# Oceanic link between abrupt changes in the North Atlantic Ocean and the African monsoon

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Published online: 30 May 2008; doi:10.1038/ngeo218

**Abrupt changes in the African monsoon can have pronounced socioeconomic impacts on many West African countries. Evidence for both prolonged humid periods and monsoon failures have been identified throughout the late Pleistocene and early Holocene epochs[1,2]. In particular, drought conditions in West Africa have occurred during periods of reduced North Atlantic thermohaline circulation, such as the Younger Dryas cold event[1]. Here, we use an ocean–atmosphere general circulation model to examine the link between oceanographic changes in the North Atlantic Ocean and changes in the strength of the African monsoon. Our simulations show that when North Atlantic thermohaline circulation is substantially weakened, the flow of the subsurface North Brazil Current reverses. This leads to decreased upper tropical ocean stratification and warmer sea surface temperatures in the equatorial South Atlantic Ocean, and consequently reduces African summer monsoonal winds and rainfall over West Africa. This mechanism is in agreement with reconstructions of past climate. We therefore suggest that the interaction between thermohaline circulation in the North Atlantic Ocean and wind-driven currents in the tropical Atlantic Ocean contributes to the rapidity of African monsoon transitions during abrupt climate change events.**

The onset of the African monsoon in the early boreal summer marks the strong seasonal transition of the eastern equatorial Atlantic Ocean. The sea surface temperature (SST) decreases from above 28 °C in May to below 23 °C in August. The development of the equatorial cold tongue in the eastern Atlantic is accompanied by a shoaling thermocline[3], which is forced by the seasonal change of the trade winds in response to the monsoon onset. The cooler SST in turn strengthens the land–sea temperature contrast, enhancing the monsoonal flow, which leads to a further decrease in SST[4]. Changes in SSTs can also influence surface evaporation over the equatorial Atlantic, which in turn can have an impact on monsoonal rainfall variability over West Africa[5]. This coupled feedback underscores the importance of the oceanic processes that maintain the equatorial cold tongue in regulating the African monsoon. Indeed, when these processes are occasionally interrupted by the visit of the Atlantic Niño—a sporadic warming that occurs interannually in the Gulf of Guinea—the monsoon circulation weakens and the monsoonal rains tend to be confined along the Guinean coast[5,6].

On geological timescales, Earth's orbital precession cycle, with a period of 23,000 years, has a dominant influence on the long-term African monsoon variation[1]. However, the palaeo proxy records indicate that the transition between the dry state and wet state of the African monsoon system can occur within decades to centuries[2], far shorter than the timescale of the precession cycle. Adding to the puzzle is the finding that the stronger-than-today monsoon during the African Humid Period (14.8 and 5.5 cal. kyr BP) was abruptly interrupted by the onset of the Younger Dryas event (12.8 and 11.5 cal. kyr BP) that brought the African climate back to near-glacial dry conditions[1,2,7]. The deactivation of the monsoon during the Younger Dryas event was accompanied by widespread cooling over much of the North Atlantic, extending deep into the southern Caribbean Sea, as recorded in a sediment core from the Cariaco Basin[8]. Interestingly, intense warming at intermediate water depths was recorded in cores from the neighbouring Caribbean island of Grenada[9]. The warming seemed to extend into the south tropical Atlantic. The reconstructed SST record off the coast of northeastern Brazil just south of the equator shows that the ocean surface warmed by approximately 2 °C (ref. 10) following the onset of the Younger Dryas event, whereas a record on the other side of the tropical Atlantic basin, extracted from a marine sediment core close to the Congo River mouth, shows a rapid SST increase of about 2.5 °C (refs 7,11). In contrast to the Caribbean records that show opposite sign temperature changes at the intermediate depth and at the surface, the surface warming in the south equatorial Atlantic is accompanied by a subsurface warming of 1–3 °C, as indicated by a nearby sediment core off the Angola coast[9]. These observations suggest that a major rearrangement of the ocean circulation in the tropical Atlantic took place during the Younger Dryas event, presumably caused by the abruptly weakened Atlantic thermohaline circulation[12,13] (ATHC) triggered by glacial meltwater discharge into the high-latitude North Atlantic. Can this ocean circulation change be a major contributing factor to the rapid shift in the African monsoon?

