Supplementary information to “Coherent high- and low latitude control of the Northwest African hydrological balance”

Regional setting core GeoB7920-2

Sediment core GeoB7920-2 (20°45.09’N-18°34.90’W, 2278 m water depth) was retrieved from the Northwest African margin off Cape Blanc during FS Meteor cruise M53/1 [1]. At present, the Northwest African arid and semi-arid regions form the world’s largest dust source that exports between 130·10^6 to 700·10^6 t/yr [21 and references herein]. Core GeoB7920-2 is located directly under the major African dust track [2] and receives therefore large amounts of windblown dust from the Sahara and the northern Sahel [3-6]. Off Cape Blanc, aeolian dust is thought to be predominantly transported by the year-round active surface trade winds [7], which also induce upwelling of nutrient rich sub-surface waters over the continental shelf and high surface water productivity [3]. Additionally, high atmospheric aerosols transported dust during boreal summer that originates from more inland sub-Saharan and northern Sahel dust sources [2]. However, the dust load of these high atmospheric aerosols is thought to largely by pass the proximal Northwest African margin to be transported much further into or even across the Atlantic Ocean [7]. The high dust flux and high biogenic productivity in surface waters result in high (~15 cm/ka) sediment accumulation rates at the location of core GeoB7920-2.

Present-day precipitation rates over Northwest Africa are controlled by the seasonal migration of the Intertropical Conversion Zone (ITCZ) and the associated surface trade-wind systems (Fig. 1). The ITCZ reaches its northern most position (20–22 °N) during boreal summer, generating a monsoonal flow over the African continent south of the ITCZ [8]. Precipitation in Mediterranean Africa and the Northwest African coastal region derives from storm tracks of the North Atlantic westerlies during boreal winter, when the ITCZ reaches its southernmost position [8].

Chrono stratigraphy

The age model of sediment core GeoB7920-2 is constructed by visual correlation of the benthic (Cibicidoides wuellerstorfi) δ¹⁸O record with that of marine sediment core MD95 2042 [9-11]. These cores are located relatively close to each other and both cores are situated in North Atlantic Deep Water (NADW). The peak-to-peak correlation was performed on the smoothed (5-point-average) δ¹⁸O records (Fig. SF1a) with the software package AnalySeries 1.1 [12]. We used the δ¹⁸O stratigraphy of MD95 2042 on the GRIP ss09sea age scale [10]. The age scale of the last ~15 ka was cross checked by comparison of the carbonate
stratigraphies of core GeoB7920 and the parallel ODP Site 658C (20°45’N, 18°35’W, 2263 m water depth) (Fig. SF1b). Site 658C contains a very well established age model consisting of 30 AMS dates on the planktonic foraminifera *G. bulloides* and *G. inflata* [13, 14]. The calibrated ^14C AMS data of Site 658C have been adjusted for a 500-year reservoir correction due to the upwelling of older surface water masses at this site [13]. There are no indications of a hiatus or a period of winnowing between 14.5 and 17.5 ka in GeoB7920-2 as have been found for Site 658C [3, 13, 14].

Figure SF1. Age model of core GeoB7920. a, Visual correlation (correlation points are indicated by black lines) of the 5-point average δ18O records (red line) of sediment core GeoB7920-2 and MD95 2042 [9-11] b, correlation of the carbonate records of GeoB7920 (red line) and Site658C (black line) [3, 13] for the last ~15 ka. The ^13C AMS age control points of Site 658C are indicated with black triangles [13, 14].

End Member Modelling

Grain-size distributions of the siliciclastic fraction of marine sediments from continental margins represents a mixture of sediment populations derived from different sources and/or transported by different mechanisms to the site of deposition. The end-member-modelling algorithm (EMMA) of *Weltje* (1997) can decompose measured grain-size distributions into
proportional contributions of an optimal set of end members [15-17]. Unlike principal component or factor analysis that express dimension less values, the end members are expressed in physical terms (µm) [16]. The EMMA assumes that the end members are linearly independent, i.e. end members can not be express as a mixture of other end members, and the grain-size distribution of the end members are fixed [15, 17]. End members are described as proportions of the constant-sum of the total number of end members, whereas each end member describes a clearly defined grain-size distribution with one dominant mode [18]. These so called dynamic populations [16, 17] resemble an assemblage of grains that occur together because they respond in a similar way to the sediment production and/or transport mechanisms.

