Dynamic topography and eustasy controlled the paleogeographic evolution of northern Africa since the mid-Cretaceous

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Abstract

Northern Africa underwent widespread inundation during the Late Cretaceous. Changes in eustasy do not explain the absence of this inundation across the remainder of Africa, and the timing and location of documented tectonic deformation do not explain the large-scale paleogeographic evolution. We investigate the combined effects of vertical surface displacements predicted by a series of mantle flow models and eustasy on northern African paleoenvironmental change. We compare changes in base level computed as the difference between eustasy and long-wavelength dynamic topography arising from sources of buoyancy deeper than 350 km to the evolution of paleoshorelines derived from two interpolated global data sets since the mid-Cretaceous. We also compare the predicted mantle temperature field of these mantle flow models at present-day to several seismic tomography models. This approach reveals that dynamic subsidence, related to Africa’s northward motion away from the buoyant regions overlying the African large low shear velocity province, amplified sea level rise, resulting in maximum inundation of northern Africa during the Cenomanian and Turonian. By the Cenozoic, decreased magnitudes of dynamic subsidence, reflecting the reduced drawdown effects of slab material beneath northern Africa associated with the impact of the Africa-Eurasia collision, combined with a comparatively pronounced progressive sea level fall resulted in ongoing region-wide regression along coastal regions. The temporal match between our preferred model and the paleoshoreline data sets suggests that the paleogeographic evolution of this region since the Late Cretaceous has mainly been influenced by the interplay between eustasy and long-wavelength dynamic topography arising from large-scale, subduction-driven, lower mantle convection.

1. Introduction

During the Late Cretaceous the paleoenvironmental setting across present-day arid northern Africa was one of widespread shallow marine inundations [Golonka, 2007, 2009; Guiraud et al., 2005; Smith et al., 1994] (Figures 1 and 2b–2f). Bata et al. [2016] discussed the critical role this Cretaceous inundation played in transporting and depositing economically important minerals (e.g., gold, diamond, and platinum) within large volumes of sediments to parts of northern Africa’s continental interior. Guiraud et al. [2005] attributed this Late Cretaceous inundation to the interplay between changes in eustasy and climate, noting also the influence of tectonic deformation on topography across the region [Jolivet et al., 2015; Frizon de Lamotte et al., 2015; Heine et al., 2013]. Yet these tectonic processes that occurred on spatial wavelengths of ~102 km cannot be reconciled with the entire extent of this inundation, which occurred on spatial wavelengths of ~103 km (Figure 1), and changes in eustasy alone do not explain the lack of preserved shallow-water sediments across the remainder of inland Africa [Golonka, 2007, 2009; Smith et al., 1994]. Earlier studies investigated the possibility that Cretaceous inundation was linked to changes in underlying mantle processes across the region, concluding that this is likely the case given that these paleoshorelines are preserved too far into northern Africa’s continental interior to be explained by another process [Sahagian, 1988]. A previous study of the influence of large-scale mantle dynamics on this Late Cretaceous flooding underpredicted its extent [Spasovic and Gurnis, 2012], which was attributed to the significant influence of tectonic processes as documented in Guiraud et al. [2005]. Investigating the emplacement of ophiolites along the northern margin of Africa associated with the closure of the Neotethys Ocean via a major obduction event during the Late Cretaceous suggests that the evolution of paleogeography across the continent could be linked to a series of switching mantle convective regimes [Jolivet et al., 2015]. A viable way to test this hypothesis would be with numerical models of mantle convection accounting for the combined influence of both slabs and plumes beneath the continent [Jolivet et al., 2015].
Figure 1. Reconstruction of continental crust (grey) in an absolute mantle frame of reference [Steinberger and Torsvik, 2008; O’Neill et al., 2005] with a focus on Africa’s plate motion history, including plate boundaries and age of oceanic crust during the Late Cretaceous and Paleogene [Müller et al., 2016]. Paleoshorelines [Heine et al., 2015] from Golonka [2007, 2009] (blue) and Smith et al. [1994] (violet) are shown. Regions of intracontinental extension [Heine et al., 2013] are shown in black during Early Cretaceous rifting and in hachured during the thermal subsidence phase.
Here we investigate the interplay between mantle dynamics and the evolution of sea level across northern Africa. We use recently developed forward geodynamic models [e.g., Hassan et al., 2015, 2016] back to the Early Cretaceous to model the history of plate motion, subduction, plume upwelling, and mantle convection across the region. We define the differences between the evolution of eustasy and surface dynamic topography, computed from the geodynamic models, as base level [Barnett-Moore et al., 2014]. We then compare the rate of change of computed dynamic topography and base level to estimates of transgression and regression across the region, which are derived from two published paleoshorelines interpolate through time [Golonka, 2007, 2009; Smith et al., 1994; see Heine et al., 2015, Figure 1]. Quasi-quantitative comparisons between base level and interpolated paleoshoreline data sets allow us to evaluate the contributions of dynamic topography and eustasy to the paleoenvironments of northern Africa since the Late Cretaceous.

