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The ocean component of the global water cycle

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Fundamentals of the Ocean Water Cycle

Introduction

The distribution of evaporation and precipitation over the ocean (its hydrologic cycle) is one of the least understood elements of the climate system. However, it is now considered one of the most important, especially for ocean circulation changes on decadal to millennial time-scales. The ocean covers 70% of the Earth's surface and contains nearly all (97%) of its free water, thus, it plays a dominant role in the global water cycle. The atmosphere only holds a few centimeters of liquid water, or 0.001% of the total. However, most discussions of the water cycle focus on the rather small component associated with terrestrial processes [*Chahine,* 1992]. This is understandable, since the water cycle is so vital to agriculture and all of man's activities. Yet current estimates indicate that 86% of global evaporation and 78% of global precipitation occurs over the oceans [*Baumgartner and Reichel,* 1975]; (Figure 1) . Since the oceans are the source of most rainwater, it behooves us to work toward a better understanding of the ocean hydrologic cycle; small changes in ocean evaporation and precipitation patterns may have dramatic consequences for the much smaller terrestrial water cycle. For example, if less than 1% of the rain falling on the Atlantic Ocean were to be concentrated in the central US, it would double the discharge of the Mississippi river!

Similarly, it is now apparent that the hydrologic cycle has a direct impact on the thermohaline circulation of the ocean (that part of the circulation driven by heat and salt differences), which is recognized as a key element of the climate system for variations on decadal to millennial time scales. The poleward transport of heat by the ocean and atmosphere serves to moderate high latitude temperatures. Meridional ocean heat transports are about equal to those in the atmosphere, and a large fraction of the ocean transport in the northern hemisphere is carried by the thermohaline overturning cell in the Atlantic [*Bryden,* 1993]. If this circulation is disrupted, as it appears to have been in the past [*Broecker,* 1987; [*Keigwin et al.*, 1991], the consequences for high and mid-latitude continental climate are severe [*Manabe and Stouffer,* 1993]. The cause of collapse of the thermohaline circulation is thought to be the "halocline catastrophe" [*Bryan,* 1986]; whereby deep convection and the formation of bottom waters can cease if surface salinity decreases sufficiently through enhanced freshwater input. Since the distribution of evaporation, precipitation, ice and continental runoff is the primary factor in the determination of surface salinity, it is essential that we adequately understand the ocean hydrologic cycle. In the following sections I review recent progress in understanding the patterns of water exchange between ocean and atmosphere, and the oceanic flows resulting from mass conservation and simple dynamics. Implications for climate models and ocean monitoring are also discussed.

Description

A rough global picture of the net water flux between ocean and atmosphere is given in Figure 2 . This is a composite picture, assembled from modern climatologies for evaporation minus precipitation (E-P) for the Atlantic [*Schmitt et al.*, 1989] and over the rest of the ocean estimates due to [*Baumgartner and Reichel* [1975]. These are based on older coast and island data for the most part and thus lack resolution over the open ocean. The *Schmitt et al.*, [1989] estimates are derived from the latent heat flux data of *Bunker* [1976] and the rainfall data of *Dorman and Bourke* [1981]. Their estimates have more detail than the seminal study of [*Baumgartner and Reichel,* 1975], since the data are based on millions of ship observations from the 1940s to the 1970s, and display much more structure. New features are seen, such as a minimum in E-P trending from northeast to southwest across the North Atlantic subtropical gyre. Also, a prominent region of net water loss near the Gulf Stream has no corresponding pattern in the North Pacific near the Kuroshio. Such a pattern would be present, if modern climatologies were to be assembled for the Pacific Ocean as well.

Despite the limitations of such a map, the major features of the water cycle are clearly seen. In the tropics, rainfall dominates over evaporation within the Intertropical Convergence Zone (ITCZ). The subtropics are characterized by an excess of evaporation, except for the South Pacific Convergence Zone, a band of net precipitation trending to the southeast away from the western equatorial Pacific. Horizontal gradients in net water flux can be large. For instance in the Atlantic at 30°W, well over 1 m/yr of net precipitation is found at 5°N, and more than 1.4 m/yr of net evaporation is estimated just 10 degrees poleward.

