Continental Margin Subsidence From Shallow Mantle Convection: Example From West Africa

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Abstract

Spatial and temporal evolution of the uppermost convecting mantle plays an important role in determining histories of magmatism, uplift, subsidence, erosion and deposition of sedimentary rock. Tomographic studies and mantle flow models suggest that changes in lithospheric thickness can focus convection and destabilize plates. Geologic observations that constrain the processes responsible for onset and evolution of shallow mantle convection are sparse. We integrate seismic, well, gravity, magmatic and tomographic information to determine the history of Neogene-Recent (< 23 Ma) upper mantle convection from the Cape Verde swell to West Africa. Residual ocean-age depths of $+2$ km and oceanic heat flow anomalies of $+16 \pm 4$ mW m⁻² are centered on Cape Verde. Residual depths decrease eastward to zero at the fringe of the Mauritania basin. Backstripped wells and mapped seismic data show that 0.4–0.8 km of water-loaded subsidence occurred in a $\sim 500 \times 500$ km region centered on the Mauritania basin during the last 23 Ma. Conversion of shear wave velocities into temperature and simple isostatic calculations indicate that asthenospheric temperatures determine bathymetry from Cape Verde to West Africa. Calculated average excess temperatures beneath Cape Verde are $> +100^\circ\text{C}$ providing $\sim 10^3$ m of support. Beneath the Mauritania basin average excess temperatures are $< 100^\circ\text{C}$ drawing down the lithosphere by $\sim 10^2$ to 10^3 m. Up- and downwelling mantle has generated a bathymetric gradient of $\sim 1/300$ at a wavelength of $\sim 10^3$ km during the last ~ 23 Ma. Our results suggest that asthenospheric flow away from upwelling mantle can generate downwelling drips beneath continental margins.

Keywords: Small-scale mantle convection, Cape Verde, West Africa,
Dynamic support, Uplift, Subsidence

1. Introduction

Tomographic studies, magmatic histories and mantle flow models suggest that variations in mantle temperature, lithospheric thickness and plate motion play an important role in generating shallow mantle convection (e.g. Pekeris, 1935; McKenzie et al., 1974; King & Ritsema, 2000; Moucha and Forte, 2011; Davies & Rawlinson, 2014; Barnett-Moore et al., 2017; Hoggard et al., 2017; Müller et al., 2017). Constraining scales and evolution of sub-plate support and its expression at Earth’s surface, particularly at passive margins, has important implications for interpretation of stratigraphy, heat flow, glacio-eustasy and for parameterizing models of mantle convection (e.g. Courtney & White, 1986; Hartley & Allen, 1994; Flament et al., 2013). We combine seismic reflection and well data along the West African margin with shear wave tomography and the uplift and magmatic history of the Cape Verde Rise to constrain thermal, spatial and temporal scales of upper mantle convection.

It is generally accepted that the mantle beneath the African plate supports short ($\sim 10^3$ km) wavelength swells and basins (e.g. Holmes, 1965; Nyblade and Robinson, 1994; Burke, 1996; Crosby & McKenzie, 2006; Burke & Gunnell, 2008; Al-Hajri et al., 2009; Jones et al., 2012). Some swells are magmatic (e.g. Afar, Hoggar, Tibesti, Cape Verde), others are amagmatic (e.g. Bié, Fouta Djallon), most have positive free-air gravity anomalies, sit atop slow upper mantle shear-wave velocity anomalies and those intersecting oceanic lithosphere have residual bathymetry of up to +2 km (e.g. Courtney & White, 1986; Lithgow-Bertelloni and Silver, 1998; Pim et al., 2008; Fishwick, 2010; Ramalho et al., 2010; Jones et al., 2012; Ritsema et al., 2012; French et al., 2013; Priestley & McKenzie, 2013; Shaeffer & Lebedev, 2013; Hoggard et al., 2017). Drainage networks splay radially from many domal swells suggesting sub-plate support exerts an important control on geomorphology (Paul et al., 2014). In contrast, negative free-air gravity anomalies and ocean-age depth residuals, subsidence patterns and anomalously fast upper mantle shear-wave velocity anomalies indicate some parts of the continent (e.g. Congo basin) are convectively drawn down (e.g. Crosby et al., 2010; Jones et al., 2012; Hoggard et al., 2017). These observations show that

mantle convection plays an important role in determining vertical motions of African lithosphere.

Continental uplift histories provide information about the timescale of positive sub-plate support. For example, laterites and lateritic gravels indicate that pre-Oligocene topographic gradients in West Africa were low (e.g. Burke and Gunnell, 2008; Chardon et al., 2016). The broadly coeval growth of carbonate reefs on deltas fringing the African continent imply low rates of clastic efflux until ~ 23 Ma (e.g. Walford et al., 2005; Paul et al., 2014). Uplifted marine terraces, over-compacted stratigraphy, sedimentary flux histories and magmatism indicate post-Oligocene epeirogeny occurred in Africa (e.g. Partridge and Maud, 1987; Walford et al., 2005; Al-Hajri et al., 2009; Guiraud et al., 2010; Rouby et al., 2016). Uplift rate histories calculated by inverting drainage networks indicate that post-35 Ma uplift of topographic swells was diachronous and staged (Paul et al., 2014; Rudge et al., 2015). This work suggests that the landscapes of the African continent are largely Neogene in age and are a result of changing patterns of sub-plate support. If upwelling mantle generated regional uplift we might expect downwelling mantle to have generated broadly coeval regional subsidence patterns.

