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Palaeo-digital elevation models for use as boundary conditions in coupled ocean–atmosphere GCM experiments: a Maastrichtian (late Cretaceous) example

Paul J. Markwick[a,*], Paul J. Valdes[b]

[a]Fugro Robertson Limited Research International, Tyn-y-coed Site, Llanrhos, Llandudno, LL30 1SA, UK
[b]School of Geographical Sciences, University of Bristol, University Road, Bristol, BS8 1SS

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Abstract

Palaeogeography (specifically palaeotopography and palaeobathymetry) provides an essential boundary condition for computer-based atmosphere and ocean modelling. It also provides the geographic context for understanding surface processes (palaeodrainage, palaeoweathering) and biotic interactions (palaeoecology, palaeobiogeography). With increased model resolution, coupled ocean–atmosphere general circulation models (GCMs) and the addition of vegetation, soil (weathering), ice and chemical modules, there is now a need for more robust, detailed palaeotopographies and palaeobathymetries that are fully integrated with the processes being modelled, especially the hydrological system.

Here, we present a new geographic information system (GIS)-based, hydrologically correct, palaeo-digital elevation model (DEM) for the Maastrichtian (late Cretaceous). We describe the methods and concepts used to construct the map, and draw attention to the limits imposed by scale and uncertainty, and how these factors must be considered as part of the error analysis of derived model results. The underlying palaeogeography is one of a series of 27 global maps that represent the stages of the Cretaceous and sub-epochs of the Cenozoic. Each map is generated at a scale of 1:30 million in ArcView® GIS and ArcInfo™, using data from the lead author's own databases of lithologic, tectonic and fossil information, the lithologic databases of the Paleogeographic Atlas Project (The University of Chicago), a survey of published literature, and DSDP/ODP data. Interpretations of elevation are derived following the methods outlined in Ziegler et al. (Ziegler, A.M., Rowley, D.B., Lottes, A.L., Sahagian, D.L., Hulver, M.L., Gierlowski, T.C., 1985. Paleogeographic interpretation: with an example from the Mid-Cretaceous. Annual Review of Earth and Planetary Sciences, 13: 385–425), an understanding of the tectonic regime and evolution of each feature, and the age–depth relationship for the ocean. The global palaeo-DEM was derived using the elevation contours from the paleogeography and the suite of hydrological tools available in ArcInfo™ GRID. The palaeo-DEM has been constrained by defining areas of internal palaeodrainage, palaeoriver mouths and known palaeoriver courses. The Maastrichtian has been completed first to provide the boundary conditions for a series of coupled atmosphere–ocean experiments.

* Corresponding author.
E-mail addresses: paul@markwick001.freeserve.co.uk (P.J. Markwick), p.j.valdes@bristol.ac.uk (P.J. Valdes).

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When integrated with the results of the coupled ocean–atmosphere model, the result is a powerful tool for understanding surface processes and an important step towards the development of a continuous series representing a fully evolving palaeolandscape.

Keywords: Palaeogeography; Geographic information system; Cretaceous; Digital elevation model; Drainage analysis; Palaeoclimatology

1. Introduction

Modern geography has long been recognised to exert an important influence on climate (Hopkins, 1842; Humboldt, 1820, 1828, 1845). This led Lyell, in the early 19th century, to postulate that changes in land-sea distributions and orography over geological time could account for the large-scale changes in palaeoclimate indicated by the fossil record (Lyell, 1830, 1872). This has remained a central tenet in palaeoclimatology (Croll, 1875; Damon, 1968; Deeley, 1919; Donn and Shaw, 1977; Frakes, 1979). However, it was not until the publication of global plate reconstruction maps in the 1970s (Scotese et al., 1979; Smith and Briden, 1977; Tarling, 1977) that global palaeogeography (the science of representing the Earth's past geographic features) could be placed in the context of plate tectonics, which provided the scale of continental changes postulated by Lyell in a series of thought experiments (see Fig. 1 p. 130 in Lyell, 1837). A series of global palaeogeographic atlases consequently appeared in the late 1970s and early 1980s (Barron et al., 1981; Ronov et al., 1984; 1989; Ziegler et al., 1979, 1983) that enabled two important developments: (1) the use of general circulation models (GCMs) to investigate pre-Pleistocene palaeoclimate (Barron, 1981, 1983, 1984; Barron and Washington, 1982; Barron et al., 1980), for which reliable global palaeogeographies were essential boundary conditions; (2) the application of both palaeoclimatology and palaeogeography to frontier exploration in the oil, gas and mineral industries (Parrish and Barron, 1986), which provided much of the subsequent motivation for palaeogeographic research.

Most published global palaeogeographies depict only palaeoshorelines and rudimentary "highlands" (Barron et al., 1981; Smith et al., 1994; Ziegler et al., 1979, 1983). In the 1980s, these appeared more than adequate as the boundary conditions for the conceptual models that were being used to predict the large-scale pattern of coastal upwelling (Miller, 1989; Parrish, 1982; Parrish and Curtis, 1982; Parrish et al., 1983; Scotese and Summerhayes, 1986). However, GCM experiments investigating the role of geography in driving climate change (an underlying assumption of conceptual models: Gyllenhaal et al., 1991; Parrish and Curtis, 1982; Patzkowsky et al., 1991) were unable to entirely replicate either the warm temperatures of the Cretaceous, or the Tertiary cooling trend, by changing large-scale land-sea distributions alone (Barron, 1985; Barron and Washington, 1982; Barron and Washington, 1984; Barron et al., 1980). This led Barron to speculate that another forcing factor was required, in addition to geographic change, of which changing atmospheric CO_2 concentrations was the most attractive (Barron, 1985; Barron and Washington, 1985; Barron et al., 1993; Chamberlin, 1898, 1899a,b,c). This is something that conceptual models cannot account for.

The central role of atmospheric CO_2 in explaining the palaeoclimatological record has been examined by many subsequent studies (Barron et al., 1993, 1994; Manabe and Bryan, 1985; Rind, 1987; Schlesinger and Mitchell, 1987; Sloan and Rea, 1995), not least because of its importance in predicting the direction and nature of future climate change (Houghton et al., 2001). CO_2, as a greenhouse gas, affects temperature, especially ocean temperatures, and this in turn influences both the global atmospheric and oceanic circulation, and the hydrological system (Manabe and Stouffer, 1980; Manabe and Bryan, 1985; Markwick et al., 2002; Otto-Bliesner, 1995). But, exactly how these effects are then partitioned on the Earth's surface is still a function of the contemporary geography; therefore, palaeogeography must be fully understood, regardless of the importance of atmospheric CO_2 as the driver of climate change.

Orography, for example, can have a major influence on the distribution of surface climate (Ruddiman and Prell, 1997). Uplands are cooler than their surrounding lowlands (lapse rate), can deflect air-flow and dictate precipitation patterns (Fig. 1): the magnitude of the effects depending on the dimensions and location of the uplift (Kutzbach et al., 1989; Raymo and Ruddiman, 1992; Ruddiman and Kutzbach, 1989; Ruddiman et al., 1989). Indeed, orography itself may directly dictate atmospheric CO_2 concentrations through silicate weathering (Berner et al., 1983; Chamberlin, 1898, 1899a,b,c). This has been used to explain the observed Tertiary cooling trend (and glaciation of Antarctica), as weathering of uplifted areas, such as Tibet, results in the draw-down of CO_2 (Berner et al., 1983; Raymo, 1991; Raymo and Ruddiman, 1992; Raymo et al., 1988; Walker et al., 1981).

DeConto and Pollard (2003) have recently shown how modelled decreases in atmospheric CO_2 are needed to generate Antarctic ice-sheets, but that the exact timing of initiation depends on the geography of ocean gateways in the Southern Ocean. Barron and Frakes (1990) showed how precipitation variability could be influenced by the geometry of shorelines and how this in turn might affect upwelling systems. Valdes (1993) and Valdes et al. (1996) demonstrated the climatic effect of seaways through continental regions, both as conduits for poleward heat transport and also as important moisture sources. Sloan showed how a moderately large lake system, such as the Green River lakes of the Eocene of the western US, could ameliorate regional climate and needed to be included in boundary conditions (Sloan, 1994). Bice et al. (2000) showed how the presence or absence of land-bridges in the North Atlantic could influence Palaeogene ocean heat transport, and also stressed the importance of well constrained palaeobathymetries (Bice et al., 1998). Horrell (1991), Otto-Bliesner and Upchurch (1997) and Upchurch et al. (1998) have highlighted the importance of prescribing vegetation in climate models by showing how a vegetated Antarctica maintains its warmth. While these studies addressed individual geographic issues, Barron (1985) postulated that regional-scale, or even local-scale, changes in geography could have far greater climatic effects if they occur consecutively or in parallel.

