submitted to Paleoceanography and Paleoclimatology March 30, 2020

Precessional Climate Variations Driven from the Tropics

J. R. Toggweiler*
Geophysical Fluid Dynamics Laboratory, NOAA, Princeton, NJ USA

* Geophysical Fluid Dynamics Laboratory, National Oceanic and Atmospheric Administration, Princeton, NJ, 08540-6649, USA (robbie.toggweiler@noaa.gov)

Key Points:

- The summer heating of northern Africa and southern Asia alters the trade winds over the tropical Atlantic.
- The changing trades combine with the AMOC to warm and cool the tropics.
- The trades and the AMOC provide a simple way for variations in the summer heating to alter the climate system.

Abstract

The Atlantic's Meridional Overturning Circulation (AMOC) is weak when the northern summer insolation is strong and vice-versa in the 23-kyr precessional frequency band. The mechanism behind this response is unknown and does not conform in any obvious way to Milankovitch theory. The link between the AMOC and precession is attributed here to the trade winds that blow out from Africa across the Atlantic. The trade winds, like the AMOC itself, are weak when the northern summer insolation is strong. The trade winds are shown here to have a direct impact on the AMOC's return flow. A return flow altered this way should warm and cool the Earth in a way that has not been previously considered.

Plain Language Summary

Variations in Earth's orbit and spin axis alter the amount of solar energy (insolation) that reaches the Earth over time and are thought to have played a major role in the ice ages of the last 2.5 million years. The orbitally-induced insolation changes are small, however, and it is still not entirely clear how they lead to warming and cooling on a global scale. This is especially true for the insolation changes brought about by the precession of the Earth's axis as it orbits the Sun. This paper shows how these changes alter the trade winds over the Atlantic and how the trade winds, in turn, alter the uptake and transport of solar energy by the Atlantic Ocean. This idea provides a new way to explain how subtle changes in Earth's orbit and spin axis come to have a large global impact.

1. Introduction

The Milankovitch hypothesis holds that the ice ages are caused by variations in the Earth's orbit and spin axis that alter the insolation at latitudes near 65°N during the summer months. It identifies three astronomical cycles with three distinct periods of variability as the main forcing mechanisms. The hypothesis was seemingly confirmed by Hayes et al. (1976), who found the predicted periods of variability in geologic records that go back 600,000 years.

The insolation variations are small, however, and it is still not entirely clear how they lead to cooling on a global scale. The original form of the Milankovitch hypothesis held that the northern ice sheets themselves are a conduit or amplifier through which small variations in the summer insolation come to have a larger impact. But the ice ages are now known to include other factors, like atmospheric CO_2, and the cooling is now known to be almost as great in the south as it is in the north – an outcome that the original Milankovitch hypothesis could not explain.

Perhaps the greatest mystery is why temperatures in the Southern Hemisphere seem to rise and fall along with the summer insolation in the Northern Hemisphere (Hayes et al., 1976; Imbrie et al., 1992; Kawamura et al., 2007). The mystery is most acute in the 23-kyr precessional band, where southern temperatures vary in-phase with the northern insolation and out-of-phase with the local insolation south of the equator. This pattern implies that the climate system is *insensitive* to the insolation south of the equator in much the same way that it seems to be especially sensitive to the insolation north of the equator.

As it turns out, there is an element of the climate system that is sensitive and insensitive in just this way − the African and Asian monsoons (Prell and Kutzbach, 1987; Pokras and Mix, 1987; Ruddiman, 2001). Monsoon winds are driven by the heating of the northern continents during the summer and the lack of heating during the winter. Atmospheric winds, moreover, are especially *insensitive* to the insolation south of the equator − simply because there is so little land in the Southern Hemisphere. The monsoons, however, are often seen as a sideshow in the ice ages because they are not thought to have a direct impact on the northern ice sheets (Imbrie et al., 1992).

Here, it is argued that the summer heating of the northern continents is the most important way for the precessional forcing to enter the climate system. The basic idea is that the summer heating alters the trade winds over the tropical Atlantic, and the altered trade winds, in turn, alter the Atlantic's Meridional Overturning Circulation, or AMOC. The AMOC then alters sea surface temperatures (SSTs) across the tropics in a way that makes the temperatures in both hemispheres rise and fall together. The origin of this idea comes from a new finding about the modern AMOC in Toggweiler, Druffel, Key, and Galbraith (2019a and 2019b, hereon TDKGa and TDKGb). The new finding is reviewed in Section 5.

Astute readers will recall that Imbrie et al. (1992) also invoked the AMOC to explain how the northern summer insolation influences the Southern Hemisphere. Imbrie et al., however, maintained a Milankovitch-like focus on the insolation at 65°N. Sections 2 and 3 below review the Imbrie et al. approach and its shortcomings. (Full disclosure: the author of this paper is one of the Imbrie et al. authors.)

2. SSTs in the North Atlantic

The original form of the Milankovitch hypothesis posited that the ice ages are driven by the insolation that falls on the northern ice sheets during the summer months. Imbrie et al. (1992) recognized, however, that a number of climate indicators seemed to respond to the northern summer insolation before the ice sheets were able to respond. Most of these indicators were from the Southern Hemisphere. Imbrie et al.'s goal was to discover how the summer insolation at high northern latitudes reaches the early responders in the south.

Imbrie et al. put together an array of 20 climate records that was designed to show how orbital information flows through the climate system. From the 20 records, Imbrie et al. developed a process model to describe how the orbital signals move from north to south. The centerpiece of the process model was the AMOC. More insolation at 65°N, they claimed, warms the northern North Atlantic and pushes back the sea ice. The reduced sea ice leads to higher salinities that increase the formation of North Atlantic Deep Water (NADW) and the increased formation of NADW strengthens the AMOC. A stronger AMOC then carries the northern warming across the equator to the high latitudes of the Southern Hemisphere. The process model made some specific predictions: SSTs in the North Atlantic should be warm, and the AMOC should be strong, at times when the summer insolation at 65°N is at a maximum.

Two of the 20 records in Imbrie et al. (1992) are SST records from the North Atlantic. Neither supported the process model. Record 6 is an SST record from core V30-97 from 41°N in the middle of the North Atlantic. Ruddiman and McIntyre (1981) had pointed out years before that the SSTs at V30-97 seemed to be anti-phased with the local summer insolation in the precessional band. So, at times when the process model expected warmer SSTs, the SSTs at 41°N were relatively cool. Record 5 is an SST record from 50°N in the subpolar North Atlantic. The SSTs at this location are strongly influenced by the northern ice sheets (Ruddiman and McIntyre, 1984). As such, the maximum and minimum SSTs at this site tend to be delayed by thousands of years in relation to the summer insolation.

Figure 1, from Chapman and Shackleton (1998), shows a number of AMOC and SST proxies from core SU90-03, which was raised at 40°N about 140 km away from V30-97, the core on which Imbrie et al.' record 6 is based. The SST proxies from core SU90-03 in the **bottom panel** show that the surface waters in this area appear to have been relatively cool 61,000, 83,000, 105,000, and 128,000 years ago. The cool intervals are 22-23 kyrs apart and coincide with times when the northern summer insolation was at a maximum.

Chapman and Shackleton noted the precessional spacing between the cool intervals in **Figure 1** but chose to emphasize that the cool intervals coincided with Heinrich Stadials 6, 7, 9, and 11, as indicated by the ice rafted debris (IRD) and higher abundances of N. pachyderma (sinistral) in the **top panel**. Of course, the IRD and higher abundances of N. pachyderma would normally be associated with a *weaker* AMOC rather than a stronger AMOC.

Imbrie et al. recognized that records 5 and 6 did not fit the process model. The mismatch led them to predict that SST records would be found further north − closer to the areas of deep-water formation − that would have SST variations that are in-phase with the summer insolation. But the expected records have never materialized.