We examine the history of sub-plate support of the West African margin where hitherto ambiguous Neogene subsidence histories contain important information about thermal and spatial scales of convective drawdown. We first determine the structure of modern lithosphere and the upper mantle using a new shear wave tomographic model and gravity data. We show that uplift and magmatism of the Cape Verde Rise is broadly coeval with rapid Neogene subsidence of the Mauritania margin constrained by well and seismic data. These uplift and subsidence patterns are then combined with temperature and simple isostatic calculations to constrain thermal and spatio-temporal scales of upper mantle convection.

2. Data Processing and Methodology

2.1. Shear wave velocities

For the mantle structure, we use an updated version of the Rayleigh wave African tomography model from Fishwick (2010), which is probably the highest resolution regional tomography model available. The shear-wave velocity structure is derived by inverting waveforms at periods between 50 and 120 s for the fundamental mode and the first four overtones. The 2010 version of the model was updated to incorporate data from recent deployments. For

our region of interest, most significant are the inclusion of data from Mid Atlantic Ridge events recorded on the PICASSO (Levander et al., 2009) and Muenster-Morocco (Thomas, 2010) deployments. These provide dense path coverage in northwest Africa. Resolution tests show that velocity perturbations of a few percent at wavelengths ~ 300 km are resolvable beneath Cape Verde and the Mauritania basin (Supplementary Information). Figure 2 shows a slice at 125 km depth through the velocity model for West Africa and its surroundings. Other depth slices are given in Supplementary Figure A.12.

2.2. Temperature from shear wave velocities and isostasy

To estimate the thermal state of the mantle beneath West Africa and Cape Verde the new shear wave model was converted into temperature. Look-up charts showing temperature as a function of depth and shear wave velocity (V_s) are shown in Figure 3 for Priestley & McKenzie (2006)’s empirical parametrization and for Goes et al. (2012)’s mineral physics approach (see Supplementary Information for detailed methodology). The difference between the two parametrizations is also shown. Calculated temperatures differ by $< 200^\circ\text{C}$ for V_s values in this study (i.e. within 0.3 km s^{-1} of AK135: see Figure 3C). Figure 3D–G shows temperature profiles calculated using the two parametrizations beneath Mauritania and the Cape Verde Rise. Solutions to the analytical expressions in Section 2 of Priestley & McKenzie (2013) yield a similar lithospheric geotherm to their 2006 parametrization. The isostatic support, U , generated by excess asthenospheric temperature, $\bar{T}$, within a layer of thickness h can be calculated using

$$U = \frac{h\alpha\bar{T}}{1 - \alpha T_o}, \quad (1)$$

where T_o is the background temperature, which we set to 1330°C , and α is the thermal expansion coefficient (Rudge et al., 2008). Note that this calculation is insensitive to assumed background temperature because $\alpha \ll 1$. For example, a 100°C change in background temperature changes calculated uplift by ~ 1 m.

2.3. Ocean age-depth residuals

Residual ocean-age depth measurements help identify regions supported by sub-plate processes and can test results obtained using Equation 1 (e.g. Richards et al., 2016). The longest seismic line in our inventory extends

114 > 400 km from the West African margin onto the Cape Verde Rise (Figure
 115 4A). The seabed, sediment-basement interface and oceanic Moho are easily
 116 identified (Figure 4B). We interpret the prominent intra-crustal impedance
 117 contrasts as igneous sills. We use Hoggard et al. (2017)’s sediment and
 118 crustal thickness corrections, which incorporate a large amount of seismic
 119 data from the Atlantic Ocean, to calculate residual depths from their plate
 120 cooling model. Residual depths were measured at the points marked on
 121 Figure 4B and are plotted in Figure 1A along with Hoggard et al. (2017)’s
 122 compilation. These values are compared to estimates of dynamic topography
 123 calculated using long wavelength (800–2500 km) components of the EIGEN6c
 124 GOCE gravity field and an admittance of 35 ± 10 mGal km⁻¹ (Figure 1; Förste
 125 et al., 2014; Hoggard et al., 2017).

126 *2.4. Subsidence history* 127 *Stratigraphy*

128 To determine the history of sub-plate support we map stratigraphy using an
 129 inventory of 8 commercial wells, 1 IODP well and 38 two-dimensional seismic
 130 lines on a 10 × 5 km grid with record lengths of 8–10 seconds two-way time.
 131 The wells are located in the north, center and south of the Mauritanian basin
 132 (Figure 1B). They provide biostratigraphic age constraints and standard N
 133 and CN schemes have been used to calibrate the planktonic foraminiferal
 134 and calcareous nannoplankton assemblages, respectively. Seven prominent
 135 horizons (seabed, 5.6 Ma, 23.8 Ma, 58.4 Ma, 70.4 Ma, 89.4 Ma and basement)
 136 have been identified using well and seismic data. Figure 4C–D shows the
 137 eastern portion of the longest line with a stratigraphic interpretation based
 138 on the well and seismic data.

139 *Backstripping for subsidence histories*

140 The subsidence history of the Mauritania basin was calculated by backstrip-
 141 ping the eight wells in our inventory. First, check-shot data from each well
 142 were used to determine compaction of sediment as a function of depth, z
 143 (Figure 5). Porosity, ϕ , was parameterized using Athy’17s (1930) formula-
 144 tion,

$$\phi = \phi_0 \exp\left(-\frac{z}{\lambda}\right). \quad (2)$$

145 The initial porosity, ϕ_0 , and compaction wavelength, λ , were constrained
 146 using a simple empirical relationship between velocity as a function of depth,
 147 $v(z)$, and compaction, such that

$$\frac{1}{v(z)} = \frac{\phi(z)}{v_w} + \frac{1 - \phi(z)}{v_{sg}}, \quad (3)$$

where v_w is pore fluid velocity and v_{sg} is solid grain velocity (Wyllie et al., 1956). For seawater, $v_w = 1.5 \text{ km s}^{-1}$. We assume that $v_{sg} = 5.5 \pm 0.5 \text{ km s}^{-1}$, which is appropriate for a mixture of sand and clay (Christensen, 1982). Since two-way time, $t = 2 \int_0^z dz/v(z)$, we can combine Equations (1) and (2) to obtain

$$\frac{t}{2} = \frac{z}{v_{sg}} + \phi_o \lambda \left(\frac{1}{v_w} - \frac{1}{v_{sg}} \right) \left[1 - \exp \left(-\frac{z}{\lambda} \right) \right]. \quad (4)$$