As GCMs, and the questions asked of them, have become more complex, currently integrating atmosphere and ocean circulation, as well as dynamic surface schemes (hydrology, vegetation, soils, ice), the need to be able to prescribe the details of the geography and other surface features has become more important. The Hadley Centre coupled ocean–atmosphere model, HadCM3 (Gordon et al., 2000; Pope et al., 2000), for which the current palae-

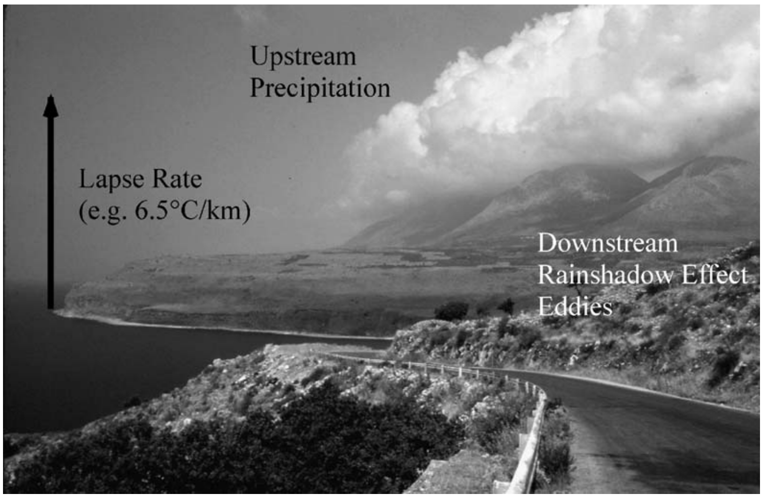

Fig. 1. The influence of orography on local and regional scales. Here, orographic clouds are shown along the upstream side of the Taygetos Mountains, Greece. At 1000 m, the temperatures at the top of these hills are cooler than at sea-level due to the lapse rate (on average this is taken to be about 6.5°/km, Barry and Chorley, 1987).

ogeography has been constructed, is uncommon among the current generation of coupled atmosphere–ocean models in that it explicitly requires outfall points for terrestrial drainage to be defined (Cox et al., 1999). This in turn necessitates an understanding of the hinterland geography and palaeodrainage. The flux of freshwater to the adjacent oceans has major implications for palaeoceanography, including the possible distribution of deep water formation (Bice et al., 1997) and salinity stratification and nutrient fluxes that may influence the formation of organic-rich mudstones (Jenkins and Williams, 1983; Krom et al., 2002; Rohling and Hilgen, 1991; Rossignol-Strick et al., 1982). HadCM3 also includes a fully interactive vegetation scheme, TRIFFID (Cox, 2000, 2001). These additions provide the opportunity to investigate the evolution of the palaeolandscape, including weathering rates and freshwater and sediment fluxes (Bogaart et al., 2002; De Roo, 1998; Hovius, 1998, 2000; Richards, 2002), but also mean that the palaeogeography must be better understood. These are also issues that have become of increasing interest to explorationists in the mineral, oil and gas industries, for whom the qualitative prediction of coastal upwelling is no longer enough, with attention now focussed on quantitative basin-scale predictions of all the components necessary for source, reservoir and seal facies.

The correct representation of geography, especially orography (and its cover), is therefore critical to climate modelling. The problem for palaeogeographers is to be able to robustly reconstruct these geographic features, which is far from easy. Lakes, for example, even well-documented ones such as those of the Eocene Green River, are generally small and fluctuate over time (Bradley, 1929; Fischer and Roberts, 1991; Franczyk et al., 1992; Roehler, 1992), such that a palaeogeographic map for a specific ‘time-slice’ may encompass numerous fluctuations. Large-scale features, such as seaways should be more robustly definable, but Ziegler and Rowley have pointed out that the extent of such seaways may have been seriously underestimated in past reconstructions (Ziegler and Rowley, 1998). Orography, in particular, is problematic since it represents areas above erosional base-level and so direct evidence is sparse in the geological record. The coarse resolution of the models themselves also poses a problem for representing palaeogeography, with mountains in early experiments being parameterised as broad curved features that could obscure internal relief changes and associated variations in climate. This may have been partly responsible for the apparent mismatch between Palaeogene mid-latitude continental interior climates interpreted from climate proxies, and those indicated by GCM experiments, which led to considerable debate in the early 1990s (Sloan and Barron, 1990, 1991, 1992; Wing, 1991).

The aim of this study is to present a new detailed palaeogeography of the Maastrichtian, as a GIS-based palaeo-DEM, for use with the HadCM3-coupled ocean atmosphere model (Markwick et al., 2002). To compile palaeogeographies to the detail required by climate modellers and modern exploration geologists, requires comprehensive databases and an understanding of the heterogeneities and scaling issues that permeate the geological record (Markwick and Lupia, 2001; Ziegler et al., 1985). We discuss these issues, while also illustrating the concepts and methods used to construct the palaeogeography (especially the generation of palaeodrainage) and the limitations in any palaeogeographic reconstruction that must be understood when model results are ‘validated’. Some of these issues are unresolvable (for instance due to the absence of any rock record) and comprise the analytical error that must be associated with every palaeogeographic map. The effect of these ‘errors’ on climate can be investigated through modelling sensitivity tests in which the palaeogeography is modified for each experiment to reflect the range of uncertainties. This process has been greatly facilitated by the development of GIS, which is utilized at every stage throughout this study.

2. Concepts

2.1. What is a time-slice?

Ideally, a palaeogeographic map should represent the geography at a specific moment in time, a “time-plane”. But, even for small geographic regions, this is rarely possible because of the heterogeneous nature of the geological record. As the spatial extent of the study increases (“extent” refers to the total spatial and/or temporal limits of the study: Markwick and Lupia, 2001), the uncertainty in dating and the temporal grain (the grain describes the minimum resolution/scale of an observation: Markwick and Lupia, 2001) increase, though not uniformly (Fig. 2a). Limiting the data on which the map is based to only temporally fine-grained information may improve temporal resolution, but at the cost of spatial coverage (Fig. 2b). The key is to balance the grain (precision) requirements, against data coverage (extent), and this will depend on the application of the resulting map (Markwick and Lupia, 2001).

Consequently, most palaeogeographic maps, especially global reconstructions, represent a "time-slice", which is usually a chronostratigraphic stage or sub-

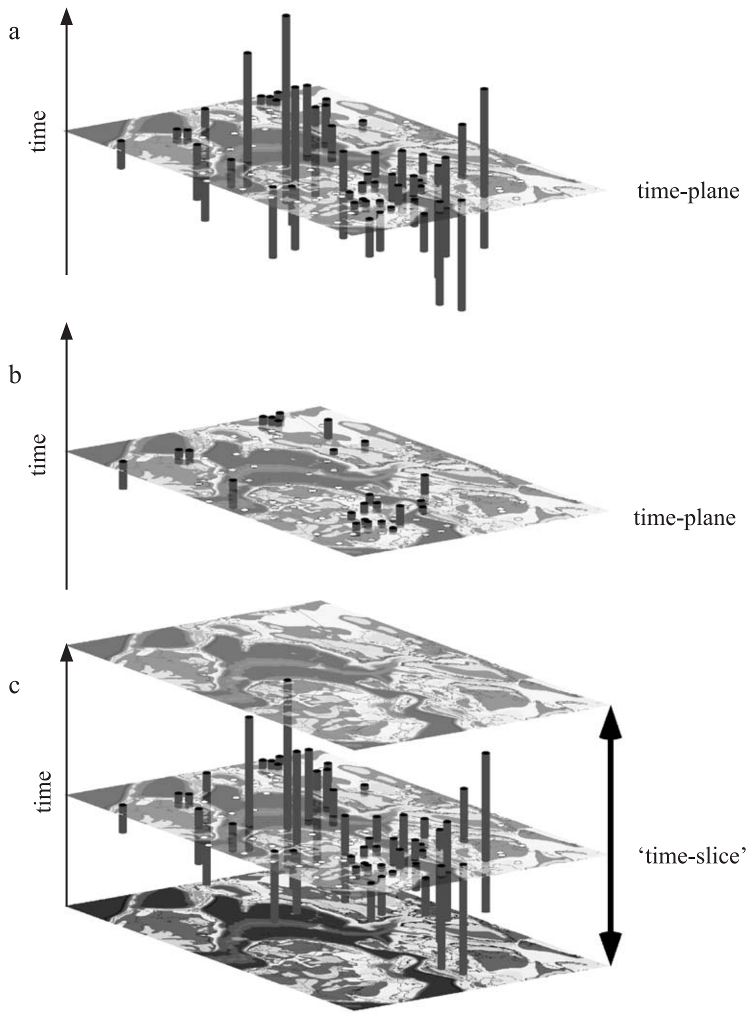

Fig. 2. (a) Palaeogeographic maps comprise data of varying temporal and spatial grains (indicated by the length of the cylinders). Here, the actual position of the time-plane is indicated along the time axis of each data point. The problem is that in reality the exact position of the "time-plane" is rarely known at any one locality, due either to imprecise (or inaccurate) dating (analytical time-averaging: Behrensmeyer and Hook, 1992) or physical mixing (taphonomic time-averaging: Behrensmeyer and Hook, 1992; Behrensmeyer and Chapman, 1993). The length of the cylinder in the figure represents the temporal time-averaging of each datapoint (whether it be a stratigraphic section, well sample or core, etc.). (b) One way to deal with this is to limit data to only information that is temporally well constrained (fine grain), but this invariably results in a loss of information as shown, leaving large areas of the map unconstrained. (c) The most parsimonious solution is to specify the grain to maximise the data density, while minimising the uncertainty, which will depend on the problem being addressed with the map. The result is a "time-slice".

