is generally not discussed. The presence of the merest veneer of sedimentary cover lying on basement provides the key constraint that the underlying basement was at the surface when that cover was deposited. Several studies have failed to take account of such basic geological constraints. Persano *et al.* (2006) reported cooling histories for outcrop samples from the Carboniferous Bathurst Batholith (eastern Australia) involving protracted cooling from *c.* 110°C in Permian times, but failed to allow for the presence of upper Permian sedimentary outliers overlying the outcropping granite. These sedimentary remnants show that the granite was rapidly exhumed following intrusion and was then reburied, which completely changes the interpretation of the data and the resulting conclusions regarding the development of the landscape. Despite this error being pointed out by Brown (2007), Gibson (2007) and Green & Duddy (2007), in their response Persano *et al.* (2007) did not acknowledge the significance of these sedimentary outliers. Similar situations where geological constraints have not been taken into account, leading to erroneous interpretations, include Johnson & Gallagher (2000; see Japsen *et al.* 2010), Persano *et al.* (2005), and Pedersen *et al.* (2012; see Japsen *et al.* 2013b). These and other examples where geological constraints have important implications for the interpretation of apatite fission-track data are discussed further in section 8.2.

While episodic heating and cooling is clearly the most realistic scenario for sedimentary basins, as witnessed by the common occurrence of unconformities in sedimentary sequences, such histories are also reasonable for basement terrains where sedimentary outliers are present. Similar comments apply where the present-day relief reveals old surfaces which have been preserved beneath a former cover (section 3.1.2). Since episodic histories apply in these basement regions, there seems to be no reason why this should not also be true in basement areas devoid of present-day cover, which could simply indicate that the former cover has been totally stripped. This is particularly the case along EPCMs where stratigraphic landform analysis suggests the presence of re-exposed peneplains.

Recognition of the presence of sedimentary outliers can lead not only to major changes in interpreted cooling rates and corresponding denudation rates, but also to more fundamental aspects of geological evolution. Weber *et al.* (2005), in a study of outcropping basement samples from the Yilgarn craton of Western Australia, showed that the preservation of thin Permian sedimentary remnants required rapid late Carboniferous to early Permian exhumation of basement to the surface and subsequent kilometre-scale reburial prior to renewed late Palaeozoic – Mesozoic exhumation. This study therefore revealed the presence of a thick former sedimentary cover over this supposedly stable cratonic region. This emphasises the common applicability of histories involving episodic cooling and reheating, even in basement regions.

The South Swedish Dome (section 3.1.2) is another key area where combining thermochronology with constraints from geology and SLA defines a history of episodic heating and cooling (burial and exhumation) through the Phanerozoic. Cambrian strata lie directly on basement in the north and east while Jurassic and mainly Cretaceous strata rest on basement in the south and west (Lidmar-Bergström 1988, figs 1, 2; Japsen *et al.* 2002, fig 15). Published apatite fission-track data (Cederbom *et al.*2000; Cederbom 2001) define late Palaeozoic cooling from temperatures around 100°C or above, reflecting erosional removal of a substantial thickness of Palaeozoic rocks deposited following formation of the Sub-Cambrian Peneplain and deposition of remnant Cambrian sediments. The apatite fission-track results also require Oligocene–Miocene cooling, following deposition of preserved Mesozoic units, implying deposition of Upper Cretaceous to Palaeogene cover. As illustrated by Japsen *et al.* (2002) this cover preserved the hilly relief surface produced during Mesozoic weathering.

Further support for the view that episodic heating and cooling provides a preferred framework for interpreting low temperature thermochronology data compared to monotonic cooling comes from a study by Flowers & Kelly (2011) of basement cores from a borehole through thin sedimentary cover overlying Precambrian basement of the US Mid-continent Region. Based on a combination of apatite fission-track data and apatite (U-Th)/He dating they showed that basement samples showed evidence of major heating and cooling within each of several major regional unconformities (Cambrian–Permian, Permian–Cretaceous and Cretaceous–Recent), defining a history of episodic heating and cooling similar to those described above. Although only *c.* 200 m of sedimentary cover represents Phanerozoic time at the borehole location in Kansas, the presence of this thin cover demonstrates that the underlying basement was not continuously exhumed through Phanerozoic time, and monotonic cooling is therefore inappropriate. Flowers & Kelly (2011) used the presence of the cover units to define periods at which the basement rocks were close to the surface. Only a slightly greater degree of erosion over the last 100 million years or so would have totally removed all the evidence that confirms this episodic history, though the real history would remain the same. This needs to be borne firmly in mind in considering likely histories for basement terrain devoid of sedimentary cover.

4.1.6 The meaning of a fission-track age

In early studies, fission-track ages from crustal sections with simple histories were commonly discussed in terms of a zonation of ages (e.g. Naeser *et al.* 1989). At shallow depths, ages were regarded as unaffected ('zone of no annealing', <70°C), while at depths greater than *c.* 3–4 km (temperatures in excess of *c.* 125°C), no tracks are retained ('total annealing zone'). Between these two extremes, fission-track ages are progressively reduced to zero through a 'partial annealing zone' (PAZ). While this zonation provided a simple conceptual basis for early studies, the combination of borehole data and laboratory experiments discussed above showed that reduction in both fission-track age and track length proceed at temperatures below 70°C, albeit more slowly than at higher temperatures. This can be seen most easily in confined track-length data due to the higher precision of these measurements compared to fission-track ages (Fig. 18).

The true meaning of a fission-track age can only be assessed in conjunction with the distribution of confined track lengths (e.g. Gleadow *et al.* 1986a, b). An apatite fission-track age will only provide a direct indication of the time over which tracks have been retained in a sample which has cooled extremely rapidly from >110°C to temperatures less than 50°C and subsequently remained at such temperatures. In such samples, track lengths will be similar to those in the age-standard apatites used to calibrate the method (Section 4.1.1). Experience has shown that such situations are relatively rare. Apatite fission-track ages therefore only rarely indicate the time over which tracks have been retained, and must be assessed in tandem with track-length data before the true significance becomes clear.

In the same way, a fission-track age rarely reflects a 'cooling age' (in the sense that cooling occurred at a particular time), and we recommend that this term should not be used. In slowly cooled terrains, ages are often referenced to a 'closure temperature', below which the daughter product is effectively retained. However, literal interpretation of this concept is of dubious validity for apatite fission-track ages because tracks formed throughout the cooling history undergo significant shortening (Fig. 19) and their contribution to the fission-track age

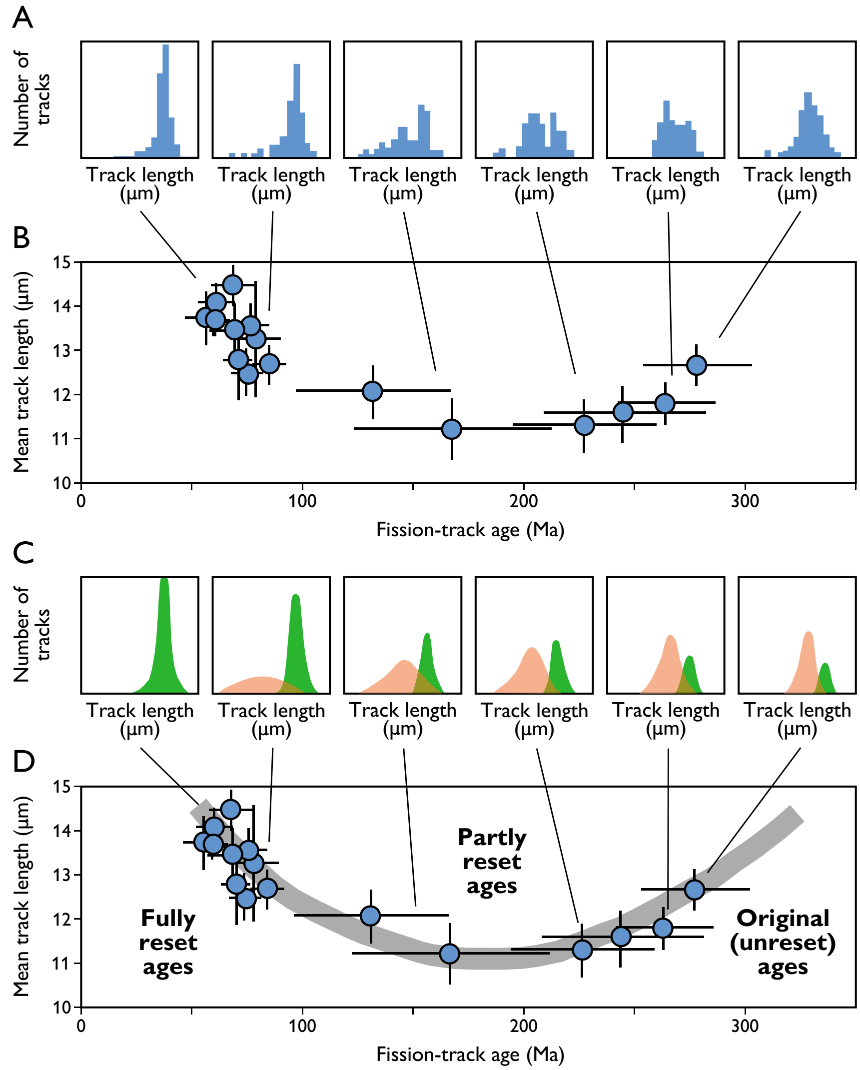


Fig. 28. Boomerang plot: relationship between mean track length and fission track age for a suite of outcrop samples from north-west England which have undergone cooling from different maximum palaeotemperatures at the same time (modified from Green 1986). **A**: Measured track length distributions. **B**: measured data. **C**: illustration of track-length distributions as mixtures of two populations. **D**: Illustration of the evolution of AFTA parameters with increasing maximum temperature (from right to left). Samples that experienced low maximum temperatures have old ages with relatively long mean track lengths, while samples in which all fission tracks were totally annealed prior to the onset of cooling give much younger fission-track ages ('reset ages'), also characterised by long (*c.* 14 μm) mean lengths. Between these two extremes, as the fission-track age decreases (representing increasing maximum palaeotemperatures prior to the onset of cooling) the mean track length decreases as the partially annealed tracks are progressively shortened. This continues until the final stages of age reduction, when the partially annealed tracks become so short that their contribution to the mean length is diminished and the mean length increases with further reduction in fission track age, trending upwards towards the long mean length characterising the reset ages. This dataset can be considered analogous to the situation illustrated in Fig. 21, but with all samples cooling to low (near surface) temperatures, such that each contains a population of long tracks formed after cooling.Modified from Green *et al.* (2002) and Green & Duddy (2013).

will consequently be reduced compared to the overall time over which tracks have been retained.

In general, an apatite fission-track age should be regarded simply as a thermal history parameter, representing the integrated balance between the production of tracks by spontaneous fission and the reduction in the probability of tracks intersecting a polished and etched grain surface due to the reduction in track length resulting from the thermal history (Figs 20, 21).

4.1.7 'Boomerang plots'

The meaning of an apatite fission-track age is further illustrated by results from north-west England. Green (1986) reported results in samples from Palaeozoic intrusions, many of which were unroofed by the end of the Devonian, and which have subsequently undergone essentially a single dominant episode of heating and subsequent cooling. The results define a systematic relationship between mean confined track length and fission-track age (Fig. 28), defining a boomerang-like trend. Samples that have undergone little thermal disturbance since the Palaeozoic have old ages with relatively long mean track lengths, while samples in which all fission tracks were totally annealed prior to the onset of cooling gave much younger fission-track ages ('reset ages'), also characterised by long (*c.* 14 µm) mean lengths. Between these two extremes, as the fission-track age decreases (representing increasing maximum palaeotemperatures prior to the onset of cooling), the mean track length decreases as the partially annealed tracks are progressively shortened. This continues until the final stages of age reduction, when the partially annealed tracks become so short that their contribution to the mean length is diminished and the mean length increases with further reduction in fission-track age, trending upwards towards the long mean length characterising the reset ages. This trend can be viewed as analogous to the depth trend in Fig. 21, except that all samples have a common long component because they all cooled to near-surface temperatures in the case of a suite of outcrop samples.

In contrast, age vs. length data from many other regions show a very different type of trend to the simple pattern reported by Green (1986). For example, results from Norway (Rohrman *et al.* 1995) show an almost opposite relationship to the classic 'boomerang' trend, while data from Africa and Brazil (Gallagher & Brown 1999b) show wide dispersion with only a slight tendency towards longest lengths associated with the youngest ages. Compared to the simple situation in north-west England where a single dominant heating/cooling episode has produced a well-defined trend, such relationships imply a much more complex history, most likely involving a series of cooling episodes, each of which may vary in magnitude across the region. Therefore, such plots should be interpreted with care, and detailed attention should be paid to the systematic change in the form of the track-length distribution through the plot, which was central to the original description by Green (1986), but which is often overlooked in other studies.

Gallagher & Brown (1997) reiterated the usefulness of plotting mean track length vs. fission-track age in considering the implications of regional apatite fission-track datasets, but within the context of monotonic cooling histories. This can lead to misconceptions, as explained in section 4.1.9.

4.1.8 Uplift rates from apatite fission-track age profiles

In early studies, the increase of fission-track age with elevation in mountainous regions was interpreted in terms of monotonic cooling, representing the progressive closure of the system and retention of fission tracks as samples moved upwards in the rock column and cooled below the closure temperature. Such age patterns were used to determine rates of uplift and erosion (e.g. Wagner & Reimer 1972) based on the assumption that the change in apatite fission-track age with elevation represented slow cooling associated with progressive denudation of the mountain range. On this basis, segments of the age vs. elevation trend with different slopes were interpreted as representing a change in the rate of denudation. However, subsequent integration of confined track-length information with the fission-track age data (Fitzgerald & Gleadow 1987) showed that the upper segment of such trends above the break- in-slope in fact represents an uplifted (exhumed) partial annealing zone (cf. Fig. 21), while samples below the break-in-slope were totally annealed prior to the onset of exhumation. Thus, the slope of the age vs. elevation profile cannot be interpreted simply in terms of slow cooling. Instead, the decrease with age through the upper section towards the break-in-slope represents increasing degrees of annealing of deeper samples prior to the onset of cooling.

4.1.9 Long term residence in the partial annealing zone vs. heating and cooling

It is often asserted that the presence of an exhumed partial annealing zone, as identified by a break-in-slope in the variation of fission-track age with elevation or depth (Fig. 21), represents a prolonged period of residence in the partial annealing zone prior to exhumation. This again reflects an attitude framed in terms of monotonic cooling. Consideration of Fig. 21 shows that long-term residence is not necessary, as data of this type are easily produced by heating (e.g. by burial) of the sequence to temperatures characterising severe annealing followed immediately by rapid subsequent cooling/exhumation. This misunderstanding reflects the viewpoint of many workers that monotonic cooling histories form the most appropriate framework for interpreting apatite fission-track data. In contrast we submit that episodic heating and cooling histories are more relevant in many situations (cf. Japsen *et al.* 2010; Green & Duddy 2010).

4.1.10 Limitations of apatite fission-track methods

As noted above, failure to recognise the limitations of the apatite fission-track system is a common cause of errors in interpreting the data. One major limitation is the fact that the data are sensitive only to key moments in the overall history. For episodic heating and cooling histories, such as illustrated in Figs 20 and 21, it should be clear that the data contain no information on the variation of temperature during the heating phase prior to the onset of cooling. It is for this reason that application of AFTA is focused specifically on determining the magnitude of the maximum temperature and the time at which cooling begins (section 4.1.4). In addition, factors such as the inherent spread in the width of the track-length distribution, resulting from the energetics of the fission-process (Green 1980), and the number of tracks counted in determining a fission-track age impose fundamental limits to the temperature and time resolution that can be achieved. A resolution of 10°C in temperature is possible for the 60 to 90°C range (equivalent in broad terms to a resolvable difference of around 1 μm in mean track length over this temperature range) while 5°C may be achievable between 90 and 110°C, due to increasingly rapid decrease in fission-track age. For samples that have been heated to less than 60°C, it is often possible only to set an upper limit to the magnitude of heating, particularly in sedimentary rocks which contain tracks formed prior to deposition. In samples that have been heated to temperatures above *c.* 110°C where tracks are totally annealed, the history prior to cooling below this temperature can of course not be constrained. Poissonian uncertainties in fission-track age (typically around 5% at best) derived from the numbers of tracks counted set a fundamental limit to resolution in time.

The consequence of these considerations is that for histories involving a series of heating and cooling episodes, AFTA will reveal only the major episodes, and may give solutions that represent the unresolved effects of multiple episodes. Solutions framed in terms of monotonic cooling, in situations where the real histories involved a series of episodes of heating and cooling, will represent only a broad approximation to the dominant palaeothermal episodes.

Given these factors, combined with the redundancy in the data (Fig. 24), it should be appreciated that in all approaches to extracting quantitative thermal history constraints from apatite fission-track data, the full detail of the underlying thermal history can never be defined. The resulting thermal history solution from any approach should therefore be considered as providing an approximation to the true history and the skill in using these techniques is to recognise the range of possible solutions that can be accommodated by a given set of data. In this context, it should be appreciated that ranges of uncertainty attributed to cooling histories derived from other approaches described in section 4.1.4 are contingent on the assumed form of the history (usually monotonic cooling), and alternative styles of history may be possible (e.g. episodic heating and cooling) for which other uncertainty limits will apply.

4.1.11 Alternative views of fission-track annealing

While the role of thermal annealing in controlling the evolution of apatite fission-track age and length is widely acknowledged, some workers have suggested that other processes can also influence data in some circumstances. Wendt *et al.* (2002) suggested that contrary to previous ideas (e.g. Fleischer *et al.* 1975) pressure could play an important role in affecting the kinetics of the annealing process. Kohn *et al.* (2003) criticised many aspects of the ideas presented by Wendt *et al.* (2002), and despite contrary arguments by Vidal *et al.* (2003), the idea that pressure can affect fission-track annealing kinetics is not a widely held view. We note that if pressure were to have a significant effect on annealing rates, it might be expected to be most obvious in results from deep drill holes in cratonic regions characterised by low thermal gradients, but results to date from such regions (e.g. Rohrman 1995; Lorencak *et al.* 2004) show no such evidence.

More recently, Hendriks & Redfield (2005) suggested that apatite fission-track ages in Finland were anomalously young because of a non-thermal annealing process due to the effects of a long-term radiation dose from alpha decay over millions of years. This suggestion, if true, has severe implications for the interpretation of apatite fission-track data. One of the lines of evidence used to justify this claim was the observation that apatite fission-track ages were younger than (U-Th)/He ages in the same apatites, contrary to expectation (see next section). But Green *et al.* (2006) showed that this is due to anomalous behaviour in the (U-Th)/He system, and not to anomalous fission-track annealing behaviour. Other evidence used to support the notion of 'radiation-enhanced' fission-track annealing can be interpreted in conventional fashion in terms of standard processes of thermal annealing (Green & Duddy 2006; Larson *et al.* 2006) and the concept of radiation-enhanced annealing currently finds little support.

4.1.12 Summary

The thermal response of fission-tracks in apatite is well understood (at least empirically), and provides results which are consistent with constraints from other techniques when applied in tandem (section 4.3). Despite often relatively low precision on thermal history constraints, the technique provides unique information yielding interpretations that cannot be obtained from other methods. While approaches used by different groups may differ in detail, the response of the system is well defined, and thermal history information provided by the technique should be reliable within the constraints and limitations of the methods used. Perceived problems with the technique, as discussed for example by Gunnell (2000), are likely to result from a failure to appreciate these limitations, rather than any fundamental problem with the technique itself. In many cases too much is expected from the technique. When attention is focused on the unique information that can be obtained, and results are integrated with regional geological constraints, the power of the technique becomes apparent.

4.2 Apatite (U-Th)/He dating

4.2.1 Historical background

The development of apatite (U-Th)/He dating (Zeitler *et al.* 1987; Wolf *et al.* 1996) potentially opened up a lower temperature window for thermochronology. Whereas fission-track dating is based on spontaneous fission-of ^{238}U atoms, (U-Th)/He dating is based on the alpha decay of uranium and thorium isotopes (and to some extent samarium, e.g. Grist & Zentilli 2005). Alpha particles, being helium nuclei, are neutralised rapidly after emission from the parent nucleus, and helium gas accumulates in the apatite lattice at a rate dependent on the U and Th content of the apatite. But once formed, the He gas is progressively lost from the apatite due to diffusion at a rate which depends on temperature, in similar fashion to the annealing of fission-tracks. This balance between the production and loss of He forms the basis of this technique.

Initial studies (e.g. House *et al.* 1997; Warnock *et al.* 1997; Wolf *et al.* 1997) illustrated the potential of the technique to provide useful information at temperatures in the 40–80°C range, possibly giving this method even greater potential than apatite fission-track methods for providing information relevant to the interpretation of landscape development, and suggesting that the combination of the two techniques should be particularly powerful.

4.2.2 Early success

Early applications of the method to elucidate thermal history information in geological conditions were based on kinetics of He diffusion in apatite and the way that this depends on temperature and grain size derived from laboratory experiments on Durango apatite (Farley 2000). In similar fashion to apatite fission-track methods, comparison of measured ages with values predicted from various thermal history scenarios allows definition of the range of thermal histories giving predictions that are consistent with the results. Comparison of (U-Th)/He ages with apatite fission-track data from boreholes in the Otway Basin, south-east Australia (House *et al.* 1999, 2002) illustrated the unique sensitivity of (U-Th)/He ages in apatite to temperatures in the range 20 to 80°C, and appeared to confirm the extrapolation of laboratory studies. Integration of apatite (U-Th)/He ages with AFTA data from the Fresne-1 well in the Taranaki Basin, New Zealand (Crowhurst *et al.* 2002) provided further demonstration of the consistency between the two techniques, and illustrated the potential power of the combination of the two methods when applied to vertical sequences of samples, where the variation of (U-Th)/He age with depth helps to further restrict the range of viable solutions defined from AFTA (Fig. 29).

Apatite (U-Th)/He dating has now been applied with apparent success in a wide range of environments from mountain belts to sedimentary basins. But while it was initially assumed that the He diffusion systematics from Farley (2000) could be applied to all common apatites, it

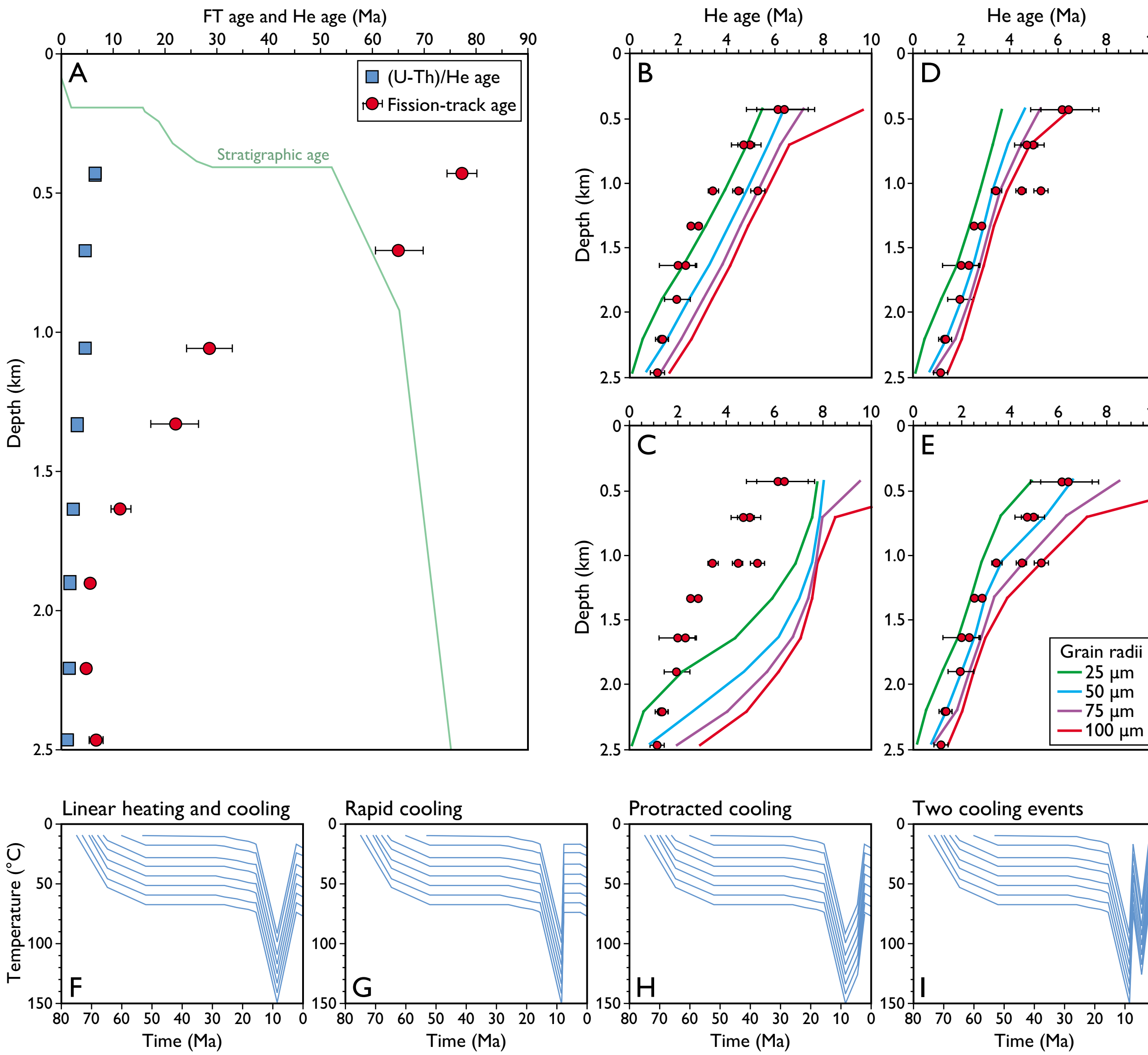


Fig. 29. Apatite (U-Th)/He and fission-track ages in samples from the Fresne-1 well, Taranaki Basin (New Zealand). **A**: Measured (U-Th)/He ages are much younger than fission-track ages in the same samples. **B, C, D, E**: Modelling the variation of (U-Th)/He age with depth using four different thermal history scenarios (**F, G, H, I**), based on the interpretation of AFTA data from these samples, allows refinement of the preferred thermal history reconstruction, favouring a scenario involving two-stage inversion (**E, I**). **FT**: Fission-track. Modified from Crowhurst *et al.* (2002).

has become clear more recently that this is not the case, as explained below.

4.2.3 Evidence of greater complexity

Despite the early success, later studies (e.g. Fitzgerald *et al.* 2006; Hansen & Reiners 2006; Danisik *et al.* 2008) showed increasing evidence of excess dispersion in apatite (U-Th)/He ages over what could be explained in terms of known controls (principally grain size). In addition to this random dispersion, studies in which (U-Th)/He data were integrated with AFTA (Green & Duddy 2006; Green *et al.* 2006) revealed a systematic discrepancy between the two techniques which increased as either the (U-Th)/He or the fission-track age increases (Fig. 30). Green *et al.* (2006) interpreted this trend to

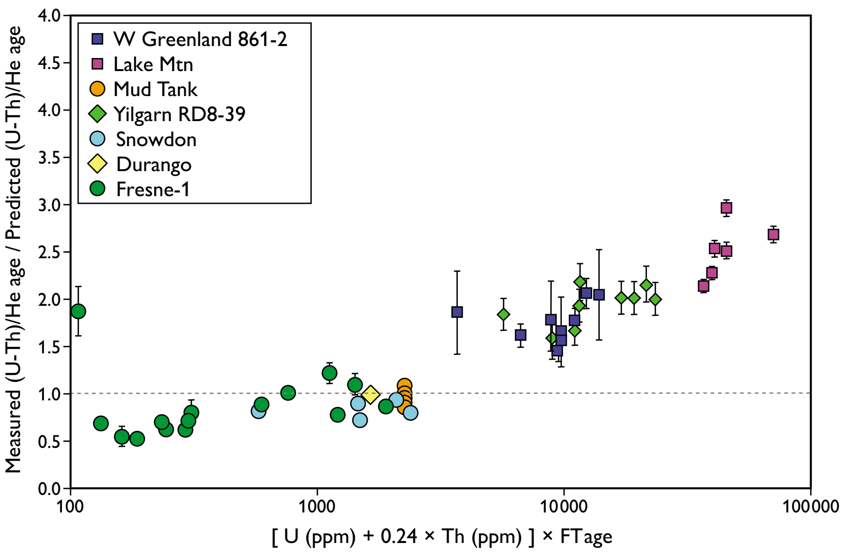


Fig. 30. Systematic change in He retention characteristics of apatite with increasing radiation- damage levels. The ratio of measured apatite (U-Th)/He age to the value predicted from the best-fit T(t) solution derived from AFTA for each grain from a number of samples with well-defined thermal histories is plotted against the product of the measured fission-track age and a function of uranium and thorium contents of each grain (after Green *et al.* 2006). This term provides an approximate measure of the number of alpha particle decays from U and Th over a time equal to the measured fission-track age. These results show a progressive departure from consistency, highlighting the enhanced He retentivity of apatite as the degree of radiation damage in the crystal lattice increases (Green *et al.* 2006). **FT**: Fission-track.

reflect enhanced He retentivity due to increasing levels of radiation damage in the apatite lattice, which results in anomalously old (U-Th)/He ages compared to the expected response based on the Farley (2000) systematics in samples with ages older than *c.* 50 Ma. Shuster *et al.* (2006) reported laboratory diffusion studies which confirmed the increase in helium retentivity as the amount of radiation damage in the apatite lattice increases.

Subsequent developments by Shuster & Farley (2009) and Flowers *et al.* (2009) have led to a more advanced understanding, in which modelling of He diffusion is integrated with modelling of fission-track retention as a surrogate for the degree of radiation damage (the 'RDAAM' model). In this model, the fission-track density is adopted as a proxy for the degree of radiation damage, and elimination of this damage is governed by fission-track annealing kinetics. This suggests that the apatite (U-Th)/He method is best employed in tandem with apatite fission-track data. Gautheron *et al.* (2009) provided further documentation of the influence of radiation damage in increasing He retentivity in apatite, and also offered an alternative quantitative model.

These approaches offer considerable promise for providing more consistent thermal history solutions from multiple methods, but as illustrated by Flowers & Kelly (2011), detailed investigation is required to ensure that the data conform to the behaviour expected by the model that is used to extract thermal history information from the data. Flowers & Kelly (2011) provided guidelines for the interpretation of 'over-dispersed' apatite (U-Th)/He data, including evaluation of the relationships between (U-Th)/He age and equivalent uranium content and grain size. Where such relationships are clear they can be used to control viable thermal history solutions. Where no such relationships exist, Flowers & Kelly (2011) recommended that such data should not be used for thermal history interpretation and further data should be acquired to explain the dispersion of the results. For datasets where such detailed investigation has not been undertaken, we suggest that apatite (U-Th)/He results should be treated with extreme caution, particularly when no apatite fission-track data are available from the same samples.

The results presented by Green *et al.* (2006) show that (U-Th)/He ages in samples where the product of the age (in Ma) and the equivalent uranium content in ppm is greater than *c.* 2500 cannot be explained in terms of the Farley (2000) systematics (Fig. 30), which should only be applied to young, low-uranium apatites. Note that this applies even though these older ages often show a high degree of reproducibility.

In the absence of quantitative modelling, (U-Th)/He ages could, in principle, be used in a qualitative sense, similar to the 'fission-track stratigraphy' approach to apatite fission-track studies described above, by comparing age with elevation trends. But in addition to the systematic effects described by Green *et al.* (2006), the excessive scatter in He ages, beyond that expected from purely analytical uncertainties described earlier, can cause serious problems in comparing He ages from different samples.

Even in the absence of such effects, because of the systematic effect of increasing levels of radiation damage, it is essential that only apatites with similar U contents and grain size should be compared.

4.2.4 He-closure temperatures and 'cooling ages'

In some studies, (U-Th)/He ages have been interpreted as representing the time at which samples cool through a closure temperature, typically around 50°C although in detail this will vary with radiation damage (Shuster *et al.* 2006). But as highlighted by Green & Duddy (2006), open system behaviour is to be expected in the upper 2 km of the crust, and long-term residence at low temperatures, or even histories involving re-burial followed by exhumation, can produce results that can be easily mistaken for slow cooling. In practice, interpretations of apatite (U-Th)/He ages are most reliable when accompanied by geological constraints (e.g. sedimentary outliers) and apatite fission-track data interpretations.

4.3 Converting thermal history information to denudation history

4.3.1 Denudation histories

In principle, conversion of thermal histories to denudation or exhumation histories is simply a matter of dividing the thermal history (actually the difference between the sample temperature and the appropriate surface temperature) by the appropriate value of thermal gradient throughout the history. This provides the variation through time of the depth of the sample with respect to the surface. In a continuous cooling scenario, this is referred to as *the denudation or exhumation history*.

Many studies explicitly set out to determine long-term denudation histories of basement regions over hundreds of millions of years, within a framework of continuous cooling. In practice, this process is not so simple, because as noted in section 4.1 the thermal history is only controlled at certain key points in the history (Fig. 24). Outside these times, the thermal history is effectively unconstrained and a wide range of alternative temperatures are allowed by the data. For this reason, rigorous control on thermal gradients is usually possible only at limited times in the history. Even if a constant heat flow is assumed, because the thermal history is effectively constrained only at key times the same is true of the resulting denudation history. Long-term denudation rates derived from such studies therefore have little meaning.

4.3.2 Palaeogeothermal gradients and removed section

Palaeotemperatures in a specific palaeothermal episode derived from AFTA data in samples over a range of depths (in sub-surface samples) or elevations (in outcrop samples from vertical rock sections) can be used to provide direct constraints on the palaeogeothermal gradient at the palaeothermal maximum. Extrapolation of the fitted gradients to an appropriate palaeosurface temperature then provides an indication of the amount of additional section that was present at the palaeothermal maximum (Fig. 31A). Bray *et al.* (1992) described how statistical procedures can be used to define the range of palaeogeothermal gradients and amounts of missing section that are consistent with palaeotemperature constraints within 95% confidence limits. The inverse correlation between these two parameters results in a hyperbolic ellipsoid region of allowed values (Fig. 31B).

The accuracy of estimates of removed section that can be derived using this approach, within the available uncertainty limits, has been confirmed in a number of different situations where independent estimates are available of the amount of former cover that has been removed (e.g. Green *et al.* 1995; Crowhurst *et al.* 2002; Japsen *et al.* 2007a).

It should be stressed that estimating amounts of removed section by extrapolating a linear palaeotemperature profile assumes that the additional section had the same average thermal conductivity as the preserved section. If independent evidence suggests that this assumption is not appropriate, then a more detailed analysis using suitable thermal conductivities is required in order to provide a more accurate solution (Fig. 32). But experience suggests that in most cases sedimentary sections are so heterogeneous that any influence of varying thermal conductivity is 'smoothed out' and a linear approximation for the palaeogeothermal gradient profile is acceptable.

It should also be emphasised that the approach illustrated in Fig. 31 only provides a constraint on the palaeogeothermal gradient at the palaeothermal maximum,

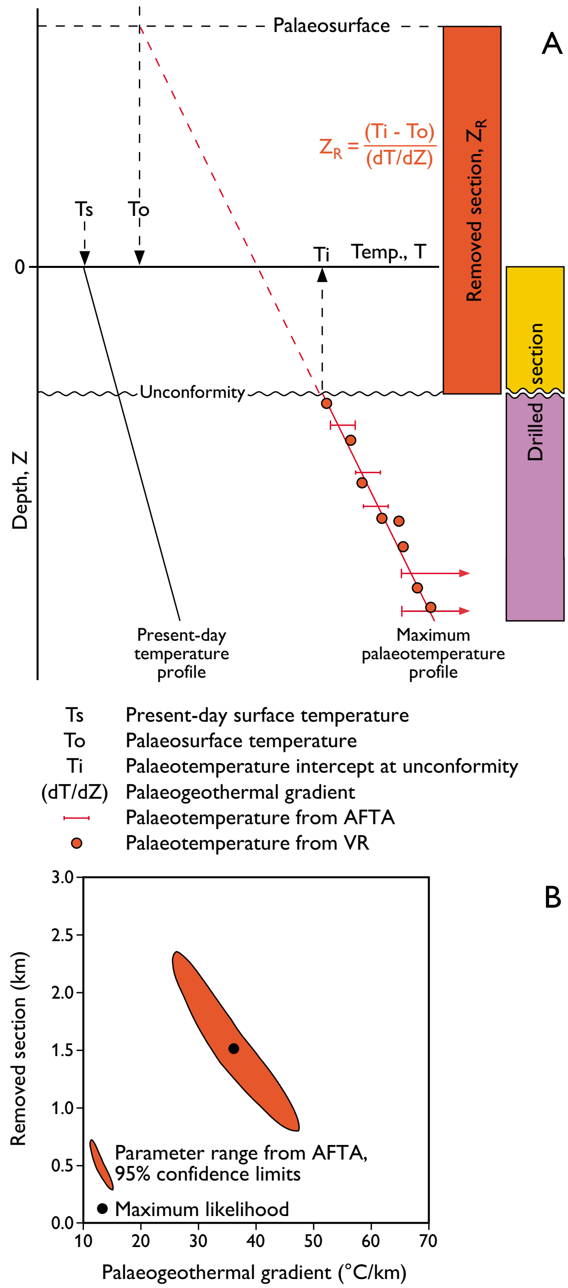


Fig. 31 Estimation of the amount of removed section from palaeothermal data. **A**: Determination of palaeotemperatures over a depth range allows definition of the range of palaeo-geothermal gradients. Extrapolation of the allowed range of palaeo-geothermal gradients then allows estimation of the amount of additional section required to explain the palaeotemperatures (equal to the amount of section that has been removed by erosion). **B**: Higher palaeo-geothermal gradients within the allowed range will require lower amounts of removed section, and lower palaeo-gradients correspond to higher amounts of removed section. Statistical procedures provide definition of the allowed range of both parameters at the 95% confidence level, as shown. Note that this analysis assumes that the thermal conductivity of the eroded sequences was the same as that of the preserved sequences. **VR**: vitrinite reflectance. Modified from Green *et al*. (2002).

and variation through time is largely unconstrained, unless multiple palaeothermal episodes can be recognised from AFTA (e.g. Green *et al.* 2004; Japsen *et al.* 2007a; Turner *et al.* 2008). Nevertheless, having a direct constraint on the palaeogeothermal gradient at the palaeothermal maximum can provide unique insights into mechanisms of heating and cooling (Bray *et al.* 1992).

4.3.3 Elevated heat flow on continental margins

The approach outlined above has provided abundant evidence of significantly elevated palaeogeothermal gradients in a wide variety of settings, e.g. south-east Australia (Duddy 1994 1997; Green *et al.* 2004), NW England (Green 2002) and West Africa (Bray *et al.* 2002; Turner *et al.* 2008), implying elevated basal heat flow. In contrast, many apatite fission-track studies, especially of continental margins, are characterised by the explicit assumption that palaeogeothermal gradients do not change appreciably through time, and that denudation is the only process capable of producing significant cooling in such areas (e.g. Gallagher & Brown 1999a, b; Brown *et al.* 2000; Gunnell 2000). This assumption is clearly erroneous, and can lead to significant overestimation of amounts of denudation. In addition to changes in basal heat flow, transport of heat due to hot fluid circulation is also a common feature in many areas (e.g. Duddy *et al.* 1994, 1998).

Gallagher *et al.* (2005) suggested a methodology for providing rigorous constraints on palaeogeothermal gradients and amounts of removed section in similar fashion to Bray *et al.* (1992). However, after applying such methods in Namibia, Raab *et al.* (2005) still assumed a constant palaeogeothermal gradient, despite a wide range of allowed values. As discussed in a later section, we suggest that failure to allow for elevated palaeogeothermal gradients has contributed to the inconsistency between denudation histories based on thermochronology and ideas on landscape development in the past.

4.4 Other methods for constraining removed section and denudation histories

4.4.1 Introduction

While many studies are based on application of apatite fission-track data in isolation, significant advantages can be obtained by combining AFTA data with results from other methods. One reason for this is the inherent redundancy in the method (Fig. 24), in that a large number of thermal histories can result in very similar AFTA parameters. In addition, the natural spread in the track-length distribution imposes limits on the recognition of low temperature events from AFTA, and can cause problems in resolving complex histories involving multiple cooling episodes. Integration with independent techniques not only provides corroboration of conclusions derived from AFTA, but can also refine the range of thermal history solutions defined from AFTA alone and provide a coherent thermal history framework. Here we discuss some of the most frequently used techniques that have been used together with AFTA and discuss the benefits that can be obtained.

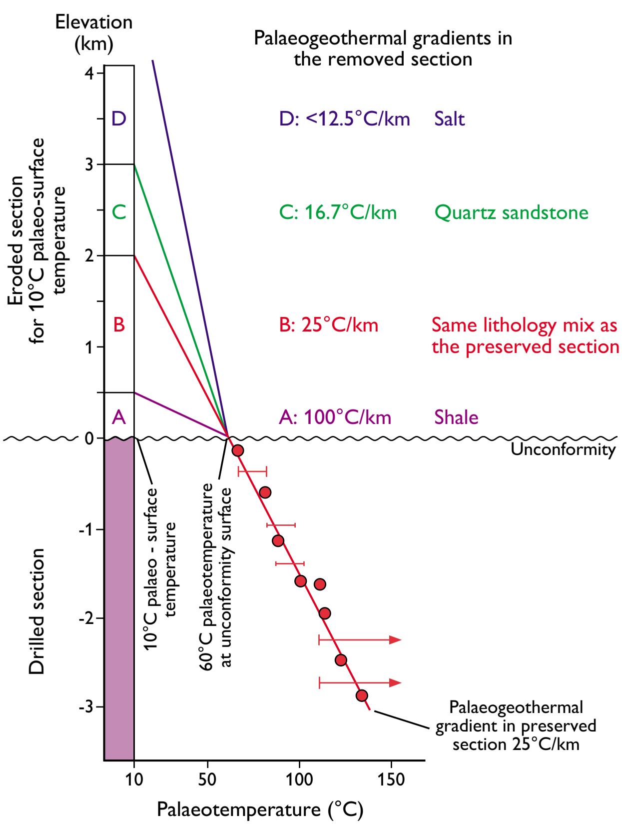


Fig. 32. Influence of the lithology of the missing section on estimation of removed thickness. Differences in thermal conductivity between the removed section and the (deeper) preserved section can produce significant non-linearity in the palaeotemperature profile, which will result in unreliable estimation of the amount of removed section using the construction shown in Fig. 31. Only where the removed and preserved sections are identical will the thermal gradient be the same throughout the entire section. However, in practice the assumption of linearity appears to give reliable results in a wide variety of situations. **VR**: vitrinite reflectants. Modified from Green *et al.* (2002).

4.4.2 Vitrinite reflectance

Vitrinite reflectance (VR), based on the increase in reflectivity of the organic maceral vitrinite (a key constituent of coal) with temperature, is the standard measure of organic maturity for hydrocarbon exploration (e.g. Tissot & Welte 1984). The kinetics of this process are well understood (Burnham & Sweeney 1989), and are very similar to those of fission-track annealing in apatite (Duddy *et al.* 1994, 1998), with VR values of 0.6% to 0.7% corresponding to total annealing of fission-tracks in typical apatites (Duddy *et al.* 1994). These factors make VR an ideal complement to AFTA data applied to sedimentary sequences, as demonstrated in a wide range of studies (e.g. Duddy 1997; Green *et al.* 2004; Japsen *et al.* 2005, 2007a, 2012b; Turner *et al.* 2008).

Integration of VR data from fine-grained units with AFTA data from sandstones also allows determination of palaeotemperatures over a wider range of depths than possible from AFTA alone. The combination of both techniques can provide much tighter control on palaeogeothermal gradients and amounts of removed section than would be possible from either technique on its own. In addition, integration of VR with AFTA data can be of great assistance in confirming earlier events soon after deposition, which may not be confidently defined from AFTA alone (e.g. Green *et al.* 2004).

4.4.3 Zircon fission-track analysis and zircon (U-Th)/He dating

Zircon is another common uranium-bearing detrital mineral amenable to fission-track analysis. From both laboratory annealing studies (Tagami *et al.* 1996) and geological evidence (Hurford 1986), fission-tracks in zircon are known to be more resistant to annealing than fission-tracks in apatite, with a closure temperature around 240°C (Hurford 1986) being generally accepted. Investigation of zircon data in samples with different levels of VR suggests significant fission-track age reduction only occurs in zircon at maximum palaeotemperatures in excess of 250°C (P.F. Green, unpublished data). Therefore zircon fission-track analysis (ZFTA) can provide information only where temperatures of this order have been reached, and are of less relevance to landscape studies than AFTA and apatite (U-Th)/He dating, although ZFTA data can be useful for providing information on sediment provenance.

The general applicability of (U-Th)/He dating of zircon remains uncertain at present. Initial studies of this technique suggested a thermal sensitivity intermediate between AFTA and ZFTA (Reiners 2005). More recent studies (Guenther *et al.* 2013) have shown that radiation damage has a profound effect on the He retentivity, which may bring the sensitivity of low-U and high U zircons closer to that of AFTA while zircons with intermediate uranium contents have sensitivities closer to ZFTA. However, this model remains largely untested with field data. Thus, while this technique has the potential to be useful in future studies of landscape development, the quantitative response of the technique remains uncertain and further work is required in order to fully understand the quantitative thermal response of (U-Th)/He ages in zircon.

4.4.4 Estimating palaeoburial using sonic velocity data

The progressive compaction of sediments with increasing burial has been widely used to determine former burial depths in exhumed basins (e.g. Marie 1975; Bulat & Stoker 1987; Hillis 1995; Japsen 1998, 2000). Comparison of compaction proxies, such as sonic velocity, in an exhumed formation with a reference curve defining the expected variation with depth in sequences of similar lithology at maximum burial depth, provides an indication of the amount of net exhumation. Selection of appropriate reference curves has been problematical in some areas, leading to erroneous conclusions regarding the extent and magnitude of exhumation (Cope 1986). More rigorous definition of the necessary reference curves for specific lithologies in recent years (Japsen *et al.* 2007b) allows more reliable estimation of former burial depths, and integration of such data with constraints from AFTA and VR data has provided consistent reconstructions of eroded section in different settings, including the North Sea Basin (Japsen *et al.* 2007a) Cardigan Bay, Western UK (Holford *et al.* 2005), Brazil (Japsen *et al.* 2012b) and the Otway Basin, south-eastern Australia (Tassone *et al.* 2013).