In subpolar latitudes precipitation dominates, though more so in the North Pacific than the North

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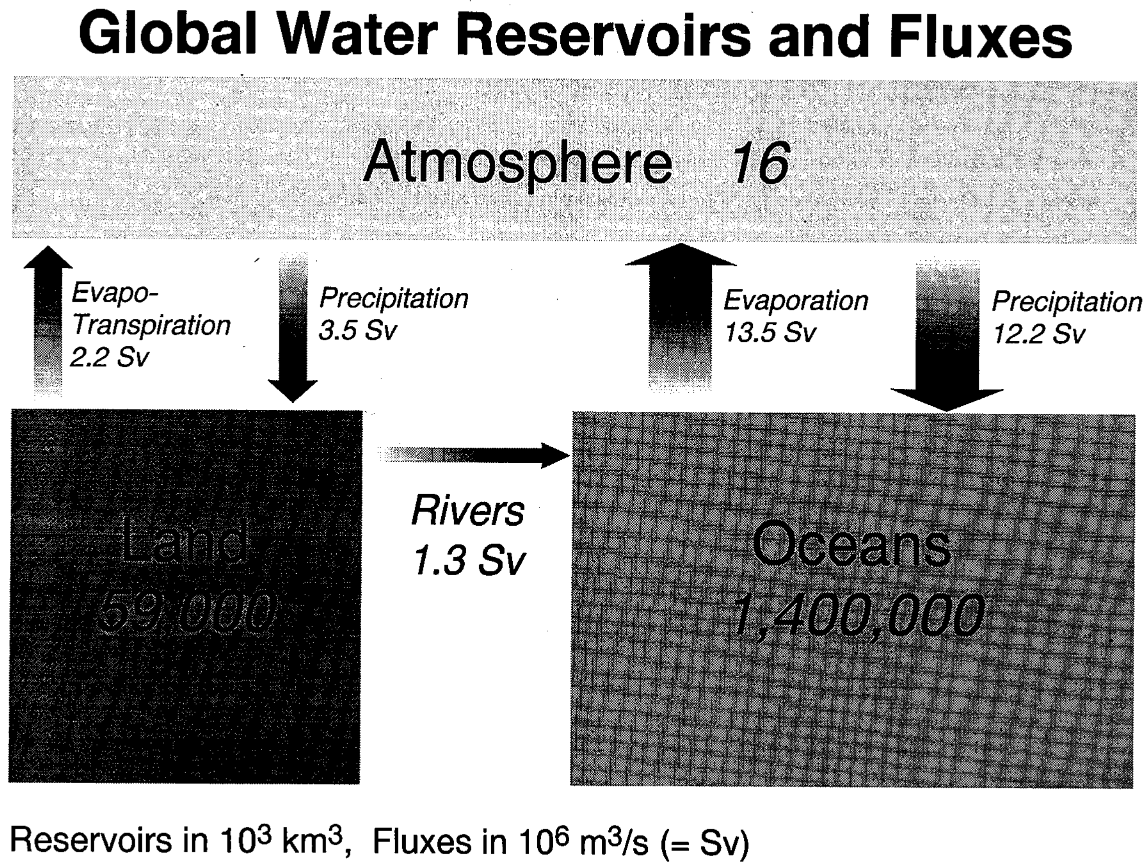

Figure 1. Estimates of the global water storage in oceanic, terrestrial and atmospheric reservoirs and the fluxes among them, using data of *Baumgartner and Reichel* [1975].

Atlantic. There is little data near the poles themselves. The amplitude of the hydrologic cycle is reduced at high latitudes by the low water vapor capacity of a cold atmosphere. There the processes of freezing, melting and transport of sea ice play an important role in the hydrologic cycle. In general, the patterns of E-P are strongly zonal, except for the North Indian Ocean, where evaporation dominates in the Arabian Sea, precipitation in the Bay of Bengal.