Epoch. This defines the temporal grain of the map. In reality, this "time-slice" may not actually represent an Earth that ever existed, but is a concatenation, or pastiche, of regional reconstructions in time and space that all fall within the bounds of the "time-slice", and which therefore comprise varying amounts of time-averaging (Fig. 2c). The stages of the Cretaceous, for example, vary in duration from 2.3 (Santonian) to 13.3 myr (Albian) (Gradstein and Ogg, 1996) and so a "time-slice" map of a Cretaceous stage may comprise considerable facies variations in any one place. This is especially true in areas of major tectonic activity, areas receiving rapid sediment flux and/or during times of rapid eustatic sea-level changes. Ziegler (1978, 1990) used "transitional" to represent areas on the map with a mix of marine and terrestrial conditions, which could therefore potentially represent the area over which the shoreline might have existed. But, this is heavily biased by the preserved record and gives little indication of how far landward a transgression may have extended, for which there is now no record. Nor does it give an indication of coincidence (or otherwise) of maximum or minimum transgression between different areas. Ziegler et al., (1983, 1985) tried to deal with this by recording the environment at the beginning and end of a time-slice in their databases. Nonetheless, their maps generally represented the maximum transgression for most of their time-slices for the Mesozoic and Cenozoic, since much of their early interest was driven by the distribution of marine source rocks. Such a decision has two major consequences: (1) it assumes that the maximum extent of the sea in all regions within a time-slice was coeval, which is unlikely to be true in a non-glacial world (see Markwick and Rowley, 1998); (2) it typically biases climate model results to a warmer and wetter climate regime, because of the larger ocean surface area available to store energy and act as a moisture source.

The definition of a "time-slice" therefore has major implications when using palaeogeographies as boundary conditions for climate and ocean models, which by definition are attempting to replicate a 'climate' (in modelling terms, the average atmospheric state over several model decades). Since this uncertainty also applies to the proxy data used to evaluate the model results (Markwick, 1998; Markwick and Lupia, 2001), care must be taken in accounting for the uncertainties involved. It is therefore important to specify the grain used for the palaeogeographic map (stored as attributes in GIS). It is also important to know how the palaeogeographer has treated facies variability within the time-slice.

In this study, we use a plate reconstruction representative of the late Maastrichtian: chron 31r; 70 Ma (Berggren et al., 1995) and 69.4 Ma (Gradstein and Ogg, 1996). Onto this map is plotted data of "Maastrichtian" age (71.3–65.0 Ma: Gradstein and Ogg, 1996). This is recorded as an attribute in the GIS for each datapoint. The 'operational' temporal grain of our map is therefore 6.3 Myr.

2.2. Regional base-level and depositional versus non-depositional systems

The palaeotopographies and palaeobathymetries depicted on palaeogeographic maps, such as those in this study, are derived from data initially compiled as palaeoenvironmental maps (Ziegler et al., 1985). At their simplest, palaeoenvironmental maps depict two major regimes (Fig. 3): contemporary depositional systems (whether subsequently preserved or eroded), which can be further divided into their respective environmental associations (e.g. deltaic, fluvial, shallow marine, etc.), and areas of net erosion (Mallory, 1972; Ronov et al., 1984; 1989; Totterdell et al., 2001; Vinogradov et al., 1967, 1968; Ziegler, 1978, 1990; Ziegler et al., 2001; Ziegler et al., 1985). Conceptually, these areas are separated by contemporary base-level (Fig. 3b) (Barrell, 1917; Wheeler, 1964). Base-level (sensu Wheeler, 1964) is important to palaeogeographers for understanding the relative distribution of contemporary relief. It is also important as it pertains to defining palaeodrainage systems (Markwick, 2001; Markwick et al., 2002), since base-level will equate to the 'equilibrium' long profile of a drainage system. As recognised by Wheeler (1964), base-level, as a surface, changes as the Earth's surface changes (at the scale of interest), but is nonetheless useful for a 'static' ('snapshot') reconstruction such as a palaeogeographic map of a specified 'time-slice'.

2.3. Uniformitariansim

The basic concept behind the topographic and bathymetric mapping, as outlined in Ziegler et al.

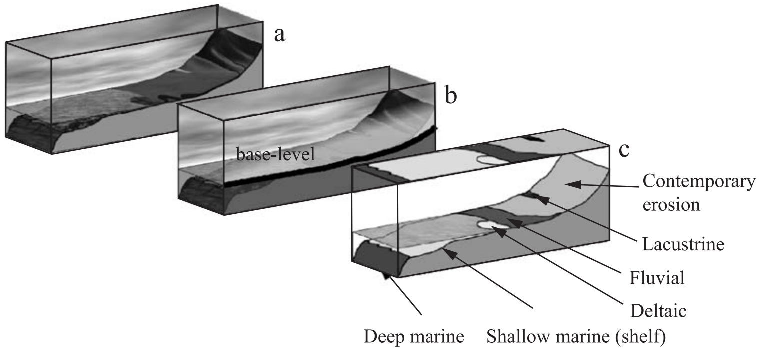

Fig. 3. The contemporary topography and bathymetry (a) can be divided into areas above and below regional base-level, which dictates the areas of contemporary deposition and erosion (b). This is represented by a palaeoenvironmental map (c) such as that of Ziegler (1990). In reality, the construction of a palaeogeographic map would progress from (c) to (a).

(1985), is that the present relationships between environmental and tectonic settings and their elevational and physiographic representation, can be applied to the past—Uniformitarianism. This is probably generally true, although care must be taken where successive tectonic events/regimes are superimposed temporally, adding complexity to the elevational (or bathymetric) signature. This emphasises the need to examine preceding and subsequent time-slices—a single geography can not be reconstructed in isolation.

2.4. Continuity of geography

Although the current study presents only a Maastrichtian palaeogeography, this is one of a series of global palaeogeographic maps spanning the Cretaceous and Cenozoic (Markwick et al., 2000). Understanding the evolution of depositional and tectonic features influences how relief, base-level and drainage are portrayed on subsequent maps. A mountain range, for example, has a longevity (once formed, orography cannot disappear 'overnight' without some definable reason) and will erode at a rate(s) depending on the underlying rock characteristics, slope, vegetation and climate, until some other specified tectonic regime (or climate change) acts upon it. Subsequent palaeogeographic maps must make historical sense—map continuity. In a fully evolving landscape model, on the temporal scale of millions of years (our ultimate aim), this should be possible through equations to govern each cell in the DEM. However, to achieve this here, all the maps have been compiled simultaneously and changes made iteratively to each in order to maintain continuity. One consequence of this is that more substantive changes may become apparent, such as the necessity for modifications to the plate reconstruction model, but given the limits of this study, these changes will be incorporated in future versions of the maps.

3. Methods

The Maastrichtian palaeogeography, used in this study to generate the palaeo-DEM, is 1 of a series of 27 Cretaceous and Cenozoic time-slice maps (Markwick, 1996; Markwick et al., 2000). These drew heavily on the work and datasets of Ziegler and his colleagues in the Paleogeographic Atlas Project (Hulver et al., 1993; Ziegler and Rowley, 1998; Ziegler et al., 1983, 1996, 1984, 1985, 1987). These maps have been used in a number of subsequent palaeoclimate and palaeontological studies. In the following sections, we describe the information included on each base map upon which palaeogeography is drawn, outlining the data we have included digitally in the GIS. We then summarize the additions and modifications we have made to the Maastrichtian palaeotopography and palaeobathymetry compared with the palaeotopographic map of Ziegler and Rowley (1998), which formed the starting point for the current map. Finally, we describe the method by which the contoured palaeotopographic map has been converted to a palaeo-DEM.

3.1. Base maps

The base maps include the geological data from which the topographies and bathymetries are interpreted (as points, lines and polygons), together with modern geographic information to help position data correctly in space. The original 1:100,000,000 scale base maps (Markwick, 1996) comprised reconstructed plate outlines, a simplified coastline and modern 5° grid, all rotated to their appropriate palaeo-positions, with data from the Paleogeographic Atlas Projects' databases for those time-slices that coincided with those of Markwick (1996, 1998). Following Ziegler et al. (1985) and Markwick and Lupia (2001), we have been careful to distinguish between data and interpretations.

In 1997, the palaeogeographies were redrawn onto new base maps at a scale of 1:30,000,000 and captured digitally in ArcInfo™ GIS. Polar orthographic projections were used for the regions from the poles to 60° latitude, and the Mollweide projection for the equatorial region between 60°S and 60°N (Markwick et al., 2000). These base maps comprise a more extensive dataset of rotated data including the following information for all of the stages of the Cretaceous and sub-epochs of the Cenozoic: a detailed coastline (Fig. 4a); country, province and state boundaries (Fig. 4b); modern bathymetric and topographic features (for the Maastrichtian these modern day features incorporate a 100-m sea-level rise commensurate with late Cretaceous values) (Fig. 4c); detailed modern rivers (Fig. 4d); modern 1° grid attributed in the GIS to differentiate between extant continental and oceanic areas (Fig. 4e).

