Paleo-Physical Oceanography with an Emphasis on Transport Rates
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PETER HUYBERS

Harvard University. email phuybers@fas.harvard.edu

CARL WUNSCH

Massachusetts Institute of Technology. email: cwunsch@mit.edu

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Abstract Paleo-physical 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 itself 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 is heavily dependent upon 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 (LGM) 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 time-scale (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 paleocirculation in any period are yet known with certainty.

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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: (1) today; (2) the past; and (3) the future. Society has a strong interest in (3), but understanding (1) and (2) are prerequisites. None of these divisions are distinct, as the ocean integrates over past events extending to thousands of years. Without adequate data, a scientist seeking to understand multi-decadal, 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 might define the domain of paleoceanography as the period prior to wide-spread instrumental observations — perhaps dating to about 1990. That late date might 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 upon 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 of the elements of modern physical oceanography (coastal dynamics, internal waves, convection, geostrophic eddies, mixed layer physics, carbon uptake, etc.) are relevant to 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 will be 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) 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, and 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 or salinity or water speed) than would be an instrumental measurement, and 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 $\delta^{18}O$ 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, salinity, or density, depending on the context and the other proxies it is paired with. 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, albeit novel elements such as isotopic fractionation effects must be included.

One way to organize a discussion of the problems of paleoceanography is to first summarize what we know of the modern circulation and how and why we think we know it. No

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

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) provide 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, fresh water, enthalpy, etc., and it is these transports which ultimately control the distributions represented in the proxies. If one is to both 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 the large modern data base, major uncertainties remain in the description and understanding of the modern ocean circulation, and the comparatively slight data base depicting the ocean of the past is a huge problem. Fig. 1 displays one of the more comprehensive core collections used for an Atlantic study in the LGM, but which is extremely limited relative to the still under-sampled 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 time scales less than about a 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 primarily grows during one season does not record an annual average value, and such seasonally sampled records can easily be misinterpreted (e.g. Huybers and 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

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[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.

simple average (e.g. McCave and 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 Fig. 2 (Shackleton et al. 1990) is an example. Analyzing data from cores confronts three major problems: (1) Measurable proxies are sometimes poorly known functions of the actual physical variables sought. (2) Records are only available from regions of significant sediment accumulation on the sea floor (see Fig. 3 for a map of thickness). Vast sections of the ocean floor are undersaturated with respect to calcium carbonate, e.g., the deep Pacific, inhibiting sediment accumulation and generally leading to significant alteration of the record which does accumulate. Further, a limit of about -90 My (e.g. Rowley 2002) exists when virtually all sediment is irretrievably lost to plate subduction. (3) 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 1987), and downslope reworking of sediments.

3   Water Masses: The Static Problem

The most prominent role of the ocean in climate is as a reservoir — containing most of the fresh water on Earth, far more carbon than exists in the atmosphere, and a very 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 fall into a few categories: (1) passive (no impact on density), (2) active (influence density), (3) conservative, (4) non-conservative, or in combinations such as passive and nonconservative. For example, the Mg/Ca ratio in foraminiferal calcite is thus a proxy for water temperature and is a conservative and active tracer, while the radiocarbon concentration in a foraminiferal shell can be a proxy for past water radiocarbon concentrations (under a series of assumptions), and is then a non-conservative and passive tracer.

Distributions of tracers in the ocean, and hence their interpretation, depend upon the tracer category, the boundary conditions, the rate of decay (if any), initial conditions, the flow field, and mixing rates. Even the so-called forward calculation (given all those pieces of information), is difficult. The inverse problem of, instead, inferring the flow field and/or mixing rates from observed tracer distributions, remains a formidable one. We begin with possibly the simplest problem — defining and inventorying the various water masses present in the ocean.

Beginning in the middle 19th Century, mapping of various properties (mainly temper-

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

ature 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. Fig. 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 indicates that a very large number of water types fills the ocean's interior, with the first ten only accounting for 40% of the ocean's volume, and the first hundred only 70% (see Fig. 5). 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 on 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 δ^{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 δ^{13}C and Cd/Ca ratios of benthic shells obtained from sediments have been shown to reflect, respectively, the δ^{13}C of dissolved inorganic carbon and the Cd content of bottom waters (Duplessy 1984, Boyle 1988, Boyle 1992). δ^{13}C and Cd/Ca concentration ratios co-vary 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 1984), Curry and Oppo (2005) have compiled foraminiferal samples of δ^{13}C along margins, seamounts, and the ocean bottom for the Last Glacial Maximum, forming a pseudo-transect in the western North Atlantic. The prevailing interpretation of the LGM δ^{13}C distribution is that nutrient-poor NADW shoaled and that a nutrient-rich Antarctic Bottom Water (AABW) filled a greater proportion of the deep Atlantic. Estimates made from mapping Cd/Ca ratios are interpreted to support this view (Marchitto and Broecker 2006). Alternative hypotheses include that the nutrient content of NADW was different during the LGM or that NADW instead filled the depths of the northeastern Atlantic. . Moreover, Figs. 4 and 5 show that quantitative inference should not be drawn from two end-member calculations. The decomposition of the ocean using only one or two tracers becomes non-unique once it is recognized that many water types are present (also see Wunsch 2007, p. 58-59).

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

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[4]δ^{13}C$= 1000 \left[R_{\text{sample}}/R_{\text{standard}} - 1 \right]$, where $R_j = \left[^{13}\text{C}\right] / \left[^{12}\text{C}\right]$.