Figure 2. (a) Present-day topography [Amante and Eakins, 2009] of northern Africa. The black square denotes extent of Tuareg Shield [Rougier et al., 2013]. RS: Red Sea and GA: Gulf of Aden. (b–f) Paleoshorelines [Heine et al., 2015] from Golonka [2007, 2009] (blue) and Smith et al. [1994] (violet). (g–l) Paleoenvironment indicator maps (see section 2) derived from the interpolation of paleoshorelines [Golonka, 2007, 2009]. Gold stars represent point locations analyzed in this study. Black outlines represent the basin outline that each point location lies within (Figure 2g). AS: Amguid Spur, TB: Taoudeni Basin, SB: Sirte Basin, and UEB: Upper Egypt Basin. (m–r) Same as Figures 2g–2l except for the paleoshorelines of Smith et al. [1994].
2. Methods

We select four point locations across northern Africa, which represent locations both within inland basins and along coastal regions (Figure 2g). These point locations were selected because they are in four basins spread over northern Africa, in which paleogeography maps indicate transgression followed by regression since the Late Cretaceous. At these locations we extract histories of (1) paleogeography (section 2.1), (2) dynamic topography (section 2.2), and (3) base level (section 3). Given the long-wavelength nature of modeled dynamic topography computed from the global mantle flow models (e.g., >1000 km) [Flament et al., 2013], additional point locations selected on a smaller spatial scale would not reveal further insights.

2.1. Paleoenvironment Indicator Maps

The two paleogeography data sets analyzed in this study [e.g., Golonka, 2007, 2009; Smith et al., 1994] were originally synthesized from proprietary data sets and global and regional paleogeographic sources published during the “predigital” era. The paleoenvironmental interpretations used to construct these maps were compiled from a variety of sources including rock outcrops, proprietary well and seismic reflection data, fossils, and earlier published paleogeographies. For more details on the construction of these paleogeographic data sets see Heine et al. [2015]. The global paleogeographic map compilation of Golonka [2007, 2009] is subdivided into 32 time intervals across the Phanerozoic, with time instances bounded by stratigraphic unconformities (e.g., Figures 2b–2f). Similarly, the compilation of Smith et al. [1994] is defined by stratigraphic stage boundaries, covering 31 time instances across the Mesozoic and Cenozoic (e.g., Figures 2b–2f). In both data sets, the time intervals associated with the paleogeographic maps are variable, ranging between 4 and 25 Myr. We therefore interpolate both of these data sets, independently, at 1 Myr intervals (Figures 2g–2r) based on the simple procedure outlined below:

1. Generate points on the sphere that sample equal area [e.g., Górski et al., 2005] in order to generate binary maps (land/ocean) for each time slice.
2. Compute the intersection $P$ of binary maps for a given pair of time slices ($S_{i-1}, S_i$). This intersection is a collection of isolated patches that represent newly submerged regions, $P_w$, and newly emerged regions, $P_t$, or $P = P_w \cup P_t$.
3. Compute a distance map of $P$ to closest land in $S_0$, based on a spherical distance metric.
4. Normalize the distance map such that within each isolated patch distances range between $[0, 1]$.
5. Relate normalized time progressions between $S_{i-1}$ and $S_i$ to the normalized distance map in step 4. That is, as time progresses between $S_{i-1}$ and $S_i$, regions featured in the distance map transition from their former state in $S_{i-1}$ to their latter state in $S_i$, based on corresponding values of distances. These transitions proceed such that regions in $P_w$ farthest from land are inundated first, whereas regions in $P_t$ closest to land emerge first.

In the absence of paleoelevation constraints, we assume a linear progression of transgressions and regressions. We then smooth these interpolated paleoshorelines over time windows of varying durations, allowing us to compute rates of changes that indicate whether a given coastline is undergoing transgression or regression. We apply a smoothing window of 20 Myr to derive long-wavelength rates of transgression or regression that can be compared to the long-wavelength changes in dynamic topography and sea level.

2.2. Numerical Models of Mantle Convection

We devise numerical models of convection within the Earth’s mantle under the extended-Boussinesq approximation [Christensen and Yuen, 1985]. The Earth’s mantle is modeled as a spherical shell with depth-dependent thermodynamic properties and temperature- and depth-dependent rheology, where the deepest lower mantle is chemically heterogeneous. We solve the equations for the conservation of mass, momentum, and energy using the parallel finite element code, CitcomS [Zhong et al., 2008], which has been extended to allow for progressive assimilation of surface plate motion and subducting slabs derived based on global plate reconstructions [Bower et al., 2015]. The underlying assumptions and the choice of model parameters employed have been outlined in earlier work [Hassan et al., 2015; Bower et al., 2013]. Table 1 lists parameters held fixed across all model cases, and additional details can be found in Hassan et al. [2015].
We use piecewise Arrhenius laws to describe the variation of viscosity with temperature, depth, and composition in the Earth's mantle, which takes the following nondimensional form:

$$
\eta(T, r) = A(r) \eta_c \exp \left( \frac{E_a(r) + (1 - r)V_a(r)}{T + T_{off}} - \frac{E_o(r) + (1 - r_{inner})V_o(r)}{1 + T_{off}} \right),
$$

(1)

where $\eta$ is the viscosity, $T$ is the temperature, $r$ is the radius, $A$ is the preexponential parameter, $\eta_c$ is the intrinsic composition-dependent prefactor, $E_a$ is the activation energy, $V_a$ is the activation volume, and $T_{off}$ is the temperature offset. For the lower mantle, we use a dimensional activation energy of 320 KJ mol$^{-1}$ and activation volume of 6.7E-6 m$^3$ mol$^{-1}$, corresponding to nondimensional units of 11 and 26, respectively, which are comparable to estimates in Karato and Wu [1993]. However, since such viscosity parameterizations lead to large viscosity variations that cause numerical difficulties, we adjust the preexponential parameter $A(r)$ and the temperature offset $T_{off}$ [Tackley, 1996] to limit the viscosity contrast to 3 orders of magnitude. The resulting viscosity profile is similar to the preferred viscosity profiles of Steinberger and Calderwood [2006]. Additional details on model setup can be found in Hassan et al. [2015].

