Paleophysical Oceanography with an Emphasis on Transport Rates

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Annu. Rev. Mar. Sci. 2010. 2:1–34

First published online as a Review in Advance on October 9, 2009

The *Annual Review of Marine Science* is online at marine.annualreviews.org

This article's doi:
10.1146/annurev-marine-120308-081056

1941-1405/10/0115-0001$20.00

Key Words

climate, ocean circulation, paleoceanography, Last Glacial Maximum

Abstract

Paleophysical oceanography is the study of the behavior of the fluid ocean of the past, with a specific emphasis on its climate implications, leading to a focus on the general circulation. Even if the circulation is not of primary concern, heavy reliance on deep-sea cores for past climate information means that knowledge of the oceanic state when the sediments were laid down is a necessity. Like the modern problem, paleoceanography depends heavily on observations, and central difficulties lie with the very limited data types and coverage that are, and perhaps ever will be, available. An approximate separation can be made into static descriptors of the circulation (e.g., its water-mass properties and volumes) and the more difficult problem of determining transport rates of mass and other properties. Determination of the circulation of the Last Glacial Maximum is used to outline some of the main challenges to progress. Apart from sampling issues, major difficulties lie with physical interpretation of the proxies, transferring core depths to an accurate timescale (the "age-model problem"), and understanding the accuracy of time-stepping oceanic or coupled-climate models when run unconstrained by observations. Despite the existence of many plausible explanatory scenarios, few features of the paleocircculation in any period are yet known with certainty.

1. INTRODUCTION

That the ocean is an important element of climate is a truism, and understanding it requires study of the three major divisions of climate: those of today, the past, and the future. Society has a strong interest in the future, but understanding the past and today are prerequisites. None of these divisions is distinct, as the ocean integrates over past events extending to thousands of years. Without adequate data, a scientist seeking to understand multidecadal, centennial, or longer-period oceanic variability can only speculate, and the year-by-year extension of the instrumental record will require a very long wait for adequate duration. Seeking insights from the record of past ocean states as embedded in paleoclimate data becomes compelling.

One may define the domain of paleoceanography as the period prior to widespread instrumental observations—perhaps dating to about 1990. That late date may be a surprise, but prior to the satellite era and other large-scale monitoring efforts of the 1990s, observations of the three-dimensional ocean were fragmentary. Physical oceanography then relied on indirect measures of the circulation—tracers such as temperature, salinity, phosphorus, radiocarbon, etc.—which are more analogous to the paleoceanographer's data types than are those from satellites and other recent innovations.

The domain of paleoceanography is so vast that there can be no pretense here to review anything but a fraction of it. One can sensibly ask what kind of ocean circulation existed billions of years ago with a fainter sun, what millennial or myriadic variability was like in the much warmer Cretaceous, or whether the tides had important effects during the Eocene, ad infinitum. All the elements of modern physical oceanography (coastal dynamics, internal waves, convection, geostrophic eddies, mixed layer physics, carbon uptake, etc.) are relevant to an understanding of the paleocirculation, but most of these are ignored here because they have been little studied in the context of the past.

The focus here is on determining the circulation, particularly the rates of movement, and on the basin-scale elements that control it. Some special attention is given to the Last Glacial Maximum (LGM, roughly 20,000 years before the present) because of the relative abundance of observations from that distinctive period. Depiction of the ocean during this interval raises many—but far from all—of the problems of paleoceanography more generally. We aim to make this review accessible both to paleoclimate scientists seeking some understanding of the problem of determining the movement of a global-scale fluid, and to physical oceanographers seeking some understanding of the particular issues involved in understanding the ocean circulation of the past. Our theme is climate, but paleoceanographic studies have many related purposes including, especially, understanding of the chemical and biological evolution of Earth as a whole.

Paleoceanographic inferences rely primarily on tracers laid down in geologic or biological deposits, which are generally referred to as "proxies." That is, the data are almost always more indirect representations of the physical field sought (e.g., a temperature, salinity, or water speed) than would be an instrumental measurement, and they are usually functions of other fields as well.[1] Thus, the ratio of concentrations of Mg and Ca in the shells of near-surface dwelling foraminifera (i.e., amoeboid protists) is commonly interpreted as representing water temperature at the time of formation of the shell in a process involving not only the temperature of the water, but also salinity and myriad biological processes (e.g., Bice et al. 2006). Another example is the δ^{18}O of a foraminiferal shell—the ratio of the mass number 18 isotope of oxygen to the more common number 16 isotope, where the anomaly is normalized by a modern standard value. This oxygen isotope ratio has been interpreted, with varying accuracy and precision, as indicative of water temperature, ice volume,

[1] Of course, direct physical variable measurements are almost nonexistent. Even the user of a mercury thermometer is measuring a length, not a temperature.

salinity, or density, depending on the context and the other proxies with which it is paired. No hard distinction exists between proxy data and more familiar modern tracer observations, particularly as preliminary efforts now exist to explicitly represent quantities such as $\delta^{18}O$ in climate models (e.g., Legrande et al. 2006), much as one attempts to model modern tritium or molecular oxygen values. Nonetheless, novel elements such as isotopic fractionation effects must be included.

One way to organize a discussion of the paleoceanographic problems is to summarize first what we know of the modern circulation and how and why we think we know it. No adequate textbook discussion of the observed modern ocean circulation exists, but its theoretical underpinnings are well served by a number of volumes (Pedlosky 1996, Thorpe 2005, Vallis 2006). Siedler et al. (2001) provided an encyclopedic overview of observational understanding.

For many purposes, essentially static descriptors of the ocean are of greatest interest: How much carbon existed as a function of depth and how did it change over long times? What was the mean salinity profile of the South Pacific? How much North Atlantic Deep Water (NADW) existed 20,000 years ago? What was the radiocarbon concentration in the deep Pacific? The pursuit of these and analogous quantities dominated physical oceanography for almost 150 years and were labeled water-mass properties, as they were the only elements that could be quantitatively described with the available technologies. Today, they are the focus of paleoceanographic study both to provide a basic description of the past oceans and, again, because such questions are the most readily accessible.

The ocean, however, is dynamically active through its transports of carbon, freshwater, enthalpy, etc., and it is these transports that ultimately control the distributions represented in the proxies. If one is both to understand why the distributions of the past were different and to calculate their influence, for example, on the global heat and water balance, the rate at which the water moves must be known.

2. SAMPLING ISSUES

A central issue is the number and distribution of observations. Despite a large modern database, major uncertainties remain in the description and understanding of the modern ocean circulation, and the comparatively slight database depicting the ocean of the past is a huge problem. **Figure 1** displays one of the more comprehensive core collections used for an Atlantic study in the LGM, but it is extremely limited relative to the undersampled modern situation. The relative paucity of data means that much of paleoceanography is unconstrained by observations and has led to sometimes fantastic speculations about how the system behaved in the past.[2]

High-frequency time variability, in this context anything with timescales less than approximately one decade, poses another problem for the interpretation of the paleorecord. Much of that proxy record comes from analyzing the molecular or atomic properties formed at discrete intervals. A plant or animal that grows primarily during one season does not record an annual average value, and such seasonally sampled records can easily be misinterpreted (e.g., Huybers & Wunsch 2003). Proxies such as grain size in deep-sea cores used to estimate water speeds at the sea bed may reflect rare, extreme events rather than any simple average (e.g., McCave & Hall 2006). More generally, any proxy whose sampling interval is determined by its environment yields a biased measure of that environment.

Paleoceanography relies most heavily on data from the cored sea floor, of which **Figure 2** (Shackleton et al. 1990) is an example. Analyzing data from cores confronts three major problems:

[2]The privately published, but widely circulated, essay by Stommel (1954) called forceful attention to the "dream-like" quality of physical oceanography in the era before development of adequate observing systems.

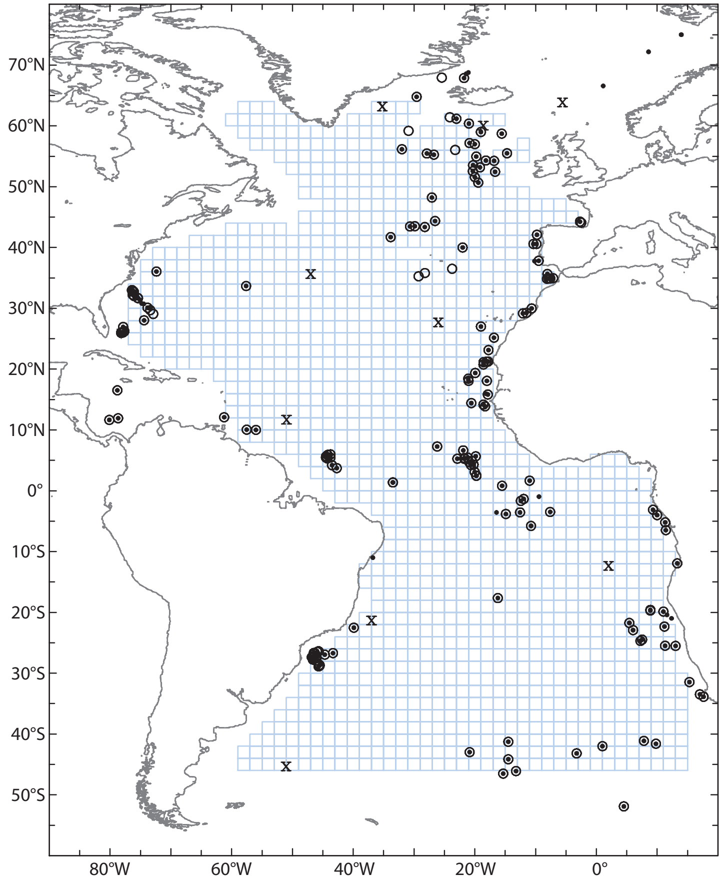


Figure 1

The distribution of cores in the North Atlantic. Solid dots are those available for the Holocene (approximately the past 12,000 years), open circles are those used for the Last Glacial Maximum, and lowercase exes denote some modern water column chemistry measurement positions. Compare the data density to that in **Figure 6** for the present-day ocean. Adapted from Marchal & Curry (2008); courtesy O. Marchal.

(*a*) Measurable proxies are sometimes poorly known functions of the actual physical variables sought. (*b*) Records are available only from regions of significant sediment accumulation on the sea floor (see **Figure 3** for a map of thickness). Vast sections of the ocean floor, e.g., the deep Pacific, are undersaturated with respect to calcium carbonate, inhibiting sediment accumulation and generally leading to significant alteration of the record, which does accumulate. Furthermore, a limit of approximately −90 My (e.g., Rowley 2002) exists when virtually all sediment is irretrievably lost to plate subduction. (*c*) The measurements are a function of depth,[3] but one generally needs time of deposition. Converting depth into age is a complex and often opaque process. Many other potential issues arise, including the stirring of sediments by burrowing animals (bioturbation, e.g., Bard et al. 1987), and downslope reworking of sediments.

[3]Even determination of depth presents difficulties, often requiring the splicing together of multiple drill cores as well as correction for sediment distortion both from in situ compaction and from the coring process.