Statistically evident end-members were calculated from the total set of grain-size measurements (N = 315) of core GeoB7920. The EMMA [15, 17] approximates a theoretical grain-size distribution from the analyzed grain-size distribution by iteratively calculation of the least square fit between the analyzed grain-size distribution and the mixture of the calculated end members. The minimum number of end members (EMs) required for a satisfactory approximation of the data is determined by the coefficient of determination ($R^2$) of each size-class (Fig. SF2b), and the mean coefficient of determination ($R^2_{\text{mean}}$) (Fig. SF2c) that can be reproduced by the grain-size distribution of the approximated end members [18]. In our case, a three-end-member model explains 88% of the variance of the total set of grain-size distributions of core GeoB7920 (Fig. SF2). The three end members have dominant modal grain-sizes of 57.8 µm (EM1), 34.6 µm (EM2), and 4.9 µm (EM3) (Fig SF2c), respectively.

The integrity of the three-end-member model was checked by comparing predicted grain-size distribution with the measured grain-size distributions (Fig. SF3) [19]. Therefore, we...
calculated the square root of the residual squared error (RS-error) for all size classes in the range of 0.4 to 300 µm. The RS error is given by

$$RS(x) = \sqrt{(GSD_1(x) - GSD_{EM}(x))^2}$$

were GSD_1(x) represents the measured and GSD_{EM}(x) the predicted proportion of size-class x. In general, the RS-errors (Fig.SF3c) of the grain-size distributions can be predicted with an error of less than 0.5 % for most grain-size classes.

Figure SF3. Time series of the measured and predicted grain-size distributions, and the corresponding RS-error of core GeoB792-2. a, Grain-size distributions measured with the Coulter LS200 Laser Particle Sizer. b, Grain-size distributions predicted with the three-end-member model. c, Difference between measured and predicted Grain-size distributions (RS-errors).

**End Member Interpretation**

On the Northwest African margin, sand- and silt-sized siliciclastics of marine surface sediments and aerosol dust samples have been related to aeolian transported dust [4, 5, 7, 16, 20, 21]. The coarsest siliciclastic particles have been found off Cape Blanc [5, 16, 20], where the surface trade winds are active throughout the year. The Northeast trade winds are seen as
the main supplier of aeolian dust onto the North African margin, whereas dust particles carried in higher atmospheric wind systems travel much further into, or even cross, the Atlantic basin [7]. Down wind sorting of the dust load carried by the surface trade winds cause an apparent proximal to distal fining of siliciclastic particles off Northwest Africa [5, 16, 21]. As shown in Figure SF4a, the grain-size distributions of EM1 and EM2 complement the aeolian end members calculated from the siliciclastic silt fraction of surface sediment off Northwest Africa [20]. Additionally, the grain-size distribution EM2 matches very well to that of a present-day aerosol sample collected off Cape Blanc (SF4b) [7]. Hence, the EM1 and EM2 are interpreted as coarse-grained proximal and fine-grained distal aeolian transported dust, respectively.

**Figure SF4.** Grain-size distributions of the end members calculated for sediment core GeoB7920-2, end members calculated for surface sediments from the Northwest Africa margin [20], and present-day dust collected off Cape Blanc [7]. a, Grain-size distributions EM1 (red), EM2 (orange) and EM3 (green) of core GeoB7920-2 (grey infill) and the three end members calculated from the siliciclastic silt fraction of surface sediments off NW Africa (white infill) [20]. Based on their regional distribution, the surface sediments are interpreted as aeolian dust (dotted black) and fluvial transported hemipelagic mud (green dotted). b, Grain-size distributions EM1 (red), EM2 (orange) and EM3 (green) of core GeoB7920-2 (grey infill) and a present-day dust sample (black, white infill) collected in the core region [7].

Fine-grained siliciclastic particles (< 6µm) in marine sediments off the Senegal River mouth have been related with fluvial transported sediment [5, 21, 22]. This characterization is further supported by grain-size distribution characterized as the fluvial end member (modal grain size 4.5µm), which is the abundant on Moroccan margin off Atlas Mountains rivers [20]. The grain-size distribution of this fluvial end member [20] is remarkably similar to that of EM3 in this study (SF4a). Although lateral transport over the shelf of these fine-grained particles can not be excluded, the spatial distribution of transported sediments indicates that abundant concentrations are only found nearby discharging rivers [5, 20]. Besides, there are strong indications of active fluvial discharge close to the core location from a western Sahara drainage basin at about 19°N [23-25] and smaller coastal rivers [6] during more humid
conditions. The influence of deeper ocean currents is neglected since sediment trap data indicates an almost undisturbed vertical transport of particles through the water column off Cape Blanc [26]. Moreover, siliciclastics of deep-sea sediments from the Arabian Sea and the subtropical Southeast Atlantic with similar grain-size distributions have been interpreted as fluvial transported hemipelagic mud [27, 28]. Hence, we interpret EM3 as non-aeolian hemipelagic mud associated with fluvial transported material.