We apply kinematic surface boundary conditions based on surface velocities derived from global plate tectonic reconstructions at 1 Myr intervals, with a linear interpolation in between. We assimilate thermal models of inferred subducting slabs into the dynamically evolving temperature field at each time step, as a model progresses toward present day, starting from a given geological time (see Bower et al. [2015] for more details).

### 2.2.2. Computation of Dynamic Topography

We compute time-dependent dynamic topography, $h$, at the surface at 5 Myr intervals as follows:

$$
h = \frac{\sigma_{rr}}{\Delta\rho g},
$$

(2)

where $\sigma_{rr}$ and $\Delta\rho$ are the radial component of stress and the density difference between the mantle and the overlying material, respectively. The radial stresses, $\sigma_{rr}$, are recomputed using Stokes flow and the temperature field at a given time. We exclude buoyancy in the top 350 km of the mantle in the Stokes flow computations in order to remove influences of data assimilation. Moreover, to exclude influences from traction induced by kinematic plate velocities (velocity boundary condition at the surface), we impose free-slip boundary conditions at the surface in these Stokes flow computations.

### 2.2.3. Parameter Space Explored

To assess the influence of model parameters and initial conditions we computed a number of models varying the density contrast ($\Delta\rho_{init}$) of a dense chemical layer above the core-mantle boundary (CMB), the plate reconstruction that dictates surface plate velocities and the location of subduction zones, and the geological age at which models are initiated (Table 2). The dense layer covering the CMB is 100 km thick in all models, which is volumetrically comparable to estimates of the volume of large low shear velocity provinces (LLSVPs) [Hernlund and Houser, 2008; Burke et al., 2008]. In most model cases, we use tectonic reconstruction TR1,

### Table 1. Model Parameters Held Fixed Across All Model Cases Presented in This Study

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rayleigh number$^a$</td>
<td>$Ra$</td>
<td>$5 \times 10^8$</td>
<td>-</td>
</tr>
<tr>
<td>Earth radius</td>
<td>$R_0$</td>
<td>6371</td>
<td>km</td>
</tr>
<tr>
<td>Density</td>
<td>$\rho$</td>
<td>3930</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>Thermal expansivity</td>
<td>$\kappa$</td>
<td>$1.42 \times 10^{-5}$</td>
<td>K$^{-1}$</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>$\kappa_0$</td>
<td>$1 \times 10^{-6}$</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>Specific heat capacity</td>
<td>$C_P$</td>
<td>1100</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>Gravitational acceleration</td>
<td>$g$</td>
<td>10</td>
<td>ms$^{-2}$</td>
</tr>
<tr>
<td>Surface temperature</td>
<td>$T_s$</td>
<td>300</td>
<td>K</td>
</tr>
<tr>
<td>Dissipation number</td>
<td>$D_l$</td>
<td>0.8</td>
<td>-</td>
</tr>
<tr>
<td>Reference viscosity</td>
<td>$\eta_0$</td>
<td>$1 \times 10^{-21}$</td>
<td>Pas</td>
</tr>
<tr>
<td>Internal heating</td>
<td>$H$</td>
<td>100</td>
<td>-</td>
</tr>
</tbody>
</table>

$^a$Note that the radius of the Earth is used instead of the thickness of the mantle to compute $Ra$, which scales it by a factor of ~11, compared to that computed based on the thickness of the mantle.
which is based on TR2 [Seton et al., 2012], but includes updates and improvements in the North Atlantic and Western Tethys [Müller et al., 2016]. In the North Atlantic Ocean, TR1 incorporates new reconstructions of rifting and breakup between Eurasia and North America, accounting for multiple episodes of extension in the region [Barnett-Moore et al., 2016]. Similarly, in the Western Tethys, this plate model incorporates an alternative plate reconstruction adapted from the work of Schettino and Turco [2011] into a global plate model [Hosseinpour et al., 2016]. In the study of Hosseinpour et al. [2016], a comparison between the reconstructed plate boundaries, constraining the newly incorporated subduction histories across the region, to several different seismic tomography models that image the mantle velocity structure shows a satisfactory match between Cenozoic subduction events in the Western Tethys. This match, however, becomes less clear for older subducted material [Hosseinpour et al., 2016], suggesting that this model of the West Tethys can be further improved.

2.3. Definition of Base Level

In geomorphology, base level is defined as the elevation at which a stream can no longer incise below (Figures 3a and 3b) [Leopold and Bull, 1979]. Here we propose that local variations in base level along a coastline, which influence shoreline migration, are controlled by the interplay between dynamic topography and eustasy (Figures 3c and 3d) [Barnett-Moore et al., 2014]. We consider long-term eustasy to be driven primarily by changes in seafloor spreading rates, with mantle-driven vertical motions contributing a second-order effect over geological timescales [Spasojevic and Gurnis, 2012]. We calculate base level, \( \phi \), by subtracting the dynamic topography history for all model cases at each considered point location (Figure S1 in the supporting information) from the smoothed global sea level curves of Haq et al. [1987] and Haq and Al-Qahtani [2005] (as presented in Müller et al. [2008]) and present base level histories from Case I and Case I(s) (Figure 4) in greater detail below in section 3. Increases in base level correspond to a transgression, while decreases correspond to a regression (Figure 4d). The time derivative of base level, \( \dot{\phi} \), determines the rate of change.