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 130m 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, North Atlantic deep water is the more saline, but they also conclude that the Southern Ocean was saltier during the LGM, approaching values of 37. While the measurements are difficult, 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, that the volume of North Atlantic Deep Water in the Atlantic decreased and that bottom density increased, the cause of the redistribution of these 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 to either their rates of production or removal separately, 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 just when the water volume is greatest). Paleoceanography addresses in several ways the problem of inferring water movement rates, but two of them dominate: (1) the use of tracers depending in some known way on time; (2) 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 about 5700 years according to the rule,

$$dC/dt = -\lambda C. \tag{1}$$ {radiocarb2}

C indicates a generic tracer concentration. The solution is,

$$C = C_0 \exp\left(-\lambda t\right), \tag{2}$$ {radiocarb3}

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,

$$t_C = -\frac{1}{\lambda} \ln\left(\frac{C}{C_0}\right), \qquad (3) \quad \{\texttt{radiocarb1}\}$$

the so-called *radiocarbon age*. Eq. (3) is the basis of radiocarbon dating of a huge variety of samples in many fields from anthropology to climate.

Eq. (3) has been applied to modern seawater, yielding radiocarbon ages ranging from a few hundred years in the North Atlantic to about 1500 years in the North Pacific at depth (see Fig. 2.4.3 of Sarmiento and Gruber 2004, Matsumoto et al. 2007).

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

$$\{\texttt{tracer1}\} \qquad \frac{\partial C}{\partial t} + \mathbf{v} \cdot \nabla C - \nabla\left(\mathbf{K}\nabla C\right) = -\lambda C + q, \qquad (4)$$

for the generic tracer, C. $\mathbf{v}$ is the three-dimensional, usually time-varying, flow field, $\mathbf{K}$ is a mixing tensor (e.g., Vallis 2006), 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). Eq. (4) can be solved by standard numerical methods if initial conditions, $C\left(\mathbf{r}, t=0\right) = C_I\left(\mathbf{r}\right)$, boundary conditions, $C_b\left(\mathbf{r}_b, t\right) = C\left(\mathbf{r}=\mathbf{r}_b, t\right)$, where $\mathbf{r}_b$ is the boundary surface — most typically the sea surface, and the flow field, $\mathbf{v}$, the mixing tensor, and source/sink, q, are all specified. If the assumption of a steady-state is justified (requiring that all terms, including the flow field, should 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, $C_\infty = C\left(C_b\left(\mathbf{r}_b\right), \mathbf{v}, \mathbf{K}, \lambda, q, \mathbf{r}\right)$, and will be a solution to Eq. (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 non-linear transformation of the tracer concentration (Eq. 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 at some 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 routes which are admissible. 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[5] (McIntosh and Veronis 1993; Wunsch 1996, 2006). But because of the forbidding data distribution problems involved with defining three dimensional tracer gradients this calculation has never been directly attempted, even in the modern ocean, except in small regions, in part because the data coverage (Fig. 6, Key et al. 2004) used

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[5]It is linear only for passive tracers not influencing the flow field or mixing tensor and when there are no uncertainties in C.

to determine the ^{14}C distribution is still limited. The generic "tracer age problem" is discussed extensively by Jenkins (1980), Wunsch (2002), Waugh et al. (2003) and others.

Determination of the radiocarbon concentration that seawater had at some point in the past is even more involved than is 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 and Boyle 1997)[6]. Bioturbation can seriously complicate paired radiocarbon estimates but this effect can be limited by sampling only at abundance peaks (e.g. Keigwin, 2004), wherein older or younger samples are less likely to have mixed. 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 atmospheric radiocarbon concentrations are time-variable, and during the last deglaciation appear to have decreased rapidly compared to the decay rate (e.g. Broecker 2007)), making it necessary to account for the influence of time-variable atmospheric radiocarbon upon ocean values.

Another tracer potentially informative of the rates of flow 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, there would be a fixed, relative burial rate between ^{231}Pa and ^{230}Th of 0.093. But ^{231}Pa does not adhere to particles as well as ^{230}Th, giving it a longer oceanic residence time prior to burial ($\sim$100-200 years as opposed to $\sim$30 years) and 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 (Marchal 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 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.

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[6]It is also necessary to correct for fractionation effects. Biological processes take up ^{12}C preferentially to ^{14}C, but assuming this selection is mass-dependent, it can be estimated and corrected for though using measurements of ^{13}C, which is stable.

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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 and Hall 2006). The collection of silt data from the LGM is perhaps most readily interpreted as indicating spatially heterogeneous flow changes relative to the late Holocene. On the basis of six silt records, McCave et al. (1995) inferred that a weaker flow existed in the deep North Atlantic (>2000m) relative to that of the late Holocene, while a more rapid circulation existed a mid-depths, (1000-2000m). 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 (4600m depth, Kuhn and Diekmann, 2002), while increased flow was found off of the East coast of New Zealand (3300m 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). 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 form,

{geost1}
$$f\rho v \quad = -g \int_{z_0}^{z} \frac{\partial \rho}{\partial x} dz + b\left(z_0, x, y\right), \tag{5}$$

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

where x, y, z are three Cartesian coordinates, f the Coriolis parameter, g is gravity, and u, v the two horizontal velocity components. ρ is inferred from measurements of temperature, salinity, and pressure through an empirical equation of state. u, v estimated in this form are called the "thermal wind". Except very close to the equator, and in very strong boundary currents, large scale flows in the ocean satisfy this balance to high accuracy[7].

  Level-of-No-Motion

  The main issue in the use of Eqs. (5) was, historically, the observationally intractable problem of determining b, (or c) and 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

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[7]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.

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 over 100 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, such as that of Ganachaud and Wunsch (2002) and 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 1000m to the bottom would represent a volume transport error of about 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 rates of meridional overturning (MOC, roughly 15 Sv) in the North Atlantic. These estimates usually differ by 2-3 Sv, and are not distinguishable given even the modern data base[8].

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 upon the large-scale structures contain 90–99% of the kinetic energy of the flow (see Ferrari and 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 — producing major sampling problems (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 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 (1) astonishing, and (2) necessarily representative of longer-term trends, neither of which is obviously true.