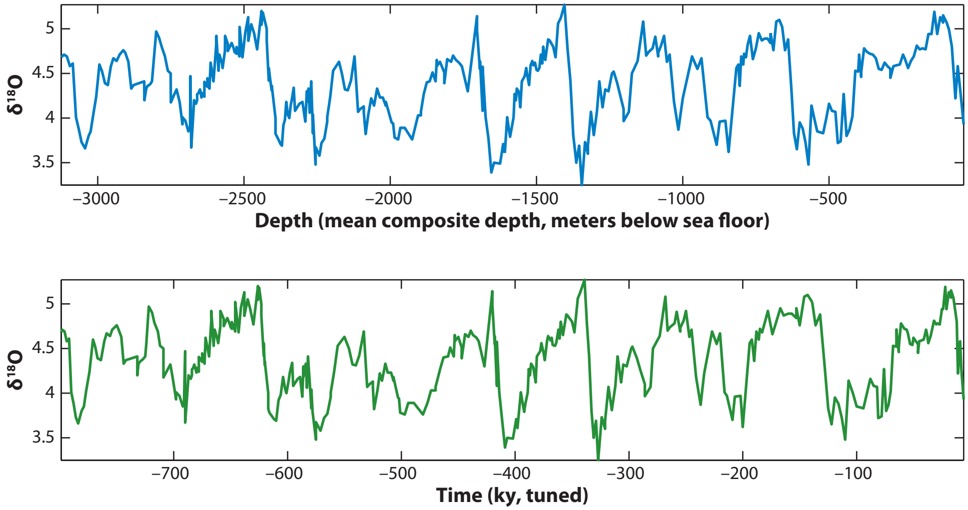


Figure 2

A $\delta^{18}O$ stratigraphy from Ocean Drilling Program site 677 in the Eastern Equatorial Pacific (Shackleton et al. 1990) plotted against depth (*blue*) and ages obtained from orbital tuning (*green*). This core is known for having a relatively constant accumulation rate, but even so, the interval spanned by the last glacial as a function of depth (between 100 and 400 meters below the sea floor) is wider, for example, than when plotted against the orbitally tuned time estimate (between 20 and 70 ky ago).

3. WATER MASSES: THE STATIC PROBLEM

The most prominent role the ocean plays in climate is as a reservoir—containing most of the freshwater on Earth, far more carbon than exists in the atmosphere, and great heat capacity. The concentrations of these and other materials dissolved in the oceans act as tracers, permitting the mapping of water-mass distributions. Tracers can be categorized as passive or active, depending on whether density is influenced, and as conservative or non-conservative, depending on whether

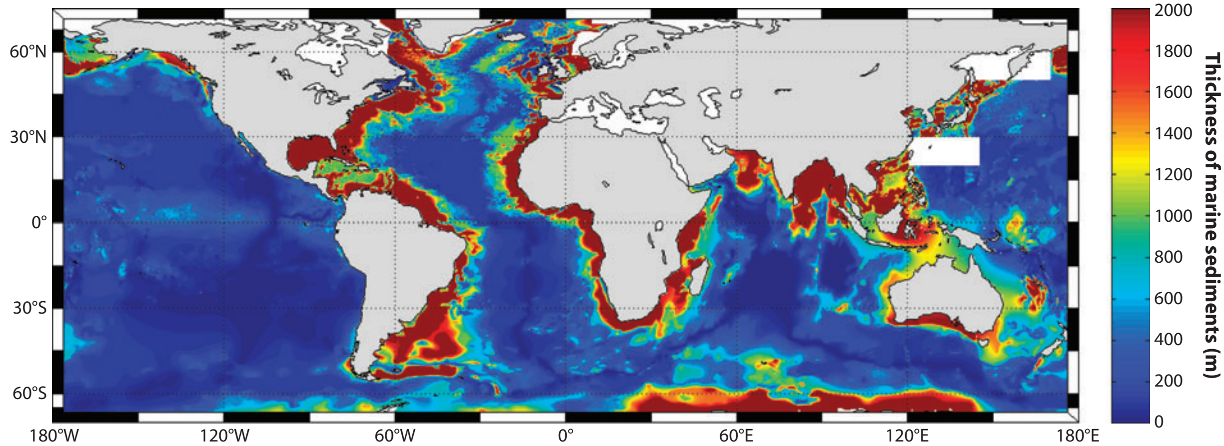


Figure 3

The thickness of marine sediments in meters (Divins 2002). Shading is saturated at 2000 meters to permit visual resolution of regions with thin sedimentary deposits—composing most of the oceans. Regions above sea level are in gray and missing data in white.

there are appreciable sources or sinks in the ocean interior. For example, the Mg/Ca ratio in foraminiferal calcite serves as a proxy for water temperature and is thus an active and approximately (over short times) conservative tracer, whereas the radiocarbon concentration in a foraminiferal shell can be a proxy for past water radiocarbon concentrations (under a series of assumptions) and is thus a nonconservative and passive tracer.

Distributions of tracers in the ocean, and hence their interpretation, depend on the tracer category, the boundary conditions, the rate of decay (if any), initial conditions, the flow field, and mixing rates. Given all these influences, even the so-called forward calculation is difficult. The inverse problem of inferring the flow field and/or mixing rates from observed tracer distributions remains a formidable one. We begin with what may be the simplest problem: defining and inventorying the various water masses present in the ocean.

Beginning in the mid-nineteenth century, charting of various properties (mainly temperature and salinity) was undertaken (see Warren 1981, Reid 1981), and today, properties including dissolved oxygen, nitrate, phosphate, and silica have been mapped (e.g., **http://www.ewoce.org**), although much coverage remains marginal. **Figure 4** shows the dominant water mass in the modern ocean—the so-called Pacific Deep Water, plus the large variety of much smaller volumes of water types displaying disparate property values. A recent decomposition (Gebbie & Huybers, submitted manuscript) (see **Figure 5**) indicates that a large number of water types fills the ocean's interior, with the first 10 accounting for only 40% of the ocean's volume, and the first 100 only 70%. Were the ocean in a true steady state, these volumes would be a consequence of a balance between water-mass formation and destruction, providing no information about either separately. (Detection of imbalances between production and destruction, however, may provide a window to changes in circulation.)

Consider the LGM by way of example. The most common proxies of the paleocean water masses are the stable carbon isotopic ratio [$^{13}C/^{12}C$ (usually expressed as $\delta^{13}C$)][4] and the cadmium-to-calcium (Cd/Ca) ratio of the carbonate shells of bottom-dwelling (benthic) foraminifera buried in marine sediments. The $\delta^{13}C$ and Cd/Ca ratios of benthic shells obtained from sediments reflect the $\delta^{13}C$ of dissolved inorganic carbon and the Cd content of bottom waters, respectively (Duplessy et al. 1984; Boyle 1988, 1992). $\delta^{13}C$ and Cd/Ca concentration ratios covary with nutrient content and are thus similar to phosphorus (see Boyle et al. 1976, Kroopnick 1985).

Updating and extending earlier efforts (e.g., Kroopnick 1985, Duplessy et al. 1984), Curry & Oppo (2005) compiled foraminiferal samples of $\delta^{13}C$ along margins, seamounts, and the ocean bottom for the LGM, forming a pseudotransect in the western North Atlantic. The prevailing interpretation of the LGM $\delta^{13}C$ distribution is that nutrient-poor NADW shoaled and that a nutrient-rich Antarctic Bottom Water filled a greater proportion of the deep Atlantic. Estimates made from mapping Cd/Ca ratios have been interpreted to support this view (Marchitto & Broecker 2006). Alternative hypotheses include ones that the nutrient content of NADW was different during the LGM or that NADW filled the depths of the northeastern Atlantic. The decomposition of the ocean using only one or two tracers becomes nonunique once it is recognized that many water types are present (alse see Wunsch 2007, pp. 58–59). Thus, as shown in **Figures 4** and **5**, quantitative inferences should not be drawn from two end-member calculations.

Sea-surface temperatures have been mapped by numerous groups (e.g., CLIMAP 1981, Mix et al. 1999, Kucera et al. 2006) and, as might be expected, they differ, with seasonal and other biases being of concern in each case. Although these provide an important upper-boundary condition,

[4] $\delta^{13}C = 1000[R_{\mathrm{sample}}/R_{\mathrm{standard}} - 1]$, where $R_j = [^{13}C]/[^{12}C]$.

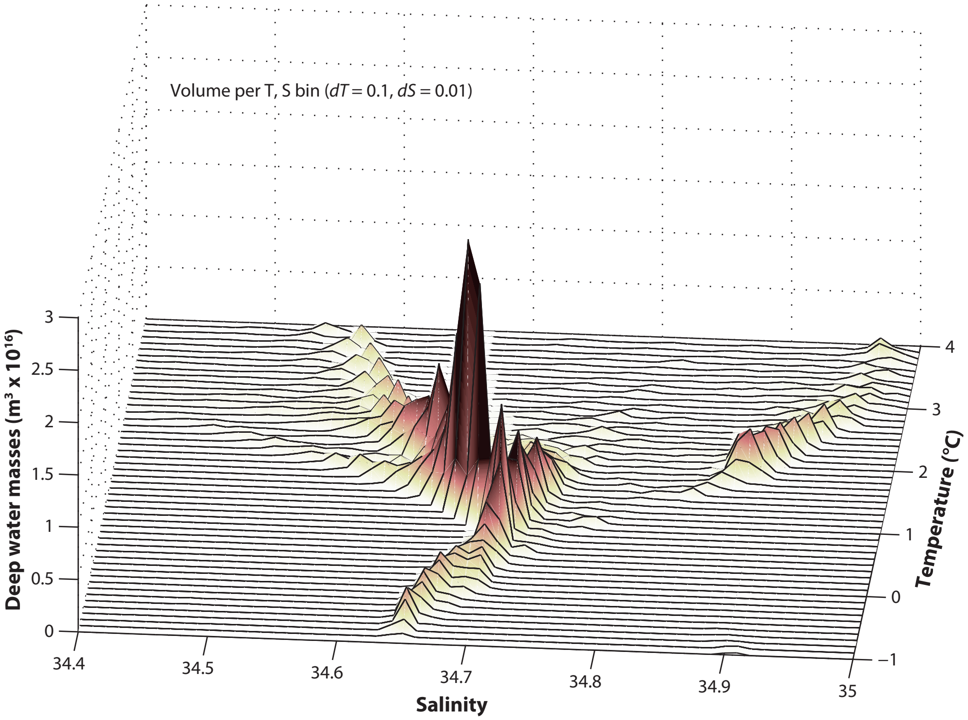


Figure 4

Approximate histogram of deep-water masses ($m^3 \times 10^{16}$) of the modern world's ocean (below potential temperature of 4°C derived from the analysis in Forget 2009; G. Forget, pers. commun.). These volumetric censuses shift with the climate state but by themselves carry no information about their rate of production or of their movement, other than in a quasi-steady state when it must be in balance with the rate of destruction. The largest peak here corresponds to the so-called Pacific Central Water of the abyss (see Worthington 1981, for a discussion of water-mass definitions and volumes). This present compilation is based on the modern hydrography resulting from the World Ocean Circulation Experiment program. Earlier versions (e.g., Worthington 1981) lacked coverage over much of the ocean, including the southern hemisphere in particular. Temperature (T) is in °C and salinity (S) is dimensionless as measured on the practical salinity scale. Note the large number of identifiable end members. The observed structures are not well understood (see McDougall & Jackett 2007).

their relationships to the three-dimensional circulation are weak and indirect, and water column data are usually used to make inferences about the ocean circulation.

Adkins et al. (2002), using the pore waters in the sediments at four deep-sea drilling sites (two in the North Atlantic and two in the Southern Ocean), concluded that the LGM abyssal ocean was more saline and more homogeneous in temperature than it is today—close to the freezing point. An LGM sea-level reduction of 130 m implies a global average salinity increase of 1.2, whereas the average change in the four data points from Adkins et al. (2002) is 1.5 ± 0.65. Today, NADW is the more saline, but Adkins et al. (2002) also concluded that the Southern Ocean was saltier during the LGM, approaching values of 37. Although the measurements are difficult to obtain, requiring pore water to be extracted from large amounts of sediment, it nonetheless seems that four data points are remarkably few for this purpose.