Meridional Fluxes

The general pattern of net evaporation in middle latitudes and net precipitation in the tropics and higher latitudes, leads to the necessity for compensating flows in the ocean. That is, the ocean must transport water into the evaporation zones and away from the precipitation regions, in order to avoid regional trends in sea level. An appreciation for the magnitude of such flows can be obtained by integrating the surface flux estimates shown in Figure 2. As the major gradients are meridional, a zonal integration in 5° latitude bands is presented, that shows the pattern of meridional transport (Figure 3, from *Wijffels et al.*, 1992). This can be compared with estimates of the water vapor transport in the atmosphere [*Peixoto and Oort*, 1983]. As noted by *Wijffels, et al.*, [1992] and *Schmitt and Wijffels* [1993], the ocean and atmosphere fluxes are about equal and opposite; meridional river transports are one or two orders of magnitude smaller. The ocean/atmosphere hydrologic cycle accounts for a significant portion of the poleward heat transport on the planet. For instance, a water transport of 0.6 Sv corresponds to 1.5 petawatts of latent heat flux (1 Sv = 1 Sverdrup = 10^6 $m^3/s = 10^9 kg/s$). At 25°N, the ocean sensible heat flux is about 2 petawatts (= 10^{15} watts = PW) [*Bryden*, 1993], so the (separate) heat transport associated with the ocean/atmosphere hydrologic cycle (0.8 PW) is quite substantial. As discussed later, ocean data alone can be used to estimate the latent heat component of meridional heat transport, thus permitting the specification of a large part of the total planetary energy budget [*Schmitt and Wijffels*, 1993].

Interbasin Transports

Notable differences exist in the amounts of water gained or lost by each ocean. There is an excess of precipitation in the North Pacific, especially in the eastern tropical Pacific, and a dominance of evaporation in the Atlantic. The Pacific-Atlantic difference is thought to be maintained by water vapor transport across Central America [*Zaucker and Broecker*, 1992] and the lack of any similar transport into the Atlantic from Africa. [*Schmitt, et al.* 1989] also point out that the Atlantic is relatively narrow, and thus under the influence of continental air over a greater fraction of its area. The western tropical Pacific also gains water vapor exported from the Indian Ocean. As a result, the North Pacific