In order to construct the palaeogeographies, the following information was added to the base maps: the environmental data from the Palaeogeographic Atlas Project (Fig. 5a); lithological, faunal and floral data from the authors own databases (Markwick, 1996) (Fig. 5a); all available DSDP and ODP data (Fig. 5b); major tectonic faults; and a synthetic isochron dataset (Fig. 5c), which automatically defines the palaeo-transform faults and which is described in more detail below. Again, all of this information is stored, attributed and referenced within the GIS.

In this study, we have preferred to compile directly onto plate reconstructions rather than onto present day

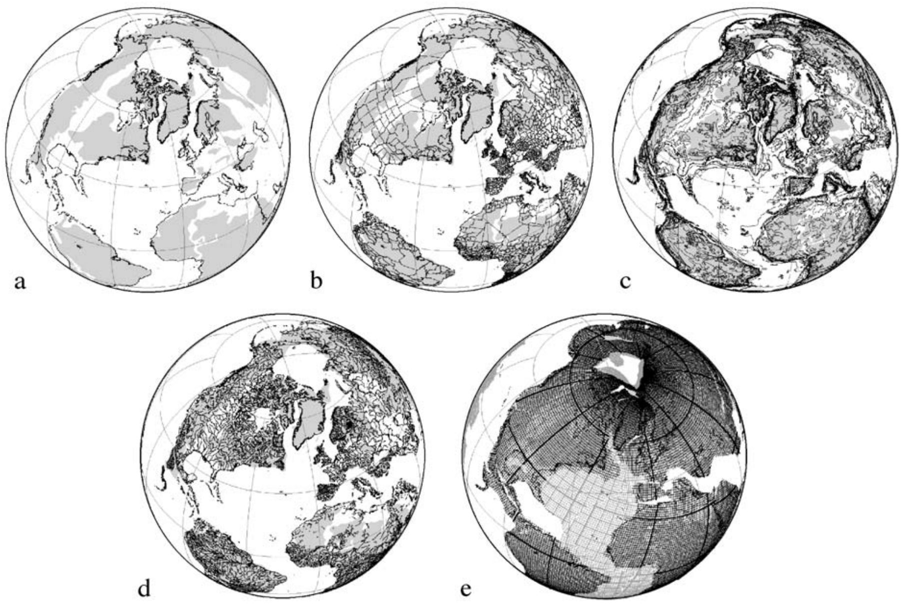

Fig. 4. The revised base-map (shown as an orthographic projection: central meridian −15°, reference latitude, +45°) generated in GIS. (a) detailed modern coastline; (b) administrative regions; (c) modern elevational data for the palaeogeographic map contour intervals, with account taken of the approximately 100 m sea-level change; (d) modern (mapped) rivers; (e) modern 1° grid differentiated into continental and oceanic regions.

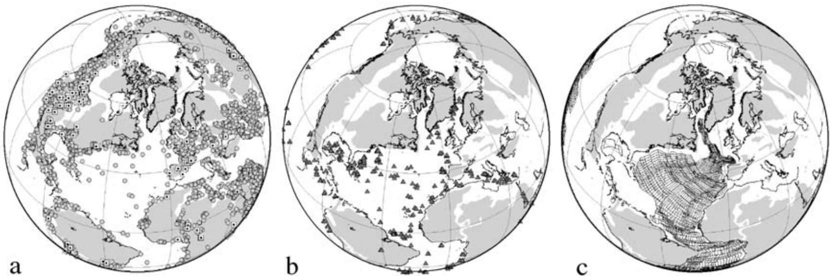

Fig. 5. Data included on the base maps: (a) lithological, palaeontological and environmental data from the databases of the Paleogeographic Atlas Project (Ziegler and Rowley, pers. comm.), grey circles, and Markwick, squares (Markwick, 1996); (b) DSDP and ODP data for ocean crust extant at the time of time-slice mapped; (c) synthetic isochrons for 5 myr increments based on the digital gridded ocean-age dataset of Müller et al. (1997).

maps, since this gives an immediate indication of the juxtaposition of features. The plate reconstructions used are those of Rowley (1995, pers. comm.). Plate reconstructions are constantly in a state of flux and invariably the process of compiling a palaeogeography, and especially its tectonic component, will suggest modifications. This is true here. However, for the Maastrichtian map, these modifications are minor in comparison with the magnitude of global geographic differences between the Maastrichtian and present day, and the effects of these differences on climate (especially given the operational resolution of the GCM, which is on the order of hundreds of kilometres: $3.75^\circ \times 2.5^\circ$ for the atmosphere, $1^\circ \times 1^\circ$ for the oceans). For the Tertiary, the details of contentious, fine-scale tectonic problems, such as the timing of ocean gateways, are important and do need to be robustly modelled, but these are not issues in the Maastrichtian. Consequently, these refinements to the plate reconstructions will be included in the next generation of maps.

Because the map compilation scale is at 1:30,000,000, this is the effective ('operational') scale for all further analysis.

3.2. Palaeotopography

Although many palaeogeographers have attempted to differentiate the erosional areas on palaeoenvironmental maps on the basis of the underlying tectonic regime (Mallory, 1972; Ronov et al., 1984, 1989; Vinogradov et al., 1967, 1968; Ziegler, 1978, 1990; Ziegler et al., 1985, 2001), only Ziegler et al. (1985) have attempted to translate this into a topography with defined elevation contours (Ziegler et al., 1996). Ziegler et al. (1985) developed systematic methodologies for reconstructing past orography based on analogy with the elevations of similar modern tectonic regimes (Ziegler et al., 1979, 1983, 1996); a method that is followed in this study. Resulting reconstructions can then be refined with data such as apatite fission track, maturity analysis and the mass-balance of eroded sediment (Hulver, 1997). Uncertainties in the exact position of palaeo-elevational contours still exist, and will vary depending on the quality and quantity of the data and underlying tectonic models. These uncertainties can be further minimised through iteratively evolving elevations through time by comparison with events in adjacent time-slices and with the ultimate (and only) elevational tie-point—the modern day.

The Maastrichtian topographic contour map, used to construct the terrestrial DEM in this study, is based largely on the palaeo-topographic map of Ziegler and Rowley (1998). We have made changes in a number of places, reflecting either consistency issues with adjacent time-slices, or additional data, which we outline below. For a fuller discussion, readers are directed to Ziegler and Rowley (1998).

3.2.1. Antarctica

For Antarctica, we have used a digitised version of Drewry's isostatically corrected map of the Antarctic bed rock surface (Drewry, 1983) as our starting point. This has then been modified to reflect the limited available Cretaceous data and the latest tectonic models. To this end, the Antarctic peninsular forms a continuous arc along the line of the subducting

Phoenix Plate as it does throughout much of the Cretaceous and Tertiary (Barker, 1982; Dingle and Lavelle, 2000; Vaughan and Storey, 1997). Marie Byrd Land was probably relatively low-lying throughout the late Cretaceous prior to the volcanic episode of the middle Tertiary (LeMasurier and Rex, 1982; 1991; LeMasurier and Landis, 1997). The palaeotopographic relief along the line of the Transantarctic Mountains is more complex. We show a relatively subdued line of mountains that reflect the relict of probable Jurassic uplift related to extension and associated silicic and basaltic magmatism: the Ferrar volcanics (Elliott, 1992). It is generally assumed that most of the uplift of the Transantarctic Mountains evident today is a late Palaeogene phenomena (Behrendt and Cooper, 1991; Busetti et al., 1999), but Fitzgerald and Stump (1997) have presented evidence for late Cretaceous uplift and denudation associated with a phase of extension in the Ross Sea. The Gamburtsev Mountains are a major orographic feature comprising an area of ca. 435,000 km^2 over 2000 m (an area comparable to that elevation over 2000 m in the US Rockies from Wyoming, Colorado and New Mexico). Veevers (1995) has suggested that this is a late Palaeozoic feature, but in the absence of any drilled data, any age assignment must be speculative. We have assumed that it is already in existence by the Maastrichtian and that it may have had an analogous tectonic history to the Hoggar Plateau of North Africa, which is a positive feature in the Palaeozoic that is subsequently reactivated at various times during the Mesozoic and Tertiary in response to changing stress regimes and mantle heterogeneities.

3.2.2. New Zealand and Australia

New Zealand has been totally redrawn following the work of King and colleagues (Hayward et al., 1989; King, 2000a,b; King et al., 1999). For Australia, we have based the map on that of the BMR palaeogeographic Atlas (BMR Palaeogeographic Group, 1990; Totterdell et al., 2001).

3.2.3. South America

We have made only three changes to Ziegler and Rowley's map of South America: (1) small alterations to the contours of NE Brazil reflecting our opinion that at least part of this uplift is Tertiary; (2) modifications around the Parecis Basin, where coarse clastics in the Late Cretaceous Parecis Formation suggest adjacent uplift (Petri and Fulfaro, 1981; Siqueira, 1989); (3) a decrease in the extent of uplift associated with the incised Guyana Shield making it more consistent with our views on its Tertiary history.