Table 2. Model Parameters Varied Across Model Cases, Showing the Parameter Space Explored

<table>
<thead>
<tr>
<th>Case</th>
<th>( \Delta \rho_{ch} )%</th>
<th>Tectonic Reconstruction</th>
<th>Start Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CI</td>
<td>2.5</td>
<td>TR1</td>
<td>230</td>
</tr>
<tr>
<td>CI(s)</td>
<td>10</td>
<td>TR1</td>
<td>230</td>
</tr>
<tr>
<td>CLI</td>
<td>3.0</td>
<td>TR1</td>
<td>230</td>
</tr>
<tr>
<td>CLII</td>
<td>3.5</td>
<td>TR1</td>
<td>230</td>
</tr>
<tr>
<td>CIV</td>
<td>3.5</td>
<td>TR2</td>
<td>230</td>
</tr>
<tr>
<td>CV</td>
<td>3.5</td>
<td>TR2</td>
<td>140</td>
</tr>
<tr>
<td>CVI</td>
<td>3.5</td>
<td>TR1</td>
<td>140</td>
</tr>
</tbody>
</table>

Figure 3. Schematic cartoon highlighting the different processes that can influence changes in base level (\( \phi \); dynamic topography minus sea level, see section 3). (a) Assuming tectonic stability, sea level rise generates a transgression, whereas a sea level fall generates a regression. (b) Assuming constant sea level, dynamic uplift results in a shoreline regression, and dynamic subsidence a shoreline transgression. (c) In the scenario where sea level rise is greater than dynamic uplift a shoreline transgression occurs. (d) In the scenario where sea level fall is greater than dynamic subsidence shoreline regression occurs.
of these predicted regressive and transgressive phases: $\hat{\phi} > 0$ indicates landward shoreline shifts (transgressions) and $\hat{\phi} < 0$ indicates seaward shoreline shifts (regressions) (Figure 5b). These rates of change can then be compared against the smoothed curves extracted from the paleoenvironment indicator maps (Figure 5).

3. Results

3.1. Dynamic Topography, Sea Level, and Base Level

We investigate the interplay between the dynamic topography predicted by a series of geodynamic models (Table 2) and the long-term sea level curves of Haq et al. [1987] and Haq and Al-Qahtani [2005], here referred to as H87 and H05, respectively. A comparison between the evolution of dynamic topography in the models, at each considered point location, shows that the dynamic topography history of the region is generally consistent across the parameter space explored in this study (Figures 4a, 4c, 4e, and 4g), suggesting that model predictions are not particularly sensitive to the choice of model parameters. A twin model for Case I, namely
Case I(s), was run in which active mantle upwellings were suppressed by imposing a dense chemical layer above the CMB with a much higher-density contrast ($\Delta\rho_{ch} = 10\%$). This results in a reduced magnitude of dynamic topography within inland northern Africa including the Taoudeni Basin (Figure 4a), and generally similar patterns of predicted dynamic topography compared to Case I, for Cretaceous and Cenozoic times across the remaining point locations (Figures 4c, 4e, and 4g). Differences computed between the dynamic topography history of Case I and other models presented here show that dynamic topography histories do not drastically differ except in the instance of Case I(s) in the Taoudeni Basin (Figure S1). We compare Case I and Case I(s) (Table 2) to understand the influence of active mantle upwellings on the results.

At the selected point locations, Case I and Case I(s) generally predict dynamic subsidence increases across the region from $\sim$120 Ma to $\sim$70–85 Ma (e.g., Figure 3b), at which time the rate of subsidence begins to decline, approaching negligible magnitudes toward present day (Figures 4b, 4d, 4f, and 4h), except for Case I(s) in the Taoudeni Basin (Figure 4b). We consider that paleoshoreline evolution is driven primarily by fluctuations in sea level when the vertical motions associated with dynamic subsidence rates are less than $-2 \text{ m Myr}^{-1}$ (Figures 4b, 4d, 4f, and 4h). The evolution of global eustasy in both H87 and H05 shows an increase from $\sim$150–180 m above present-day sea level at 120 Ma to $\sim$210–250 m at $\sim$80 Ma, at which time both curves imply an onset in sea level fall that continues until present day (Figures 4a, 4c, 4e, and 4g). Across the region, the rate of sea level fall is greater than the rate of dynamic subsidence for the last 80 Ma, suggesting a transition from a transgressive to a regressive environment. In the case of the Taoudeni Basin, the very low rate of dynamic subsidence predicted by Case I(s) suggests that eustatic sea level evolution would have been the dominant mechanism shaping shoreline regression during the Late Cretaceous. Calculated base level for both Case I and Case I(s) confirms this interplay between eustasy and dynamic topography (Figure 3d) across all locations, illustrating how the dynamic drawdown of the region amplified sea level rise with maximum flooding occurring across northern Africa during the Cenomanian and Turonian times ($\sim$100–90 Ma; increasing base level), followed by a slow transition into a region-wide regression that corresponds to an decrease in

Figure 5. (a, c, e, and g) Paleoenvironment indicators, where 0 denotes water, and 1 denotes land, based on paleogeographies by Golonka [2007, 2009] in yellow and Smith et al. [1994] in light purple. Both are smoothed over a 20 Myr window so as to compare to long-wavelength dynamic topography through time. (b, d, f, and h) Evolution of the rate of base level change for Case I(s) and Haq and Al-Qahtani [2005] ($\dot{\phi}$ dark green), Case I(s) and Haq et al. [1987] ($\dot{\phi}$ pale green), Case I and Haq and Al-Qahtani [2005] ($\dot{\phi}$ dark blue), and Case I and Haq et al. [1987] ($\dot{\phi}$ cyan). Order of point locations and basins are the same as Figure 4. Landward shoreline shifts (transgressions) occur when $\dot{\phi} > 0$ and seaward shoreline shifts (regressions) occur when $\dot{\phi} < 0$. Arrows indicate the onset in the Late Cretaceous-Cenozoic regression at each point location as estimated by the PEI maps and base level calculations. Note the agreement between Case I(s) (solid arrows) and the PEI maps (solid arrows) across most point locations.
base level as early as the Late Cretaceous for Case I(s) in the Taoudeni Basin (~80 Ma; Figure 4b) and during the Eocene for Case I (~40–50 Ma; Figures 4b, 4d, 4f, and 4h).