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

4.4 Paleoceanographic Geostrophy

Efforts have been made to employ geostrophy during the LGM (Legrand and Wunsch 1995, Ortiz et al. 1997, Lynch-Stieglitz et al. 1999, Lynch-Stiegltiz 2001, Lynch-Stiegltiz 2006) and these all depend foremost upon the success with which paleodensity can be reconstructed. The ability to constrain past density comes from the fortuitous, albeit imperfect, happenstance that the oxygen isotopic ratios in calcite shells ($\delta^{18}O_{\text{calcite}}$), 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, $\delta^{18}O_{\text{calcite}} \sim aT + b + cS + d$, 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 and O'Neil, 1997), consistent with the fractionation expected under thermodynamic equilibrium, but that b varies depending on which foraminiferal species is analyzed. Co-variation between $\delta^{18}O_{\text{calcite}}$ 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 and 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 $\delta^{18}O_{\text{calcite}}$ exists because their dependencies upon T and S differ (Gebbie and 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 $\delta^{18}O_{\text{calcite}}$ and $\delta^{18}O_{\text{seawater}}$.) There is also the question of whether the proxy data properly average flows such as seen in Fig. 8.

Lynch-Stieglitz et al. (1999) used sequences of sediment cores taken across the sloping margins on either side of the Florida Straits to reconstruct horizontal gradients in $\delta^{18}O_{\text{calcite}}$ during the LGM, and assuming a constant relationship with density, 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: there is no physical relationship between shear and absolute velocity[9]. Lynch-Stieglitz et al. (1999) assumed that a level-of-no-motion existed at the sea floor and concluded that LGM transports were reduced to between 14–21 Sv relative to Holocene values of about 30 Sv. Taking a cross-section of 100 km times a mean depth of about 500 m, the approximately 10 Sv difference could be accounted for if there had been a bottom velocity of about 20cm/s during the LGM.

Was the bottom flow in the Florida Straits greater during the LGM than today? Fig. 7

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[9]Needler's (1985) formula produces a relationship for the absolute velocity in terms of the density, but 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.

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 about 10cm/s everywhere[10]. We do not know whether bottom velocities were larger in the past[11], but there appears no theoretical or observational reason to rule such a change out, and given a lowering of sea level by 120 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 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.[12] The ability to determine heat transport requires knowing not only the mass transport, but also the change in heat content integrated along the flow path — a much harder problem. 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.

An attempt was also made by Lynch-Stieglitz et al. (2006) to constrain the vertical shear at depths between 200 and 2000 meters 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 $\delta^{18}O_{\text{calcite}}$ and density is poorly constrained (Lynch-Stieglitz et al. 2006, Gebbie and Huybers, 2006).

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[10]The Florida Straits were the site of G. Wüst's important demonstration of the utility of geostrophic balance. Warren (2006) points 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 indeed, but with some crucial exceptions

[11]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.

[12]Use of geostrophy in this region is not straightforward. The two profiles used to compute the shear do not lie transverse to the Stream, but 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.

4.5   Inverse Methods for Estimating Transports

Box inverse methods, which were developed in oceanography beginning about 1975 (e.g., Wunsch 1996, 2006), have become the natural way to combine *all* observational information with dynamical principles in order to make transport estimates, while permitting inferences to be drawn regarding large oceanic volumes. The volume issue enters because for many (not all) situations, large-volume integrations tend to subdue observational and model noise. The first attempt at such a calculation for the LGM was apparently made by Legrand and Wunsch (1995) who modeled δ^{13}C as a passive conservative tracer and δ^{18}O as an active conservative tracer under the assumption that gradients in δ^{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.[13]

Marchal and Curry (2008) used a more extensive collection of δ^{18}O and δ^{13}C data and accounted for the effects of organic matter remineralization on δ^{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 δ^{13}C of ambient dissolved inorganic carbon and, thus, introduces the necessary time-dependent terms into the representation of δ^{13}C. This carbon flux places additional constraints on the rate of flow, but whose accuracy depends on poorly constrained rates of remineralization. Like Legrand and Wunsch (1995), Marchal and Curry (2008) were also *unable* to show that the the LGM circulation differed from the modern. If, however, artificially small uncertainties are assumed for the δ^{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, while not yet achieved, are thus conceivable.

Gebbie and 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 δ^{18}O$_{\text{calcite}}$ 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: δ^{18}O$_{\text{calcite}}$ complemented with temperature estimates from Mg/Ca ratios, as well as porewater estimates to constrain past values of salinity and δ^{18}O$_{\text{seawater}}$.

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 δ^{13}C and Cd/Ca data,

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[13]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. The distinction is fundamental.

accounting for the effects of organic matter production and remineralization; adjustments to surface salinity were used, and 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. Such efforts 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 non-unique.

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. 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, almost a factor of two increase over what was available to Legrand and Wunsch in 1995, but trifling relative to the billions of data used to constrain the modern problem (e.g. Wunsch and Heimbach, 2007). Third, paleoproxies tend to only reflect 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 and 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 fail to explain how such motion was set up, what determines its overall strength, nor 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, or Vallis 2006).

5.1   Winds

Ocean circulation theory begins with the wind field, usually written as $\tau = (\tau^{\mathbf{x}}, \tau^{\mathbf{y}})$ 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 and Chelton, 2008) and estimates obtained from meteorological models ("reanalyses," see Kistler et al. 2001), a very great space/time complexity of the wind over the ocean has been documented (Fig. 9).

By the middle of the 20th century, it had become apparent (see e.g. Pedlosky 1996) that the large-scale movement of water in the ocean, excepting near-equatorial regions, depended not directly upon τ, but upon its derivatives in the form,

{tau2}
$$\nabla_h \times \tau = \left(\frac{\partial \tau^{\mathbf{y}}}{\partial \mathbf{x}}, -\frac{\partial \tau^{\mathbf{x}}}{\partial \mathbf{y}} \right). \tag{7}$$

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 (time scales 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 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 paleo-wind 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 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 SSTs and ICE-4G reconstruction (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 and Valdes 1998, Kageyama et al. 1999, Kageyama and Valdes 2000). But in contrast to the earlier simulations, Li and Battisti (2008), using a model run from the Paleoclimate Model In-

tercomparison Project 2 (CCSM3, 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 first-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.