3.2.4. Africa

We have added relief to the Guinea Highlands and intermittent uplifts along the coast from Sierra Leone to Nigeria (Kwahu Plateau, Akwapin Togo Range and Plateau of Yurubaland). This reflects our view of the reactivation history of these features along this margin during the Santonian–Maastrichtian (Bennett and Rusk, 2002), which are probably due to changes in the Atlantic poles of rotation and which are also associated with the inversion of many of the basins along the Central Africa Shear Zone. This probably also affected the northern outlet of the Congo River (from the Cuvette Centrale) into the Benue Trough. However, we show this palaeo-Congo still draining into Nigeria during the Maastrichtian, but acknowledge that this may have changed around this time. This northern route has been postulated for the Cretaceous palaeo-Congo River based on micro-diamond provenance (Censier, 1990) and current directions in the Carnot Sandstone Formation (Censier and Lang, 1999).

3.2.5. North America

We have followed Ziegler and Rowley's (1998) map of North America entirely. This shows a North American palaeogeography dominated by the remnants of the Western Interior and Hudson Bay seaways. These are bounded in the west by the consequences of the Laramide uplifts and in the east by the eroding Appalachian mountain belt.

3.2.6. Greenland

For Greenland, we have used a gridded dataset of modern ice thickness and bedrock depth (Douglas MacAyeal, pers. comm., The University of Chicago, see also Bamber et al., 2001). From this, we have calculated the isostatically corrected topography assuming simple Airy isostasy, but ignoring flexural responses (Fig. 6a–c). Modifications to this map were then made to remove the Tertiary uplifts (Birkelund and Håkansson, 1983; Boyd, 1993; Eldholm and

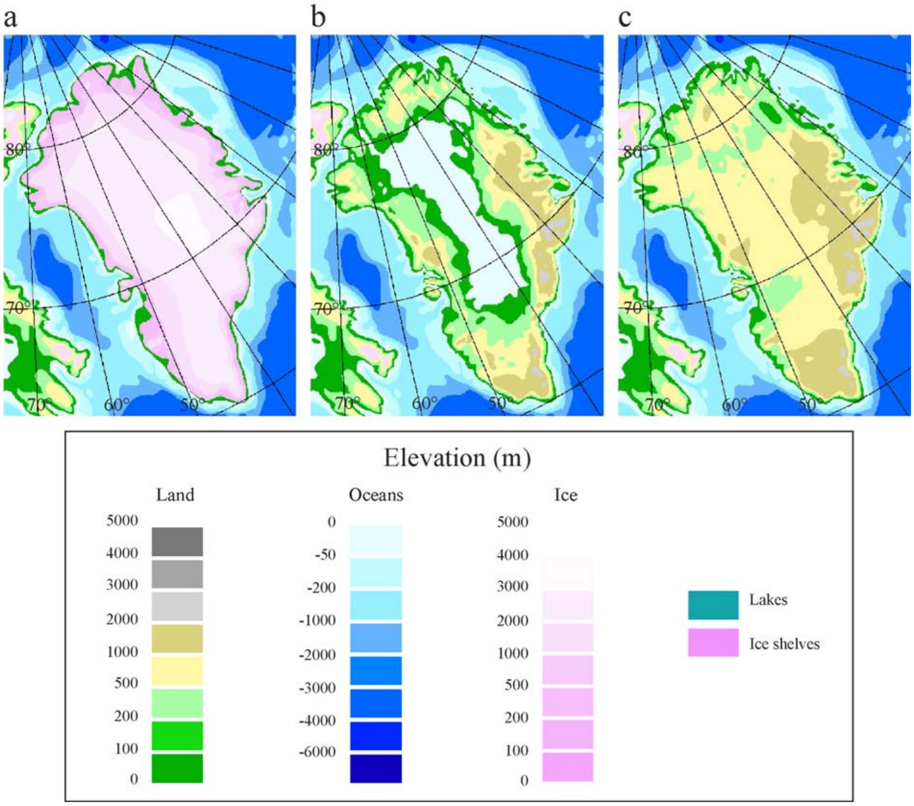

Fig. 6. Maps of Greenland compiled at the same resolution as the palaeogeographies in order to facilitate direct comparisons, showing: (a) modern topography with the ice sheet; (b) sub-glacial topography; (c) isostatically corrected sub-glacial topography.

Theide, 1980; Eyles, 1996; Larsen and Saunders, 1998; McKenna, 1983; Thiede and Myhre, 1996). This gridded data was then contoured in ArcInfo™ GRID.

3.2.7. Europe

The Pontide uplift area has been added in the central Tethys (Nikishin et al., 2001; Yilmaz et al., 1997; Ziegler, 1990; Ziegler et al., 2001), although its exact extent is speculative. We have also included a 1000-m contour around the highlands of the Atlantic margin of Norway. In Iberia, we have expanded the land area following Ziegler (1990) and Stampfli et al. (2001).

3.2.8. Asia

In Asia, we have again followed the map of Ziegler and Rowley (1998), except that we have modified the southward extent of the former area of the Lhasa and Quiangtang Blocks, in order to ensure the correct collisional history in the Eocene (Rowley, 1996).

3.3. Palaeobathymetry

The Maastrichtian map compiled by Ziegler and Rowley (1998) included only the shelf break for the oceans. It is therefore the palaeobathymetry that has been greatly augmented in this study, given its potential to affect ocean circulation in the coupled ocean–atmosphere model experiments. The bathymetry of the ocean crust is based on the following sources: a polygonized and rotated version of Müller et al.'s (1997) gridded ocean age dataset (contoured for 5 myr increments); a rotated modern ocean feature dataset; ODP-DSDP locations and data. We modelled tectonic features for areas of ocean floor that have been subducted subsequent to the Maastrichtian, assuming symmetry across spreading centres. Complexity exists in areas of hotspots (guyots, seamounts) or other ocean plateaus. These features are superimposed on the age–depth curve as positive features (taking into account flexural loading) and their past depths calculated accordingly. Additional information on palaeo-depths has also come from the extensive DSDP-ODP datasets, as well as other published sources.

The relationship between the depth of the ocean crust and its' age is well understood and equations to define the cooling curve are prominent in the published literature. We have used the original equation from Parsons and Sclater (1977) for ocean crust younger than 80 Ma:

$$d = 2500 + 350\sqrt{t} \quad (1)$$

For crust older than 80 Ma, we have used the following equation (Kearey and Vine, 1996; Parsons and McKenzie, 1978):

$$d = 6400 - 3200\exp\left(\frac{-t}{62.8}\right) \quad (2)$$

In each case, t is the age (myr) of the contemporary crust and d is its depth (m).

Using these equations, we have adopted two methods for reconstructing the palaeobathymetry. In both cases, we use the published ocean-age gridded dataset of Müller et al. (1997) as the starting point, with all analyses and manipulations performed in ArcInfo™ and ArcView® GIS. The first method converts the gridded data to points, which are then rotated to their palaeo-positions and then re-gridded. However, problems can occur due to projection-related grid size issues as cells move latitudinally. To avoid such issues, our preferred method for the current study was to contour the modern dataset for each 5 million-year interval (based on the GPTS time-scale of Cande and Kent, 1995) to create a set of 'synthetic' isochrons. These were then rotated to their palaeo-positions these with their associated tectonic plates and a new attribute calculated in the GIS to give the contemporary ocean age. To simplify the maps, only synthetic isochrons that are the same age or older than the time-slice are rotated for each relevant map (Fig. 5c). From the cooling curve (Eq. (1)), the bathymetric contours used in this study can be read off and plotted automatically.

Ziegler et al. (1985) used the following bathymetric contours for their palaeogeographies: 0, −50, −200, −4000 and −6000 m. The last three represent the average depths of the modern shelf break, base of the rise and edge of subduction trenches, respectively, and were therefore relatively robust to define. To these, we have also added contours for −1000, −2000 and −3000 m.

3.3.1. Indian Ocean

The palaeobathymetry of the Indian Ocean is dominated by series of interconnected spreading ridges separating Madagascar–India–Australia–Antarctica (Fig. 7a). In the northwest, the Madagascar–Seychelles ridge was in the process of jumping to its present position between India and the Seychelles (Plummer and Belle, 1995) associated with plume activity of the Réunion hotspot that forms the Deccan Traps around the K-T boundary (Duncan, 1990). We speculate that the jump around the Mascarene Bank may have occurred slightly later, based on the geometry of the plates at this time and the suggestion that the Mascarene Bank and Seychelles are separated by a distinct transform. Much of the Mascarene Bank may be a direct result of these volcanics, with the Réunion hotspot subsequently creating the Laccadive and Maldive ridges and the Chagos Bank (Duncan, 1990).