Down core proportional variations of the three end members of GeoB7920-2 (Fig. SF5a) indicate abundant proportions of the coarse aeolian EM1 during the last glacial during maximum glacial conditions and stadial cold events. This has been related to intensification of the surface trade winds as a result of latitudinal shifting of climate belts and increasing meridional pressure gradient during glacial conditions [4, 21, 28]. Alternatively, exposure of the relative broad shelf off Cape Blanc during the last glacial strongly reduced the distance between the African continent and the core location. An estimation of this distance between 20° and 21° N over the past 120 ka, suggest that a distal migration of coarse-grained proximal dust can partly explain the high proportions EM1 during glacial conditions (Fig. SF5). There is no obvious relation between the calculated distance and the proportional variation of EM3 (Fig. SF5).

Figure SF5. Time series of a, proportional contribution of the end member in core GeoB-7920-2, and b, the estimated distance between GeoB-7920-2 and the African continent between 20° and 21° N indicating a minimum and maximum core-to-coast distance throughout the last glacial-interglacial cycle. The distance is represented by the interception of the present-day shelf morphology at 20° and 21° N and the global sea level curve of Wealbroeck et al. (2002) [29].
The proportions of the fluvial end-member EM3 anti-correlate with the sum of the two aeolian end members EM1 and EM2. Since the three-end-members are expressed as constant-sum proportions, their proportional variations are relative. For that matter, increasing proportions of fluvial EM3 can be explained by either enhanced fluvial discharge or decreased aeolian input due to expansion of the continental vegetation cover. Although, both scenarios indicate an increase of the continental humidity, changes in the aeolian dust supply could probably have a stronger imprint on the proportional end member contribution since it dominates siliciclastic sedimentation off Cape Blanc. However, to better account for the relative variation of all three end members we use the log ratio of the fluvial EM3 and the aeolian EM1 and EM2 as a continental humidity index (log(EM3/[EM1 + EM2])).

CLIMBER-2 model and simulations
CLIMBER-2 encompasses a 2.5-dimensional statistical dynamical model of the atmosphere, a multibasin, zonally averaged ocean model that includes sea ice dynamics, and a dynamical model of terrestrial vegetation, which are coupled by fluxes of energy, water, and momentum [30-32]. The atmospheric model has a coarse resolution of 10° in latitude and about 51° in longitude. The transport equations for temperature and humidity are solved on 10 vertical levels utilizing universal vertical profiles for temperature and humidity. The short-wave and long-wave radiation fluxes are calculated for 16 vertical layers accounting for evolving stratus and cumulus cloud coverage and average aerosol and ozone concentrations. The ocean model has zonally averaged basins for the Atlantic Ocean, the Indian Ocean, and the Pacific Ocean.
which are connected by the Antarctic circumpolar current (Fig. SF6). The meridional resolution is 2.5°, and the vertical is resolved by 20 layers. Land grid cells have fractions of glacial cover and ice-free surface. The latter is separated into forest, grass and bare soil fractions, which interactively evolve under changes in climate. CLIMBER-2 has been validated against the present-day climate, tested against comprehensive general circulation models, and used successfully for a variety of paleoclimate studies (see the list of references in the Table of EMICs [33]).