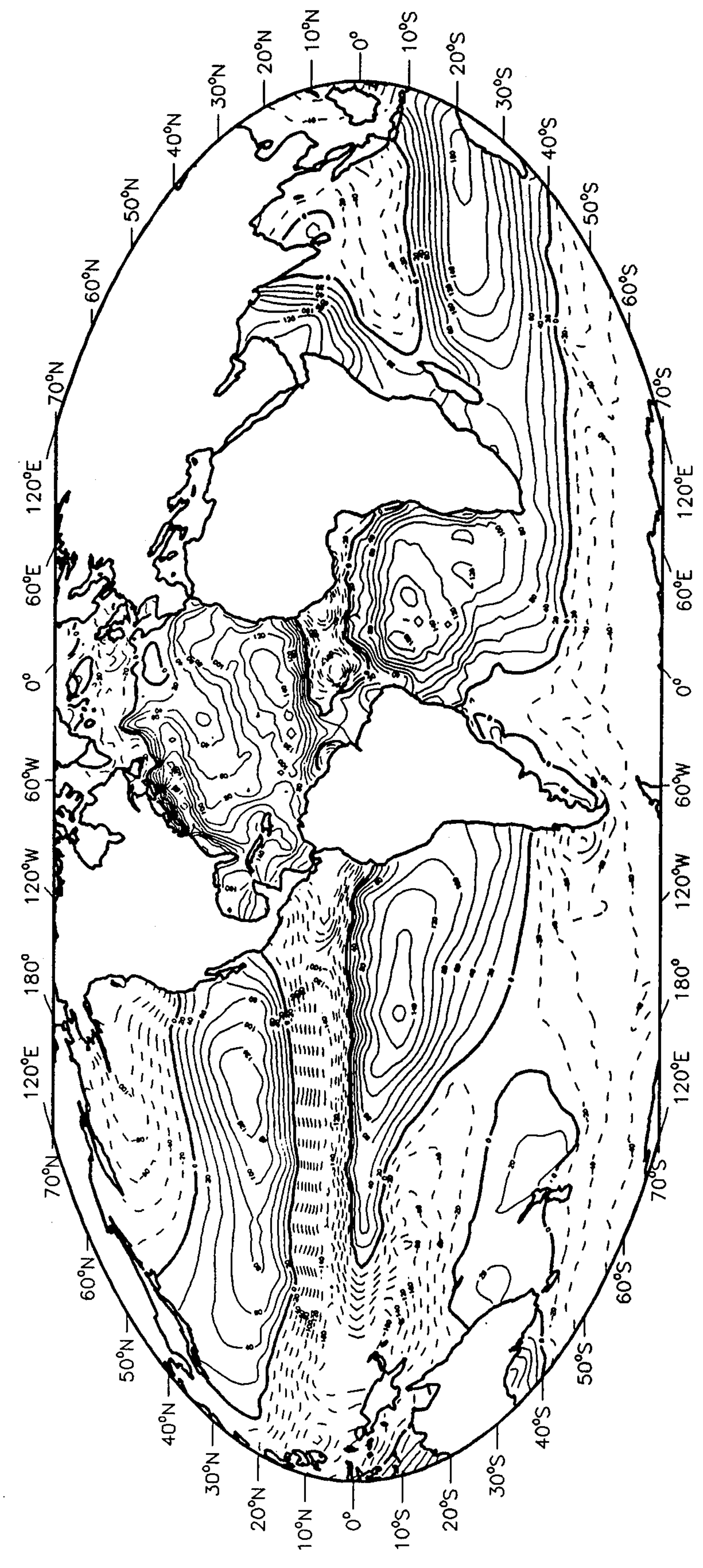

Figure 2. A map of evaporation minus precipitation (E-P) over the global ocean, in centimeters per year. This is a composite of *Schmitt et al.* [1989] for the Atlantic and *Baumgartner and Reichel* [1975] for the other oceans.