To the southeast, Kerguelen and Broken Ridge still formed one large feature (Coffin, 1992; Inokuchi and Heider, 1992; Schlich and Wise, 1992; Shipboard Scientific Party, 2000b). To the east the Australia–Antarctica ridge formed only a narrow bathymetric high towards Tasmania, which in the Maastrichtian was connected by shallow (<200 m) sea with Victoria Land (Shipboard Scientific Party, 2000a). The South Tasman Rise was subaerial (Shipboard Scientific Party, 2000a). Spreading between Antarctica and Australia remained slow until the Palaeogene (Müller et al., 2000), with an active India–Australia ridge system extending northeast to the subducting margin of SE Asia. Ninetyeast Ridge was actively forming at this time as India moved north (Royer et al., 1991).

3.3.2. Pacific Ocean

The palaeo-Pacific in the Maastrichtian was dominated by a number of plates that have now largely disappeared, although debate exists about exactly how many. In the northwest, the Sea of

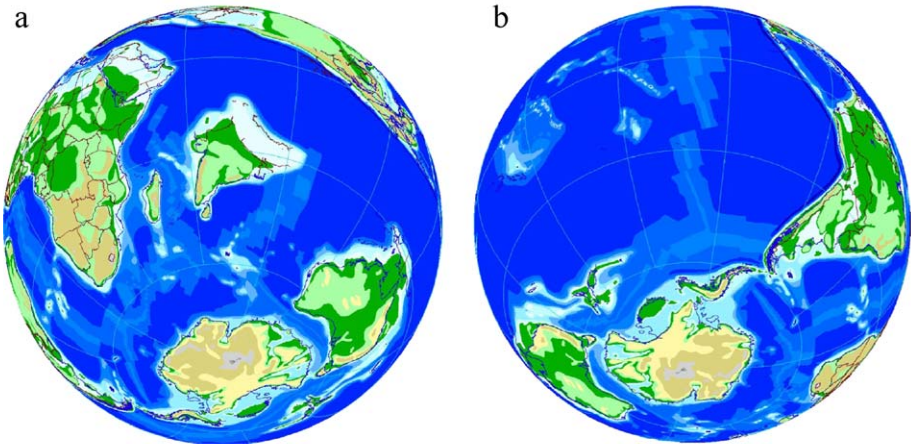

Fig. 7. (a) Orthographic view of Indian Ocean region showing the complexity of spreading ridges and bathymetric highs that result in a series of isolated deep basins. (b) Orthographic view of south east Pacific region showing bathymetric highs in the contemporary central Pacific (including the Ontong Java Plateau and Mid-Pacific Mountains). For legend, see Fig. 6.

Okhotsk block had docked by the Maastrichtian (Nokleberg et al., 2000). In the north, the Kula plate was subducting under Alaska, and in this version we have suggested that this continues south with the Pacific-Kula ridge to intersect the East Asian margin to the east of SE China. However, there is evidence to suggest that a third plate exists here, the Izanagi Plate, and that it is this plate that is subducting under eastern China and Japan, with the Izanagi–Kula ridge intersecting the coast further north (Engebretson et al., 1987; Sager et al., 1988). In the northeastern Pacific, Haeussler et al. (2001) and Roeske et al. (2002) have postulated the existence of an additional triple junction with the Resurrection Plate separating the Kula from the Farallon Plates during at least the Palaeocene and Eocene.

The remaining palaeo-bathymetric features are based on an assessment of the modern bathymetry using available DSDP and ODP data and their probable subsidence histories (Fig. 7b): mid-Pacific Mountains (Haggerty and Silva, 1995; Thiede et al., 1981; Winterer, 1976), Hess Rise (Vallier et al., 1981a,b), Shatsky Rise (Bralower et al., 2002; Sager et al., 1988; Sharman and Risch, 1988), Line Islands (Haggerty and Silva, 1995; Jackson and Schlanger, 1976; Schlanger and Silva, 1981; Winterer, 1976), Marshall Islands (Haggerty and Silva, 1995), Manhiki Plateau (Jackson and Schlanger, 1976; Jenkyns, 1976; Larson et al., 2002; Winterer, 1976) and Ontong Java Plateau (Berger et al., 1991; Kroenke and Yan, 1993; Kroenke and Wessel, 1999; Sliter and Leckie, 1993; Yan and Kroenke, 1993).

In the southwest Pacific, our tectonic model follows that of Kroenke (1984), Kroenke and Yan (1993) and Yan and Kroenke (1993). In the Maastrichtian, the picture is not as complex as it becomes in the Cenozoic. The Tasman Sea is still actively spreading (Gaina et al., 1998; Müller et al., 2000) and the adjacent Lord Howe Rise is probably relatively shallow.

3.3.3. Atlantic Ocean

In the central Atlantic, we have reconstructed the palaeobathymetry of the New England and Corner Seamounts, with the Crusier and Great Meteor seamounts (Grevemeyer, 1999) sitting astride the ridge at this time. To the south, the Ceara Rise and Sierra Leone Rise also lay over the mid-Atlantic ridge in the Maastrichtian.

In the South Atlantic, the Walvis ridge was still a major positive bathymetric feature in the Maastrichtian (Moore et al., 1984). To the south of the Aghulas fracture zone, the spreading centre lay further to the east than it does today, with the jump back to its modern position occurring in the Palaeocene, based on the ocean-age data (Müller et al., 1997).

3.4. Ice area

The contemporary ice area in the geological past is the subject of considerable debate, especially for the

Cretaceous (Abreu, 2000; Abreu et al., 1998; Frakes and Francis, 1988; Frakes et al., 1992; Kemper, 1987; Markwick and Rowley, 1998; Miller et al., 1999; Rowley and Markwick, 1992). We do not show ice on Antarctica on this map, although a small ice-sheet has been added above 1000 m for the purpose of a separate GCM sensitivity test. Miller et al. (1999) suggest the possibility of a "moderate-sized ice sheet" in the early Maastrichtian (ca. 71 Ma, Cande and Kent, 1995) based on sequence stratigraphic interpretations from New Jersey and oxygen isotope measurements from DSDP sites 305 and 463 in the Pacific and ODP site 690 off Antarctica. Those authors correlate this 'glaciation' with a global sea-level change recorded on the Haq et al. curve of about 60 m (Miller et al., 1999), which they suggest should really lie between 20 and 40 m (using the isotopic records and a transfer function of 0.008‰/m). However, which global change is being correlated with is unclear since the Haq et al. third-order curve, as presented on the Gradstein et al. (1994) time-scale (Hardenbol et al., 1998), shows falls at ca. 69 Ma (ca. 50 m) and 66 Ma (ca. 85 m), but a high at ca. 71 Ma. Nonetheless, our calculations (based on a revised version of the equations given in Rowley and Markwick, 1992) suggest that if the Haq curve is correct, then it requires an average ice sheet area in the Maastrichtian of about 4,366,000 km^2 (36% modern Antarctica). This, in turn, would be enough to glaciate everything above 1000 m on the isostatically corrected ice-free reconstruction (Markwick and Rowley, 1998). A literal interpretation of the third-order curve drop at 66 Ma would require an ice sheet of almost 1.5× modern Antarctica, which seems untenable given the lack of direct evidence around Antarctica (Rowley and Markwick, 1992). Following the arguments of Rowley and Markwick (1992), we accept that the magnitudes of the Haq et al. curve are probably in error, and while a small ice-sheet on the Gamburtsev Mountains is possible, this would only account for a eustatic change of 10–20 m (Markwick and Rowley, 1998; Rowley and Markwick, 1992). This is less than that required by Miller et al. (1999).

The isotopic data is even more problematic, since given currently quoted bottom water temperatures of 10 °C (D'Hondt and Arthur, 2002), a literal interpretation of the oxygen isotope record (Zachos et al., 2001) would require an even larger ice-sheet in the Maastrichtian (taking the mean benthic value from Zachos et al., would require all of the isostatically corrected sub-glacial topography to be glaciated, which again is not supported by geological data).

3.5. Digital elevation models and palaeo-drainage

Digital elevation models are gridded representations of the Earth's topography and bathymetry, which greatly facilitates calculations and analyses in studies of the landscape and its interactions with climate, hydrological and biological systems. DEMs of the present day (e.g. ETOPO 5, Edwards, 1986, and GLOBE, GLOBE Task Team et al., 1999; USGS, 2001) are widely used for this sort of work, but there have been few attempts to define past drainages within an integrated, evolving palaeo-DEM (Lillegraven and Ostresh, 1988).

Our aim in this study is to generate a global palaeo-DEM for the Maastrichtian that can be used directly by the HadCM3-coupled ocean–atmosphere climate model as the geographic boundary condition, incorporating defined palaeodrainage. Eventually, we hope to be able to generate a fully evolving landscape model. HadCM3 is uncommon among current CGCMs in that the drainage basins must be explicitly defined complete with an outfall point, with model-derived freshwater runoff being discharged at this outfall for interaction with the ocean (Cox et al., 1999). By defining the palaeo-DEM as the boundary condition, rather than simply smoothing broadly defined elevational data as has been done previously in climate modelling, we can more precisely partition surface dependent climatic variables, especially the surface runoff. This will have major importance in oil, gas and mineral exploration, where modelled sediment fluxes (in part, a reflection of surface runoff) can be used to predict the nature and distribution of reservoir facies (Markwick, 2001; Markwick and Valdes, 2003; Markwick et al., 2002).