Even if one accepts the prevailing interpretation of these data, i.e., that the volume of NADW in the Atlantic decreased and that bottom density increased, the cause of the redistribution of these

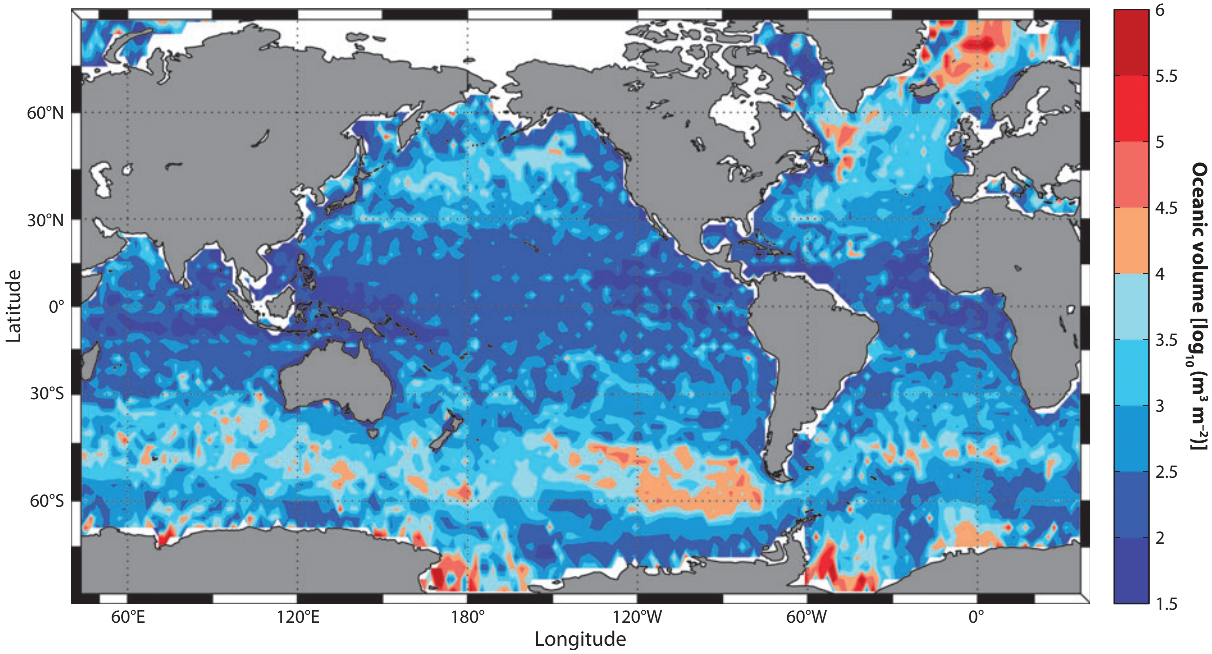


Figure 5

The volume of the ocean that originated from—i.e., last interacted with—surface locations resolved into 2° by 2° grid boxes (Gebbie & Huybers 2009). The color scale corresponds to volume in units of $\log_{10}$ (m^3 m^{-2}). Note the detailed structure and large number of surface points that contribute to the interior, defying any interpretation of the interior ocean as being filled from a small number of deep-water formation sites.

water-mass properties remains unclear. The multitude of mechanisms potentially contributing to the production, movement, and destruction of conservative tracer fields so far prevents conclusive inferences regarding the underlying changes in the circulation.

4. THE TRANSPORT PROBLEM

Water-mass volumes are not connected in any known way either to their rates of production or, separately, to removal, the "standing crop" being an accumulated difference of these two processes. Although many papers have simply assumed that an increased volume of fluid was the result of a larger production rate, there is no basis for such an inference—indeed there are counter examples (e.g., in a bathtub, the tap is usually turned off when the water volume is greatest). Paleoceanography addresses in several ways the problem of inferring water-movement rates, but two of them dominate: (*a*) the use of tracers depending in some known way on time and (*b*) geostrophic balance via the thermal-wind equations.

4.1. Radiometric Tracers

Consider ^{14}C, which is produced in the atmosphere when nitrogen-14 absorbs a neutron. This radiocarbon then decays back into stable nitrogen-14 with a half-life of approximately 5730 years according to the following rule:

$$dC/dt = -\lambda C. \tag{1}$$

As C indicates a generic tracer concentration, the solution is

$$C = C_0 \exp(-\lambda t), \tag{2}$$

where C_0 is the initial radiocarbon concentration. In the simplest case, if C_0 is known and no sources or sinks of carbon exist, the elapsed time is

$$t_C = -\frac{1}{\lambda} \ln\left(\frac{C}{C_0}\right), \tag{3}$$

i.e., the so-called radiocarbon age. Equation 3 is the basis of radiocarbon dating of a huge variety of samples in many fields from anthropology to climate.

Equation 3 has been applied to modern seawater, yielding radiocarbon ages ranging from a few 100 years in the North Atlantic to approximately 1500 years in the North Pacific at depth (see figure 2.4.3 in Sarmiento & Gruber 2004; Matsumoto 2007).

In a moving, mixing fluid, the governing equation has the far more complicated form

$$\frac{\partial C}{\partial t} + \mathbf{v} \cdot \nabla C - \nabla (\mathbf{K} \nabla C) = -\lambda C + q, \tag{4}$$

where C is the generic tracer; $\mathbf{v}$ is the three-dimensional, usually time-varying, flow field; $\mathbf{K}$ is a mixing tensor (e.g., Vallis 2006); and q represents any interior source or sink (e.g., remineralization, if one is discussing carbon isotopes, or exchange with particles in reactive species as in the thorium series). Equation 4 can be solved by standard numerical methods if initial conditions $C(\mathbf{r}, t = 0) = C_I(\mathbf{r})$ and boundary conditions $C_b(\mathbf{r}_b, t) = C(\mathbf{r} = \mathbf{r}_b, t)$, where $\mathbf{r}_\mathrm{b}$ is the boundary surface (most typically taken as the sea surface), the flow field ($\mathbf{v}$), the mixing tensor ($\mathbf{K}$), and the source/sink (q), are all specified. If the assumption of a steady state is justified (requiring that all terms, including the flow field, be unchanging), then the resulting solution will tend to be independent of the initial conditions through diffusive loss of structure. But it will still be a function of the remaining externally imposed functions and parameters, $C_\infty = C(C_b(\mathbf{r}_b), \mathbf{v}, \mathbf{K}, \lambda, q, \mathbf{r})$, and it will be a solution to Equation 4 without the first term. Given all the dependencies of C_∞, no quantitative interpretation of its distribution is simple.

What is usually required is not the apparent age, which is a nonlinear transformation of the tracer concentration (Equation 3), but the flow field, $\mathbf{v}$, and the mixing tensor, $\mathbf{K}$. That is, even with an estimate of the time since a water mass was present at some other location, an estimate of the distance the mass moved is needed to compute a rate, and lower and upper bounds on the rate will follow from the most direct or most tortuous admissible routes. If C is known sufficiently accurately everywhere, one can seek $\mathbf{v}$ and $\mathbf{K}$ so as to determine the rates of water movement. The problem is then a tractable linear, static, inverse problem (which is linear only for passive tracers not influencing the flow field or mixing tensor and when there are no uncertainties in C) (McIntosh & Veronis 1993; Wunsch 1996, 2006). But because of the forbidding data-distribution problems involved with defining three-dimensional tracer gradients and in part because the data coverage (**Figure 6**) (Key et al. 2004) used to determine the ^{14}C distribution is still limited, this calculation has never been directly attempted, even in the modern ocean, except in small regions. The generic tracer-age problem is discussed extensively by Jenkins (1980), Wunsch (2002), Waugh et al. (2003), and others.

Determining the concentration of radiocarbon that seawater had at some point in the past is even more involved than the modern problem, as it requires correcting a given sample for subsequent radiocarbon decay. One method is to make paired radiocarbon measurements on planktic (surface-dwelling) and benthic (bottom-dwelling) foraminifera and use the planktic radiocarbon concentrations to correct for the aging effects in the benthic values (Broecker et al. 1984, Adkins & Boyle 1997). It is also necessary to correct for fractionation effects. Biological processes take up

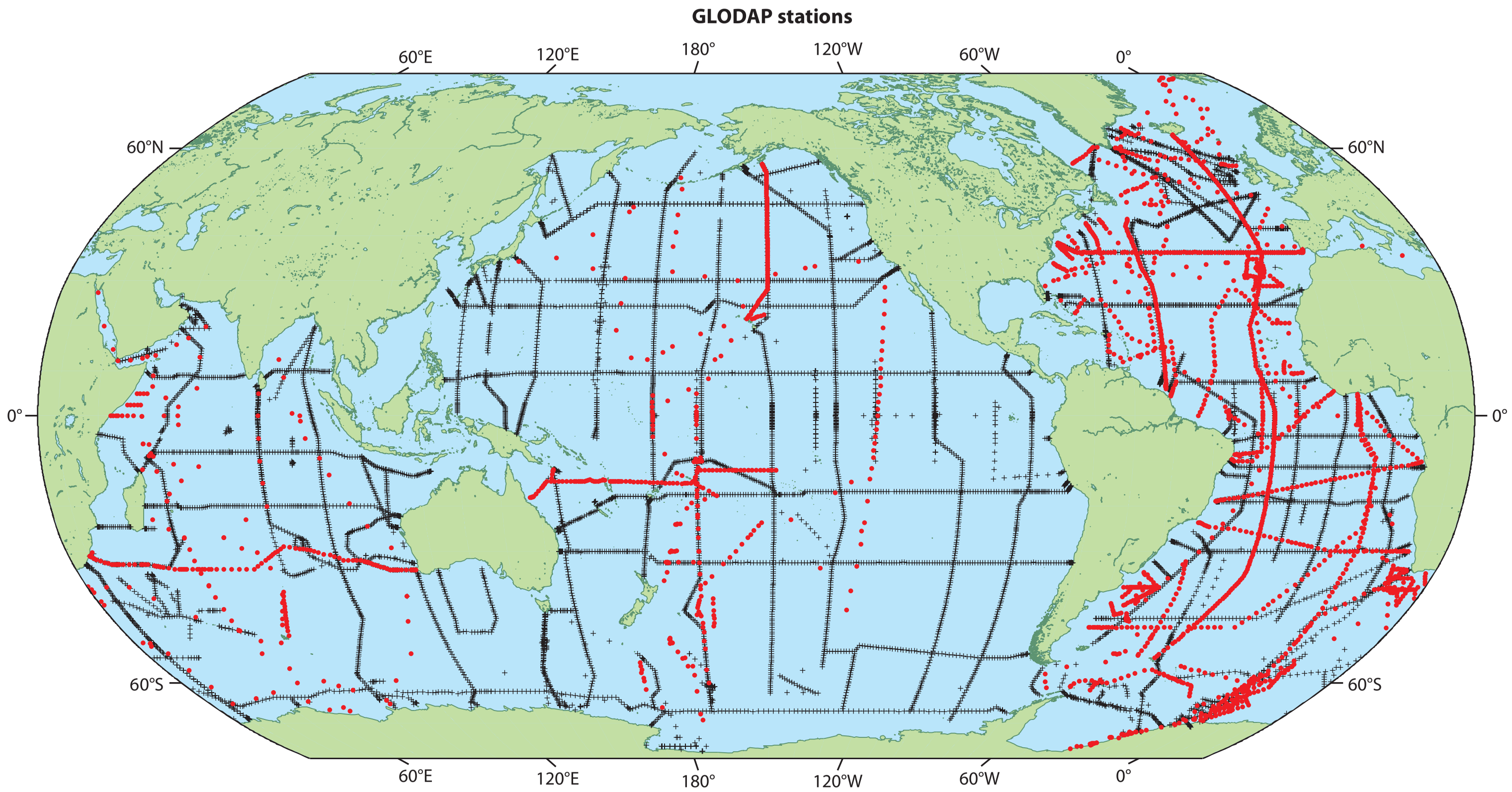


Figure 6

The distribution of stations used to compute the modern radiocarbon age (Key et al. 2004; courtesy of R. Key). In the Atlantic, compare it to the positions in **Figure 1**. Red dots denote data obtained prior to 1990 using measurement standards somewhat lower than those indicated by the black crosses.

^{12}C preferentially to ^{14}C, but assuming this selection is mass dependent, it can be estimated and corrected for by using measurements of ^{13}C, which is stable. Bioturbation can seriously complicate paired radiocarbon estimates, but this effect can be limited by sampling only at abundance peaks (e.g., Keigwin 2004), wherein the mixing of older or younger samples may have less influence upon the result. It is also possible to pair radiocarbon measurements in corals with age estimates determined from uranium series decay (e.g., Robinson et al. 2005). Another issue is that both the atmospheric radiocarbon concentration, and its partitioning with the ocean, appear highly variable. For example, during the last deglaciation, the atmospheric radiocarbon concentration appears to have decreased rapidly compared with the decay rate (e.g., Broecker & Barker 2007), making it necessary to account for the influence of time-variable atmospheric radiocarbon on ocean values.