Since these compaction-based methods are controlled primarily by maximum burial depths (effective stress), in situations where they provide consistent indications of former burial depths with those derived from palaeothermal methods such as AFTA and VR, the results can be regarded with great confidence. In addition, while results from AFTA and VR can often be explained by a range of palaeogeothermal gradients and amounts of removed section, additional constraints from compaction-based methods can significantly reduce the range of viable solutions, as illustrated for example by results from the Hans-1 well, offshore Denmark (Japsen *et al.* 2007a).

4.5 Constraints from basic geological data and stratigraphic landscape analysis

In addition to other analytical techniques, as discussed above, one form of information that is essential in ensuring that results obtained from low temperature thermochronology and these other techniques is meaningful, but is often overlooked, is basic geological data. AFTA data in isolation can often be explained by such a wide range of histories (e.g. Fig. 24) that unless some independent constraints on the history can be defined, little or no meaningful thermal history information can be obtained. For sub-surface samples, interpretation of AFTA data begins by construction of a default thermal history (section 4.1.4), which is the history that can be reconstructed based on the preserved stratigraphic section and present-day thermal regime. This provides a context within which information from techniques such as AFTA and VR can be assessed (see e.g. Green *et al.* 2004; Japsen *et al.* 2007a for more details). For outcrop samples of sedimentary rock, two points of the temperature–time history provide a framework within which the post-depositional history can be evaluated: the depositional age (a time at which the apatite was demonstrably at the surface) and the present-day temperature. Similar principles apply to basement immediately underlying sedimentary units. While this may seem straightforward, several examples have been cited earlier (section 4.1.5) where such basic constraints have been ignored.

Information provided by SLA (chapter 3) can also be used to aid in interpretations of AFTA data, for example by recognising the presence of re-exposed peneplains indicating prior exposure at the surface (section 8.2), generations of incised valleys which may indicate staged uplift (section 5.5) and/or near-horizontal peneplains that can be used as stratigraphic markers (section 6.4.4).

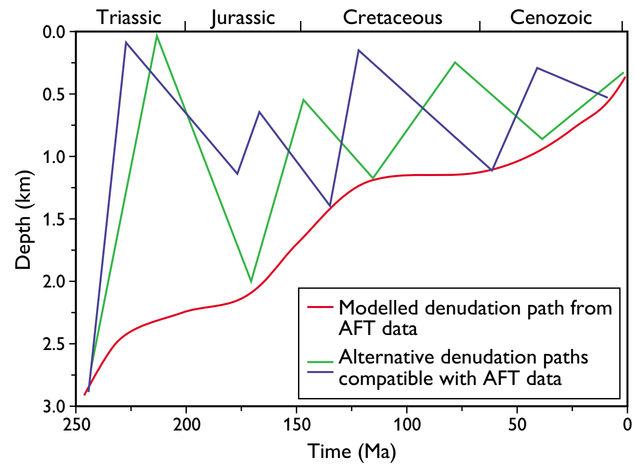


Fig. 33. Monotonic vs. episodic cooling histories. Given the high degree of redundancy in apatite fission-track (**AFT**) data, as illustrated in Fig. 24, it is not possible to discriminate between the continuous denudation history, shown in red, from possible alternative histories such as those shown in blue and green, involving episodic burial and exhumation. For this reason, average denudation rates integrated over the entire history have very little meaning, and the data may equally well be explained in terms of a number of episodes involving much more rapid denudation in each.

4.6 Summary

Derivation of thermal history information from fission-tracks in apatite, as well as other methods of low-temperature thermochronology, is not straightforward. Different approaches have been developed, each with their own advantages and limitations. But whichever approach is adopted, only the broad form of the history can be defined, and the degree to which the temperature–time, or T(t), solution approaches the real history depends on the complexity of the history. It is therefore essential to appreciate the limitations of the technique.

Interpretations of data in basement outcrop samples which assume monotonic cooling are problematical because the high degree of redundancy in the data results in a wide range of allowed histories. Such histories also often conflict with geological evidence. Basement samples overlain by or adjacent to sedimentary cover (however thin it may be) can provide a framework defined by geological constraints, resulting in more reliable histories. Evidence from SLA for periods of denudation and preservation of surfaces (by burial) can provide further constraints. Results from a wide range of settings provide consistent evidence of multiple episodes of burial and exhumation (heating and cooling), and we suggest that this type of scenario provides a more suitable paradigm for interpreting data.

While constraining thermal histories is difficult, defining denudation or exhumation histories (effectively palaeodepth through time) is even more challenging. In addition to uncertainty in thermal histories discussed above, factors such as variation in basal heat flow through time and other forms of heat transport (especially hot fluids, igneous intrusions) introduce further complications (although application of multiple techniques can assist to some degree). Because thermal history solutions are dominated by certain key points in the history (Fig. 24), the same is true of denudation/exhumation histories (Fig. 33), and outside these key moments the denudation/exhumation history is only very poorly defined. For this reason, long-term denudation rates derived from such data by averaging over long-term cooling histories have little meaning.

5. The West Greenland margin; a consistent synthesis of geological data, stratigraphic landscape analysis and low-temperature thermochronology

5.1 An integrated approach

The West Greenland continental margin shares characteristics with elevated, passive continental margins around the world, with asymmetrical highs reaching 1–2 km a.s.l., steeper on the oceanward side with incised valleys, sloping more gently inland, and topped with low-relief planation surfaces (Figs 34, 35). A key aspect of the region that allows a wide variety of methods to be employed in reconstructing the development of the margin (in terms of the history of subsidence and uplift) is the presence of Cretaceous to Eocene sedimentary and volcanic sequences exposed in the onshore Nuussuaq Basin (Fig. 36). The evidence from these sequences themselves, and their relationship to basement rocks and landscapes developed in basement rocks, provides major insight into events during and after rifting that is not available on many other margins where cover rocks are absent. In this chapter we illustrate how an integrated approach, in which landscape analysis and palaeothermal methods are combined with geological evidence, has led to a consistent regional synthesis.

5.2 Geological background: the Nuussuaq Basin and the offshore record

5.2.1 Onshore exposures of the Nuussuaq Basin

Sediments and volcanic rocks of the Nuussuaq Basin are exposed in mountains capped by plateaux reaching up to 2 km a.s.l. on the peninsulas of Nuussuaq and Svartenhuk and the island of Disko (Figs 34, 36). Chalmers *et al.* (1999) described the structural development of the basin (both onshore and offshore) and provided a summary of its stratigraphic history. Figure 37 shows a summary of the stratigraphy of the basin presented by Dam *et al.* (2009) and Fig. 38 shows a summary of the subsidence and uplift history of the basin, based on the exposed stratigraphy.

Figure 39 shows examples of geological evidence for subsidence and uplift in the Nuussuaq Basin. Upper Cretaceous fluvial and deltaic sediments dip east (tectonic stratigraphic sequence 2, TSS2; Dam *et al.* 2009), due to rotation of originally flat-lying sediments within fault blocks (Figs 39A, B, C). These deltaic sediments are unconformably overlain by latest Maastrichtian and Danian sediments (TSS4–6, Fig. 39A) and are in turn overlain by Paleocene picritic and basaltic hyaloclastites and lavas (TSS7, Figs 39C, D, TSS6 not visible; Clarke & Pedersen 1976; Chalmers *et al.* 1999; Dam *et al.* 2009; Henriksen *et al.* 2009). The youngest rocks surviving within the onshore part of the basin are Eocene lavas (Storey *et al.* 1998; Schmidt *et al.* 2005) below an extensive planation surface that can be traced bevelling rocks from within the Nuussuaq Basin into adjacent basement terrain over distances of several hundreds of kilometres (Figs 40, 41; Bonow *et al.* 2006a, b).

Rifting had started in the basin by the Albian (the Kome Formation, TSS1; Dam *et al.* 2009), although reflection seismic evidence (Chalmers *et al.* 1999) suggests that there may be older sediments than this deep in the basin. The rifting was followed by a period of thermal subsidence and the deltaic Atane Formation (TSS2) was deposited into the accommodation space formed (Figs 37, 39A, B, C). A period of tectonism and uplift occurred in the Campanian (Dam *et al.* 2009), but no equivalent event is known from the Labrador Sea and the origin of this episode is not understood. The eastern margin of the basin is defined by a major fault on Nuussuaq. Basement rocks are exposed on one side of the fault more than a kilometre above the top of Cenomanian sediments, which underlie covering basalts on the other side of the fault. Pulvertaft (1989) has shown that the Cenomanian sediments were laid down in a low-energy fluviatile environment that shows no evidence of significant nearby topography, meaning that the fault-scarp did not exist at that time. On the other hand, farther north, mid-Paleocene lavas pass across the fault without deflection by it, showing that all the kilometre-scale movement of the fault must have taken place between the Cenomanian and the mid-Paleocene.

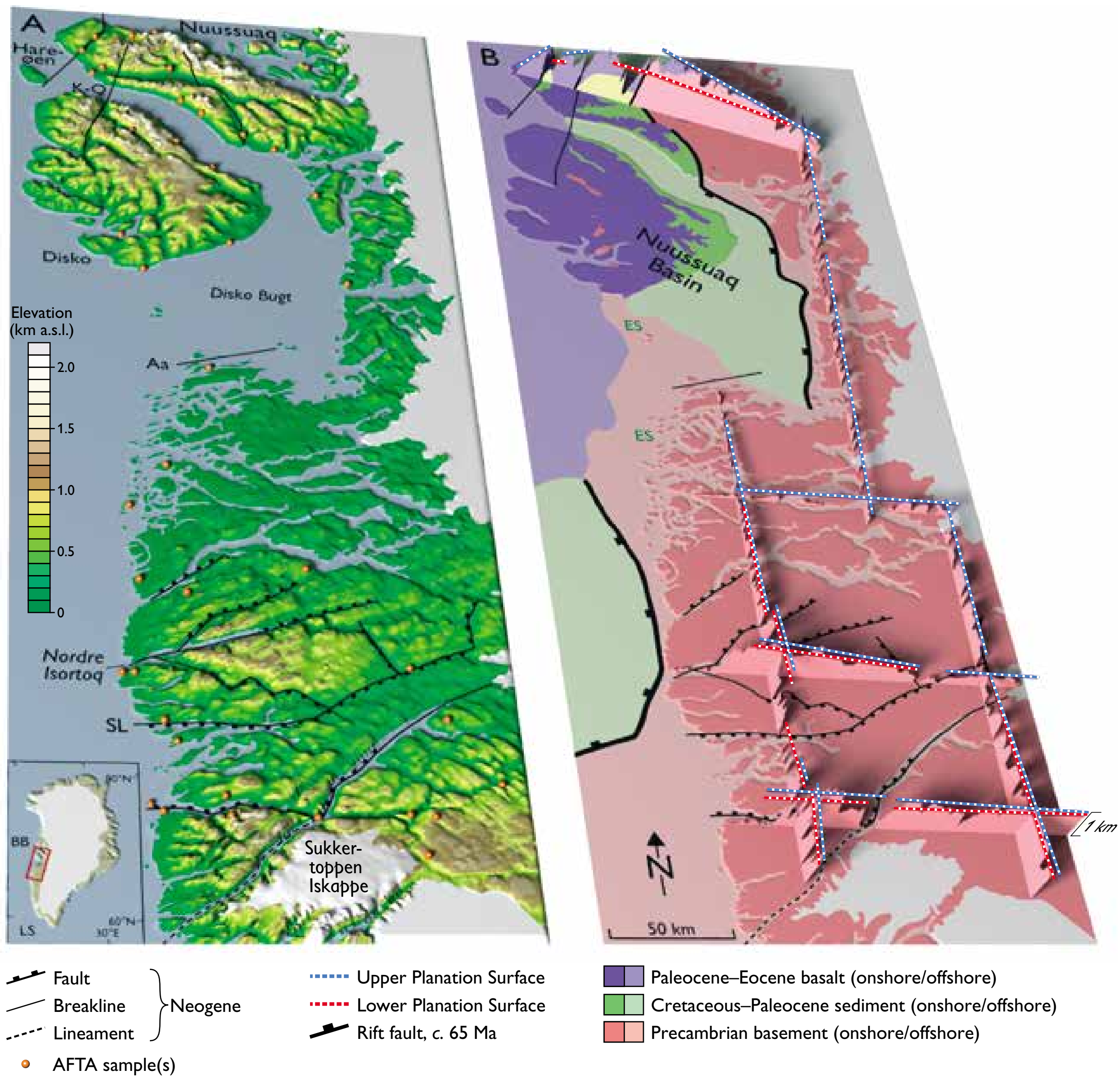
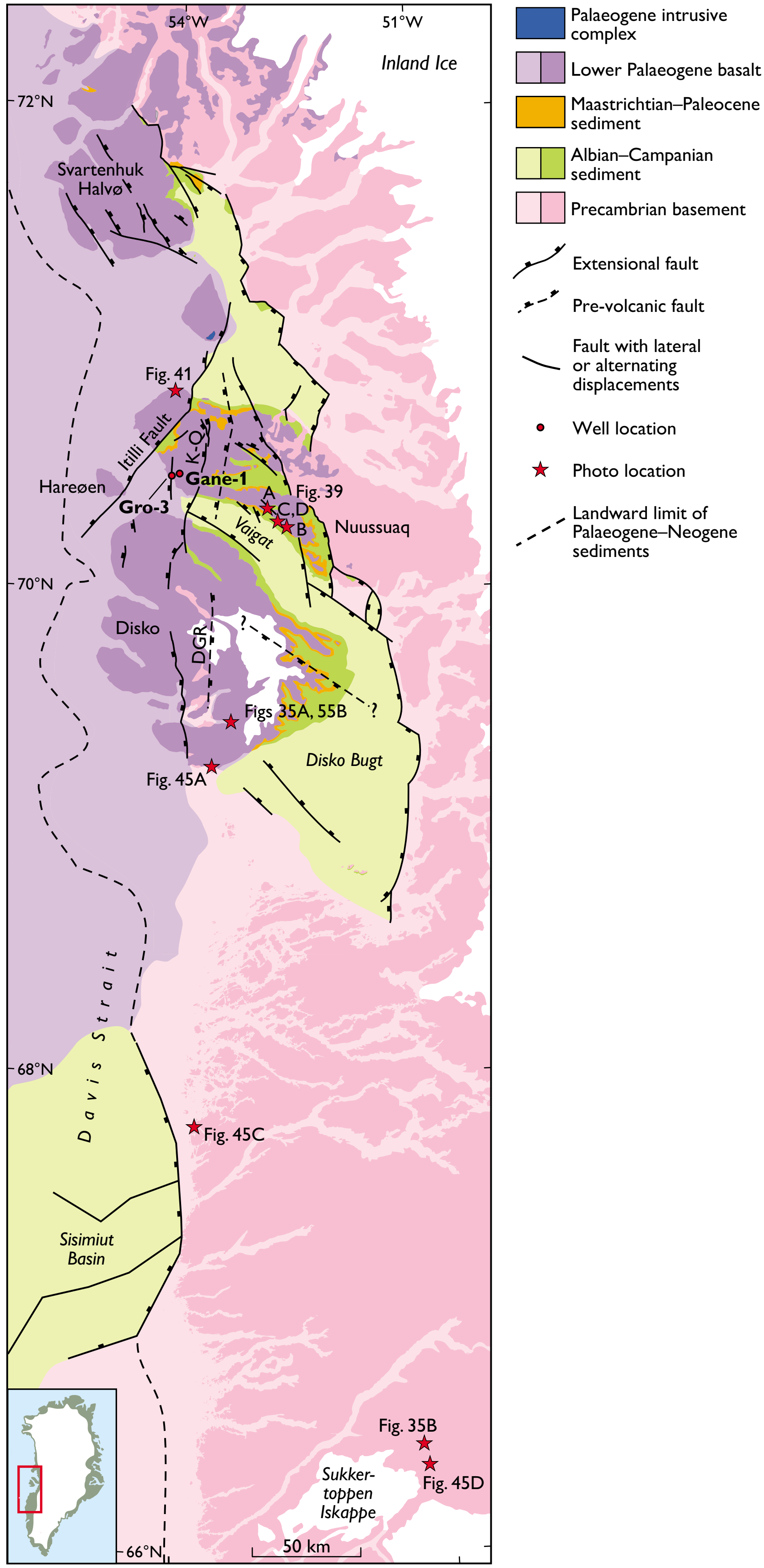


Fig. 34. Topography (**A**) and geology (**B**) of central West Greenland. Also shown on **B** are topographical profiles with interpretation of two planation surfaces (Upper Planation Surface, UPS (blue lines) and Lower Planation Surface, LPS (red lines)). These surfaces are generally inclined to the east but also to the west on western Nuussuaq and to the north, south of Disko Bugt (Bonow *et al.* 2006a, b). They are separated in the highest areas by up to 1 km in the west but merge in the east to one surface. The surfaces cut across both Precambrian basement and mid-Eocene volcanic rocks and are therefore of post-mid-Eocene age. Mapping of the tilted and broken planation surfaces led to the identification of faults formed in connection with the uplift and tilts of the planation surfaces. Three significant faults and breaklines (changes in slope gradient) relative to the planation surfaces are (1) the N–S Kuugannguaq–Qunnilik (**K–Q**) Fault on Disko and Nuussuaq, (2) an E–W fault just north of Aasiaat (**Aa**) where orthogneisses are separated from supracrustal rocks to the north; this fault separates the south-dipping planation surface on Disko from the north-dipping surface south of Disko Bugt, and (3) the E–W Sisimiut Line (**SL**) that coincides with the Precambrian Ikertôq thrust zone. **ES** depicts a hilly etch-surface, re-exposed mainly from below Cretaceous cover rocks but also from Paleocene volcanic rocks on southern Disko and northern Nuussuaq (see Bonow 2005). **BB**, **LS** (on insert map): Baffin Bay and Labrador Sea. From Japsen *et al.* 2006.

Fig. 35. The elevated plateau landscape of the Upper Planation Surface in West Greenland at about 900 m a.s.l. **A**: southern Disko across Paleocene basalts and **B**: northeast of Sukkertoppen Iskappe across Precambrian basement. Glacial reshaping has mainly affected the incised valleys, which have been widened and deepened. Locations shown in Fig. 36. Photos: N. Nielsen, University of Copenhagen (**A**) and K. Secher (**B**). From Japsen *et al*. 2006.

Fig. 36. Geological map of central West Greenland. The Cretaceous–Palaeogene Nuussuaq Basin includes onshore and offshore areas from Svartenhuk Halvø to Disko Bugt. **DGR**: Disko Gneiss Ridge; **K–Q**: Kuugannguaq–Qunnilik Fault. Modified from Chalmers *et al.* (1999), Chalmers & Pulvertaft (2001), Bonow *et al.* (2007b) and Dam *et al.* (2009).



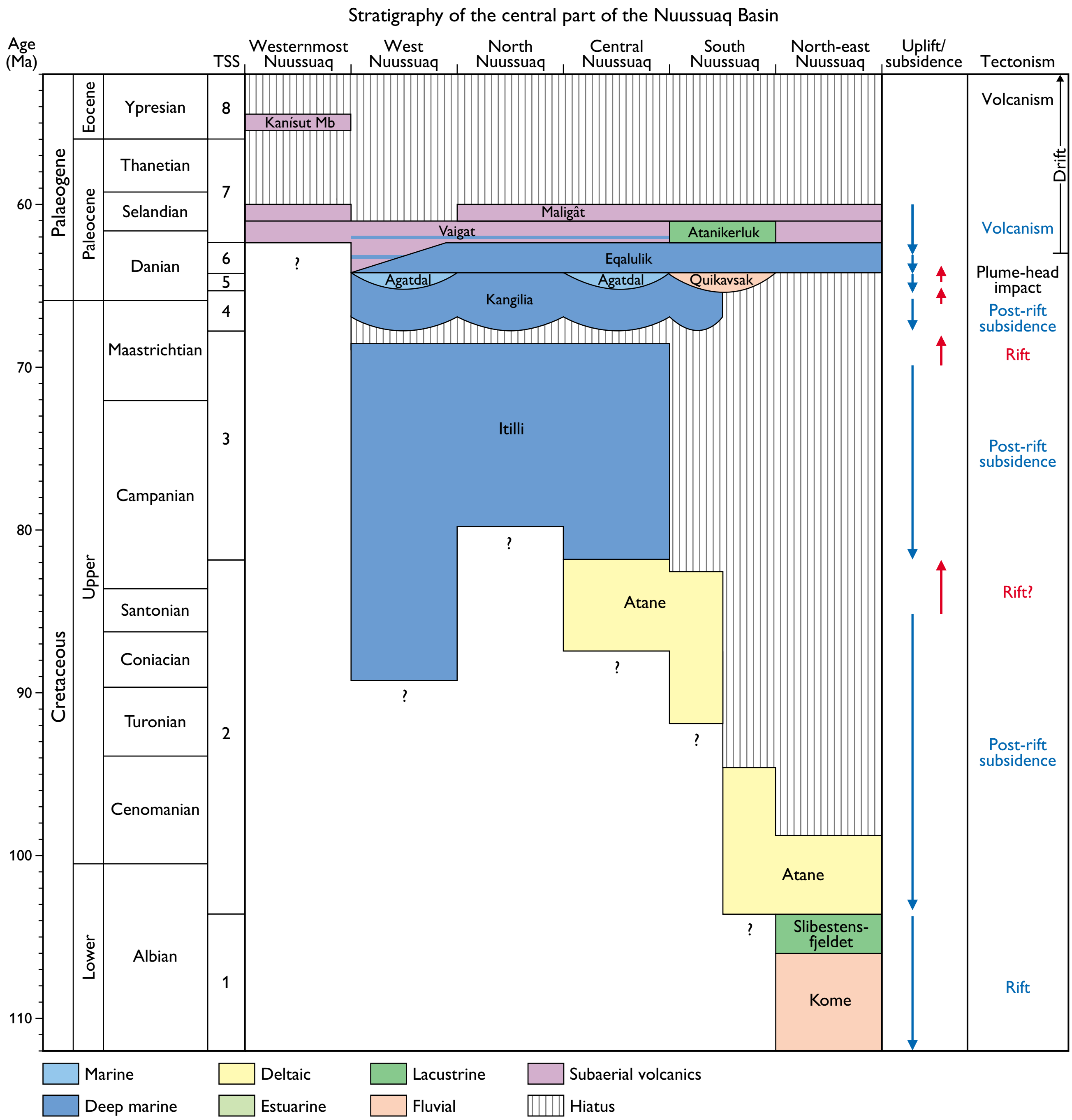


Fig. 37. Stratigraphy and tectonic events along a W–E line through the central part of the Nuussuaq Basin, West Greenland, corresponding roughly to the dip direction produced by the late Cretaceous faulting. Note that the impact of the plume head in the Danian resulted in both uplift and subsidence. Tectonic stratigraphic sequences (**TSS**) and formation names (central part of scheme) defined by Dam *et al.* (2009). Arrows indicate episodes of uplift (red) and subsidence (blue). Modified from Storey *et al.* 1998, Dam *et al.* (2009), L.M. Larsen (personal communication 2013) and G. K. Pedersen & H. Nøhr-Hansen (personal communication 2013).

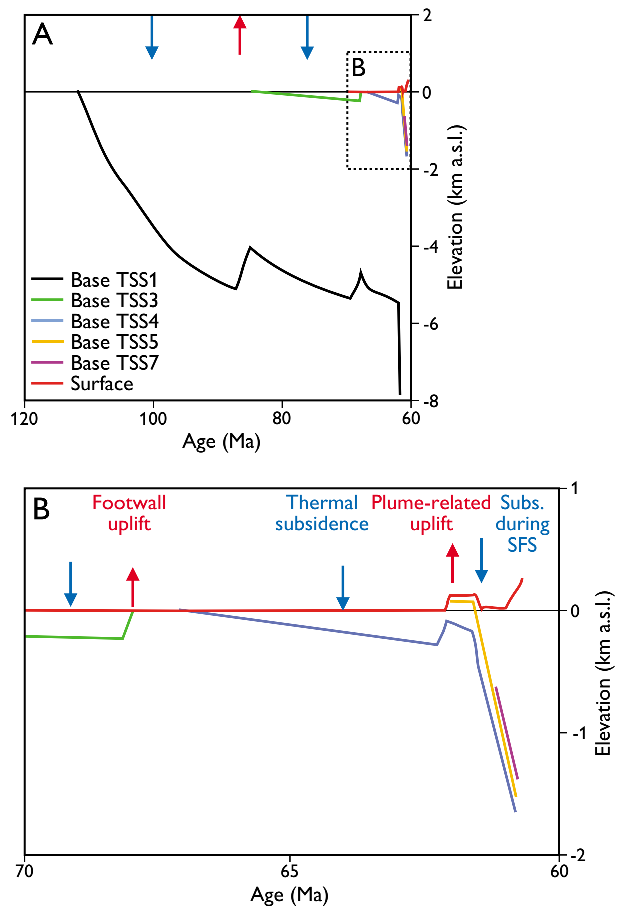


Fig. 38. Summary of the subsidence and uplift history of the Nuussuaq Basin. **A**: Schematic diagram showing estimates of uplift and subsidence derived from the geological record that affected the Nuussuaq Basin between Early Cretaceous rifting and shortly after mid-Paleocene break-up (Fig. 37). **B**: Enlargement of part of **A**. The record is from Dam *et al.* (2009) except where otherwise cited and refers to an area on the south coast of Nuussuaq marked on Fig. 36 (photos in Fig. 39). An episode of rifting took place probably in the Aptian (depositing TSS1) followed by thermal subsidence that started in the Albian (Chalmers *et al.* 1999). The Atane Formation (TSS2) was deposited during the latter event. It is at least 3 km thick but a reflection seismic line on the south coast of Nuussuaq (Chalmers *et al.* 1999, fig. 9) shows that the sediments at this location are at least 6 km thick and may be 8 km thick. An episode of uplift followed by renewed subsidence and deposition of the Itilli Formation (TSS3) took place in the Campanian. Faulting, uplift and erosion of major submarine channels in the Maastrichtian led to the complete erosion of TSS3 at some localities such as the one represented here. The Kangilia Formation (TSS4) was deposited in the renewed subsidence after this event (Fig. 39A). Two episodes in the late Danian uplifted the surface to above sea level and the fluvial Quikavsak Formation (TSS5) was deposited in the river channels (Fig. 39B). Submarine volcanism commenced in the western Nuussuaq Basin at this time. The volcanism built up a volcanic island and became subaerial at the same time as the eastern Nuussuaq Basin subsided by 600–700 m. Subaerial lava flows from the west flowed into this basin (TSS7), filling it as a hyaloclastite delta (Fig. 39C). Subsidence continued, however, as shown by the deposition of a succession of lava flows with hyaloclastite bases and subaerial tops (Fig. 39D) and eventually entirely sub-aerial lavas. Deposition of a later (Eocene) subaerial volcanic succession (TSS8) followed a hiatus in the volcanism. Arrows indicate episodes of uplift (red) and subsidence (blue). **Subs. during SFS**: Subsidence during sea-floor spreading. **TSS**: tectonic stratigraphic sequences in Fig. 37 (Dam *et al.* 2009).

Two phases of channel incision have been observed beneath the flood basalts in the Nuussuaq Basin (Dam & Sønderholm 1994, 1998; Dam *et al.* 1998; Dam & Nøhr-Hansen 2001; Dam 2002; Dam *et al.* 2009). The earliest channels form part of the latest Maastrichtian unconformity, and Dam *et al.* (1998) suggested that they were eroded as submarine canyons on the crests of the fault blocks that rotated the Cretaceous sediments to dip eastwards. This movement thus defines the timing of latest rifting in the Nuussuaq Basin close to the Cretaceous–Cenozoic boundary, in agreement with the known constraints on the timing of the latest rifting offshore (Aram 1999; Chalmers & Pulvertaft 2001; Christiansen *et al.* 2001; Dalhoff *et al.* 2003; Chalmers 2012).

Post-rift Danian subsidence and deposition of the Kangilia Formation (TSS4) was followed by uplift during the mid-Paleocene, probably in response to the impact of the head of the Iceland plume (Dam *et al.* 1998), when new channels were eroded as subaerial valleys, which are still filled with fluviatile Paleocene sediments (Quikavsak Formation, TSS5, 6, Fig. 39B). Volcanism commenced shortly afterwards, during Chron 27n (Riisager & Abrahamsen 1999), corresponding to the interval 61.65 to 61.98 Ma (Gradstein *et al.* 2004).

Volcanism started on a submarine slope to the west of the area where the major channels were eroded. The volcanic pile (TSS7) was built up above sea level so that lavas began to be erupted subaerially. However, the lavas that entered the sea farther east formed eastward-prograding Gilbert-type delta structures with cross-bedded hyaloclastite sets up to 600 or 700 m thick (Fig. 39C; Pedersen *et al.* 1993), indicating that the basin had subsided by at least this amount during the few hundred thousand

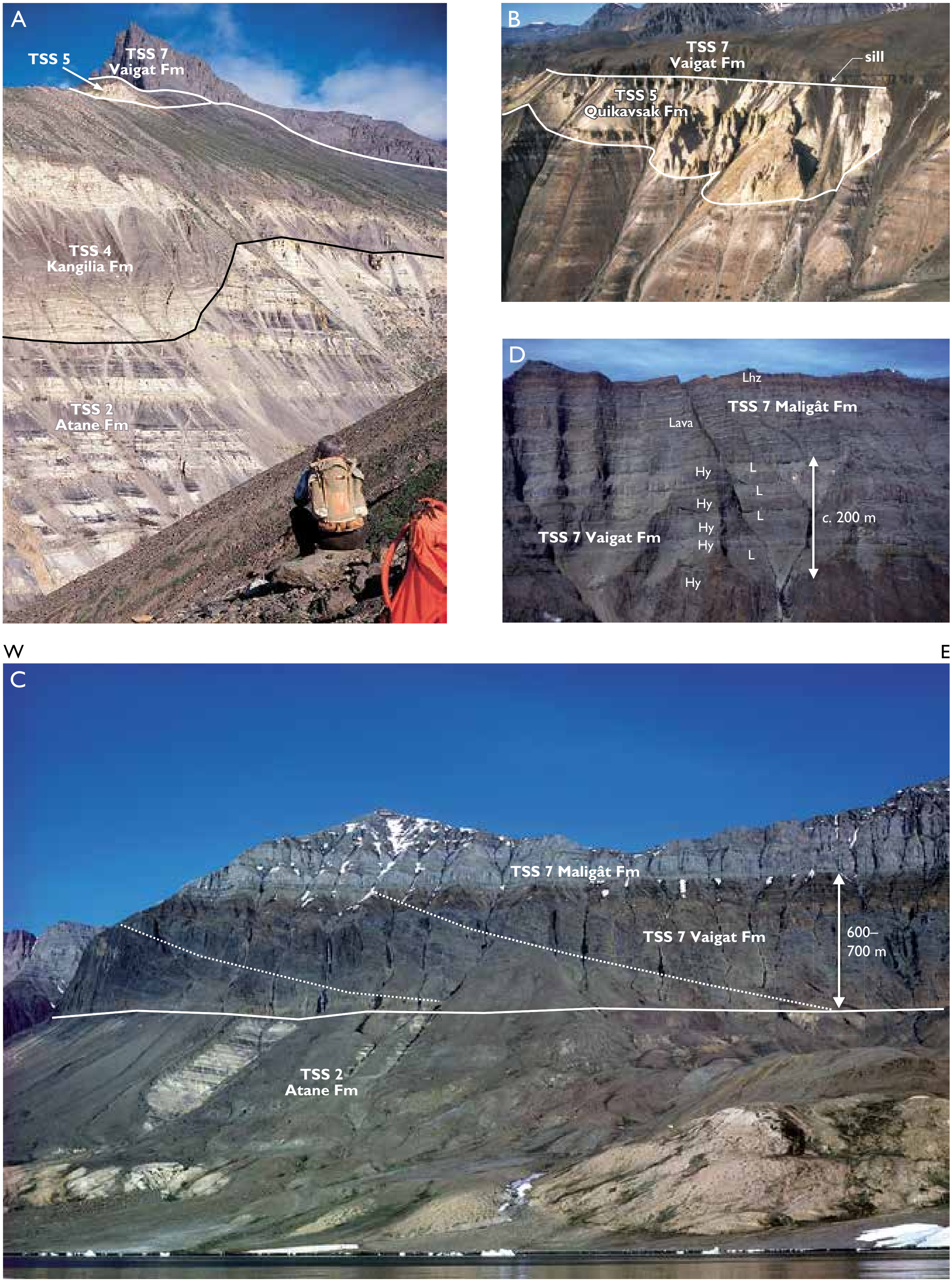

A
TSS 5
TSS 7
Vaigat Fm
TSS 4
Kangilia Fm
TSS 2
Atane Fm
B
TSS 7
Vaigat Fm
sill
TSS 5
Quikavsak Fm
D
Lhz
TSS 7 Maligât Fm
Lava
Hy
L
Hy
L
L
Hy
Hy
L
Hy
c. 200 m
TSS 7 Vaigat Fm
W
E
C
TSS 7 Maligât Fm
TSS 7 Vaigat Fm
600–
700 m
TSS 2
Atane Fm

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Fig. 39. Examples of geological evidence for subsidence and uplift in the Nuussuaq Basin. See Fig. 37 for stratigraphy and Dam *et al*. (2009) for stratigraphic details. Photos from the south coast of Nuussuaq (locations on Fig. 36). **A**: Uplift and incision during Maastrichtian rifting (TSS4). The fluvial and deltaic sediments of TSS2 (Atane Fm; Dam *et al*. 2009; also shown in **B** and **C**) were probably laid down in a basin subsiding thermally after a rift phase in the Aptian–Albian (Chalmers *et al*. 1999; Dam *et al*. 1998, 2009). The unconformity and deep channel were formed by erosion and listric faulting during the Maastrichtian that appears to have removed the whole of TSS3 (Itilli Fm), perhaps 250 m thick at this location, plus some of TSS2 (Atane Fm). This tectonic episode affected much of southern West Greenland (Figs 42, 43). The channel and unconformity were buried below transgressive marine sediments of TSS4 (Kangilia Fm), indicating renewed subsidence. There is a slight glimpse of the effects of the next phases of uplift (TSS6 and TSS7, volcanic rocks), but they are very foreshortened from this viewpoint. To see this mountainside from a much higher viewpoint, see Dam *et al*. (2009, fig. 14). Photo: T.C.R. Pulvertaft. **B**: Uplift and incision due to mid-Paleocene plume impact (TSS5, TSS6). Renewed uplift in two phases (TSS5 and TSS6) during the mid-Paleocene, uplifted and eroded TSS4 (Kangilia Fm), forming new unconformities that separate the fluvial sediments of TSS5 (Quikavsak Fm) from TSS2. The Quikavsak rivers flowed westwards into the sea where western Nuussuaq is today. This phase of erosion is widespread over the whole of offshore southern West Greenland (Dalhoff *et al*. 2003; Japsen *et al*. 2005) and Dam *et al*. (1998) suggested that it is due to uplift from the rise and impact of the Iceland plume head onto the base of the lithosphere. **C**: Volcanism following early post-plume subsidence (TSS7). Submarine volcanism built up a volcanic island west of the present-day Nuussuaq Peninsula. By that time, 600–700 m of subsidence had taken place farther east, sufficient to form a marine basin into which the peridotitic lavas flowed as and eastward prograding, hyaloclastite Gilbert delta in the Vaigat Fm of TSS7 (darker volcanic rocks). Dashed lines show examples of hyaloclastite foresets. The succeeding lighter volcanic rocks are the Maligât Fm. Photo: L.M. Larsen. **D**: Volcanism during continued subsidence. Subsidence continued in the late Paleocene after the marine basin (shown in **C**) was full, shown by the 200 m thick succession of lava flows with hyaloclastite bases (**Hy**) and subaerial tops (**L**), indicating that subsidence and magma production were keeping pace. Eventually, either subsidence slowed or lava production increased until the volcanic succession became sub-aerial (**Lhz**). Photo and interpretation: A.K. Pedersen, University of Copenhagen. **TSS**: tectonic stratigraphic sequences in Fig. 37 (Dam *et al*. 2009).

years after deposition of the fluviatile sediments of the Quikavsak Formation.

The basin continued to subside during subsequent deposition of a 200 m thick succession of subaerial lavas alternating with five horizons of foreset-bedded hyaloclastites on Nuussuaq (Fig. 39D; Pedersen *et al*. 2002), a succession recording an almost perfect balance between aggradation of the lava pile and subsidence of the basin. This succession is overlain by a 160–180 m thick zone of subaerial lavas across which the transition from Chron 27n to 26r is recorded (Riisager & Abrahamsen 1999; Pedersen *et al*. 2002), and above which Piasecki *et al*. (1992) found marine dinoflagellates at a present-day height of 1176 m a.s.l., indicating subsidence was followed by substantial uplift after continental break-up in the Labrador Sea during Chron 27 (mid-Paleocene; Chalmers & Laursen 1995).

Thus the uppermost Maastrichtian and Paleocene sedimentary and volcanic rocks in the Nuussuaq Basin record uplift events that Dam *et al*. (1998) estimated had removed up to 1.3 km of Cretaceous section (Fig. 39B), followed by subsidence of at least 1000 m and probably considerably more than that (Figs 39C, D). At least 900 m of this subsidence took place during Chron 27n (Pedersen *et al*. 2002), which lasted 330 ka, so the minimum subsidence rate was 3 km/Myr. The presence of submarine rocks today at heights of more than a kilometre above sea level shows that the events that lifted them to those heights must have taken place after the Paleocene (Fig. 38). The stratigraphic record onshore does not record those events, but the stratigraphic record offshore does give some clues.

5.2.2 The geological record off southern West Greenland

Dalhoff *et al*. (2003) reported the presence of a major unconformity that separates mid-Paleocene and younger sediments from Campanian and older sediments over the whole of the sedimentary basins offshore southern West Greenland (Figs 42–44). They attributed this unconformity to uplift and erosion caused by impact of the head of the Iceland plume, because renewed sedimentation onto the unconformity (and presumably therefore renewed subsidence) started contemporaneously with the onset of volcanism in the Nuussuaq Basin. The formation of the base Cenozoic unconformity offshore was followed by subsidence that created sufficient accommodation space in the Sisimiut Basin (Fig. 36) to contain a 2.5 km thickness of sediment deposited during 21 million years of late Paleocene to mid-Eocene time (Fig. 42B; Dalhoff *et al*. 2003). Seismic sequence analysis calibrated by borehole control shows that sedimentation in the northern part of the basin was constantly sufficient to

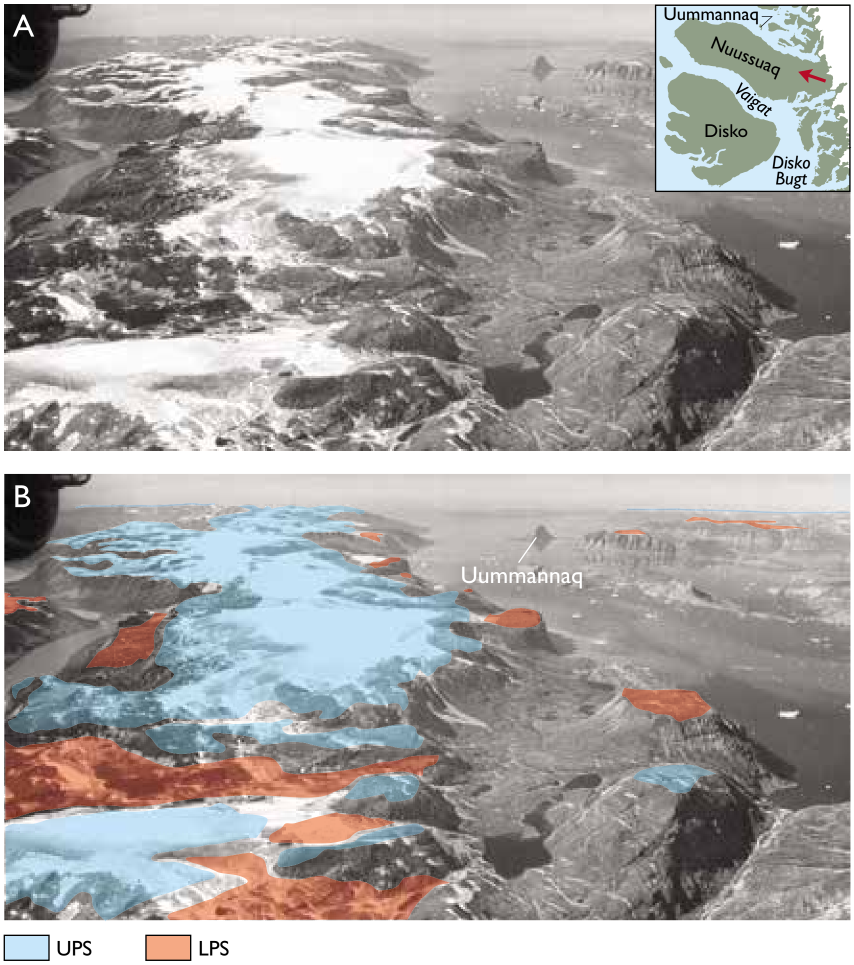


Fig. 40. The Upper Planation Surface (UPS) and the Lower Planation Surface (LPS) developed across Precambrian basement (foreground) and Paleocene basalt (background) on Nuussuaq (view from east towards west). © Danish Geodata Agency. Modified from Bonow *et al*. (2006a) and Green *et al*. (2011).

fill the available accommodation space, indicated by the continuous deposition of delta-top facies. This indicates an average subsidence rate of 120 m/Myr.

Thus the sedimentary and volcanic record preserved both onshore and offshore West Greenland shows evidence for substantial mid-Paleocene uplift that was probably associated with the impact of the Icelandic plume, but the record also indicates that this uplift was short-lived and was followed by rapid and substantial subsidence. This uplift event could not, therefore, have been what formed the present-day mountains in West Greenland. That event must have taken place later than the mid-Palaeocene break-up. How and when this uplift and tilting took place has been analysed by integrating landscape analysis and AFTA, as described in the following sections.

5.3 Results of SLA in West Greenland

By identifying and mapping re-exposed peneplains (sub-Cretaceous and sub-Paleocene etch surfaces; i.e. hilly peneplains), high-level flat peneplains (planation surfaces), and a deep valley generation in West Greenland and combining these observations with the stratigraphic record, Bonow *et al*. (2006a, b) defined a relative denudation chronology for the development of the West Greenland margin.