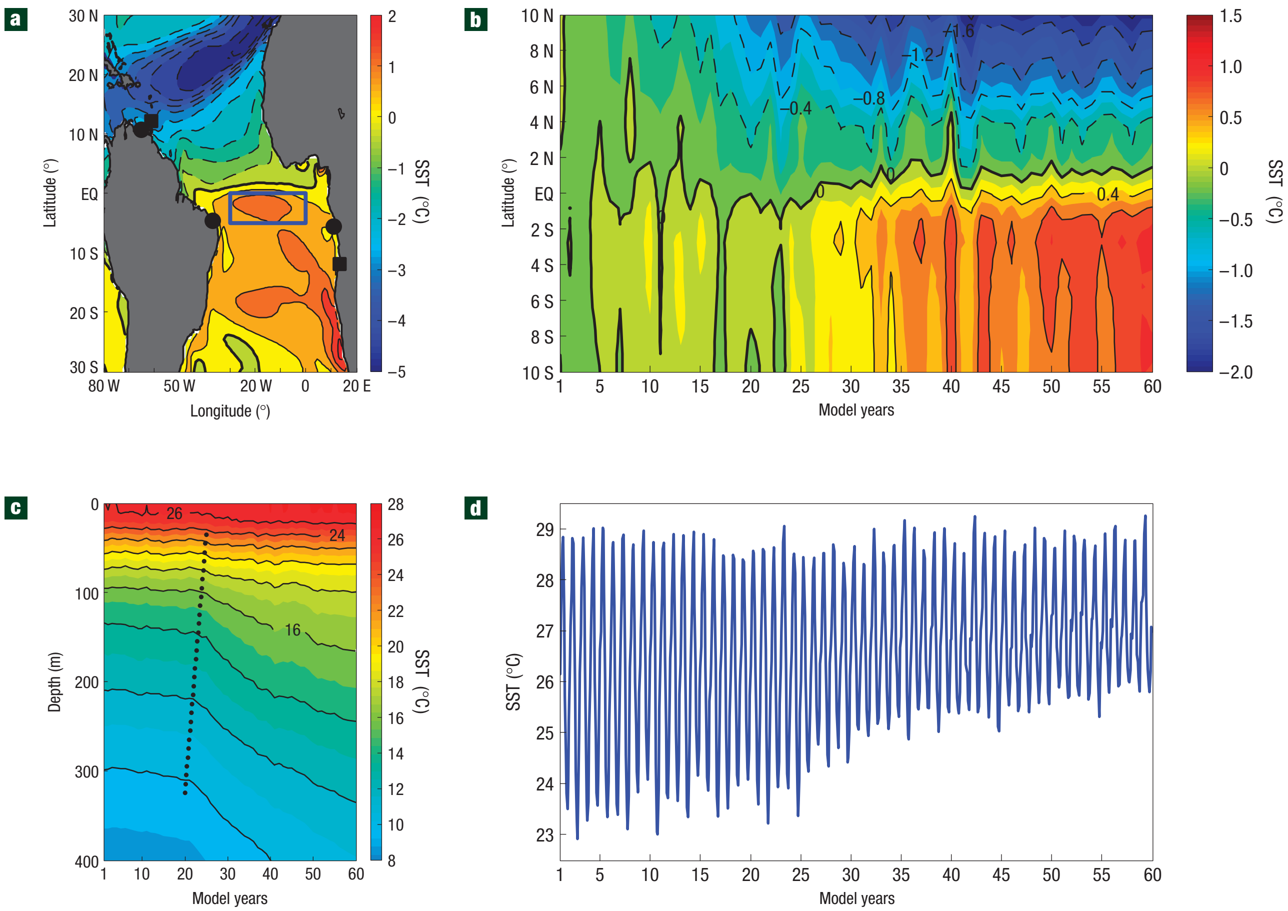


**Figure 1 Simulated ocean temperature change caused by freshwater input in the high-latitude North Atlantic. a**, SST difference between the water-hosing and control simulations averaged over the last 20 years of a 60 year model run showing cooling in excess of 4 °C (warming in excess of 1 °C) in the north (south) tropical Atlantic Ocean. **b**, SST difference between the water-hosing and control simulations zonally averaged across the Atlantic basin as a function of time and latitude. **c**, Upper ocean temperature averaged over the area between 30° W and 0° E in longitude and 5° S and 0° N in latitude indicated by the rectangle in **a** as a function of depth and time. The dotted line indicates the abrupt changes in temperature. **d**, Time evolution of the SST averaged over the area indicated by the rectangle in **a** as a function of time. Rapid equatorial warming is seen about 20–25 years into the simulation in **b**–**d**. Ensemble averaging has been applied to all of the fields. The black circles in **a** indicate the locations of the proxy records that show surface cooling[8] (warming[7,10,11]) in the north (south) tropical Atlantic during the Younger Dryas event and the squares indicate the locations of rapid subsurface warming[9].

The cooling (warming) in the North (South) Atlantic revealed by the palaeo proxy records during the Younger Dryas event has been simulated by a wide range of climate models as a robust response of the SST to changes in freshwater input in the high-latitude North Atlantic[14], lending support to the hypothesis that the hydrological cycle changes in the high latitudes can trigger basin-to-global scale abrupt climate shifts through a series of oceanic and atmospheric processes and feedbacks[15,16]. Even though