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 time scale of the LGM and the subsequent deglaciation, the major tidal shift would have been the result of the 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 Peltier (2004) ice-sheet governed ocean topography. Inferences about changes in deep mixing are, however, so dependent upon 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 time-scales, 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 are directly dependent upon 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 which 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 and 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 no direct relationships are known 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 and 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). 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 which 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 like the ocean, which is heated and cooled at the same level (near the sea surface), has a 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 doesn't the ocean simply fill up with that cold-dense water?

Sandström (1908; see Kuhlbrodt 2008 for a translation) 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 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 unfortunately became known as "Sandström's Theorem,"[14] but like 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 discussion 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

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[14]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).

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. 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[15].

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 upon buoyancy forcing. Parts of it are necessarily supported by turbulent mixing (see Wunsch and 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, can be expected to have a different circulation (see Wunsch and Ferrari 2004, Huang 2004, Ferrari and Wunsch 2009). Only the wind field and the tides appear to be viable as major turbulence energy sources, albeit 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 paleo counterparts to all aspects of the modern ocean circulation problem, and it can generally be safely assumed that determining past ocean circulation will be no simpler than determination of the modern. Here we briefly touch on some of the more interesting or important issues.

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 and 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 truly 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 about 1.5m. Anything larger than this value, absent very large changes in the forcing fields, is due to changes in fresh water content derived from continental ice, as depicted in the various sea level reconstructions through the last deglaciation (e.g. Bard

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

et al. 1989, Fairbanks and Peltier, 2006). The idea that ocean volume changed greatly on very long time scales because of 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 already discussed), 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 fresh water injection at local scales during melting phases, and an excess of evaporation during glacial build up.

Melting ice sheets will also change the gravity field and, thereby, influence regional sea level. For example, removal of West Antarctica would actually lead to 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   Age 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 the last 40,000 years at most, but 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). Approximate uranium series dates are also attainable from corals (e.g. Gallup et al. 2002, Thompson and Goldstein 2005) or even directly from sediments (Henderson and Slowey 2000).

Other time-markers are available, usually through correlation with major events identifiable in cores and that have been dated in non-marine material. Examples include matching abrupt changes observed in sediment properties to changes dated in speleothems (e.g. stalagmites or stalactites) or ice core records (Marchitto et al. 2007), inferring the timing of glacial terminations from the dating of coral terraces (e.g. Thomson, 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 and Wunsch 2004).

Another common approach is so-called "orbital tuning", which depends on the assumption that variations in proxies can be matched 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, Shackleton et al. 1995), but can instead result from standard tuning methods independent of any true orbital signal (Neeman 1993, Huybers and Aharonson, 2009).

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

Given the irreducible uncertainty in the timing of paleoclimate records, there is an urgent need for statistical methods which 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 1996, Mudelsee et al. 2009), calibrating age using uncertain radiocarbon dates (Buck, 2004), and testing for covariance between time-uncertain records (Haam and Huybers, submitted). Time-uncertainty merits serious attention from the signal processing and statistical community.

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[16].

Ample evidence shows that the water mass structures of the North Atlantic were different during the LGM — unsurprising given the very different 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 those differences were. Some further comments about the nature of North Atlantic behavior during the last deglaciation can be found in Wunsch (2007) who suggested that the North Atlantic area is too small to dominate the global climate system.

A number of authors (e.g. Toggweiler and Samuels, 1995, Toggweiler and 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 which 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 MOC control, rather than 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[17].

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)

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[16]That much paleoclimate interest initially focussed 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.

[17]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 time scales of millions of years.

General circulation models of the ocean, as sub-components of more general models of the climate system, are essential tools that have proven highly useful in depicting and understanding the ocean circulation. These models are seductive — time is speeded up enormously, unobserved phenomena can apparently be computed, it can be done on dry land, etc. — but they are immensely complicated machines, involving hundreds of thousands of lines of computer code assembled by numerous individuals over the decades since about 1955. Like all powerful tools, considerable skill is required to successfully use them.

In particular, they are called "models" and not "reality" because they are necessarily imperfect approximations to the ocean. Among many problems, 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[18]. But there are many other sources of error, including the numerical approximations to the Navier-Stokes equations, inaccurate parameterization of sub-grid scale processes (commonly mixing, internal waves etc.), error-prone initial and boundary conditions, among others. 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 like $t^{1/2}$ (a random walk). That is, all non-trivial time-stepping models can be expected to accumulate errors as the time-horizon of integration grows. Unhappily, this phenomenon is rarely remarked. 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 the 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. The last 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 flux in a model may have a near constant error, while the movement of a passive "dye "injected into a model could diverge from reality with time, at least until approaching

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[18]Basili et al. (1992) report rates of about one error for every two thousand lines of FORTRAN code in operational programs used for flight dynamics. This rate represents a decline from four per two thousand lines in the testing phase of the models.

a relatively uninteresting final state of uniform dispersal.

That a great variety of errors occur in ocean models can hardly be doubted. Consider Figs. 10, 11 (Hecht et al. 1995) who introduced a dye patch into an extremely simple oceanic circulation model (Stommel's 1948 analytical one). Fig. 11 shows the analytical solution (computed by quadrature) for the position and shape of the patch after 1.5×10^8s (about 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 such 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 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 depiction and understanding of 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 and 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. A suite of nine models subjected to LGM forcing and boundary conditions under the auspices of the Paleoclimate Modeling Intercomparison Project 2 showed four with an increase in overturning, four a decrease, and one with 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) conclude that,

> "Based 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 we do not yet know from observations how the overturning circulation changed during the LGM, 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 elements of those data are robust and some have extremely fragile interpretations that rest only upon a series of assumptions that may once have seemed vaguely plausible — but through familiarity have been elevated to the status of facts. Distinguishing these extremes is the heart of the paleoceanographic problem.

Three reasons exist for an oceanic focus as here. First, it 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. Enough has been learned to show that without a doubt the system shifts on all time scales. 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 and Heimbach, 2008).