A sub-Cretaceous etch surface (ES) formed by kaolinitic clay weathering of gneiss is re-exposed from below Cretaceous sedimentary strata and a sub-Paleocene ES from below volcanic rocks on Disko and Nuussuaq (Pulvertaft & Larsen 2002; Bonow 2005). The weathering products below the Paleocene cover on southern Disko, the saprolites, are less mature than those below the Cretaceous cover, but the hills are of about the same height as those on the sub-Cretaceous ES. It seems there-

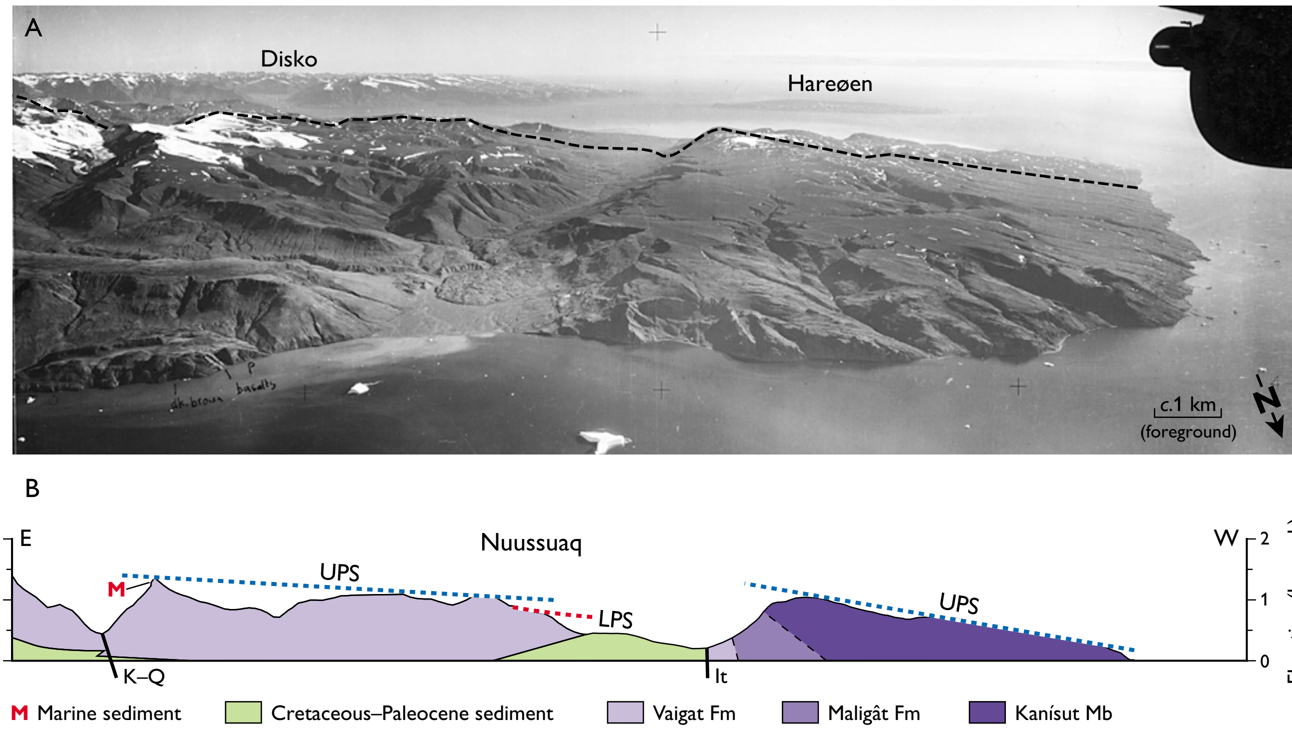


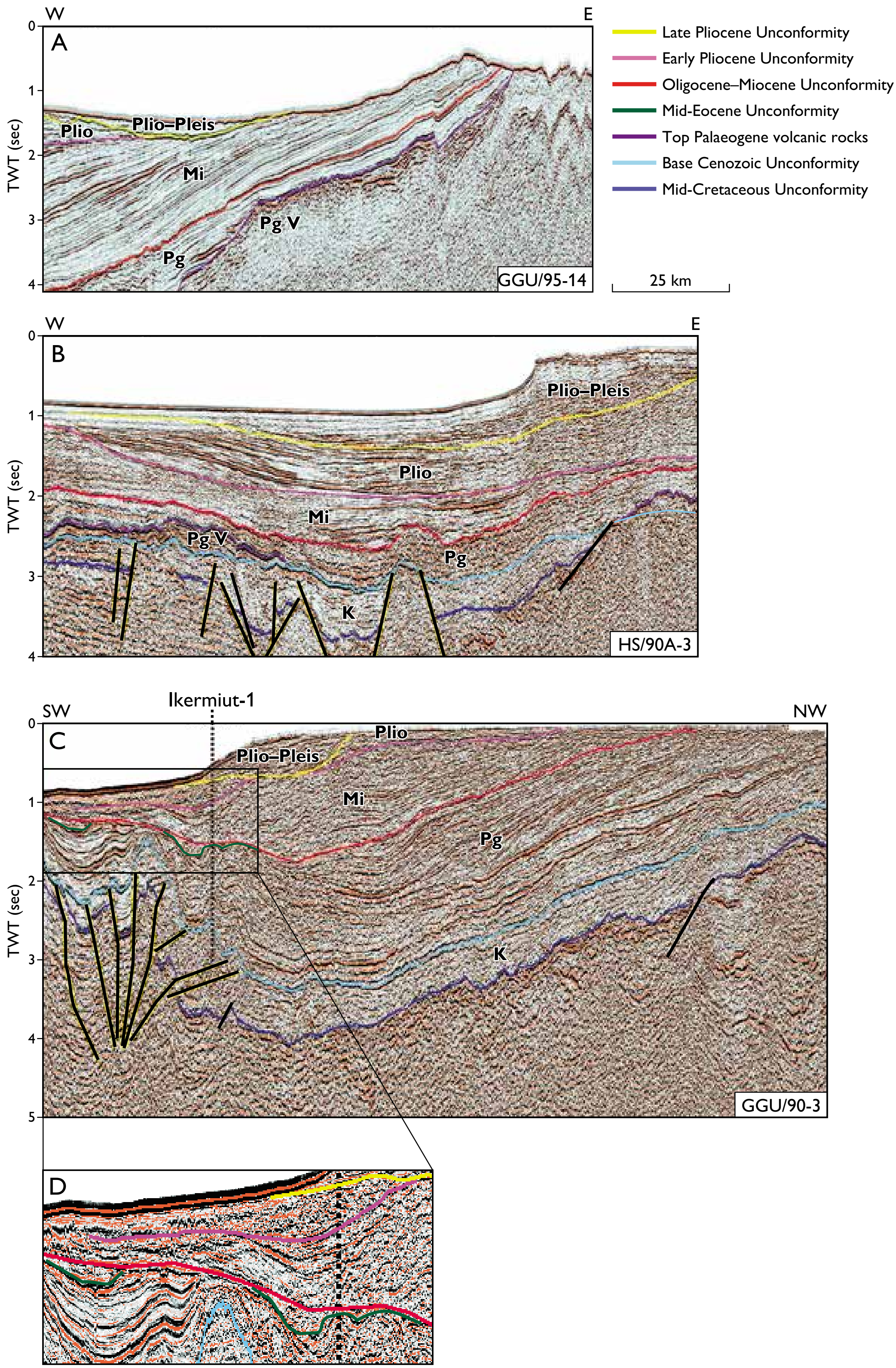
Fig. 41. **A**: Well-preserved Upper Planation Surface (**UPS**) across steeply dipping, Palaeogene volcanic rocks on westernmost Nuussuaq (view from north towards south). **B**: Palaeosurfaces and geology along a profile (dashed line in **A**). Note the very recent fluvial incision. The low position has preserved the planation surface from being destroyed by glaciers and cirques. Location shown in Fig. 36. Volcanic stratigraphy in Fig. 37. **It**: Itilli Fault. **K–Q**: Kuugannguaq–Qunnilik Fault. **LPS**: Lower Planation Surface. **M**: marine sediment (Piasecki *et al.* 1992). © Danish Geodata Agency. Modified from Bonow *et al.* (2006b).

fore likely that the old kaolinitic saprolite was stripped, and that the present relief belongs to a common late Mesozoic ES (Fig. 45A, B). The hills are limited by fracture systems trending NW–SE and by ENE–WSW schistocity. These lineaments are all affected by the deep weathering. Another N–S fracture system probably formed later than the ES as those fractures are not affected by the deep weathering (Bonow 2005).

Bonow (2005) used reflection seismic lines to identify the hilly surface at the sea bed south of Disko and found that the surface continues over the exposed basement areas south of Disko Bugt (Fig. 45C). The ES is tilted, rising from the coast with higher elevations to the south and east. The inclined ES is cut off by a younger peneplain (Fig. 5), which rises slowly towards the south, where it gradually splits into two distinct peneplains, which we refer to as the Upper and Lower Planation Surface (UPS and LPS, respectively; Fig. 34).

Bonow *et al.* (2006a, b) traced the UPS across volcanic rocks as young as mid-Eocene (39 Ma; Schmidt *et al.* 2005) on Nuussuaq and could thus conclude that the UPS is of post-Eocene age. The LPS is developed as a system of wide and shallow valleys cut below the UPS; e.g. in the area north-east of Sukkertoppen Iskappe (Figs 45D–F), but merges with the UPS; e.g. south of Disko Bugt and on western Nuussuaq. Both surfaces are of regional extent (>500 km; Figs 34, 40). The UPS can be traced across the entire area. Since both surfaces are continuous across rocks of different resistance to erosion, and since no single geological surface exists to which these surfaces could have been graded, the most likely explanation is that they were each graded to a general base level at the time of their formation. As the coast is known to have been near the present onshore areas throughout the Cenozoic (Nielsen *et al.* 2001; Dalhoff *et al.* 2003), the most likely base level is sea level. A lower valley generation, glacially reshaped to a considerable extent, is deeply incised below the LPS (Figs 45E, F). As both the UPS and LPS have lost their original base of formation, they must have been uplifted and are now palaeosurfaces.

The UPS forms the summit surfaces, from just above sea level on western Nuussuaq to over 2000 m, of crustal blocks that have been uplifted and tilted by different amounts (Figs 34, 41), indicating a tectonic element in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1
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4
Late Pliocene Unconformity
Early Pliocene Unconformity
Oligocene–Miocene Unconformity
Mid-Eocene Unconformity
Top Palaeogene volcanic rocks
Base Cenozoic Unconformity
Mid-Cretaceous Unconformity
25 km
B
Plio–Pleis
Plio
Mi
Pg V
Pg
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Plio
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Fig. 42. Reflection seismic lines from offshore southern and central West Greenland showing examples of the major unconformities in the basin. See Fig. 43 for stratigraphy and Fig. 44 for locations of these lines. Rifting took place prior to deposition of the Kangeq Formation (**K**) and pre-Cretaceous sediments may also be present below this unconformity. See Chalmers & Pulvertaft (2001) and Chalmers (2012) for details. The whole of the Maastrichtian and most of the Danian successions are missing between the Kangeq and Palaeogene (**Pg**) successions (Dalhoff *et al*. 2003; Japsen *et al*. 2010). The Pg succession consists of upper Paleocene and Eocene deltaic (on the seismic line in **C**) or pro-delta marine (on the seismic line in **B**) sediments (Dalhoff *et al*. 2003). They are separated from overlying upper Miocene (**Mi**) sediments by a hiatus from which the whole of the Oligocene and lower Miocene successions are missing. The Miocene and overlying Pliocene (**Plio**) sediments are mostly contourites and associated sediments. Both are truncated by a late Pliocene unconformity (or at the sea bed), after which thick wedges of coarse sediment prograded across the shelf edge. **A**: Seismic line GGU/95-14. The line runs west from near the western end of Nuussuaq. The top of the Palaeogene lavas that reach an elevation over 2 km a.s.l. onshore dip westwards to more than 3 km below sea level. Note that all the pre-Pleistocene Cenozoic sediments also dip west and are eroded at the sea bed just west of the coast, indicating that the Eocene (**Pg**) and Miocene (**Mi**) sedimentary strata must have extended farther east and must have covered at least parts of present-day Nuussuaq. Redrawn from Chalmers (2000, fig. 4). **B**: Seismic line HS/90A-3. Note that the unconformity between the upper Miocene (**Mi**) and Eocene (**Pg**) sediments shows little or no angular truncation, indicating that uplift (and any erosion) during the Oligocene and early Miocene must have been very uniform over a large area (Fig. 43). The platform area capped by Palaeogene lavas at the eastern end of this line is the Nukik Platform. **C**: Seismic line GGU/90-3 through the Ikermiut-1 borehole. The sediments at the western end of this line have been folded in compressive movements along the Ungava Fault system (Oakey & Chalmers 2012), and those in the eastern part of this line have been uplifted and eroded after deposition of the Miocene succession. **D**: Detail of the area west of the Ikermiut-1 borehole showing the Oligo–Miocene unconformity (red) and an earlier mid-Eocene unconformity (green) are quite distinct at this location, being separated by upper Eocene sediments. This latter unconformity truncates sediments folded during compressive movements along the Ungava Fault system in the Eocene (Oakey & Chalmers 2012). The mid-Eocene unconformity was dated at 45 Ma (Dalhoff *et al*. 2003) and upper Eocene sediments are present between it and the Oligo–Miocene Unconformity. The existence of these two separate unconformities was ignored by McGregor *et al*. (2012), who came to the erroneous conclusion that the Oligo–Miocene Unconformity was related to plate tectonic movements in the Labrador Sea and Baffin Bay. Those movements had, however, ceased 10 million years earlier. **Plio–Pleis**: Pliocene–Pleistocene sediments. **Plio**: Pliocene sediments. **Mi**: upper Miocene sediments. **Pg**: Palaeogene sediments. **Pg V**: Palaeogene lava.

the uplift (Bonow *et al*. 2006a, b). The occurrence of marine Paleocene sediments 1200 m a.s.l. (Piasecki *et al*. 1992) close to the UPS on western Nuussuaq is consistent with the elevation of the UPS being a measure of the amount of rock uplift since the formation of the UPS at this location.

The initial uplift raised the UPS, which was simultaneously broken along faults, thereby forming different tectonic blocks. This phase triggered incision of new fluvial valleys to a base at sea level that ultimately led to the formation of the partly developed LPS. The floors of these valleys are preserved today at a maximum elevation of 1 km a.s.l. and at a lowest level of 400 m a.s.l., revealing a second phase of differential uplift. The two uplift phases resulted in three central high areas within the study area; Nuussuaq and Disko <2000 m, east of Nordre Isortoq <1500 m and Sukkertoppen Iskappe <1800 m a.s.l. (Fig. 34). Removal of cover rocks resulted in re-exposure of the ES at low elevations around Disko Bugt after the second uplift.

Bonow (2004) used geology and maps of elevation contours and slopes (Fig. 46) to reconstruct a palaeodrainage system across Nuussuaq and Disko with a primarily southeastern direction, connected to the lower plateau remnants of the LPS. The present drainage of these valleys on Nuussuaq is generally to the east, but they turn westwards into the major Affarsuaq Valley, forming agnor (fish-hooked) valleys, indicative of an eastward movement of the water divide in connection with the glaciations. The maps of Disko also indicate an originally south-easterly direction of the major valley systems. A major valley with a south-east-heading drainage has probably also occupied Vaigat, the sound which today separates Nuussuaq and Disko. The south-east-heading palaeodrainage starts along the Kuugannguaq–Qunnilik Fault (K–Q in Fig. 46A), while the palaeodrainage on the western side of this fault was towards the west. This interpretation is supported by the erosional pattern of the Paleocene basalts, as areas with Cretaceous and Precambrian rocks exhumed from below the Paleocene cover, narrow north-westwards along the suggested palaeodrainage. The headward erosion of the valleys has reached and slightly passed the Kuugannguaq–Qunnilik Fault. The palaeodrainage helps in defining tilt direction in certain areas, e.g. in the areas between the Kuugannguaq–Qunnilik Fault and the main Cretaceous boundary fault system on Nuussuaq that separates the Cretaceous basin from

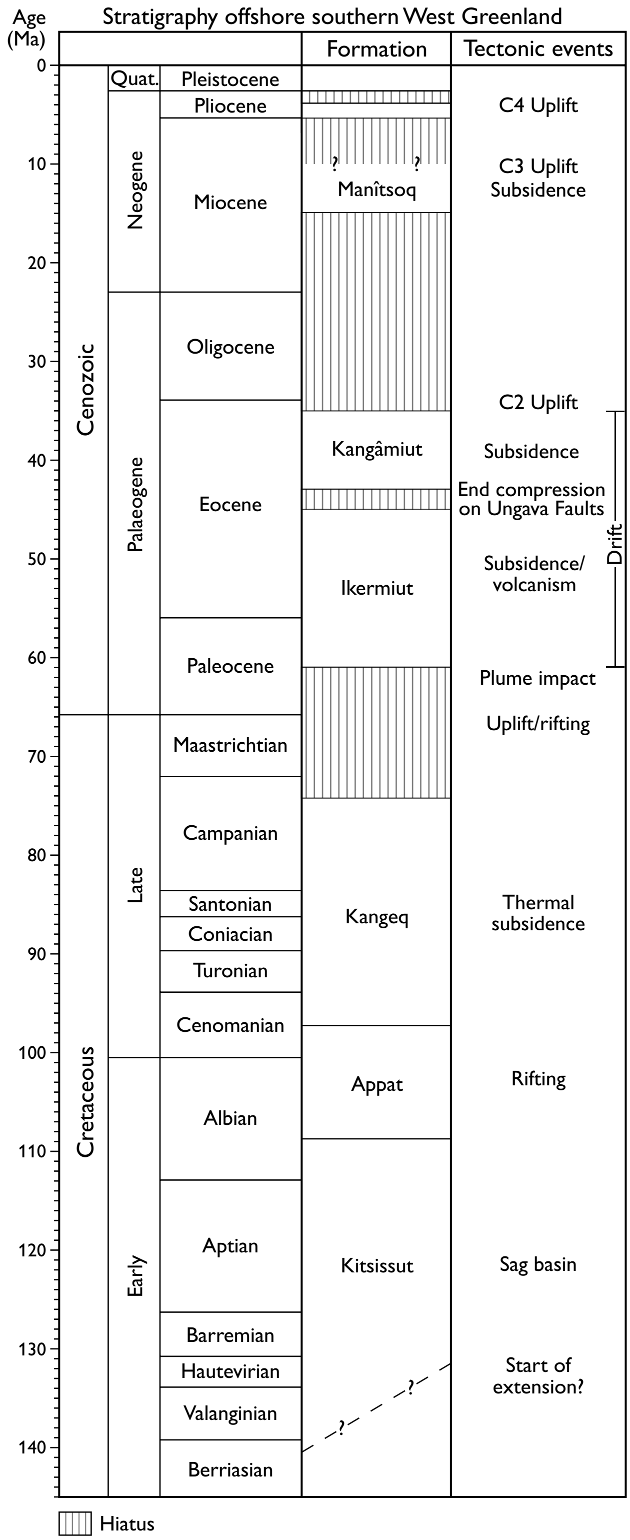


Fig. 43. Stratigraphy offshore southern West Greenland based on Piasecki (2003) and Sørensen (2006). Note the base Cenozoic, Oligo–Miocene and late Cenozoic unconformities. Dalhoff *et al.* (2003) attributed the base-Cenozoic unconformity to uplift and erosion caused by impact of the head of the Iceland plume, because renewed sedimentation onto the unconformity started contemporaneously with the onset of volcanism in the Nuussuaq Basin. A thick succession of reflections ('Deep Sequence', Chalmers & Pulvertaft 2001), below lower to mid-Cretaceous sequences, offshore West Greenland, and sea bed samples from the Davis Strait containing Ordovician and Jurassic sediments (Dalhoff *et al.* 2006), indicate that Palaeozoic to Jurassic deposits may be present offshore West Greenland as indicated in Fig. 42 (see Japsen *et al.* 2010). For C2–C4 see Table 2.

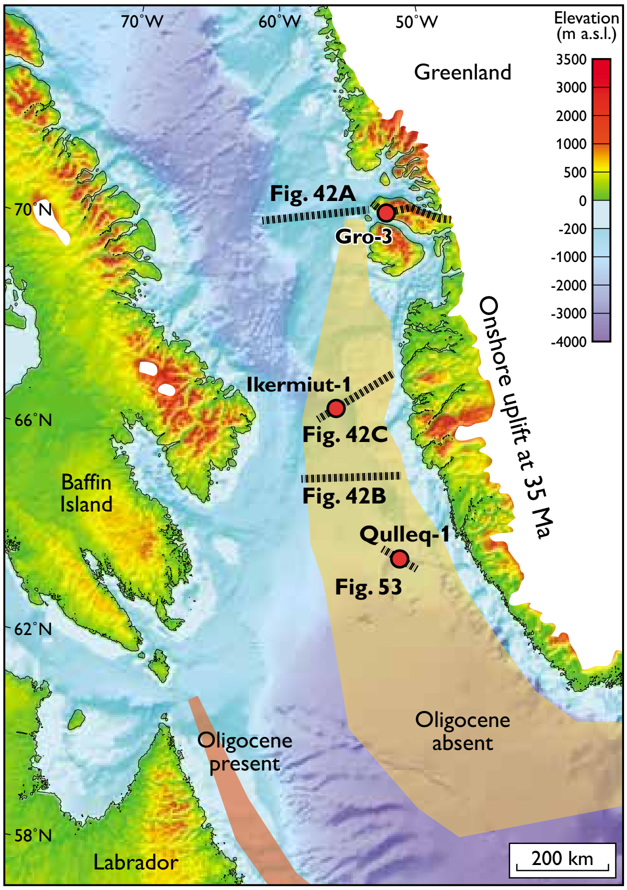


Fig. 44. Outline of the extent of the Oligocene – lower Miocene hiatus offshore southern and central West Greenland based on seismic and well data (Sørensen 2006; see the offshore stratigraphic column in Fig. 43). Oligocene strata have been encountered east of Labrador on the Canadian side of the Labrador Sea. No information is available for the area between the two areas where Oligocene is marked as absent and present, respectively. The Oligocene hiatus thus represents the time interval during which the Oligo–Miocene peneplain, UPS, was graded to sea level onshore after the uplift event that began at the Eocene–Oligocene transition. Bathymetry and bedrock topography in metres. Icecaps indicated as white. Seismic lines and cross-sections shown in Figs 42 and 53 are indicated. Modified from Japsen *et al.* (2010).

the basement block (Fig. 46A). Here the UPS cannot be identified due to glacial obliteration of the surface, but the direction of palaeodrainage suggests a tilt towards the south-east (Fig. 46C).

Glacial erosion has to a large degree reshaped valleys and other areas in low positions along downfaulted blocks. Not only Disko Bugt but also Vaigat hosted major west-heading outlet glaciers that have widened and deepened the originally east-heading river valleys and also ultimately eroded through the high area around the Kuugannguaq–Qunnilik Fault (Fig. 36). The wide Affarsuaq Valley on Nuussuaq was also largely deepened and widened westwards by glacial action. Another area of pronounced glacial erosion is the outer, western parts of the low tectonic block between the two high areas between Nordre Isortoq and Sukkertoppen Iskappe. Several areas here have been eroded below sea level. On the other hand, the plateaux inland are well preserved (Fig. 47). In positions close to the coast and above 1000 m, an alpine relief has, in some places, developed due to erosion by valley glaciers and cirques (Fig. 5). In the tectonic block east of the Kuugannguaq–Qunnilik Fault on Nuussuaq, the UPS is totally destroyed in this way. In contrast, the UPS is well preserved in low positions on westernmost Nuussuaq (Fig. 41).

5.4 Use of palaeothermal methods to define an absolute chronology of uplift and exhumation events

The results of SLA presented in section 5.3 provide clear definition of multiple uplift events of West Greenland. A period of prolonged uplift and erosion took place after Paleocene break-up west of Greenland and led to development of a regionally extensive peneplain that cuts across Palaeogene basalt and older rocks. This peneplain (UPS), now recognised as plateaux up to 2 km a.s.l., is broken and tilted in different directions and now extends from sea level to the highest summits. A less tilted, lower planation surface (LPS) occurs between 400 and about 1000 m a.s.l. (Fig. 40). The deep valleys incised below the LPS are more or less glacially reshaped. Thus, three episodes of uplift and erosion occurred after break-up: the first episode led to formation of the UPS and two subsequent episodes lifted the UPS to its present elevation. Low-temperature thermochronology studies, undertaken to place these episodes into an absolute timescale (Fig. 48), were described by Japsen *et al.* (2005, 2006, 2009) with detailed results presented as online supplementary data files. We present a summary of results in the following.

AFTA data were collected in a large number of outcrop samples from the Nuussuaq Basin and adjacent basement regions (65–72°N), resulting in definition of a series of Phanerozoic cooling episodes (Table 2). One pre-Cenozoic cooling episode is of importance for this study, namely the Late Jurassic event which began between 160 and 150 Ma (cooling episode C1). This event is indicated in AFTA data from most samples from the basement areas, and preceded the onset of known rifting offshore and in the Nuussuaq Basin (Chalmers & Pulvertaft 2001). The ES is the probable end-result of land-form evolution following this cooling episode, which thus must have involved exhumation.

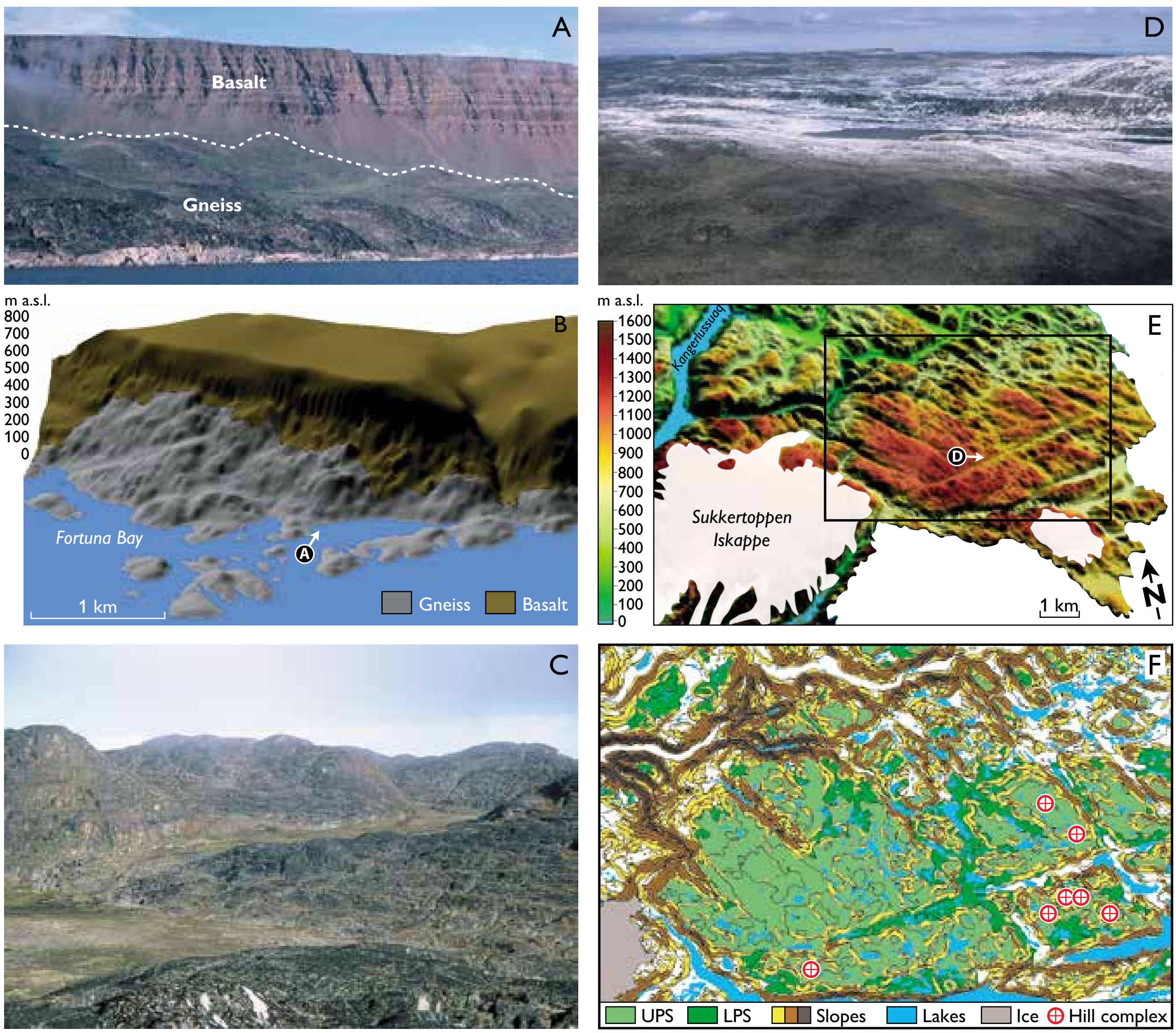


Fig. 45. Four distinct landscape types in West Greenland. **A**, **B:** The re-exposed sub-Paleocene etch surface on southern Disko. The stippled line shows the approximate contact between Paleocene basalt and Precambrian gneiss. The summit plateau is the Upper Planation Surface (**UPS**). **C**: The hilly etch surface on the mainland, just south of Disko Bugt. The hills are up to 100 m high. **D**: The UPS and the Lower Planation Surface (**LPS**), east of Sukkertoppen Iskappe (location shown in **E**). **E**: 3D model of the landscape east of Sukkertoppen Iskappe with the UPS, LPS and incised valleys. **F**: Interpretation of surfaces east of Sukkertoppen Iskappe (location shown in **E**) with UPS, LPS and incised valleys, marked with brown slopes with more than 12° inclination. Photo locations are also marked in Fig. 36. From Bonow *et al*. 2007b.

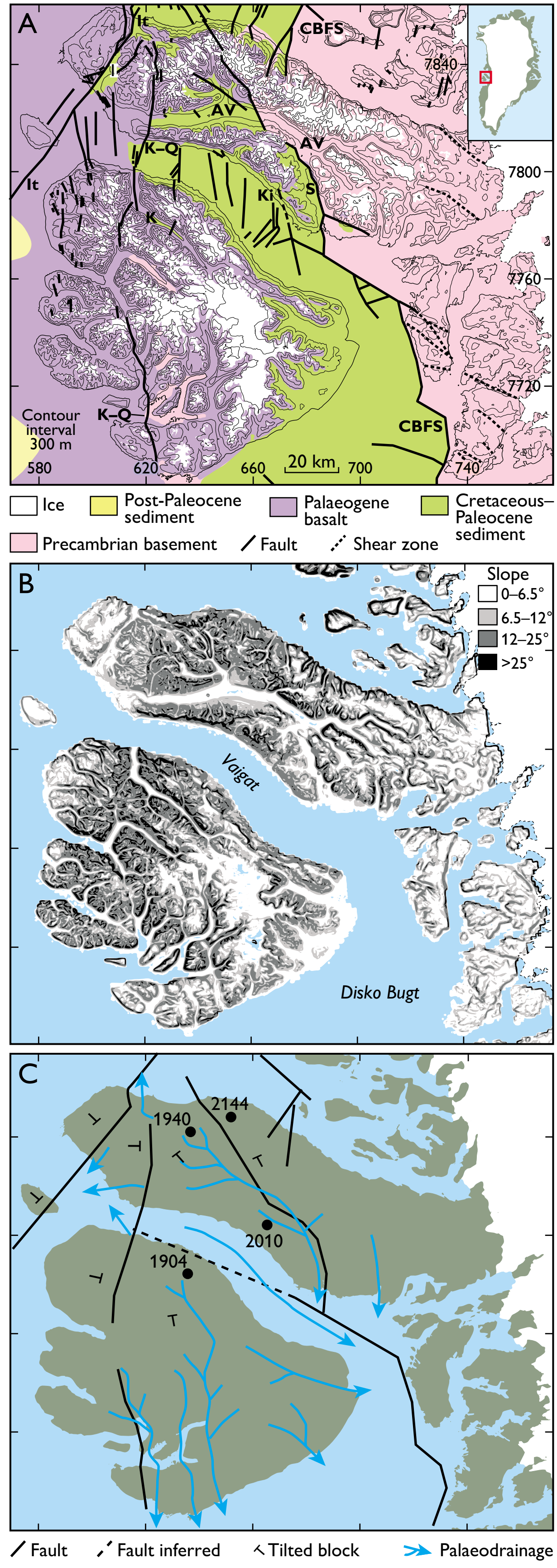


Fig. 46. Reconstruction of the palaeodrainage for Nuussuaq and Disko. **A**: Geology and relief. **B**: Slopes. **C**: Reconstructed palaeodrainage. **AV**: Affarsuaq Valley. **CBFS**: Cretaceous boundary fault system. **It**: Itilli Fault. **Ki**: Kingittoq Fault. **K–Q**: Kuugannguaq–Qunnilik Fault. **S**: Saqqaq Valley. UTM coordinates (km) zone 22N. From Bonow 2004.

The most detailed constraints on the Cenozoic history were obtained from analyses in and around the Nuussuaq Basin, where AFTA data were supplemented by VR data in Cretaceous and Palaeogene sedimentary rocks. Cretaceous and Palaeogene sedimentary and volcanic rocks rest directly on basement in the Nuussuaq Basin, and this immediately shows that the basement was at the surface at the time of deposition. These conditions provide useful geological constraints on the Mesozoic–Cenozoic cooling/exhumation and heating/burial history of the basement rocks.

The tightest constraints on the timing and magnitude of three episodes of Cenozoic cooling were obtained from AFTA and VR data in Palaeogene and Upper Cretaceous sediments in two boreholes (Gro-3 and Gane-1) down to 3 km depth in western Nuussuaq (Figs 36, 49). Combined with results from regional outcrop samples, these data show that the sedimentary units began to cool from their maximum post-depositional palaeotemperatures at some time in the interval 36 to 30 Ma (C2, Eocene–Oligocene transition), with two subsequent cooling episodes beginning in the intervals 11 to 10 Ma (C3, late Miocene) and 7 to 2 Ma (C4, latest Miocene–Pliocene). The Eocene–Oligocene cooling episode is also recognised in AFTA data from outcrop samples of basement and sedimentary rocks across the region, with the highest palaeotemperatures prior to the onset of cooling observed at locations at the western end of Nuussuaq (Fig. 50), defining a distinct 'hot spot' close to the location of the Gro-3 borehole. This 'hot spot' reflects an elevated palaeogeothermal gradient at this location compared to surrounding regions. Note that the two most recent episodes are only recognised in deeper AFTA samples which cooled from palaeotemperatures around 80°C or more in these events (Fig. 51A). In surface and near-surface samples, peak temperatures in these episodes are of insufficient magnitude to be resolved in the AFTA data.

The Eocene–Oligocene cooling episode started 30 million years after the final rifting which occurred around the K/T boundary, more than 25 million years later than peak volcanism during the Paleocene and *c.* 10 million years later than the youngest recorded volcan-

Fig. 47. The well-preserved inland plateau (the Upper Planation Surface) across Precambrian basement and cut by deep valleys and fjords; view towards south-west from a position just south of Sukkertoppen Iskappe (Fig. 34). Plateaux inland are well preserved, although glacial erosion has stripped all shallow saprolites from the surfaces and lakes have formed, but no deep erosion has occurred. The photo illustrates how surfaces of resistant bedrock remain flat even after glacial erosion. Elevation of the plateau about 1 km a.s.l. © Danish Geodata Agency.

ism, yet the AFTA data show that the Upper Cretaceous sedimentary units were hotter at the end of the Eocene than they were during the volcanic and rifting phases. Maximum palaeotemperatures derived from AFTA and VR data in the two boreholes are highly consistent (Fig. 51A), noting that the VR data only record the maximum temperature event. The availability of the VR data provides more reliable definition of the palaeotemperature profile characterising this event than would be possible from AFTA alone which only provides minimum limits on the maximum palaeotemperature in deeper samples.

The combined dataset for the Gro-3 and Gane-1 boreholes define an Eocene–Oligocene palaeogeothermal gradient in the range 39 to 44°C/km, and require that an additional *c.* 1925 m (between 1750 and 2100 m) of section must have been present above the borehole location (close to sea level) at the palaeothermal maximum (Fig. 51B; Japsen *et al.* 2005, note added in proof). Some of this section certainly consisted of basalts that are still present today in the mountains around the borehole (Fig 51C). However, it is unlikely that all of it was basalt and part of the former section above today's highest plateaux may well have consisted of Eocene sediments that transgressed from the west. Such a section is seen on offshore seismic data, but today it is truncated by a shallow unconformity or the sea bed some distance offshore (Fig. 42A; Chalmers 2000).

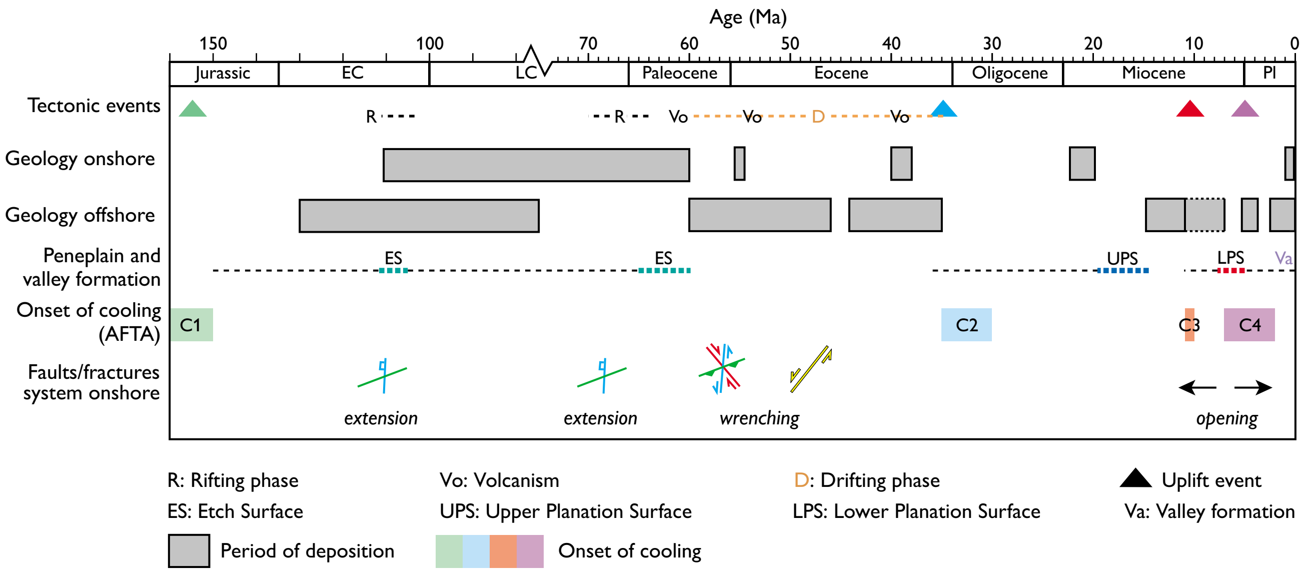


Fig. 48. Summary of stratigraphy offshore and onshore West Greenland, cooling events identified onshore from AFTA, peneplain formation and valley incision, and key regional tectonic events as seen in the structural systems onshore. Onshore and offshore unconformities correspond to regional tectonic events that are identified by stratigraphic landscape analysis and constrained in time by the cooling events defined by AFTA. Weathering and erosion of the etch surface (**ES**) in basement rocks took place following Late Jurassic exhumation (cooling event **C1**, Table 2) and prior to Early Cretaceous burial, and renewed development took place prior to Paleocene volcanism. Following post-breakup subsidence and burial, a first phase of uplift and exhumation that began at the Eocene–Oligocene transition (**C2**), led to the formation of the Upper Planation Surface (**UPS**) during the Oligo–Miocene (see Fig. 54). This surface was offset by reactivated faults, resulting in megablocks that were tilted and uplifted to present-day altitudes of up to 2 km in two phases that began in the Late Miocene and in the latest Miocene–Pliocene, **C3** and **C4**, respectively. The C3 uplift led to incision below the uplifted UPS and thus to formation of the Lower Planation Surface (**LPS**). The peneplains and incised valleys thus reveal information on erosion from periods in the past from where there exist no other geological data. Dotted line: maximum age range of sediments. Onshore stratigraphy from Fig. 37 and Schmidt *et al.* (2005), Pedersen *et al.* (2006). Offshore stratigraphy from Fig. 43. **EC**: Early Cretaceous. **LC**: Late Cretaceous. **Pl**: Plio–Pleistocene. Modified from Bonow *et al.* 2007b and Japsen *et al.* (2009).

5.5 Discussion of AFTA data from West Greenland

Redfield (2010) questioned the interpretation of AFTA data and the supporting information from West Greenland presented by Japsen *et al.* (2005, 2006, 2009), claiming that many factors had been ignored which could have significant impact on the data. However, Redfield (2010) did not mention a large number of publications in which these factors, and the way in which they are accounted for in the interpretative methods employed by Japsen *et al.* (2005, 2006, 2009), were discussed at length (many of these publications are discussed and referenced in sections 4.1 and 4.2). For example, Redfield (2010) claimed that the two Neogene cooling events defined from the AFTA data are 'model artefacts'. However, as discussed above these events are also clearly documented in the landscape and in the geology of the region. The timing constraints on these episodes are derived from AFTA data in samples from depths >1 km which show that the samples were heated to >120°C (supported by VR data). Apatites in these samples have thus only retained tracks since Neogene cooling. Redfield (2010) accepted that regional planation and subsequent uplift/exhumation clearly postdates Eocene basalts which are truncated by the UPS, and further acknowledged that Neogene uplift and erosion "may well have occurred". Green *et al.* (2011) emphasised that their interpretation is supported by a wealth of independent information, and therefore considered their interpretation to be highly reliable.

5.6 Integration of palaeothermal data with SLA

A major problem in previous landscape studies was the difficulty in dating the formation of epigene surfaces and

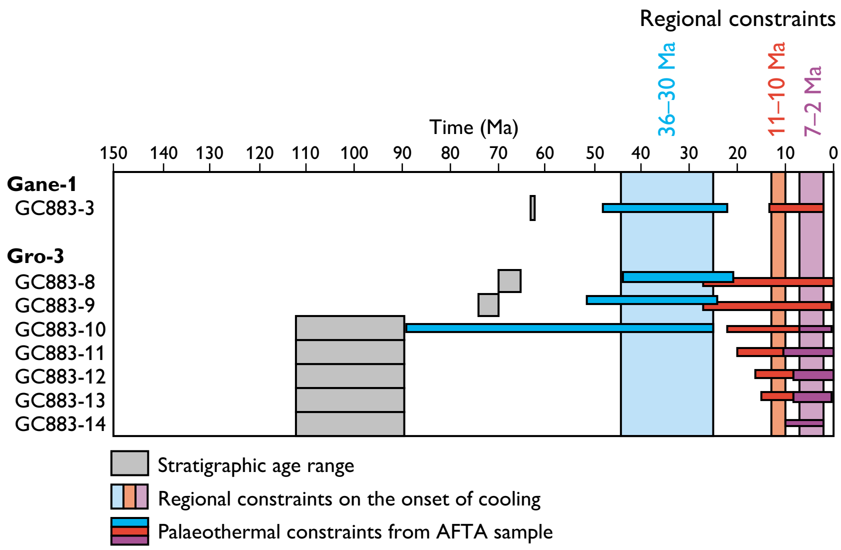


Fig. 49. Palaeothermal episodes recognised from AFTA data in the Gro-3 and Gane-1 boreholes. Synthesis of all data suggests three discrete episodes of cooling (vertical bands). Analysis of a larger dataset from the Nuussuaq Basin allows the onset of cooling episodes to be refined further to 36–30 Ma (C2, Eocene–Oligocene transition), 11–10 Ma (C3, Late Miocene) and 7–2 Ma. (C4, latest Miocene – Pliocene); see Table 2. Modified from Japsen *et al.* (2005).

their uplift. This difficulty has been overcome in West Greenland by combining the relative chronology derived from the landscape with application of AFTA and VR data to define the absolute timing and magnitude of the corresponding erosional episodes as discussed in the preceding section as illustrated in Fig. 52. Here we explore the relation between the three post-breakup episodes of cooling and exhumation revealed by AFTA data and the steps in the landscape defined by the UPS and the LPS.

The cooling episode that began at the Eocene–Oligocene transition is recognised in AFTA results from almost every sample analysed from within the Nuussuaq Basin and in the surrounding basement areas (Japsen *et al.* 2006). The samples recording this event are distributed across an area similar in extent to the UPS that, as described earlier, has been dated independently to be post-mid-Eocene (Fig. 34). Japsen *et al.* (2006) therefore considered that both the formation of the UPS and the cooling at the Eocene–Oligocene transition are expressions of the same episode of exhumation.

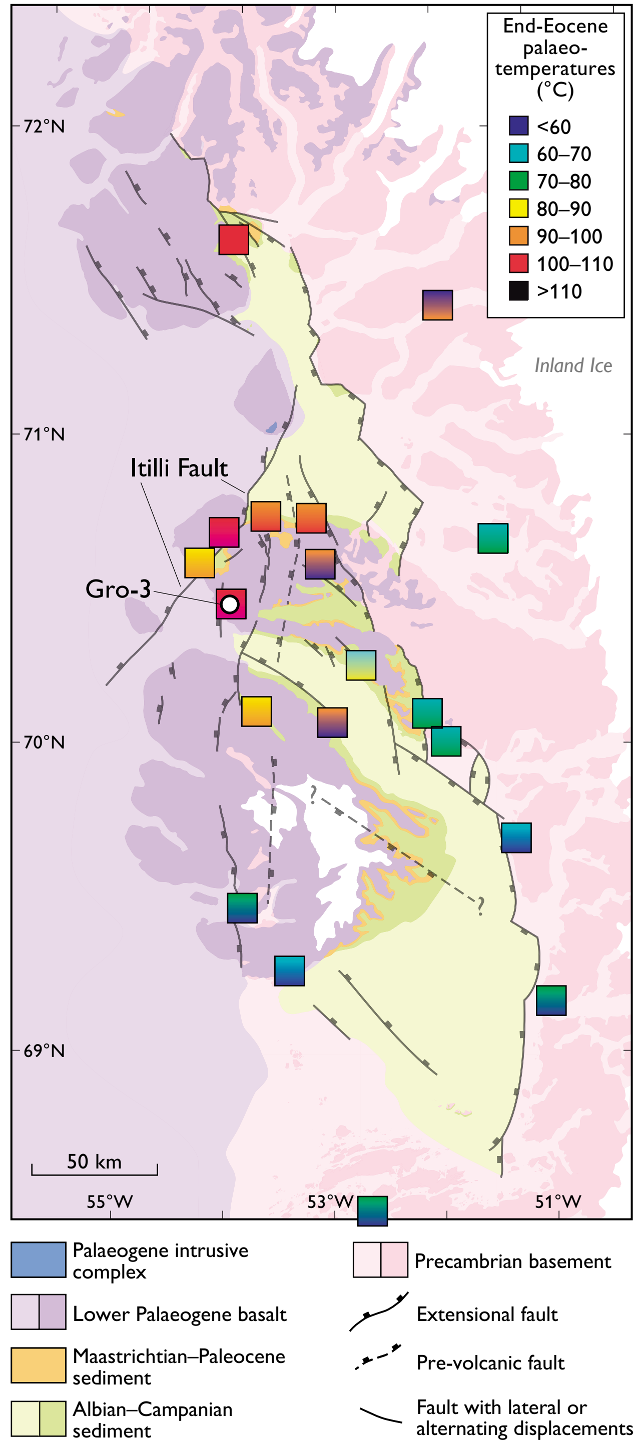
The Gro-3 borehole was sited on westernmost Nuussuaq in a valley bottom near sea level in a mountainous area. This allows the possibility of investigating the relation between the cooling events defined from the Gro-3 data and the elevation of the UPS and LPS that we have mapped in this area (Bonow *et al.* 2006b). As shown in Fig. 51C, the UPS forms the local summit level at a height of *c.* 1250 m a.s.l. (1100–1400 m a.s.l.) in the vicinity of the Gro-3 borehole. At the onset of cooling from the Eocene–Oligocene C2 palaeothermal maximum, a cover of *c.* 1925 m (1750–2100 m) was present above the drill site (Fig. 51B, point 1). Only part of the removed cover can be explained by the basalt present in the adjacent mountains (Fig. 51C) and an additional *c.* 675 m (350–1000 m) of section must have been present above the present-day summit level (UPS) at the palaeothermal maximum. This additional section must have been deposited after the youngest preserved basalt was laid down and prior to the onset of C2 exhumation in the

Table 2. Intervals for onset of cooling episodes in West Greenland

Onset of cooling (Ma)	Stratigraphic interval	Cooling episode
560–500	Latest Neoproterozoic – Cambrian	-
370–355	Late Devonian – Early Carboniferous	-
230–220	Late Triassic	C0
160–150	Late Jurassic	C1
36–30	Eocene–Oligocene transition	C2
11–10	Late Miocene	C3
7–2	Latest Miocene – Pliocene	C4

Intervals defining the beginning of episodes of regional cooling derived from AFTA data in 69 samples from outcrops and boreholes in West Greenland. Cooling episodes after Japsen *et al.* (2006, 2009) and Bonow *et al.* (2007b).

Fig. 50. Eocene–Oligocene (C2, 36–30 Ma, Table 2) palaeotemperatures from AFTA in samples from Nuussuaq and surrounding regions. The map shows a pronounced 'hot spot' on western Nuussuaq close to Itilli Fault (and the location of the Gro-3 borehole). This reflects an elevated palaeogeothermal gradient at this location compared to surrounding regions. Boreholes are represented by the shallowest sample in each case. While samples were taken from different elevations, these results have not been adjusted to a common datum. Because the range of elevations is less than *c.* 500 m this will not significantly distort the regional pattern. Values in individual samples and further discussion in supplementary data files to Japsen *et al.* (2005, 2006). Modified from Green *et al.* (2011).



interval 36 to 30 Ma, and subsequently removed during the formation of the UPS.