We seek minimum root-mean-squared (rms) misfit between observed and theoretical time-depth values by varying ϕ_o and λ (Figures 5 & 6). We remove the effects of loading (i.e. decompact) by calculating thickness of sedimentary layer $z_1 z_2$ when deposited at the surface, $z_3 z_4$. At the surface $z_3 = 0$. As the solid component must remain constant during compaction we can integrate

$$\int_{z_1}^{z_2} (1 - \phi) dz = \int_0^{z_4} (1 - \phi) dz, \quad (5)$$

to obtain

$$z_4 = z_2 - z_1 + \phi_o \lambda \left[\exp \left(-\frac{z_2}{\lambda} \right) + \exp \left(-\frac{z_1}{\lambda} \right) + 1 - \exp \left(-\frac{z_4}{\lambda} \right) \right]. \quad (6)$$

z_4 is solved by the Newton-Raphson method. We calculate water-loaded subsidence assuming a fixed sea level for each well using a simple isostatic calculation,

$$S_w = S_s \left(\frac{\rho_a - \rho_s}{\rho_a - \rho_w} \right) + W_d, \quad (7)$$

where water loaded subsidence, S_w , is expressed as a function of sediment-loaded subsidence, S_s (Sclater & Christie, 1980). Asthenospheric density, $\rho_a = 3200 \text{ kg m}^{-3}$, sediment density, $\rho_s = 2400 \text{ kg m}^{-3}$, and density of water, $\rho_w = 1000 \text{ kg m}^{-3}$. We apply paleo-water depth corrections, W_d , based on benthonic biofacies (e.g foraminifera assemblages) documented in well reports. Table 1 summarizes the paleo-water depth corrections, which are consistent with Vear (2005). Water-loaded subsidence and chronostratigraphy for all wells are shown in Figure 7.

171 *Theoretical thermal subsidence*

172 We compare measured subsidence patterns to theoretical post-rift (i.e. ther-
173 mal) subsidence. The thermal subsidence curves, $S_t(t)$ in Figure 7 were
174 produced using

$$S_t = E_o r \left[1 - \exp \left(-\frac{t}{\tau} \right) \right], \quad (8)$$

175 where

$$E_o = \frac{4l\rho_m\alpha T}{\pi^2(\rho_m - \rho_w)}, \quad r = \frac{\beta}{\pi} \sin \left(\frac{\pi}{\beta} \right). \quad (9)$$

176 Parameters and their values are as follows, t = time, τ = thermal time con-
177 stant = $l^2/\pi^2\kappa$, where l = lithospheric thickness (125 km), thermal diffusivity,
178 $\kappa = 10^6 \text{ m}^2 \text{ s}^{-1}$. Mantle density, $\rho_m = 3300 \text{ kg m}^{-3}$, $\alpha = 3.3 \times 10^{-5}/^\circ\text{C}$,
179 and temperature of the asthenosphere = 1330°C (Fowler, 2006). We assumed
180 that the end of rifting (i.e. $t = 0$) occurred at 170 Ma (Vear, 2005). The
181 best-fitting stretching factors, β , are annotated in Figure 7.

182 *Solid sediment thickness*

183 We examined the spatial distribution of sedimentary rock across the Maurita-
184 nia margin using our inventory of reflection seismic lines. First, sedimentary
185 horizons were mapped in two-way time, t . Secondly, check shot data from
186 all wells were stacked and averaged to determine the best-fitting regional ve-
187 locity model: $z(t) = 0.0017t^2 + 0.9t - 34$ where t = two-way travel time in
188 ms. This model was used to calculate the depth of each horizon and solid
189 sediment thickness, T_{sol} , between seismic horizons (e.g. seabed, 5.6 Ma, 23.8
190 Ma; Figure 4) was calculated using

$$T_{sol} = z_2 - z_1 + \phi_o \lambda \left[\exp \left(-\frac{z_2}{\lambda} \right) - \exp \left(-\frac{z_1}{\lambda} \right) \right], \quad (10)$$

191 which corrects for the effects of compaction. We used the compaction param-
192 eters, $\phi_o = 0.75 \pm 0.1$, $\lambda = 2.66 \pm 1.2 \text{ km}$, that best-fit the mean time-depth
193 pairs of all check shot data in our inventory (Figure 5). z_1 and z_2 are depth
194 of the top and bottom of a seismic interval, respectively (e.g. Walford et al.,
195 2005). Solid sediment thickness of Neogene rock is shown in Figure 8.

3. Results

3.1. Tomography

The Mauritania margin sits between the Cape Verde swell and the West African Craton. The Cape Verde swell is underlain by slow mantle ($V_s < 4.4$ km s⁻¹) between 75 and 200 km depth (Figure 2 & Supplementary Figure A.12). A protuberance of fast velocities ($V_s > 4.4$ km s⁻¹) extends from the West African Craton beneath the Mauritania margin between ~ 100 and 150 km depth (see Supplementary Information). This protuberance extends beneath the stretched lithosphere of the West African margin, and is located beneath the central portion of the Mauritania basin (the region covered by the wells in our inventory). Two receiver function studies from Cape Verde have shear wave velocities that differ by > 0.5 km s⁻¹ at depths < 100 km, which makes them difficult to interpret (cf. Lodge & Helfrich 2006; Vinnik et al., 2012). Vinnik et al. (2012)’s study indicates that shear wave velocities are up to 0.4 km s⁻¹ slower than AK135 at depths of 100–200 km, which is broadly similar to the velocity structure in our regional tomographic model.