the ATHC is widely perceived as a leading physical cause for the SST response, recent modelling studies show that atmospheric processes interacting with the ocean mixed layer can also play a significant role in transmitting the high-latitude SST changes to the tropics[17], raising the debate on the teleconnection mechanisms between the high latitude and tropics during the abrupt climate events. At issue is whether the ocean circulation change plays an active role in tropical Atlantic abrupt climate events, such as the deactivation of the African monsoon during the Younger Dryas event, or is mainly a passive response to atmospheric changes.

We attempt to address these issues by analysing an ensemble of state-of-the-art coupled climate model simulations where fresh water is hosed into the high-latitude North Atlantic to suppress the ATHC, mimicking meltwater discharge during abrupt climate change events, such as the Younger Dryas event (hereafter referred to as the water-hosing experiment; see the Methods section for model and experiment descriptions). Throughout the study, ensemble averages are used to suppress the internal variability of the coupled model, enabling a better depiction of the transient response to the freshwater forcing.

Figure 1a shows the model SST change, in reference to the control simulation, averaged over the last 20 years of the water-hosing simulation in the tropical Atlantic basin. Consistent with the palaeo proxy records and other model water-hosing experiments[14], a dipole-like SST response emerges with strong cooling in excess of 4 °C over the north tropical Atlantic, and warming in excess of 1 °C over much of the equatorial and south tropical Atlantic. Figure 1b shows a rapid development of the cooling signal in the North Atlantic immediately following the