Constraints on palaeogeothermal gradients and removed section for the two more recent episodes are less well defined (Fig. 51B), largely because these are only defined from AFTA and palaeotemperature constraints are relatively broad (Fig. 51A). But as indicated by point 3 in Fig. 51B, for a palaeogeothermal gradient of *c.* 30°C/km, the latest Miocene – Pliocene (C4) palaeotemperatures can be explained by around 850 m of additional section, which corresponds with the height of the LPS at this location. In other words, the 'Pliocene' cooling episode, which began between 7 and 2 Ma, can be explained simply by incision and excavation of the present relief below the LPS within the valley where the borehole is located. This event can therefore be interpreted as representing the onset of uplift that took the LPS to its current altitude.

The question then remains as to the evolution between points 1 (C2 maximum burial at *c.* 35 Ma) and 3 (onset of latest Miocene – Pliocene C4 uplift and incision) in Fig. 51B. One option involves a scenario in which there was no cover above the UPS at *c.* 10 Ma when the late Miocene C3 cooling phase began (point 2). At point 2, the cover required above the ground surface to explain the late Miocene palaeotemperatures corresponds to the present-day elevation of the UPS, for a palaeogeothermal gradient of *c.* 40°C/km (i.e. the amount of rock still preserved above the level of the drill site in the sides of the valley). In this scenario, the 10 Ma cooling episode can be explained in terms of excavation of the valley in which the well is sited, and the onset of cooling at 10 Ma would therefore indicate the beginning of uplift leading to incision below the UPS.

With the wide range of possible combinations of removed section and palaeogeothermal gradient allowed by the late Miocene palaeotemperatures, a range of alternative scenarios is possible, as shown by the dashed trajectories in Fig. 51B. Each of these alternative trajectories involves some degree of cover on the UPS at the onset of late Miocene cooling, with the precise amount depending on the value of late Miocene palaeogeothermal gradient. In the previous section, we argued that the UPS obtained its final shape after protracted denudation

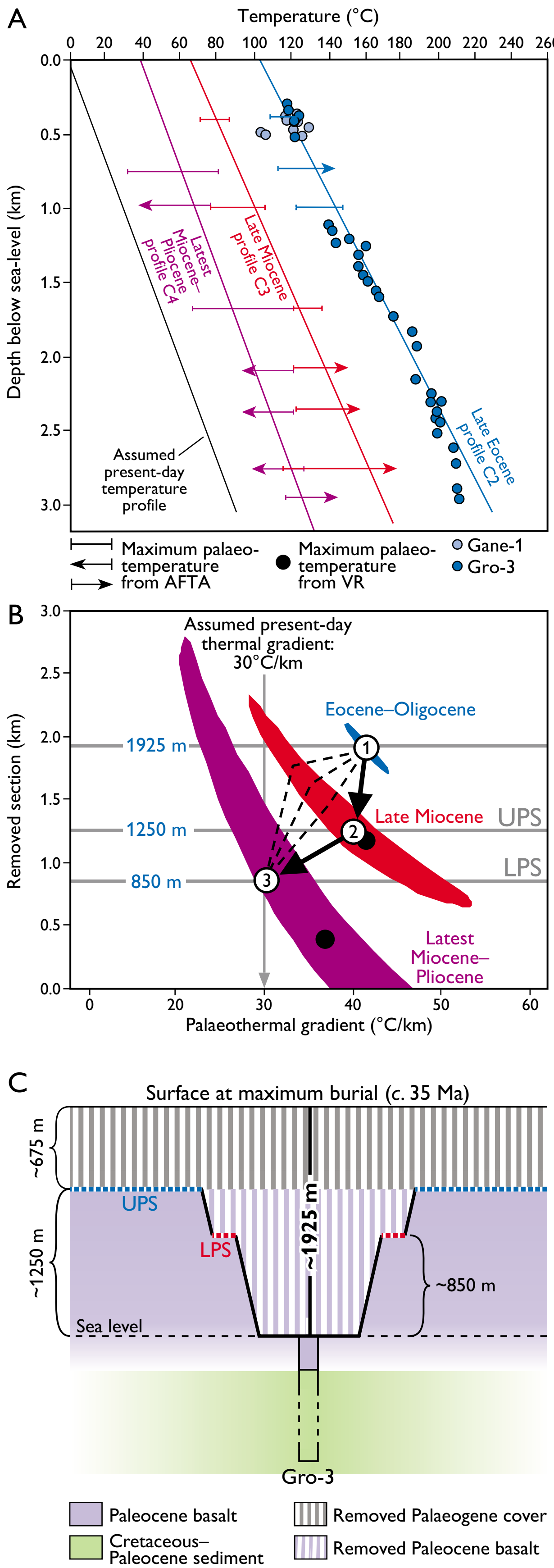
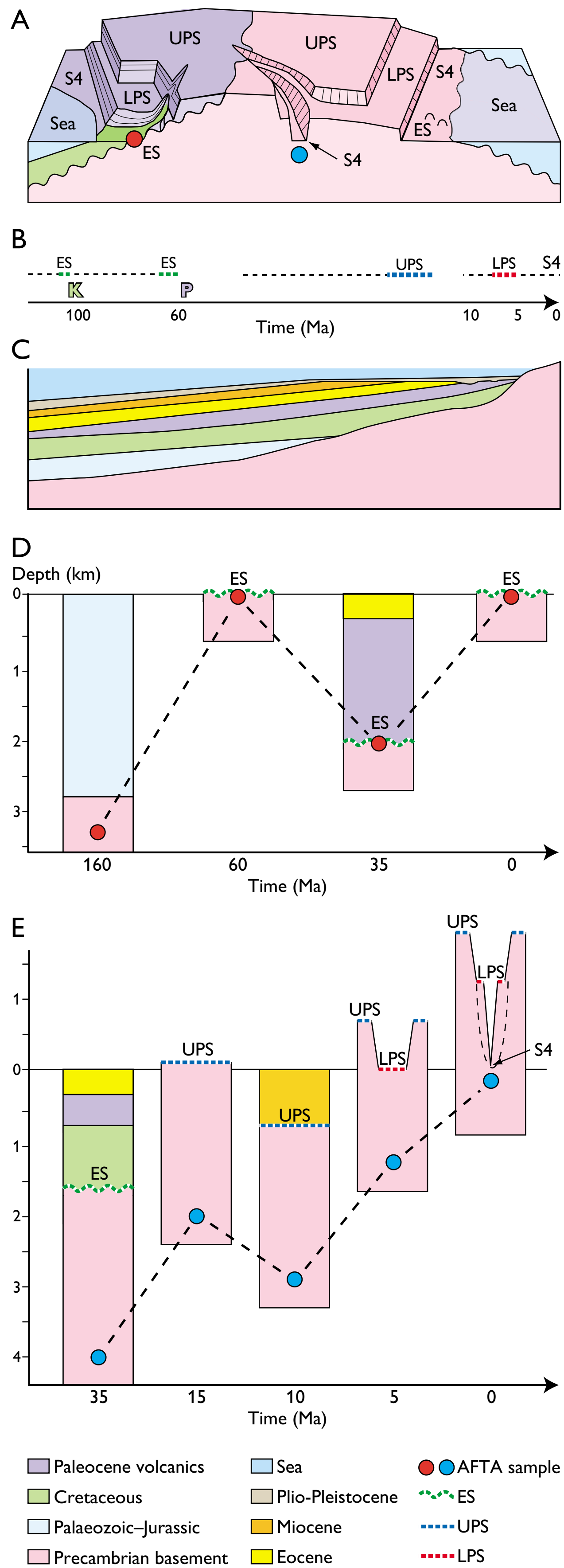


Fig. 51. Reconstruction of the burial, uplift and exhumation history around the Gro-3 borehole, western Nuussuaq, based on AFTA and stratigraphic landscape analysis (location shown in Fig. 36). **A**: Palaeotemperatures in three episodes identified from AFTA and vitrinite reflectance (VR) in the Gro-3 and Gane-1 boreholes; Eocene–Oligocene (C2, onset between 36 and 30 Ma), late Miocene (C3, onset between 11 and 10 Ma) and latest Miocene – Pliocene (C4, beginning between 7 and 2 Ma; Table 2). **B**: Ranges of palaeogeothermal gradient and removed section required (banana-shaped areas) to explain the palaeotemperatures in each episode from **A**. The well-defined palaeothermal maximum (point **1**) corresponds to a cover above the valley floor of *c.* 1925 (1750–2100) m for a palaeogeothermal gradient between 39 and 44°C/km. Point **3** in **B** corresponds to a cover above the valley floor of *c.* 850 m thickness for a palaeogeothermal gradient of *c.* 30°C/km, corresponding with the height of the LPS at this location, and thus the latest Miocene – Pliocene cooling episode can be explained simply by incision of the present relief below the Lower Planation Surface (**LPS**). Several trajectories are indicated between points **1** and **3**. Point **2** represents a scenario in which there was no cover above the Upper Planation Surface (**UPS**) when the late Miocene cooling phase began. At point **2**, the cover required above the ground surface to explain the late Miocene palaeotemperatures corresponds to the present-day elevation of the UPS (*c.* 1250 m or 1100–1400 m a.s.l.), for a palaeogeothermal gradient of *c.* 40°C/km. In this scenario, the 10 Ma (C3) cooling episode can be explained in terms of excavation of the valley in which the well is sited. The onset of cooling at 10 Ma would therefore indicate the beginning of uplift leading to incision below the UPS. The dashed lines in **B** indicate the wide range of possible combinations of removed section and palaeogeothermal gradient allowed by the late Miocene palaeotemperatures. Each of these alternative trajectories involves some degree of cover on the UPS after its final formation, at the onset of late Miocene cooling, with the precise amount depending on the value of late Miocene palaeogeothermal gradient. **C**: Reconstruction of the total amount of rock present above the site of the Gro-3 borehole at the Eocene–Oligocene palaeothermal maximum relative to the elevation of the UPS and LPS in the adjacent mountains; *c.* 1250 (1100–1400) and *c.* 850 (800–900) m a.s.l., respectively. The amount of section present at the Eocene–Oligocene palaeothermal maximum extends above the local summit level (the UPS) by *c.* 675 (350–1000) m; consequently, the formation of the UPS involved removal of that section. Modified from Japsen *et al.* (2005, 2009).

Fig. 52. Conceptual model highlighting the relationships between stratigraphic landform analysis, AFTA and onshore/offshore geology in central West Greenland. The red and the blue dots in sketch **A** mark two AFTA samples with different exhumation histories (see **D** and **E**). **A**: Sketch of landforms in central West Greenland showing four different erosion surfaces from ES (oldest), UPS, LPS to S4 (youngest). The age of a re-exposed etch surface (**ES**) is well constrained due to protective cover rocks. The Upper and Lower Planation Surfaces (**UPS** and **LPS**) cut across both strata of different age and the ES. **S4** is a newly formed surface. The surfaces and the relationships to geology can be directly observed in the landscape. **B**: A relative event chronology from the classical geomorphological interpretation of the landforms. Cover rocks (**K**: Cretaceous, **P**: Palaeogene) constrain the timing of the final formation of surfaces (ES), but may also give the maximum age of surfaces (UPS). The dashed lines indicate the approximate time needed for completion of a surface. The dashed lines indicate the uncertainties for their formation in time. Time line from geology and AFTA data. **C**: The sedimentary sequences along the passive margins are tilted and truncated below the Plio–Pleistocene strata due to late uplift of the landmass. The oldest sequence is thus exposed closest to the coast. A significant unconformity spans late Eocene to late Miocene. Glaciers expanding to the shelves have also contributed to the erosion of offshore sequences. **D**: Uplift and burial history of the red sample in **A** (exposed at ES near Paleocene basalt). AFTA data record Late Jurassic onset of cooling (exhumation) from palaeotemperatures of about 75°C corresponding to burial below a cover of about 2 km depending on the palaeogeothermal gradient. Near Sukkertoppen Iskappe, Palaeozoic sediments were present near the present surface in the Late Jurassic (Japsen *et al.* 2009), implying that the cover at that time included Palaeozoic–Jurassic rocks as in the case shown here. Because the ES formed in exposed basement, this overburden must have been removed prior to deposition of Paleocene basalts (this case) or Lower Cretaceous sediments (elsewhere in the Nuussuaq Basin) that has preserved the ES at certain locations. The ES was again re-exposed at the surface due to uplift and erosion, recorded as onset of cooling by AFTA and as truncation of the sedimentary sequences offshore. **E**: Burial and uplift history of the blue sample in **A**, exhumed at the valley bottom (S4). Formation of the UPS included erosion of both overburden and basement, as the ES was obliterated by the planation event. Reburial of the UPS is suggested by AFTA data (see Fig. 51). Uplift results in the development of a new planation surface, the LPS by valley incision below the UPS, tilted sedimentary sequences offshore and AFTA recording onset of cooling. Further uplift raised the UPS and LPS to their present elevation. The fluvial V-shaped valleys were reshaped by glaciers. The development of the new S4 surface was initiated. Modified from Bonow *et al.* (2007a).



during Oligo–Miocene time, but prior to late Miocene (*c.* 10 Ma) uplift that initiated the dissection. The scenarios involving some degree of cover on the UPS indicated by the dashed lines in Fig. 51B thus require that sediment accumulated on the UPS after its final formation and before late Miocene uplift. Our preferred solution involves some degree of reburial of the UPS. We base this interpretation on the presence of a thick Miocene sequence, tilted and truncated just offshore to the west of Nuussuaq, which provides strong evidence that Miocene strata once extended across at least western Nuussuaq (see section 8.7).

Combining results from Nuussuaq and the basement areas to the south demonstrates that the UPS and LPS were formed by denudation across the entire area of central West Greenland, even in the basement area where no remnants of cover rocks are preserved. Interpretation of AFTA data from outcrop samples up to 1.8 km a.s.l. near Sukkertoppen Iskappe (Fig. 34) shows that the UPS is the end-product of post-Eocene denudation even in basement areas, and that Phanerozoic sediments – most likely of Cretaceous–Palaeogene age – must have been present above this level prior to the onset of denudation (Japsen *et al.* 2009).

In conclusion, the results reviewed above suggest that the three post-breakup events of uplift and exhumation identified in the AFTA data reflect the events that led to the distinct levels in the landscape. (1) Uplift and exhumation that began at the Eocene–Oligocene transition led to formation of the UPS. (2) Uplift and exhumation that began in the late Miocene led to formation of the LPS by incision below the uplifted UPS (after removal of any Miocene cover). (3) latest Miocene – Pliocene uplift and exhumation led to formation of the present-day fjords and valleys below the uplifted LPS. This interpretation implies that the UPS was graded to base level through Oligocene and Miocene times until the late

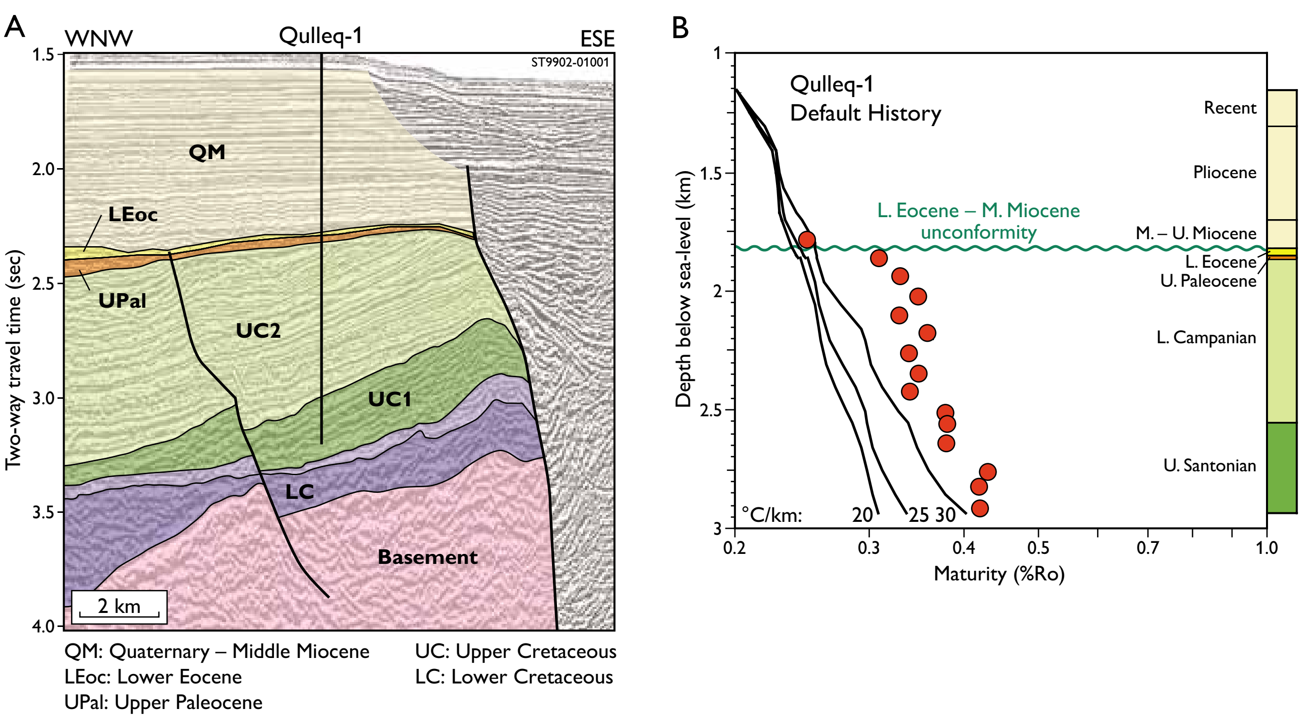


Fig. 53. The Qulleq-1 well, offshore West Greenland. **A**: Seismic line through the Qulleq-1 well. **B**: Plot of vitrinite reflectance (VR, red dots) data vs. depth for the Qulleq-1 well together with the drilled succession. In **A,** note the pronounced angular unconformity below the Neogene succession that truncates lower Eocene, Paleocene and Upper Cretaceous strata. In **B**, black curves indicate predicted VR trends based on the default history (preserved stratigraphy and geothermal gradients ranging from 20 to 30°C/km). The discrepancy between observed and predicted VR values indicates that the section has been hotter in the past, most likely due to deeper burial below a section that was subsequently removed. Only the VR value for the sample above the base-Neogene unconformity matches the default history. This suggests that the early Eocene – middle Miocene hiatus represents the removal of sediments of that age. Location shown in Fig. 44. Modified after Christiansen *et al.* (2001) and Japsen *et al.* (2010).

Miocene, when a phase of uplift initiated the destruction of the UPS by river incision.

5.7 Correlation with the offshore record

Four sedimentary successions of Cenozoic age have been recognised offshore southern West Greenland (Dalhoff *et al.* 2003; Piasecki 2003; Sørensen 2006): mid-Paleocene to late Eocene, middle to late Miocene, early Pliocene and late Pliocene to Pleistocene (Figs 42, 43). These are separated by hiatuses of Oligocene to early Miocene, late Miocene, and early Pliocene age. Deposition of the first part of the earliest sequence corresponds with the period of subsidence recognised in the Nuussuaq Basin from the exposed stratigraphy and palaeothermal data.

The Oligocene to middle Miocene hiatus formed during the same period as the UPS (beginning around the end of the Eocene, *c.* 35 Ma). There is little evidence of truncation of sedimentary reflectors at this unconformity (Fig. 42), so the uplift that formed the UPS seems to have been fairly uniform over a large area of the continental shelf as well as the present-day onshore (Japsen *et al.* 2006). Correlating the two most recent hiatuses with the two phases of late Neogene uplift suggests that the late Miocene hiatus formed during the uplift that began at *c.* 10 Ma and resulting in formation of the LPS, while the early Pliocene hiatus provides a more precise timing for the onset of the final uplift phase at *c.* 4 Ma (Figs 43, 48). The total magnitude of Neogene vertical motions along the coast of West Greenland can thus be defined by the depth of the base Miocene offshore and the elevation of the UPS onshore.

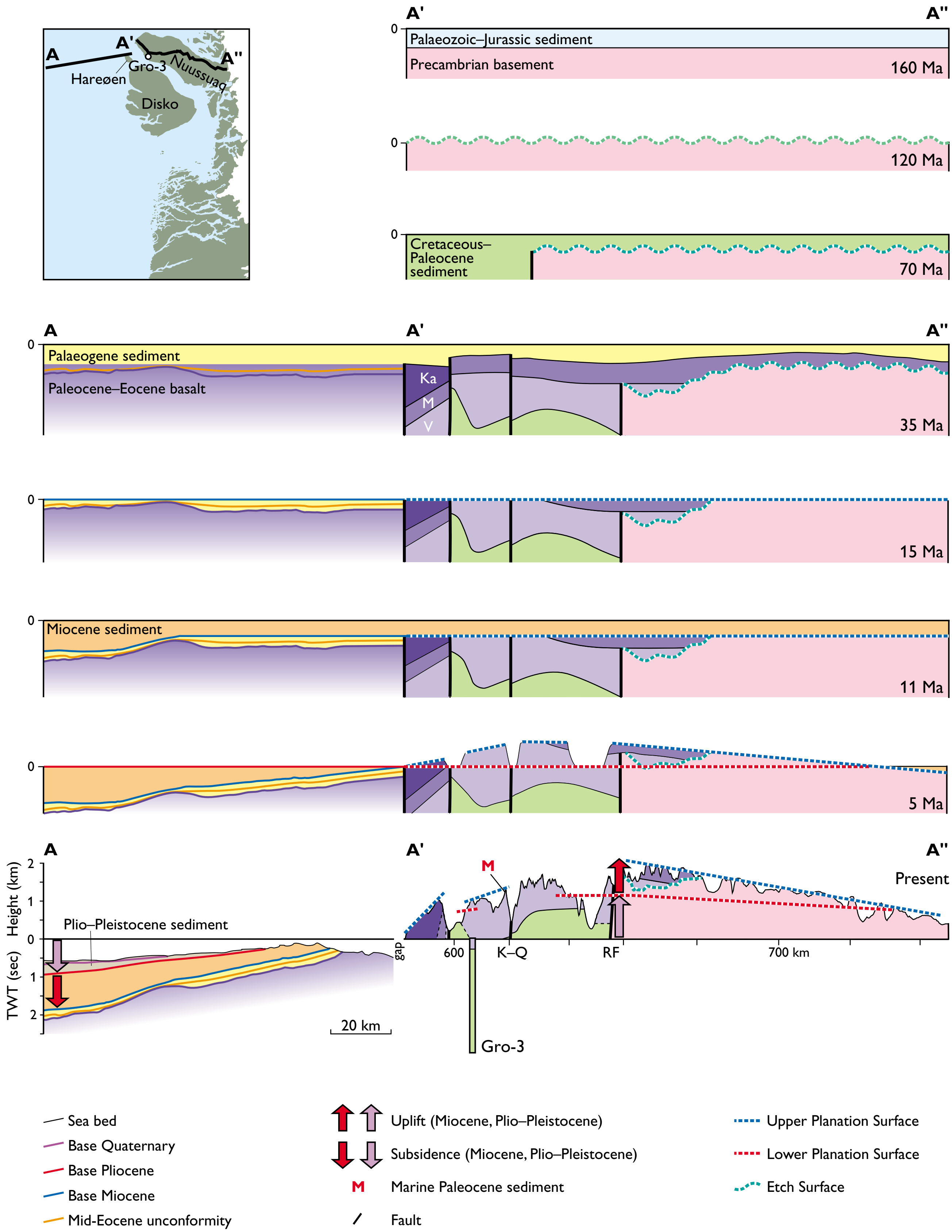
Because of its regional importance, we will discuss the nature of the Oligocene hiatus at some length in the next section.

5.7.1 The nature of the Oligocene hiatus offshore southern West Greenland

Oligocene sediments are absent on the shelf off central and southern West Greenland (Sørensen 2006) and, north of about 66°N, upper Eocene sediments are separated from lower Eocene sediments by another pronounced hiatus (Figs 43, 44; Dalhoff *et al.* 2003). South of about 66°N, the two hiatuses merge into one and upper Miocene sediments lie directly on lower Eocene sediments (Fig. 42; cf. Nøhr-Hansen 2003; Piasecki 2003; Sønderholm *et al.* 2003; Sørensen 2006). The earlier hiatus coincides with the substantial decrease in the speed of sea-floor spreading in the Labrador Sea after Chron 21 and the latter hiatus with the final cessation of sea-floor spreading between Greenland and North America (Chalmers & Pulvertaft 2001).

It is not possible to use only reflection seismic data to evaluate whether a hiatus represents a period of non-deposition or if sediments have been deposited and then removed. The Oligocene hiatus is seen as a very low-angle unconformity on seismic data from much of the Greenland shelf, and it can thus be difficult to judge if it is an erosional unconformity (Figs 42A, B). Such low-angle unconformities are conventionally regarded as representing intervals of stability, but their character seems to be different from the third- and higher order unconformities studied by sequence stratigraphy. Dalhoff *et al.*'s (2003) sequence stratigraphic analysis of the mid-Paleocene to mid-Eocene succession of southern West Greenland identified 11 third-order sequences between the base Cenozoic unconformity and the mid-Eocene unconformity, each separated by a third-order unconformity that probably formed in response to variations in sea-level. Sedimentation in at least the northern Sisimiut Basin was, however, continuous enough to keep the accommodation space filled. The base Cenozoic, Oligocene–Miocene and probably the late Miocene and Plio–Pleistocene unconformities appear to have been the result of a process that interrupted the subsidence of the basin and probably involved uplift and erosion of previously deposited sediment. Evidence for this is available from the Qulleq-1 exploration well.

The Oligocene to middle Miocene unconformity in the area penetrated by the Qulleq-1 exploration well is clearly angular and truncates lower Eocene, Paleocene and Upper Cretaceous strata (Fig. 53A; Christiansen *et al.* 2001). VR data from the Qulleq-1 well suggest that the pre-Neogene succession in the well has been more deeply buried in the past. VR data from the well (Fig. 53B) are seen to plot above curves of predicted VR trends based on the 'default history' derived from the preserved stratigraphy and based on a likely range of geothermal gradients between 20 and 30°C/km (in this history, hiatuses are assumed to represent periods of non-deposition, not erosion). The discrepancy between observed and predicted VR values indicates that the drilled section below the unconformity has been hotter in the past, most likely due to deeper burial below a section that has been removed.



A
A'
A"
Gro-3
Nuussuaq
Hareøen
Disko
0
Palaeozoic–Jurassic sediment
Precambrian basement
160 Ma
120 Ma
Cretaceous–
Paleocene
sediment
70 Ma
Palaeogene sediment
Paleocene–Eocene basalt
Ka
M
V
35 Ma
15 Ma
Miocene sediment
11 Ma
5 Ma
M
Present
Height (km)
TWT (sec)
2
1
Plio–Pleistocene sediment
gap
600
K–Q
RF
700 km
20 km
Gro-3
Sea bed
Base Quaternary
Base Pliocene
Base Miocene
Mid-Eocene unconformity
Uplift (Miocene, Plio–Pleistocene)
Subsidence (Miocene, Plio–Pleistocene)
Marine Paleocene sediment
Fault
Upper Planation Surface
Lower Planation Surface
Etch Surface

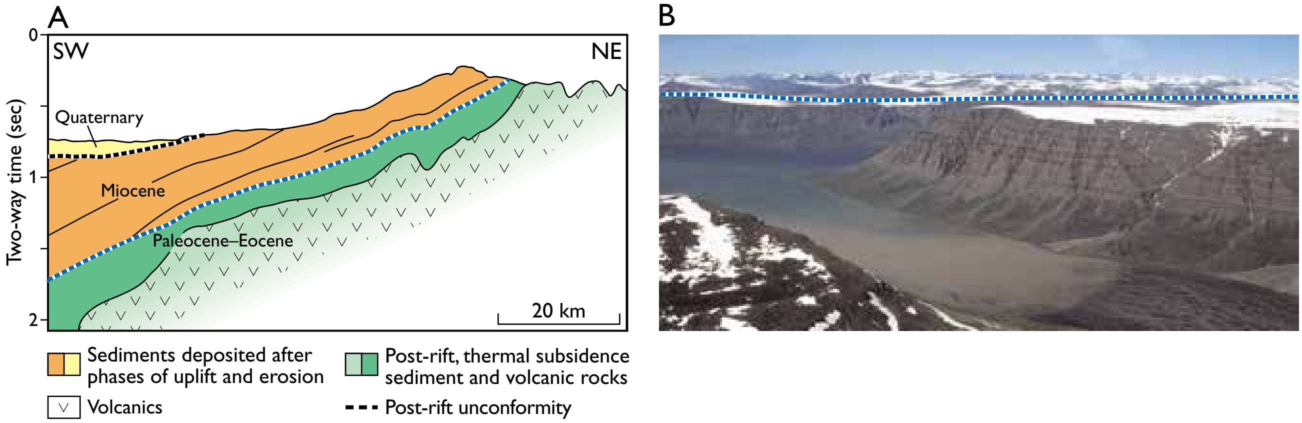


Fig. 55. Offshore-onshore correlation along the elevated, passive continental margin (EPCM) of West Greenland. **A**: Highlighted Oligo–Miocene hiatus (blue dashed line) on a seismic section off Nuussuaq (see Fig. 42A; location shown in Fig. 44). **B**: Highlighted Oligo–Miocene peneplain (the Upper Planation Surface, UPS, blue dashed line) across Paleocene basalts on southern Disko (location shown in Fig. 34). The present-day high topography of the West Greenland EPCM formed since the late Miocene, by uplift and dissection of the UPS, the remnants of which can now be recognised as plateaux at elevations up to almost 2 km close to the sea and as an unconformity at depths of more than 1 km offshore.

This possibility is strengthened by the observation that the VR value for the only Neogene sample matches the default history, indicating that this sample is now at its maximum burial depth. Further studies based on palaeothermal (AFTA and VR) data and palaeoburial (sonic) data and constrained by the preserved stratigraphy, are needed to evaluate the timing, magnitude and extent of the exhumation related to the Oligocene hiatus offshore West Greenland. Such studies would provide firm constraints on both the thermal and the burial history of the preserved sedimentary section, allowing definition of those areas where maximum burial was reached during mid-Cenozoic times.

Studies of the burial and exhumation history west of Greenland are also important for understanding the tectonic evolution of the Arctic during the Cenozoic within a wider context: a hiatus similar to that encountered on the West Greenland shelf is present in the Sverdrup Basin in the eastern Canadian Arctic (Harrison *et al.* 1999), and a similar stratigraphic break was unexpectedly penetrated on the Lomonosov Ridge (central Arctic Ocean) by the Integrated Ocean Drilling Program (IODP) Expedition 302 (Backman *et al.* 2008). The drilled core documented a 26 million years hiatus, separating middle Eocene (*c.* 44 Ma) from lower Miocene sediments (*c.* 18 Ma). Sangiorgi *et al.* (2008) were unable to determine whether that hiatus was generated by sediment erosion or by non-deposition, but emphasised that it conflicts with classical post-rift thermal subsidence models for passive margins. The temporal correlation between this Oligocene hiatus and the time interval, during which the Oligo–Miocene peneplain, UPS, was graded to sea level onshore West Greenland, emphasises the regional controls on such processes, as also highlighted by Green & Duddy (2010).

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Fig. 54. Reconstruction of tectonic events in central West Greenland from the Palaeozoic to the present-day, with focus on the late Cenozoic development (cf. Fig. 48). Late Palaeozoic to Mesozoic cover was removed after Late Jurassic uplift and prior to development of a sub-aerial weathered etch surface (Bonow 2005; Japsen *et al.* 2006, 2009). This surface was covered by Cretaceous sediments and Paleocene basalts. After post-rift subsidence and renewed uplift, a regional Upper Planation Surface (UPS) developed across both basement and basalts. Then followed a period of subsidence, with probable development of a Miocene cover across the UPS; notice the tilted and truncated Miocene sequence just west of Nuussuaq (Fig. 42A) and the freshness of the UPS on the western tip of Nuussuaq (Fig. 41). The subsequent uplift of the UPS resulted in removal of the Miocene sediments and development of the Lower Planation Surface (LPS) until about 5 Ma when a last phase of uplift led to incision below the LPS. **TWT**: two-way travel time. From Japsen *et al.* 2006.

5.8 Tectonic evolution of the West Greenland EPCM

Synthesis of information derived from SLA, palaeothermal methods and the stratigraphic record of the West Greenland margin has defined the tectonic development of its modern day continental margin (Figs 38, 54). The present-day high topography was formed predominantly since the late Miocene, by differential uplift (involving disruption into tilted blocks) and dissection of a peneplain, the remnants of which can now be recognised as plateaux at elevations up to almost 2 km close to the sea and as an unconformity at depths of more than 1 km offshore (Fig. 55). Our results show that an interval of 20 to 25 million years was sufficient to form a regionally extensive peneplain such as the UPS, and imply that erosion over such a period of time was sufficient to erase older surfaces, where they were not protected by cover rocks. On the other hand, an interval of about 6 million years was insufficient time for the valleys of the LPS to coalesce into a regional peneplain.

The development of the West Greenland margin involved multiple cycles of subsidence and burial and subsequent uplift and erosion (Fig. 54). The formation of the elevated terrain in West Greenland postdates the cessation of rifting by *c.* 50 million years, and the initial onset of uplift and erosion which led to formation of the UPS close to sea level appears to correlate with the termination of sea-floor spreading in the adjacent ocean basin (Srivastava 1978). These conclusions contradict the common assumption that continuous uplift, denudation and cooling histories are appropriate for EPCMs, and also suggest that the present landscape of this EPCM is dominated by processes subsequent to rifting, break-up and continental separation. The similarity of the West Greenland EPCM to those in other parts of the world leads us in turn to speculate that the same might be true of other margins. We discuss evidence from a number of margins in support of this speculation in the next chapter.

6. Contrasting views on the development of EPCMs in other areas from landscape studies and thermochronology

6.1 Introduction

The previous chapter demonstrated how integration of low-temperature thermochronology and landscape analysis with basic geological constraints defines the evolution of the continental margin of West Greenland involving multiple episodes of subsidence/burial and subsequent uplift and denudation. In contrast, many published studies of other EPCMS which attempt to integrate results from these three disciplines within a common framework have been less successful. Such studies have commonly been carried out within a paradigm of monotonic cooling and continual denudation, and have in some cases resulted in major inconsistencies between the different approaches. In this chapter we review evidence from EPCMs in southern Africa, south-east Australia and Brazil, classic areas where extensive investigations have been carried out from both thermochronology and landscape analysis, and then briefly review results from other EPCMs. Inconsistencies between interpretations derived from the two approaches in previous studies are highlighted, and the underlying reasons are investigated. Evidence is presented which suggests that the evolution of many EPCMS was in fact more similar to that described for West Greenland than previously envisaged, raising the possibility that this type of development may be the rule, rather than the exception. This has major implications for the nature of the underlying processes, as discussed in chapter 7.

6.2 Southern Africa

6.2.1 Classic landscape studies

Southern Africa is a classical area for landform analysis, with detailed studies resulting in the description of stepped low-relief surfaces graded to the general base level (sea level) formed at a successively upwarped margin (e.g. King 1967, 1972). In these works, and in more recent work by Partridge & Maud (1987), the emphasis is on a history involving continuous uplift and denudation since the eruption of the Lower Jurassic basalts of the Drakensberg Group (the youngest preserved stratigraphic unit of the upper Carboniferous to Lower Jurassic Karoo Supergroup; e.g. Tankard *et al.* 2009), which forms the highest topography in South Africa (Fig. 2). The focus in these studies is on definition and dating of regional planation surfaces representing the staged uplift of southern Africa. King (1967) identified two surfaces above the Drakensberg Escarpment, namely Gondwana and post-Gondwana, and regarded them to be of Jurassic and early-mid Cretaceous age, while he thought the major summit planation surface below the escarpment and inland of it, the African Surface, was formed from the Late Cretaceous to the early Miocene with most erosion in the Late Cretaceous (Table 3; cf. Fig. 2). Three lower landscape generations were interpreted to be the results of Miocene and Pliocene uplift episodes, by correlation with coastal or offshore deposits. Partridge & Maud (1987) came to a different conclusion to that of King (1967), finding no evidence for the existence of preserved Mesozoic surfaces above the African Surface (Table 3). They regarded the African Surface as being present on both sides of the escarpment, and to be the result of a single cycle of erosion from the time of rifting to the early Miocene, with most erosion during the Jurassic to Cretaceous. They further identified two Post-African Surfaces and two Neogene uplift events. In contrast to these earlier studies in which events were seen as extending continent-wide, Moore *et al.* (2009) described the high-level plateau of southern Africa as defined by three separate divides representing 'axes of flexure' dating from Early Cretaceous, mid-Cretaceous and Palaeogene times.

One aspect of the South African landscape that dominates much of the debate regarding the development of the present-day landscape is the Great Escarpment (e.g. King 1962). Ollier & Marker (1985, p. 49) succinctly described the major geomorphic features of southern Africa as "a plateau… bounded by the Great Escarpment, and younger erosional features between the escarpment and the sea". As discussed in section 2.2, Ollier & Marker (1985) interpreted the plateau as an old land surface (the

Table 3. Development of African palaeoplains as described by different authors

King (1972, 1982)	Partridge & Maud (1987)	Partridge (1998)
Gondwana, Jurassic	Non-existing	Non-existing
Uplift		
Post-Gondwana Early Cretaceous; proto-Drakensberg	Non-existing	Non-existing
Uplift		
African Surface, Late Cretaceous (main erosion) – early Miocene; below Drakensberg escarpment	**African Surface**, one cycle of erosion: rifting – early Miocene; deep weathering profiles; most erosion: Jurassic–Cretaceous	Massive denudation (2–3 km) during the Early Cretaceous resulting in the African surface; silcretes; Eocene peripheral cover
Uplift 20 Ma	Uplift	Uplift, early Miocene
Rolling landscape Miocene sediments	Post-African I Miocene – end of Pliocene	Post-African I
Uplift; end of Miocene		
Widespread landscape, Pliocene		
Uplift	Uplift; end of Pliocene	Major Neogene uplift
Youngest landscape Quaternary (Valley of a Thousand Hills)	Post-African II	Post-African II and gorges

'palaeo-plain'), with significant denudation restricted to the region between the escarpment and the coast. This concept was developed further by Ollier & Pain (1997, p. 1) who described the palaeoplain as "little changed from the land surface that existed before continental break-up", correlating with a 'basal unconformity' in the offshore sedimentary section. Within this concept, this elevated ancient surface has never been covered by a younger sedimentary section and has not undergone significant denudation.

Kempf (2010) provided an extensive review of alternative previous viewpoints in regard to the Great Escarpment of southern Africa, pointing out that in addition to the view of the Great Escarpment as a remnant of rifting, others have regarded it as a post-rift feature. Kempf (2010) concluded on the basis of morphological and geological evidence that the Great Escarpment of north and south Namibia is the end-product of denudation resulting from tectonic uplift of the continent following rifting and break-up and formation of the South Atlantic Ocean.

In these and other classical studies of southern African landscapes, amounts of denudation are limited essentially to those required for 'infilling' of the landscape to the level of these planation surfaces, and no consideration is given to the possibility that extensive sedimentary sequences may have been deposited and subsequently removed, in some cases even when sedimentary outliers are preserved. However, application of low-temperature thermochronology to southern Africa provides a quite different view, as reviewed below.

6.2.2 Low-temperature thermochronology studies

Numerous apatite fission-track studies across southern Africa over the last 20 years (Brown *et al.* 1990; Gallagher & Brown 1999a, b; Brown *et al.* 2000, 2002; Raab *et al.* 2002, 2005; Kounov *et al.* 2008, 2009; Tinker *et al.* 2008a) have established that the region has undergone major denudation during the Cretaceous. A map of apatite fission-track ages, based on these studies, is shown in Fig. 56. A more extensive dataset, showing similar features, is illustrated by Gallagher *et al.* (1998), but full details of this dataset have not yet been published. Across much of the region, including areas extending more than 500 km inland from the west and south coasts, apatite fission-track ages are less than 145 Ma. These fission-track ages cannot be interpreted directly as indicating the timing of any specific event (section 4.1.6), but because the analysed samples are Jurassic or older, these results can be regarded as indicating Cretaceous (or later) cooling from palaeotemperatures around 100°C or above over a wide region.

The cooling revealed by the apatite fission-track ages shown in Fig. 56 has been explained by the authors of

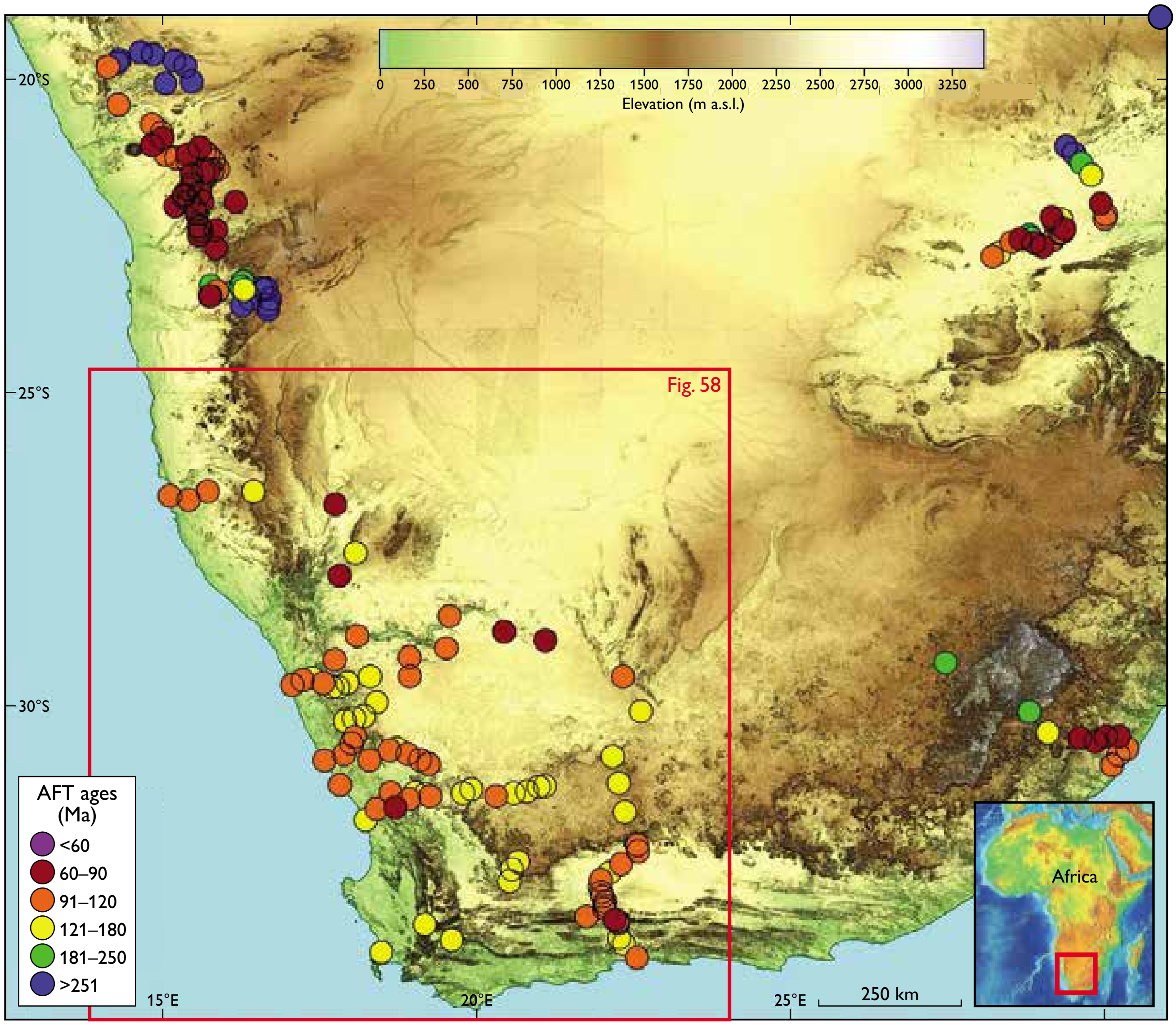

Fig. 56. Map of elevation and published apatite fission-track ages across southern Africa (see section 6.2.2 for sources). Ages as young as 120 Ma or less, which clearly post-date rifting and continental separation, are found hundreds of kilometres inland and bear no obvious relationship to the topography, including the location of the Great Escarpment.

the respective studies solely in terms of denudation following Early Cretaceous rifting, on the basis that denudation is the only mechanism capable of producing significant cooling at rifted margins (Gallagher & Brown 1997; 1999a, b; Brown *et al.* 2000, 2002). In Namibia, estimates of the total amount of section removed following rifting are around 3 km or more across a region extending 300 km or more inland and as much as 5 km at the coast (Gallagher & Brown 1999a, b; Brown *et al.* 2000, Raab *et al.* 2002, 2005), with a major phase of accelerated cooling in the Late Cretaceous (*c.* 70 Ma). On the Natal coast in the south-east (see Fig. 2), Brown *et al.* (2002) estimated that a minimum of 4.5 km of section has been removed since 130 Ma. Farther inland, but still seaward of the Drakensberg Escarpment, Brown *et al.* (2002) estimated around 3 km of denudation since *c.* 91 Ma, while west of the Lesotho Highlands, they estimated around 1.7 km of denudation since 78 Ma, with accelerated denudation through the Late Cretaceous and much lower rates through the Cenozoic at both locations. Tinker *et al.* (2008a) interpreted their results in terms of two dominant phases of denudation, in the intervals 140 to 120 Ma and 100 to 80 Ma, with between 2.5 and 3.5 km of section removed in the Late Cretaceous phase.