3.2. Temperature from shear wave velocities

The 1330°C isotherm calculated using Priestley & McKenzie (2006)’s and Goes et al. (2012)’s parametrizations is at a depth of 90–105 km beneath Cape Verde and 145–175 km beneath the Mauritania basin (Figure 9c). Conversion of shear wave velocities into temperature yield average values of $1486 \pm 26^\circ\text{C}$ and $1248 \pm 34^\circ\text{C}$ for a 100–200 km deep layer beneath the Cape Verde Rise and Mauritania basin, respectively. The uncertainties in calculated temperatures encapsulate the range of values determined from the Priestley & McKenzie (2006) and Goes et al. (2012) parametrizations. Uncertainties in temperatures from V_s based on uncertainties in experimental elastic and anelastic parameter values have been estimated to be about 50°C (Cammarano et al. 2003). Using a simple isostatic relationship between excess temperature of the layer and uplift yields $+643 \pm 88$ m for the Cape Verde Rise and -180 ± 114 m for the Mauritania basin (Equation 1). Calculated temperature (and isostatic support) depends strongly on the data distribution and inversion method used to derive the shear velocity model. For example, isostatic support of Cape Verde and the Mauritania basin calculated using Schaeffer & Lebedev (2013)’s shear wave model are $+466 \pm 20$ m and -827 ± 16 m, respectively. These calculations indicate that a change

231 in sub-plate support of order $+10^3$ m to -1710^3 m exists from Cape Verde
232 to West Africa.

233 *3.3. Ocean age-depth*

234 Residual ocean-age depths decrease from $+1.5 \pm 0.5$ km on the Cape Verde
235 Rise eastward to 0 km where oceanic lithosphere abuts the Mauritania mar-
236 gin (Figure 1). These observations are consistent with Courtney & White’s
237 (1986) heat flow anomalies, which decrease to normal oceanic values from
238 $+16 \pm 4$ mW m $^{-2}$ at the center of rise. The change in residual depth anoma-
239 lies is coincident with an increase in shear wave speeds from < 4.2 km s $^{-1}$
240 beneath the Cape Verde Rise to > 4.4 km s $^{-1}$ beneath the Mauritania basin
241 (Figure 2).

242 *3.4. Stratigraphy and subsidence*

243 There is at least 8 km of Cenozoic and Mesozoic sedimentary rock in the
244 Mauritania basin. Paleocene-Eocene stratigraphy is characterized by low
245 rates of clastic deposition. Apart from the northernmost part of the study
246 area, Eocene to Late Oligocene stratigraphy is dominated by extensive car-
247 bonate deposition. Proximal sedimentary rock has been eroded, which was
248 probably caused by Oligocene-Neogene uplift and erosion of continental to-
249 pography (e.g. Rudge et al., 2015; Chardon et al., 2016). This denudation
250 is most clearly expressed at well Chinguetti 4-3 where Miocene stratigraphy
251 (~ 23 Ma) is underlain by a prominent regional unconformity that marks a
252 significant change from carbonate to clastic input (Figure 4). Foraminiferal
253 assemblages indicate that Cenozoic paleo-water depth was neritic (0–150 m)
254 at proximal wells and middle bathyal (500–2000 m) at distal wells (Table
255 1; Figure 7; Leckie & Olson, 2003; Vear, 2005; Immenhauser, 2009). The
256 foraminiferal data indicate that dramatic Neogene changes in water depth
257 did not occur, which could be a result of high sedimentation rates and the
258 post-rift setting.

259 Cenomanian to Campanian water-loaded subsidence is 0.4–0.8 km for all
260 wells in the study area. Best-fitting theoretical thermal subsidence curves
261 indicate that stretching beneath all wells was modest ($\beta < 1.5$; Figure 7).
262 Cenozoic subsidence at the northern and southern fringes of the Mauritania
263 basin is very low (Figure 7). In contrast, wells in the center of the basin record
264 water-loaded subsidence of up to 0.8 km from 23.8–0 Ma (~ 0.03 mm a $^{-1}$).
265 This subsidence pattern is a significant departure from predicted thermal
266 subsidence patterns. Reflection data does not contain evidence of significant

Neogene stratigraphic growth (Figure 4). The thick Neogene sedimentary pile, mapped using the seismic dataset, indicates this subsidence pattern is regional and extends ~ 500 km from the center of the Mauritania basin (Figure 8). We estimate clastic sedimentation rates using the isopachs of solid sediment thickness shown in Figure 8. In the Mauritania basin solid sedimentary flux was high: $0.2^{+0.2}_{-0.1} \times 10^3 \text{ km}^3 \text{ Ma}^{-1}$ between 23–5.6 Ma, and $1.9^{+2.0}_{-1.4} \times 10^3 \text{ km}^3 \text{ Ma}^{-1}$ from 5.6–0 Ma. The sedimentary flux errors were calculated by propagating uncertainties in compaction parameters. A Neogene increase in clastic deposition is reported elsewhere in Africa (e.g. Paul et al., 2014).