The derivation of DEMs from palaeogeographical maps follows a similar process to that adopted by cartographers and hydrologists working with modern data (Jenson and Domingue, 1988; Martz and Garbrecht, 1998; O'Callaghan and Mark, 1984). The major difference is that the data resolution (grain) is far poorer for palaeo-maps. Consequently, with only nine terrestrial contours defined globally, it is untenable to use the palaeo-DEM to generate the palaeorivers and palaeodrainage basins automatically. These have to be defined independently based on geological evidence.

3.5.1. Defining palaeodrainage systems

Some of the major palaeodrainage systems of the late Cretaceous are relatively well constrained. The palaeo-Amazon has long been known to have drained westward prior to the late Miocene (Hoorn, 1993; Hoorn et al., 1995). In Africa, the palaeo-Congo is believed to have drained to the northwest into the Benue Trough prior to the late Cretaceous or Palaeogene (Censier, 1990; Giresse, 1982) and inversion along the Central Africa Shear Zone. In southern Africa, the Limpopo and Zambezi have followed approximately their modern courses since the early Cretaceous (Dingle et al., 1983), although Dingle et al., do suggest changes in the Orange drainage system. In Central and Southeast Asia, however, the pattern of most of the major river systems probably originated from Tertiary tectonic activity; even the large systems such as the Huang He (Lin et al., 2001) or the Red River, so their Maastrichtian topology is more speculative. In Europe and North America, terrestrial drainage is greatly influenced by the seaways that covered large areas of the continents, whose retreat during the Palaeogene dictated the subsequent drainage history.

For the Maastrichtian, the palaeodrainage systems are based on this existing data, as well as the constraints defined by the palaeogeograhy itself. Globally, we have taken the modern major river system (Fig. 4d) and traced each drainage system back through time to understand the point at which tectonics forces us to reconsider the drainage geometry. This is clearly an iterative process, which requires palaeogeographic reconstructions for the intervening time-slices (Markwick et al., 2000). For HadCM3, the critical definition is the geometry of the drainage basin and the outfall point of each. The result is that the Maastrichtian has been divided into 605 drainage basins (Fig. 8a), of which 13 are internally draining (mainly in eastern Asia). The general river topology has been added in Fig. 8b, but with the caveat that probably only the major trunk streams are robustly definable. Strahler numbers have then been appended to each river segment (Strahler, 1957) and nodes at river source, confluence and out-fall have been attributed in the GIS (Fig. 8c). The drainage basins have then been gridded for use in the HadCM3 together with the positions of each relevant out-fall nodes.

3.5.2. Generating the palaeo-DEM

The palaeo-DEM is created entirely within ArcInfo™ GIS using the GRID extension and its suite of hydrological tools (ESRI, 1992). The results can be displayed within ArcView® GIS or ArcGIS™,

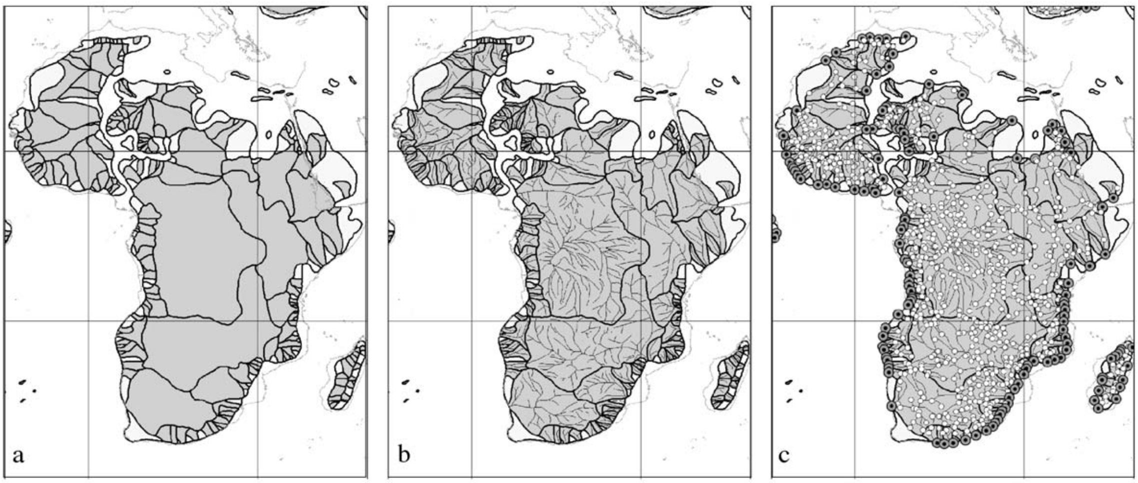

Fig. 8. Detail of the Maastrichtian of Africa illustrating (a) palaeodrainage basins, (b) palaeorivers and (c) drainage nodes.

exported to any other raster program, or directly to the climate model. The basic input into GRID is the palaeogeographic contours data (Figs. 9a). This contour dataset is then interpolated to the grid cell size required (0.1° in these experiments) using the TOPOGRID command (Fig. 9b). For the palaeo-DEM shown in this study, the grid cells were generated using a geographic coordinate system. This means that calculations made using the grid must be made with care since the absolute area of each grid cell will change with latitude. For quantitative analyses, the grids must be generated on an equal area projection, which requires that the world be broken down into regions that can be manageable in this way, such as by continent using a Lambert Equal area projection (Steinwand et al., 1995). Gridding using both marine and terrestrial contours creates spurious results because of the sparcity of contours (especially in the

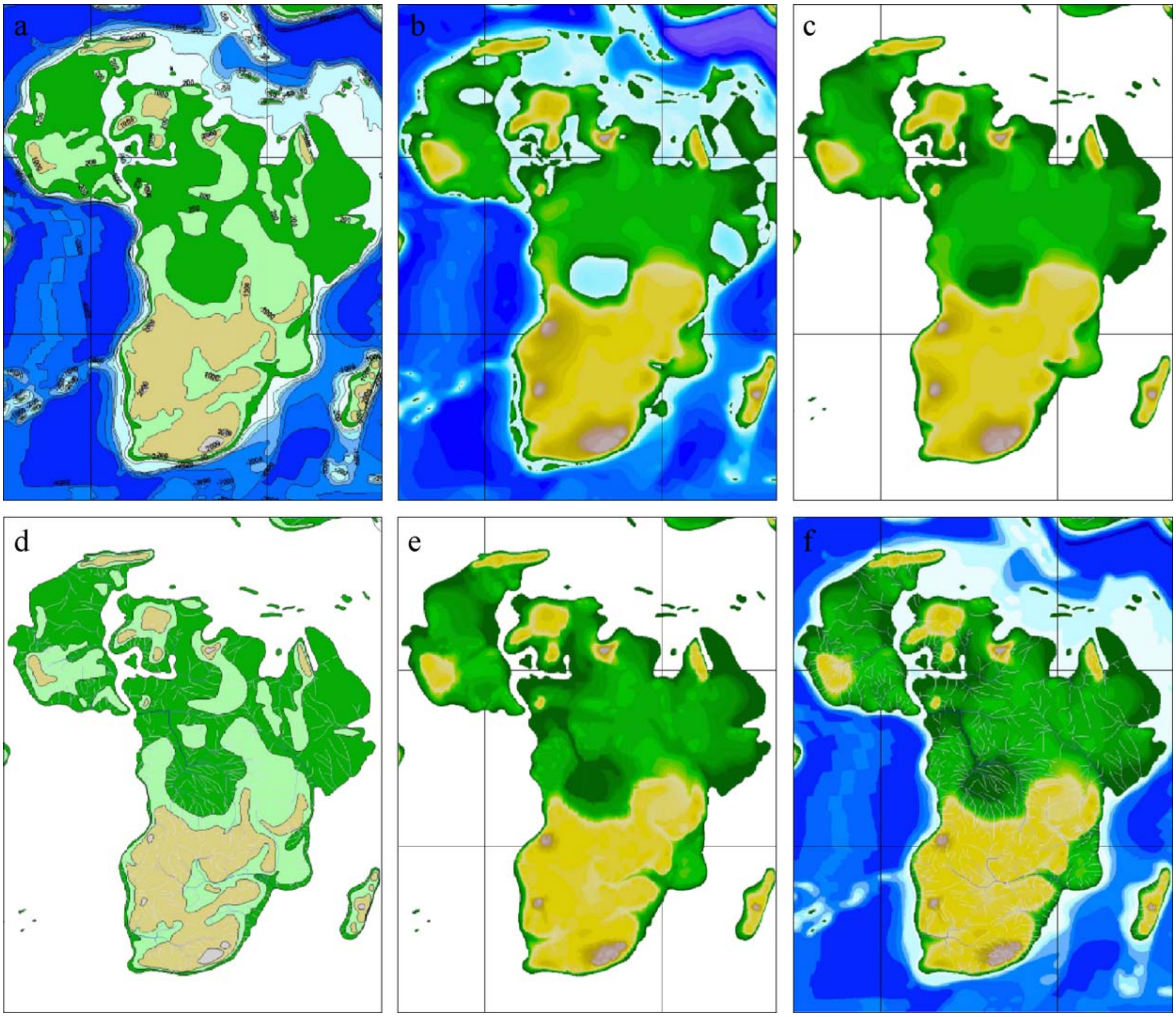

Fig. 9. The process of generating the DEM: (a) the contour map used as primary input; (b) DEM generated with no correction leads to problems in areas of low relief (e.g. East Africa); (c) ocean and land generated separately using masks for each; (d) river coverage on contour map; (e) DEM generated using the land-only mask, the rivers coverage and minimum elevation set to 0 m; (f) final DEM with ocean. For legend, see Fig. 6 (for a,d and oceans) and Fig. 10 (for b,c,e,f).