3. Strength of the AMOC

A widely used indicator of AMOC strength is the $\delta^{13}C$ of the shells of bottom-dwelling (benthic) foraminifera from the deep Atlantic. A positive $\delta^{13}C$ shift is seen as an indication that a stronger AMOC filled more of the deep Atlantic with high-$\delta^{13}C$ NADW (Raymo et al, 1990; Curry, 1996). The most definitive work on this topic has been done by Lisiecki et al. (2008), who developed a four-core stack of mid-depth Atlantic records (3000-4010 m) to show how the strength of the AMOC has varied over the last 425,000 years in the obliquity and precessional bands.

The AMOC variations in the obliquity band are consistent with an origin in high northern latitudes. The summer insolation at maximum tilt leads to warmer temperatures, reduced sea ice, and increased evaporation in the areas where NADW is formed. Higher salinities then lead to a stronger AMOC. Lisiecki et al. also show that the AMOC is strongest several thousand years after maximum tilt. Basically, the northern ice sheets shrink in response to maximum tilt and their smaller size seems to augment the warming and the higher salinities. The augmentation is delayed, however, and the delay helps explain the lag in the AMOC's response.

But, as one might expect from **Figure 1**, the situation in the precessional band is different. Lisiecki et al. show that the AMOC appears to be weak, i.e. mid depth $\delta^{13}C$ values are relatively light, when the insolation during northern summers is most intense. Moreover, there is no lag in the precessional band; *minimum* AMOC seems to be in-phase with maximum summer insolation.

Imbrie et al. (1992) imagined that the changes in the northern North Atlantic would produce similar changes in the AMOC in both the obliquity and precessional bands but this expectation turns out not to be true. Lisiecki et al. had no explanation for the precessional response except to say that the AMOC "must be sensitive to factors other than Milankovitch forcing and ice volume."

4. Trade Winds and the AMOC

Imbrie et al.'s expectation about the AMOC was not unreasonable. So why has it turned out to be so wrong? As will be argued below, a key part of the AMOC is altered by the trade winds that blow across the tropical Atlantic, and the trade winds are strongly altered by the precessional heating of northern Africa and southern Asia. More heating weakens the trade winds and weakens this part of the AMOC. This effect then overwhelms the changes in the northern North Atlantic and reverses the sign of the precessional response.

The trade winds over the Atlantic are separated into northern and southern bands by the Intertropical Convergence Zone (ITCZ). The northern trades blow from the coast of Africa toward the Caribbean Sea to the north of the ITCZ, while the southern trades, or southeast trades, blow up from southern Africa south of the ITCZ toward the area where the equator meets the east coast of South America. Both the northern and southern trades are impacted in a negative way by the monsoon winds generated by the summer heating of northern Africa and southern Asia.

Battisti et al. (2014) carried out two model simulations for the periods 207,000 and 218,000 years ago, when Earth's orbital eccentricity and the impact of the precessional forcing were especially strong. The Earth was relatively far from the Sun during northern summers 207,000 years ago and was relatively close to the Sun during northern summers 218,000 years ago. **Figure 2** shows the change in 850 mb summer winds between the two model solutions (218 kyr minus 207 kyrs). The figure shows that the low-latitude summer winds over all three ocean basins are altered in a major way when the summer heating of the northern continents is especially strong or weak. One sees, in particular, that the wind anomalies over the tropical Atlantic are directed back toward Africa in a way that opposes the modern trades. Figure 10 in Bosmans et al. (2015) shows something very similar.

Most importantly, the wind anomalies in **Figure 2** are as large as the mean winds in these areas today (~5 ms^{-1})! This means that the summer heating of northern Africa and southern Asia 218,000 years ago was potentially strong enough to have eliminated the summer trade winds over much of the tropical Atlantic.

As will be shown below, the trade winds are important because they draw the return flow of the AMOC up to the surface. The return flow begins with the upwelling of deep water south of the Antarctic Circumpolar Current (ACC) and the northward transport of the upwelled water in the surface Ekman layer (Toggweiler and Samuels, 1995; Gnanadesikan, 1999). The upwelled water is then incorporated into two water masses that are formed north of the ACC − Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) (Iudicone et al., 2008; Sloyan et al., 2010). SAMW and AAIW are the water masses from which NADW is eventually made in the North Atlantic.

Figure 3 gives two contrasting descriptions of what happens next. The diagram in the **top panel** (adapted from Gnanadesikan, 1999) shows the upwelled deep water being added to a "bowl" of low-density water north of the ACC. The low-density water added in the south is then removed when NADW is formed in the North Atlantic. The water added to the bowl has a relatively low density because the dense water upwelled south of the ACC is warmed and freshened as it is pushed to the north in the surface Ekman layer. As depicted in the **top panel** (and in Gnanadesikan's original diagram), nothing much happens to the water in the bowl between its entry point in the south and the area where NADW is formed.

In the **bottom panel of Figure 3**, the bowl is divided into warm and cool layers separated by the 10°C isotherm. SAMW and AAIW are added to the cool lower layer, as they are in the real ocean. But, in this case, the added water is shifted up into the warm surface layer as it passes through the tropical Atlantic. NADW is then formed mainly from the water in the warm upper layer. The upshift in the **bottom panel** has been well documented in the oceanographic literature but has received scant attention in the ocean-modeling and climate-modeling communities. According to Wunsch (1984), Schmitz and Richardson (1991), and Lumpkin and Speer (2003), about 60% of the subantarctic water that enters the South Atlantic today as part of the return flow is shifted up to the surface within the tropical Atlantic.

The claim here is that the upshift varies over time in response to the trade winds over the tropical Atlantic. When the northern summer insolation is strong and the trades are weak, as they were 218,000 years ago, the upshift is minimal and the return flow passes through the tropics without being drawn up to the surface. Conversely, when the northern summer insolation is weak and the trades are strong, as they were 207,000 years ago, the upshift is strong and most of the return flow is drawn up to the surface. The situation today falls somewhere in between because the northern summer insolation today is relatively weak but not nearly as weak as it was 207,000 years ago.

5. Impact of the Upshift on Earth's Climate

The upshift in **Figure 3** should have a big impact on the heat engine that carries heat and moisture poleward in the atmosphere. Basically, the water entering the tropical Atlantic in **Figure 3** is cool while the surface waters leaving the tropical Atlantic are warm. So, when the cool inflow is transformed into a warm outflow, the ocean removes heat that would otherwise be available for warming the tropics. The heat is then carried away to a location (the North Atlantic) where it has less impact on the heat engine.

When the trade winds are weak, more of the heat from the Sun remains in the tropical Atlantic, and the extra heat leads to warmer SSTs. The warming is then quickly transmitted to the rest of the tropics so that the warmest areas of the tropics are warmed as well (Wallace, 1992; Sobel et al., 2002). The atmospheric convection in these areas extends more vigorously into the upper troposphere. An upper troposphere warmed and moistened in this way should then give rise to stronger midlatitude weather systems that carry the extra heat and moisture poleward in both hemispheres.

Conversely, when the trade winds and the upshift are strong, more of the heat from the Sun is removed and carried away to the North Atlantic. SSTs across the tropics are thereby cooler than they would otherwise be. This leads to a cooler and drier upper troposphere and weaker mid-latitude weather systems that deliver less heat and moisture to the polar regions.

Something similar happens during ENSO events today. SST variations in the cool eastern equatorial Pacific (EEP) are known to influence the tropical troposphere and remote areas of the ocean (e.g. Bjerknes, 1969, Klein et al, 1999). More specifically, the SSTs in the warmest parts of the tropical ocean are known to vary along with the SSTs in the relatively cool EEP (Wallace, 1992; Sobel et al., 2002). The eastern South Atlantic is relatively cool like the EEP. It follows that the warmest SSTs in the ocean would warm and cool over precessional cycles along with the eastern South Atlantic in a similar way.

The summer insolation that alters the trade winds varies over time in the same way as the summer insolation over the ice sheets at 65°N. This means that the growth and retreat of the northern ice sheets in the 23-kyr precessional band could just as easily be a response to the warming and cooling in the tropics as a response to the insolation overhead.