4. Discussion

Cenomanian to Campanian subsidence occurred across the West African margin. Wells at the northern and southern fringe of the basin indicate that Neogene subsidence was < 50 m (Figure 7). In contrast, stratigraphy in the center of the Mauritanian basin records up to 0.8 km of Neogene subsidence. Interpretation of a two-dimensional seismic reflection dataset shows that Neogene subsidence occurred over an area of $\sim 250 \times 10^3 \text{ km}^2$ (Figure 8). Seismic data does not contain evidence of significant Neogene stratigraphic growth, which indicates that this phase of subsidence was unrelated to stretching or salt movement (Figures 1 and 4). Neogene subsidence is an order of magnitude larger than, and the opposite sign to, estimated glacio-eustasy (Figure 7; Haq et al., 1987; Miller et al., 2005).

The backstripping performed (e.g. Figure 7) assumes that the Mauritania margin has negligible elastic strength, which is consistent with Watts (1994)’s suggestion that lithosphere along parts of the northwest African margin is weak ($T_e < 5$ km). To assess the impact of loading by Cape Verde on subsidence of the Mauritania margin we estimated the position of maximum uplift (i.e. flexural forebulge, $x_b = \pi\alpha_f$), where flexural parameter, α_f , is a function of elastic thickness, T_e . We assume an infinite elastic plate and that the load is concentrated at Boa Vista, ~ 700 km away from the backstripped wells in the Mauritania basin. The flexural parameter is a function of flexural rigidity, D , thus

$$\alpha_f = \left[\frac{4D}{g(\rho_m - \rho_w)} \right]^{1/4}, \quad D = \frac{ET_e^3}{12(1 - \sigma^2)}. \quad (11)$$

299 We set mantle and water densities to $\rho_m = 3300$ and $\rho_w = 1000 \text{ kg m}^{-3}$,
300 respectively, $g = 9.81 \text{ m s}^{-2}$. Young'17s modulus $E = 70 \text{ GPa}$ and Poisson's
301 ratio, $\sigma = 0.25$ (e.g. Turcotte & Schubert, 2002). Using Pim et al. (2008)'s
302 estimate of $T_e = 30 \text{ km}$ from wide-angle seismic data and gravity modelling
303 indicates that the crest of the forebulge is 230 km away from Cape Verde.
304 This small circle encloses the moat of negative free-air gravity anomalies
305 surrounding the islands at the center of the rise. If $T_e = 60 \text{ km}$, $x_b = 392 \text{ km}$
306 (cf. Wilson et al., 2013). If $T_e = 100 \text{ km}$, $x_b = 574 \text{ km}$. These calculations
307 show that loading from Cape Verde could not have generated the Neogene
308 subsidence observed $\sim 700 \text{ km}$ away in the Mauritania basin (Figure 8D).

309 Thus, loading, salt withdrawal, thin-skinned tectonics or glacio-eustasy
310 cannot explain the observed localized Neogene subsidence of the Mauritania
311 basin. Instead, shear wave tomography, residual ocean-age depth measure-
312 ments and long wavelength free-air gravity anomalies indicate that a gradient
313 in sub-plate support exists from Cape Verde to Mauritania (Figure 9). Calcu-
314 lated excess temperatures in the asthenosphere are large enough to generate
315 isostatic support of a few hundred meters. Note that our estimate of sub-
316 plate support does not include that generated by viscous stresses (i.e. the
317 dynamic problem). Close agreement between observed and theoretical ther-
318 mal subsidence patterns and fast shear wave velocity anomalies beneath the
319 plate indicate that sub-plate support has developed since 23 Ma. $^{40}\text{Ar}/^{39}\text{Ar}$
320 and K/Ar dated Cenozoic volcanism indicates that growth of the Cape Verde
321 swell has been sporadic since $\sim 26 \text{ Ma}$ (e.g. Mitchell et al., 1993; Holm et
322 al., 2008; Madeira et al., 2010). Growth of the swell probably precedes onset
323 of Neogene subsidence in the Mauritania basin by a few million years. These
324 observations combined with a gradient in residual ocean age depth measure-
325 ments and shear wave velocity anomalies indicate small-scale ($\sim 1000 \times 17200$
326 km) upper mantle convection beneath Cape Verde and the West African mar-
327 gin since at least 23 Ma. Neogene infilling of accommodation space, which
328 we suggest was created by mantle drawdown, has resulted in hydrocarbon
329 maturation in the Mauritania basin (Vear, 2005).

330 Models of small-scale (e.g. edge-driven) convection often invoke plate
331 motion as an important cause of mantle flow (e.g. Shahnas & Pysklywec,
332 2004; Davies & Rawlinson, 2014). However, that mechanism cannot explain
333 our observations because the African plate has been relatively stationary
334 since $\sim 30 \text{ Ma}$ (e.g. Burke, 1996). Instead, Neogene subsidence patterns,
335 the history of magmatism at Cape Verde, and tomographic and isostatic
336 modeling indicate that hot upwelling mantle and consequent asthenospheric

337 flow can trigger lithospheric instabilities at the continental margin (Figure
338 10).

339 5. Conclusion

340 We calculate the subsidence history of the Mauritania basin, offshore
341 West Africa, by backstripping 8 wells and mapping 53000-line-km of two-
342 dimensional seismic reflection data. Cenomanian–Campanian subsidence is
343 observed in wells across the basin and is followed by post-rift thermal sag.
344 Stretching factors are modest, $\beta < 1.5$. In the center of the basin an addi-
345 tional 0.4–0.8 km of Neogene water-loaded subsidence is observed. The am-
346 plitude of subsidence indicates that glacio-eustasy is not responsible for gen-
347 erating the regional increase in accommodation space observed. The wave-
348 length, location and amplitude of subsidence preclude loading as a viable
349 mechanism for generating observed subsidence. Stratal geometries in seismic
350 data intersecting wells contain little/no evidence of growth in the shallow
351 section, which indicate that Neogene salt withdrawal or thin-skinned tecton-
352 ics did not generate the observed subsidence patterns. Measured ocean-age
353 depth residuals, calculated subsidence histories, magmatic histories and a
354 new shear wave tomographic model suggest that upper mantle convection
355 has generated an east-west bathymetric gradient of $\sim 1/300$ during the last
356 ~ 23 Ma (Figure 10). Our observations are consistent with a lithospheric
357 drip below the Mauritanian part of the West African margin that formed in
358 response to the flow of hot asthenosphere away from the upwelling mantle
359 beneath Cape Verde.