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 under-constrained problem, unless and until, there is a technical or intellectual breakthrough. (Perhaps ancient or modern DNA will permit for the decoding of past climates, e.g. Waller et al, 2007) Breakthroughs and their influence upon the field cannot be predicted and one can best proceed under the assumption that great uncertainties will linger indefinitely. In such a situation, where the data are sparse and ambiguous, it is essential to avoid focusing on one possibly 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[19]. For example, among the many hypotheses would be those that make the ocean circulation the "trigger" of climate change and that such triggers lie 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.

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[19]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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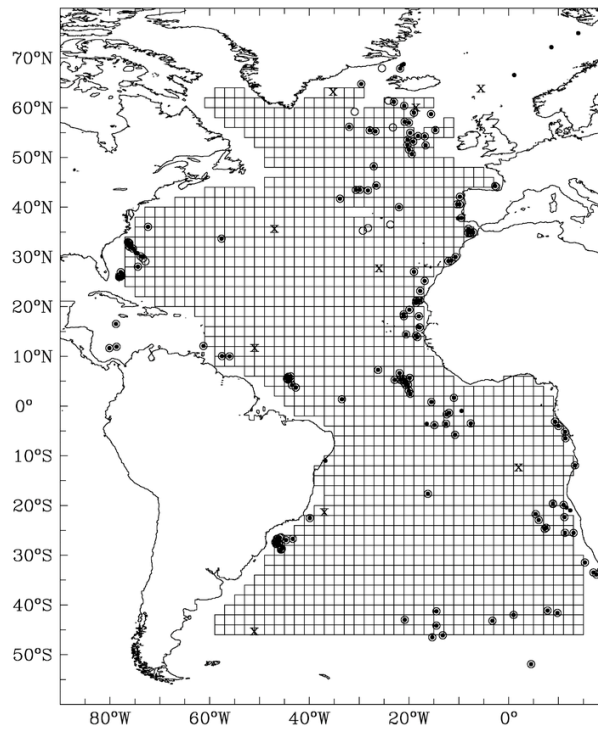
Figure 1: From Marchal and Curry (2008; courtesy O. Marchal). The distribution of cores in the North Atlantic. Solid dots are those available for the Holocene (about the last 12,000 years), open circles are those used for the LGM, and the 'x' denote some modern water column chemistry measurement positions. Compare the data density to that in Fig. 6 for the present-day ocean.
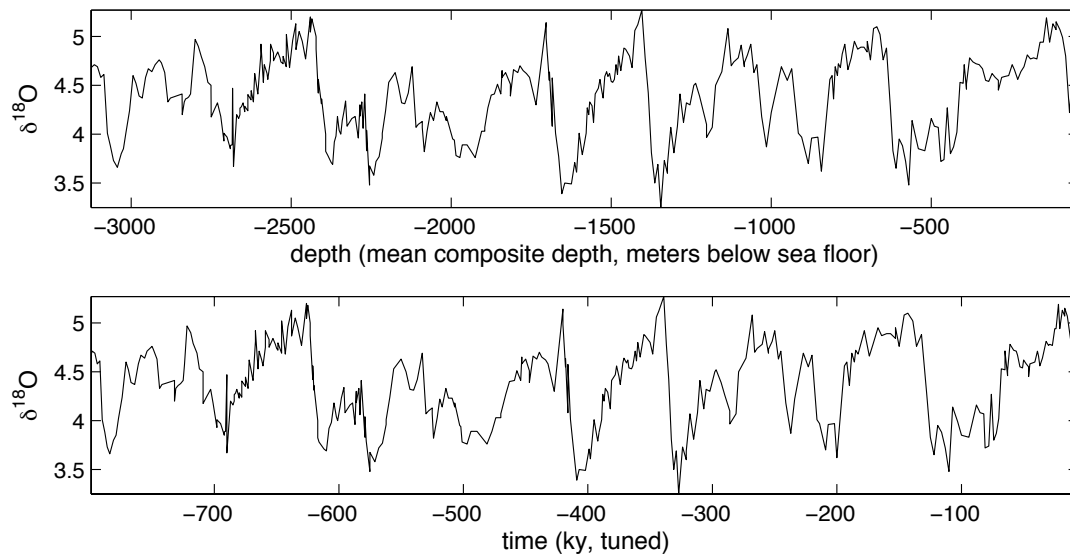
_curry_data_dist}

Figure 2: A δ^{18}O stratigraphy from Ocean Drilling Program site 677 in the Eastern Equatorial Pacific (Shackleton et al. 1990) plotted against depth (top) and ages obtained from orbital tuning (bottom). This core is known for having 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, then when plotted against the orbitally-tuned time estimate (between 20 and 70 ky ago).

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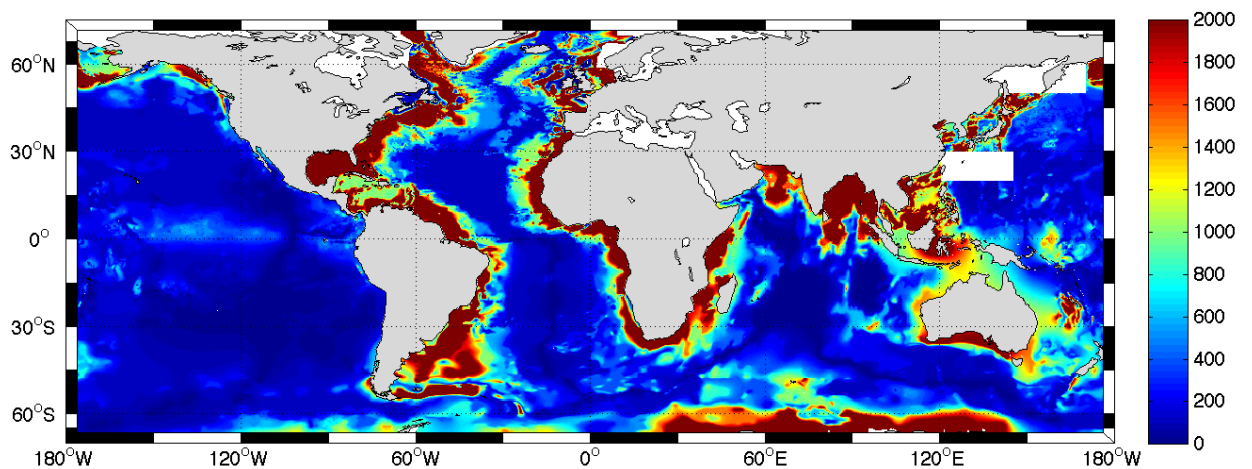
Figure 3: The thickness of marine sediments in meters (Divins, 2002). Shading is saturated at 2000 meters to permit visual resolution of regions with thin sedimentary deposits — comprising most of the oceans. Regions above sea level are in gray and missing data in white.