6. A Brief Description of How the Upshift Works

Inverse models of the ocean's large-scale overturning show that ~8-10 Sv of cool subsurface water is upwelled to the surface in the tropical areas of each ocean basin (e.g. Wunsch, 1984; Lumpkin and Speer, 2003). The cool water, moreover, is known to have a subantarctic origin, like the SAMW and AAIW that are added to the lower layer in the **bottom panel of Figure 3**. The inverse models are not very specific, however, about how and where the upwelling occurs.

TDKGa constructed a set of maps of the ocean's surface $\Delta^{14}C$ to investigate the upwelling. Their maps show that surface waters with subantarctic $\Delta^{14}C$ values are found in upwelling areas along the margins in all three ocean basins. The spreading of the subantarctic water away from the upwelling areas was then used to show that the upwelling in the inverse models must be taking place along the margins in the upwelling areas.

In the companion paper, TDKGb tried to simulate the surface $\Delta^{14}C$ in two versions of a modern climate model and found that the models could only produce the 2-3 Sv of upwelling driven by the winds along the coast. They could not produce the 8-10 Sv of upwelling seen in the inverse models and the $\Delta^{14}C$ maps. In seeing how the models fell short, TDKGb were able to come up with a description of how the larger volumes of upwelling occur. Their description starts with the winds that blow across the Atlantic and Pacific basins near the equator.

The easterly winds near the equator push the local surface waters across the basins and, in so doing, take water from the east and pile it up in the west. The easterlies also produce a divergence that draws up water from the Equatorial Undercurrent (EUC). Thermocline water flows into the equatorial zones from the north and south to feed the

EUC and the divergence (Liu et al., 1994; McCreary and Lu, 1994). But the thermocline water upwelled from the EUC is not very cool, ~19-24°C.

Some of the warm water that is piled up in the west, however, is drawn away to become NADW (Lu et al., 1998), and the withdrawal puts a different conceptual frame around the upwelling in the east. *Without* the withdrawal, all the water pushed across the basins by the easterlies returns to the east via the EUC. As pointed out above, none of this water is very cool, or low in $\Delta^{14}C$. *With* the withdrawal, on the other hand, the amount of water that returns via the EUC is reduced – because the withdrawal in the west lowers the pressure head that drives the EUC. As a result, more water is pushed across the basins by the winds than returns via the EUC.

According to TDKGb, the additional water that is pushed across the basins is drawn up from the cool subantarctic layers that outcrop along the eastern margins. More water is drawn up from these layers when more of the warm water piled up in the west is drawn away. The models examined by TDKGb could not produce the upwelling in the east because *none* of the warm water piled up in the west was drawn away.

As will be shown in more detail below, the key to the upwelling along the eastern margins is the way that the *push* from the easterly winds near the equator connects with the *draw* from the North Atlantic. The models examined by TDKGb could not make this connection.

7. Temporal Variations in the Equatorial Atlantic and Caribbean Sea

The $\Delta^{14}C$ maps in TDKGa provide a snapshot of the upwelling in the modern ocean but they have nothing to say about how it varies in time. The $\delta^{13}C$ evidence in Lisiecki et al. (2008), meanwhile, shows how the AMOC varies in time but only in areas that are remote from the tropics. So, how do we know that the upwelling in the tropical Atlantic varies along with the precessional forcing and the AMOC?

McIntyre et al. (1989) developed SST proxies for the warm and cool seasons in the equatorial Atlantic and were able to show that the "cool season" SSTs exhibit a strong response in the precessional band. The cool season is northern summer when the south-east trades are strongest. McIntyre et al. found that the cool season SSTs in the equatorial

Atlantic are relatively warm when the summer heating of northern Africa is enhanced. They attributed the changes to monsoon winds that weaken the trades, as is argued here. Schneider et al. (1995 and 1996) show that the SSTs off southern Africa vary in much the same way. Neither McIntyre et al. nor Schneider et al. linked the warming and cooling to the AMOC, however.

Straub et al. (2013), meanwhile, found distinct precessional cycles in the $\delta^{15}N$ of the water that flows into the Caribbean Sea via the AMOC (**Figure 4**). The $\delta^{15}N$ signal was derived from planktonic foraminifera that grew in the western Caribbean Sea near ODP site 999 (12.8°N). $\delta^{15}N$ is an upwelling indicator like the surface $\Delta^{14}C$; both preserve information about the upwelling that is carried well beyond the upwelling areas.

The $\delta^{15}N$ variations in Straub et al. (2013) are a response to nitrogen fixation that is stimulated by the phosphate in the upwelled subantarctic water. As shown in the **top panel of Figure 4**, the surface waters at site 999 are isotopically light (smaller positive numbers) due to more nitrogen fixation during times when the northern summer insolation is weak and the trade winds are strong. The $\delta^{15}N$ variations at site 999 thereby provide clear empirical evidence that the upshift was, in fact, stronger. There is more upwelling off Africa and more transport of equatorial surface waters into the Caribbean Sea when the northern summer insolation is weak and the trade winds are strong.

8. Discussion

The Earth tends to be warm when the summer insolation over the Northern Hemisphere is enhanced and the AMOC is often invoked to explain why this is so, e.g. Imbrie et al. (1992). The evidence compiled above suggests, however, that the AMOC is actually relatively weak during the warm intervals.

As argued here, the relative warmth and weak AMOC are related. But it is not the weak AMOC *per se* that causes the warming. It is rather the upshift in the AMOC's return flow. The upshift produces a pervasive cooling that extends to both hemispheres. The claim here is that the Earth warms when the upshift is weakened by the insolation. The first section of the Discussion below describes the physics of the upshift in more detail. The same physics is then used to show why the AMOC and the upshift are strong and weak at the same times. The remainder of the Discussion shows how the pervasive cooling accounts for important features in the paleo record.

8a. Physical Basis for the Upshift

As shown by TDKGb, the upshift is not easily reproduced in 3-D models. This was shown to be the case in the two models examined in TDKGb, and it would seem to be true for other models as well (e.g. Mechoso et al., 1998; Wang et al., 2014). The resolved physics in these models, therefore, cannot be used to show how the upshift works. A lower-order description is invoked here instead − an offshoot of the simple low-order model in Gnanadesikan (1999).

Gnanadesikan's model describes the mass balance in a bowl of low-density water that resides in the upper ocean north of the ACC. Low-density water is added to the bowl via the northward Ekman transport associated with the westerly winds over the Southern Ocean and the diapycnal mixing at the bottom of the bowl. Water is removed from the bowl when NADW is formed. The rate at which NADW is formed is proportional to the thickness of the bowl. So, when more water is added to the bowl via the forcing terms, more water is removed from the bowl via NADW.

The low-order model considered here is a mechanical energy balance that works in much the same way. The thickness of the bowl is related to the amount of available potential energy (APE) associated with the bowl of low-density water. As in Gnanadesikan's model, the southern westerlies and diapycnal mixing put energy into the APE pool. The formation of NADW also takes energy from the pool and thins the bowl. The extracted energy is then dissipated by the currents that make up the overturning. (Readers interested in a more complete description of APE and the ocean's mechanical energy budget are referred to Toggweiler and Samuels, 1998.)

The energy balance also includes the buoyancy of the water in the different layers of the bowl, i.e. there is more buoyancy/more APE associated with the warm layers near the top of the bowl and less with the cool layers near the bottom. It also includes the buoyancy/energy that is added to or taken out of the bowl via the heat exchange with the

Sun and atmosphere. A key question for the upshift is: how does the warm buoyant water near the top of the bowl flow out of the bowl to the North Atlantic?

Viewed from above, Gnanadesikan's bowl is a mound of elevated sea surface height (SSH). The warm buoyant water in the tropical Atlantic is seen, in this context, to leave the bowl via a cut or a trough in the mound. The cut begins in the northern North Atlantic where the SSH is low due to the formation of NADW. It then extends down the coast of North America and around the Gulf of Mexico/Caribbean Sea to the area off northern Brazil, where it reaches the equatorial zone at ~5°N. Water from the equatorial zone leaves the bowl as a geostrophic flow along the offshore edge of the cut.