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	Well	WD (m)	Cenozoic PWD	Benthonic biofacies (e.g)
A	NSO-1	79	Neritic	<i>Cibicides Lobatulus</i>
B	Loup De Mer-1	172	Upper Bathyal	<i>Gabonisporis Vigouroxi</i>
C	Labedina-1	1266	Middle Bathyal	<i>Bulmina Costata</i>
D	Chinguetti 4-3	281	Upper Bathyal	<i>Cibicidoides grosseperforatus</i>
E	Tevet-1	490	Upper Bathyal	<i>Epinooides Vicksburgensis</i>
F	Merou-1	1776	Middle Bathyal	<i>Bulmina Costata</i>
G	Tiof-4	1120	Middle Bathyal	<i>Bulmina Striata</i>
H	Flamant-1	1414	Middle Bathyal	<i>Cibicidoides Grimsdale</i>

Table 1: Modern (WD) and palaeo-water depth (PWD) of wells on Mauritania margin (Figure 1). Example is given of benthic biofacies (e.g. foraminifera) data from well reports used to determine PWD. Neritic: 0–150 m, upper bathyal: 150–500 m, middle bathyal: 500–2000 m (Immenhauser, 2009).

Figure 1: (A) Bathymetry of West African margin. Black polygon on inset globe = location of panel A. Red/black/blue contours = positive/zero/negative long wavelength (800–2500 km) free-air gravity anomalies from GOCE dataset (Förste et al., 2014); contour interval = 10 mGal. Circles/upward/downward pointing triangles = accurate/minimum/maximum residual ocean-age depth estimates (this study; Hoggard et al., 2017). Black/white diamonds = commercial/IODP-DSDP wells. X—X' = cross-section shown in Figure 9. CV = Cape Verde Rise, MB = Mauritania basin, WAC = West African craton and FD = Fouta Djallon. (B) Geological map of West African margin with location of seismic data (black polygon) and wells (diamonds). Wells: a = NSO-1, b = Loup De Mer-1, c = Labedina-1, d = Chinguetti 4-3, e = Tevet-1, f = Merou-1, g = Tiof-4, h = Flamant-1. Gray polygons = salt; black lines with tick marks = thin-skinned normal and thrust faults. Dashed lines labeled COB = continent-ocean boundary.

Figure 2: Slice through new shear wave tomographic model at 125 km depth. Color scale = absolute shear wave velocities. Red/black/blue contours = positive/zero/negative long wavelength (800–17172500 km) free-air gravity anomalies from GOCE dataset; contour interval = 10 mGal. Circles and upward/downward pointing triangles = residual ocean-age depth measurements and minimum/maximum estimates (this study; Hoggard et al., 2017). Black/white diamonds = commercial/IODP-DSDP wells. X–X' = cross section shown in Figure 9. Dashed line encloses region of fast velocities anomalies ($> 4.4 \text{ km s}^{-1}$). Note protuberance of fast velocities beneath the Mauritania basin.

Figure 3: Look-up charts showing temperature as a function of depth and shear wave velocity (V_s) for (A) Priestley & McKenzie (2006)'s empirical parametrization (T_{PM06}). (B) Goes et al. (2012)'s mineral physics parametrization with anelastic velocity correction (T_G). (C) Difference between the two parametrizations. Gray line = global average (modified AK135 from Schaeffer & Lebedev, 2013); dashed lines = global average ± 0.3 km s⁻¹. (D) Black curve = absolute shear wave velocities, V_s , beneath Mauritania from new tomographic model. Gray line = global average. (E) V_s converted to temperature as a function of depth using Priestley & McKenzie (2006)'s parametrization (black line) or the Goes et al. (2012) parametrization corrected for anelasticity (gray line). Circles = 1330°C used to map lithosphere-asthenosphere boundary (see Figure 9). Dashed line = dry peridotite solidus (Fitton et al., 1991). (F–17G) Velocities and calculated temperatures beneath the Cape Verde Rise.

Figure 4: (A) Seismic dip-line 1212-NWA from TGS dataset (see inset map). Imagery is generally excellent away from salt. Black circles = location of residual ocean age depths measurements. (B) Interpreted seismic line. Chronostratigraphy is based on seismic and well data (see body text). Note this line intersects oceanic lithosphere. (C-D) Eastern portion of line. Note thick Miocene to Recent sedimentary rock with no/minor stratigraphic growth.

Figure 5: (A-H) Check shot data for wells on the Mauritania margin (Figure 1). Best-fitting compaction parameters (initial porosity, ϕ_0 , and compaction wavelength, λ) were determined by minimizing misfit between observed check shot data and calculated time-depth curves (red lines; Figure 6).

Figure 6: (A-H) Compaction parametrization. Root-mean-squared (rms) misfit between observed and calculated time-depth curves plotted as a function of initial porosity, ϕ_0 , and compaction decay length, λ . Red crosses = compaction parameters with lowest residual misfit to time-depth data (Figure 5); dashed lines = contours of misfit used to estimate compaction uncertainties.