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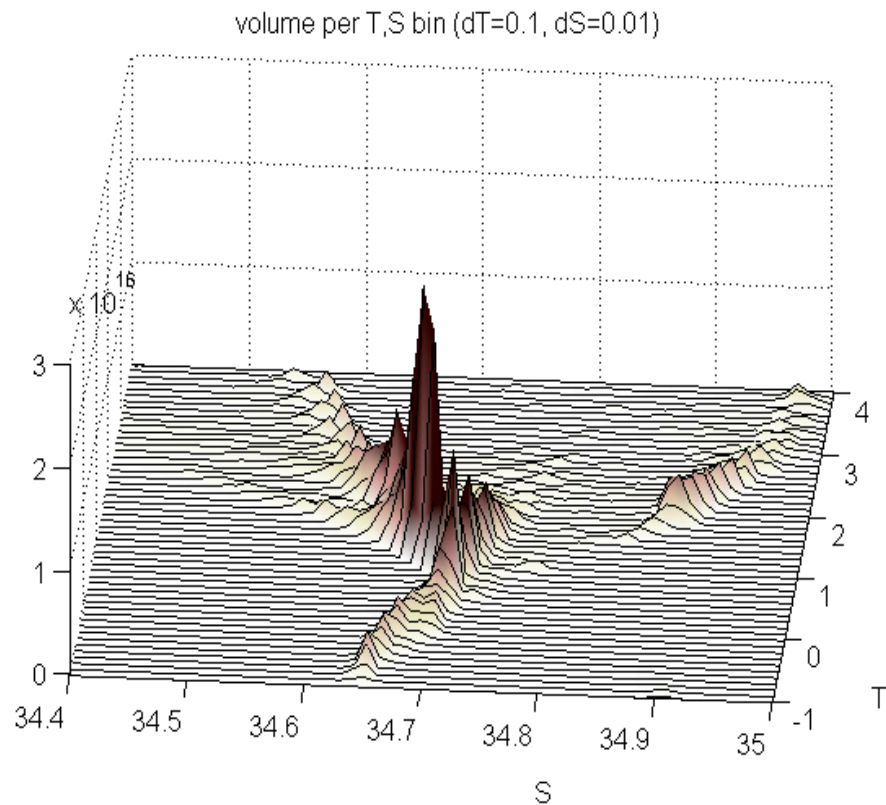
Figure 4: Approximate histogram (m^3) of deep water masses of the modern world ocean (below potential temperature of 4°C derived from the analysis in Forget (2008); G. Forget, private communication, 2008). These volumetric censuses shift with the climate state but by themselves carry no information about their rate of production or of their movement, other than in a quasi-steady state that it must be in balance with the rate of destruction. The largest peak here is largely 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 upon the WOCE-era modern hydrography. Earlier versions (e.g., Worthington, 1981) lacked coverage over much of the ocean, including especially, the southern hemisphere. Temperature is in °C and salinity is dimensionless as measured on the practical salinity scale. Notice the large-number of identifiable end members. The observed structures are not well understood (see McDougall and Jackett 2007).
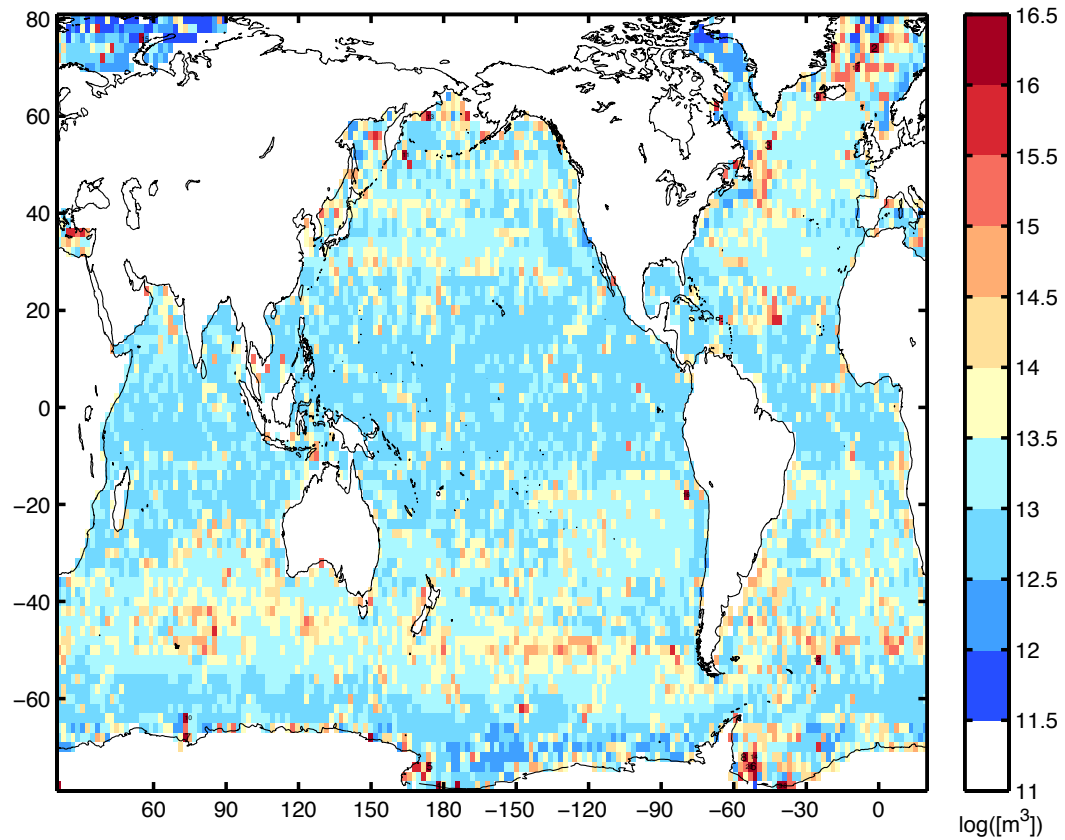
{forget_fig2.p