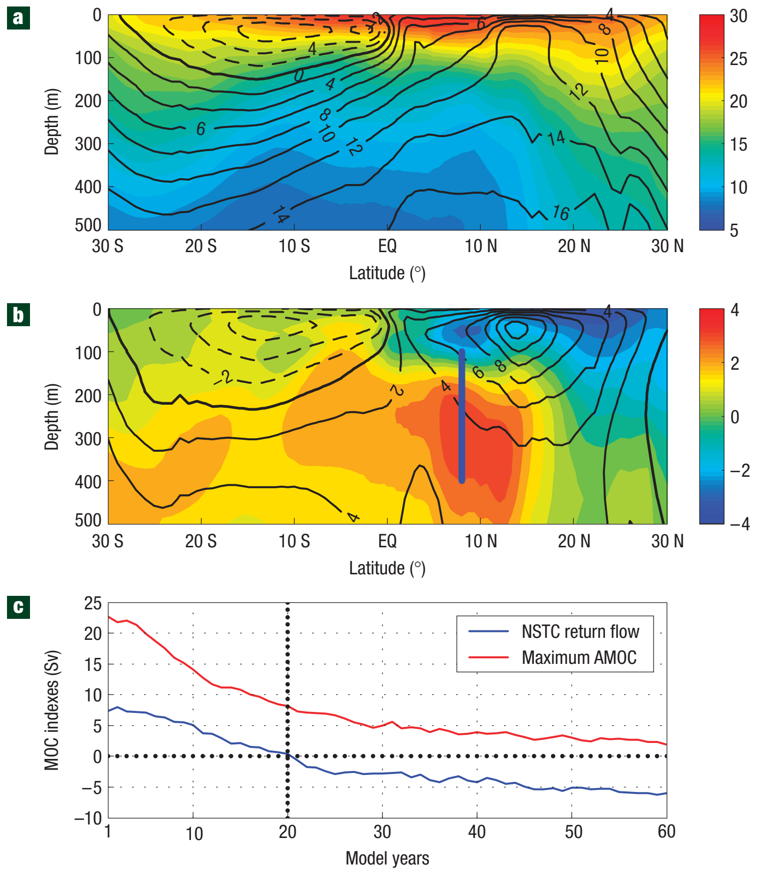


**Figure 2 Relationship between AMOC and temperature change. a**, AMOC streamfunction (contour) that shows the ocean circulation in the meridional plane and zonally averaged temperature (colour) in the upper tropical Atlantic ocean from 30° S to 30° N and 0–500 m averaged over 100 year of the control simulation. **b**, AMOC streamfunction (contour) and zonally averaged temperature change (colour) in reference to the control simulation averaged over the last 40 years of the water-hosing simulation. In **a**,**b**, the contour interval is 2 Sv and temperature is in °C. **c**, The maximum AMOC streamfunction as a function of time and the mass transport in the low branch of the NSTC averaged across the basin along 8° N from 100 to 400 m indicated by the blue vertical line in **b**. The latter is dominated by the North Brazil Current along the western boundary and it reverses direction at year 20 as indicated by the vertical dotted line in **c**. The transport is measured in Sv, where 1 Sv = $10^6$ m$^3$ s$^{-1}$.

onset of the freshwater forcing. The cooling reaches the deep tropics in less than a decade. This fast response time is probably attributed to atmospheric processes[18] and interactions between surface heat fluxes and the ocean mixed layer[19].

The warming south of the equator behaves in a more perplexing way, and it seems to develop in two stages. In the first stage, a weak warming of the order of 0.2–0.3 °C forms within the first two decades of freshwater forcing. This warming is caused by the planetary wave adjustment of the ocean in response to the freshwater forcing. As the deep-ocean convection reduces over the high-latitude North Atlantic owing to the enhanced stratification, cold water is trapped near the surface, causing warming beneath the upper ocean. The local pressure change associated with the warming quickly develops into Kelvin waves that propagate rapidly towards the equator along the western boundary and then along the equatorial waveguide[20]. The subsurface warming manifests itself as a surface warming signal near the equator, because of the close association between the SST and thermocline in the equatorial upwelling zone[21].

More drastic warming takes place 25–35 years after the onset of the freshwater forcing (Fig. 1b). During this second stage, the boreal summer equatorial SST increases more abruptly and by more than 2 °C. The surface warming is clearly preceded by an abrupt weakening in the upper ocean stratification (Fig. 1c). It is this later warming that is so critically important for the change in the African monsoon, as the weakened upper ocean stratification has a direct impact on the Atlantic cold tongue SST during the boreal summer (Fig. 1d).

The abrupt warming results from a nonlinear response of the tropical Atlantic coupled system to interactions between the ATHC and the wind-driven subtropical cells (STCs): the two components of the Atlantic meridional overturning circulation (AMOC). An analysis of the AMOC streamfunction indicates that the STCs in the upper tropical ocean incur the most striking AMOC changes. In the control simulation, the streamfunction indicates a 12 sverdrup (Sv) interhemispheric flow across the equator in the upper 300 m, which is within the observed estimates[22]. A major portion of the interhemispheric flow is from the upper return branch of the ATHC that interacts with the STCs[23,24]. The interaction is particularly strong with the northern subtropical cell (NSTC), where the equatorward (southward) return flow below the surface is essentially blocked by the northward ATHC flow[24,25], leaving it invisible in the upper AMOC streamfunction (Fig. 2a). In sharp contrast, the NSTC in the hosing simulation emerges as a closed cell (Fig. 2b), carrying warm subsurface north subtropical gyre waters to the equatorial zone. This