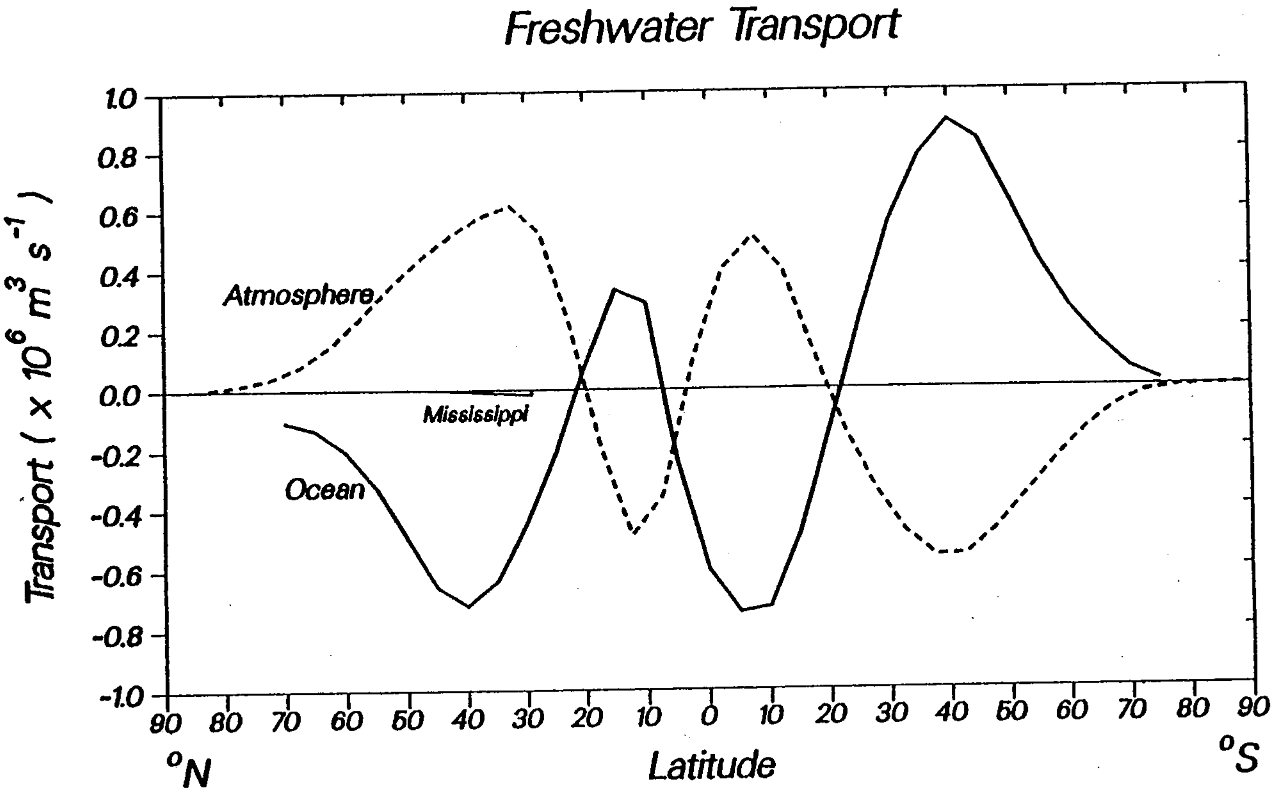

Figure 3. The meridional transport of water by the global ocean and atmosphere, in Sverdrups ($= 10^6$ m^3/s). The oceanic estimate derives from the integration of the E-P value shown in Figure 2, plus river discharges into the ocean. The actual meridional transport by rivers alone is small, an estimate for the Mississippi is shown. The atmospheric transport of water vapor comes from *Peixoto and Oort* [1983].

is notably fresher than the other oceans, especially the North Atlantic.

The differences in surface water fluxes also mean that there is a necessity for interbasin transport in the ocean. This could be accomplished by a number of different paths, given the multiply- connected nature of the ocean basins. *Baumgartner and Reichel* [1975] constructed a scheme which assumed zero water transport across the Atlantic equator, and integrated their surface flux estimates relative to that point. However, *Wijffels, et al.* [1992] showed that this arbitrary assumption could be replaced by direct oceanographic transport measurements in Bering Strait. The scheme for ocean water transports they developed differs dramatically from the earlier work (Figure 4), which had confused the flux convergence in the Arctic basin with a flux. Indeed, they find that most of the North Pacific water gain (nearly a Sverdrup), is transported through Bering Strait. An assumption of zero water flux across the Pacific Equator appears to be nearer the truth.

The interbasin water transport is accompanied by a salt transport. Since there is no significant atmospheric pathway for salt, the salt flux through coast to coast sections within a given sub-section of the interconnected oceans must be the same. That is, the northward salt flux across 24°N in the Pacific is equal to that through Bering Strait and the southward transport across any zonal section in the Atlantic. This constraint allows us to calculate the freshwater flux divergence between sections. Such direct ocean estimates can be used to evaluate integrations of surface fluxes [*Schmitt and Wijffels*, 1993]. When complemented by western boundary current measurements and Ekman transport (upper-ocean, wind-driven flow) estimates, the salinity and geostrophic velocity fields (orthogonal to the pressure distribution) from a hydrographic section can be used to calculate the salt flux. As noted by [*Schmitt and Wijffels* 1993], the salt flux can be subtracted from the mass flux to give the net (fresh) water flux divergence between sections in the same basin (or channel). Two examples are given there, one for the freshwater gain between Bering Strait and 24°N in the Atlantic, the other for the gain between 24°N in the Pacific and 24°N in the Atlantic. Estimates of the flow through Bering Strait [*Coachman and Aagaard*, 1988] and 24°N data analyzed by *Hall and Bryden* [1982] allow a calculation of the freshwater input between the sections. This gain is estimated to be about 0.1 Sv. In contrast, the climatologies sum to about zero water gain in the Arctic and Atlantic north of 24° N. This would indicate the necessity to increase high latitude precipitation or runoff estimates (or decrease evaporation) in the climatologies, though the discrepancy is not substantial compared to the uncertainties which must ap-