Figure 5: The oceanic volume traceable to each 2° by 2° surface location from Gebbie and Huybers (2009). The colorscale corresponds to volume in units of $\log_{10}((m)^3)$. Note the detailed structure and large number of surface points which contribute to the interior, defying any interpretation of the interior ocean as being filled from a small number of deep water formation cites.
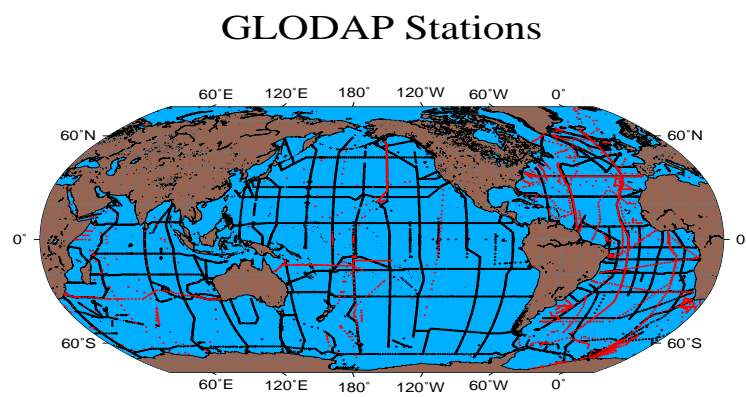
{Gebbie.pdf}

GLODAP Stations

Figure 6: The distribution of stations used to compute the modern radiocarbon age (Key et al. 2004; courtesy R. Key). Compare it, in the Atlantic, to the positions in Fig. 1
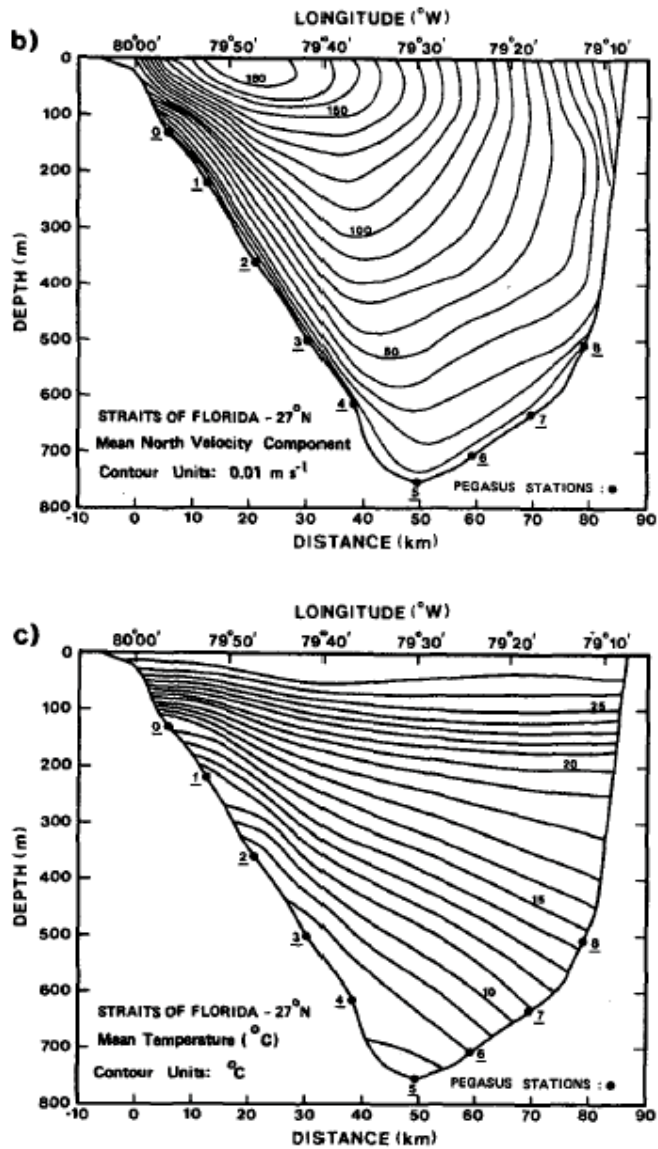
{key_et_al_200