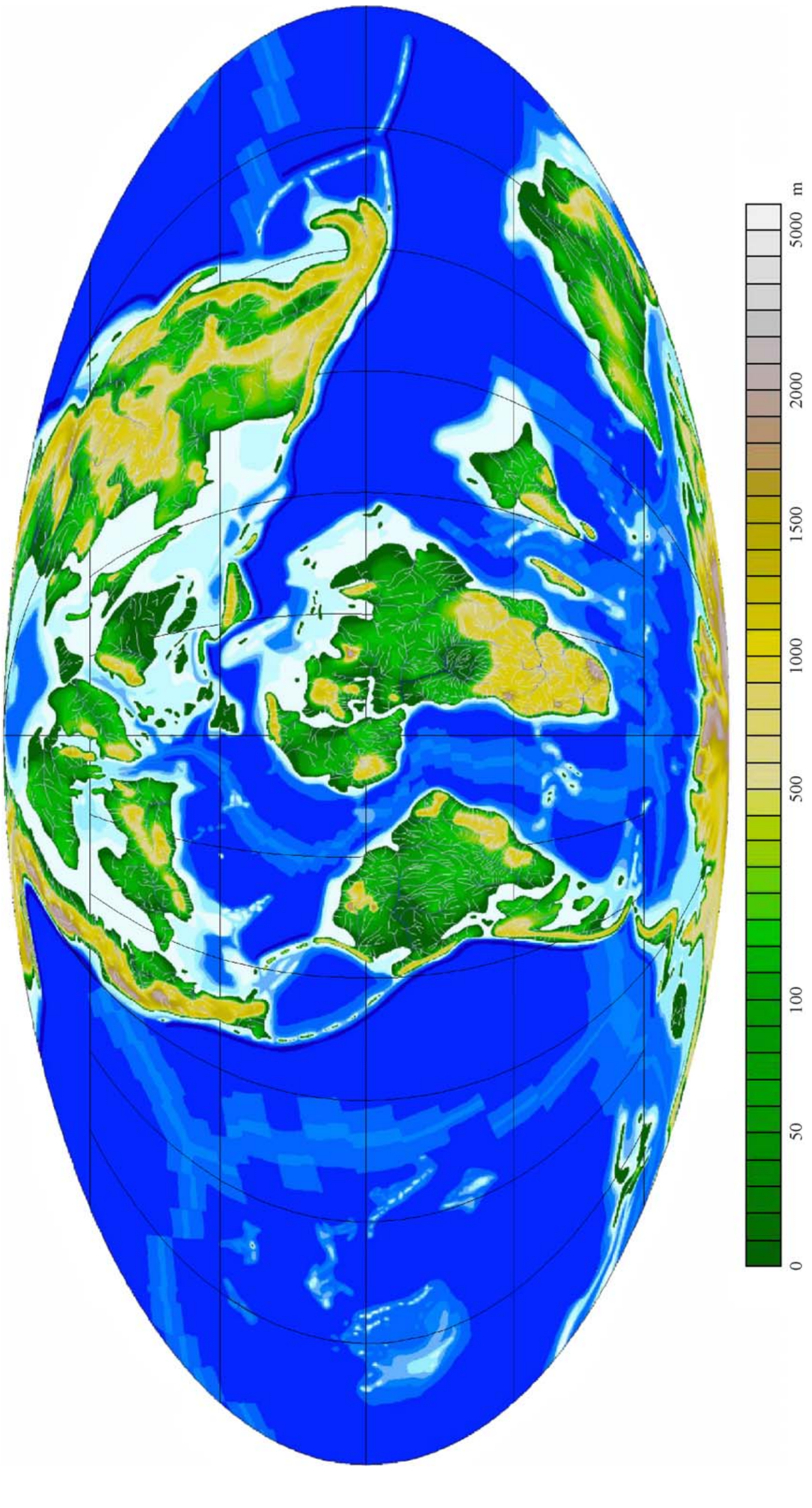

Fig. 10. Final Maastrichtian palaeo-DEM. Mollweide projection.

deep marine environment and in lowland areas) and the use of a spline fit through the data. This is a common problem in going from contour data to a DEM and is found even in well-established DEMs of the Recent, such as ETOPO5 (Edwards, 1986) and GLOBE (GLOBE Task Team et al., 1999). To counter this, the shoreline has been used as an anchor contour with the oceans and land being treated separately through the use of masks representing the land and sea, as defined by the 0-m (shoreline) contour. In addition, and in order to prevent spurious results in large terrestrial flat areas such as the Congo basin (Fig. 9b), the minimum terrestrial elevation limit is also set at 0 m (Fig. 9c) (see Martz and Garbrecht, 1998 for a discussion of the problem of flat areas and depressions in DEMs). In this paper, we concentrate on describing the method as applied to topography, but the same method is also applied to bathymetry.

The output in Fig. 9c is still relatively coarse with numerous large sinks (depressions; areas of internal drainage), such as the Congo (Cuvette Centrale), which are not supported by geological observations. This is again a reflection of the limited contour resolution, and requires additional information, which is supplied by the palaeorivers coverage generated above (Fig. 9d). This removes many of the large spurious sinks (Fig. 9e). The final stage is to specify the purported sink areas (Fig. 9f). The use of the palaeorivers coverage in generating the DEM explicitly means that all resulting long profiles will be approximately equilibrium profiles, which is unlikely to be the case (in reality) in areas of major tectonic activity. Again, this is a problem that will be dealt with in future studies.

In order to check that the resulting DEM is "hydrologically correct" (the drainage systems are consistent with the DEM) the derived palaeo-DEM is used to generate a drainage system, which is compared with the original palaeorivers dataset. This method follows that used by USGS in generating the Hydro1K dataset (Verdin, 2001) and described in Tarboton et al. (1991). Areas of internal drainage (sinks) must be specified. However, the derived DEM also includes analytically derived sinks (usually small) that must be first removed. This is done by first filling all sinks in the data to create a flow system that can only outlet to the major ocean system. Once all of the sinks have been removed, the required sinks (representing known internally draining basins) are added back to the dataset.

To obtain the flow channels using the palaeo-DEM, streamlines are constructed by calculating the flow direction and the flow accumulation for each grid cell. The results are then compared with the original palaeoriver dataset and modifications made through several iterations until the two datasets (DEM derived rivers and independently derived palaeorivers) are comparable. With the revised DEM and river networks, the final stage is to define the drainage basins and watersheds using GRID, as a further check on the consistency between geologically derived drainage basins (Fig. 8a) and what the generated palaeo-DEM implies. This is again an iterative process.

Hypsometric curves can be generated from the resulting surface and compared with the smoothed versions based solely on the drawn contours. The effects of changing sea-level in Maastrichtian are also now easier to analyse (for use with climate model sensitivity experiments), which may indicate further modifications to the DEM model or guide further questions of the stratigraphy itself.

4. Conclusions

Palaeogeography (palaeotopography and palaeobathymetry) provides a critical boundary condition for models of surface processes, especially the new generation of coupled ocean–atmosphere models. The need for more robust geographies has led us to develop a GIS-based palaeo-DEM for the Maastrichtian (late Cretaceous) for use by climate modellers especially those using the HadCM3 model with its requirement for specifying palaeodrainage systems (Fig. 10). The underlying palaeogeography draws heavily on the Maastrichtian palaeotopography of Ziegler and Rowley (1998), to which we have made some modifications and added a detailed palaeobathymetry. This process has directed us to refinements that will be included in later versions, as part of an ongoing project to complete a series of stage-level palaeogeographic maps for the Cretaceous and Cenozoic (Markwick et al., 2000).

Acknowledgements

The Maastrichtian palaeogeographic map presented here is one of a series of Cretaceous and Cenozoic time-slice maps, based on the work of the Paleogeographic Atlas Project, and begun as part of Markwick's PhD at the University of Chicago, where funding was through a teaching assistantship. Markwick is indebted to Fred Ziegler, David Rowley and Mike Hulver of the Paleogeographic Atlas Project (The University of Chicago) without whose guidance this work would not have been continued. Markwick also gratefully acknowledges Fugro Robertson Limited for providing access to their digitisers and ArcInfo™ and ArcView® GIS software. We thank Matthew Huber, Roy Livermore and an anonymous reviewer for their constructive comments.

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