Another potentially informative tracer of flow rates in past oceans is the $^{231}Pa/^{230}Th$ ratio of bulk sediment (Yu et al. 1996, Marchal et al. 2000, McManus et al. 2004). In the absence of oceanic transport or variations in the particulate flux settling through the oceans, there would be a fixed ratio of 0.093 between the rates at which ^{231}Pa and ^{230}Th are buried at the bottom of the ocean. But in the presence of circulation, the fact that ^{231}Pa does not adhere to particles as well as does ^{230}Th becomes important. ^{231}Pa has a longer oceanic residence time prior to burial, $\sim$100–200 years as opposed to $\sim$30 years for ^{230}Th, making the amount of ^{231}Pa buried in sediments more sensitive to oceanic transport. Relative burial rates of Pa and Th are also sensitive to the amount and composition of the particles settling through the oceans (e.g., Chase et al. 2002), and this can greatly complicate the interpretation of burial ratios (Marchal et al. 2000, Gherardi et al. 2005, Siddall et al. 2007). Initial results from sites near Bermuda (McManus et al. 2004) and Portugal (Gherardi et al. 2005) indicated that the $^{231}Pa/^{230}Th$ burial ratio increased there between 18–15 ky and again near 12 ky, suggestive of decreased meridional overturning circulation during these periods. However, acquisition of three more cores from the North Atlantic (Hall et al. 2006, Gherardi et al. 2008) has demonstrated more heterogeneous lateral and depth variations, as is anticipated from the spatial and temporal complexity of the known modern circulation (see, e.g., Wunsch 2007) and particle fluxes. Space-time sampling problems are omnipresent.

4.2. Sortable Silts

Measurements of sortable silts in marine cores (e.g., McCave 2007) provide a direct constraint on the rate of past bottom flows, though issues involving sediment source effects, spatial and temporal variability of flow over bedforms, and the influence of ice-rafted detritus all complicate interpretation (see McCave & Hall 2006). On the basis of six silt records, McCave et al. (1995) inferred that a weaker flow existed in the deep North Atlantic (>2000 m) relative to that of the late Holocene, while a more rapid circulation existed at mid-depths, (1000–2000 m). A more recent study, however, finds little change across these time periods at mid-depths along the Reykjanes Ridge (Praetorius et al. 2008). In the Southern Hemisphere, decreased flow rates were found at a deep site in the South Cape Basin during the LGM (4600 m depth; Kuhn & Diekmann 2002), while increased flow was found off of the east coast of New Zealand (3300 m depth; Hall et al. 2001). In the Amirante Passage of the Indian Ocean, little change was found between the LGM and Holocene (McCave et al. 2005). The collection of silt data from the LGM is perhaps most readily interpreted as indicating spatially heterogeneous flow changes relative to the late Holocene. These and other studies appear promising, but as with $^{231}Pa/^{230}Th$, a more comprehensive understanding of the space-time variability and temporal averaging seems necessary before drawing conclusions regarding widespread changes in past circulation.

4.3. Geostrophy

Observational understanding of modern ocean circulation rests on knowledge of the density field, ρ, and so-called geostrophic-hydrostatic balance in the forms

$$f\rho v = -g \int_{z_0}^{z} \frac{\partial \rho}{\partial x} dz + b(z_0, x, y), \tag{5}$$

and

$$f\rho u = g \int_{z_0}^{z} \frac{\partial \rho}{\partial y} dz + c(z_0, x, y), \tag{6}$$

where x, y, and z are three Cartesian coordinates; f is the Coriolis parameter; g is gravity; and u and v are the two horizontal velocity components. ρ is inferred from measurements of temperature, salinity, and pressure through an empirical equation of state. u and v estimated in this form are called the thermal wind. Except for very close to the equator, and in very strong boundary currents, large-scale flows in the ocean satisfy this balance to high accuracy. Assertions that in such a balanced state the flow is "driven" by the density (or corresponding pressure) gradients represent a misunderstanding. In a balanced state, one can equally well argue that the density differences are driven by the flow. The physics of the establishment of the balanced state involve many more processes, and elaborate calculations are required before any assertions of "cause" could be justified.

Level-of-no-motion. The main issue in the use of Equation 5 or 6 was, historically, the observationally intractable problem of determining b (or c), which led to the assumption that if z_0 were sufficiently deep, then $b, c \approx 0$, implying no flow at $z_0(\theta, \lambda)$. Although arguments persisted in the literature over exactly how to choose $z_0(\theta, \lambda)$, the need to assume the existence of a "level-of-no-motion" to proceed in calculating the thermal wind led to this assumption becoming embedded in the textbook literature as an article of faith. The reference-level-velocity problem distorted physical oceanography for more than 75 years. A discussion of its solution and its modern interpretation can be found in Wunsch (1996, 2007).

By the late 1970s, the development of inverse and related methods largely solved the level-of-no-motion problem, primarily through the use of additional constraints derived from explicit conservation requirements for mass, salt, etc., and dynamical tracers such as potential vorticity. These methods also permit calculation of solution uncertainties. Estimates now exist (e.g., Ganachaud & Wunsch 2002, Ganachaud 2003) of quantities such as the enthalpy or nutrient transport, which have sufficiently small formal uncertainties as to be useful. Some of the challenge in arriving at accurate estimates can be understood by noting that a 1 mm/s error in the flow field, extending the width of the Pacific Ocean from 1000 m to the bottom, would represent a volume transport error of approximately 30×10^6 m^3/s (30 Sverdrups, Sv), approximately equal to the volume transport of the Gulf Stream at Florida. Thus, very small velocity errors can have large oceanographic implications. A disputatious literature exists concerning the modern rates of meridional overturning circulation (roughly 15 Sv) in the North Atlantic. These estimates usually differ by 2–3 Sv, and they are not distinguishable given even the modern database.[5]

Eddies and the problem of scales of motion. The existence of large-scale and contourable tracer structures led to the misleading inference that the flow field giving rise to them must also be large scale and steady. It is now clear that the intense turbulence superimposed on the

[5]Distinguishing decadal changes of this magnitude presents a formidable problem for groups now attempting to forecast the MOC.

large-scale structures contain 90–99% of the kinetic energy of the flow (see Ferrari & Wunsch 2009). Ignoring it is equivalent to discussing climate under the assumption that the existence of weather is unimportant. Two separate problems arise: In the presence of eddies, climate records are noisy—making it difficult to identify and extract climate signals of interest (Wunsch 2008)—and the eddies can have important, even dominant, influence on the nature and behavior of the much larger-scale space-time property structures.

Almost no component of the ocean circulation is time invariant. After nearly 150 years of regarding the ocean as having an essentially fixed, time-independent circulation and properties, the discovery that everything was changing to a certain degree produced an intellectual turmoil not yet recognized by many investigators: Decadal records showing, e.g., trends in salinity or heat content are still published as though they are (*a*) astonishing and (*b*) necessarily representative of longer-term trends, neither of which is obviously true.

4.4. Paleoceanographic Geostrophy

Efforts have been made to employ geostrophy during the LGM (Legrand & Wunsch 1995; Ortiz et al. 1997; Lynch-Stieglitz et al. 1999a, 1999b, 2006; Lynch-Stieglitz 2001) and these all depend foremost on the success with which paleodensity can be reconstructed. The ability to constrain past density comes from the fortunate, albeit imperfect, happenstance that the oxygen isotopic ratios in calcite shells ($\delta^{18}O_{calcite}$), when precipitated in equilibrium with seawater, tend to increase when salinity, S, rises and temperature, T, declines (Lynch-Stieglitz et al. 1999b), giving the approximate expression, $\delta^{18}O_{calcite} \sim aT + b + cS + d$, where the constants b and d are kept distinct to facilitate discussion. Laboratory and field studies indicate that a is approximately $-0.2‰$/degree C (Kim & O'Neil 1997), consistent with the fractionation expected under thermodynamic equilibrium, but that b varies depending on which foraminiferal species is analyzed. Covariation between $\delta^{18}O_{calcite}$ and S arises because evaporation and precipitation tend to influence salinity and the $\delta^{18}O$ of seawater similarly, though different regions of the ocean show differing slopes, c, (e.g., Craig & Gordon 1965, Broecker 1986) and intercepts, d. Modeling (LeGrande et al. 2006) and observational studies (Adkins et al. 2002) indicate that the relationship between salinity and the $\delta^{18}O$ of seawater will change with the climate. Furthermore, no unique relationship between density and $\delta^{18}O_{calcite}$ exists because their dependencies on T and S differ (Gebbie & Huybers 2006) and, to a lesser degree, because of the nonlinearities in the equation of state. (Unless obvious from the context, subscripts are used to distinguish $\delta^{18}O_{calcite}$ and $\delta^{18}O_{seawater}$.) There is also the question of whether the proxy data properly average flows such as seen in **Figure 7**.

Lynch-Stieglitz et al. (1999a, 1999b) used sequences of sediment cores taken across the sloping margins on either side of the Florida Straits to reconstruct horizontal gradients in $\delta^{18}O_{calcite}$ during the LGM, and assuming a constant relationship with density, they used these to reconstruct density gradients. From the thermal-wind relationship, it was then found that the shear was reduced relative to the values seen in the Florida Straits today. But absent further information, one has the same reference-level-velocity problem as discussed above: No physical relationship exists between shear and absolute velocity.[6] Lynch-Stieglitz et al. (1999a, 1999b) assumed that a level of no motion existed at the sea floor and concluded that LGM transports were reduced to between 14 Sv and 21 Sv relative to Holocene values of approximately 30 Sv. Taking a cross section of 100 km times a mean depth of ~500 m, the approximately 10-Sv difference could be accounted for if there had been a bottom velocity of ~20 cm/s during the LGM.

[6]Needler's (1985) formula produces a relationship for the absolute velocity in terms of the density, but it applies in principle only to an eddy-free oceanic interior and involves such high derivatives of the density field as to be of theoretical interest only.