The AOV-IC simulation uses external forcing of orbital induced insolation variations [35], prescribed atmospheric CO₂ concentrations [36], and prescribed inland ice variations. At the last glacial maximum, the inland ice was assumed to cover the areas indicated in reconstructions by Peltier (1994) [37]. Changes in total ice volume were supposed to follow global sea-level changes according to Waelbroeck et al. (2002) [29, 38], which have been tuned to the orbital timescale of Martinson et al. [39] for this study. The area covered by the ice sheets was assumed to follow ice volume being scaled by a power of 2/3. This relatively simple assumption implies a fast change in area covered by the ice sheets when the ice volume is small and a slower change with increasing ice volume, which is in line with modeling results by Calov et al. [40]. Given the coarse geographical resolution of CLIMBER-2, this simple scenario of past inland ice variations can be viewed as sufficiently realistic. As in AOV-IC, the AOV-IC-f simulation uses orbital induced insolation variations [35] and prescribed atmospheric CO₂ concentrations [36]. Besides changes in inland ice, the AOV-IC-f simulates uses an additional freshwater forcing that triggers Dansgaard/Oeschger events and Heinrich events (not introduced in AOV-IC) [41]. The ice sheet at the LGM was prescribed according to Peltier (1994) [37], which corresponds to lowering of ~105 meters with respect to the present-day global sea level (Fig. SF7). The global inland ice volume at 60 ka BP was estimated by a global sea level lowering of ~50 meters with respect to the present-day sea level, which is in general agreement with reconstructions of global sea level change [29]. The volume of the Laurentide and Fennoscandian ice sheets between 60 ka BP and the LGM were varied as depicted in Fig. SF7. This rather schematic variation of ice sheets implies that the Eurasian ice volume almost vanishes at 60 ka BP which might be unrealistic. However, for the scope of this study of the effect of Heinrich events and Dansgaard/Oeschger events on North African climate this simplification is warranted. In the model, Heinrich events and Dansgaard/Oeschger events are triggered by pulses of freshwater input into the North Atlantic (see below), not by a steady trend in inland ice volume and associated change in sea level. The trend in inland ice volume is prescribed to introduce a slow, steady cooling towards the
Glacial Maximum. This cooling affects the frequency of Dansgaard/Oeschger events, but has only a marginal impact on their strength. Ice volume changes of the Laurentide ice sheet due to ice surge events (Heinrich events) were simulated by associated freshwater fluxes of 0.15 Sv (1 Sv = 10^6 m^3/s) into the Northern North Atlantic with a prescribed periodicity of 7500 years. Accordingly, the change in ice mass during Heinrich events led to a change in sea level of some 10 meters, which corresponds to a reduction of 750 m of the Laurentide ice sheet between 40 and 60N [41].

Figure SF7. Transient forcing to trigger D/O and Heinrich events in simulation AOV-IC-f. a, Changes in inland ice volume I in metres of sea level change (m SL) compared to today (dashed line is volume Eurasian ice sheet, solid line is global ice volume). b, Freshwater forcing in the North Atlantic in Sverdrup (1 SV = 10^6 m^3 s^-1). The top curve is the freshwater forcing to trigger Heinrich Events, and the lowest curve is a random white noise forcing at high northern latitude which triggers D/O events). The middle curve is a small sinusoidal forcing which itself is too weak to trigger D/O events, but which synchronizes D/O events via the mechanism of stochastic resonance [42-44].

Heinrich- and Dansgaard/Oeschger events in the AOV-IC-f simulations are triggered by three components of freshwater forcing, as described by Ganopolski and Rahmstorf (2001, 2002) [42, 43] (Fig SF7). All three components were applied to the North Atlantic between 50° N and 70° N with a maximum amplitude at 60° N [42, 43]. The first component prescribes regular pulses of freshwater with an amplitude of 0.15 Sv and a prescribed periodicity of 7500 years, as described above. These freshwater pulses simulate the hydrological impact of Heinrich events and cause a temporal complete shutdown of the thermohaline circulation [43]. Secondly, we introduced white-noise freshwater anomalies that trigger Dansgaard/Oeschger events. This component represents the ubiquitous internal variability of the atmosphere-ocean.
system which is not explicitly simulated in our model. The white-noise freshwater flux has a standard deviation of 0.035 Sv and is strong enough to trigger randomly varying Dansgaard/Oeschger events. Finally, the Dansgaard/Oeschger events were synchronized by a small freshwater forcing of 0.01 SV and a period of 1500 years. These small freshwater pulses cause synchronization of Dansgaard/Oeschger events via the mechanism of stochastic resonance [43], but are too weak to generate Dansgaard/Oeschger events by themselves. Although the latter freshwater pulses are hypothetic and its forcing is under debate, inclusion of such forcing allows reproducing several important features of the Dansgaard/Oeschger events known from paleoclimatic records [45]. The combination of different forcing components is used to explore the effects of Heinrich events and Dansgaard/Oeschger events on the global climate system [43].

References


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