It is easy to see, from this perspective, how the upshift works. The trade winds over the tropical Atlantic push a large volume of equatorial water across the basin and pile it up in an area where it is juxtaposed with the cut. With this set up (low SSH along the coast, high SSH offshore) the piled-up water is able to continue on toward the Caribbean Sea and beyond. Then, as the trade winds push more water across the basin and more piled up water is drawn away, more cool water is drawn up to the surface off Africa. So, more of the heat/buoyancy coming into the bowl is removed and less remains. The isotherms in the bowl are displaced toward the surface. The upshift is thus much like the formation of NADW: when it is active, it thins the bowl and takes advantage of mechanical energy that would otherwise go unused.

Without the juxtaposition, on the other hand, the heat/buoyancy coming into the equatorial zone remains stranded in the bowl and the bowl is thicker. This is what seems to be happening in climate models. According to TDKGb, the culprit is poorly resolved (partly ageostrophic) flows in the Yucatan Channel and/or Florida Straits that prevent information about the SSH in the northern North Atlantic from reaching the equatorial zone. In other words, the flows through the Caribbean Sea and around Florida in the models are poorly resolved in a way that eliminates the cut at ~5°N.

The Gnanadesikan (1999) model has one more component that has not been considered thus far − the baroclinic eddies in the ACC. The eddies allow the light water added to the bowl by the southern westerlies to leak back out to the south. The leak is greater when the bowl is thick and the ACC is strong and is smaller when the bowl is thin and the ACC is relatively weak.

This is important because the AMOC should be stronger with a smaller leak. As shown in Gnanadesikan (1999), more of the water that is added to the bowl by the southern westerlies is removed as NADW when less of the added water leaks out in the south. This tendency should be enhanced when the trade winds over the Atlantic are strong. (The logical progression here is: stronger trades = stronger upshift = thinner bowl = weaker ACC = less vigorous eddies = smaller leak = more NADW for a given wind stress in the south.) This may help explain why the AMOC itself is strong when the northern summer insola-tion is weak per Lisiecki et al. (2008).

8b. Impact of the Upshift on Southern Temperatures

When the AMOC is suppressed in climate models, the North Atlantic cools and a complimentary warming develops across much of the Southern Hemisphere. The equator, meanwhile, tends to be an area of zero temperature change (e.g. Crowley, 1992; Zhang and Delworth, 2005). This pattern reflects the fact that the AMOCs in climate models, with no upshift, are only able to move heat across the equator from one hemi-sphere to the other.

When the AMOC in the real ocean is suppressed the tropics should warm (because the upshift would be suppressed as well), and the tropical warming should spread to both hemispheres. The Northern Hemisphere should then warm somewhat less (because less heat is being carried into the North Atlantic), and the Southern Hemisphere should warm more (because less heat is being given up to the other hemisphere). The opposite should occur when the AMOC speeds up. The tropics and both hemispheres should cool, with the north cooling less and the south more. The net result is that changes in the AMOC should stand out especially well in Southern Hemisphere temperature proxies.

The **top panel of Figure 5** (adapted from Kawamura et al., 2007) shows two proxies for the air temperatures over and around Antarctica over the last 350,000 years (orange and black curves). (The black curve better represents the temperature variations away from Antarctica.) The summer insolation at 65°N is given by the red curve. Blue vertical lines have been added to mark the insolation minima, which are labeled according to the marine isotope stages that they represent. The stage labels are aligned with the insolation, as in Kawamura et al. (2007), e.g. '5b' and '5d' refer to the insolation minima centered on 94 and 116 kyrs. (As noted previously, the summer insolation at 25°N has the same temporal pattern as the insolation at 65°N in the red curve.)

As is clear from the black curve, the area around Antarctica is cool when the northern summer insolation is weak (blue lines). These are the times when the cooling effect of the upshift was most pronounced. The temperatures around Antarctica are also seen to rise and fall more dramatically during the first halves of the three 100-kyr eccentricity cycles in the figure. These are times when the orbit was more eccentric and the precessional forcing was stronger.

The **second panel** shows the variations in atmospheric CO_2 over this time period. As emphasized by Kawamura et al., atmospheric CO_2 exhibits less variability in the precessional band and lags more clearly behind the forcing. The two **lower panels** show the changes in sea level and global ice volume. As emphasized by Imbrie et al. (1992), sea level and ice volume lag the northern summer insolation by about a quarter cycle.

8c. Isotope Stage 5d

The most telling interval, in this regard, is isotope stage 5d (~116,000 years ago). The air temperatures around Antarctica were nearly as cold during this interval as they were during stage 6, the previous glacial maximum. The cooling, moreover, seems to have been almost entirely a response to the precessional forcing. As shown in the **second panel of Figure 5**, atmospheric CO_2 remained fairly high and is therefore not available to explain the cooling. Global ice volume also remained fairly low.

Along these lines, Koutavas (2018) produced a stack of five SST records from the eastern equatorial Pacific (EEP) that extends through stage 6. The Koutavas stack shows that the cooling around Antarctica during stage 5d coincided with a ~3°C cooling in the EEP. According to the **fourth panel of Figure 4** the SSTs at site 999 in the Caribbean

Sea (from Schmidt et al., 2004) cooled just as dramatically. These records are important because they show that the 5d cooling in Antarctica occurred in the tropics as well.

Koutavas links the cooling in the EEP to a mechanism that is much like the one put forward by Imbrie et al. (1992). He calls for cooling in the northern North Atlantic that is transmitted to Antarctica via the AMOC. Koutavas then goes on to argue that the cooling around Antarctica is transmitted to the EEP through an "ocean tunneling" mechanism in the South Pacific. This is a rather convoluted way to explain the cooling in the EEP. The claim here is that the 5d coolings in both areas were simple responses to a stronger upshift in the tropical Atlantic.

Four of the five cores in Koutavas's SST stack are from the area north of the cold tongue in the equatorial Pacific. The SSTs in these records have minima during stages 4, 5, and 7 that are aligned with the temperature minima in Antarctica. The precessional variability in the record from within the cold tongue, however, is not as distinct. This finding is relevant to the topic taken up in Section 8e.

8d. Holocene Conundrum

Stage 5d is also of interest because the Earth's orbital position at that time was much like it is right now. (The Earth is farther from the Sun now in the same way that it was farther from the Sun during northern summers 116,000 years ago.) The orbit is less eccentric today, however, so the insolation is not as low today as it was during stage 5d. The less-reduced insolation still seems to be having a modest impact.

A reconstruction of the Earth's global mean temperature over the last 11,000 years shows that the Earth was relatively warm in the early Holocene (10,000 to 6,000 years ago). It then cooled by about a half degree over the last 6,000 years (Marcott et al., 2013). The relatively warm temperatures during the early Holocene are noteworthy because the northern ice sheets had not yet melted back. The amount of CO_2 in the atmosphere also paused at 260 ppm between 10,000 and 8,000 years ago and then increased by another 20 ppm over the rest of the Holocene (Indermühle et al., 1999). The lower level of atmospheric CO_2 10,000 years ago, like the extant ice sheets, should have made the early Holocene relatively cool.

Indeed, climate models, when given these forcing factors, predict that the Earth should have been relatively cool during the early Holocene and should have warmed by a half degree over the last 10,000 years. This is the opposite, however, of the tendency seen in the observations (Liu et al., 2014). Liu et al. call this discrepancy the "Holocene temperature conundrum." Some process (that is not active in climate models) seems to have overwhelmed the impacts of the retreating ice and rising CO_2 and cooled the Earth over the course of the Holocene.

The northern summer insolation was at a maximum 11,000 years ago. The heating of northern Africa and southern Asia was therefore relatively strong and the trade winds over the Atlantic were relatively weak. This would have made the tropics and mean climate warmer, despite the still extant ice sheets and lower CO_2. The summer insolation, meanwhile, is at a minimum now and the trade winds over the Atlantic are stronger. This should have made the tropics and mean climate cooler during preindustrial time in relation to 11,000 years ago.

8e. A Brief Note on the Pacific

The AMOC has a Pacific component and, as shown in TDKGa, the Pacific has an upshift that is much like the upshift in the Atlantic. The upwelling, in this case, takes place mainly off Peru. The upshift in the Pacific, however, does not vary in time in the same way. This is because the summer heating of the northern continents has the opposite effect on the trade winds over the Pacific.