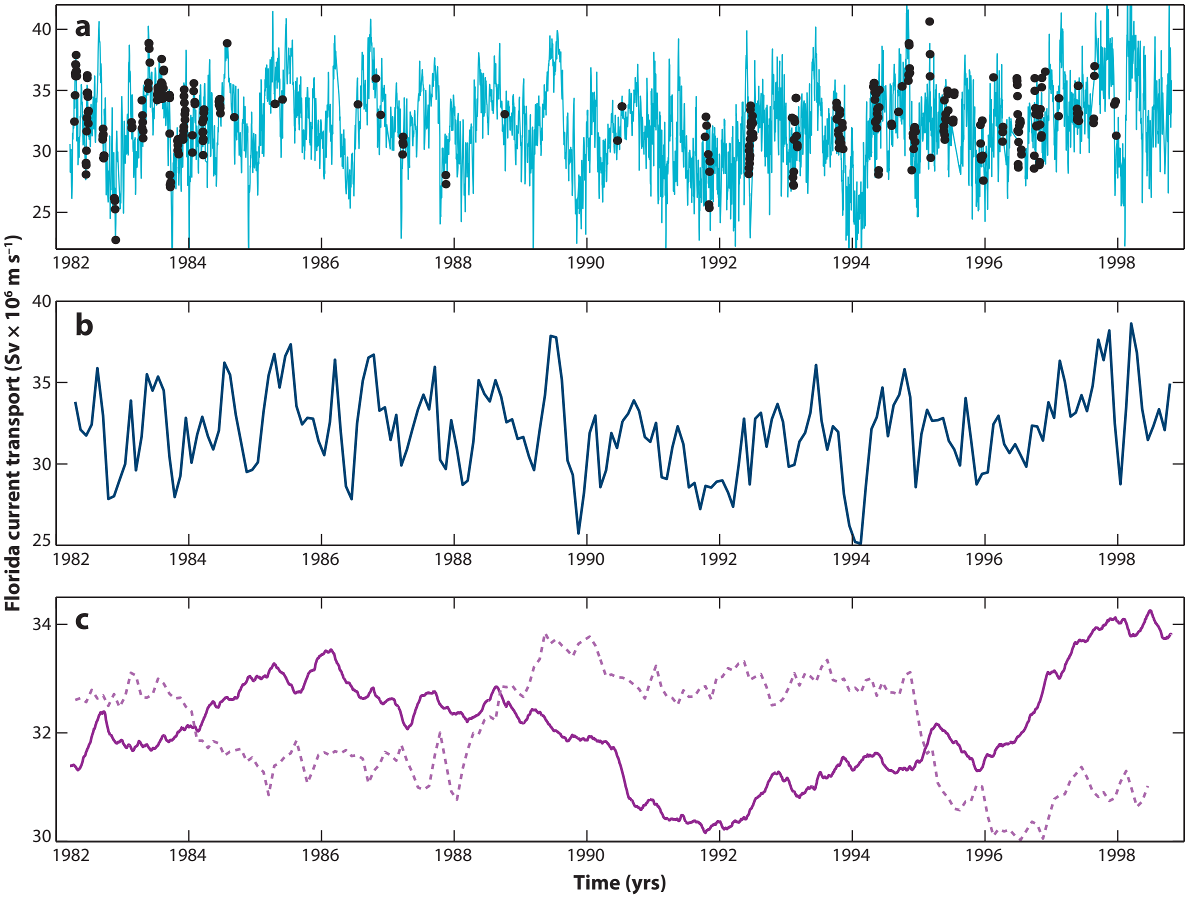


Figure 7

Measured values through time of the volume transport through the Florida Straits (Baringer & Larsen 2001; courtesy M. Baringer). Values reach as high as 40 Sv and as low as 25 Sv, raising the question of what values are reflected in the temperature proxies of the sediments. Panel (*a*) displays the inferred daily transports, with the black dots indicating the calibration times and values; (*b*) displays the running monthly mean transports; and (*c*) shows the running annual means and the North Atlantic Oscillation (NAO) index. Direct connection to wind forcing is obtained only on timescales of many years.

Was the bottom flow in the Florida Straits greater during the LGM than today? **Figure 8** shows a modern estimate of the transport, averaged over two years (Leaman et al. 1987), whose time mean is 32 ± 3 Sv. It is apparent that the flow at the bottom is finite and approximately 10 cm/s everywhere.[7] We do not know whether bottom velocities were larger in the past,[8] but

[7]The Florida Straits were the site of G. Wüst's important demonstration of the utility of geostrophic balance. Warren (2006) pointed out, however, that much of his inference does not withstand scrutiny. Fortunately, later tests of geostrophy (see references in Wunsch 1996, p. 76) show it to be a very good approximation, albeit with some crucial exceptions.

[8]We note the potential for sortable silt to constrain bottom velocity and therefore overcome some of the problems referenced with respect to identifying a geostrophic level of no motion. Such silts are scoured off the deepest sections of the Florida Straits (W. Curry, personal communication), but perhaps the distribution of silts along the margins would be of some use.

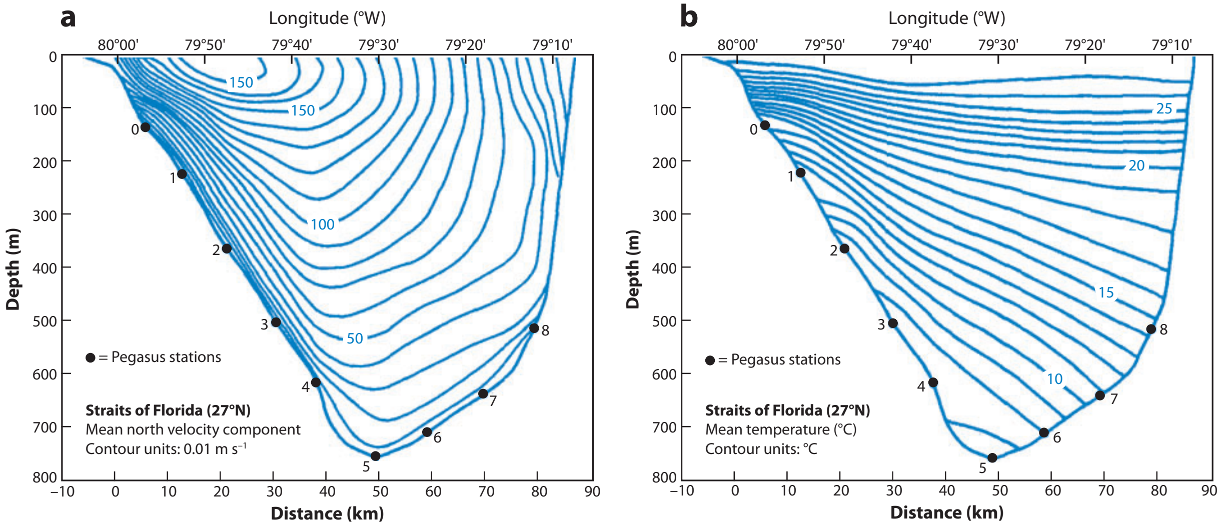


Figure 8

(*a*) Time-average velocity and (*b*) temperature field in the modern Florida Current near Miami (Leaman et al. 1987). In (*a*), the bottom velocity is 10 cm/s almost everywhere. In (*b*), note the complex structure of the temperature field, with significant shifts in the inter-isotherm distances on the east. Both graphs also display the standard deviations of these fields. Little seems to be understood of how the temporal variability affects the paleotemperature proxies. "Pegasus" is a direct velocity measuring device.

there appears no theoretical or observational reason to rule out such a change, and given a lowering of sea level by 130 m, an increased bottom flow during the LGM seems plausible.

Lund et al. (2006) made a similar estimate with similar conclusions for the past millennium, but they also extrapolated the inference of a reduced mass transport to speculate that the entire North Atlantic Ocean had a reduced heat transport during the Little Ice Age.[9] The ability to determine heat transport requires knowing not only the mass transport, but also the change in heat content integrated along mass conserving sections everywhere along the flow paths. Indeed, Murakami et al. (2008) found that three coupled climate models all showed a reduction in the volume transport during the LGM, but that they also all showed an increase in the oceanic heat transport. Volume and property transports are distinct. (Model problems are taken up below.) More generally, no reason exists to think that changes in the ocean circulation were uniformly of one sign—parts may have slowed while other parts intensified, as seen in the modern world.

Lynch-Stieglitz et al. (2006) also attempted to constrain the vertical shear at depths between 200 and 2000 m in the South Atlantic. Here, one has the advantage of being able to employ the additional constraint that the total transport across the South Atlantic is near zero, save for a contribution flowing across the Bering Strait and minor contributions from the net of evaporation, precipitation, and runoff. Their reconstruction of density suggested a decrease, or even reversal, in shear at mid-depths. The result is still inconclusive, however, because the contribution from a surface Ekman flux is unknown, the density estimates from the eastern margin are distributed over the latitude range 4°S to 34°S, and the exact relationship between

[9] Use of geostrophy in this region is not straightforward. The two profiles they used to compute the shear do not lie transverse to the stream, but instead are far displaced in the streamwise direction too, so that the Bernoulli effect on pressure cannot be ignored (see Chew et al. 1982). The downstream profile was also obtained in a region where the current is undergoing a near-right angle turn, thus introducing cyclostrophic effects.

$\delta^{18}O_{calcite}$ and density is poorly constrained (Lynch-Stieglitz et al. 2006, Gebbie & Huybers 2006).

4.5. Inverse Methods for Estimating Transports

Box inverse methods, which were developed in oceanography beginning around 1975 (e.g., Wunsch 1996, 2006), have become the natural way to combine all observational information with dynamical principles to generate transport estimates, while permitting inferences to be drawn regarding large oceanic volumes. The volume issue enters because for many (but not all) situations, large-volume integrations tend to subdue observational and model noise. The first attempt at such a calculation for the LGM was made by Legrand & Wunsch (1995), who modeled $\delta^{13}C$ as a passive conservative tracer and $\delta^{18}O$ as an active conservative tracer under the assumption that gradients in $\delta^{18}O$ primarily reflect gradients in density. However, they found that their collection of data, when combined with the thermal wind and the volume and tracer conservation equations, was insufficient to distinguish the LGM circulation from that of the modern ocean. Their result has been erroneously quoted as showing that the modern circulation was no different from that of the LGM. To the contrary, they showed only that the then available data did not require a change from the present flows—this distinction is fundamental.

Marchal & Curry (2008) used a more extensive collection of $\delta^{18}O$ and $\delta^{13}C$ data and accounted for the effects of organic matter remineralization on $\delta^{13}C$ when inverting for the circulation in the North Atlantic. Oxidation of organic matter, which has a relatively low $^{13}C/^{12}C$ ratio, tends to reduce the $\delta^{13}C$ of ambient dissolved inorganic carbon and, thus, introduces the necessary time-dependent terms into the representation of $\delta^{13}C$. This carbon flux thus places additional constraints on the rate of flow, although the constraint is weak because of uncertainty in the rate of organic carbon remineralization. Like Legrand & Wunsch (1995), Marchal & Curry (2008) were also unable to show that the LGM circulation differed from the modern. If, however, artificially small uncertainties are assumed for the $\delta^{13}C$ data and the influence of mixing is assumed weak, the data would then require some change to the circulation. Useful constraints on the LGM circulation, although not yet achieved, are thus conceivable.

Gebbie & Huybers (2006) undertook a more limited inverse calculation, using one vertical section across the South Atlantic, in an effort to better understand the implication of the $\delta^{18}O_{calcite}$ data of Lynch-Stieglitz et al. (2006). Again, the overturning rate could not be usefully determined using existing data. Their results suggested, however, that a transect of sediment cores along a single latitude band might accurately determine the rate if multiple forms of measurements were made: $\delta^{18}O_{calcite}$ complemented with temperature estimates from Mg/Ca ratios, as well as pore water estimates to constrain past values of salinity and $\delta^{18}O_{seawater}$.

Formal comparison to the modern circulation is not required. Winguth et al. (1999) attempted a direct estimate of the LGM circulation, using an inverse method to bring an ocean general circulation model into better consistency with LGM $\delta^{13}C$ and Cd/Ca data, accounting for the effects of organic matter production and remineralization. Adjustments to surface salinity were used, and they caused a shoaling and reduction in the flux of NADW. However, it is unknown whether a comparable fit could also be obtained through adjustment of other model parameters—without similar consequences for the flux of NADW. Efforts such as that by Winguth et al. (1999) are legitimate, but the limited data type and volume, and the very restricted choice of adjustable parameters, so far mainly show that the solutions obtained will be highly nonunique.

Using a simple rectangular model domain, Huybers et al. (2007) explored the question of why the LGM meridional circulation is so uncertain relative to that of the modern. There are several reasons. First, much of the uncertainty is associated with paleoclimate measurement inaccuracies. For example, paleodensity estimates are 100 times more uncertain than their modern counterparts. Proxies for wind speed and current velocity, insofar as they exist, are even more uncertain. Second is the dearth of data. Marchal & Curry (2008) employed ~400 data points in their study, a factor of almost two increase over what was available to Legrand & Wunsch (1995), but trifling relative to the billions of data used to constrain the modern problem (e.g., Wunsch & Heimbach 2007). Third, paleoproxies tend to reflect only properties near the surface or the sediment-water interface, making it difficult to constrain the conditions in the interior. (See Lynch-Stieglitz et al. 2007 for a review of LGM data.)

In oceanographic usage, the terminology inverse problem often refers to static situations, but no such restriction is necessary (see Wunsch 2006; Wunsch & Heimbach 2007) and problems in which most elements are time dependent have been addressed. These methods are not discussed here as they have yet to be used in the paleoceanographic context.

5. SETTING RATES OF MOTION

Both geostrophic balance and inversions of equations governing tracer distributions are diagnostic of motion, but they fail to explain how such motion was set up, what determines its overall strength, or how it is maintained against friction. Only deviations from geostrophic balance lead to such determinations. A review of ocean circulation theory here is an impossibility, and some very general remarks must suffice (see Pedlosky 1996, Vallis 2006).