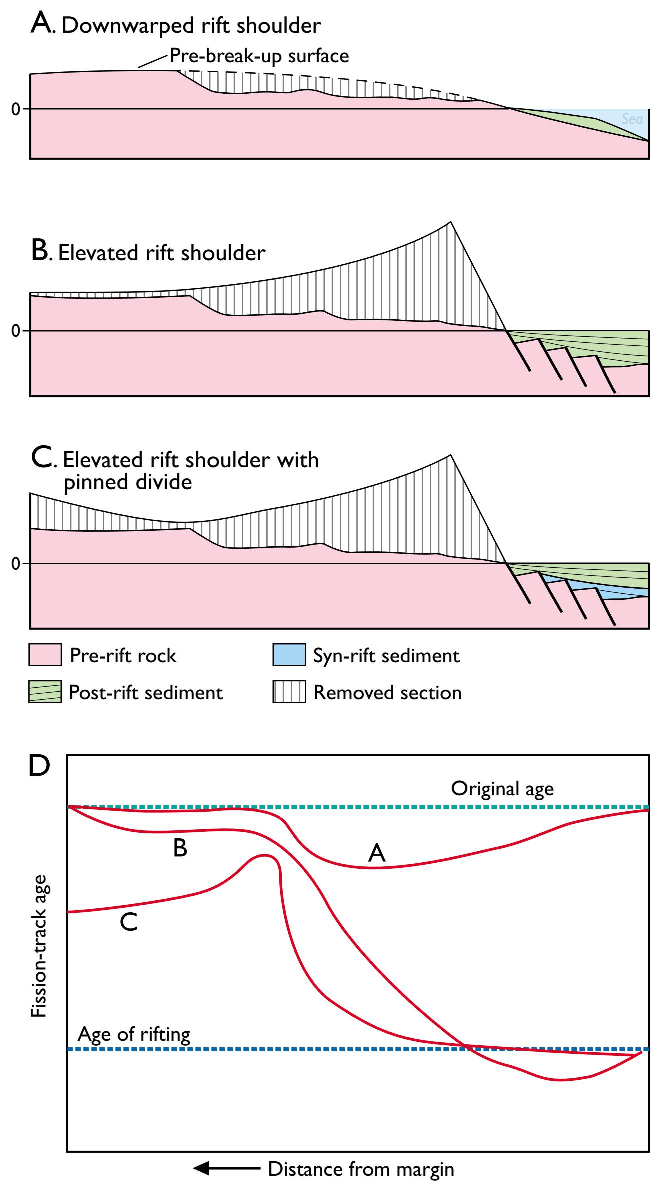

Fig. 57. Geomorphological models that are commonly used in discussions of the development of elevated, passive continental margins (EPCMs; after Gallagher *et al.*1998). **A**: Down-warped rift shoulder (Ollier & Pain 1997). **B**: Scarp retreat from an elevated rift shoulder (e.g. Gilchrist & Summerfield 1991). **C**: Down-wearing of an elevated rift shoulder with pinned divide (e.g. Brown *et al.* 2002; Persano *et al.* 2006). **D**: Variation of fission-track age vs. distance expected from models **A**, **B** and **C** (red curves) compared with 'Original age' (green) and 'Age of rifting' (blue). In these models significant post-rift denudation (and hence apatite fission-track ages close to or younger than the time of rifting) is limited to the coastal side of the escarpment. But as shown in Figs 56, 58 and 60, results from two much-studied EPCMs suggest that young fission-track ages, denoting significant post-rift cooling, extend much farther inland than the escarpment, and these models do not provide an accurate description of the denudation histories of EPCMs.

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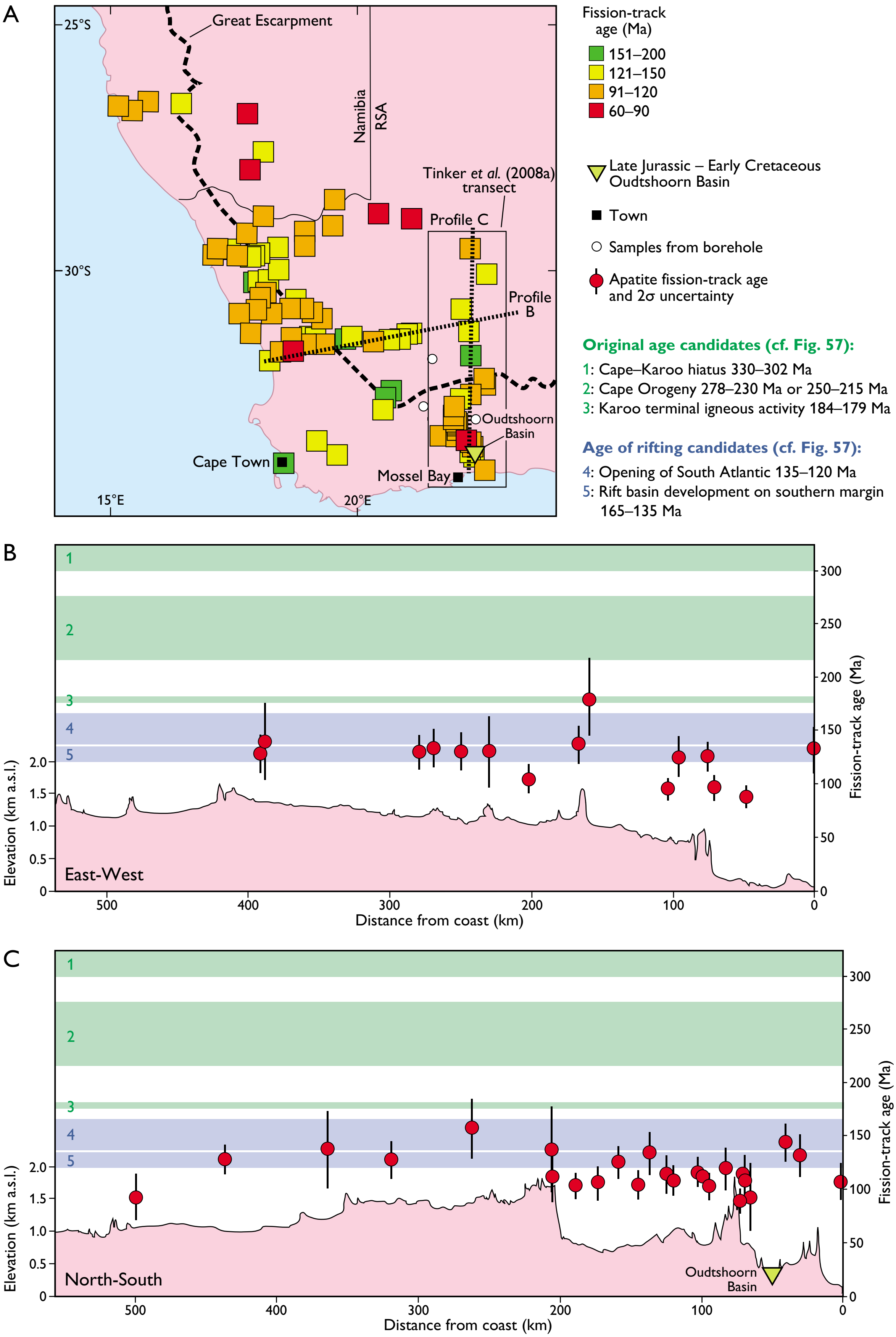
Fig. 58. Variation of apatite fission-track age with elevation across the south-west African elevated, passive continental margin (EPCM). **A**: Map showing a close-up of the fission-track age map of southern Africa in Fig. 56. Profiles **B** and **C**: Data points (red dots) extending up to 500 km or more into the continent allow a direct comparison with the concepts represented in Fig. 57 (data from Tinker *et al.* 2008a and Kounov *et al.* 2009). These panels are colour coded to match the representation of parameters in Fig. 57D, viz: 'Original age' (green), and 'Age of rifting' (blue). Three options are illustrated for the 'Original age' which effectively corresponds to the time at which track retention began in the apatites. The most likely candidate is the extensive Karoo igneous event (candidate 3 in panels B and C; 184–179 Ma, Duncan *et al.* 1997; Svensen *et al.* 2012) while other possibilities are the Cape Orogeny (candidate 2), dated to either 278–230 Ma (Newton *et al.* 2009) or 250–215 Ma (Tankard *et al.* 2009) or the hiatus between deposition of the Cape and Karoo Groups which Tankard *et al.* (2009) assign to the interval 330–302 Ma (candidate 1). Ages of rifting on the southern and western margins are also indicated (candidates 4 and 5 from Macdonald *et al.* 2003). Measured fission track ages do not show the type of behaviour shown by any of the notional trends in Fig. 57, and we conclude that these models do not provide an accurate depiction of the evolution of EPCMs. **RSA**: Republic of South Africa.

6.2.3 Attempts to reconcile landscape studies and thermochronology in southern Africa

The results of these and many other low-temperature thermochronology studies of EPCMs are commonly discussed within a framework in which the modern topography is assumed to be related to continental rifting and break-up, and the onshore margins are assumed to have remained elevated since break-up, with significant denudation restricted to the region seaward of the Great Escarpment. Gallagher *et al.* (1998) outlined three models embodying these concepts, as illustrated in Fig. 57A–C. These models are expected to produce different patterns of apatite fission-track age variation (Fig. 57D), which could be diagnostic in understanding the nature of the processes controlling the development of EPCM landscapes.

Although these models have been widely employed as a conceptual framework for understanding low-temperature thermochronology data from passive margins (e.g. Bishop 2007; Campanile *et al.* 2008; Burke & Gunnell 2008), the results shown in Fig. 56 do not display the patterns of age variation expected from these models. Gallagher & Brown (1999a, p. 48) concluded that "data are broadly consistent with a model where the majority of



A
Great Escarpment
25°S
30°S
15°E
20°E
Namibia
RSA
Tinker et al. (2008a) transect
Profile C
Profile B
Oudtshoorn Basin
Cape Town
Mossel Bay
Fission-track age (Ma)
151–200
121–150
91–120
60–90
Late Jurassic – Early Cretaceous Oudtshoorn Basin
Town
Samples from borehole
Apatite fission-track age and 2σ uncertainty
Original age candidates (cf. Fig. 57):
1: Cape–Karoo hiatus 330–302 Ma
2: Cape Orogeny 278–230 Ma or 250–215 Ma
3: Karoo terminal igneous activity 184–179 Ma
Age of rifting candidates (cf. Fig. 57):
4: Opening of South Atlantic 135–120 Ma
5: Rift basin development on southern margin 165–135 Ma
B
East-West
Elevation (km a.s.l.)
Fission-track age (Ma)
Distance from coast (km)
C
North-South
Oudtshoorn Basin
Elevation (km a.s.l.)
Fission-track age (Ma)
Distance from coast (km)

denudation occurs in what is now the low-elevation coastal plain, seaward of the topographic escarpment..., with regional anomalies in the interior and along-strike of the margin related to post-breakup reactivation of regional structures." But as is clear from Fig. 56, ages as young as 100 Ma or less occur up to 500 km inland, well away from any structures that may have been reactivated.

This is further emphasised in Fig. 58, where measured fission-track ages from the south-west of South Africa along two transects shown in Fig. 56 are plotted against distance from the coast, together with the respective topographic profile along the transect. For purposes of comparison with Fig. 57, various candidates for the 'Original age' are shown in green, and two possibilities for 'Age of rifting' are shown in blue, while data are shown in red to match the colour coding of lines in Fig. 57. Measured fission-track ages are consistently lower than the various candidate 'Original ages' across the whole length of both transects, over distances of over 500 km. The conceptual models illustrated in Fig. 57 therefore provide little guide to the processes actually responsible for the observed pattern of fission-track ages across southern Africa.

Estimates of the amount of section removed by Cretaceous denudation derived from apatite fission-track data have traditionally proven difficult to reconcile with results from studies based on analysis of landforms. Swart (2006) has drawn attention to these problems in Namibia, considering interpretations of apatite fission-track data in terms of up to 5 km of post-Early Cretaceous denudation at coastal locations and up to 3 km or more inland to be totally unrealistic on the basis of the preserved geological section. Aizawa *et al.* (2000) provided a view of the evolution of onshore Namibia based on stratigraphic and geomorphic constraints which bears little relation to the history suggested by the interpretations derived from the apatite fission-track data, with only small thicknesses of section removed since continental break-up. Some syntheses of onshore denudation histories based on such 'conventional approaches' (Bluck *et al.* 2005; Goudie 2005) fail to mention the apatite fission-track data and their interpretations at all, highlighting the disparity between conclusions derived from the two approaches and the difficulties of reconciliation.

Fission-track ages as young as 100 Ma or less in samples from inland of the Great Escarpment in Fig. 56 are incompatible with the idea that the inland plateau represents a pre-rift land surface (Ollier & Pain 1997) or any trace thereof. Partridge (1998) attempted to reconcile the requirement for 3 km of missing section to explain apatite fission-track data from South Africa with geological and geomorphological information by suggesting that all erosion was essentially completed by the end of the Early Cretaceous. However, as highlighted above, this is inconsistent with the timing of denudation defined by these apatite fission-track studies, which require significant amounts of Late Cretaceous denudation. Studies of the morphology of kimberlite pipes also suggest much lower amounts of denudation than indicated by the apatite fission-track interpretations, suggesting a maximum of *c.* 1.8 km of post-Early Cretaceous denudation (Hawthorne 1975), while studies of different types of xenoliths in kimberlites of different age suggest less than 1 km of denudation since *c.* 85 Ma over most of the region (Hanson *et al.* 2009).

Burke & Gunnell (2008) proposed a two-stage model to explain the pattern of apatite fission-track ages across southern Africa, involving scarp retreat from initial rifting (similar to Fig. 57A) followed by uplift of a "Great Swell" and resulting erosion (equivalent to Fig. 57B) over the last 30 million years. Burke & Gunnell (2008) suggest that this will produce a pattern similar to trend C in Fig. 57D, referencing Gallagher *et al.* (1998) as suggesting that this trend provides the best description of the measured age variation. The rationale behind this two-stage model is not clear, as Burke & Gunnell (2008) also say that erosion over the last 30 million years in their scenario will not have much impact on the apatite fission-track data. So it remains unclear how a pattern of ages similar to trend C in Fig. 57D, which shows partially reset ages extending inland of the Great Escarpment, will result from this model. As shown in Fig. 56 neither the model espoused by Burke & Gunnell (2008) nor any of the models illustrated in Fig. 57 can explain the observed pattern of apatite fission-track ages, which shows pervasive, uniform, young fission-track ages close to or younger than the time of break-up extending hundreds of kilometres inland from both the southern and western continental margins.

6.2.4 Possible alternative explanations

In considering why apatite fission-track studies and other approaches lead to such disparate views of the denudation history of south-west Africa, one factor that requires reassessment is the assumption in previous studies that denudation is the only process that causes significant cooling at continental margins (i.e. palaeotemperatures solely represent greater depth of burial, and palaeogeothermal gradients were similar to present-day values). Elevated heat flow associated with rifting would significantly reduce the amount of additional burial (and subsequent denudation) required to explain the palaeotemperatures indicated by the apatite fission-track data. AFTA data from a number of margins, including southeast Australia (Duddy *et al.* 1994; Duddy 1997; Green *et al.* 2004), West Greenland (Japsen *et al.* 2005, 2006) and West Africa (Bray *et al.* 2002; Turner *et al.* 2008), have provided abundant evidence of significantly elevated palaeogeothermal gradients, in most cases associated with continental rifting and break-up. However, this is far from universal, as noted at the UK North Atlantic margin by Green *et al.* (1999) and the Atlantic margin of Brazil by Japsen *et al.* (2012b; section 6.3). Evidence in support of elevated Cretaceous heat flow along the Atlantic margin of south-west Africa was reported by Whitehead *et al.* (2002) who concluded, from the mineralogy of mantle xenoliths in Upper Cretaceous intrusives, that the Late Cretaceous heat flow in coastal locations was twice the value in the vicinity of the Gibeon kimberlites 300 km inland. In addition, Tinker *et al.* (2008b) reported that while the timing of denudation inferred from apatite fission-track data in outcrop samples from the south-western part of South Africa (Fig. 58) showed a good match to that of accelerated deposition offshore, the amount of sediment preserved in offshore basins is less than half of that expected on the basis of the denudation required to explain the apatite fission-track data. This mismatch can be readily explained if the Cretaceous heat flow was higher than the present day value, although erosional removal of offshore sediments (e.g. McMillan 2003) also contributes to some degree in explaining the discrepancy.

Tinker *et al.* (2008b) also drew attention to the temporal coincidence between episodes of Cretaceous igneous activity and enhanced denudation in South Africa. The contribution from intrusive activity to the thermal histories revealed by apatite fission-track data in Namibia has not been considered in published studies to date. Numerous episodes of igneous activity are recognised across Namibia, including Lower Cretaceous anorogenic complexes (e.g. Milner *et al.* 1995), Upper Cretaceous bodies such as the Gross Brukkaros structure and the regionally extensive Gibeon kimberlite field with ages of *c.* 77 Ma and between 72 and 79 Ma, respectively (Reid *et al.* 1990), as well as a number of alkaline plugs (Whitehead *et al.* 2002). Cenozoic igneous activity includes the Klinghardt Phonolites around Luderitz (*c.* 37 Ma, Marsh 1975; Lock & Marsh 1981). Possible heating mechanisms that may be manifested in the AFTA data include elevated heat flow, local exhumation related to thermal doming or hydrothermal effects, which are particularly pronounced around the Gross Brukkaros intrusion (Miller 2008).

AFTA and VR data from the heavily intruded sequences of the UK Atlantic margin have shown that the thermal effect of intrusive bodies can be more widespread than that expected on the basis of simple conductive heating (Parnell *et al.* 1999; Duddy *et al.* 1994, 1998). By comparison it seems likely that the multiple episodes of igneous activity that have affected Namibian coastal regions may well have produced profound palaeothermal effects, which may have been interpreted in terms of deeper burial in previous studies.

A key factor in many of the published apatite fission-track studies of southern Africa is that they do not take into account geological constraints on the underlying thermal and denudational histories, and in many cases the interpretations derived from apatite fission-track data appear to conflict with geological evidence. For example, Late Cretaceous cooling reported by Raab *et al.* (2002) in outcrop samples from northwest Namibia in and around the Damara fold belt was interpreted solely in terms of denudation (as above), involving removal of several kilometres of section. Such an explanation is difficult to reconcile with geological evidence, given the presence in that area of outcropping Jurassic sediments at Waterberg and Mt Etjo (see Miller 2008), as well as the Lower Cretaceous Etendeka volcanics closer to the coast, which show that the present-day surface was also close to the surface during the Mesozoic. This implies that much of the rock that was removed during Late Cretaceous exhumation across the region must have been first deposited on top of these Mesozoic units before being subsequently removed.

Recently, Dauteuil *et al.* (2013) presented a history of the Namibia margin in which the palaeothermal effects defined from the apatite fission-track data were explained by deeper burial in the Early Cretaceous by up to 5 km of Etendeka volcanics, subsequently eroded in the Late Cretaceous. This explanation is more compatible with stratigraphic constraints than removal of basement or Karoo Supergroup rocks as suggested by Raab *et al.* (2005), although it appears to be open to speculation as to whether the Etendeka basalts ever attained such thicknesses.

Inconsistencies between the results of apatite fission-track and landscape studies in southern Africa should not detract from the evidence in Fig. 56 that rocks now outcropping across much of southern Africa were at 100°C or more during the Cretaceous prior to the onset of exhumation. Since such palaeotemperatures are unlikely to occur within 1 km of the surface for any reasonable geothermal gradient, these data suggest that significant amounts of section must have been removed across the region, and geomorphological studies that fail to account for the cooling reflected in the apatite fission-track data cannot be considered realistic. This implies that the major planation surfaces that now define the elevated topography across southern Africa were formed long after rifting and break-up, by erosion that included removal of substantial thicknesses of sedimentary cover.

Plausible interpretations of the apatite fission-track data across southern Africa will only emerge by considering the thermal history data within the context imposed by independent geological and geomorphological constraints. As we illustrate below, consideration of existing data in this light suggests a more complex evolution than the continuous denudation histories employed in all the studies discussed above.

6.2.5 Evidence for post-breakup subsidence and burial of the southern Africa margin

All of the apatite fission-track and geomorphological studies referred to above are interpreted in terms of continual cooling and long-term denudation. However, as illustrated here, data presented by Tinker *et al.* (2008a) from South Africa suggest that the development of the southern margin of Africa may have been more similar to that of West Greenland presented in chapter 5.

Tinker *et al.* (2008a) reported apatite fission-track data in outcrop samples, collected along a roughly north-south transect (Fig. 56), extending from the south coast and crossing the Great Escarpment into the elevated inland region to a point *c.* 500 km from the coast. The region south of the escarpment contains considerable relief, making up the mountains of the Cambrian to Carboniferous Cape Supergroup reaching over 2000 m a.s.l., but the erosional base of the landscape descends from about 800 m a.s.l. south of the escarpment to sea level at the coast. Tinker *et al.* (2008a) analysed samples of the Cape Supergroup and older (Pre-Cape) units in the south, and various Permian and Triassic units of the Karoo Supergroup in the north of their transect, plus additional samples from three deep boreholes in the Karoo sequence south of the escarpment. Results from these samples were interpreted as defining major Late Cretaceous (between 100 and 80 Ma) cooling, while samples from a fourth borehole north of the escarpment were interpreted as showing an Early Cretaceous phase of cooling.

Tinker *et al.* (2008a) attributed Late Cretaceous cooling to denudational removal of a once much thicker Karoo sequence extending in time to the Jurassic Drakensberg volcanics (*c.* 183 Ma; Duncan *et al.* 1997). For Karoo Supergroup samples from the northern part of their transect (Fig. 58C), Tinker *et al.* (2008a) showed that the required thickness of former cover can easily be accommodated by the thicknesses of younger Karoo units preserved elsewhere in the basin. But at locations in the south of the transect, the presence of Upper Jurassic to Lower Cretaceous sedimentary units of the Uitenhage Group, deposited mainly between *c.* 151 and *c.* 135 Ma (Shone 2009), in extensional basins such as the Oudtshoorn Basin (location shown in Fig. 58), rules out such an interpretation. While the northern margins of these basins tend to be fault-controlled, at southern basin margins and elsewhere the sedimentary units of the Uitenhage Group rest directly on Palaeozoic metasedimentary rocks of the Cape Supergroup and/or older basement. Therefore, even if any units of the Karoo Basin sequence were deposited in this region, for which there is no evidence (Johnson *et al.* 2009), they must have been removed prior to the Late Jurassic.

Late Cretaceous (100 to 80 Ma) temperatures of up to *c.* 100°C are expressed in the fission-track data of Tinker *et al.* (2008a), in outcropping samples of Cape Supergroup and older units across this region, including one sample close to the southern margin of the Oudtshoorn Basin where Uitenhage Group sediments are in depositional contact with underlying units. These rocks must have been at or close to the surface in the Late Jurassic to Early Cretaceous, and therefore burial of the Palaeozoic and older rocks by a much thicker sedimentary cover post-dating the preserved sedimentary units of the Uitenhage Group is required in order to achieve the estimated temperatures prior to the onset of Late Cretaceous denudation.

Subsequent work (Green *et al.* 2011) has shown that the outcropping Uitenhage Group sedimentary rocks across the south-western part of South Africa were also heated to palaeotemperatures around 80°C or more prior to exhumation which began between 85 and 70 Ma. These results confirm that the region was buried by up to 2 km of Lower Cretaceous section prior to Late

Cretaceous exhumation. The presence of sedimentary remnants such as these Late Jurassic-Early Cretaceous extensional basins along the south coast of South Africa therefore provides a key geological constraint on the likely denudation history of the region. None of the published apatite fission- track studies of southern Africa have taken account of such constraints, and as a result these studies have failed to reveal the true nature of the development of the margin.

It is also worth pointing out that as illustrated in Fig. 59, the interpretation by Tinker *et al.* (2008a) that a thickness of *c.* 3 km or more of section was removed since the Late Cretaceous from the region below the Great Escarpment, where the boreholes are located, implies that unless massive faulting or flexure can be invoked, the Great Escarpment must also have been buried by around 2 km of section at that time. And if an Early Cretaceous phase of denudation affected this region as suggested by Tinker *et al.* (2008a), the present-day summit of the Great Escarpment must have been buried even more deeply at that time. This makes it difficult to envisage a scenario in which the Great Escarpment represents a fundamental control on post-rift denudation as embodied in the comparison of eroded thicknesses with offshore sedimentation by Tinker *et al.* (2008b). Instead, these observations suggest that the present-day Great Escarpment is the end result of the denudational history, rather than representing a fundamental control on the process of denudation. Its location today appears to owe much to the presence of resistant Karoo dolerite sills.

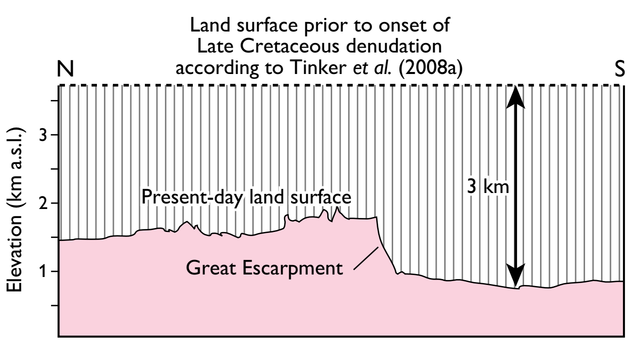


Fig. 59. Conceptual representation of the northern portion of the N–S profile in Fig. 58C, around the position of the Great Escarpment. Tinker *et al.* (2008a) interpreted their results from boreholes south of the escarpment (shown in Fig. 58) in terms of Late Cretaceous denudation, during which around 3 km of section was removed. This diagram emphasises that this interpretation requires that the present-day escarpment was buried by *c.* 2 km of section prior to the onset of Late Cretaceous exhumation. It is therefore clear that the present-day landscape is the product of the Late Cretaceous denudation history, and preserves no information on the pre-rift configuration (cf. Fig. 57).

6.2.6 Summary

Integrating geological constraints with published apatite fission-track data shows that the development of the modern topography characterising the southern margin of Africa was much closer to that of West Greenland outlined in chapter 5 than previously envisaged. Following break-up, the region subsided and was buried by up to 2 km of section, which was removed by uplift and denudation that began much later. The present-day topography is therefore not related to processes of continental rifting or break-up, but is instead the result of post-breakup processes, similar to the conclusion of Kempf (2010) in relation to the escarpment of Namibia (section 6.2.1).

6.3 South-east Australia

6.3.1 Thermochronology

The classic study by Moore *et al.* (1982) of apatite fission-track data from south-east Australia was one of the earliest applications of low-temperature thermochronology to a rifted margin, defining for the first time a now-familiar pattern of young apatite fission-track ages (around 100 Ma) along the coast, increasing inland to values around 300 Ma in the interior highlands. Similar patterns of apatite fission-track ages have since been identified in many other areas of the world, as reviewed e.g. by van der Beek *et al.* (1995), but as more data have become available it has become clear that in many areas young ages also occur hundreds of kilometres inland, as in Figs 57, 60.

Kohn *et al.* (2002) compiled available apatite fission-track data from south-east Australia (Fig. 60) and, after rejecting the possibility that elevated heat flow played any significant role in the heating revealed by the data, interpreted their results as reflecting only heating due to additional depth of burial and cooling due solely to denudation, with up to 3 km of section removed from coastal locations since 130 Ma. But even 100 km or more inland, Kohn *et al.* (2002) suggest that around 1–2 km of section has been removed since *c.* 130 Ma. This, together with apatite fission-track ages around 100 Ma or

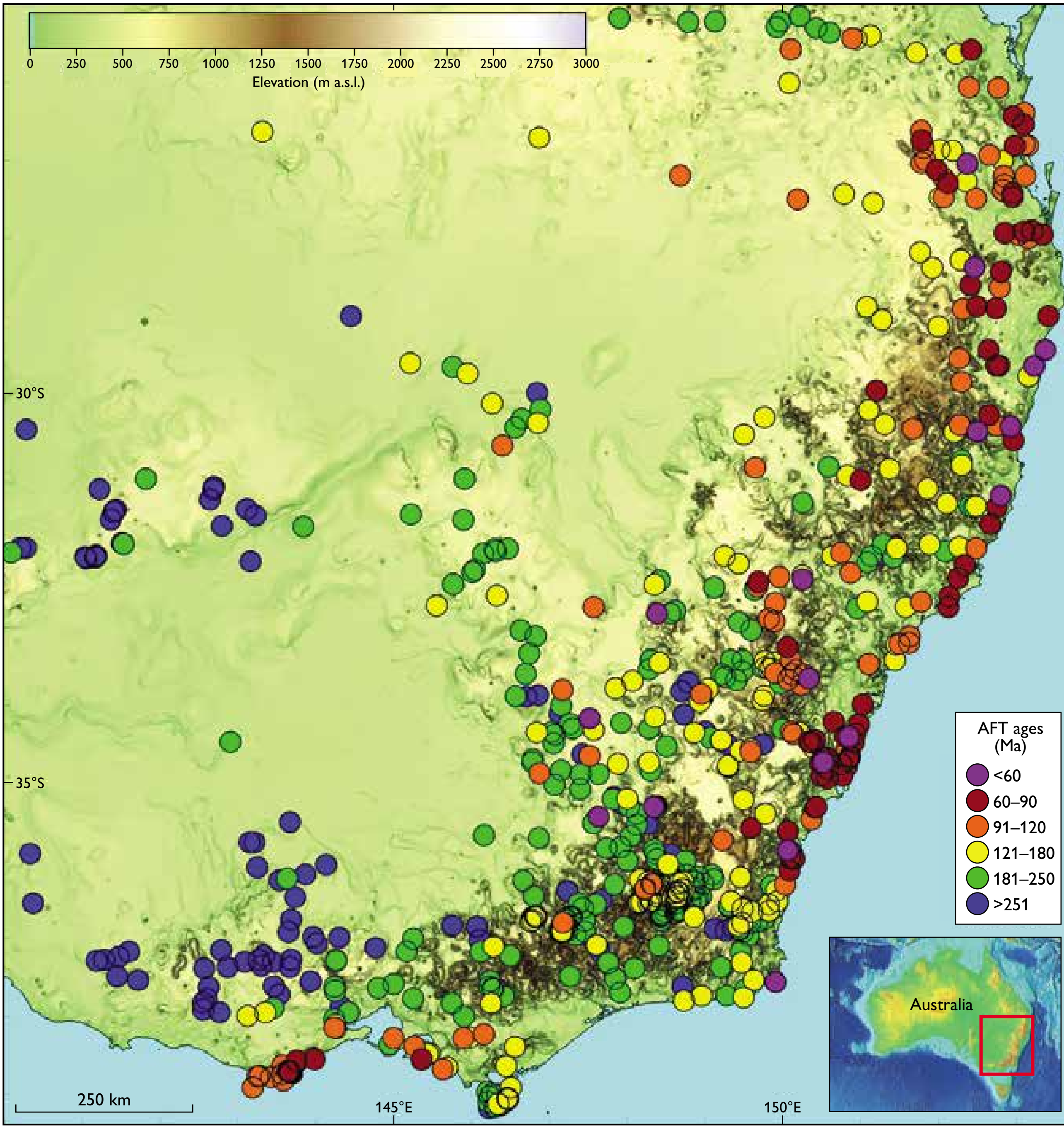


Fig. 60. Map of published apatite fission-track ages across south-east Australia (Kohn *et al.* 2002). Most of these results represent analyses of rocks of Devonian age or older. As in southern Africa, young ages are found hundreds of kilometres inland, and again bear no obvious relationship to the topography, including the location of the Great Escarpment.

less many hundreds of kilometres inland in Fig. 60, again emphasises the failure of the conceptual models in Fig. 57 to describe real data.

6.3.2 Landscape analysis

The landscapes of south-east Australia have been a subject of intense study over the last 100 years or more. In earlier morphological interpretations the high plains of south-east Australia (regions of low relief at elevations above 1 kilometre) were regarded as evidence of late Pliocene uplift following planation at sea level (Andrews 1910) or alternatively Oligo–Miocene uplift (King 1967). Craft (1933) also interpreted the incised valleys (corresponding to the Great Escarpment; Ollier 1982) as witness of Pliocene uplift. Over time, opinions shifted towards regarding the high plains as old, remnant features, and interpretations favouring recent uplift were replaced by the idea that the south-eastern Highlands represent the remnants of the Palaeozoic Lachlan Fold Belt (Lambeck & Stephenson 1986). More recently, opinion has once again shifted towards emphasising Cenozoic uplift (Holdgate *et al.* 2008), assisted by a growing body of evidence of late Cenozoic tectonism across south-east Australia (Dickinson *et al.* 2002; Sandiford 2003).

Bishop & Goldrick (2000) provide an extensive review of previous work, and comment on the particular interest of the eastern Australian margin in view of its 'general tectonic stability' and 'generally low rates of denudation'. A key aspect of geomorphological investigation in south-east Australia has been the quest to identify landscape elements of Mesozoic age (as reviewed by Bishop & Goldrick 2000; also see for example Hills 1975; Hill 1999), coupled with the interpretation of aspects of the present-day landscape as representing long-term stable axes (Persano *et al.* 2006). Another strong emphasis has been on recognition of the impact of rifting on the landscape, with the Great Escarpment (high margin with deeply incised valleys; Fig. 3) commonly interpreted as a direct consequence of the rifting process, as also thought by many to be the case in South Africa (e.g. Ollier 1982, 1985, 1995, Ollier & Marker 1985; Ollier & Pain 1997).

6.3.3 Attempts to reconcile thermochronology and landscape analysis

The notion of south-east Australia as tectonically stable does not fit comfortably with the interpretation of kilometre-scale denudation suggested by apatite fission-track data (above). As summarised succinctly by Bishop & Goldrick (2000, p. 246): "the various thermochronologic approaches ... suggest that the present land surface is ... the result of kilometre-scale denudation in the late Mesozoic, whereas some geomorphological interpretations indicate the preservation of Mesozoic landscape elements at the present land surface". On this basis, it is very difficult to reconcile these conflicting views whilst remaining faithful to both.

The apatite fission-track data summarised in Fig. 60 again provide no support for the conceptual models shown in Fig. 57, with ages as young as 120 Ma or less in Palaeozoic basement rocks hundreds of kilometres inland from the coast. In addition, apatite fission-track data (O'Sullivan *et al.* 1995) and VR and palaeomagnetic data (Middleton & Schmidt 1982), show that the Early Cretaceous palaeothermal signature along the south-eastern Australian margin shows no obvious correlation with the presence and location of the coastal escarpment. The interpretation by Middleton & Schmidt (1982) that their VR and magnetisation measurements indicated up to 3 km of additional section at coastal locations in the Permian–Triassic Sydney Basin reawakened an old debate. Branagan (1983) reviewed the history of debate regarding the 'vanished sequence' of the Sydney Basin, and concluded on the basis of evidence from preserved stratigraphy and geomorphology that less than 1 km of section could have been deposited and subsequently removed in post-Triassic time.

More recently, apatite fission-track data from south-east Australia have been the focus of further discussion and controversy (e.g. Nott & Purvis 1995; Leaman 2003; Osborne *et al.* 2006) without any satisfactory resolution being reached. Geomorphological studies tend to focus on continuous denudation, and interpret the absence of any sedimentary cover as indicating that none was ever deposited. On the other hand, thermochronological studies have historically tended to ignore constraints imposed by geology. A recent review from a geomorphological perspective of methods for dating land surfaces, drawing heavily on Australian examples (Watchman & Twidale 2002) fails to mention apatite fission-track methods, despite including a wide range of other methods from radiocarbon to Ar-Ar dating, perhaps illustrating a scepticism surrounding apatite fission-track-based interpretations in general.

As with south-west Africa, one aspect of the apatite fission-track interpretation that may help to bring these opposing views closer together is the possibility of significantly elevated heat flow associated with rifting along the south-east Australian margin, which would reduce the amount of removed section required to explain the apatite fission-track data. Middleton & Schmidt (1982) comment that VR data require higher palaeogeothermal gradients at coastal locations in Sydney Basin wells, and elevated Early Cretaceous palaeogeothermal gradients have been documented in the Otway Basin (Duddy

1994, 1997; Cooper 1995; Cooper & Hill 1997; Green *et al.* 2004), so it seems reasonable to suppose that this may also have been the case along the south-east Australian margin.

Nevertheless, as with the case of south-west Africa, for any reasonable value of palaeogeothermal gradient, kilometre-scale denudation is required to explain the apatite fission-track data, both along the coastal strip but also in the highland regions. To date, no convincing mechanisms have been proposed by which this can be achieved, and this has no doubt contributed to the difficulty in accepting this interpretation.

Persano *et al.* (2002, 2005) attempted to integrate thermochronological data more directly with evidence from geomorphology in south-east Australia. Persano *et al.* (2002) suggested, based on new apatite (U-Th)/He ages from the same region of south-east Australia originally studied by Moore *et al.* (1982), that the upland plateau region had undergone only limited denudation (<1 km of removed section) over the last 180 million years, compared to much greater denudation of the coastal plain region (several kilometres of removed section) since break-up in the Early to mid-Cretaceous. This contrasts markedly with previous interpretations derived from apatite fission-track data (above) that the upland region has also undergone kilometre-scale denudation since break-up. But as reviewed earlier (section 4.2), it has recently become clear that interpretation of apatite (U-Th)/He ages is not as straightforward as previously assumed. Thus, it is likely that the reason for this mismatch between the interpretation from apatite fission track and the conclusions derived by Persano *et al.* (2002, 2005) from their apatite (U-Th)/He ages probably arises because their apatites have retained more He than expected due to increased levels of radiation damage. Hence, in terms of the systematics used in their interpretation the apatite (U-Th)/He ages are 'anomalously old'. We should note that Persano *et al.* (2005) did consider the possibility of elevated heat flow along the margin, related to rifting, in contrast to previous studies.

In summary, despite large amounts of apatite fission-track data from south-east Australia and over 100 years of geomorphological investigation, no reconciliation has yet been achieved between the two approaches or within them. The nature of the processes responsible for the heating and cooling revealed by the apatite fission-track data remains the subject of debate, which has no doubt contributed to the lack of reconciliation with other lines of evidence. Firm constraints on palaeogeothermal gradients at the Early Cretaceous palaeothermal maximum, together with a precise chronology for the onset of cooling, would provide a major boost to the understanding of the underlying processes. To echo the comments of Bishop & Goldrick (2000), kilometre-scale denudation as required by apatite fission-track data from the upland region is very difficult to reconcile with the preservation of Mesozoic landscape elements, and true reconciliation of the two approaches will require multi-disciplinary investigations in which unambiguous independent constraints are available to constrain interpretations from both approaches, as in the West Greenland example discussed in chapter 5.

Fig. 61. Eocene muds overlying former river gravels and covered by Eocene basalt near the summit of Mount Hotham in the south-east Australian highlands. As discussed by Holdgate *et al.* (2008), the presence of these muds at an altitude of *c.* 1800 m above the present-day sea level is strong evidence of post-Eocene uplift of the present-day mountains.

6.3.4 Evidence for episodic burial and uplift on the south-east Australian margin

Interpretation of landscape elements as dating from the Mesozoic in previous landscape studies of south-east Australia is based chiefly on inference and unconstrained extrapolation of unconformity surfaces. Recently, Holdgate *et al.* (2008) drew renewed attention to the presence of Eocene muds and sands preserved below a cover of basalts (former valley fills) forming summits of high-level plains at almost 2 km a.s.l. in the south-east Australian highlands (Fig. 61). Holdgate *et al.* (2008) highlighted the lithological and other similarities between these Eocene sediments and deposits of similar age in the subsurface within the Gippsland and Murray basins, to the south and north of the present-day mountains. They suggested that these similarities indicated that the Eocene sediments were deposited within a low-relief landscape at or near sea level, which was uplifted and dissected in post-Eocene times. These observations suggest that the mountains of south-east Australia are relatively young (post-Eocene), favouring the earliest views on this subject, as reviewed in section 6.3.2. This scenario is strikingly similar to that outlined for West Greenland in chapter 5. The firm constraints on the timing of uplift and dissection provided by Holdgate *et al.* (2008) offer the promise of a more rigorous integration of information from thermochronology and landscape studies in the future.

While not relating directly to post-break up events, reset fission-track ages and high VR values in sedimentary rocks of Triassic age in the Sydney Basin (O'Sullivan *et al.* 1995; Middleton & Schmidt 1982) emphasise that heating to the palaeotemperatures required to explain the apatite fission-track and VR data must have involved additional burial by Upper Triassic to Lower Cretaceous sediments which have been subsequently eroded away. The amount of additional section required is uncertain because of the lack of rigorous constraints on palaeogeothermal gradients. But the precise amount of additional burial is not as important as the fact that burial is required prior to the onset of cooling (i.e. denudation). Thus, thermal histories involving monotonic cooling, as favoured by most thermochronological studies (as discussed earlier) are not relevant to this situation. This argument can be extended to coastal regions beyond the southern limit of the Sydney Basin sedimentary cover, where basement outcrops must also have been buried by a Permian to Lower Cretaceous sedimentary cover. This emphasises the importance of sedimentary outliers in helping to define the evolution of the margin, as illustrated elsewhere throughout this paper.

6.4 Brazil

6.4.1 Introduction

The Atlantic margin of Brazil is a classic area for studies focused on the origin of the extensive plateaux that characterise the country, even far from the coast (Fig. 62; e.g. King 1956a, 1967). Plateaux – or planaltos (elevated plain) in Portuguese – are key aspects of the Brazilian geography: the capital Brasilia was developed on such a planalto about 1200 m a.s.l., and today the address of the web site of the Brazilian government is 'planalto.gov.br'.

The geological record in north-east Brazil allows direct evaluation of the vertical motion of the margin following break-up of the South Atlantic at the Aptian–Albian transition (Torsvik *et al.* 2009). The highly fossiliferous Santana Formation – in particular famous for fossil fish – crops out on the flanks of the Araripe Plateau. It is of early Albian age and part of the post-rift succession of the extensional Araripe rift basin (Fig. 63; Assine 2007). The Santana Formation is represented largely by non-marine or quasi-marine strata, but according to Martill (2007), the occurrence of echinoids (sea urchins) in the Santana Formation, defining an unambiguously marine horizon, has been known and documented since 1966. Arai (2000) also reported additional indicators of mid-Cretaceous marine environment across the interior of Brazil. The presence of these marine, post-rift strata in the interior of Brazil at elevations up to 600 m a.s.l. (Morais Neto *et al.* 2006) testifies to both post-rift subsidence and subsequent significant uplift.

Furthermore, Morais Neto *et al.* (2006) presented AFTA data which document that post-rift sediment from the Araripe Plateau reached palaeotemperatures of 80–100°C in the Late Cretaceous, and thus that burial of the Araripe rift continued after the deposition of the Cenomanian deposits that are the youngest sediment preserved today (Assine 2007).

Despite this clear evidence of post-rift subsidence and much later uplift and erosion, scientific controversy abounds in the literature surrounding the antiquity of landscapes on the Brazilian EPCM. Many Brazilian as well as overseas authors emphasise that the formation of the present-day landscape took place millions of years after break-up (e.g. King 1956a, 1967; Almeida

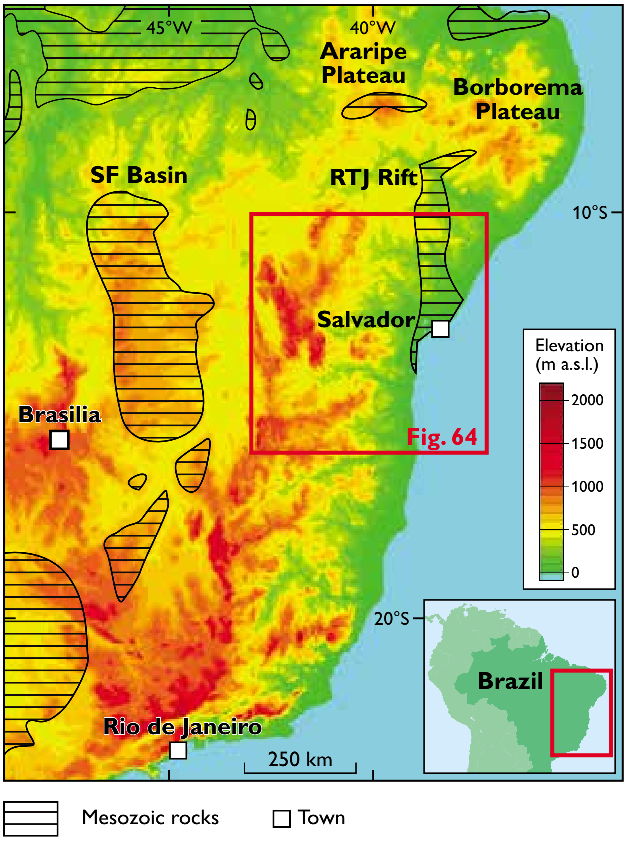


Fig. 62. Elevation and extent of Mesozoic rocks in eastern Brazil. Note the scattered occurrence of plateaux even far from the Atlantic margin. **SF**: São Francisco. **RTJ**: Recôncavo-Tucano-Jatobá.

and Carneiro 1998 (and references therein); Valadão 1998; Ab'Saber 2000; Cobbold *et al.* 2001; Riccomini *et al.* 2004; Zalán & Oliveira 2005). Other workers follow the most widely used approach at the present day, as described in section 2.4, in claiming that the dominant features in the landscape of Brazil date back to rifting and break-up of the Atlantic margin (e.g. Gilchrist & Summerfield 1990; Gallagher *et al.* 1994, 1995, 1998; Ollier & Pain 1997; Sacek *et al.* 2012). We seek to resolve the apparent conflict between these opposing views in the following discussion.

6.4.2 Studies focused on landscape evidence

Lester King (King 1956a, 1967) presented the most complete geomorphological analysis of eastern Brazil so far. King identified surfaces from a combination of fieldwork and early topographical maps and used geological evidence to place time constraints on the palaeosurfaces he mapped (e.g. King 1967, p. 233). He regarded the highest surface in the landscape as the oldest and the lower surfaces to be younger, and identified four denudation surfaces in eastern Brazil, which he interpreted as representing cyclic, low-relief, base level-governed erosional surfaces. He interpreted the age of the two oldest (and highest) surfaces in eastern Brazil to be of Mesozoic age, and he named them Gondwana and post-Gondwana. He considered the Gondwana Surface to be a pre-break-up surface that had developed across South America and Africa. In Brazil, the Gondwana and post-Gondwana Surfaces are only preserved in the highest areas in the interior; e.g. small remnants in Chapada Diamantina and in parts of Planalto da Conquista (see Fig. 64). More extensive, according to King (1967), is the Sul-Americana (South America) Surface that is well preserved, dominating the highlands of Chapada Diamantina as well as parts of the Planalto da Conquista. King attributed an early Cenozoic age to the Sul-Americana Surface because it cuts across silicreted sands of early Cenozoic age in locations west of the São Francisco River. King (1967) named a smooth surface with occasional inselbergs the Velhas Surface; e.g. on the flanks of Planalto da Conquista, which he regarded to be late Cenozoic. King (1956a, 1967) assigned large parts of coastal and near-coastal areas to belong to the youngest erosional cycle, the Paraguaçu, and he thought that this cycle was due to recent tilting along the present coastline.