3.2. Comparing Base Level Changes to Paleoenvironment Indicator Maps

Here we compare paleoshoreline migrations from paleoenvironment indicator maps [Golonka, 2007, 2009; Smith et al., 1994] to rates of base level change (section 2.3), here referred to as $\phi_{H05}$ and $\phi_{H87}$, and their rate of change curves, respectively. $\phi_{H05}$ and $\phi_{H87}$ (Figures 4 and 5 and S1). We use the paleoenvironment indicator curves smoothed across a 20 Myr window in the comparison, given the long-wavelength nature of the modeled dynamic topography. In the Taoudeni Basin (Figure 2g) the rates of base level change calculated for Case I indicate maximum flooding (e.g., Figure 3c) at ~91 Ma, which is 4 Myr earlier than the paleogeography of Golonka [2007, 2009] (Figures 5a and 5b). The regression (e.g., Figure 3d) calculated for Case I indicated by base level change starts at ~40 Ma, which is 40 Myr later than the early onset in regression of Golonka [2007, 2009] at ~80 Ma (Figures 5a and 5b). The rates of base level change predicted for Case I(s) better match the early onset of this regression at ~80 Ma (Figures 5a and 5b). In the Amguid Spur (Figure 2g), $\phi_{H05}$ in Case I indicates maximum flooding at ~90 Ma, which is ~10 Myr later than the paleogeography of Golonka [2007, 2009], at ~100 Ma. The peak in $\phi_{H87}$ of Case I at ~90 Ma (Figure 5e) is consistent with a maximum flooding event at this time inferred from Smith et al. [1994] (Figure 5f). A comparison between the onset in regression from the rates of base level change for Case I at ~39 Ma and the paleogeographies at ~60 and ~55 Ma from Golonka [2007, 2009] and Smith et al. [1994], respectively, indicates an ~13 Myr temporal mismatch in this basin. Again, the rates of base level change computed for Case I(s) better match this onset in regression between ~55 and 60 Ma as deduced from both paleogeography data sets (Figures 5e and 5f). In the Sirte Basin, rates of base level change for both Case I and Case I(s) (Figure 2g) show maximum flooding between ~90 Ma ($\phi_{H05}$) and 100 Ma ($\phi_{H87}$). Comparably, both paleoenvironment indicator results from Golonka [2007, 2009] and Smith et al. [1994] estimate this at 95 and 99 Ma, respectively, illustrating a temporal offset of only ~3–5 Myr. The onset in regression estimated from the base level rates of change for both Case I and Case I(s) at ~35 Ma (Figure 5d) is also consistent with the paleoenvironment indicator results of Smith et al. [1994] at ~34 Ma and ~10 Myr earlier than Golonka [2007, 2009] (Figure 5c). Finally, in the Upper Egypt Basin rates of base level change for both Case I and Case I(s) show maximum flooding at ~90–100 Ma, ~25 Myr later than that inferred from the paleogeographies of Golonka [2007, 2009] and Smith et al. [1994], at ~118 Ma (Figure 5g). Comparatively, the onset in regression (Figure 3d) from the rates of base level change for Case I at ~45 Ma (Figure 5h), against that of Golonka [2007, 2009] and Smith et al. [1994] at ~48 Ma (Figure 5g), shows a minor temporal lag of only ~2 Myr, whereas the onset of regression for Case I(s) is ~8 Myr later than the onset of regression for Case I at ~37 Ma (Figure 5h).

In summary, in the context of the timing in Late Cretaceous inundation, Case I shows the best temporal matches between rates of base level change and paleoenvironment indicators, with temporal discrepancies <10 Myr in the Taoudeni Basin, Sirte Basin, and Amguid Spur, while Case I(s) best matches paleogeography constraints in the Sirte Basin and Amguid Spur. In the far east of the study area both model matches are comparably poor in the Upper Egypt Basin (~25 Myr). In the context of the timing in the subsequent major regression, the paleoenvironment indicators importantly indicate the onset in this regression occurs inland at ~80 Ma and later in time for locations approaching the coastal regions. This is best reflected in the changes in base level calculated for Case I(s), which predicts a temporal evolution of regression from inland to coastal region of northern Africa that is very similar to that inferred from the paleogeography. Comparably, temporal discrepancy between the rates of base level change for Case I and paleoenvironment indicators is largest farther inland in the Taoudeni Basin (~40 Myr) and becomes progressively smaller approaching the coastal regions as highlighted in the Amguid Spur (~13 Myr), then Sirte Basin (>10 Myr), and finally Upper Egypt Basin (~2 Myr; Figure 5). This comparison between the base level histories predicted by Case I(s), Case I, and the paleoenvironmental indicator maps shows that Case I(s) best matches the spatial and temporal paleogeographic evolution of northern Africa.