As seen in **Figure 2**, the easterlies over the western Pacific, in particular, are stronger when the northern summer insolation is enhanced. This is the opposite of the situation in the Atlantic, where the easterlies are weaker. This means that the tendency of the trade winds to pile up equatorial water in the western Pacific is out-of-phase with the formation of NADW and the strength of the AMOC. Indeed, **panel f of Figure 4** shows that $\delta^{15}N$ variations in the western equatorial Pacific tend to be 180° out-of-phase with the Atlantic variations in **panel a**. Thus, one would not expect to see the same warming and cooling off Peru that one sees in the tropical Atlantic.

So, what happens when the draw from the North Atlantic is relatively strong and the trade winds over the Pacific are relatively weak? As will be shown in a forthcoming paper, the draw from the North Atlantic seems to enhance the upwelling in the subarctic North Pacific instead. As a result, the near-surface flow through the Indonesian Seas should have more of a low-salinity northern source when the northern summer insolation is weak and more of a high-salinity equatorial source when the insolation is strong, as seen, for example, during the Holocene in Stott et al. (2004).

9. Conclusions

The insolation falling on the Northern Hemisphere during the summer months has been shown time and time again to be especially important for Earth's climate variability. According to Milankovitch theory, the summer insolation has its greatest impact at 65°N, where the northern ice sheets reside. The claim here is that the impact at 25°N may actually be more important. This is because the heating of the northern continents at 25°N has a large asymmetric impact on atmospheric winds that is not matched by the impact of the summer heating at 25°S. The net result is that the trade winds over the tropical Atlantic are very strong and very weak during opposite phases of the precessional cycle. These trade wind variations should have a direct impact on tropical SSTs, which can warm and cool the polar regions of both hemispheres.

The origin of this idea comes from modern observations which show that the return flow of the AMOC is drawn up to the surface as it passes through the tropical Atlantic (the upshift). The upshift produces a strong cooling that is readily felt by the rest of the tropics. The upshift, moreover, is modulated in a simple and direct way by the insolation at 25°N (via the trade winds) and the cooling effect is most pronounced when the northern summer insolation is weak and the trade winds over the Atlantic are strong. The Earth warms when the northern summer insolation is strong and the cooling effect is diminished.

The upshift, unfortunately, is not part of the overturning that is generated by modern climate models. So, even though the trade winds in the models might respond to the insolation in a realistic way, without the upshift, the warming and cooling seen in paleo records does not occur.

Acknowledgements

Much of the work on this paper was carried out while the author (JRT) was a member of the scientific staff at the Geophysical Fluid Dynamics Laboratory, which is part of National Oceanic and Atmospheric Administration (NOAA). The paper was completed after JRT retired in May 2019. He greatly appreciates the fact that GFDL has made his ongoing work possible with office space and computer support. The author would also like to thank Elizabeth Sikes, Nadir Jeevanjee, Danny Sigman, and Eric Galbraith for their very helpful internal reviews. Data were not used, nor created for this research.

References

Battisti, D. S., Q. Ding, and G. H. Roe, Coherent pan-Asian climatic and isotopic response to orbital forcing of tropical insolation, JGR Atmospheres, 119, 11,997–12,020, doi:10.1002/2014JD021960, 2014.

Bjerknes, J., Atmospheric teleconnections from the equatorial Pacific, Mon. Weath. Rev., 97(3), 163-172, 1969.

Bosmans, J. H. C., S. S. Drifjhout, E. Tuenter, F. J. Hilgen, and L. J. Lourens, Response of the North African summer monsoon to precession and obliquity forcings in the EC-Earth GCM, Climate Dynamics, 44, 279-297, 2015.

Chapman, M. R. and N. J. Shackleton, Millenial-scale fluctuations in North Atlantic heat flux during the last 150,000 years, Earth Planet. Sci. Lett., 159, 57-70, 1998.

Crowley, T. J., North Atlantic Deep Water cools the Southern Hemisphere, Paleoceanography, 7(4), 489-497, 1992.

Curry, W. B.., Late Quaternary deep circulation in the Western Equatorial Atlantic, In *The South Atlantic: Present and Past Circulation*, edited by G. Wefer, W. H. Berger, G. Siedler, and D.J. Webb, Springer-Verlag, Berlin, 577-598, 1996.

Gnanadesikan, A., A simple predictive model for the structure of the oceanic pycnocline, Science, 283, 2077-2079, 1999.

Hays, J. D., J. Imbrie, and N. J. Shackleton, Variations in the Earth's orbit: Pacemaker of the Ice Ages, Science, 194, 1121-1132, 1976.

Imbrie, J. et al., On the structure and origin of major glaciation cycles, 1, Linear responses to Milankovitch forcing, Paleoceanography, 7(6), 701-738, 1992.

Indermühle, A. et al., Holocene carbon-cycle dynamics based on CO_2 trapped in ice at Taylor Dome, Antarctica, Nature, 398, 121-126, 1999.

Iudicone, D., S. Speich, G. Madec, B. Blanke, The global conveyor from a Southern Ocean perspective, J. Phys. Oceanogr., 38, 1401-1425, 2008.

Kawamura, K. et al., Northern Hemisphere forcing of climate cycles in Antarctica over the past 360,000 years, Nature, 448, 912-916, 2007.

Klein, S. A., B. J. Soden, N.-C. Lau, Remote sea surface temperature variations during ENSO: Evidence for a tropical atmospheric bridge, J. Climate, 12, 917-932, 1999.

Koutavas, A., Temperature correlations between the eastern equatorial Pacific and Antarctica over the past 230,000 years, Earth Planet. Sci. Lett., 485, 43-54, 2018.

Lisiecki, L. E., M. E. Raymo, and W. B. Curry, Atlantic overturning responses to late Pleistocene climate forcings, Nature, 456, 85-88, 2008.

Liu, Z., S. G. H. Philander, and R. C. Pacanowski, A GCM study of tropical-subtropical upper-ocean water exchange, J. Phys. Oceanogr., 24, 2606-2623, 1994.

Liu, Z. et al., The Holocene temperature conundrum, PNAS, 111, E3501-E3505, doi: 10.1073/pnas.1407229111, 2014.

Lu, P., J. P. McCreary, and B. A. Klinger, Meridional circulation cells and the source waters of the Equatorial Undercurrent, J. Phys. Oceanogr., 28, 62-84, 1998.

Lumpkin, R. and K. Speer, Large-scale vertical and horizontal circulation in the North Atlantic Ocean, J. Phys. Oceanogr., 33, 1902-1920, 2003.

Marcott, S., J. D. Shakun, P. U. Clark, and A. C. Mix, A reconstruction of regional and global temperature for the past 11,300 years, Science, 339, 1198-1201, 2013.

McCreary, J. P., and P. Lu, Interaction between the subtropical and equatorial ocean circulations: the subtropical cell, J. Phys. Oceanogr., 24, 466-497, 1994.

McIntyre, A., W. F. Ruddiman, K. Karlin, and A. C. Mix, Surface water response of the Equatorial Atlantic Ocean to orbital forcing, Paleoceanography, 4(1), 19-55, 1989.

Mechoso, C. R. et al., The seasonal cycle over the tropical Pacific in coupled ocean-atmosphere general circulation models, Monthly Weather Review, 123, 2825-2838, 1995.

Pokras, E. M. and A. C. Mix, Earth’s precession cycle and Quaternary climate change in tropical Africa, Nature, 326, 486-487, 1987.

Prell, W. L. and J. E. Kutzbach, Monsoon variability over the past 150,000 years, J. Geophys. Res., 92(D7), 8411-8425, 1987.

Raymo, M. E., Ruddiman, W. F., N. J. Shackleton, and D. W. Oppo, Evolution of Atlantic-Pacific $\delta^{13}C$ gradients over the last 2.5 m.y., Earth Planet. Sci. Lett., 97, 353-368, 1990.

Ruddiman, W. F., *Earth's Climate: Past and Future*, W. H. Freeman and Co., New York, 465 pp, 2001.