Figure 7: (A-H) Cenozoic subsidence of Mauritania margin. Solid lines = decompacted, water-loaded subsidence; line thicknesses = compaction uncertainties (< 50 m). Circles = decompacted, water-loaded subsidence corrected for palaeo-water depth (error bars); dashed-lines = best-fitting thermal subsidence profiles (see body text). Red lines = anomalous Neogene subsidence (0.03 mm a⁻¹). (I-J) Gray/black curves = Haq/Miller eustatic curves (Haq et al., 1987; Miller et al., 2005).

Figure 8: Neogene solid sediment thickness of Mauritania margin and flexural loading of Cape Verde. (A/B/C) Solid thickness between 23.8–5.6/5.6–0 Ma/23.8–0 Ma. (D) Flexure of an infinite elastic plate for a point load centred on Boa Vista (solid circle). Colour scale = free-air gravity data from EIGEN6c GOCE dataset (Förste et al. 2014). Dashed lines = position of maximum uplift (forebulge) for $T_e = 30$ km or 60 km (cf. Pim et al., 2008; Wilson et al., 2013). Black diamonds = backstripped wells on West African margin. White diamonds = IODP & DSDP wells.

Figure 9: Summary of results along cross-section X—X' (see Figures 1 & 2). (A) Circles and downward pointing triangles = residuals from ocean-age depth projected onto line. Diamonds = Neogene–Recent (23.8–0 Ma) subsidence calculated from wells in the centre of Mauritania basin. Gray band and black line = dynamic topography calculated using long wavelength (800 ± 2500 km) free-air gravity anomalies from GOCE dataset and an admittance of 35 ± 10 mGal km⁻¹. Arrow = continent-ocean boundary (COB) mapped using reflection seismic data. (B) Topographic swath and thickness of sedimentary rock in Mauritania basin: solid/dashed/dotted white lines = 58.4/23.8/5.6 Ma horizons (Figure 4). (C) Surface wave tomography. Dash-dot/dash lines = 1330°C isotherm calculated using (P) Priestley & McKenzie (2006)/(G) Goes et al. (2012) parametrizations. (D) Squares = calculated average temperature between 100–1717200 km depth. Error bars = range from P and G parametrizations. (E) Black squares = isostatic support calculated using excess temperatures (Equation 1).

Figure 10: Cartoon summarizing results (see Figures 1 and 9 for annotation). Black circle beneath Cape Verde = Moho estimated from wide-angle seismic data (Pim et al., 2008). Arrows indicate shallow mantle flow.

506 **Appendix A. Tomography model**

507 Fishwick (2010)’s surface wave tomography model was updated to incor-
508 porate data from networks throughout Africa. Inclusion of data from Mid
509 Atlantic Ridge events recorded on the **PICASSO** (Levander et al., 2009) and
510 **Muenster-Morocco** (Thomas, 2010) deployments provide dense path coverage
511 across the Northwest African margin. Waveforms with low signal-to-noise
512 ratios between periods of 50–120 s were removed from the dataset prior to
513 inversion. Path coverage used to constrain the tomographic model is shown
514 in Figure A.11A-B. The resolution of the tomographic model was examined
515 at 150 km depth by inverting synthetic data constructed using the checker-
516 board velocity model shown in Supplementary Figure A.11C. Path average
517 velocities were calculated using the inventory of African paths shown in Sup-
518plementary Figure A.11A and noise was added to the synthetic dataset com-
519mensurate with uncertainties in the real data. Supplementary Figure A.11D
520 shows the recovered velocity structure obtained by inverting the noisy syn-
521thetic data. Importantly for our purposes, velocity perturbations of a few
522 percent at wavelengths of ~ 300 km are resolvable beneath Cape Verde and
523 the Mauritania basin. Supplementary Figure A.12 shows the velocity model
524 for West Africa and its surroundings.

Figure A.11: (A) Path coverage (gray lines), seismic stations (triangles) and events (circles) used to constrain African tomographic model. (B) Coverage across Mauritania margin. (C–D) Recovery of synthetic velocity perturbations at 150 km depth. (C) Checkerboard pattern of input velocity perturbations of $\pm 6\%$ at 4° separation. (D) Recovered velocity model from inversion of synthetic data calculated using path coverage shown in panel A and noise commensurate with uncertainties in real data at 150 km depth. Dashed line encloses non-zero synthetic velocity perturbations in input.

Figure A.12: Slices through new shear wave tomographic model from 75–200 km depth. Color scale = absolute shear wave velocities. See Figure 2 for annotation. Note protuberance of fast velocities beneath the Mauritania basin (arrows in panels B and C).

525 Appendix B. Temperature from shear wave velocities

526 We estimate mantle temperatures from shear waves using three techniques
527 adapted from Priestley & McKenzie (2006, 2013) and Goes et al. (2012).

528 Appendix B.1. Priestley & McKenzie (2006)

529 Priestley and McKenzie (2006)’s approach uses mineral-physics based ex-
530 pressions, but with parameters chosen to reconcile tomographic velocities
531 with thermal constraints from the oceanic plate cooling model and cratonic
532 xenoliths. In their updated conversion they show that parameter values
533 may need to vary depending the global tomographic model used (Priestley
534 & McKenzie, 2013). We generate a look-up table of temperatures, T , as a
535 function of pressure, P , and shear wave velocities, V_s , using Equation 5 in
536 Priestley & McKenzie (2006). We assume that pressure is lithostatic (i.e.
537 $P = \rho g z$, where $\rho = 3300 \text{ kg m}^{-3}$ is density, $g = 9.81 \text{ m s}^{-2}$, $z = \text{depth}$)

$$V_s = \{1 + b_v[z - 50]\} \left\{ mT + c + A \exp \left[-\frac{E + PV_a}{R(T + 273.15)} \right] \right\}, \quad (\text{B.1})$$