4. Discussion

4.1. Comparing Predicted and Tomographically Imaged Mantle Structures

To gauge model reliability, we qualitatively compared predicted present-day mantle temperature for Cases I and I(s) to tomographic images (as in Shephard et al. [2014], Flament et al. [2014], and Zahirovic et al. [2012])
Figure 6. Longitudinal (P1 and P2) and latitudinal (P3 and P4) cross-sections through Case I (light green slabs) and Case II (pink slabs) across northern Africa and surroundings (inset map). Magenta dots at 5° intervals in map view correspond to magenta dots along seismic tomography cross sections. Green contours represent mantle predicted to be 5% colder than the average at a given depth, considered to represent subducted slabs.
including MIT-P08 [Li et al., 2008], GypsumS [Simmons et al., 2010], GypsumP [Simmons et al., 2010], and s40rts [Ritsema et al., 2011]. This collection of seismic tomography models was included to qualitatively account for variations between the different mantle structures imaged in these models. We took four cross sections across northern Africa through Case I and Case I(s) (Figure 6; P1–P1", P2–P2", P3–P3", and P4–P4") and qualitatively assessed the match between predicted cold anomalies (considering model slabs to be represented by mantle 5% colder than the average at a given depth) and fast seismic anomalies.

For the two north-south cross sections, both Case I and Case I(s) predict similar ancient slabs in both the upper and lower mantle at depths comparable to fast seismic anomalies in the four global seismic tomography models (Figure 6; P1 and P2). We attribute anomalies between ~1500 and 2000 km depth to Mesozoic Alpine Tethys subduction [Schettino and Turco, 2011] and anomalies down to ~1300 km to Cenozoic Alpine Tethys subduction [Schettino and Turco, 2011]. Section P1 shows the best visual match between the predicted and seismically imaged slabs (Figure 6). However, both Case I and Case I(s) do not capture the pronounced mantle anomalies tomographically imaged between 500 and 1000 km depth, extending from 10°S to 25°N along P1 and P3 in MIT-P08 (Figures 7b and 7c). Caution is warranted when interpreting the origin of these mantle anomalies (Figures 6b and 6c), as their geometries are not clearly imaged in other considered tomography models (Figure 6; P1). One possibility is that these higher-velocity anomalies (or cold downwellings) imaged beneath cratonic regions in central and western Africa reflect the effects of...
edge-driven convection developed in the upper mantle beneath the transition of thick cratonic lithosphere and thin lithosphere as described for this region in King and Ritsema [2000]. The two east-west cross sections reveal two higher-velocity anomalies at a depth of ~1500–2000 km related to the closure of the Neotethys and Vardar oceans, which ceased during the Early Cretaceous [Hosseinpour et al., 2016] (Figure 6; P3 and P4). These cross sections show how model slabs can buckle and pile as they descend deeper into the mantle [Loiselet et al., 2010], which is best illustrated in section P3 (Figure 6). These piles on section P3 are consistent with a large elongated high-velocity anomaly modeled in GypsumS, GypsumP, and s40rts and less so in MIT-P08, preserved at very similar depths of ~1500–2000 km and extending from ~5°W to beyond 50°E also linked to the subduction of the Vardar and Neotethys oceans (Figure 6; P3). Both Case I and Case I(s) predict the Vardar slab to be at ~1200–2000 km depth, between ~5°W and 15°E, and the Neotethys slab at depths of ~1500–2000 km, from ~20°E and extending eastward beyond 50°E. These model slabs are in agreement with two prominent seismic anomalies in MIT-P08 at similar depths, and to a lesser extent in GypsumS, GypsumP, and s40rts, which image the elongated structure of P3 (Figure 6; P4).

Overall, this qualitative comparison does not allow us to quantify the subducted masses at present-day depths; however, it suggests that both Case I and Case I(s) reasonably predict cold thermal anomalies, representing major subducted slabs, at similar depths to fast seismic velocity anomalies as imaged in several different seismic tomography models. This comparison lends support to the combined imposed subduction history and mantle rheology that underpins the modeled evolution of dynamic topography used in this study (section 2.2.3).