Ruddiman, W. F. and A. McIntyre, Oceanic mechanisms for amplification of the 23,000-year ice-volume cycle, Science, 212, 617-627, 1981.

Ruddiman, W. F. and A. McIntyre, Ice-age thermal response and climate role of the surface Atlantic Ocean, 40°N to 63°N, Geol. Soc. Am. Bull., 95, 381-396, 1984.

Schmidt, M. W., H. J. Spero, and D. W. Lea, Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation, Nature, 428, 160-162, 2004.

Schmitz, W. J. and P. L. Richardson, On the sources of the Florida Current, Deep-Sea Res., 38, Suppl. 1., S379-S409, 1991.

Schneider, R. R., P. J. Müller, and G. Ruhland, Late Quaternary surface circulation in the east equatorial South Atlantic: Evidence from alkenone sea surface temperatures, Paleoceanography, 10(2), 197-219, 1995.

Schneider, R. R., P. J. Muller, G. Ruhland, G. Meinecke, H. Schmidt, and G. Wefer, Late Quaternary surface temperatures and productivity in the east-equatorial South Atlantic: Response to changes in trade/monsoon wind forcing and surface water advection In *The South Atlantic: Present and Past Circulation*, edited by G. Wefer, W. H. Berger, G. Siedler, and D.J. Webb, Springer-Verlag, Berlin, 527-551, 1996.

Sloyan, B. M., L. D. Talley, T. K. Chereskin, R. Fine, and J. Holte, Antarctic Intermediate Water and Subanatarctic Mode Water formation in the southeast Pacific: the role of turbulent mixing, J. Phys. Oceanogr., 40, 1558-1574, 2010.

Sobel, A. H., I. M. Held, and C. S. Bertherton, The ENSO signal in tropical tropospheric temperature, J. Climate, 15, 2702-2706, 2002.

Stott, L., K. Cannariato, R. Thunell, G. H. Haug, A. Koutavas, and S. Lund, Decline of surface temperature and salinity in the western tropical Padific Ocean in the Holocene epoch, Nature, 431, 56-59, 2004.

Straub, M., D. M. Sigman, H. Ren, A. Martinez-Garcia, A. N. Meckler, M. P. Hain, and G. H. Haug, Changes in North Atlantic nitrogen fixation controlled by ocean circulation, Nature, 501, 200-203, 2013.

Toggweiler, J. R. and B. Samuels, Effect of Drake Passage on the global thermohaline circulation, Deep-Sea Res. I, 42, 477-500, 1995.

Toggweiler, J. R. and B. Sauels, On the ocean's large-scale circulation near the limit of no vertical mixing, J. Phys. Oceanogr., 28, 1832-1852, 1998.

Toggweiler, J. R., E. R. M. Druffel, R. M. Key, and E. D. Galbraith, Upwelling in the ocean basins north of the ACC: 1. On the upwelling exposed by the surface distribution of $\Delta^{14}C$, JGR Oceans, 124, https://doi.org/10.1029/2018JC014794, 2019.

Toggweiler, J. R., E. R. M. Druffel, R. M. Key, and E. D. Galbraith, Upwelling in the ocean basins north of the ACC: 2. How cool subantarctic water reaches the surface in the tropics, JGR Oceans, 124, https://doi.org/10.1029/2018JC014795, 2019.

Wallace, J. M., Effect of deep convection on the regulation of tropical sea surface temperature, Nature, 357, 230-231, 1992.

Wang, C, L. Zhang, S.-K. Lee, L. Wu, and C. R. Mechoso, A global perspective on CMIP5 model biases, Nature Climate Change, 4, 201-205, 2014.

Wunsch, C., An estimate of the upwelling rate in the equatorial Atlantic based on the distribution of bomb radiocarbon and quasi-geostrophic dynamics, J. Geophys. Res., 89(C5), 7971-7978, 1984.

Zhang, R. and T. L. Delworth, Simulated tropical response to a substantial weakening of the Atlantic thermohaline circulation, J. Climate, 18, 1853-1860, 2005.

Figure Captions

Figure 1. **(top)** Abundance of N. pachyderma (sinistral) and IRD grains over the last 150,000 years from North Atlantic core SU90-03. **(bottom)** Summer sea surface temperatures (black diamonds) and winter sea surface temperatures (open diamonds) derived from faunal indicators. Copied from Chapman and Shackleton (1998).

Figure 2. Change in summer precipitation (red shades positive, blue shades negative) and change in 850 mb winds (vectors) between 218 kyr BP, when the northern summer insolation was particularly strong, and 207 kyr BP, when the insolation was particularly weak. Adapted from Battisti et al. (2014).

Figure 3. Impact of the trade winds on the return flow of the AMOC. **(a)** Illustration from Gnanadesikan (1999) showing how the return flow passes through a bowl of low-density fluid in the upper ocean. The bottom of the bowl in the diagram is the 5° isotherm. **(b)** Modification in which the bowl is divided into upper and lower layers separated by the 10° isotherm. SAMW and AAIW are added to the cool lower layer. NADW is removed from the warm upper layer. The trade winds bring about a shift of SAMW and AAIW from the lower layer into the upper layer.

Figure 4. **(a)** Shell-bound $\delta^{15}N$ in planktonic foraminifera from ODP Site 999 in the western Caribbean Sea plotted along with the precessional forcing (upside down). **(b)** $\delta^{18}O$ and **(d)** Mg/Ca SSTs from G. ruber in the same core. Panels **f** and **g** show $\delta^{15}N$ variations in the western equatorial Pacific and off the west coast of Mexico. Copied from Straub et al. (2013).

Figure 5. Antarctic air temperature (orange and black curves in **panel A**), atmospheric CO_2 (**panel B**), and sea level/global ice volume (**C,D**) over the last 340,000 years. Red curve in **panel A** is the summer insolation at 65°N; the black numbers above and below are isotope stages. The blue vertical lines denote minima in northern summer insolation. Adapted from Kawamura et al. (2007).