Peulvast & Claudino-Sales (2004) identified two erosional levels of regional extent in north-east Brazil: A low plain between 0 and 300 m a.s.l. (the Sertaneja Surface or Sertão) and the discontinuous remains of a high plain between 750 and 1100 m a.s.l.; including the Araripe and Borborema plateaux. Peulvast *et al.* (2008) suggested that these plateaux are remnants of a continuous and low-lying, Late Cretaceous rift flank which was uplifted in post-Cenomanian time, following deposition of the youngest rocks now on the plateau. However, Valadão (1998) regarded the stepped landscapes in north-east Brazil to reflect mainly the Cenozoic development. Valadão (1998) thus found that the plateau surfaces in eastern Brazil coincide with the regional erosion surface that truncates Maastrichtian sediments in the Sanfranciscana Basin (Fig. 62; Campos & Dardenne 1997).

Bonow *et al.* (2009) and Japsen *et al.* (2012b) analysed landforms in north-east Brazil (10–15°S; Figs 64, 65) based on a digital elevation model from which they constructed a contour map (cf. Bonow 2004; Bonow *et al.* 2006a). The mapping was supported by profiles along

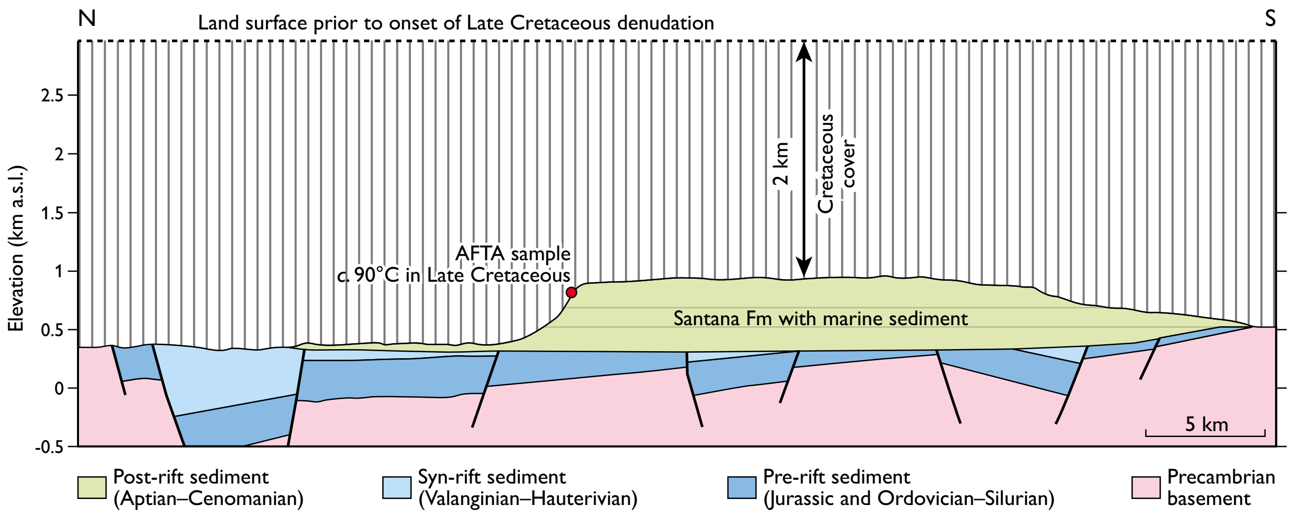


Fig. 63. Profile across the Araripe Plateau (see Fig. 62) with the Early Cretaceous rift section above Jurassic and Palaeozoic pre-rift sediments (Assine 2007). The post-rift Santana Formation of early Albian age contains unambiguous marine strata (e.g. echinoids) at an elevation of *c.* 600 m a.s.l. (Morais Neto *et al.* 2006; Assine 2007; Martill 2007). These observations testify (1) to post-rift subsidence and (2) to subsequent significant uplift. Thermal history interpretation of AFTA samples from the post-rift sequence of the Araripe Basin shows that this sediment reached palaeotemperatures of 80–100°C in the Late Cretaceous in close agreement with vitrinite reflectance values of 0.56% corresponding to a palaeotemperature of 93°C (Morais Neto *et al.* 2006). These observations show that burial of the Araripe rift continued after the deposition of the Cenomanian sediments that are the youngest preserved today. The thickness of the cover removed above the present-day surface of the plateau since the onset of Late Cretaceous denudation thus amounts to *c.* 2 km (1.8–2.5 km for a palaeogeothermal gradient of 30°C and a palaeosurface temperature of 25°C). Based on Morais Neto *et al.* (2006).

a square grid where topographical profiles were plotted along the grid lines together with maximum and minimum heights within a swath. In this way they identified two surfaces of low, relative relief and of regional extent, the Higher Surface (HS) and the Lower Surface (LS); cf. Figs 65, 66. There is also a coastal plain of limited extent, and above the level of the HS there are distinct hills along ridges of particularly resistant rock. The HS is preserved on high ground and includes the plateaux of Chapada Diamantina and Planalto da Conquista. These plateaux define a coherent surface, dipping slightly seawards. The HS cuts across Precambrian rocks that are frequently deeply weathered. The HS and the Cenozoic laterites (CPRM 2003) are preserved where the drainage system is unaffected by the fluvial system related to the younger LS at lower elevations. The LS is well defined across wide areas in the interior. The LS cuts across Precambrian basement in the west, the intracontinental Recôncavo-Tucano-Jatobá (RTJ) Rift, and Precambrian basement again in the east. It truncates the post-rift sediment of the Aptian Marizal Formation within the rift (Silva *et al.* 2007), but it also truncates the lower Miocene, marine Sabiá Formation within the Recôncavo Basin (Viana *et al.* 1971).

Bonow *et al.* (2009) found that both the HS and the LS are erosional features (peneplains) because they cut across rocks of different ages and resistance to erosion. The HS must originally have extended across a larger area than that of the present-day plateau remnants, as the LS has developed at its expense. The authors used geological constraints to conclude that the ages of the HS and LS are Palaeogene and Neogene, respectively, and that sea level is the most likely base level to which the surfaces were graded. The LS is rapidly being dissected by river incision into the sediment of the RTJ Rift due to a recent change in base level and thus producing a coastal plain.

A Palaeogene age for the HS across the highlands of Chapada Diamantina and Planalto da Conquista is consistent with the interpretation of King (1967) who mapped these plateaux as part of the Palaeogene Sul-Americana Surface. The highly dissected HS on the flanks of Planalto da Conquista corresponds to the Neogene Velhas Surface of King, whereas the LS and the coastal plain are equivalent to his Paraguaçu Surface within the study area. Further north, Peulvast *et al.* (2008) also identified two erosional levels of regional extent: a surface at low elevation and the discontinuous remains of a high plain (including the Araripe and

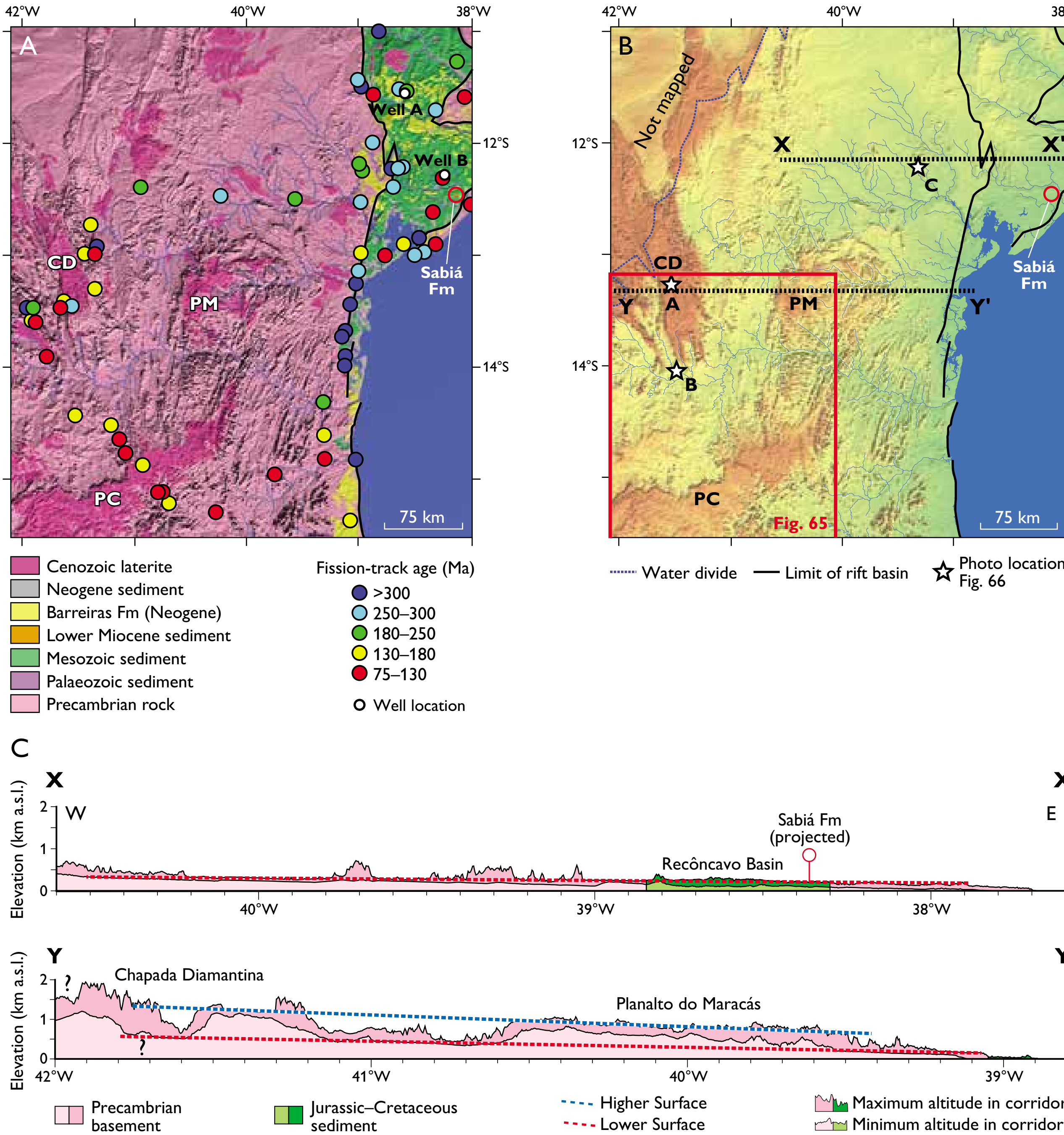


Fig. 64. **A**: Geology and **B**: Elevation of the area of north-east Brazil studied by Bonow *et al.* (2009) and Japsen *et al.* (2012b). Location shown in Fig. 62. Note that the presence of the Early Cretaceous rift only has a limited expression in the landscape, and that the plateaux coincide with extensive laterite covers. Based on geological maps of Brazil and of the state of Bahia (CPRM 2001, 2003). **C**: Two profiles to illustrate mapping of low-relief surfaces and interpreted peneplains, i.e., Higher Surface and Lower Surface (Bonow *et al.* 2009). Plateaux are part of one, seaward-dipping surface (profile YY'). Locations of profiles shown in **B**. Mapped, low-relief surfaces cut across basement (consisting of rocks that have different resistance to erosion) as well as the sedimentary sequence of the Recôncavo-Tucano-Jatobá Rift. Hence, surfaces are erosional features. The Higher Surface (HS) formed as a peneplain by fluvial erosion to near sea level during the Eocene–Oligocene. The HS was subsequently buried at the Oligo–Miocene transition and then uplifted and re-exposed in the Miocene. Uplift caused rivers to incise, after which the Lower Surface (LS) formed below the HS. **CD**: Chapada Diamantina. **PC**: Planalto da Conquista. **PM**: Planalto de Maracás. The location of the marine deposits of the Lower Miocene Sabiá Formation is indicated. Modified from Japsen *et al.* (2012b).

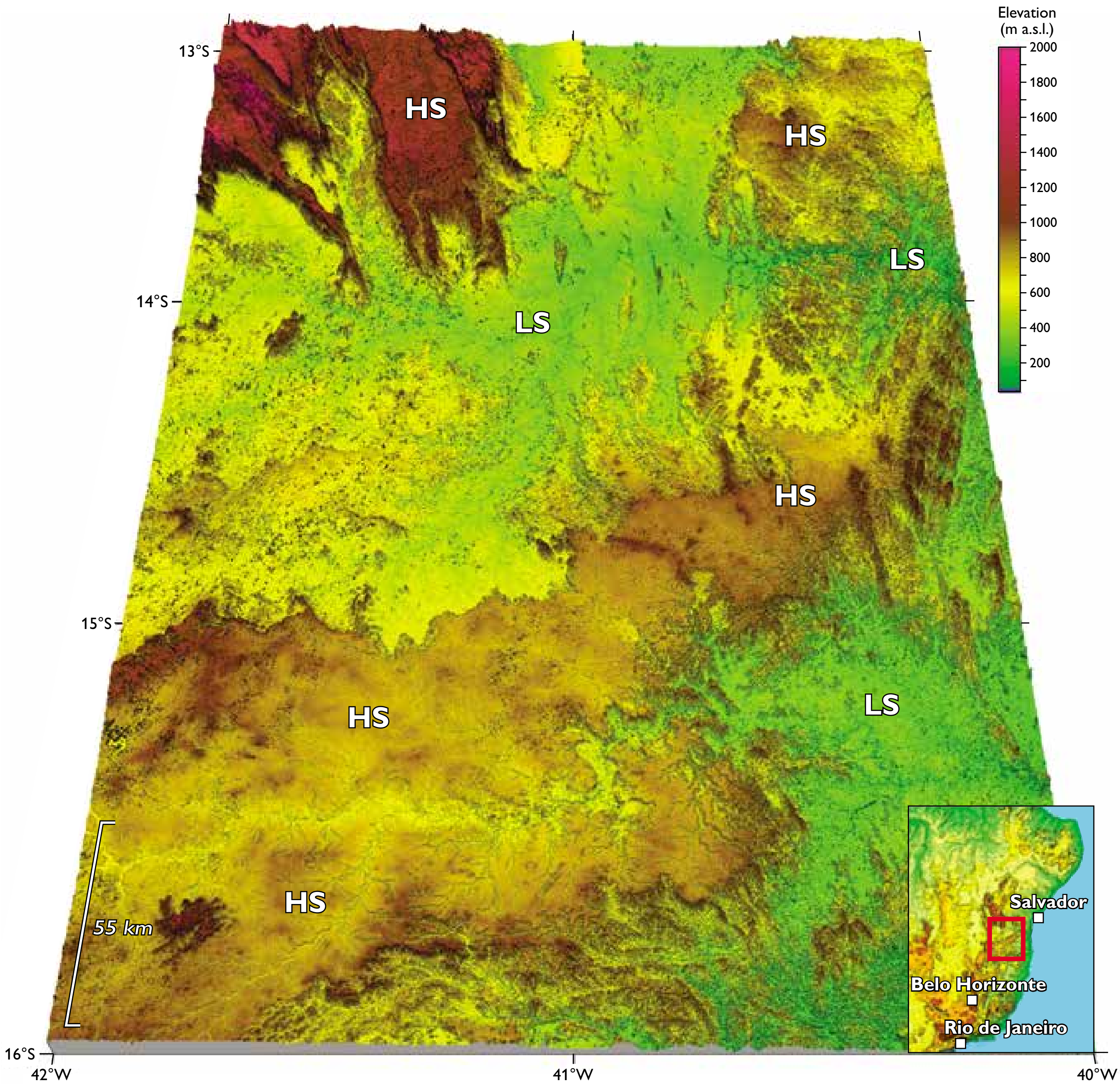


Fig. 65. Altitude between Chapada Diamantina in the north and Planalto da Conquista in the south with interpretation of two peneplains, the Higher Surface (**HS**) and the Lower Surface (**LS**). The LS is between 300 and 400 m a.s.l. and dips slightly eastward. Sharp, erosional, winding escarpment separates the LS from HS. The HS is particularly well preserved on Planalto da Conquista (*c.* 900 m a.s.l.) where the surface covers an area of *c.* 18,000 km² within the map frame; note the wide and shallow valley that trends towards east. The HS is also clearly defined on Chapada Diamantina (*c.* 1200 m a.s.l.). Location shown in Fig. 64. Modified from Japsen *et al.* (2012b).

Borborema plateaux). There are thus two regional peneplains in north-east Brazil, a lower surface and remnants of a higher surface that were formed by erosion during Neogene and Palaeogene, respectively (see also Japsen *et al.* 2012b).

In south-east Brazil, an extensive, low-relief erosion surface known as the Japi Surface, characterises much of the landscape at an elevation of 1200–1300 m a.s.l., but it locally reaches elevations of 2 km a.s.l. (e.g. Almeida & Carneiro 1998 (and references therein); Riccomini *et al.* 2004). According to Almeida & Carneiro (1998)

Fig. 66. Landscapes and interpretation of peneplains in north-east Brazil (Bonow *et al.* 2009; Japsen *et al.* 2012b). **A**: The Higher Surface (HS) in Chapada Diamantina, here at 1200 m a.s.l. with wide shallow valleys. **B**: The erosional Chapada Diamantina Escarpment between the HS and Lower Surface (LS). The LS is here at *c.* 700 m a.s.l. **C**: The LS with wide and shallow valleys north-west of Salvador. The surface is here at 250 m a.s.l. Photo locations shown in Fig. 64B.

the Japi Surface truncates a number of well-dated Upper Cretaceous intrusive complexes but not Maastrichtian complexes, leading these authors to conclude that the formation of the Japi Surface was completed before the end of the Late Cretaceous.

6.4.3 Studies focused on thermochronology

Gallagher *et al.* (1994, 1995) presented apatite fission-track analysis results from basement and sediment samples from the margin of south-east Brazil. Their fission-track ages broadly increase inland from between 60 and 90 Ma on the coastal plain to >300 Ma in the hinterland, and they concluded that the data reflected protracted denudation since the opening of the South Atlantic. They based their analysis on what they considered to be the empirical scarp retreat model of Gilchrist & Summerfield (1990; see Fig. 57), and concluded that their fission-track data were broadly consistent with that model. Gallagher *et al.* (1998) noted that the youngest apatite fission-track ages along the south-east Brazilian margin are significantly younger than the age of the rifting that they assumed led to the formation of the margin, and thus that substantial amounts of post-rift denudation had occurred. Franco-Magalhães *et al.* (2010) presented young apatite fission-track ages (66–6 Ma) from the south-east Brazilian margin that they interpreted as reflecting post-rift activation of the margin in Late Cretaceous, Eocene and Miocene times. However, they did not discuss how these very young ages relate to the truncated strata of Ordovician to Jurassic sediments and Lower Cretaceous flood basalts that are prominent in their study area. Nor did they discuss their results relative to the post-rift Japi Surface that characterises the planaltos of their study area (cf. Almeida & Carneiro 1998). Further, they did not discuss their apatite fission-track results in terms of timing and magnitude of discrete exhumation episodes and how this relates to formation of the Japi Surface relative to burial and exhumation of their study area.

Significant post-rift erosion has also been inferred from apatite fission-track studies of outcrop samples along the margin of north-east Brazil even at considerable distance from the coast (Harman *et al.* 1998; Turner *et al.* 2008; Morais Neto *et al.* 2008). Harman *et al.* (1998) identified two main phases of cooling, the first in the Early Cretaceous (*c.* 130 Ma) and the second in the Late Cretaceous (60 to 80 Ma). Harman *et al.* (1998) estimated a total cooling to the present-day along the continental margin of *c.* 100°C since 130 Ma, and a cooling in the interior of their study area since 80 Ma of 50–70°C. Turner *et al.* (2008) also identified two major cooling episodes which began in the intervals from 110 to 100 Ma (from palaeotemperatures higher than 75°C) and 40 to 10 Ma (from palaeotemperatures between 40 and 80°C). Morais-Neto *et al.* (2008) interpreted their data in terms of two dominant episodes of cooling that began in the intervals from 100 to 90 Ma (from palaeotemperatures higher than 80°C) and after 20 Ma (from palaeotemperatures between 45 and 85°C), but they also reported evidence for an intermediate event in the interval from 65 to 50 Ma. Despite the differences in the assessments of the timing between these studies, it is apparent that rocks that are now at the surface in north-east Brazil have cooled by 50°C or more since the mid to late Cretaceous. Combined with the presence of Cretaceous sedimentary units in the region, these results imply that kilometre-thick deposits once covered the rocks now exposed on present-day plateau surfaces, and that these deposits have been removed since the Cretaceous. Consequently the present plateau surfaces must be post-Cretaceous erosional features.

6.4.4 Integration of approaches

The studies described in section 6.4.3 were focused primarily on interpretation of thermochronology data, without taking much account of other constraints. In contrast, Cobbold *et al.* (2001), Cogné *et al.* (2011, 2012) and Japsen *et al.* (2012b) published studies of the post-breakup development of the Brazilian margin based on integration of thermochronological data and geological observations, and in the case of Japsen *et al.* (2012b) also with analysis of large-scale landforms. These studies lead to very different conclusions relative to the earlier thermochronological studies.

Cobbold *et al.* (2001) studied a variety of geophysical and geological observations as well as apatite fission-track data from the obliquely rifted margin of south-east Brazil (20–27°S), and documented Late Cretaceous, Eocene and Neogene reactivation of older structures, attributed to the combined effects of far-field stresses and hot-spot activity. In particular, Cobbold *et al.* (2001) found that the coastal mountains underwent block faulting and uplift in the Neogene.

Cogné *et al.* (2011, 2012) studied the elevated margin of south-east Brazil (22–24°S) where Upper Cretaceous/Palaeogene intrusions provide good evidence for post-breakup tectono-magmatic activity, and focused particularly on the Cenozoic onshore basins in the area; e.g. the Taubaté Basin (cf. Riccomini *et al.* 2004). They did not, however, include analysis of the landscape in their work, e.g. the extensive Japi Surface (Almeida & Carneiro 1998), and were thus unable to judge when the margin was close to sea level (compare Figs 67 and 68). Modelling of the thermochronological data together with geological observations led Cogné *et al.* (2011, 2012) to conclude that three periods of post-rift accelerated cooling had affected the margin; during the Late Cretaceous, the Palaeogene and the Neogene. The areas near the Cenozoic basins also experienced a period of burial prior to the Neogene; see Fig. 67. Cogné *et al.* (2012) summarised that the post-breakup evolution of south-east Brazil reflects a combination of structural inheritance, magmatic activity and plate-wide stress, leading to post-rift episodic uplift, rather than erosion of rift-related, uplifted relief.

Japsen *et al.* (2012b) reported the outcome of an integrated study of landscape development and thermo-tectonic evolution of the EPCM in north-east Brazil, focusing on the Early Cretaceous RTJ Rift and the extensive plateaus in the hinterland (Fig. 64; mainly 10–15°S; cf. Magnavita *et al.* 1994). Bonow *et al.* (2009) and Japsen *et al.* (2012b) analysed the landforms in the study area and identified two surfaces of low, relative relief and regional extent, referred to as the Higher Surface (HS) and the Lower Surface (LS) as discussed in section 6.4.2; cf. Figs 65, 66.

Japsen *et al.* (2012b) reported AFTA data from outcrop samples from north-east Brazil and from samples of sediment from boreholes down to 5.3 km below ground level in the RTJ Rift (Fig. 64), and defined nine regional cooling episodes from a synthesis of thermal history solutions derived from AFTA data in all samples (Table 4). The events date back to the Palaeozoic, but Japsen *et al.* (2012b) focused on the (post-rift) Cretaceous and Cenozoic development. The AFTA data define cooling episodes which began at *c.* 120 Ma and in the intervals from 80 to 75, 48 to 45 and 18 to 15 Ma (Aptian, Campanian, Eocene and Miocene cooling episodes). An Albian event (beginning between 110 and 105 Ma) is only recognised in samples east of the RTJ rift.

The AFTA and VR data from wells in the RTJ Rift show that the syn-rift sequences began to cool from their

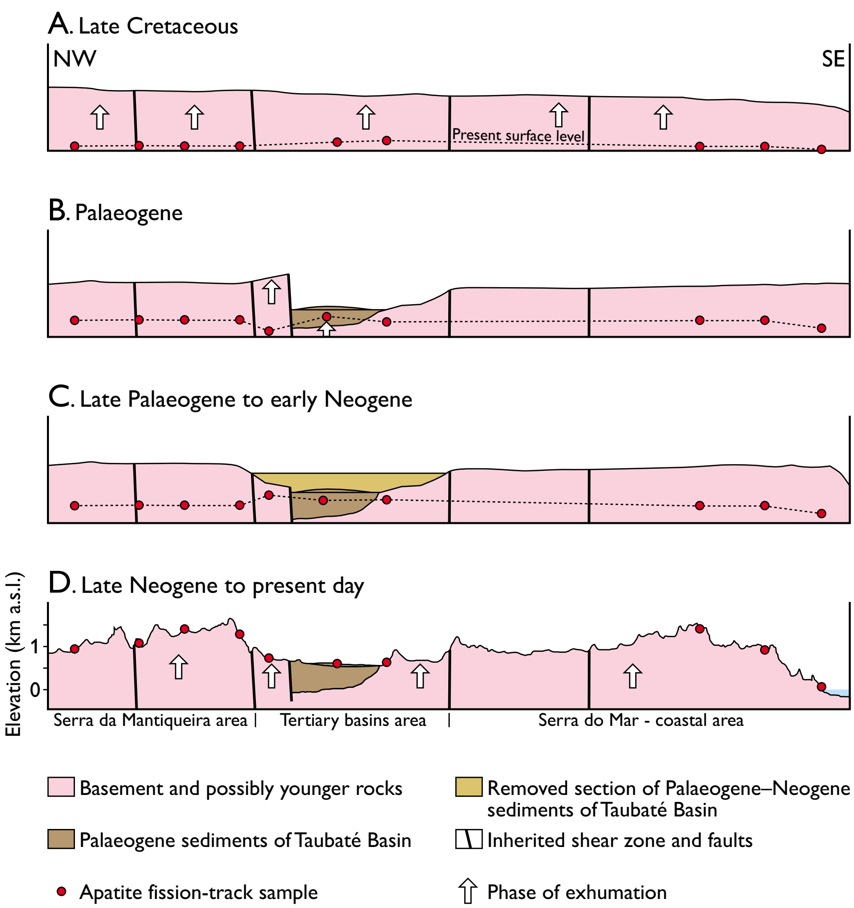


Fig. 67. The post-rift development of the south-east Brazilian margin (23°S) based on thermochronological data and thermal history modeling according to Cogné *et al.* (2012). **A**: In the Late Cretaceous, a considerable rock column covered the present-day surface prior to a first phase of post-rift uplift and erosion. **B**: During the early Palaeogene, a second phase of cooling/exhumation occurred only within the area of Cenozoic basins. **C**: During the late Palaeogene and early Neogene, the area around the Cenozoic basins was buried under sediments. **D**: From the late Neogene until the present day, a third phase of uplift and erosion affected the entire area leading to the formation of the present-day topography. Cogné *et al.* (2012) concluded that the high topography along the margin of south-east Brazil is the consequence of post-rift episodic uplift, rather than of erosion of rift-related, uplifted topography. Sacek *et al.* (2012) presented a different conclusion involving no post-rift activation as discussed in the text.

maximum post-depositional palaeotemperatures in the Campanian, followed by two further cooling phases in the Eocene and Miocene (Fig. 69; Japsen *et al.* 2012b). Analysis of the variation of palaeotemperature with depth in the wells shows that heat-flow conditions were close to those of the present day during all three episodes, implying that cooling in all three episodes was due to exhumation. Campanian palaeotemperatures were attributed to additional (post-rift) burial by *c.* 3 km of post-rift section, while Eocene and Miocene palaeotemperatures indicate burial by *c.* 2 km and *c.* 1 km of additional section, respectively.

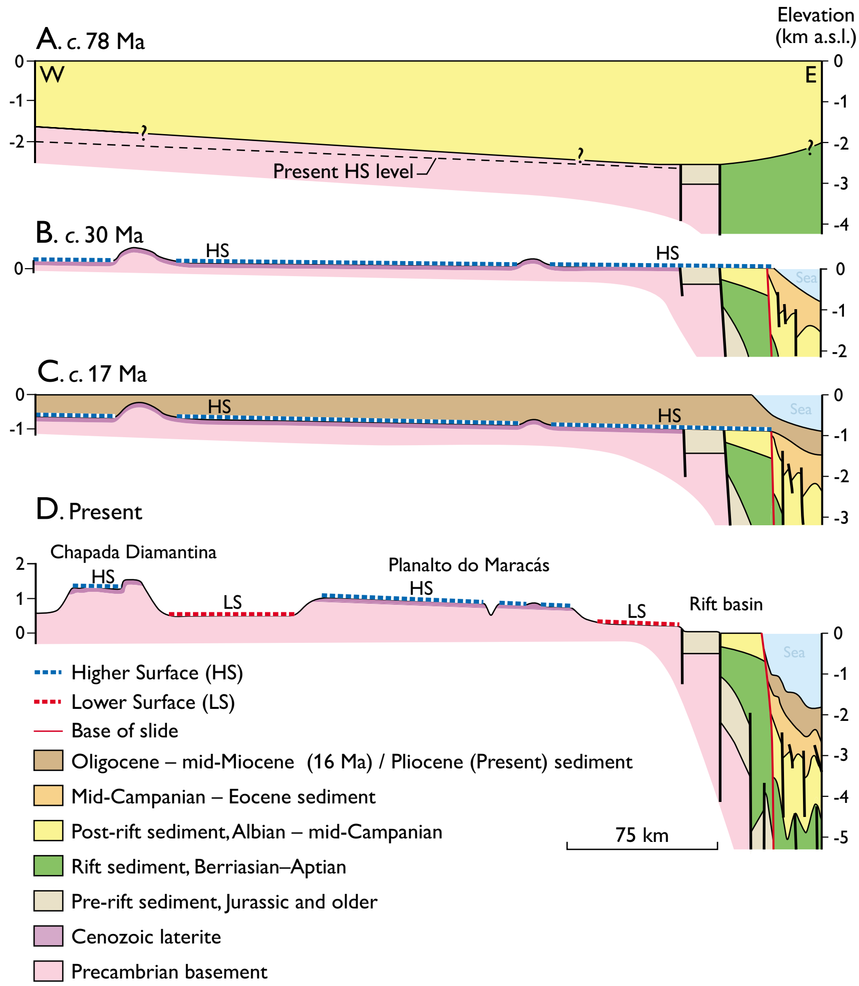
The conclusions of earlier studies described above (Harman *et al.* 1998; Turner *et al.* 2008; Morais Neto

Table 4. Intervals for onset of cooling episodes in north-east Brazil

Onset of cooling (Ma)	Stratigraphic interval	Areas/rocks with cooling episode identified
450–410	Ordovician–Devonian	*
320–300	Carboniferous	*
230–220	Late Triassic	*
180–170	Middle Jurassic	*
c. 120	Aptian	Most basement samples
110–105	Albian	Only samples in restricted area
80–75	Campanian	Sediment and basement samples
48–45	Eocene	Deep well samples
18–15	Miocene	Sediment and basement samples

Intervals defining the beginning of episodes of regional cooling based on AFTA data in 131 samples from outcrops and boreholes in north-east Brazil (Japsen *et al.* 2012b).
* Pre-Cretaceous episodes recognised only in restricted areas where effects of more recent events (particularly in the Aptian) are low enough to preserve evidence of earlier history. The early episodes likely affected much of the region.

Fig. 68. Burial and exhumation history along a profile in north-east Brazil across Chapada Diamantina and the Early Cretaceous rift basin, based on a topographic profile (B–B' in Fig. 64) and a geological cross section offshore (Menezes & da Silva Milhomen 2008). **A**: *c.* 78 Ma: Campanian maximum burial of the Lower Cretaceous synrift sequence below a Cretaceous cover that most likely extended over the basement terrains from the Atlantic margin into the Sanfranciscana Basin (Fig. 62). **B**: *c.* 30 Ma: Final formation of the Higher Surface (HS) by erosion to base level as a peneplain with deep weathering profiles and laterites after Campanian and Eocene phases of uplift and erosion. Major sliding offshore took place after the Campanian and Eocene uplift events (Cobbold *et al.* 2010). **C**: *c.* 17 Ma: Oligo–Miocene burial of the interior highlands and of the coastal zone. **D**: Present: Following 1) Miocene uplift which caused re-exposure of the HS and formation of the Lower Surface (LS) by river incision and valley widening with weathering, and 2) minor uplift in the Quaternary which led to incision below the lower surface and to formation of a coastal plain. Timing of events of cooling and exhumation from AFTA are listed in Table 4. Modified from Japsen *et al.* (2012b)



et al. 2008) are broadly compatible with the results of Japsen *et al.* (2012b). All studies emphasise the significance of regional cooling in the Late Cretaceous, even though earlier studies did not resolve the full complexity of the variation in the thermal history across the region. The greater detail obtained by Japsen *et al.* (2012b) was possible because they also analysed samples from deep wells in the rift. The three main cooling episodes identified by Japsen *et al.* (2012b) in north-east Brazil match the Late Cretaceous, Palaeogene and Neogene events identified by Cogné *et al.* (2012) in south-east Brazil.

Japsen *et al.* (2012b) presented a synthesis of geological data, stratigraphic landscape analysis and palaeothermal and palaeoburial data, defining a four-stage history of post-rift episodes of burial, uplift and exhumation that shaped the Atlantic margin of north-east Brazil (Fig. 68): (A) After Early Cretaceous break-up, the margin underwent burial beneath a thick sedimentary cover; (B) uplift episodes in the Campanian and Eocene led to almost complete removal of these deposits and formation of a large-scale, low/relief erosion surface (the Higher Surface, HS); (C) the HS was deeply weathered and finally reburied at the Oligocene–Miocene transition; and (D) Miocene uplift and erosion produced a new, lower-level peneplain (the Lower Surface, LS) by incision of the uplifted and re-exposed HS. Japsen *et al.* (2012b) noted that the uplift phases in Brazil were synchronous with uplift phases in Africa and the Andes (Fig. 70).

This four-stage model agrees with the results of previous studies which found that the RTJ rift was buried below a kilometre-thick cover prior to exhumation (Magnavita *et al.* 1994), that the plateau surface (HS) in north-east Brazil was fully developed by the end of the Palaeogene (King 1967; Valadão 1998), that the Oligo–Miocene transition was characterised by subsidence of the coastal regions throughout Brazil (Viana *et al.* 1971; Schobbenhaus & Brito Neves 2003; Rossetti *et al.* 2013), that plateaux elsewhere in north-east Brazil are covered

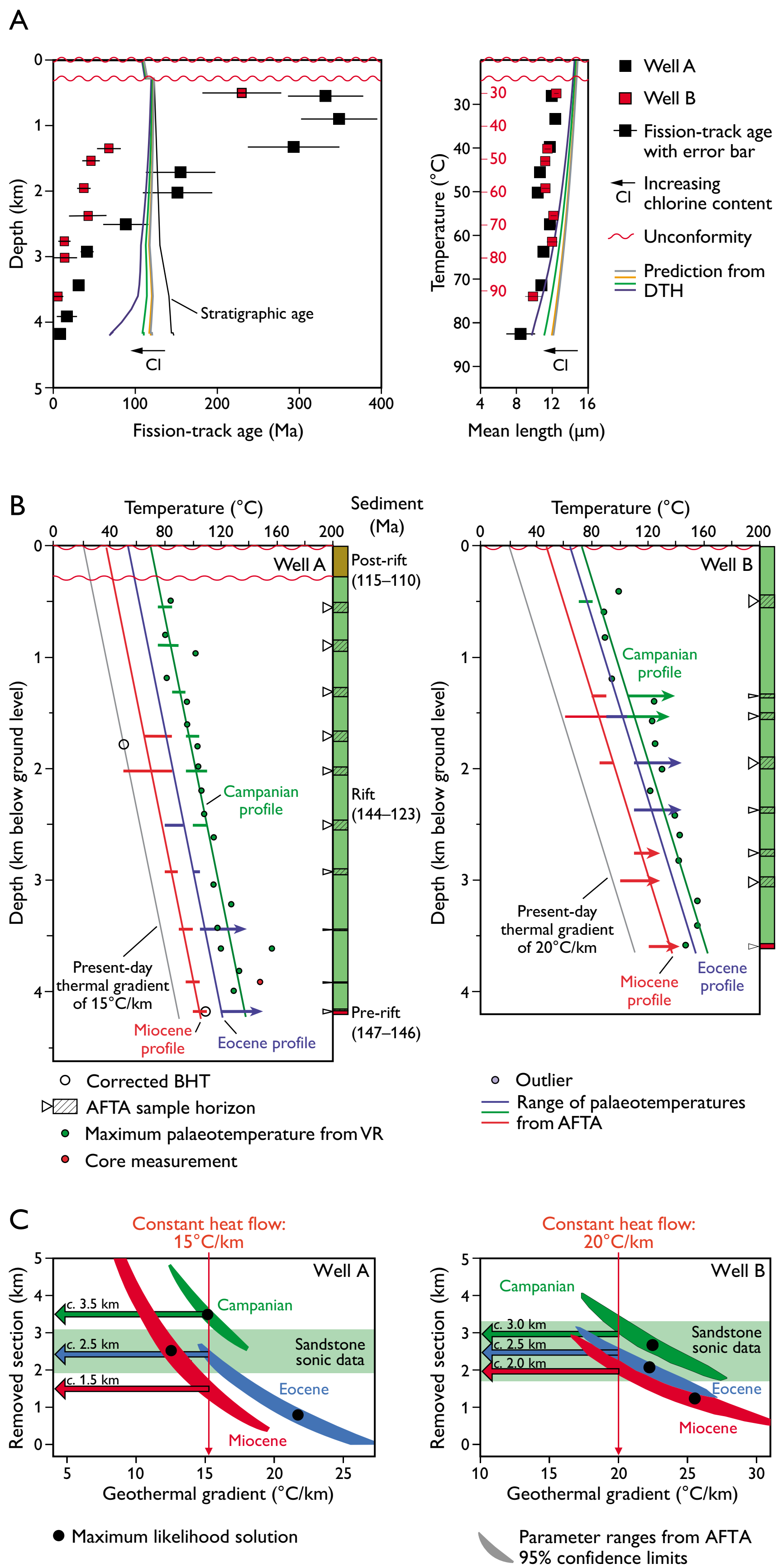


Fig. 69. Palaeothermal data from two wells defining burial and exhumation histories in the Recôncavo-Tucano-Jatobá Rift (location shown in Fig. 64A). **A**: Apatite fission-track analysis (AFTA) parameters for samples from wells A and B (drilled in the Tucano and Recôncavo Basins, respectively) plotted against depth and present-day temperature. The black line in the left panel shows the increasing stratigraphic age with depth. Coloured lines show the predicted patterns of fission-track age and mean track length for apatites containing <0.1, 0.5, 1.0, and 1.5 wt% Cl from the default thermal history (**DTH**). The default thermal history was derived from the preserved sedimentary section and the present-day thermal gradient calculated from corrected borehole temperatures (**BHT**) in each well. The fission-track ages decrease systematically with depth. At depths greater than 2 km, the ages are much younger than the values predicted from the default thermal history. This shows that the sampled units have been hotter in the past (Green *et al.* 2002). The pattern of decrease in fission-track age with depth is characteristic of a section that has undergone major cooling, with the 'break-in-slope' at a depth of *c.* 2.5 km representing the transition between partial and total annealing of fission tracks formed prior to the onset of cooling (Fig. 21). The corresponding fission-track age of *c.* 75 Ma represents the onset of exhumation. **B**: Palaeotemperature constraints vs. depth in wells A and B for the Campanian, Eocene, and Miocene palaeothermal episodes (Table 4). Drilled stratigraphy for each well is shown to the right. Constraints for each episode in both wells define linear profiles, sub-parallel to the present-day temperature profile, characteristic of heating predominantly due to deeper burial. **VR**: vitrinite reflectance. **C**: Ranges of amount of removed section and palaeogeothermal gradients (banana-shaped areas) required to explain palaeothermal profiles in wells A and B within 95% confidence limits. Limits on the amount of removed section were also estimated from sonic data from sandstone units (rectangular areas). Interpretations based on constant geothermal gradients corresponding to present-day conditions are also indicated, with palaeothermal and palaeoburial (based on sonic data) approaches giving highly consistent results in both wells (cf. Japsen *et al.* 2007a, 2012b). The present-day temperature profile for well B is based on corrected BHT data and temperatures revised on the basis of the AFTA data. Modified from Japsen *et al.* (2012b).

with non-fossiliferous, continental sediment (e.g. the Palaeogene Serra do Martins Formation on Borborema Plateau; Morais Neto *et al.* 2008) and that uplift affected much of Brazil in the Miocene (King 1967; Valadão 1998; Cobbold *et al.* 2001). Aspects that have remained unrecognised in previous studies, however, are the absolute timing of the episodes of uplift and the magnitudes of the burial and exhumation episodes that followed Early Cretaceous rifting and Eocene–Oligocene peneplanation.

6.4.5 Continuing controversy

While the various thermochronology studies discussed above led to differing conclusions in regard to the evolution of the continental margin of Brazil, they all consistently provide evidence of kilometre-scale denudation across a wide region. Where syn- or post-rift sedimentary remnants are preserved, this implies deposition of significant thicknesses of sedimentary cover that has been subsequently eroded. But some authors regard such interpretations as unrealistic.

Peulvast *et al.* (2008) discussed the apparent conflict in north-east Brazil, between the amount of cover that has been removed following Atlantic break-up based on landscape studies and amounts derived from apatite fission-track data. The idea that a substantial thickness of section could have been deposited and then eroded was regarded as unrealistic, despite evidence from apatite fission-track data (Harman *et al.* 1998; Morais Neto *et al.* 2006, 2008) showing that samples now at surface were at 70–85°C in the Late Cretaceous. Peulvast *et al.* (2008) found that the thickness of the eroded post-rift cover indicated by these studies by far exceeded their estimates based on the exposed strata, and preferred to dismiss the apatite fission-track results as either reflecting low-temperature artefacts in fission-track annealing models or effects of Cenozoic magmatism. But the results presented by Japsen *et al.* (2012b) document that the high Late Cretaceous palaeotemperatures revealed by apatite fission-track data from across north-east Brazil are not artefacts of the annealing models since they are confirmed by VR data and also by interpretation of sonic data from deep boreholes in the RTJ Rift (Fig. 69C). An explanation of post-rift palaeotemperatures in and around the RTJ Rift in terms of Cenozoic magmatism can be ruled out, since none has been reported from that area. In addition, the palaeothermal data from the deep boreholes in the RTJ Rift show that the palaeogeothermal gradient has been low (10–30°C/km) since the Campanian (Fig. 69C).

There are also diverging views between studies of the burial and exhumation history of south-east Brazil. As referred to above, Cogné *et al.* (2012) combined apatite fission-track data with geological observations to conclude that the high topography along the margin of south-east Brazil is the consequence of post-rift episodic uplift, rather than of erosion of rift-related uplifted topography. Furthermore, these authors showed that the basement rocks now exposed at the surface were buried below a kilometre-thick cover in the Late Cretaceous, and that they remained buried until the late Neogene outside the Taubaté Basin. In contrast, Sacek *et al.* (2012) used numerical modelling to investigate the same segment of the Brazilian margin as that studied by Cogné *et al.* (2012) but came to conflicting conclusions, as they found that the present-day morphology along that margin can be explained as resulting from rift-related vertical motions alone, without requiring significant post-rift "rejuvenation" (their quotation marks). Sacek *et al.* (2012) interpreted their modelling results to show that the combination of rapid erosion of the coastal (rift) escarpment and differential subsidence of the margin favours the emergence of a secondary bulge and its establishment as the main, 'permanent' drainage-divide escarpment, while the coastal escarpment progressively disappears. Sacek *et al.* (2012) did not consider the possibility that rocks now exposed away from an initial escarpment have been more deeply buried at any time since rifting and break-up.

Reconciliation of thermochronological data with the geological record and landscape evidence – particularly in terms of recognising the former presence of covers that are now removed – remains a matter of controversy in discussion of the development of the Brazilian margin. As a result, the extent, origin, age and importance of the Brazilian planaltos remain controversial in the tectonic analysis of the margin.