5.1. Winds

Ocean circulation theory begins with the wind field, usually written as $\tau = (\tau^x, \tau^y)$ for the two components of stress (force/unit area) exerted on the ocean and which are functions of space (x, y) and time (t). Historically, determination of the wind field over the ocean was done by compilation of ship reports of wind velocities and only very crude space-time means were found. But with modern satellite measurements (Risien & Chelton 2008) and estimates obtained from meteorological models ("reanalyses"; see Kistler et al. 2001), a great space/time complexity of the wind over the ocean has been documented (**Figure 9**).

By the mid-twentieth century, it had become apparent (see e.g., Pedlosky 1996) that the large-scale movement of water in the ocean, excepting coastal and near-equatorial regions, depended not directly on $\boldsymbol{\tau}$, but on its derivatives in the form

$$\nabla_h \times \tau = \left(\frac{\partial \tau^y}{\partial x}, -\frac{\partial \tau^x}{\partial y} \right). \tag{7}$$

The wind field and its derivatives set the overall magnitude of observed flows so that, for example, volume transports of the western boundary currents are 30 Sv, not 0.3 or 300 Sv. Magnitudes of the corresponding interior flows tell us that spatial variations in time-mean sea level will be of order 1 m and not 0.1 or 10 m. Because the eddy field is generally believed to arise primarily through instability of the "mean" currents, the wind field strength also sets bounds on the magnitudes of the small-scale variability.

Elementary theory also shows that to make rapid changes in the ocean circulation, the wind field is the most efficient mechanism—surprisingly, very high frequency fluctuations (timescales of days), will be felt almost instantaneously at the sea floor, much more effectively than slower changes for which the stratification intervenes and greatly slows the abyssal response. The literature showing that, for example, the North Atlantic Oscillation tends to control most observed

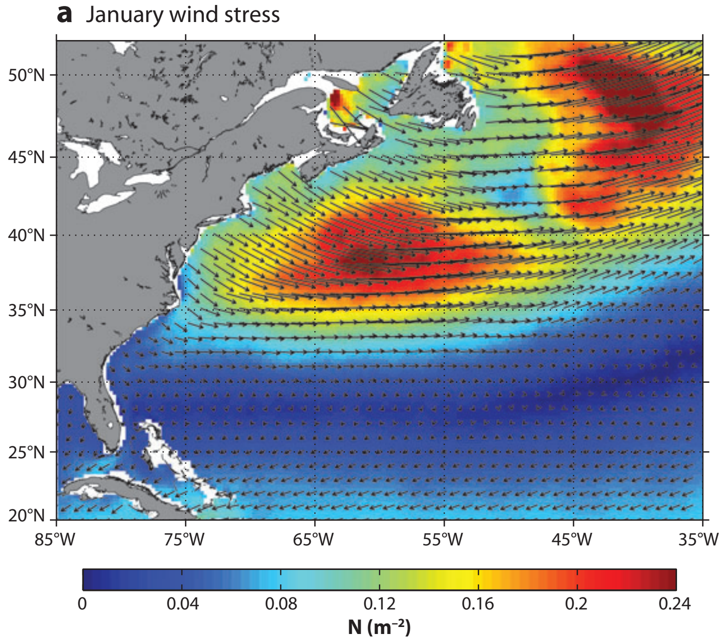


Figure 9

(*Continued*)

modern regional fluctuations, is fully consistent with the notion that the ocean circulation is first and foremost the result of driving by the wind fields.

Ignorance of the paleowind field is one of the greatest obstacles to understanding of past oceanic circulations. Dust- and pollen-concentration variations in ice and sediment cores are sometimes interpreted as proxies for wind strength and direction, but they can also reflect time-variable source areas (e.g., desertification), episodic events, and changing pathways (e.g., Biscaye et al. 1997, Stuut et al. 2002). Conceivably, determination of the wind field will be best made from observations concerning the ocean circulation (e.g., Lynch-Stieglitz 2001)—an interesting inverse problem.

Conflicting reports exist in the modeling literature regarding the LGM wind stress. In the 1990s, the Paleoclimate Model Intercomparison Project 1 forced a collection of models with given LGM boundary conditions and, in particular, the CLIMAP (1981) estimates of sea-surface temperatures and the ICE-4G estimate of ice-sheet topography (Peltier 1994), both of which have subsequently been revised. The model results generally indicated a stormier world, as might be anticipated from the stronger equator-to-pole temperature gradients (Hall et al. 1996, Dong & Valdes 1998, Kageyama et al. 1999, Kageyama & Valdes 2000). But in contrast to the earlier simulations, Li & Battisti (2008), using a model run from the Paleoclimate Model Intercomparison Project 2 (see Otto-Bliesner et al. 2006), found a stronger and steadier Atlantic jet extending into the Northern and Eastern Atlantic but with diminished wintertime atmospheric eddy activity relative to today. Apparently, the difference between these model runs lies not with any fundamental difference in model physics, but with the choice of ice-sheet orography and the sea-surface temperatures either imposed or derived from the model. That such lowest-order changes arise from details of the model configuration calls into question whether an accurate representation of the wind forcing can be obtained as a response to imposed conditions within a model.

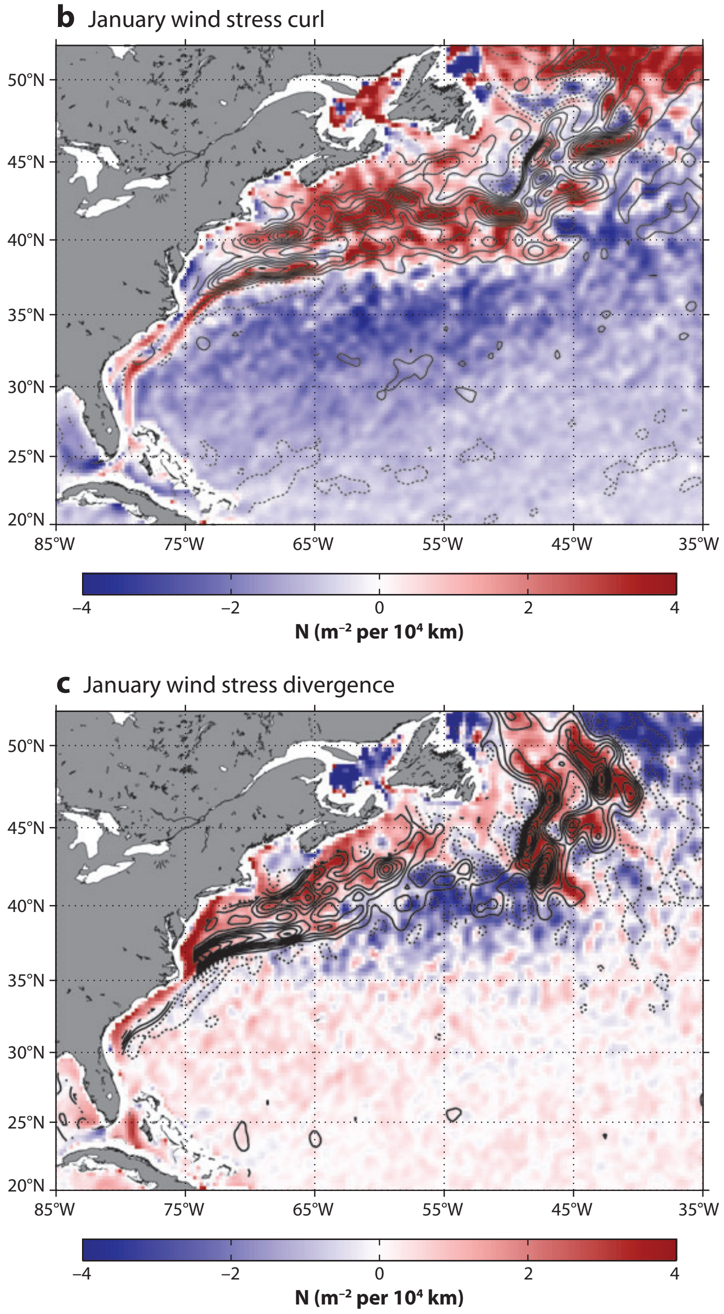


Figure 9

Time average over 8 years of estimated wind stress (*a*) and its curl (*b*) in the western North Atlantic from scatterometer measurements (Risien & Chelton 2008; courtesy C. Risien). The spatial complexity has a direct influence on the resulting ocean circulation. A comparable time dependence exists (not shown). Some of the small-scale "mottling," especially visible in the curl, is probably noise in the measurements.

The modern ocean circulation is believed to be baroclinically and barotropically unstable (Gill et al. 1974, Smith 2007). Thus an increase in the wind strength, driving the large-scale flows harder, would not lead directly to an increase in the circulation strength, but merely a faster transfer of wind energy into the geostrophic eddy field. Something like this behavior is seen in models of the Southern Ocean (e.g., Olbers et al. 2007) where the transports do not respond in any simple way to changes in wind strength.

5.2. Tides and Mixing

Speculations exist about tidal changes over specific periods (Egbert et al. 2004, Wunsch 2005, Arbic et al. 2008, and others). On the timescale of the LGM and the subsequent deglaciation, the major tidal shift would have been the result of a change in sea level (roughly 130 m lower during the LGM). Reduction in modern continental shelf area could lead to an increase in the amplitude of the deep-water tides, and hence their mixing. Egbert et al. (2004) computed the global tidal distribution under the hypothesis of lowered sea level and the ice-sheet-governed ocean topography by Peltier (2004). Inferences about changes in deep mixing are, however, so dependent on the assumptions concerning deep stratification that very little can said about what to expect. Arbic et al. (2008) have suggested that parts of the deglacial history of the North Atlantic could be the result of extremely large tides forming near the ice-sheet edges and destabilizing them. On very long timescales, over which continents move, the tides will likely have to be analyzed in a statistical sense, as movements in and out of resonance appear to be common and of great significance to regional tidal power inputs.

Various speculations exist (Simmons et al. 2004) about how to parameterize tidal motions as mixing physics. It is now thought that mixing takes place intensely in special regions and is generally weak elsewhere. But the parameterizations are not well understood and depend directly on the poorly known deep stratification over topographic features. That mixing rates in the past were the same as today and had the same spatial distributions is very unlikely. Ocean models that do not account for changes in mixing are suspect, but we are not in a position to say what values and distributions should be used (see, e.g., Saenko 2006).

The turbulence, whether wind or tidally powered, is represented in theory and models as a vertical (or, commonly, diapycnal) mixing coefficient, K_z. Bryan (1987), Scott & Marotzke (2002), Nilsson et al. (2003), and others discuss the relationship between values of K_z and the intensity of the oceanic overturning (mass transports) and, in a few cases, the meridional enthalpy transports. A summary, however, would be that there are no known direct relationships that operate on a global scale. For example, the modeling study of Montenegro et al. (2007) did not produce any simple connection between enhanced mixing in confined regions and the overall structure or strength of the flow. The efficacy and influence of mixing depends strongly on the stratification in which it is operating and that, itself, is a consequence of the mixing, among many other influences, and is not locally determined.

The suggestion by Huntley & Zhou (2004) and Dewar et al. (2006) that biological populations have a significant impact on ocean mixing and hence on the circulation is an intriguing, extremely controversial idea (see Visser 2007, Gregg & Horne 2009). If proven important, it enormously complicates the problem of modeling past oceanic states.

5.3. Buoyancy

High-latitude convection and the formation of dense water that sinks to intermediate or great depths is one of the central elements of the ocean climate system. Unfortunately, the existence of this flow, which is not in doubt (e.g., Warren 1981, Saunders 2001), led to the misconception that the ocean was a convective system like the atmosphere, but one with the top and bottom interchanged. The atmosphere is heated from below and cooled above as in the classical Rayleigh-Bénard situation. A fluid such as the ocean, which is heated and cooled at the same level (near the sea surface), has very different physics, which is perhaps best understood by asking the question: If deep water forms at the sea surface and sinks to the sea floor, why does the ocean not simply fill up with that cold-dense water?