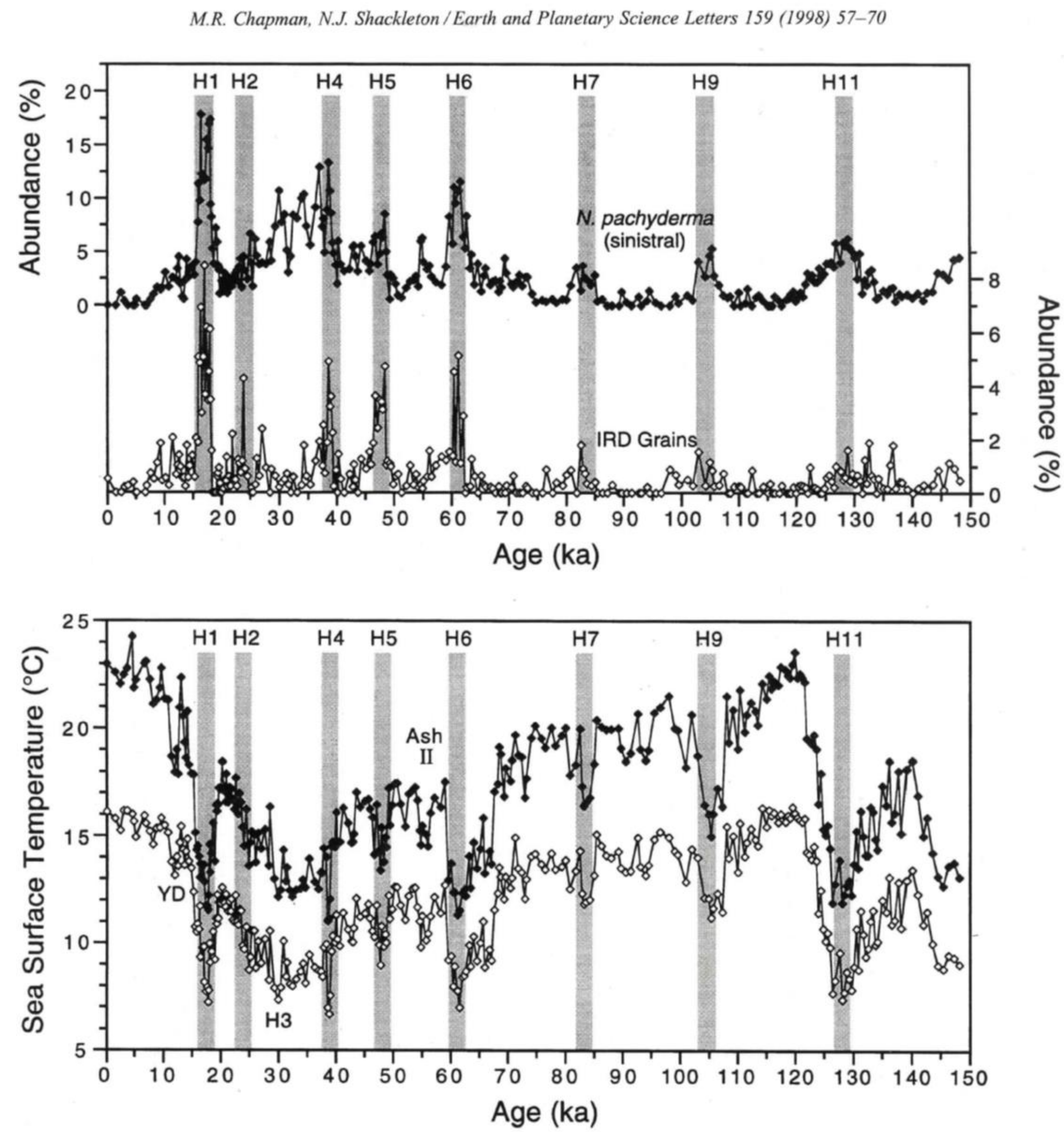

Figure 1. **(top)** Abundance of N. pachyderma (sinistral) and IRD grains over the last 150,000 years from North Atlantic core SU90-03. **(bottom)** Summer sea surface temperatures (black diamonds) and winter sea surface temperatures (open diamonds) derived from faunal indicators. Copied from Chapman and Shackleton (1998).

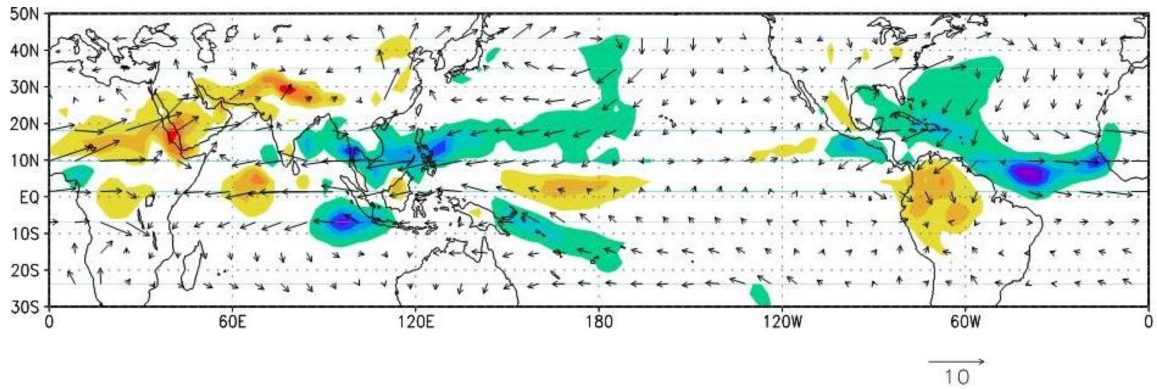

Figure 2. Change in summer precipitation (red shades positive, blue shades negative) and change in 850 mb winds (vectors) between 218 kyr BP, when the northern summer insolation was particularly strong, and 207 kyr BP, when the insolation was particularly weak. Adapted from Battisti et al. (2014).

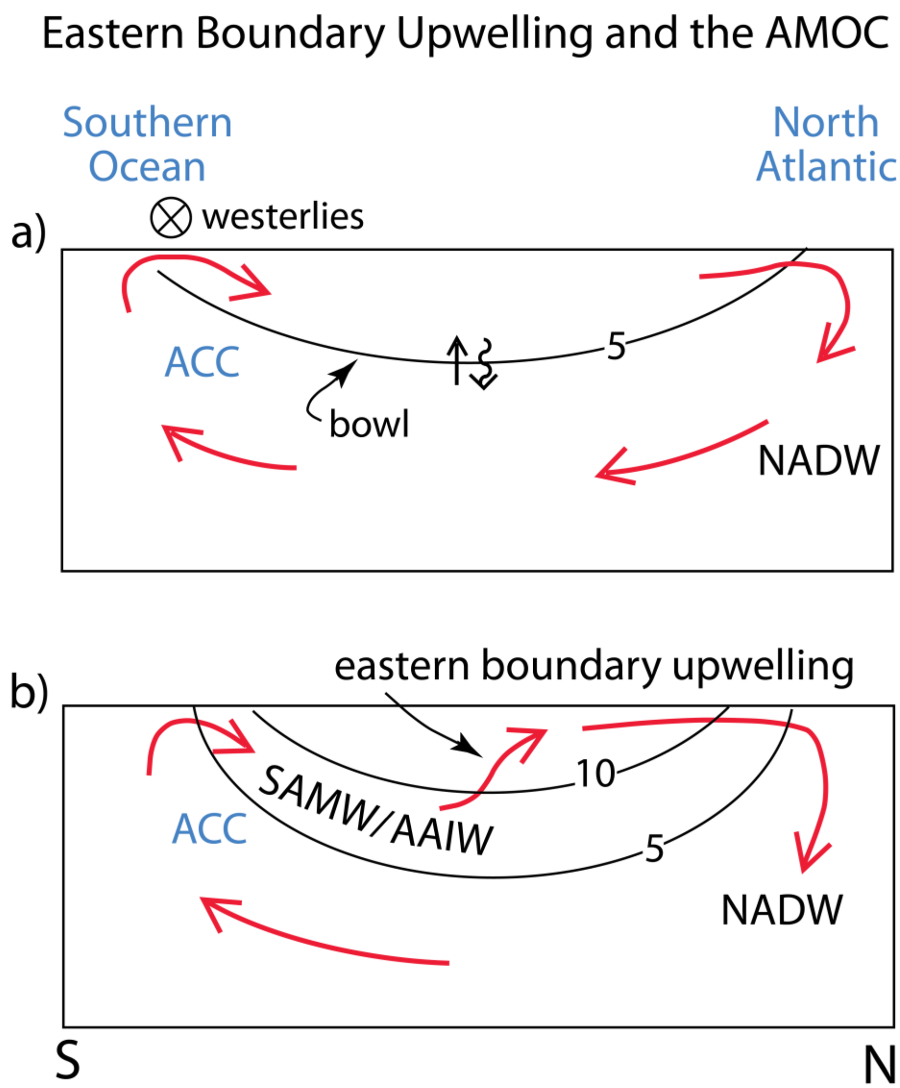

Figure 3. Impact of the trade winds on the return flow of the AMOC. **(a)** Illustration from Gnanadesikan (1999) showing how the return flow passes through a bowl of low-density fluid in the upper ocean. The bottom of the bowl in the diagram is the 5° isotherm. **(b)** Modification in which the bowl is divided into upper and lower layers separated by the 10° isotherm. SAMW and AAIW are added to the cool lower layer. NADW is removed from the warm upper layer. The trade winds bring about a shift of SAMW and AAIW from the lower layer into the upper layer.