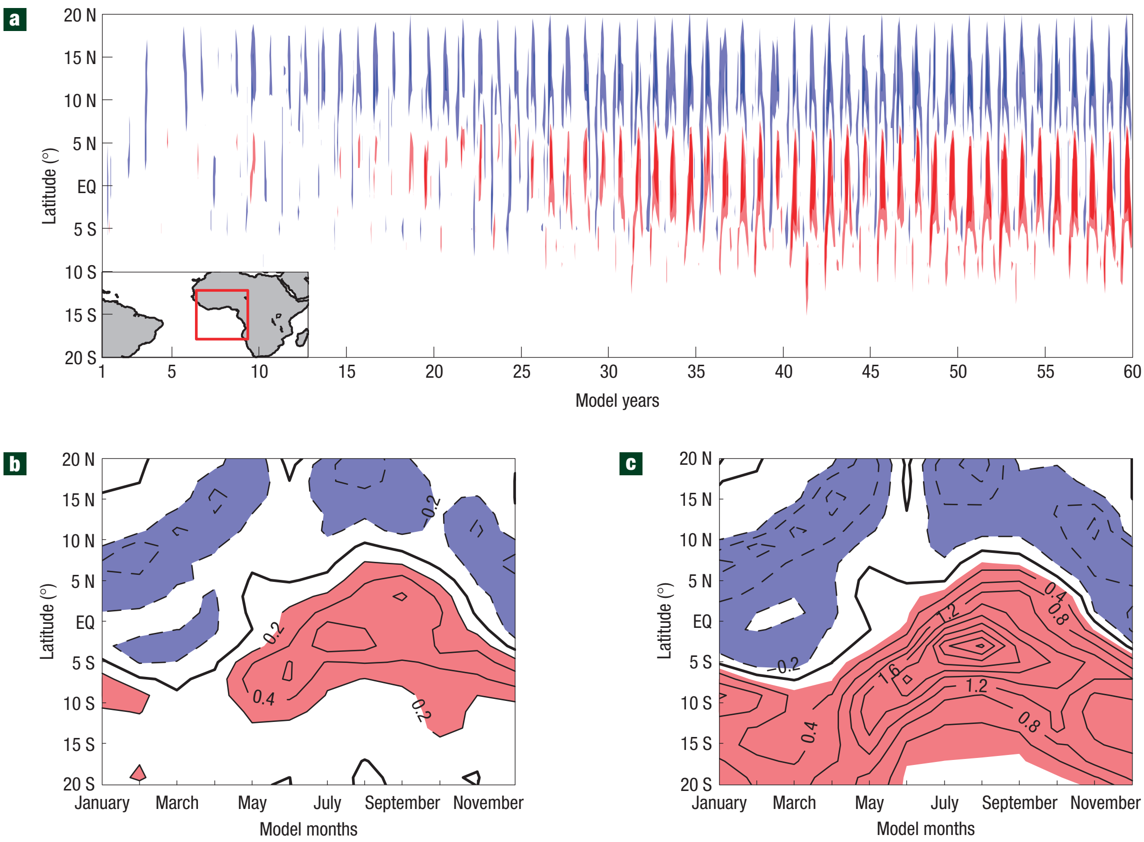


**Figure 3 Changes in the monsoonal rains.** Zonally averaged precipitation over the Sahel and Gulf of Guinea (the square in the inset in **a**) between the water-hosing and control runs. **a**, Monthly mean precipitation change during the 60 year model simulation, showing a drastic increase of the summer monsoon rainfall south of 5° N about 20–25 years into the simulation following the onset of the equatorial warming (Fig. 1d). The red (blue) shading denotes an increase (a decrease) of precipitation in a range between 0.5 m yr$^{-1}$ and 3.5 m yr$^{-1}$. **b**, Fractional change of the precipitation annual cycle during the first 20 years of the simulation. **c**, The same fractional change during the last 20 years. The fractional change is defined as the precipitation difference between the two runs divided by the precipitation of the control run.

occurs only when the ATHC is weakened beyond a threshold (8 Sv in this case, Fig. 2c). The North Brazil Current off the northeastern coast of South America, where the oceanic warming originates, holds a key for the ATHC–STC interaction (see the Supplementary Information).