Sandström (1908; for an English translation, see Kuhlbrodt 2008) and a number of subsequent authors, notably Jeffreys (1925), analyzed systems in which the heating and cooling were at different levels relative to each other. Sandström (1908) discussed an idealized Carnot cycle (see Defant 1961) and concluded that a system like the ocean would have a very sluggish circulation if this forcing acted in isolation. The result became known as Sandström's theorem,[10] but as with all results for real fluids, it is an approximation, not a mathematical theorem. It has in recent years given rise to a remarkably argumentative and confusing literature (see, e.g., the discussions in Young 2005, Marchal 2007).

Much of the confusion arises because buoyancy exchange with the atmosphere is not the only force acting, and so the ocean circulation does not resemble what one would anticipate from Sandström's approximation, leading some authors to conclude that it is irrelevant. Furthermore, in the laboratory, it is possible to produce measurable, sometimes striking, flows directly proportional to the molecular coefficients of diffusion. (And laboratory Reynolds numbers are many orders of magnitude smaller than oceanic ones.) Sandström's inference, consistent with Jeffreys's interpretation, is not that there is no abyssal flow from surface buoyancy forcing—merely that it will be weak.[11]

How buoyancy forcing acts in concord with other forces is the essence of understanding the circulation. The only simple inference one can make is that, although buoyancy exchange has a major influence on the structure of the circulation and is a major determinant of the transports of heat and other properties, it cannot drive the circulation in the sense of providing the energy required to sustain it.

The deep ocean does have a significant observed flow, not directly dependent on buoyancy forcing. Parts of it are necessarily supported by turbulent mixing (see Wunsch & Ferrari 2004). The intensity and spatial distribution of that turbulence is a determinant of the possible circulation patterns. Turbulence requires energy to support it, and in a paleocean where abyssal turbulence is different from its values today, it can be expected to have a different circulation (see Wunsch & Ferrari 2004, Huang 2004, Ferrari & Wunsch 2009). Only the wind field and the tides appear to be viable as major turbulence energy sources, even though the partitioning of their roles in the modern ocean remains incomplete and controversial and their roles will shift with a changing climate.

6. MISCELLANEOUS ISSUES

In addition to some novel issues, there exist paleocounterparts to all aspects of the modern ocean circulation problem, and it can be safely assumed that determining past ocean circulation will be no simpler than is determination of the modern. Here we briefly touch on some of the more interesting or important issues.

[10]A terminology apparently due to V. Bjerknes, who regarded Sandström's result as a form of his own circulation theorem (see Bjerknes et al. 1933).

[11]The general fluid dynamics of "horizontal convection" remains disturbingly unsettled; see, for example, Siggers et al. (2004).

6.1. Sea-Level Change

Fluctuations in regional sea level over years and decades are seen, having magnitudes of up to 10 mm/y (see Cazenave & Nerem 2004, Wunsch et al. 2007, and many others). Changes in sea level at any location of order 10 cm or less can arise either from local shifts in the general circulation or from global changes. If the temperature of the ocean should change on average by the large value of 10°C, the shift in the sea-surface height would be approximately 1.5 m. Anything larger than this value, absent very large changes in the forcing fields, is due to changes in freshwater content derived from continental ice, as depicted in the various sea-level reconstructions through the last deglaciation (e.g., Bard et al. 1989, Peltier & Fairbanks 2006). The volume of ocean basins can also change on very long timescales, but the idea that this resulted from changes in sea-floor spreading rates is now out of favor (see Rowley 2002).

The implication of changes in sea level for physical oceanography involves primarily the tides (as discussed above), shifts in interbasin connections (Bering Straits, Drake Passage, Isthmus of Panama, etc.), opening and closing water-mass mixing pathways, and topographic changes in shallow waters. Glacial volume changes imply concomitant salinity shifts and the injection or removal of fluid labeled by isotopic ratio anomalies (e.g., $\delta^{18}O$). Dynamical (fluid flow) considerations also arise with respect to freshwater injection at local scales during melting phases and an excess of evaporation during glacial buildup.

Melting ice sheets will also change the gravity field and, thereby, influence regional sea level. For example, removal of the West Antarctic ice sheet would lead to a sea-level reduction in the immediate vicinity and an increase along the U.S. coast 30% in excess of the eustatic rise (Mitrovica et al. 2009). Both sea and land ice processes enter strongly into the discussion, but space dictates that we leave the subject here.

6.2. Time Uncertainty

But [when were] the snows of yesteryear?

(modified from François Villon)

Physical oceanographers studying the modern ocean rarely cope with a data series whose timing is uncertain. In most of paleoceanography, however, the independent variable of measurement is depth in cores rather than time. Determining the absolute or relative timing of events is crucial if the sequencing and, one hopes, the causes of climate changes are to be inferred from the proxy record.

Radiocarbon dates are available over approximately the past 40,000 years, but they are subject to the tracer uncertainties discussed above. Errors on the order of millennia are easily possible during the last deglaciation (e.g., Waelbroeck et al. 2001), and during parts of the last deglaciation, the relationship between radiocarbon years and calendar years is multivalued (e.g., Hughen et al. 2004). Imperfect uranium series date estimates are also attainable from corals (e.g., Gallup et al. 2002, Thompson & Goldstein 2005) or even directly from sediments (Henderson & Slowey 2000).

Other time markers are available, usually through correlation with major events identifiable in cores and that have been dated in nonmarine material. Examples include matching abrupt changes observed in sediment properties to changes dated in speleothems (e.g., stalagmites or stalactites) or ice-core records (e.g. Marchitto et al. 2007), inferring the timing of glacial terminations from the dating of coral terraces (e.g., Thompson & Goldstein 2005), or the Brunhes-Matuyama and other geomagnetic reversals from the dating of volcanic flows (e.g., Raymo 1997). Ages can be interpolated between absolute estimates using sediment accumulation models, and the error growth arising from unsteady accumulation rates can be controlled by averaging across many such estimates (e.g., Raymo 1997, Huybers & Wunsch 2004).

Another common approach is so-called orbital tuning, which depends on the assumption that variations in proxies can be matched, commonly at zero-phase lag, against variations in Earth's insolation. The method has had notable successes, such as predicting an older date for the Brunhes-Matuyama magnetic reversal (Johnson 1982, Shackleton et al. 1990), though the possibility of circular reasoning is omnipresent. For example, eccentricity-like amplitude modulation found in the precession band of orbitally tuned records has been cited as evidence for the accuracy of orbitally tuned ages (e.g., Shackleton et al. 1990, 1995), but it can instead result from standard tuning methods independent of any true orbital signal (e.g., Neeman 1993, Huybers & Wunsch 2004).

If absolute ages cannot be determined in records, it is often still useful to constrain their relative timing by aligning events that are believed to be contemporaneous. Benthic $\delta^{18}O$ is commonly used for the synchronization of marine sediment cores (e.g., Lisiecki & Raymo 2005), but even where event identification is truly unambiguous, multimillennial errors can be introduced by the long equilibration times in the ocean (e.g., Wunsch & Heimbach 2008) and the likelihood of nonuniform changes in temperature and $\delta^{18}O_{seawater}$ across various oceanic regions (e.g., Skinner & Shackleton 2005). Another technique is to align the global component of geomagnetic field intensity variations that are preserved in marine sediment cores (Stoner et al. 2002); although even under best-case circumstances, inaccuracies of several thousand years result (McMillan & Constable 2006).

Given the irreducible uncertainty in the timing of paleoclimate records, there is an urgent need for statistical methods that can be applied to time-uncertain series of data. The problem has only begun to be explored—for example, with respect to estimating spectra in the presence of time uncertainty (Thomson & Robinson 1996, Mudelsee et al. 2009), calibrating age using uncertain radiocarbon dates (Buck & Millard 2004), and testing for covariance between time-uncertain records (Haam & Huybers 2009). Time uncertainty merits serious attention from the signal processing and statistical communities.

6.3. The North Atlantic Obsession

Like the Genesis story, the idea that the North Atlantic Ocean meridional overturning circulation is the major controller of the climate system has taken on an almost mythic status. It supposes that large freshwater discharges into the North Atlantic uniformly greatly weakened the circulation and gave rise to a major climate shift, at least hemispherically and often globally. It is thus worth attempting to distinguish evidence for ocean changes (which are inevitable) from evidence that the North Atlantic Ocean is the controlling or dominant factor in the response.[12]

Ample evidence shows that the water-mass structures of the North Atlantic Ocean were different during the LGM—unsurprising given the distinct atmospheric and biological conditions at the surface. That major changes occurred in the wind and buoyancy exchange fields is guaranteed, but little is known of what they were. For the reasons already described, no concrete knowledge exists determining overall rates of North Atlantic flow during the LGM or any other time prior to the modern period. They were assuredly different in many aspects, but we do not know what

[12]That much paleoclimate interest initially focused on the North Atlantic is readily explained: It is comparatively small and surrounded by North American and Western European oceanographic institutions, has a relatively high sedimentation rate and better preservation, and has the highest modern data density. Whether it truly dominates the climate system is less obvious.

those differences were. Some further comments about the nature of North Atlantic behavior can be found in Wunsch (2007), who suggested that its area is too small to dominate the global climate system.

A number of authors (e.g., Toggweiler & Samuels 1995, Toggweiler & Russell 2008) have concluded that the wind field of the Southern Ocean is important to the meridional overturning rates in the North Atlantic (the so-called Drake Passage effect). Such results are plausible and emphasize that, in a fluid, effects can appear at large separations in distance and time from their proximate cause. Absent wind-field estimates, it will be difficult to produce quantitative theories either of the large-scale ocean circulation or of the small-scale mixing that helps determine it.

With the recent recognition (e.g., Brauer et al. 2008, Steffensen et al. 2008) that some elements of the climate system can shift far faster than the large-scale ocean circulation—best regarded as basically a fly wheel—perhaps the notion of North Atlantic meridional overturning circulation as the control, rather than as the response and feedback, will finally be challenged. The most volatile elements of the climate system are the wind field—major changes can occur in hours—and sea ice (e.g., Stroeve et al. 2007), which has a huge seasonal range. They are the most likely explanations of abrupt climate change. One needs mechanisms capable of providing both rapid change and stability over long periods in the new state. The ocean provides stability, perhaps as part of a response/feedback mechanism; by itself, it is unlikely to produce the rapid transitions thought to occur.[13]

7. GENERAL CIRCULATION MODELS AND ISSUES ARISING

The sciences do not try to explain, they hardly even try to interpret, they mainly make models.
(von Neumann 1955, p. 492)

General circulation models of the ocean, as subcomponents of more general models of the climate system, are essential tools that have proved highly useful in depicting and understanding ocean circulation. These models are seductive—time is sped up enormously, unobserved phenomena can apparently be computed, calculations can be done comfortably at home, etc.—but they are immensely complicated machines, involving hundreds of thousands of lines of computer code assembled by numerous individuals over the decades since around 1955. As with all powerful tools, considerable skill is required to use them successfully.

Notably, they are called "models" and not "reality" because they are necessarily imperfect approximations to the ocean. The probability that such enormous collective assemblages are free of coding errors approaches zero. Errors make the code differ from that intended by the programmer: Basili et al. (1992) reported rates of approximately one error for every 2000 lines of FORTRAN code in operational programs used for flight dynamics. This rate represents a decline from four per 2000 lines in the testing phase of the models. But there are many other sources of error, including the numerical approximations to the Navier-Stokes equations and inaccurate parameterization of subgrid scale processes (commonly mixing, internal waves, etc.), error-prone initial and boundary conditions. No time-stepping technique, except when applied to the most trivial sort of problem, can be expected to run forward in time without gradually accumulating certain forms of error. In simple systems, systematic errors tend to grow linearly with time, t, and stochastic errors as $t^{1/2}$ (a

[13] The meaning of rapid is a relative one, and in paleoceanographic studies, its use ranges from those characterizing Dansgaard-Oeschger events (changes of order 10 years and less) to anything changing more swiftly than geological timescales of millions of years.

random walk). That is, all nontrivial time-stepping models can be expected to accumulate errors as the time horizon of integration grows. Unhappily, this phenomenon is rarely remarked upon. Note that even steady-state solutions are usually obtained by time stepping through transients.