From the above review, it seems clear that the Brazilian plateaux and their present elevation are post-rift features, and that they were graded towards the general base level, which in the case of the post-rift development of north-east Brazil was determined by sea level of the adjacent Atlantic Ocean (e.g. King 1967; Almeida & Carneiro 1998; Valadão 1998; Zalán & Oliveira 2005). Yet many studies of the development of the Brazilian margin do not integrate the presence of these large-scale peneplains into their analysis (e.g. Gallagher *et al.* 1998; Franco-

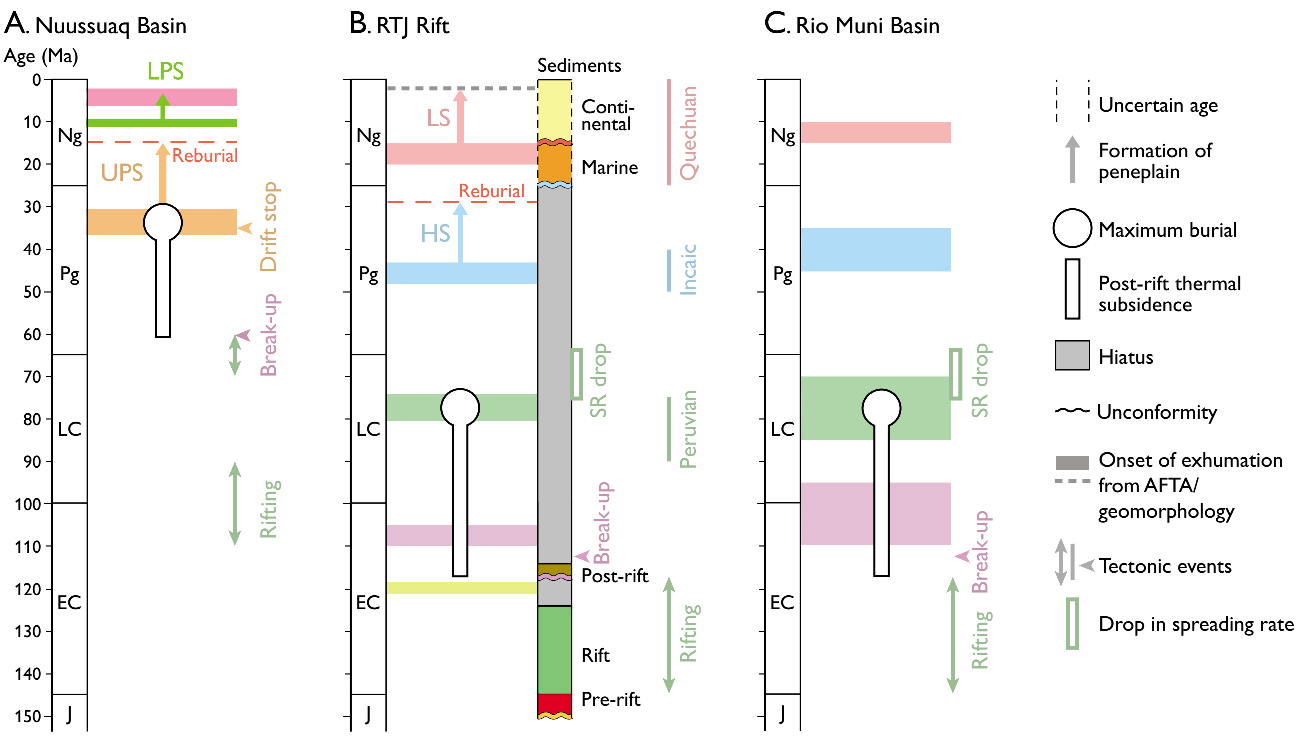


Fig. 70. Comparison of timing of post-rift uplift events, formation of peneplains, and tectonic episodes along margins in the Atlantic domain. **A**: Nuussuaq Basin (*c.* 70°N), West Greenland (Japsen *et al.* 2006). The Upper Planation Surface (**UPS**) defines the plateau whereas the Lower Planation Surface (**LPS**) is a system of palaeovalleys (Figs 34, 35; Bonow *et al.* 2006a, b). Events of subsidence and uplift along the West Greenland margin, prior to break-up are shown in Fig. 38. **B**: Recôncavo-Tucano-Jatobá Rift (**RTJ**; *c.* 12°S), north-east Brazil. The Higher Surface (**HS**) defines the plateau. The Lower Surface (**LS**) is an extensive low-relief surface that has developed at the expense of the HS (Figs 65, 66). Stratigraphy is after Viana *et al.* (1971) and Silva *et al.* (2007). Main phases of Andean orogeny: Peruvian, Incaic, and Quechuan (Pardo-Casas & Molnar 1987; Cobbold *et al.* 2001, 2007). **C**: Rio Muni Basin (*c.* 2°N), West Africa (Turner *et al.* 2008); conjugate margin to north-eastBrazil. The colours of the horizontal bands indicate onset of uplift events and the interpreted correlation between events on the conjugate margins in north-east Brazil and west Africa. The present high topography with plateaux in north-east Brazil and West Greenland were formed millions of years after break-up when regional peneplains were uplifted and dissected during the Neogene. Changes in plate motion after Chalmers & Pulvertaft (2001) and Torsvik *et al.* (2009). **AFTA**: apatite fission-track analysis. **EC**: Early Cretaceous. **J**: Jurassic. **LC**: Late Cretaceous. **Ng**: Neogene. **Pg**: Palaeogene. **SR**: spreading rate. Modified from Japsen *et al.* (2012b).

Magalhães *et al.* 2010; Cogné *et al.* 2012; Sacek *et al.* 2012).

In contrast, Almeida & Carneiro (1998) stressed that the dating of the extensive Japi Surface as a stratigraphic marker is of crucial importance for understanding the development of the Brazilian margin. This statement emphasises the importance of the accurate identification of such erosion surfaces. For example, can the Japi Surface at *c.* 1200 m be confidently correlated with the flat summits at *c.* 2 km a.s.l. as suggested by Almeida & Carneiro (1998), and is it the same surface that truncates pre-Maastrichtian but not Maastrichtian intrusive complexes? To our knowledge, no such mapping in Brazil has yet been made. Careful mapping of the extent of erosion surfaces such as the Japi Surface becomes a cornerstone for evaluating the relation between the surface and geological markers and hence for incorporating the presence of elevated peneplains in our understanding of the development of EPCMs.

6.4.6 Summary

In contrast to studies which assumed that the Atlantic margin of Brazil was uplifted at the time of rifting and remained as a positive region since, integration of thermochronology with observations of geology and landforms have shown that the margin subsided after break-up and was buried by up to 3 km of post-rift sediment

(Magnavita *et al.* 1994; Morais Neto *et al.* 2006, 2008; Japsen *et al.* 2012b). This cover was subsequently almost entirely removed in a series of episodes of uplift and erosion that first led to formation of an extensive erosional surface near sea level and then to its uplift and dissection, leaving the present-day plateau remnants. These studies thus document that the present-day high topography along the Brazilian EPCM formed long after break-up. This supports conclusions from early studies based purely on observations in the field (e.g. King 1956a; Almeida & Carneiro 1998). A significant difference is that the early studies did not consider the possibility of burial below post-rift sedimentary cover which was subsequently removed.

The evolution of the north-east Brazil EPCM defined above is similar in many respects to that defined for West Greenland in chapter 5, and as also inferred for southern Africa (section 6.2) and south-east Australia (section 6.3). We therefore suggest that this style of development, involving post-breakup subsidence and subsequent uplift and erosion in a series of episodes, describes the evolution of many EPCMs and may indeed provide a general description of their evolution.

6.5 Other areas

Comments similar to those in relation to southern Africa, south-east Australia and Brazil, discussed above, also apply to a number of other areas where apatite fission track and geomorphology have traditionally proved difficult to integrate. Nowhere is the inconsistency between the two approaches more pronounced than in northern Australia, where landforms regarded by some as having remained unaffected from early Palaeozoic times or even earlier have yielded much younger apatite fission-track ages, suggesting kilometre-scale denudation over the last 300 million years (Belton *et al.* 2004). Numerous geomorphological studies (as reviewed by Belton *et al.* 2004) have contributed to the idea that Australia preserves some of the world's oldest landscapes, reflecting the tectonic stability of the cratonic cores of the continents. However, extensive mapping of apatite fission-track data across the continent (Gleadow *et al.* 2002) has shown an absence of apatite fission-track ages in excess of 500 Ma, suggesting that the continent has experienced a much more tectonically active history than previously supposed. In the Yilgarn Block of Western Australia, Weber *et al.* (2005) reported fission-track ages generally between 200 and 300 Ma in samples from present-day outcrops, refuting the idea of the long term stability of this Archean craton. Indeed, the results presented by Weber *et al.* (2005) actually show that the supposedly stable shield was re-exposed from below a sedimentary cover and reburied during the Permian, as discussed in chapter 4.

The tendency to see topography as old is well illustrated by Persano & Dobson (2009) who sought to provide an answer to Bob Dylan's musical query about erosional timescales. They presented an interpretation of apatite fission-track and apatite (U-Th)/He data from two vertical profiles in north-west Scotland in terms of long-term, post-Caledonian, monotonic cooling reflecting denudation over 400 Ma. However, the presence of Late Devonian volcanics and sedimentary units near one of their transects and Triassic sandstones at outcrop over the wider region shows that these areas were close to the surface at various times in the post-Caledonian history. Holford *et al.* (2010) showed that taking these sedimentary remnants into account results in a much more active history involving multiple episodes of burial/deposition and uplift/exhumation than that suggested by Persano & Dobson (2009).

Northern England provides an example of a situation where apparently incompatible results from AFTA and landscape/geological investigations have been brought into a consistent regional framework. In initial interpretations of AFTA data from the area (Green 1986, Lewis *et al.* 1992), samples which cooled from greater than 110°C in the Paleocene (*c.* 60 Ma) were interpreted in terms of burial by >3 km of section that was removed during Cenozoic denudation. Such thicknesses of former cover were regarded by many as unrealistically high (e.g. Holliday 1993). However, later, more detailed studies involving AFTA data in samples over vertical sections from both wells and outcrop sections revealed that palaeogeothermal gradients were seriously underestimated in previous studies, and use of more appropriate values around 50 to 60°C/km as defined from measured AFTA and VR data suggested much lower amounts of eroded section (up to *c.* 1.5 km), allowing the two approaches to be readily reconciled (Green 2002).

While the focus of discussion in this chapter has been to highlight inconsistencies in the interpretation of landscape and thermochronology studies, it should also be noted that some examples exist where the two approaches have shown a much greater degree of consistency, including studies of south-east Asia (Schoenbohm *et al.* 2004) and Corsica (Kuhlemann *et al.* 2005). The key to successful integration of the different approaches in these areas lies in data from both approaches being given equal weight, with the evidence inferred from landscape studies used to provide prime constraints on the evolution, similar to the studies in West Greenland and Brazil discussed here.

We suggest that an approach similar to that outlined in chapter 5, involving detailed thermochronology analyses in carefully collected samples, with data interpreted within a framework constrained by geological observations and stratigraphic landscape analysis, will ultimately lead to reconciliation of results from different approaches in southern Africa, south-east Australia and elsewhere, and to an improved understanding of the development of EPCMs.

Similar comments apply to cratonic regions. As shown in chapter 3, re-exposed peneplains with characteristic relief (the flat sub-Cambrian and the hilly sub-Mesozoic peneplains) in basement rocks are of major importance in the present landscape of south Sweden and this despite several Quaternary glaciations. Such re-exposed forms also make up much of the relief farther north (Lidmar-Bergström 1995, 1996; Lidmar-Bergström *et al.* 2013). Similar observations are also made on the Canadian shield with a flat sub-Ordovician surface (Ambrose 1964) and a hilly sub-Mesozoic surface (Lidmar-Bergström & Jansson 2005). However, such observations are seldom acknowledged in recent studies dealing with long-term erosion. Integration of stratigraphic landscape analysis and thermochronology in regions such as these will provide an improved understanding of subsidence and uplift of shield areas, previously widely regarded as stable (cf. the north Australian and Western Australian examples reviewed above).

6.6 Numerical modelling of EPCM development

In recent years, considerable effort has been devoted to development of numerical models describing the geomorphological development of rifted margins (e.g. van der Beek *et al.* 1994, 1995; Braun & van der Beek 2004; Sacek *et al.* 2012). These models commonly seek to explain the form of EPCMs as a result of erosion acting on already elevated rift flanks formed at high elevations that were already extant at the time of break-up, within a framework of progressive denudation. They further assume that denudation is focused seaward of the coastal escarpment, with high-level plains being much less affected, in similar fashion to the conceptual models illustrated in Fig. 57. In this way, the need for a tectonic component of uplift associated with (or following) rifting and break-up is eliminated (although the high elevation remains unexplained), and the morphology is then explained purely in terms of erosional processes (moderated by changes in climate) and the corresponding isostatic response, within a framework of progressive emergence and monotonic cooling. An extension of these studies is the three-dimensional, thermal-kinetic modelling package PECUBE, the current status of which, together with a large number of applications, is reviewed by Braun *et al.* (2012). This allows prediction of vertical and horizontal patterns of ages from various thermochronological methods, but is focussed towards upward crustal movements involving progressive denudation of basement regions.

As discussed above, evidence from southern Africa, south-east Australia, Brazil and other regions, in addition to West Greenland (chapter 5), defines a different style of behaviour, revealing major post-breakup burial and subsequent exhumation, extending hundreds of kilometres inland from the escarpment, often in multiple episodes. So the relevance of the numerical models described above to real world situations is far from clear. Such models include no tectonic input to the development of the margin. However, in regions such as south-east Australia, southern Africa, Brazil and West Greenland, where additional burial is required to produce the palaeotemperatures prior to the onset of cooling defined from thermochronological data, it is also necessary to explain the transition from subsidence/burial to uplift/exhumation. In such circumstances, a tectonic input would appear to be mandatory.

In summary, models of EPCM development based on continuous denudation do not provide an accurate description of the processes involved. New approaches are required which include post-breakup subsidence and burial, in addition to subsequent denudation and uplift, in order to provide an accurate description.

6.7 Concluding remarks

In those situations discussed above where conclusions derived from thermochronology conflict with those from landscape analysis, often little attempt has been made to integrate both approaches with equal weighting. More typically, proponents from each side have looked at the region with a mindset based on their own experiences and ideas, and failed to take into account what appears to be conflicting evidence from another point of view. Common examples include the adoption of constant heat flow in thermochronological studies despite abundant evidence that this is untenable, failure to acknowledge the information provided for example by erosion surfaces (peneplains) and valley generations in landscape studies and their relevance to the interpretation of thermochronology data, and attempts to 'shoe-horn' data into compliance with simple conceptual models. The omission of any discussion of apatite fission-track data in many landscape studies, as discussed earlier, amounts to effectively discounting the relevance of the apatite fission-track approach. We suggest that if observations are based on reliable science, then a satisfactory reconciliation must be possible, and that all that is required to achieve this is an attitude that is receptive to novel outcomes, rather than demanding that results conform to expectation and accepted models.

The next chapter discusses various mechanisms which might explain the episodic development of EPCMs described in the preceding chapters.

7. What processes drive the formation of EPCMs?

Plate-tectonic theory accounts well for mountain ranges on continents formed at convergent or collisional plate margins, whether they are due to subduction of an oceanic plate below a continental plate or to the collision of two continents. Plate-tectonic theory does not, however, account well for the mountain ranges discussed in this paper; those reaching 1–2 km a.s.l. found inland from many passive (extensional) continental margins (e.g. Japsen & Chalmers 2000; Japsen *et al.* 2012a). Such EPCMs have been recognized for many years (chapter 1), but many explanations in the literature for their presence are unconvincing. The special properties of only one EPCM are commonly used to try to account for its existence, despite their common aspects (e.g. asymmetrical topography and high plains), while others make assumptions about properties of EPCMs that are not in accordance with observations. We review here some of the more popular hypotheses.

7.1 Hypotheses that apply to one or only a few EPCMs

Some hypotheses concerning the formation of individual EPCMs rely on some particular property that is not shared by others. For example, much of the Norwegian EPCM coincides with the Scandinavian part of the Caledonian orogenic belt. Mountain ranges formed by continental collision develop deep roots that hold their topography uplifted by isostasy (e.g. Watts 2001, pp. 339–361). Nielsen *et al.* (2009) assumed that the Norwegian mountains are merely the eroded remnants of former Caledonian topography and that they are held uplifted by a remnant crustal root. This hypothesis has now been tested and disproved (Stratford & Thybo 2011; Ebbing *et al.* 2012; Maupin *et al.* 2013). No root is present beneath the Northern Scandes (Ebbing *et al.* 2012) and, while there appears to be a small root under the Southern Scandes (Stratford & Thybo 2011), it is not sufficient to support the main mountain range isostatically and is displaced from it. There is also good evidence that the Scandinavian Caledonides collapsed in the Devonian (e.g. Dewey *et al.* 1993) while the branch of the Caledonian Orogeny under northern Poland and Germany is buried beneath Palaeozoic and younger sediments. See Lidmar-Bergström & Bonow (2009) and Chalmers *et al.* (2010) for a more comprehensive discussion of Nielsen *et al.* (2009).

Similarly, Pedersen *et al.* (2012) suggested that the high topography in East Greenland is an erosional remnant of topography originally developed during the Caledonian orogeny. Pedersen *et al.* (2012) used apatite fission-track data in an attempt to demonstrate a monotonic cooling history for that region, representing continual uplift and denudation, since the end of the Caledonian orogeny. However, as discussed by Japsen *et al.* (2013a) the geological record of East Greenland shows that the Caledonian mountains there had collapsed by the late Palaeozoic (see section 8.2) and the region was, at least partially, buried below deltaic and marine Carboniferous sediments. Surlyk *et al.* (1984) showed that a peneplain had formed at sea level over much of northern East Greenland by the mid-Permian, and the area was again reburied under Mesozoic sediments. Haller (1971, p. 321) summarised the demise of the Caledonian fold belt in the following way: "In Central East Greenland, which seems to have been the heart of the ancient mountain belt, the vigorous late movements lasted into the Permian. The molasse-laden Caledonian mountain stumps then became completely bevelled and during the late Permian the sea advanced from the east. This transgression marked the end of the Caledonian orogeny. Hence all subsequent events are to be included in the post-Caledonian history".

A similar style of long-term evolution was suggested for the highlands of south-east Australia by Lambeck & Stephenson (1986) who suggested the present-day mountains were a residual of the Palaeozoic Lachlan Fold Belt, with the topography maintained since *c.* 250 Ma by the isostatic response to erosional unloading. This model is inconsistent with thermochronological evidence of major early Cretaceous denudation across the region as reviewed in section 6.2, and also fails to explain the presence of Eocene, low-level, fluvio-lacustrine muds at *c.* 1.8 kilometres above sea level, while similar muds also occur in adjacent basins at depths of up to 2 km below sea level (Holdgate *et al.* 2008), as discussed in section 6.3.4.

Another hypothesis that applies to only a few EPCMs is based on the observation that some of them coincide with, or are at least near, large igneous provinces (LIPs). It appears to be the case that, while mantle upwelling (plume emplacement) immediately prior to emplacement of a LIP causes uplift (e.g. Dam *et al.* 1998; Saunders *et al.* 2007), there is no evidence that such uplift persists after cessation of volcanism, as has been suggested by some authors (e.g. Mackay *et al.* 2005). Subsidence during and after volcanism greater than the immediately preceding uplift has, however, been demonstrated in some LIPs such as West Greenland (section 5; Japsen *et al.* 2005) and East Greenland (Brooks 2011; Bonow *et al.* 2014, in press). In addition, while some EPCMs contain LIPs, others do not (e.g. south-east Australia and Norway), and the West Greenland and Baffin Island EPCMs run continuously from a LIP to a non-volcanic margin (Chalmers, 1997; 2012).

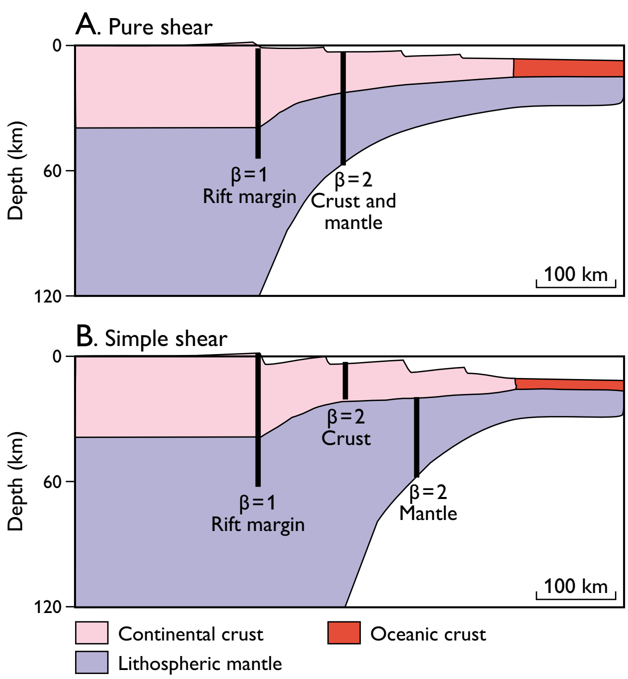


Fig. 71. Cross-sections showing the difference between **A**: pure shear extension and **B**: simple shear extension. In **A** the amount of extension of the upper mantle and crust is the same at the same place. In **B**, extension of the upper mantle takes place much farther basinward than extension of the crust, shown by β = 2 being at different locations in the crust and mantle. Post-rift cooling causes subsidence of the basin margin in **A** but not in **B**, so any flexural uplift of the margin produced during rifting will also subside in **A** but not in **B**. Extensive modelling of subsidence of basins with low beta-factors (<2) by the hydrocarbon industry has shown that McKenzie's (1978) pure shear model of rifting is a good first approximation for the development of these basins, although extension at high beta factors appears to take place as in **B**.

7.2 Permanently uplifted rift margins?

As discussed in chapter 2, much of the literature (e.g. Gallagher *et al.* 1998; Brown *et al.* 2002; Persano *et al.* 2006; Swift *et al.* 2008; Sacek *et al.* 2012) assumes that EPCMs are erosional modifications of rift margins that were uplifted at the time of rifting/onset of sea-floor spreading and have remained uplifted since. Some attempts to account for EPCMs as geodynamic features have also assumed that they have been present since the time of rifting or break-up (e.g. Weissel & Karner 1989; Gilchrist & Summerfield 1990; Chéry *et al.* 1992; ten Brink & Stern 1992; Watts 2001). However, as Weissel & Karner (1989) showed and Japsen *et al.* (2012a) discussed, rift margins will remain uplifted indefinitely only if there is little or no extension of the mantle below the rift margin at the time of rifting (Fig. 71); i.e. when the extension approximates simple shear.

Understanding of rifts and rift margins has progressed enormously in recent decades. Early debates on rift formation concentrated on two end-member models; pure shear (e.g. McKenzie 1978) and simple shear (e.g. Wernicke 1985). Understanding of tectonic controls on sedimentation within a rift has also developed from the early days of sequence stratigraphy (e.g. Payton 1977), when it was assumed that the relative changes in sea level, necessary to generate the third and higher-order unconformities that form sequence boundaries, are eustatic (e.g. Haq *et al.* 1987 and references therein).

Studies since then have shown that simple shear and pure shear are probably end members of a continuum of extensional styles, and that extensional style and amount can vary not only between the crust and mantle (Cloetingh & Ziegler 2009), but at different levels within the crust (e.g. depth-dependent stretching; Davis & Kusznir 2004).

While it is undoubtedly the case that a number of factors, such as variations in the total net amount of oceanic crust and variations in the terrestrial ice budget, can cause eustatic sea level to vary, it has also become apparent that modest stress fluctuations on a regional scale can lead to relative sea-level changes of the same order (100–200 m) and on time-scale characteristics (*c.* 3 million years) for third order cycles (Cloetingh *et al.* 1985).

Evidence has also been accumulating steadily that, in addition to the modest relative changes in sea level recorded by third and higher-order cycles, rifted basins and margins have been subjected to breaks in subsidence and sedimentation that last 10 million years and more (e.g.

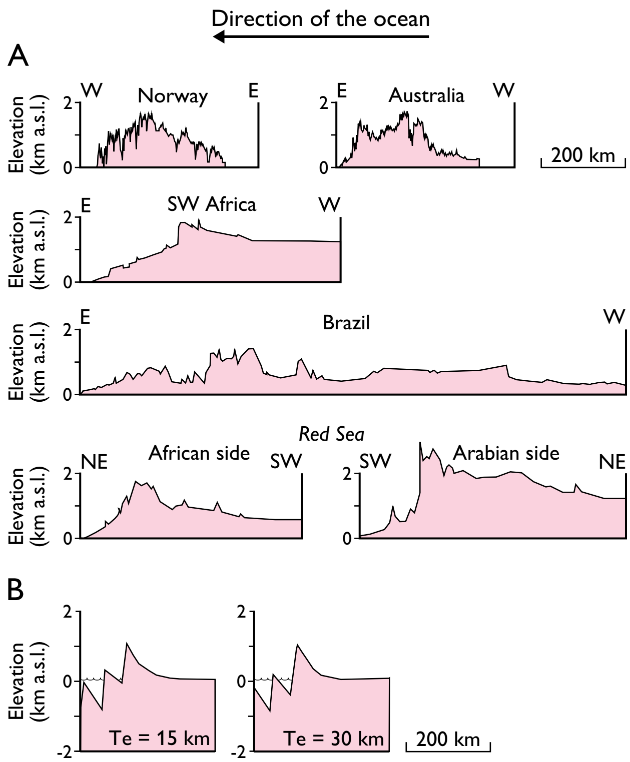


Fig. 72. Comparison of cross-sections of elevated, passive continental margins (EPCMs) with the expected amount of flexural uplift of a rift margin. **A**: Cross-sections of five different EPCMs (all at the same horizontal and vertical scale). The sections are oriented so that the nearest ocean is to the left and the continental hinterland to the right. **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. Modified after Japsen *et al.* (2012a).

Ziegler & Cloetingh 2004; Cloetingh & Ziegler 2009). The unconformities that result from these events are second order and unconformities of this type in the southern West Greenland Basins have already been described in section 5.6.

Extensive modelling of subsidence of marginal basins by the hydrocarbon industry has shown that McKenzie's (1978) pure shear model of post-rift subsidence approximates the actual depositional record quite well in basins where extension factors are less than about 2, except where the sedimentary record is interrupted by second order unconformities. The subsidence curves predicted by the McKenzie (1978) pure shear model are offset across these unconformities (see e.g. fig. 10 in Cloetingh & Ziegler 2009). The offsets are not what would be expected if sedimentation merely ceased, but are more consistent with the erosion of a substantial thickness of sediment prior to renewed subsidence.

If a rift forms by processes approximating simple shear, the lithospheric mantle underlying the proximal part of the rift (at extension factors less than about 2) extends less than the overlying crust (Fig. 71B). The finite strength of continental crust causes the area just outside the rift to rise because of its flexural response (see e.g. Watts 2001, pp. 286–339). The proximal part of the rift basin will not subside thermally after rifting ceases because the underlying lithospheric mantle is not warmer than it was pre-rifting, so the uplifted rift margin will not subside either, but will only be lowered by erosion. On the other hand, the lithospheric mantle underlying a rifted margin that extends by processes approximating pure shear extends by the same amount as the overlying crust. In this case the rift will subside because the thinned and therefore heated underlying lithospheric mantle cools and subsides, causing the overlying proximal rift to subside too (McKenzie 1978). As in the case of simple shear, the finite strength of continental crust causes the margin just outside the rift to rise because of its flexural response, but it will subside again as the rift subsides because of the flexural response to both the finite strength of the thermally subsiding rift and to loading by sediments deposited in the basin. These effects, together with erosion of the margin while uplifted, will cause the margin outside the rift to subside below sea level and sediments will transgress across the rift margin forming a so-called 'steer's-head' geometry in cross-section (Watts *et al.* 1982; White & McKenzie 1988; Braun & Beaumont 1989; Roberts & Yielding 1991).

The width of uplifted area caused by flexural uplift of a rift margin is also generally much smaller than the observed widths of EPCMs. Figure 72 shows the calculated widths of uplifts caused by flexural uplift using elastic thicknesses of 15 km and 30 km compared with cross-sections through actual EPCMs. It is readily apparent that the actual EPCMs are all much wider than the calculated flexural uplifts. Much wider flexural uplifts can be generated by assuming larger elastic thicknesses, as great as 115 km (e.g. Chéry *et al.* 1992; ten Brink & Stern 1992), but those are much greater than the elastic thickness calculated independently for most EPCMs (Watts 2001, fig. 8.30). Elastic thicknesses corresponding to that of the lithosphere could arise if the crust and lithospheric mantle were locked and extended together, i.e. in pure shear, but then the rift margin should subside and form a 'steer's head' after cessation of rifting. It is difficult to envisage where such large elastic thicknesses might arise in the case of simple shear, because that failure mode requires a weak layer somewhere in the deep crust or upper mantle to allow extension of the crust and mantle to be decoupled. In that event, the elastic response of the crust and mantle should be independent and observed elastic thicknesses should be of the same order as the strong layer of the crust.

Extension to form rifts at extension factors of less than about 2 appears to be approximated by simple shear, interrupted by episodes of uplift that form second order unconformities. Post-rift subsidence is close to the McKenzie (1978) model, and the rift margin should subside by flexural response to sediment loading to form 'steer's head' geometry. Since this appears to be how the proximal parts of most passive margins extend, 'steer's head' geometry should be normal on a passive margin, and there should be no permanently-uplifted rift margin.

7.3 Uplift by underlying hot and/or upwelling mantle

A number of authors have proposed that EPCMs may be uplifted by mechanical support from an underlying mantle upwelling or plume or by isostatic uplift from low-density asthenosphere originating from such plumes. Jones *et al.* (2012), for example, used the admittance, which is the ratio of the coherent parts of the gravity and topography (McKenzie & Bowin 1976; Watts 2001), to suggest that the uplifted margins around southern Africa are supported by several mantle diapirs. Only the longer wavelength part of admittance reveals mantle isostatic anomalies. Shorter wavelength anomalies show isostatic anomalies arising from the crust and have been used to calculate elastic thicknesses of the lithosphere (Watts 2001), but McKenzie & Fairhead (1997) have shown that estimates of elastic thickness using calculations of admittance are erroneous in areas of smooth topography, because there is insufficient signal from the topography. Much of the interior of southern Africa is very flat, although at an altitude of between 1 and 2 km, whereas the topography of the margins is much more rugged. It might be that Jones *et al.*'s (2012) calculations show edge effects of a much larger mantle diapir under the whole of southern Africa which is lifting an area around 2000 km in diameter (Ni *et al.* 2002).

In any case, it is difficult to see why mantle plumes should preferentially arise under passive continental margins as suggested by Jones *et al.* (2012). The margins on opposite sides of the South Atlantic are both uplifted, but are moving apart. There is evidence for uplift of southern Africa by a mantle diapir (Ni *et al.* 2002), but there is no large-scale, free-air gravity anomaly across the South Atlantic that suggests that any mantle diapir extends there. So any plume under eastern Brazil would have to be independent of the African one, but would yet have to keep pace with the moving South American plate. This is special pleading. It might be that the presence of a passive margin could induce mantle convection. This idea has been investigated by King & Anderson (1998) who showed that any induced convection in such a place in the presence of horizontal asthenospheric flow would cause downwelling of the mantle on the continental side of the passive margin, not upwelling.

Recently, Rickers *et al.* (2013) published a high-resolution, S-wave velocity interpretation of the North Atlantic region that showed a low-velocity layer beneath much of the oceanic lithosphere, consistent with the long-wavelength, bathymetric high of the North Atlantic. The low-velocity layer extends locally beneath the continental lithosphere of the southern Scandinavian Mountains, the Danish Basin, part of the British Isles and eastern Greenland. Rickers *et al.* (2013) argued that the spatial correlation between the low-velocity layer and uplifted regions suggested dynamic support by low-density asthenosphere originating from the Iceland and Jan Mayen hotspots. Rohrman & van der Beek (1996) had earlier proposed that the Norwegian mountains could be supported by a diapiric intrusion into the lithosphere of some anomalously warm asthenospheric material flowing from the Icelandic hotspot. This hypothesis was tested by Pascal & Olesen (2008), who used heat-flow measurements over Norway to show that there is no abnormally hot mantle under the Norwegian mountains, and this observation would seem to suggest that Rickers *et al.*'s (2013) modelling may not account for all of the factors uplifting the Scandinavian mountains.

Estimates of the amounts of dynamically supported uplift of a continental margin are based on the amounts calculated from adjacent, abnormally shallow oceanic crust. Extension of the calculations to the adjacent continents suggests that there should be up to a kilometre of dynamic support in onshore central East Greenland, smaller amounts in onshore Norway and the British Isles and negligible amounts in West Greenland and Baffin Island (e.g. Steinberger 2007), the exact amounts depending on assumptions about the density and rheology of the continental crust and upper mantle. A kilometre of dynamic support is insufficient on its own to explain the heights of the East Greenland EPCM, and there is insufficient uplift even when the effect of dynamic uplift is combined with the calculated amount of flexed uplift caused by erosion of the fjords and glacial valleys (Medvedev *et al.* 2013). Japsen *et al.* (2013a), however, drew attention to the much higher elevation of East Greenland compared to West Greenland and suggested that this difference could be due to dynamic support in the east from the Iceland plume.

Other EPCMs that probably contain elements of dynamic support from present-day mantle upwelling are the margins around the Red Sea (Al-Hajri *et al.* 2009). There is, however, no *a priori* reason to invoke present-day mantle upwelling under eastern Australia, western India and eastern Brazil if other explanations for the presence of EPCMs there can be found.

7.4 Uplift due to compression

Passive continental margins are, by definition, produced under conditions of extensional stress. Continental crust extends, thins and eventually breaks, and oceanic crust forms in the break and widens such that the now riven continental fragments move apart. The extended margin then cools and subsides and sediment is deposited onto the subsiding continental and oceanic crust. The lithosphere must continue to extend at the mid-ocean ridge, implying net extensional stresses there, but there is increasing evidence that the continental margins on either side of the ocean, and even oceanic crust, may come under compression after breakup (e.g. McAdoo and Sandwell 1985; Cloetingh & Wortel 1986; papers in Doré *et al.* 2002; Cobbold *et al.* 2007; Holford *et al.* 2011). The World Stress Map Project (Heidbach *et al.* 2008) shows that, where data exist, all EPCMs are under compression. A number of authors (e.g. Løseth & Henriksen 2005; Pedoja *et al.* 2011; Japsen *et al.* 2012a) have speculated that the uplift of EPCMs may be caused by compression, but have not proceeded to a quantitative model of how such compression may affect a passive continental margin.

Cloetingh & Burov (2010 and references therein) have argued that continental crust in many parts of the world is under sufficient compression that sinusoidal, crustal-scale folds have formed, and Cloetingh *et al.* (2008) extended the discussion to EPCMs and suggested that they formed by the effects of compression. The folding described by Cloetingh & Burov (2010) cannot arise from strains originating in ridge push, because even the weakest continental crust is too strong. Strain from ridge push amounts to around 40 MPa, though this can rise to around 100 MPa if a large mantle plume head underlies the ridge (Bott 1993).

Compression stress sufficient to induce folding of continental crust depends on both the chemistry of the crust and its temperature regime. There is general agreement that the uppermost 12-20 km of continental crust consists predominantly of wet, quartz-rich lithologies, such as granite and granitic gneiss. Weak lower crust can arise if the lower crust is also granitic or if it consists of more basic materials and the temperature gradient is high; so high that uppermost mantle would be melting and basaltic volcanism would be expected in such areas. Continental crust could fold under an imposed stress of around 300 MPa if its strength resides primarily in the uppermost *c.* 15 km, but the resulting folds would have wavelengths of 100 km or less; much shorter than the typical 200–250 km wavelength reported by Cloetingh & Burov (2010).

There would be no detachment between continental crust whose lower part consists of basic or ultrabasic rock and the upper mantle. In such a case, the entire lithosphere would have to fold as a unit. This would require compressional stresses approaching 2 GPa and would give rise to folds with a wavelength around 400–500 km. McAdoo & Sandwell (1985) have reported folded oceanic lithosphere in the Indian Ocean that would require such stresses, where Cloetingh & Wortel (1986) model a concentration of compressional stress. It is, however, possible to fold continental lithosphere at much lower stresses if the lower crust is of intermediate composition.

Folds with wavelengths of the order of 200–250 km, as reported by Cloetingh & Burov (2010), can arise if the lower crust is of intermediate composition and is detached from the underlying mantle by a weak zone at the base of the crust (see e.g. Burov 2009). To fold such crust requires stress of the order of 1GPa; stress of the same order as that required to drive an orogeny. It is therefore possible that stress originating from an orogeny on one part of a plate may be sufficient to cause folding in other parts of the same plate including uplift of a formerly passive margin on the opposite side of the continent from the orogeny.

Cobbold *et al.* (2001, 2007) observed that post-rift uplift phases in Brazil in the Campanian, Eocene and Miocene are synchronous with the Peruvian (90–75 Ma), Incaic (50–40 Ma) and Quechuan (25–0 Ma) uplift phases in the Andes. Subsequent studies confirmed this result (Cogné *et al.* (2012) and extended the synchroneity of all three events to the West African margin (Japsen *et al.* 2012b; see Fig. 70). The Andean phases coincided with rapid convergence on the western margin of South America and uplift in the Campanian coincided with a decline in spreading rate at the Mid-Atlantic Ridge (Torsvik *et al.* 2009). Because the uplift phases in Brazil and Africa are common to the margins of two diverging plates, Japsen *et al.* (2012b) suggested that the driving forces can transmit across the spreading axis, so their origin must therefore be in the asthenosphere, and that the common cause for both vertical movements and lateral changes in the motion of the plates is lateral resistance to plate motion.

Other authors have noted similar correspondence elsewhere. Cloetingh *et al.* (1990) showed that rapid late Neogene subsidence and sedimentation around the North Atlantic were consistent with rapid changes of intraplate stress that most likely also gave rise to major changes in plate motions at that time. Janssen *et al.* (1995) found a correlation between changes in plate motions and the evolution of rifted basins in Africa. A phase of uplift and erosion of margins on both sides of the North Atlantic (Green & Duddy 2010; Japsen *et al.* 2010) at the Eocene–Oligocene transition (*c.* 35 Ma) happened at the same time as a reorganisation of the spreading ridge there (Gaina *et al.* 2009). These observations again indicate the presence of an influence that can transmit across a spreading ridge, and that uplift events of EPCMs correlate with changes in plate motion.

The most likely source of these effects and of large stresses within the crust is basal traction from asthenospheric currents. Alvarez (2010) argued that the continued collision of India with Asia, after the descending slab had detached from the subducting continent, cannot be explained by slab pull and that ridge push alone is insufficient. Thus some basal traction must also be involved. Husson *et al.* (2012) argued that uplift of the Andes can be explained by basal traction arising from an upwelling plume under southern Africa pushing the South American plate against the subducting Nazca plate. This model implies that the entire South American plate is in compression (cf. Cobbold *et al.* 2007), and this compression may be sufficient to fold the continental crust and cause uplift of eastern Brazil. Horizontal, asthenospheric flow on its own does not produce uplift, unless the flowing material is anomalously hot (when the uplift is due to Pratt isostasy). The horizontal flow simply carries the overlying plate with it (like a boat on a river). The flow moves past the base of the lithosphere, creating basal traction and compression, only when the plate meets resistance (the boat hits a rock).

In contrast to the correspondence of events on either side of the South Atlantic, the Miocene events of regional uplift and exhumation around the North Atlantic began at quite different times in West Greenland and southern Scandinavia. In Greenland, the uplift of the UPS began at *c.* 10 Ma (section 5.5) whereas the onset of Miocene uplift and exhumation around southern Norway occurred in the earliest Miocene at *c.* 23 Ma (Japsen *et al.* 2007a; Rasmussen 2008, 2010), and this movement is reflected in a regional, base-Miocene unconformity in the NW European margin (Stoker *et al.* 2005). These observations indicate that the driving forces for the uplift of southern Scandinavia in the earliest Miocene are related to forces transmitted within the Eurasian plate rather than in the asthenosphere, which is why they did not affect West Greenland.

None of the effects of mantle-driven uplift or compression are, however, long lasting. Changes in mantle up- and downwelling may take place in very short times, less than 1 million years (e.g. in the Palaeogene of West Greenland, see chapter 5 and in the Faroe-Shetland region of the UK North Atlantic Margin, see Shaw-Champion *et al.* 2008). Changes in stress necessary to produce third-order sequences take place on a time-scale of 3 million years or less (Cloetingh & Ziegler 2009 and references therein) and changes in the compressional stresses large enough to cause crustal-scale folding must also take place on timescales less than 10 million years (see section 8.1).

7.5 Summary

The shared characteristics of EPCMs (high topography with inland sloping plateaux cut by incised valleys and steeper oceanward decline) suggest that EPCMs have similar underlying causes. We thus reject the assertion by Maupin *et al.* (2013, p. 20) that "each mountain chain has its own characteristics, making it unique and leading to different preferred mechanisms for their tectonic evolutions". Causes that require special conditions that apply to only one or a small number EPCMs are unlikely to provide satisfactory explanations for the formation of EPCMs in general. The numerous observations presented here indicate that EPCMs form long after rifting and break-up as a result of episodic burial, uplift and exhumation.

As discussed in section 7.4, it seems likely that some EPCMs are, at least partially, uplifted by upwelling mantle diapirs and/or hot underlying asthenosphere derived from them. There is no evidence, however, for present-day upwelling mantle diapirs near e.g. the Brazil or east Australian EPCMs, so this explanation seems insufficient on its own to account for all EPCMs. Some other effect seems to be necessary, and one likely explanation may be lithosphere-scale folding combined with flexed isostatic response to local erosion onshore and sediment loading offshore (Cloetingh *et al*. 2008). The lithosphere-scale folding is caused by compression, itself derived either from orogenies elsewhere on a plate or from basal drag of the lithosphere by horizontal asthenospheric flow. The effects of these two sources of the forces driving vertical motion along EPCMs may explain why both peneplains and events of cooling/exhumation detected by AFTA are regional in extent and also explain why some uplift events are common to margins of divergent plates whereas some are not.

8. Key issues concerning the development of EPCMs

8.1 Steady-state or transient landscapes, ancient landscapes or young landscapes?

That landscapes are in a steady state is a view that is held by many researchers modelling isostatic uplift of continental margins (see references in Bishop 2007), and Summerfield (2000) and others have cast doubt over the idea that landscapes have a memory indicating former base levels. The results discussed in chapters 5 and 6 question these views. Even though it is clear that a landscape is all the time developing toward a steady state, there are reasons to believe that there is a considerable time lapse between cause and effect in the development of landscapes (cf. Brunsden 1993), giving rise to what are now sometimes called transient landscapes (see Bishop 2007). Formation of extremely low-relief landscapes such as the Sub-Cambrian Peneplain demonstrates that it is possible to produce a peneplain as the end result of erosion (section 3.1.1). A tectonic uplift event is needed to change from a situation of subsidence to re-expose the old landscape, and tilting will produce valley incision (section 3.1.5).

Isostatic compensation for erosion does not, however, mean that isostasy maintains the height above sea level of a landscape despite erosion. Erosion always reduces the *average* height of a landscape, although isostacy causes the reduction to be less than the amount eroded; to reduce the mean elevation of a region by a certain amount, a much larger thickness of rock must be removed. Assuming Airy isostasy, a thickness of rock $\Delta t = \Delta h(1-\rho_c/\rho_m)^{-1}$ must be eroded to lower the height above sea level by Δh, where ρ_c and ρ_m are the densities of material being eroded and mantle respectively. Figure 73 shows that to denude a region to a low elevation in basement (i.e. to form a peneplain) a considerable thickness of rock must be removed, whereas to achieve the same in sedimentary rocks a much smaller removed thickness is required, because of the lower density of the eroded rocks. This may help to explain why in many areas, creation of peneplains in basement is associated with exhumation from palaeotemperatures >110°C, as recorded in AFTA data, whereas subsequent deposition of sedimentary cover and removal to form a new peneplain often involves removal of a smaller thickness of section, and lower palaeotemperatures. For example, the presently exposed basement of e.g. West Greenland was metamorphosed at depths >20 km during the late Proterozoic (*c.* 1600 Ma), and that amount of upper crust had been removed prior to the Ordovician (400 Ma). In contrast, subsequent events have involved deposition and removal of much smaller thicknesses of Phanerozoic cover. The history of the continental margin of North-East Greenland illustrated in Figs 74, 75 provides another illustration of these points.

In the old studies of continental margins, the summit plateaux were thought to be graded to pre-uplift sea level and uplift was regarded to be of Neogene age (e.g. Reusch 1901; Craft 1933; Ahlmann 1941), while more recent studies were based on the idea that these surfaces were equivalent to the pre-break-up surface and thus of Mesozoic age (e.g. Ollier 1985). Crosscutting relationships between peneplains (Fig. 11) are important to determine, in order to draw conclusions on how far the re-exposed peneplains extend and to ascertain where erosion has cut deeper than a re-exposed peneplain. The old ages

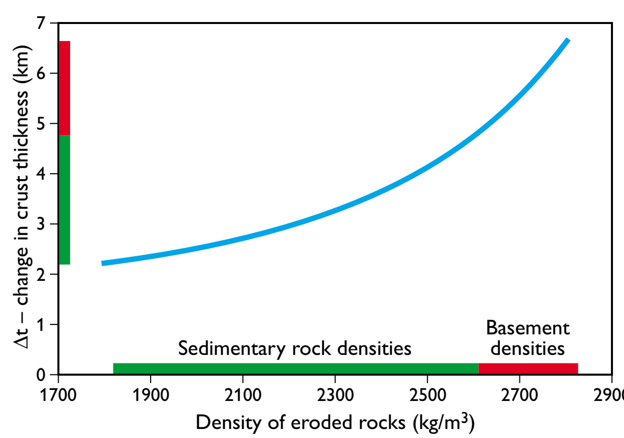


Fig. 73. Plot of the amount of erosion (change in crustal thickness) needed to reduce the mean elevation of an uplifted area by 1 km vs. the density of the eroded rocks (assuming local Airy isostasy and a mantle density ρ_m = 3300 kg/m^3). Erosion always leads to a reduction in elevation, but, because of isostasy, the reduction is less than the amount eroded. Erosion and isostasy do thus not lead to a 'steady state'. If basement rocks of density ρ_c = 2800 kg/m^3 are eroded, 5.7 km of erosion must take place (i.e. the crust thins by this amount) for every km of elevation reduction. In contrast, removal of only *c.* 2.5 km of sedimentary cover (ρ_c = 2000 kg/m^3) is needed to lower the elevation above sea level by 1 km.