4.2. The Paleogeography of Northern Africa Since the Mid-Cretaceous
4.2.1. The Interplay Between Mantle Dynamics and Sea Level Across Northern Africa
For both Case I and Case I(s), dynamic subsidence generally gradually increases in northern Africa from the mid-Aptian (~120 Ma) until the Campanian (~70–80 Ma; Figure 4; orange curve). However, Case I(s) predicts a relatively consistent low rate of dynamic subsidence since the mid-Aptain until present day for the inland Taoudeni Basin (Figure 4). We attribute the increase in dynamic subsidence across the remainder of the region to its relative northward motion [Müller et al., 2016] away from the dominant buoyant effects of the African large low shear velocity province (LLSVP), which has been imaged in seismic tomographic studies [e.g., Su and Dziewonski, 1997; Masters et al., 2000; Ishii and Tromp, 1999] and shown to provide anomalous dynamic support beneath the Africa continent [Hager et al., 1985; Lithgow-Bertelloni and Silver, 1998; Moucha and Forte, 2011] and into a zone of dynamic subsidence overriding the descent of relict Tethyan slabs (Figures 7d–7i and section 4.1). During this time the evolution of dynamic subsidence amplified the effects of sea level rise (Figure 3), resulting in the maximum inundation of northern Africa in the Cenomanian and Turonian between ~100 and 90 Ma (Figure 5). This predicted maximum inundation is consistent with the findings of Guiraud et al. [2005], who proposed the main marine transgression culminated during these times. Preserved remnants of Cretaceous marine sediments across the inland Tuareg Shield (Figure 2a; black box), now uplifted and interspersed with preserved Cenozoic volcanism, suggest subsidence and marine inundation across parts of inland northern Africa during the Cretaceous, followed by a drop in relative sea level to expose these sediments [Rougier et al., 2013]. The pattern of continental inundation is not spatially consistent across northern Africa (Figures 1 and 2). Areas of maximum flooding may reflect topographic lows that formed in proximity to regions of Early Cretaceous extension (Figure 1) [Jolivet et al. 2016; Frizon de Lamotte et al., 2015; Heine et al., 2013]. Importantly, the temporal pattern of regression derived from the paleoenvironment indicator maps indicates its onset is earliest at ~80–75 Ma for inland regions of northern Africa, commencing progressively later approaching coastal regions (Figure 5). This early onset in inland regression is best captured in the base level evolution of Case I(s) (Figures 4b and 5b), indicating dynamic topography of deep origin as a likely influence given the large spatial scale (>1000 km) of the transgression [e.g., Flament et al., 2013]. In both model cases, across the remaining parts of northern Africa, predicted dynamic subsidence rates begin to decline by the latest Cretaceous (~80–70 Ma), reaching negligible magnitudes (section 3.1) by ~30 Ma at most considered locations (Figures 4d, 4f, and 4h). This is similar to previous efforts that modeled mantle convection backward in time, limited to the past 30 Myr across all of Africa, which predicted no major changes in the evolution of dynamic topography beneath the northern part of the continent [Moucha and Forte, 2011]. From the latest Cretaceous we suggest that this steady onset in base level regression inland and decrease in dynamic subsidence rates in coastal regions (Figure 5) could reflect Africa’s slowing northward motion due to the earliest
onset of its convergence with Eurasia [Lepêrêtre et al., 2015; Jolivet et al., 2015]. We suggest that during this
time, changes in the mantle convective regime introduced a reduction in the slab-pull forces on the
African plate along its northern margin and a decrease in dynamic subsidence associated with the
Tethyan slabs beneath its northern region (Figures 5 and 7) [Jolivet et al., 2015; Jolivet and Facenna,
2000]. Model results from Case I(s) imply that sea level fall was ultimately the primary mechanism driving
shoreline regression since the Late Cretaceous as a result of decreased subsidence rates across inland
northern Africa (Figure 5b) and continued decreasing rates of dynamic subsidence along coastal regions
(Figures 5d, 5f, and 5h), which are ongoing. The late Eocene onset in inland regression predicted by Case
I is at odds with paleoenvironmental indicators (Figures 4 and 5). At present day, parts of northern
Africa’s topography have been argued to be the manifestation of regional uplift associated with short-
wave-length mantle flow expressed as a number of magmatic and amagmatic domes [Jones et al., 2012;
Ebinger and Sleep, 1998], including the Hoggar and Tibesti domes (Figure 2a) that were likely uplifted in
the last ~30 Myr [e.g., Paul et al., 2014; Rougier et al., 2013; Roberts and White, 2010]. These domes, located
in inland northern Africa, are proposed to have uplifted ~30 Myr after the onset of regression as shown by
paleoenvironment indicators across this inland region (latest Cretaceous; Figure 5a), suggesting that the
inland regional regression cannot be reconciled with the shallow mantle processes responsible for their
topography. Second, these domes are too localized (<~300 km) to explain this regional (>1000 km) regres-
sion, but their uplift could have contributed to the Cenozoic coastal regression as a second order process.
Other documented tectonic processes occurring contemporaneously with this regression have previously
been linked to playing a part in the evolution of paleogeography across isolated parts of northern Africa
[e.g., Spasojevic and Gurnis, 2012]. These include (1) compression in the Atlas Mountains interpreted to have
occurred in two phases in the late Eocene (~40 Ma) and Pleistocene (~3 Ma) [Frizon de Lamotte et al., 2000,
2009] (Figure 1a); (2) tectonic inversion along the southern coast of the East Mediterranean [Frizon de
Lamotte et al., 2011]; (3) extension in the Sirte Basin from ~70 Ma to ~50 Ma [Capitanio et al., 2009]
(Figure 2g), interrupted by a short phase of Eocene uplift proposed to be related to intraplate stresses pro-
gressed from changes in the kinematics of the Atlantic [Van Der Meer and Cloetingh, 1993]; (4) extension
in the Gulf of Aden-Red Sea, proposed to have initiated in the early to middle Oligocene (~30 Ma) [Fantozzi
and Sgavetti, 1998; Watchorn et al., 1998] (Figure 2a); and (5) a major compressional event in the early late
Eocene along the African-Arabian plate margin extending beneath the Red Sea and Gulf of Aden [Guiraud
et al., 2005] (Figure 2a). These tectonic events are likely to have played a secondary role, as the concentra-
tion of deformation during the Eocene, Oligocene, and Miocene, as listed above, cannot be reconciled with
the paleoenvironmental indicator maps. The earliest onset in this regression during the Late Cretaceous
across inland northern Africa temporally matches the extension in the Sirte Basin. However, we argue that
the small spatial extent of this basin and its distance from the inundated regions of northern Africa (Figure 2g)
make the link between the two events implausible. We also acknowledge the similarity in timing between
the brief uplift in the Sirte Basin associated with compression related to intraplate deformation during the
Eocene [Van Der Meer and Cloetingh, 1993] and the timing in the coastal regions regression. However, the
spatial extent of this basin cannot be reconciled with the extent of this regression (Figure 2g) nor the time-
dependent evolution of this regression, which commenced in the Late Cretaceous across inland northern
Africa. Furthermore, the modeling approach does not account for any vertical motions related to
concurrently evolving tectonic processes; therefore, our understanding of how they interact with mantle con-
vective flow to modulate Earth’s surface topography remains poor (e.g., propagation of intra plate stresses;
section 2.2). The results suggest a scenario in which the interplay between a fall in eustasy and changes in
mantle convection, occurring on spatial scales of 10^3 km, can account for the large-scale regional paleoge-
graphic evolution of northern Africa since the mid-Cretaceous.