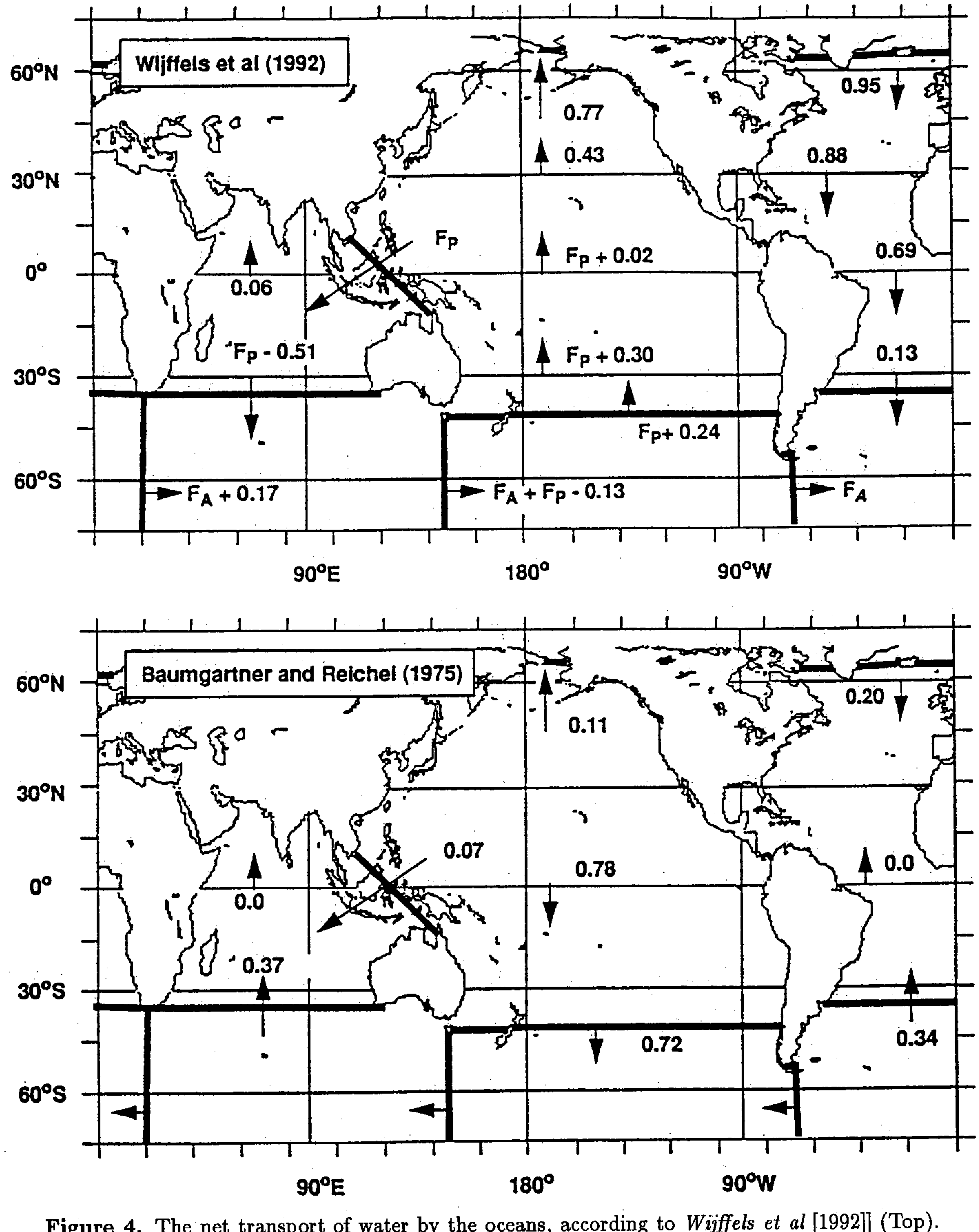

Figure 4. The net transport of water by the oceans, according to *Wijffels et al* [1992]] (Top). Contrast this scheme with that of *Baumgartner and Reichel* [1975] (Bottom).

ply to both types of data (Figure 5). Also shown is the recent freshwater flux estimate of *Friedrichs and Hall* [1994] for 11°N. This latitude is at an extrema in water transport, and also indicates that the climatologies have insufficient water gain to the north. However, because of the uncertainties associated with stronger, and more variable, Ekman and eddy fluxes at 11°N, the error bars are larger than at 24°N. Field studies in the dynamically interesting tropics will be very valuable for improving error estimates there.