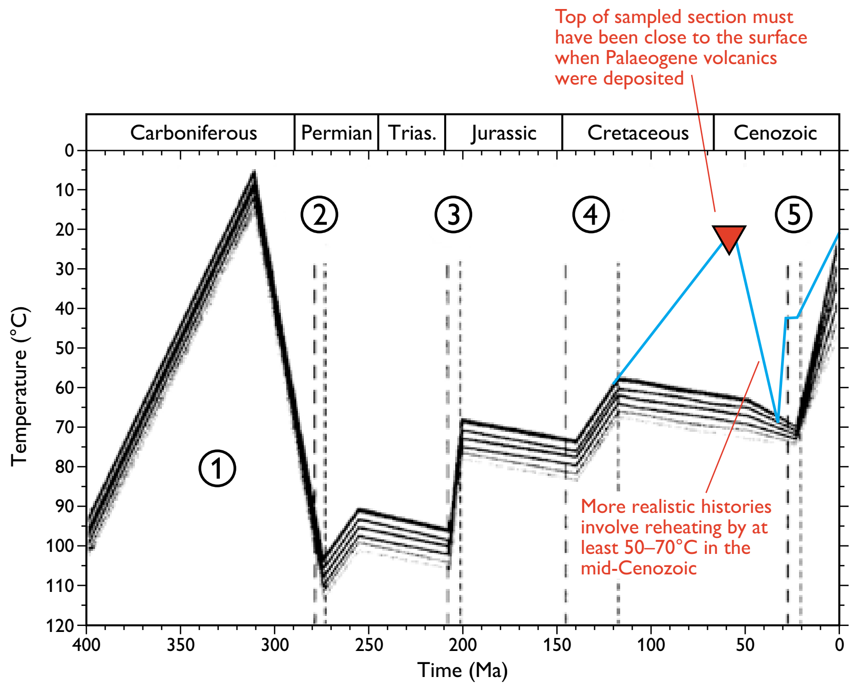


Fig. 74. Thermal history solutions for a vertical section of Carboniferous sandstones on Clavering Ø, East Greenland (*c.* 74°N; location shown in Fig. 75), from Johnson & Gallagher (2000). Also shown is a more realistic interpretation obtained by incorporating the presence of the Palaeogene volcanics at the highest elevations above the section sampled by Johnson & Gallagher (2000). These constraints show that the uppermost sample must have been at near surface temperature in the Palaeogene, implying a major degree of re-burial by additional volcanics and/or post-volcanic Palaeogene sediments prior to the onset of cooling from the mid-Cenozoic palaeothermal peak of about 70–80°C. The alternative cooling history indicated matches two episodes of cooling at c. 35 and 10 Ma (Thomson *et al.* 1999). Each line represents the temperature history of an individual sample. In section (**1**), fission tracks are produced and annealed in the source region of the sediment prior to deposition; in section (**2**), maximum temperatures at 274 Ma (mid-Permian); in section (**3**), cooling at 206 Ma (Early Jurassic); in section (**4**), poorly constrained cooling at 140 Ma (Early Cretaceous); and in section (**5**), Cenozoic reheating and maximum temperatures at 23 Ma. Modified after Japsen *et al.* 2010.

suggested in many morphological studies of peneplains of the Gondwana continents (Ollier 1982; Twidale 1985, 2007) might be explained by extrapolating too far from covers without noting changes in inclination of the peneplains and/or interpreting such changes as caused by warping instead of deeper erosion in the basement and formation of a younger peneplain. The older idea of young (late Palaeogene/Neogene) ages of high-level peneplains (e.g. Reusch 1901; Ahlmann 1919, 1941) seems to be more realistic when tested against AFTA (chapters 5 and 6).

Japsen *et al.* (2006, 2012b) investigated the time available for the formation of peneplains in West Greenland and north-east Brazil and found that an interval of 20 million years is sufficient, after rejuvenation of relief during an initial uplift event (see Fig. 70). Japsen *et al.* (2006, 2012b) showed that in West Greenland, the present-day relief formed after uplift that began in the late Miocene (*c.* 10 Ma; see Fig. 54) whereas the relief in north-east Brazil formed after uplift that began in the early – middle Miocene (between 18 and 15 Ma; Fig. 68).

It seems possible that in many places around the world, much of the present-day relief is not older than the Neogene/late Palaeogene (cf. Thornbury 1969), unless it has been re-exposed from below a former cover. How far back in time the present landscape has a memory seems to depend on bedrock resistance, climate, and location (Twidale 1976; Brunsden 1993; Bishop 2007).

Recently, Egholm *et al.* (2013) drew renewed attention to a long-standing debate (also see e.g. Baldwin *et al.* 2003) concerning the apparent enigma of the long-term preservation of kilometre-scale mountainous relief in ancient Palaeozoic orogenic belts which are tectonically inactive today, citing examples including the Caledonides of Greenland and Scandinavia and the Lachlan orogen of south-east Australia. In the light of the results from these areas presented here, this enigma can be simply resolved by acknowledging that the mountains in these classic

EPCM regions and others are the result of much more recent rejuvenation and uplift, unrelated to the ancient orogenies. As described in chapters 2 and 3, the form of landscapes should be regarded as a source of information on their development. This information can then be used to design more accurate modelling studies, rather than designing such studies to explain a particular view by choosing appropriate parameters.

8.2 Landscape evolution and thermochronology: progressive emergence (continual cooling) vs. episodic burial and exhumation (heating and cooling)

As discussed in preceding chapters, a common assumption in thermochronological studies of EPCMs is that the rocks comprising the uplifted areas have cooled monotonically. This presumably reflects the widespread notion that rocks in such regions reach the present-day surface through a long history of relatively slow denudation. This assumption is difficult to reconcile with the observations from West Greenland presented in chapter 5, where the exposed geology documents a series of episodes of burial and exhumation. As discussed in chapter 6, results from other EPCMs, including Brazil, South Africa and south-east Australia can be interpreted in terms of a similar episodic style of evolution. In addition, SLA in southern Scandinavia provides clear evidence of episodic evolution (chapter 3).

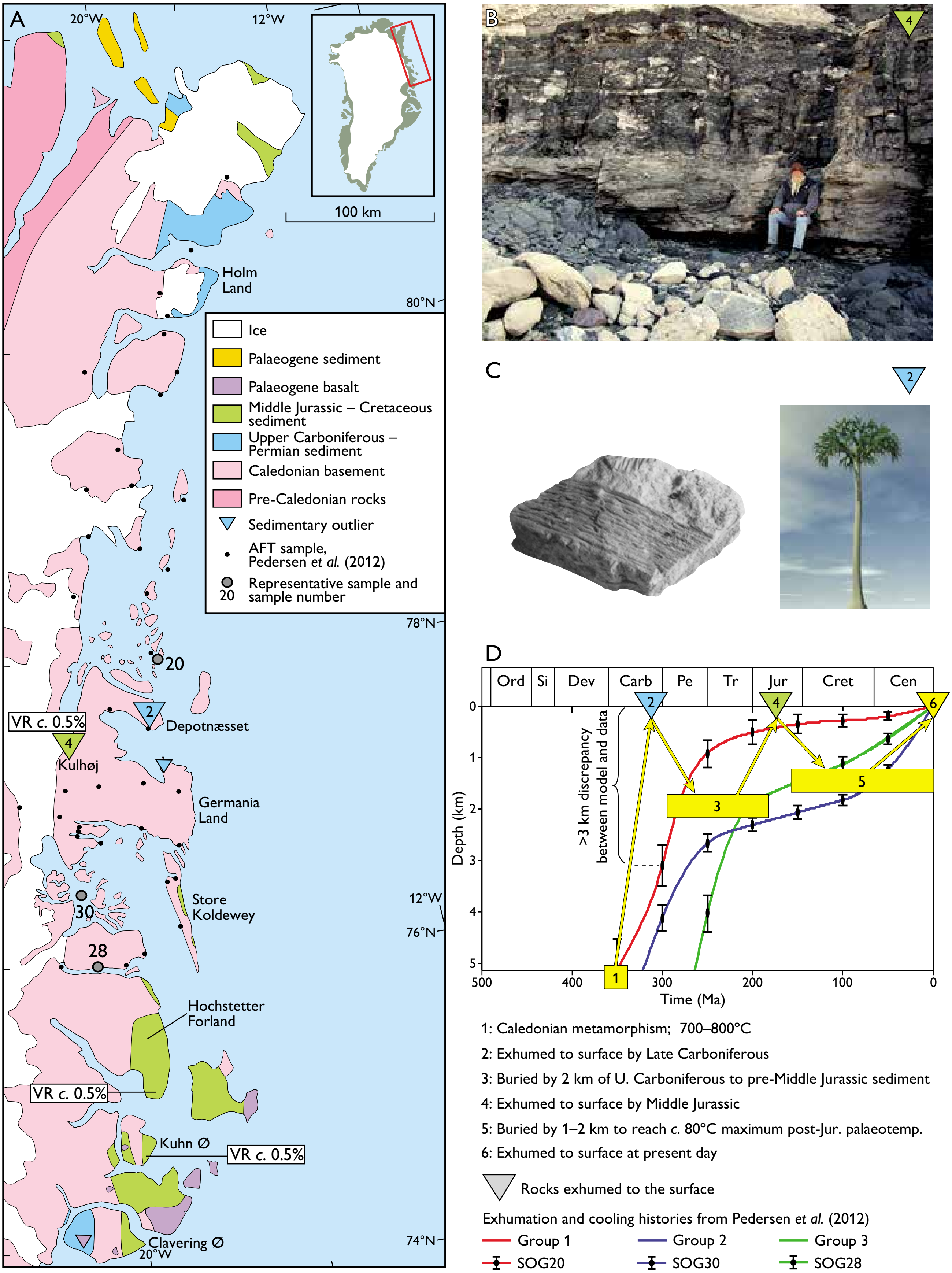
The concept of monotonic cooling appears to derive from classical models of landscape development. Major landscape forms such as low-relief erosion surfaces in stepped sequences led to the idea of continuous uplift interrupted by periods of quiescence, and the formation of vast peneplains (Davis 1899) or pediplains (King 1951). King (1962, 1967, 1983) acknowledged the former existence of covers, e.g. a former Cretaceous cover of the Brazilian margin, and noted that there had been periods of subsidence and burial. Yet, working with landforms, he mainly discussed the uplift events. In Europe the idea of a continuous lowered base level since the Cretaceous was used by Baulig (1935) to argue that this was the cause of erosional steps seen in the topography (the 'eustatic theory' referred to by Chorley 1965; also see chapter 2) as by then the continents were thought to be stable. Thus the highest surfaces were considered the oldest; it was thought that either uplift had taken place or that sea level had been lowered. In general, the old studies did not consider re-burial as an important factor in landscape development, because in those days there was no way of documenting whether thick sections of rock might have been removed. Thus landscapes were analysed with the assumption of a monotonic, though intermittent rise relative to sea level.

We know now that eustatic sea-levels have never varied by more than a few hundred metres (Miller *et al.* 2005), and evidence has now accumulated that disproves the notion of continental stability (e.g. Moucha *et al.* 2008). Despite these changes in understanding, the notion of continual emergence, which was originally devised as an essential component of the eustatic explanation, has remained from times that pre-date the acceptance of a more mobile crust dominated by mantle-driven processes in a plate-tectonics environment (see chapter 7; e.g. Steinberger 2007; Roberts & White 2010; Jones *et al.* 2012).

Re-exposed surfaces of different age (Ambrose 1964; Twidale 1985) are indicators of former covers. Yet such surfaces have not been commonly used in the past to reconstruct landscape development. On land, such covers are just remnants and originally must have been both thicker and extended farther, even outside the present distribution of re-exposed relief, which is commonly cut off by a younger peneplain (Figs 11, 14). Such covers seem to have been the norm. Covers of different age directly on Precambrian basement in Fennoscandia indicate that the major erosional phase of basement rocks occurred before the Cambrian (Lidmar-Bergström 1997). Major periods of covering in Scandinavia were then the Palaeozoic, the late Mesozoic, and probably the Eocene, while most of the Mesozoic, the Paleocene and post-Eocene were periods of uplift and erosion in the basement (Lidmar-Bergström 1995, 1996). Lidmar-Bergström *et al.* (2013) thus showed that the relation between relief in basement and cover rocks of different age directly on basement in Scandinavia provides important information about the Phanerozoic uplift and subsidence history of the region.

In the approach favoured by some geomorphologists (section 2.4), re-exposed relief is often assumed to be evidence of removal of just a thin cover, and this assumption may lead to difficulties in accepting the results of apatite fission-track data (Gunnell 2000; Peulvast *et al.* 2008; also see section 6.4).

Where sedimentary outliers are present on basement rocks, they provide an important constraint, defining times when the basement surface was previously at or very close to the land surface. In such circumstances,


A
20°W
12°W
100 km
Holm Land
80°N
Ice
Palaeogene sediment
Palaeogene basalt
Middle Jurassic – Cretaceous sediment
Upper Carboniferous – Permian sediment
Caledonian basement
Pre-Caledonian rocks
Sedimentary outlier
AFT sample, Pedersen et al. (2012)
20 Representative sample and sample number
78°N
20
VR c. 0.5%
4
Kulhøj
2
Depotnæsset
Germania Land
Store Koldewey
12°W
76°N
30
28
Hochstetter Forland
VR c. 0.5%
Kuhn Ø
VR c. 0.5%
Clavering Ø
20°W
74°N
B
4
C
2
D
Ord
Si
Dev
Carb
Pe
Tr
Jur
Cret
Cen
Depth (km)
>3 km discrepancy between model and data
Time (Ma)
500
400
300
200
100
0
1
2
3
4
5
6
1: Caledonian metamorphism; 700–800°C
2: Exhumed to surface by Late Carboniferous
3: Buried by 2 km of U. Carboniferous to pre-Middle Jurassic sediment
4: Exhumed to surface by Middle Jurassic
5: Buried by 1–2 km to reach c. 80°C maximum post-Jur. palaeotemp.
6: Exhumed to surface at present day
Rocks exhumed to the surface
Exhumation and cooling histories from Pedersen et al. (2012)
Group 1
Group 2
Group 3
SOG20
SOG30
SOG28

interpretations involving monotonic cooling histories will produce meaningless results. This may seem obvious, but a number of published examples have failed to take the presence of cover rocks into account (Johnson & Gallagher 2000; Persano *et al*. 2006; Pedersen *et al*. 2013, as illustrated in Figs 74–77. Clearly, thermal histories involving continuous cooling are inappropriate in areas where cover rocks are preserved. Integration of the constraints provided by cover rocks is essential in order to obtain realistic thermal histories.

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Fig. 75. Conflicting interpretations of the post-Caledonian development in North-East Greenland illustrating the importance of stratigraphic constraints in the interpretation of thermochronology data. **A**: Geology of North-East Greenland and location of apatite fission-track (**AFT**) samples analysed by Pedersen *et al*. (2012). Outliers of upper Carboniferous – Permian and Jurassic–Cretaceous sedimentary rocks (triangles coloured according to the legend) occur in patches between 75° and 80°N, while these sedimentary units are more extensive farther south and farther north. The presence of these sedimentary rocks demonstrates that the underlying basement was at the surface prior to their deposition. **Grey circles**: samples SOG20, 28, 30 that Pedersen *et al*. (2012) chose as representative of their three groups of samples. **B**: Outcrop of Middle Jurassic coal located at triangle 4 on the map. **C**: Upper Carboniferous deposit located at triangle 2 on the map; fossil and reconstruction of Lepidodendron (Piasecki *et al*. 1994). **D**: Exhumation paths (red, blue and green) based on modelling by Pedersen *et al*. (2012) of apatite fission-track data in three representative samples chosen by these authors. Superimposed curves (yellow) illustrate the geological constraints on the burial and exhumation history of basement rocks around Germania Land. (**1**) Caledonian basement with eclogitic inclusions reflects high-pressure metamorphism of Devonian age (700–800°C, 410–390 Ma; Gilotti *et al*. 2008). (**2**) The presence of upper Carboniferous sediment overlying the basement shows that the basement was exhumed to the surface by the late Carboniferous (Piasecki *et al*. 1994). (**3**) Maturity of sporomorphs in these deposits indicates that they were buried below a cover, 1.5–2 km thick (Piasecki *et al*. 1994). (**4**) The presence of Middle Jurassic sediment overlying the basement shows that the upper Carboniferous – Middle Jurassic cover was partly removed by the Middle Jurassic (e.g. Bojesen-Koefoed *et al*. 2012). (**5**) Vitrinite reflectance (VR) values of *c*. 0.5% for the Jurassic deposits show that they were buried below a sedimentary cover, 1–2 km thick (Bojesen-Koefoed *et al*. 2012). (**6**) Today the Caledonian basement is exposed at the surface over most of the region. Modified from Japsen *et al*. (2013b).

8.3. EPCMs: permanent highs or the result of late uplifts?

The development of the relief of the Natal monocline after formation of the African Surface (below the Drakensberg Escarpment) resulted in lower steps in the landscape (several generations of peneplains and valleys; Fig. 2). The only explanation presented so far for these stepped surfaces and incised valleys is grading to lowered base levels following Neogene uplift events (King 1972, 1983; Partridge & Maud 1987). The presence of both several high surfaces and lower valley steps is neglected in the prevailing approach to such matters as discussed in section 2.4. Ollier & Marker (1985), for example, who studied the Great Escarpment of southern Africa, only identified one palaeoplain that they argued to be of Mesozoic age. They assumed that the margin had remained high since break-up, and suggested two models for its development, marginal downwarp and rift-shoulder uplift (Fig. 57). These ideas were adopted in the models created and tested by e.g. Brown *et al*. (2000) and Persano *et al*. (2002). These authors adopted the new methods of thermochronology and focussed on the classic areas of Lester King in southern Africa and of Cliff Ollier in eastern Australia. Their models were based on three assumptions formulated by Gilchrist & Summerfield (1994). First, that the present topography has an origin related to continental break-up. Second, that the marginal upwarps of mature passive margins cannot be explained by the dynamic effect of rifting as the axis of maximum uplift is now located 100 km or more inland and therefore they regard isostatic response as an important factor (cf. King 1956b). Third, that the new continental margins were originally at high elevation at break-up due to either rift-related surface uplift or pre-existing residual high terrain. Their idea, which many have accepted, is that the continental margins reached their present elevations before break-up of Gondwana and have remained elevated ever since. The idea of an originally high margin is nowadays often taken for granted (e.g. Campanile *et al*. 2008).

In contrast, early observations (section 8.1) of the relationships between high-level peneplains and incised valleys gave a different explanation of EPCM development as a mainly Neogene story. This description is consistent with our results from West Greenland and elsewhere (chapters 5 and 6). As shown in these chapters, there is now a strong body of evidence to show that passive margins have moved up and down and that their recent elevation is coupled to and maintained by geodynamic processes hitherto not completely understood.

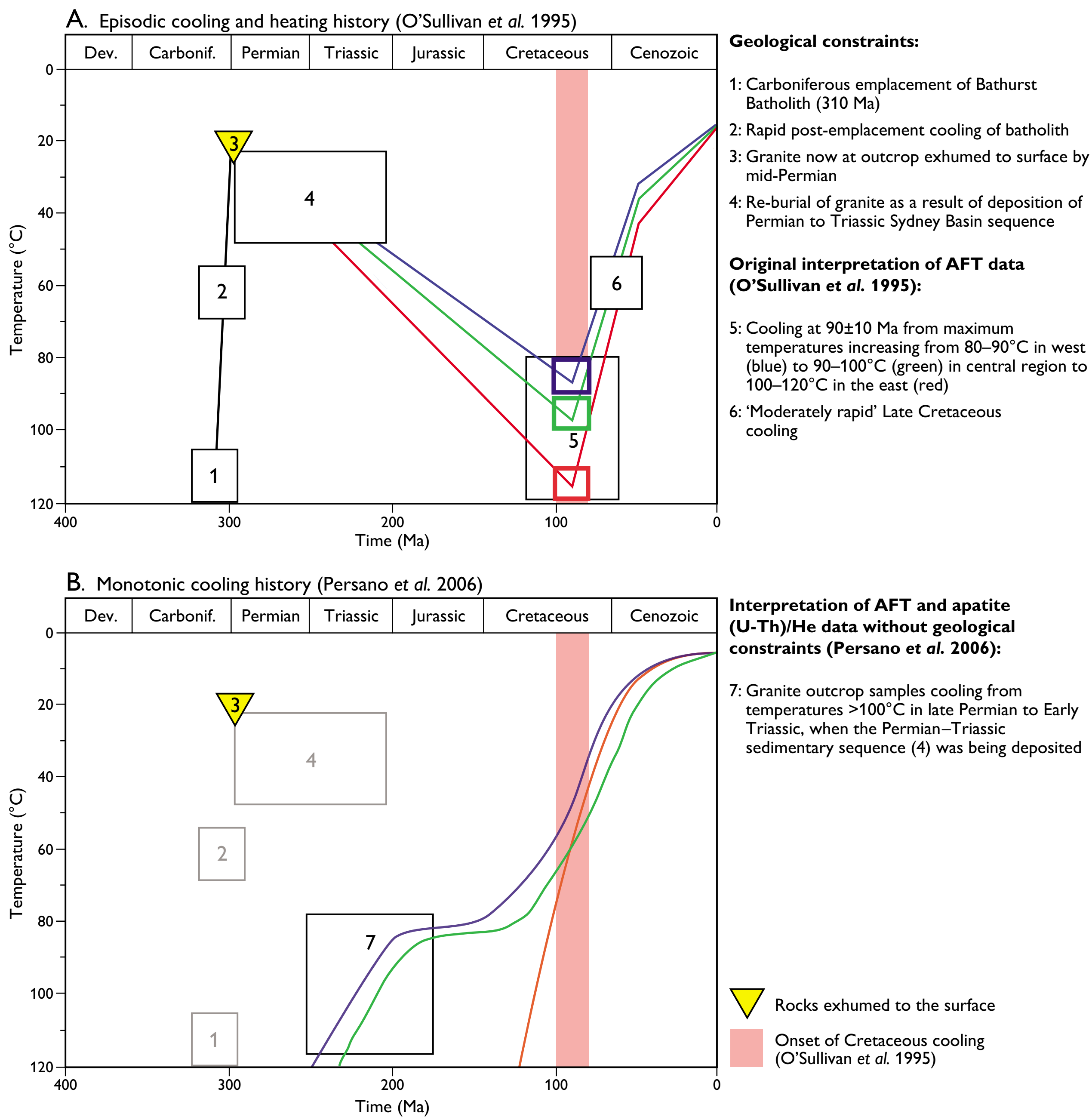


Fig. 76. Two conflicting interpretations of the post-Carboniferous development in south-east Australia, illustrating the importance of stratigraphic constraints in the interpretation of thermochronology data. **A**: Thermal history interpretation of granite samples from the Bathurst Batholith, west of Sydney, New South Wales, from O'Sullivan *et al.* (1995), together with constraints provided by geological evidence, viz: (**1**) emplacement of the Batholith at *c.* 310 Ma (Facer 1978); (**2**) rapid post-emplacement cooling as demonstrated by (**3**) granite now at outcrop (and sampled by O'Sullivan *et al.* 1995) overlain by Permian sedimentary units of the Sydney Basin sequence; (**4**) deposition of Permian to Triassic sedimentary units of the Sydney Basin sequence above present-day granite outcrops. O'Sullivan *et al.* (1995) interpreted apatite fission-track (**AFT**) data in samples of both granite and overlying Permian sedimentary remnants as defining heating to Early Cretaceous temperatures (**5**) from 80–90°C in the west to 90–100°C at central locations and 110–120°C in the easternmost outcrops where the granite disappears beneath the sedimentary cover, followed by rapid cooling (**6**) which began in the interval 90±10 Ma. **B:** Persano *et al.* (2006) integrated apatite (U-Th)/He ages from samples of outcropping granite with the apatite fission-track data of O'Sullivan *et al.* (1995) and interpreted the results in terms of slow cooling, with samples in the west and centre of the study region cooling through *c.* 100°C in Permian times (**7**), while samples farther to the east only cooled below *c.* 100°C in the Early Cretaceous. The geological constraints (1–4) show that the granite now at outcrop was close to the surface during the late Permian and being buried by the Sydney Basin sequence. Persano *et al.* (2006) did not take the geological constraints into account, resulting in unrealistic conclusions (see Green *et al.* 2007; Brown 2007; Gibson 2007 for further details).

8.4 The meaning of 'escarpment'

The notion of a 'Great Escarpment' is often confusing in studies of passive-margin development. It sometimes denotes the whole oceanward slope of the EPCM, sometimes the deeply incised valleys below the base of the high-level landscapes and sometimes a structural feature such as the Drakensberg Escarpment in South Africa (Fig. 2). We use the word escarpment in its broadest sense, as a steep slope. Conspicuous, high escarpments are often made up of resistant rocks; e.g. the Drakensberg Escarpment (Moore & Blenkinsop 2006), the escarpment around Beaufort West in South Africa dominated by Karoo Sills and the escarpment along the south-eastern Australian coast south of Sydney in Triassic sandstones.

The notion 'Great Escarpment' was used by Ollier (1982) for the highly irregular backwall caused by the incised valleys, which abruptly cut into the elevated and arched plateau surface of eastern Australia. There are valleys incised into the African Surface in southern Africa of the same nature as those making up the so called Great Escarpment in eastern Australia, but these valleys are incised into the African Surface and thus start below the Drakensberg Escarpment which forms part of Great Escarpment defined by Ollier & Marker (1985) in southern Africa. This difference in the definition of Great Escarpment in the two areas is not noted in recent literature on modelling of EPCMs, and the existence of a stepped landscape on the oceanward side (Fig. 2) is ignored. In the King (1956b, 1972, 1983) scheme, the Drakensberg Escarpment is just one, though prominent, of a series of steps. Ollier (1985), however, did not acknowledge the relevance of a stepped topography and only discussed the case of one high-level plain and one Great Escarpment forming the general coastal slope. It is this description that most modern modellers have used even though it is often not applicable.

The review of the landscapes at different EPCMs presented here suggests that while different EPCMs share many characteristics (e.g. the presence of high-level, low-relief surfaces cut by deeply incised valleys), many differences in detail exist, which provide further information on the development of specific margins. High-level surfaces are often present at several elevations (different for each margin), but often there is a lowest major peneplain below which rivers have cut deep gorges down to the coastal plain. At some locations the valleys have widened to give rise to an escarpment of high sinuosity, as seen in eastern Australia (Fig. 3), probably due to deeply weathered bedrock. In glaciated areas, incised valleys have often been elongated backwards and form long fjords. At some locations resistant rocks have formed a wall, a real escarpment, but this wall can be just one step of a margin with several steps. Thus climate and rock structure influence escarpment appearance. For conclusions on the tectonic development of EPCMs, it is the high-level, low-relief surfaces that provide the most important insights, whereas the exact appearance of the escarpment is of minor importance.

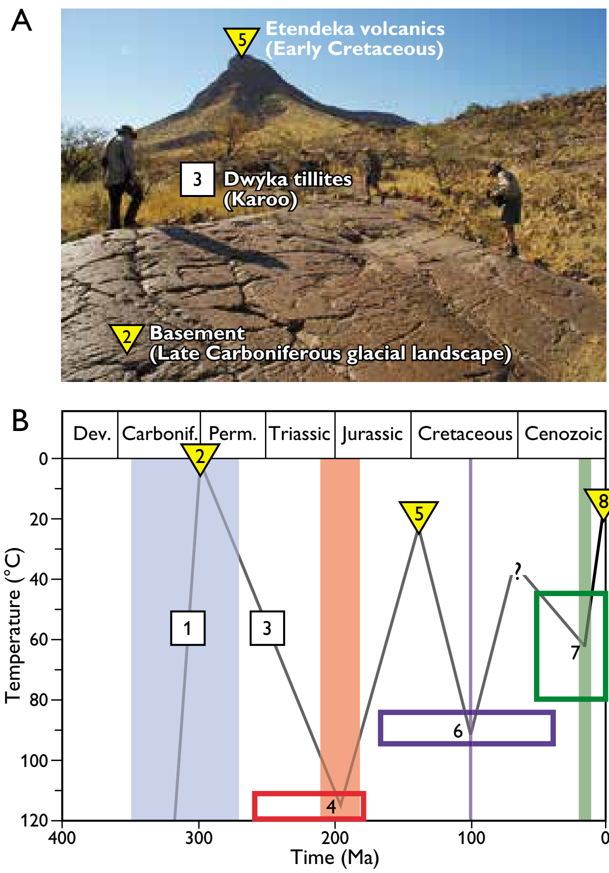
8.5 Down-wearing, back-wearing and scale

There is an important difference between the processes acting on high-level palaeoplains and those that shape deeply incised valleys. High-level palaeoplains are characterised by slow down-wearing processes (the general lowering of a land surface by denudation), while there is an intensified weathering along incised valleys, due to reactivation of groundwater flow, accelerated slope processes and backward retreat at valley heads (Ollier 1982; Thomas 1994).

Some researchers regard deep weathering/stripping of saprolite as the major agent in landscape formation. They refer to these processes as downwearing and consider base level to be of no importance (Thomas 1994; Phillips 2005). 'Etch surface' is then a label for the entire landscape, which is regarded as being in a steady state as denudation (weathering and erosion) keeps pace with uplift, but maybe with a change of saprolite thickness due to climatic changes (Thomas 1989). In SLA, base level is instead regarded as decisive also for etch-surface formation, as erosion of the landscape is limited by this base level. The etching (deep weathering) is a secondary feature, dependent on climate.

Japsen *et al.* (2005, 2006, 2009) showed that in West Greenland the uplifted UPS formed during Oligocene–Miocene times. Over the comparatively short time span until today, a slow down-wearing process of the original peneplain, possibly after stripping of any unconsolidated cover, can be regarded as negligible (cf. Phillips *et al.* 2006). Thus the peneplain level can be used for estimation of rock uplift since peneplain formation.

Scarp retreat is the classical back-wearing concept (King 1962, 1967; Ahnert 1982, 1998). In recent papers, the concept of down-wearing has been used to describe long-term river incision and valley widening which results in major erosion of continental margins (Gallagher *et al.* 1998; Persano *et al.* 2002), thus giving down-



1: Granite exhumed during Late Carboniferous
2: Formation of glacial landscape
3: Deposition of Karoo cover
4: Basement reheated to >110°C below Karoo cover; cooling began *c.* 200 Ma
5: Removal of most of the Karoo cover and deposition of Etendeka basalts
6: Basement and younger cover reheated to *c.* 90°C from which cooling began at *c.* 100 Ma
7: Later episode of cooling
8: Exposure of basement and remaining cover at the surface today

Palaeothermal constraints from AFTA sample
Regional constraints on the onset of cooling
Rocks exhumed to the surface

Fig. 77. Stratigraphic constraints on the thermal history interpretation of AFTA data from a basement outcrop in north-west Namibia (18.184°S, 12.762°E). **A**: Late Carboniferous – early Permian glacial landscape, as evidenced by the roche moutonnée on which the geologists are standing, which is actively being exhumed from beneath a former cover of the late Carboniferous – early Permian Dwyka Group of the Karoo Supergroup. This is overlain, in turn, by basalts of the Lower Cretaceous Etendeka volcanics, comprising the distant peak. **B**: Extraction of thermal history information from AFTA data in apatites from the granite of the roche moutonnée reveals three episodes of cooling, as shown by the coloured box outlines (**4**, **6** and **7**). AFTA data from samples across the region provide constraints on the onset of four episodes of cooling/exhumation, as shown by the vertical, coloured bands. (**1**) Samples farther inland preserve evidence of early cooling, representing exhumation of the granite to the surface. (**2**) Geological evidence shows that the granite had been exhumed to the surface, and was therefore at low temperature, prior to deposition of the Karoo Supergroup. (**3**) The AFTA data show that the rock was then reheated to >110°C, implying deposition of a thick Karoo cover, of which the surviving Dwyka sediments are just a remnant. (**4**) AFTA data show that the granite sample cooled below 110°C at around 180 Ma. (**5**) The presence of the Lower Cretaceous Etendeka basalt only *c.* 100 m above the basement outcrop shows that by *c.* 132 Ma the sample must have cooled to near-surface temperatures, implying that the former Karoo cover had been largely removed by this time. (**6**) AFTA data define another phase of cooling from *c.* 90°C which began at *c.* 100 Ma (based on regional data). This implies re-burial by some thickness of Lower Cretaceous section. (**7**) A later episode of cooling is also defined from the AFTA data. (**8**) Today the granite of the roche moutonnée is exposed at the surface. Integration of AFTA with stratigraphic constraints defines the episodic nature of the thermal history. Interpretation of the AFTA data in terms of monotonic cooling clearly conflicts with geological evidence at this location. Note that if erosion had been more pronounced, and the cover rocks removed, as over a large area of the surrounding country, no evidence would be preserved of this episodic heating/cooling history

wearing the same meaning as back-wearing in classical studies. The idea of a pinned (stationary) divide along the line of the present escarpment of EPCMs was suggested by Gallagher *et al.* (1998), meaning that its position was due to different rates of denudation between the divide (slow denudation), the inland areas (higher denudation), and the coastal area (highest denudation; Fig. 57C). Yet it is evident from profiles by King (1972, 1983) and a profile by van der Beek *et al.* (2002) that the Drakensberg Escarpment is far northwest of the uplift axis of the Natal monocline (section 8.4). Erosion has caused considerable retreat of the escarpment and mainly across less resistant rocks. Moore & Blenkinsop (2006, p. 599) noted that the Drakensberg Escarpment is not restricted to the ocean-ward side and they demonstrated that this feature is due to headward retreat. They write: "Headward retreat of large waterfalls on the major rivers over considerable distances (10–100 km) provides clear field evidence that scarps formed by resistant lithologies will not invariably degrade as a result of the existence of an inland drainage divide, as has been argued on the basis of surface process modelling." That gorge-head migration is a most important process as a bedrock incision mechanism is also demonstrated in two papers from the East Australian margin (Seidl *et al*. 1996; Nott *et al*. 1996).

In their models of development of EPCMs, Gallagher *et al*. (1998) and Persano *et al*. (2002, 2005) present new ideas on how fluvial landscapes develop by river incision and valley widening along a pinned divide without headward retreat. Their denial of headward retreat also seems to be the reason why the fluvial process is looked upon as down-wearing in these models, although they accept the process of valley widening which is part of a general scarp retreat. These are severe inconsistencies.

In the long time span needed for formation of a full peneplain, the question of down-wearing contra back-wearing is not an important issue, compared to what is seen from the thermochronology constrained by sedimentary covers. No significant topography is left. This means that the present difference in process rates between the high plains and the incised valleys, is a relatively new feature. There is thus a scale difference between those processes causing a slow down-wearing of high-level plains and the rapid processes causing back-wearing by river incision and valley widening along EPCMs.

8.6 Are epigene surfaces sometimes re-exposed?

The discussion in chapter 2 distinguishes between re-exposed and epigene peneplains; the former we identify at the contact with the cover, in the case of the latter, no cover is present. However, results discussed in chapters 5 and 6 suggest that the apparently epigene peneplains, the UPS in West Greenland and the Higher Surface in north-east Brazil, have had covers (note that a Palaeogene cover is preserved on some remnants of the Higher Surface in Brazil; section 6.4.4). Epigene peneplains are generally preserved after uplift because of their location far from incising rivers that destroy the peneplains by backwearing. How is it possible that the suggested Oligo–Miocene covers in West Greenland and north-east Brazil can be stripped over large areas and the underlying surface at the same time is totally preserved? Difference in resistance to downwearing between a relatively soft cover and a resistant base provides a possible answer.

The tilted UPS of West Greenland has almost no residual relief and might represent a re-exposed unconformity. The reason why the erosion surface exposed on westernmost Nuussuaq is so well preserved could therefore be that it has been protected by the Miocene cover which is preserved immediately offshore farther west (Figs 41, 42A). In north-east Brazil, the Palaeogene Serra do Martins Formation covers parts of the Palaeogene Higher Surface on the Borborema Plateau but not on Planalto da Conquista (see section 6.4.4). This indicates that the surface is in fact a re-exposed unconformity. If the, by definition, epigene surfaces are sometimes re-exposed from temporary unconsolidated, easily eroded covers, the label epigene peneplain just means that no covers are found on it.

8.7 Episodic development of EPCMs

In preceding chapters, evidence has been presented to show that a number of EPCMs have been uplifted significantly later than the time of rifting and continental break-up, and often in multiple episodes. Japsen *et al*. (2012a) have pointed out that unconformities offshore are commonly extensions of the erosion surfaces present at high level in EPCMs. These unconformities commonly truncate earlier sedimentary successions that generally dip oceanward (Fig. 6), showing that the sedimentary units once extended across the onshore region. These structural relations thus indicate episodic, positive and negative post-rift vertical movements at EPCMs.

One simple explanation why many workers have assumed EPCMs to be permanent uplifts that are somehow related to rifting and break-up, may be that many EPCMs appear to be located above thicker crust/lithosphere in close juxtaposition to thinner crust/lithosphere; i.e. along formerly active rifts (Japsen *et al*. 2006, 2012a; Osmundsen & Redfield 2011). This condition occurs landward of where continental crust starts to thin as rift basins towards oceanic crust (for example in Brazil, south-east Australia and Scandinavia). But it also occurs where thicker cratonic crust adjoins thinner extended continental crust within a continent, such as in southern Sweden where 40 km thick crust under the South Swedish Dome is juxtaposed to 30 km thick crust on the south-western side of the Sorgenfrei–Tornquist Zone (Japsen *et al*. 2012a, fig. 8).

Redfield & Osmundsen (2013) also suggested that one single uplift event cannot have created Scandinavia's stepped topography. They found the evidence to be more consistent with a series of uplifts occurring episodically throughout the Cenozoic, each followed by a period of stability during which erosional surfaces were incised to a relatively low base level. In their view, the progressive and episodic tilting of the Norwegian hinterland caused slope-dependent landforms to become rejuvenated, reinforcing or overprinting the preceding landscape until a new equilibrium became established. Redfield & Osmundsen (2013) did not consider the possibility that the Norwegian margin has been buried since Late Jurassic rifting. There is, however, evidence that sediment accumulated over the present onshore areas in Jurassic times and possibly later; VR values for erratic blocks from a Jurassic basin in a mid-Norwegian fjord suggest exhumation of the basin from below a cover of *c.* 2 km (Sommeruga & Bøe 2002). This observation is in agreement with the presence of re-exposed sub-Mesozoic relief on the mid-Norwegian coast (Fig. 7).

The episodic nature of vertical movements at EPCMs presented here helps to clarify various issues that have long been debated. For example, in southern Africa Burke and Gunnell (2008), Tinker *et al.* (2008a) and Paton (2012) have all highlighted an apparent conflict between evidence of Cretaceous 'uplift' defined from thermochronology and evidence for Cenozoic 'uplift' provided by more conventional methods, as discussed in chapter 6. A period of Cretaceous exhumation (leading to formation of low-relief surfaces across much of southern Africa) followed by Cenozoic uplift with only limited accompanying or subsequent erosion (taking the surfaces to their present elevations) provides a plausible explanation of the modern-day topography, and there need be no conflict between the rival interpretations (see section 6.2). This discussion highlights occasional confusion in the literature in understanding the nature of events revealed by low-temperature thermochronology, and that a minimum of two post-rift uplift events are needed to shape the present-day relief with elevated plateaux cut by deep valleys.

In summary, the development of EPCMs involves a series of positive and negative post-rift, vertical movements, in which a considerable amount of rock is both deposited and removed during the period subsequent to rifting and break-up.

The episodic development described here raises the question of the origin of the sedimentary cover required to bury a region by kilometre-scale thicknesses of cover. Given the nature of events described, a likely explanation is that while some areas are subsiding and accumulating sedimentary cover as a result of negative movements, other areas are being uplifted and eroded due to positive movements. Identification of such paired events remains a challenge for future definition of the nature of the events responsible for the episodic evolution of EPCMs.

A similar question regards the ultimate destination of the rocks eroded in such events. Several attempts have been made to compare the sediment budget in offshore basins with the denudation history of onshore margins (e.g. Campanile *et al.* 2008; Tinker *et al.* 2008b; Rouby *et al.* 2009; Anell *et al.* 2010). However, such analyses do not take into account the possibility that the offshore sedimentary basins may have themselves been eroded. In many of the cases discussed here, offshore basins have undergone significant amounts of denudation, although the resulting unconformities show little or no angular discordance, and are interpreted as representing periods of non-deposition. In addition, it is sometimes difficult to determine the ultimate resting place of eroded sediments, which can be transported over wide distances. Ager (1973) noted that at any one time, most of the sediment within a basin is moving laterally through the basin, rather than being retained. This being so, the question of mass balance between eroded sediment from a margin and thicknesses of sedimentary rocks in an adjacent basin seems far from simple.

8.8 Observation vs. theory in understanding the evolution of EPCMs

Stratigraphic landscape analysis is based on observations of landscapes and geological information. When integrated with information from thermochronology, the combined approach has led to a description of the evolution of EPCMs in terms of episodic burial and exhumation (subsidence and uplift) which is fundamentally different to accepted ideas of EPCM evolution, as illustrated in preceding chapters. Yet this new description continues to draw criticism from many quarters (Nielsen *et al.* 2009; Redfield 2010; Pedersen *et al.* 2012, 2013), largely on the grounds that no mechanism is available by which this style of development can be explained.

In several areas of geology over the last 150 years, conclusions drawn from basic observations have been denied because theory 'proves' that those conclusions cannot be valid. Examples include the occurrence of past ice ages, deep geological time, and continental drift. Micro landforms, such as striae, puzzled researchers for many years, but their observations demanded an explanation, which was ultimately obtained with the theory of the ice age. The value of 100 million years (later 20 million years, Stacey 2000) for the age of the Earth derived by the future Lord Kelvin (Thomson 1863) for many years constituted a major problem for acceptance of natural selection (England *et al.* 2007). Macro landforms such as the shape of the continents gave the idea of continental drift. The concept was supported by fossils on both sides of the Atlantic which belonged to the same type of environments while nearby fossil communities were different, as well as numerous similarities in the geology of the two continents across the ocean. As such the process was clear to southern hemisphere geologists (e.g. Du Toit 1937). But geophysicists (mainly located in the northern hemisphere) argued that continental drift was physically impossible. Such attitudes provided a major obstacle to understanding, instead of initiating a search for the underlying processes.

Acceptance of ideas for which evidence had become overwhelming was delayed for many years, until ultimately the underlying mechanism became clear. In the case of continental drift, it is ironic that the mechanism was not discovered as a result of a careful programme of investigation in an attempt to discover the mechanism, but only through a serendipitous combination of circumstances (Allègre 1988). Similar comments apply to the recognition of the role of radioactivity in providing a longer geological timesecale and revealing the depth of geological time.

All the above examples are characterised by a failure to recognise that accepted theories no longer provided an explanation of available data, and that new ideas were required. In the case of the evolution of EPCMs, evidence has been presented in previous chapters which leads to the conclusion that prevailing notions are no longer acceptable, and a paradigm shift is required. Yet studies involving monotonic cooling and continuous denudation continue to appear, the simplistic models shown in Fig. 57 continue to be popular (Bishop 2007; Campanile *et al.* 2008; Burke & Gunnell 2008), and the idea of removal of a significant thickness of cover continues to be regarded as unrealistic (e.g. Peulvast *et al.* 2008). Numerical modelling techniques have advanced in recent years. The notion of the continuous elevation of EPCMs, however, remains as a foundation (e.g. Sacek *et al.* 2012), despite abundant evidence to the contrary (e.g. Cobbold *et al.* 2001).

In an influential paper on 'Geomorphology as a science: the role of theory' Rhoads & Thorn (1993, p. 287) define science as an activity that "seeks to discover knowledge through a two-stage activity involving the creation and justification of ideas (theory)". They go on to argue that "the primacy of observations in science is a myth and that all observations are theory-laden in the sense that the act of observation inherently involves interpretation and classification". While this is true, it is also clear that if all observations are interpreted purely in terms of accepted theories, then no new discoveries will ever be made, so the mind must remain open to new concepts when existing models fail. In the examples cited above involving ice ages, the age of the Earth and continental drift, adherence to existing theories constituted an obstacle to progress.

A wealth of information shows that traditional theories for the development of EPCMS are no longer satisfactory. It is a first-order observation that high-level regions of low relief characterise EPCMs. Such regions of low relief, cutting across rocks of different age and resistance, can only be created by erosion to base level, and since EPCMs are formed adjacent to oceans, in the absence of resistant lithologies over wide areas, the most likely base level is sea level. As shown here, EPCM landscapes demonstrate an episodic evolution involving a series of positive and negative vertical movements. However, instead of seeking models and theories that honour these observations, their significance is denied because accepted theories cannot explain them. This argument therefore seems to be back to front. New theories are required which attempt to explain the wealth of observations pointing to the episodic upward and downward movements of EPCMs, as well as the significant amounts of rock that are deposited and removed following continental break-up before the present-day EPCM landscapes were developed.

9. Summary and conclusions

The elevated, passive continental margin (EPCM) of West Greenland developed through a series of vertical movements (both positive and negative) following continental break-up (chapter 5). The topography of the West Greenland margin shares many characteristics with other EPCMs (elevated plateaux 1–2 km a.s.l. with deeply incised valleys, gently declining inland and a much steeper oceanward slope terminating at a coastal plain) suggesting a common overall style of development. However, the still prevailing view of EPCM development is that they retain their elevation and characteristic landforms from break-up or even earlier. This contrasts starkly with the history of episodic subsidence/burial and uplift/denudation described for the West Greenland Margin in chapter 5.

Previous thermochronological studies favouring the long-term elevation of margins have been based on assumptions of progressive uplift/denudation and continual cooling. However, a review of geomorphological and thermochronological studies from other EPCMs (chapter 6) shows that these assumptions often conflict with geological evidence, and histories involving repeated episodes of burial and exhumation (heating and cooling), similar to that defined in West Greenland, are more realistic.

A wealth of evidence supports the conclusion that the elevated, low-relief regions that characterise EPCMs are uplifted peneplains originally graded to sea-level in the adjacent, opening ocean. The peneplains were subsequently uplifted to their present levels by movements that took place long after rifting. In many cases the peneplains were preserved prior to uplift by a sedimentary cover, which was subsequently removed during uplift. Following uplift the peneplains were further modified by valley incision and valley widening, sometimes in multiple episodes, leading to the development of additional peneplains at lower levels. This results in characteristic, stepped high surfaces and a low generation of incised valleys, sometimes coalescing to an escarpment. However, the description of the oceanward part of a margin in terms of a single Great Escarpment is over-simplified and most margins are more complex (chapter 8). Overall, the typical EPCM landscapes should be viewed as reflecting the integrated effects of their post-rift development, and are not related to the rifting process.

While the shared characteristics of EPCMs suggest that they are likely to have formed as a result of essentially similar processes, the form of each EPCM is different in detail, and the differences between margins can be used to provide conclusions regarding the specific development of each, using SLA in concert with geological evidence and thermochronology. While the nature of the controlling process(es) remains unclear, the regional extent of the vertical movements documented here suggests a plate-scale control, possibly related to compressive stresses resulting from distal orogenies, basal drag of the lithosphere by horizontal asthenospheric flow or from changes in plate motion (chapter 7). Such explanations may explain why some episodes are common to margins of divergent plates whereas some are not.

The results presented here provide a consistent body of evidence to demonstrate that EPCMs have moved vertically, both downwards (subsidence/burial) and upwards (uplift/exhumation) long after continental break-up. Definition of these movements in time and space is essential to a full understanding of the nature of the underlying processes involved in EPCM development. More fundamentally, identifying common aspects between different EPCMs has the potential to allow investigation of the properties of the lithosphere and mantle which control these vertical movements of rifted margins, long after continental break-up.

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