4.2.2. Comparing Base Level Change for Different Model Cases to the Paleoenvironment
Indicator Maps

When comparing the base level evolution of Case I and Case I(s) and the paleoenvironment indicator time
series, we consider a maximum temporal mismatch of up to ~10 Myr in the timing of transgressions and
regressions to be acceptable given the coarse, variable temporal resolution of the original paleogeographic
maps (see section 2.1). Discrepancies between the onset in regression of Case I and the paleoenvironment
indicator evolutions within inland regions of northern Africa suggest that the inclusion of active mantle
upwellings in the modeling approach results in an overprediction of modeled dynamic topography as seen in
the comparison between the dynamic topography histories of Case I (Figures 7d–7f) and Case I(s)
against the paleoenvironment indicator evolutions (~25 Myr) along coastal regions, and because it uses a model in our main comparison as its very low rate of dynamic subsidence results in large temporal offsets. We do not consider this model in our main comparison as its very low rate of dynamic subsidence results in large temporal offsets against the paleoenvironment indicator evolutions. With the exception of Case VI, comparing the remaining model cases to the paleoenvironment indicator evolutions shows that varying the plate reconstructions and model start age also gives similar <10 Myr temporal offsets in the onset in major regressions along coastal regions, with the largest offsets for the onset in this regression further inland. The large temporal difference inland is attributed to active upwellings included in these additional models, similar to Case I (Figure S1). Case VI predicts the lowest estimate of dynamic topography across all point locations through time (Figure 4). Compared to Case V, which has the same model initiation age and density of the basal thermochemical layer, Case VI has a comparatively younger ocean floor in the Tethys (~140 Ma) linked to the reconstructions of TR2, compared to that of TR1 (section 2.2.3) used in Case V. As the age of the subducting oceanic lithosphere controls the buoyancy of sinking slabs, younger ages result in less negatively buoyant slabs; therefore, the net flux of cold material into the deeper mantle is reduced and so is the return flow of warm material toward the surface, resulting in the reduced magnitude of dynamic topography. We do not consider this model in our main comparison as its very low rate of dynamic subsidence results in large temporal offsets against the paleoenvironment indicator evolutions (~25 Myr) along coastal regions, and because it uses a relatively young model start age (140 Ma). Case III also confirms that the topography evolution of Case VI would be much closer to Case I if the model start age was 230 Ma because Case III uses the same model parameters as Case VI with the exception of an earlier model start age at 230 Ma.

The finer spatial resolution of the paleoenvironment indicator maps captures the progressive change in paleoshorelines, distinguishing clearly between the earlier onset in transgression and regression for regions of northern Africa (e.g., Taoudeni Basin; Figure 5a, or Upper Egypt Basin; Figure 5g) and later onsets (Amguid Spur, Figure 5e) in coastal regions (Sirte Basin and Upper Egypt Basin, Figures 5c and 5g). This indicates that base level histories are distinguishable between inland and coastal regions, as shown for the base level evolution of Case I(s). This is an important observation in the modeling approach described here, as it suggests that deep Earth, long-wavelength mantle convection was the primary driving force in the paleogeographic evolution of this region, modulated by eustatic sea level change.

5. Conclusions

The interplay between the evolution of modeled dynamic topography and global sea level fluctuations could explain the continent-wide Late Cretaceous inundation and then regression of northern Africa. Our preferred mantle flow model results suggest that the long-wavelength dynamic topography, arising from sources of buoyancy deeper than 350 km, of inland northwest Africa remained relatively low since the mid-Cretaceous, while toward the east and to the north along its coastal regions, subsidence rates increased until the Campanian. This is consistent with the region’s northward motion away from the buoyant effects of the African Superswell and into a region overriding relict Tethyan slabs. Comparing predicted changes in base level for our preferred model case with two sets of paleogeographic maps indicates that during the Cenomanian and Turonian this dynamic subsidence amplified the effects of a large first-order sea level rise by drawing down significant parts of northern Africa, resulting in widespread transgression. Shortly after this time, during the Late Cretaceous, inland parts of northwest Africa began to undergo a regression, reflecting the interplay between an onset in sea level fall and already low rates of dynamic subsidence beneath this part of the region. Across the remaining parts of northern Africa, the progressive onset of this regression toward coastal regions through time reflects the decline in dynamic subsidence rates as a response to Africa’s slowing northward plate motion driven by its early convergence with Eurasia and its increasing distance from the buoyant effects of the African LLSVP. By the early Cenozoic, across coastal regions, continued decreasing rates of dynamic subsidence allowed a pronounced global sea level fall to dominate shoreline migrations, driving an ongoing region-wide regression. This explanation can suitably account for the Cretaceous transgression recorded across northern Africa, except in its far east, and can also account for the progressive onset in regression inferred in both inland and coastal regions of northern Africa since the Late Cretaceous. A qualitative comparison between the predicted depths of model slabs at present-day and several tomography models shows good agreement in general, suggesting that model results are robust. The paleogeographic
evolution of northern Africa since the Late Cretaceous has primarily been influenced by the interplay between deep Earth, long-wavelength mantle convection, and changes in eustatic sea level. We propose that the large differences in the observed inundation history between northern Africa and the rest of the continent would not exist for this period in the absence of the dynamic subsidence predicted by the presented models.

References


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