An abrupt change of monsoonal rains follows the rapid oceanic warming (Fig. 3a). Before the warming, the change in the rainfall seasonal cycle is of an order of 20%–40% (Fig. 3b). This change is attributed to the cooling of the north tropical Atlantic that develops fully within the first two decades (Fig. 1b). During this period, much of the rainfall change is associated with the southward shift of the mean Intertropical Convergence Zone position and the rainfall change within the summer monsoon months is less pronounced. After the warming, the drastic increase in the boreal summer SST in the Gulf of Guinea (Fig. 1d) amplifies the seasonal rainfall changes to above 60%. In particular, there is a threefold increase in the boreal summer precipitation over the Gulf of Guinea (Fig. 3c), indicative of a major change in the monsoon circulation. Therefore, the rapid warming in the equatorial Atlantic is crucial in the pronounced regime shift of the summer monsoon rainfall. Furthermore, the interannual SST variability associated with Atlantic Niño is reduced more than 50% after the onset of the oceanic warming[26], leading to a permanent Atlantic Niño-like condition in the equatorial region (see the Supplementary Information).

The finding reported here is corroborated by similar water-hosing experiments using other climate models, indicating the robustness of the proposed oceanic mechanism. The ATHC threshold behaviour may contribute to the bimodal behaviour of the African monsoon recorded during the latest Pleistocene and early Holocene. It is probable that during the Younger Dryas event the ATHC may be weakened so much that it exceeded the threshold, triggering an abrupt weakening in the upper ocean stratification of the equatorial South Atlantic. This then causes a major reduction in the boreal summer cooling of the Atlantic cold tongue and drastically amplifies the African monsoon response to the high-latitude climate change. Future studies with enhanced model resolutions are required to reveal the full temporal and spatial features of the African monsoon response, as these may be sensitive to model physics and resolutions. If this hypothesis is proved correct, then the imposed threshold in the change of the ATHC provides a useful measure of how significantly changes in Atlantic Ocean circulation may affect the African monsoon in future climate change.

## METHODS

### CLIMATE MODELS

The primary model used here is the fully coupled ocean–atmosphere global general circulation model (CM2.1) developed at the Geophysical Fluid

Dynamics Laboratory (GFDL). The ocean model includes an explicit free surface and a true freshwater flux exchange between the ocean and atmosphere. The ocean model has 50 vertical levels (22 levels within the top 220 m) and 1° horizontal resolution with the meridional resolution reducing to 1/3° within the tropics. The atmosphere model has 24 vertical levels and horizontal resolution of 2° latitude × 2.5° longitude.

### CONTROL EXPERIMENT

The model was first spun up for 220 years using constant AD 1860 radiative forcing conditions, including the solar irradiance, and then integrated for 2,000 years. The control integration produces a stable realistic simulation of pre-industrial climate without flux adjustments[27]. Five 60 year segments were taken from the 2,000 year run at a 50 year interval from the beginning of the run and then subject to an ensemble average to form a 60 year record to facilitate a direct comparison with the water-hosing runs.

### WATER-HOSING EXPERIMENT

An idealized strong extra freshwater forcing of 0.6 Sv ($1\ \mathrm{Sv} = 10^6\ \mathrm{m^3\ s^{-1}}$) was uniformly added to the GFDL CM2.1 over the high-latitude North Atlantic (55°–75° N, 63° W–4° E) and run for 60 years. An ensemble of five such integrations was made, each of which starts with an initial condition that corresponds to the beginning state of the five 60 year segments from the control simulation[28].

Received 6 February 2008; accepted 1 May 2008; published 30 May 2008.

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Supplementary Information accompanies this paper on www.nature.com/naturegeoscience.

## Acknowledgements

We wish to thank J. Yin for providing us with the extra water-hosing experiment output of the GFDL CM2.1 model. This work is supported by the National Science Foundation, Grant No. OCE-0623364 and ATM-0337846, the Climate Program Office, US National Oceanic and Atmospheric Administration, Grant No. NA050AR4311136 and the Office of Science (BER), US Department of Energy, Grant No. DE-FG02-07ER64443.

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