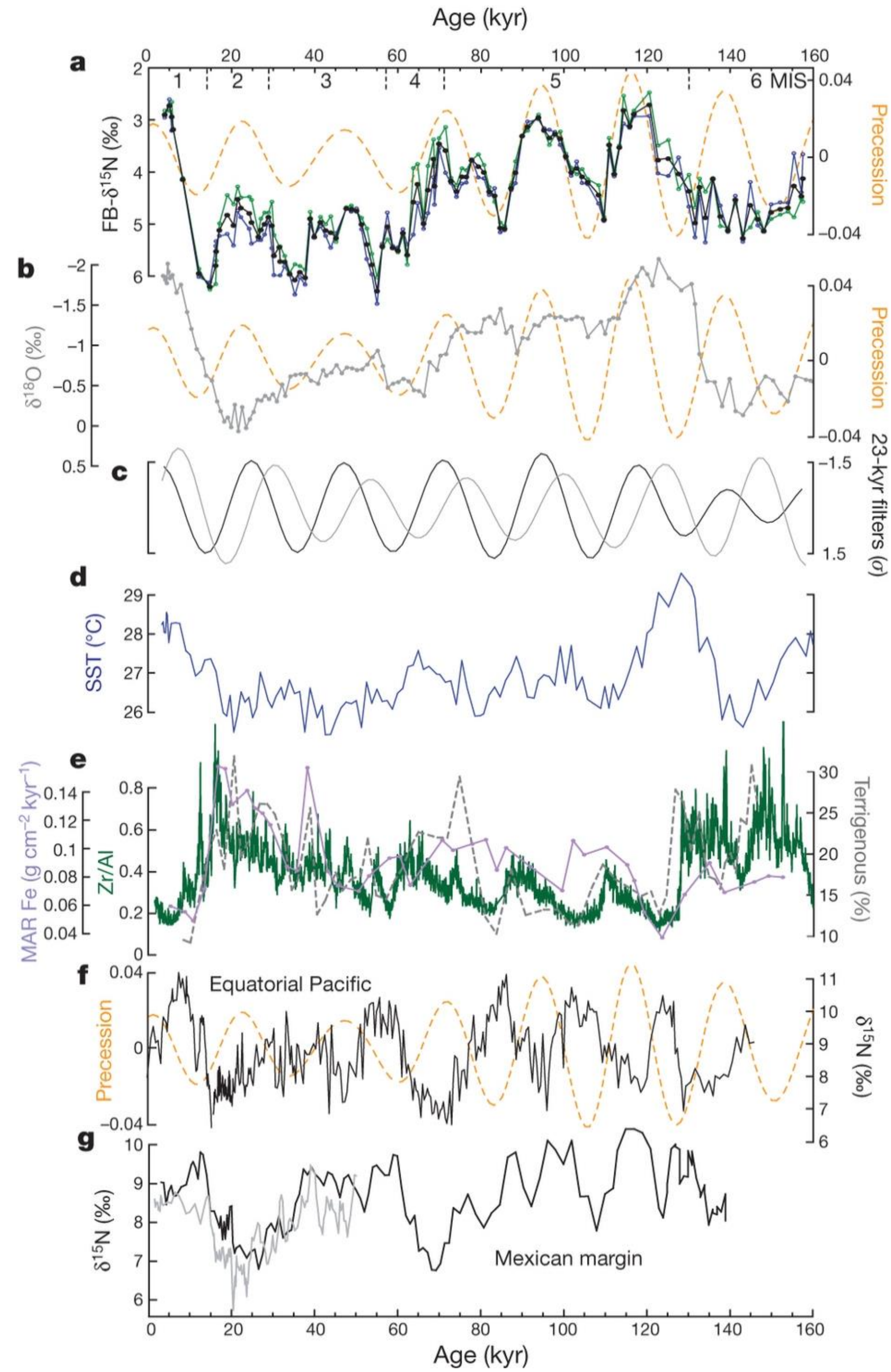

Figure 4. **(a)** Shell-bound $\delta^{15}N$ in planktonic foraminifera from ODP Site 999 in the western Caribbean Sea plotted along with the precessional forcing (upside down). **(b)** $\delta^{18}O$ and **(d)** Mg/Ca SSTs from G. ruber in the same core. Panels **f** and **g** show $\delta^{15}N$ variations in the western equatorial Pacific and off the west coast of Mexico. Copied from Straub et al. (2013).

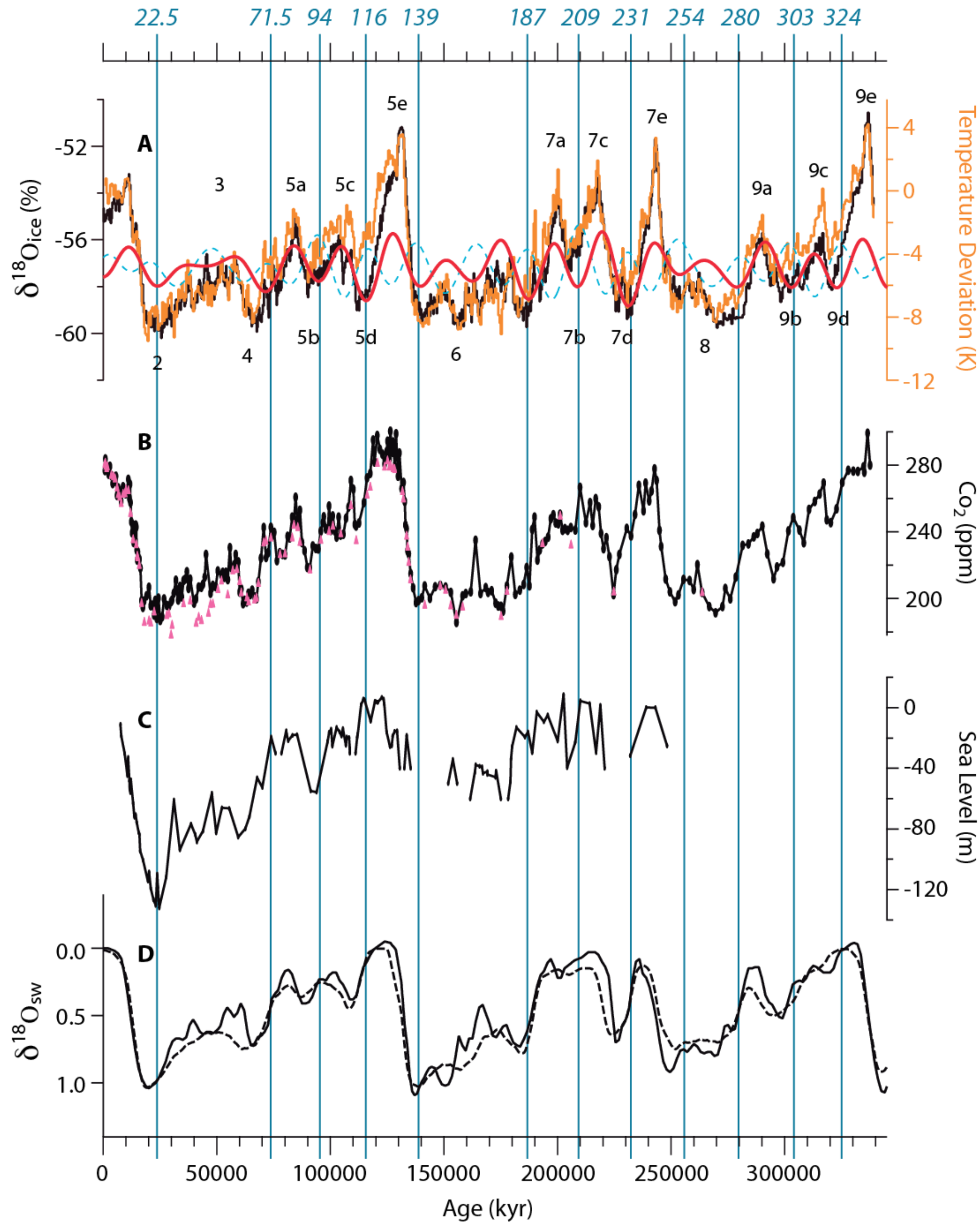

Figure 5. Antarctic air temperature (orange and black curves in **panel A**), atmospheric CO_2 (**panel B**), and sea level/global ice volume (**C, D**) over the last 340,000 years. Red curve in **panel A** is the summer insolation at 65°N; the black numbers above and below are isotope stages. The blue vertical lines denote minima in northern summer insolation. Adapted from Kawamura et al. (2007).