Consider, as an analogue, the problem of launching a spacecraft from Earth to land on Mars. Although orbital dynamics are far simpler than those governing climate, and trajectory computations have been used for 400 years, no engineer would expect to hit a landing spot without an entire series of on-course corrections. Those corrections would account for errors in the launch angles and velocities, simplifications in the bodies of the solar system (e.g., Venus and the Sun treated as spherical, omission of general relativity), random fluctuations in the solar wind, imperfect representation of the controls, truncated numerics, etc.

It is sometimes asserted that, because ocean or climate models contain feedbacks and constraints, errors do not grow without bound. This statement is undoubtedly true, but the degree to which the accumulation of error destroys or distorts the representation of certain quantities must be determined on a case by case basis. For example, the Gulf Stream transport in a model may have a near-constant error, whereas the movement of a passive "dye" injected into a model could diverge from reality with time, at least until approaching a relatively uninteresting final state of uniform dispersal.

That a great variety of errors occur in ocean models can hardly be doubted. Consider **Figures 10** and **11** (from Hecht et al. 1995), who introduced a dye patch into Stommel's (1948) extremely simple oceanic circulation model. **Figure 11** shows the analytical solution (computed by quadrature) for the position and shape of the patch after 1.5×10^8 s (approximately 5 years). The remaining panels display the position and shape from eight conventional upwind advection-diffusion schemes used in oceanic models. Even the best of them has a distorted dye patch, and the worst have dye in physically impossible places. That these errors arise after such short periods and with such a simple (steady, linear, flat bottom, etc.) model is at least strongly suggestive that model error needs to be estimated before solutions run for long times can be taken seriously (What happens to the distribution of carbon in such models?). An ocean model useful for making

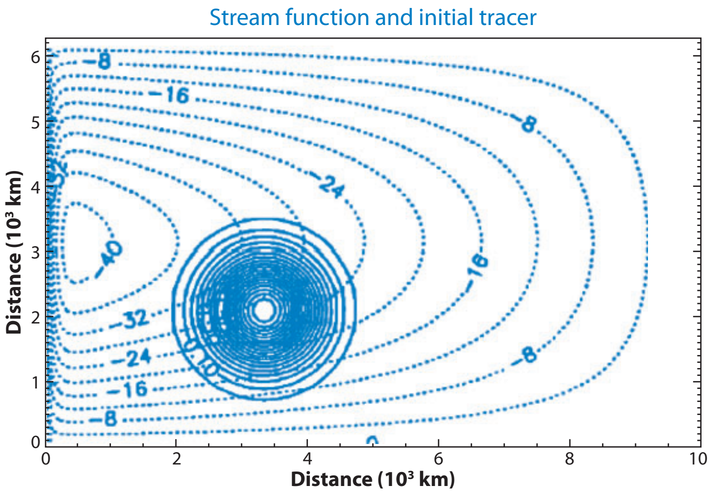
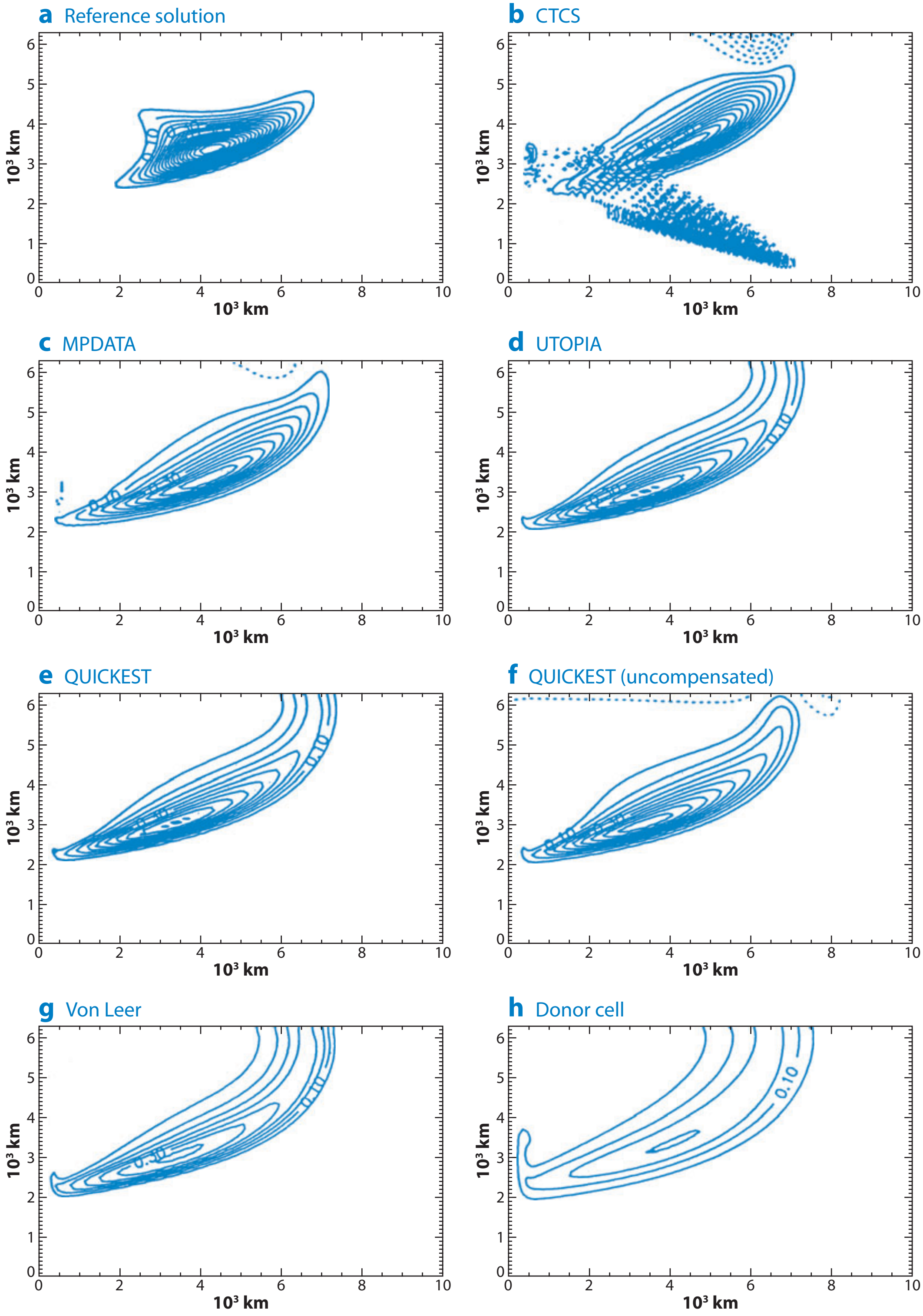


Figure 10

Initial conditions for a dye-patch calculation discussed by Hecht et al. (1995). The patch is embedded in a Stommel (1948) steady, linear dynamics gyre, and then its trajectory is followed in time. See **Figure 11** for its exact structure and position after five years.


a Reference solution
b CTCS
c MPDATA
d UTOPIA
e QUICKEST
f QUICKEST (uncompensated)
g Von Leer
h Donor cell
0.10
10^3 km
10^3 km
0
1
2
3
4
5
6
0
2
4
6
8
10

a 10-year prediction may be useless for a 1000-year one. Quantitative understanding of error growth in such models is essential.

These remarks must not be interpreted as implying that ocean models should not be used. To the contrary, they are vital tools in depicting and understanding the climate system. But they are only tools, ones that neither explain nor represent the complete physical system. We need a quantitative theory of models—one that would parallel the uncertainty statements routinely provided for data—that will permit models to evolve beyond their role as a novelization of the climate system.

Because so little is known of atmospheric conditions in the distant past, most emphasis in oceanic modeling has been in the context of coupled systems, in which the atmospheric state, and hence exchange with the ocean, is computed as a consequence. Of the suite of nine models subjected to LGM forcing and boundary conditions under the auspices of the Paleoclimate Modeling Intercomparison Project 2, four had an increase in overturning, four a decrease, and one had essentially no change (Weber et al. 2007). Even a random selection of nine numbers would generally give the (false) impression of suggesting at least some predictive power. Weber et al. (2007) concluded that "[b]ased on these results, it seems inconclusive whether existing climate models have the accuracy to simulate AMOC [Atlantic meridional overturning circulation] changes in response to future increases in greenhouse gas levels."

Regarding future climate change, there does exist an expectation reflected in the literature that the overturning circulation will weaken, but here too, there is widespread agreement that processes critical for the representation of the overturning are inadequately represented in the models. Thus, perversely, it could be considered fortunate that observational determination of the overturning circulation during the LGM has been forestalled, permitting a more objective model test when and if such observations become available.

8. CONCLUDING REMARKS

The past isn't dead. It isn't even past.

(William Faulkner, "Requiem for a Nun," Act 1, 1951)

Because marginally adequate, near-global observational systems of the ocean begin only in the 1990s, one is driven to understand the record of past ocean states as embedded in the paleoclimate proxies. The gist of this review is that some interpretations of those data are robust while others are extremely fragile, rest on a series of vaguely plausible assumptions, and confuse fact with mere familiarity. Distinguishing these extremes is the heart of the paleoceanographic problem.

Three reasons exist for an oceanic focus as provided here. First, the ocean is a major component of the climate system. Second, the important role of deep-sea cores in documenting the past climate of the entire Earth system means that the medium through which those records accumulated must be understood. Third, the modern ocean state cannot be understood without reference to the geologic past as the equilibration time of some parts of the ocean interior extend out to many millennia (Wunsch & Heimbach 2008). Enough has been learned to show that without a doubt the system shifts on all timescales.

Figure 11

(*a*) The reference solution shows the exact solution for the dye-patch concentration after approximately five years found by Hecht et al. (1995). Remaining panels (*b–h*) show the concentrations calculated by seven conventional numerical methods. All have errors, some very large. Note how short the integration time and how simple the flow field are. See also Hecht et al. (1998).

No guarantee exists that the goal of determining past climate can be fulfilled. Present understanding of proxies, and the restricted regions that can ultimately be drilled and sampled on the sea floor, strongly suggest that determination of the past state will remain a hugely underconstrained problem, unless and until there is a technical or intellectual breakthrough. [Perhaps ancient or modern DNA will permit the decoding of past climates, e.g., Waller et al. (2007).] Breakthroughs and their influence on the field cannot be predicted, and one can best proceed under the assumption that great uncertainties will linger. In such a situation, where the data are sparse and ambiguous, it is essential to avoid focusing on one plausible story to the exclusion of all others. One should deliberately try to construct a range of solutions that exhibit the different possibilities consistent with the data and models. Recall Chamberlin's (1890) insistence on the need for retaining multiple working hypotheses.[14] For example, among the many hypotheses would be those that make the ocean circulation the "trigger" of climate change, with such triggers lying largely in the North Atlantic, as opposed to possibilities that the ocean responds primarily to disturbances from the coupled atmosphere and ice distributions and that cause and effect are largely global.

DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

ACKNOWLEDGMENTS

The authors are supported in part by National Science Foundation grant OCE-0645936. We are grateful for the particularly helpful comments, suggestions, and corrections of O. Marchal, G. Gebbie, J. Sachs, J. Lynch-Stieglitz, and E. Bard. Remaining errors and unhappy omissions are all ours, however. Many interesting topics have been painfully omitted for lack of space.

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[14]In another paper, which was notable for being ahead of its time, Chamberlin (1906) discussed the potential role of the ocean in the glacial cycles, warm Pliocene climates, and scenarios similar to snowball earth, with particular attention to the "abysmal" circulation. He concluded by noting that the ocean could not be the fundamental cause of major climate shifts. (We are indebted to E. Bard for the reference.)

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Annual Review of Marine Science

Volume 2, 2010

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Errata

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