538 where $b_v = 3.84 \times 10^{-4} \text{ km}^{-1}$, $m = -2.8 \times 10^{-1} \text{ m s}^{-1} \text{ }^\circ\text{C}^{-1}$, $c = 4720 \text{ m}$
539 s^{-1} , $A = -1.8 \times 10^{16} \text{ m s}^{-1}$ are empirically determined constants. Activation
540 energy $E = 409 \times 10^3 \text{ J mol}^{-1}$, activation volume $V_a = 10^{-6} \text{ m}^3 \text{ mol}^{-1}$ and
541 the gas constant $R = 8.3145 \text{ J mol}^{-1} \text{ K}^{-1}$. Solutions to Equation (B.1) can
542 be easily recast to extract $T(P(z), V_s)$.

543 Appendix B.2. Priestley & McKenzie (2013)

544 A modified approach based on McCarthy et al. (2011)’s scaling of seismic
545 wave frequencies, f , by Maxwell relaxation times, τ , is given in Priestley &
546 McKenzie (2013). The Maxwell time scale $t = \eta/\mu$, where η and μ are vis-
547 cosity and the shear modulus, respectively. A dimensionless frequency is ex-
548 pressed as $f' = f\tau$ and expressions for $V_s(f')$ can then be obtained. Priestley
549 & McKenzie (2013) combine this approach with an empirical parametriza-
550 tion similar to that in their 2006 paper. Combining their Equations 4–7, 9,
551 10 and 12

$$V_s = \sqrt{\rho^{-1} \mu(P, T) F(f')}, \quad (\text{B.2})$$

552 where

$$\mu(P, T) = \mu_0 + (\partial\mu/\partial T)T + (\partial\mu/\partial P)P, \quad (\text{B.3})$$

$$F(f') = \begin{cases} \sum_{k=0}^6 a_k (\ln f')^k & \text{for } f' \leq 10^{13} \\ 1 & \text{for } f' > 10^{13} \end{cases} \quad (\text{B.4})$$

a_k were determined empirically by McCarthy et al. (2011): $a_0 = 0.55097$,
 $a_1 = 0.054332$, $a_2 = -0.0023615$, $a_3 = -5.7175 \times 10^{-5}$, $a_4 = 9.9473 \times 10^{-6}$,
 $a_5 = -3.4761 \times 10^{-7}$, $a_6 = 3.9461 \times 10^{-9}$. We tested seismic frequencies $f =$
0.01 to 0.05 Hz and calculated the dimensionless frequencies using $\tau = \eta/\mu$,
where $\eta = \eta_0/a^*$, $a^* = \exp[(E + P_r V_a)/RT_r]/\exp[(E + P V_a)/RT]$. Values
for the following parameters are from their Table 1: $\mu_0 = 72.66 \times 10^9$ Pa,
 $\partial\mu/\partial T = -0.871 \times 10^7$ Pa, $\partial\mu/\partial P = 2.04$, $\log_{10} \eta_0 = 22.38 \log_{10}(\text{Pa s})$,
 $E = 402.9 \times 10^3$ J mol⁻¹, $V_a = 7.81 \times 10^{-6}$ m³ mol⁻¹, $R = 8.3145$, $T_r = 1473$
K, $P_r = 1.5 \times 10^9$ Pa. Note that Equations 18–22 in their paper can also be
used to determine $V_s(P, T)$.

We find that calculated lithospheric geotherms are similar to those from
their 2006 parametrization. However, calculated sub-lithospheric tempera-
tures are significantly greater. Takei et al. (2014)’s experimental results,
which extend the frequency range of McCarthy et al. (2011)’s experiments
to incorporate seismic frequencies, indicate that $T(V_s, P)$ determined using
Priestley & McKenzie (2013)’s results are probably too high at low wave
speeds (see also Richards et al., 2016). For these reasons we use their earlier
parametrization, which yields broadly similar results to Goes et al. (2012)’s
approach.

Appendix B.3. Goes et al. (2012)

Goes et al. (2012)’s parametrization of V_s was produced for a pyrolytic
composition. We used an attenuation model similar to their Qg and assumed
a wave frequency of 0.05 Hz. The velocity to temperature conversion includes
the effects of anelasticity. V_s is calculated as a function of temperature,
pressure, composition (X) and frequency

$$V_s(T, P, X, f) = V_{\text{anh}}(T, P, X) \left(1 - \frac{Q_g^{-1}(f, T, P, C_{OH})}{2 \tan(\pi\alpha/2)} \right), \quad (\text{B.5})$$

where

$$Q_g(f, T, P, C_{OH}) = A_g \left(\frac{C_{OH}^{\text{ref}}}{C_{OH}} \right)^s f^\alpha \exp \left(\frac{\alpha \gamma T_m}{T} \right). \quad (\text{B.6})$$

579 C_{OH} is hydration and C_{OH}^{ref} is a reference hydration, in this case we assume
 580 a dry mantle, $C_{OH}^{\text{ref}} = 50 \text{ H}/10^6 \text{ Si}$. Exponent s is determined empirically,
 581 $A_g = 0.1$, $\alpha = 0.15$, $\gamma = 38$. T_m is melting temperature; T_m/T is the inverse
 582 of the homologous temperature. Note that $\gamma T_m = E + PV_a$ (Goes et al.,
 583 2012). The anharmonic velocities, V_{anh} , are determined using Goes et al.
 584 (2012)'s thermodynamic approach. These values were used to generate the
 585 look-up chart shown in Figure 3b.