Figure 7: Time average velocity (upper) and temperature field in the modern Florida Current near Miami (Leaman et al. 1987). The bottom velocity is almost everywhere 10cm/s and note the complex structure of the temperature field, with significant shifts in the inter-isotherm distances on the east. They also display the standard deviations of these fields. Little seems understood of how the temporal variability affects the paleotemperature proxies.
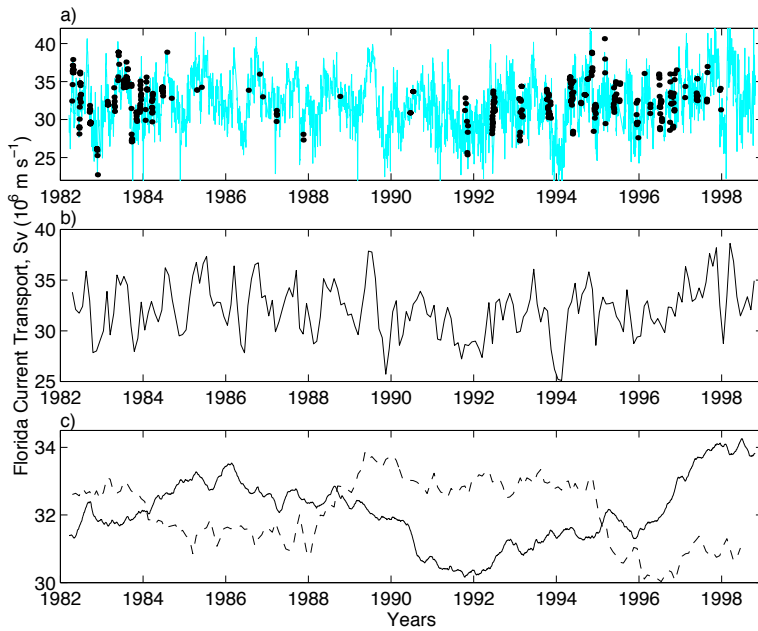
{leaman_etal.p

Figure 8: Measured values through time of the volume transport through the Florida Straits (Baringer and Larsen 2001, courtesy M. Baringer). Values reach as high as 40Sv and as low as 25Sv, raising the question of what values are reflected in the temperature proxies of the sediments.
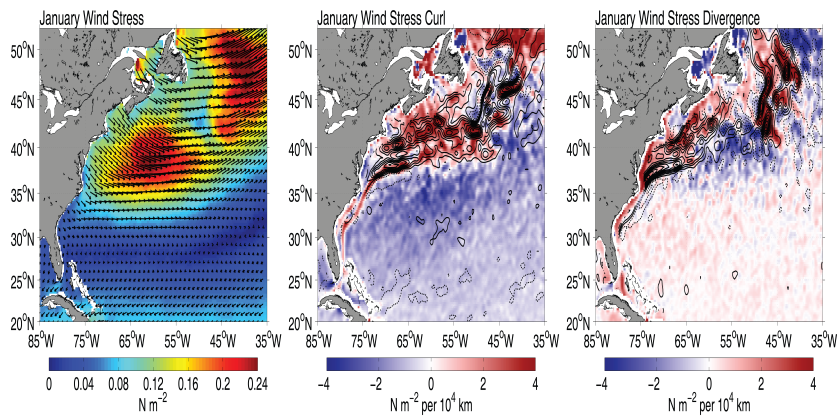
{baringer_lars

Figure 9: Time average (over 8 years) of estimated wind stress and its curl in the western North Atlantic (Risien and Chelton 2008; courtesy C. Risien); from scatterometer measurements. The spatial complexity has a direct influence on the resulting ocean circulation. A comparable time-dependence exists, not shown here. Some of the small-scale "mottling," especially visible in the curl, is probably noise in the measurements.

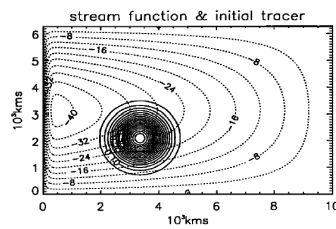ton_natlantic.pdf}

stream function & initial tracer

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 followed in time. After five years, its exact structure and position are shown in Fig. 11.

_etal_1995_1.pdf}

reference solution

CTCS

(a)

(b)

MPDATA

UTOPIA

(c)

(d)

QUICKEST

QUICKEST (uncompensated)

(e)

(f)

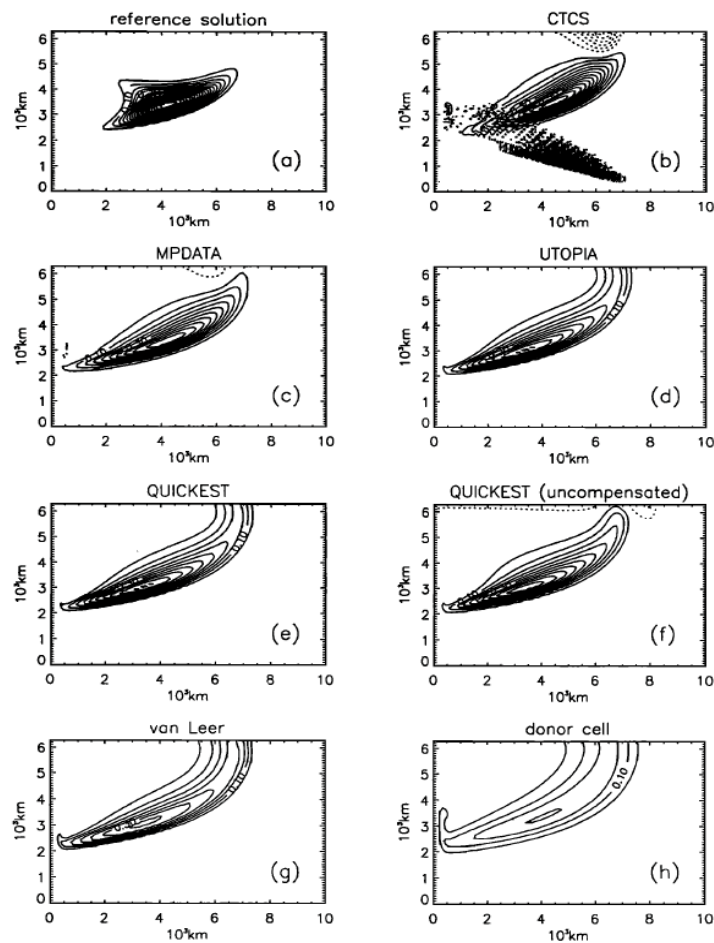van Leer

donor cell

(g)

(h)

Figure 11: Upper left panel shows the exact solution for the dye patch concentration after about five years found by Hecht et al. (1995). Remaining panels show the concentrations calculated by eight 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).

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