A more extensive area of ocean can be evaluated by considering sections at 24°N in both the Atlantic and Pacific. Since the salt flux in these two sections must be equal (though opposite in sign), by virtue of the

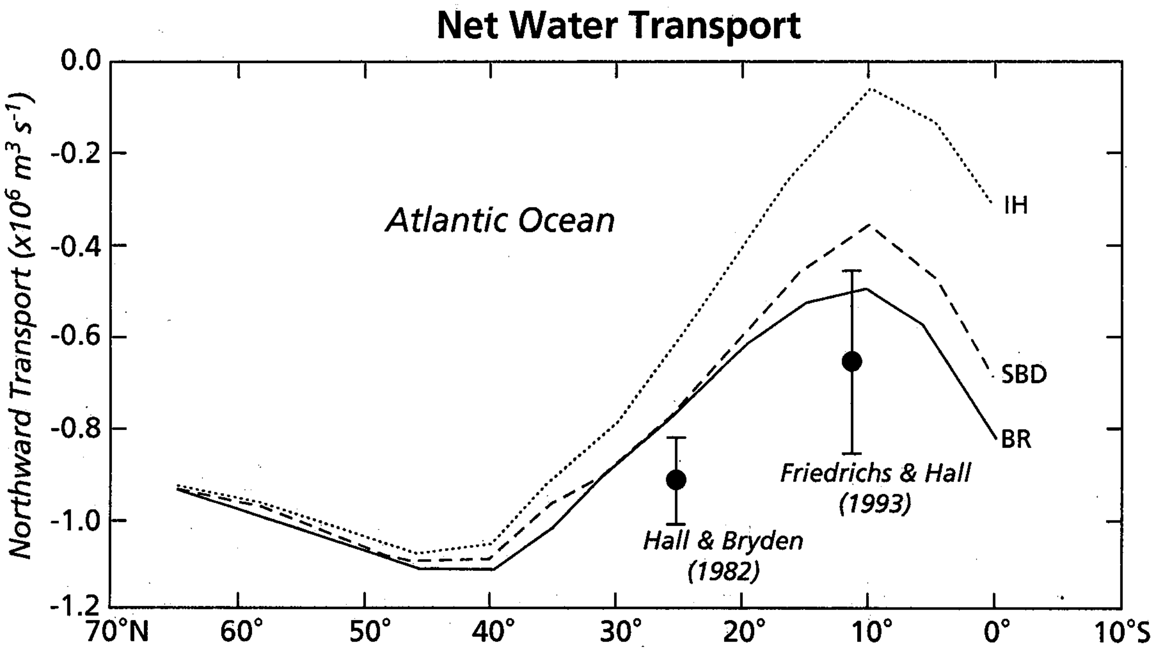

Figure 5. Meridional tranport of water in the Atlantic ocean, according to three different surface flux climatologies. All are summed relative to an estimated Arctic southward export due to the Bering Strait throughflow and the water budget of the Arctic itself. The climatologies are those of *Baumgartner and Reichel* [1975] (solid line), *Schmitt et al* [1989] (dashed line) and the combination of *Isemer and Hasse* [1987] evaporation estimates with *Dorman and Bourke* [1981] precipitation values (dotted line). Also shown are ocean based estimates at 24°N by *Hall and Bryden* [1982] and 11°N by *Friedrichs and Hall* [1993].

Bering Strait through-flow, it is possible to calculate the total ocean freshwater gain north of 24°N. Only the section data of *Hall and Bryden* [1982] (Atlantic) and *Bryden, et al.*, [1991] (Pacific) are necessary for this calculation, the precise magnitude of the Bering Strait through-flow is not critical. The result is a net southward transport of water by the ocean of 0.31 Sv. Combined with southward meridional river flows of about 10% of this value, the ocean/river return flow is in reasonable agreement with atmospheric water transport at this latitude [*Peixoto and Oort,* 1983]. This is an interesting result, as there has been speculation that the "missing petawatt" [*Bryden,* 1993] resulting from the discrepancies among the meridional heat flux estimates due to the Earth's radiation budget, atmospheric transport and ocean transport, might be found in unresolved latent heat transport in the atmosphere. However, since the return flow would be in the ocean, the data do not support an increase of more than double the presently estimated meridional transport of water at 24°N, which would be required for the extra petawatt [*Schmitt and Wijffels,* 1993].

The more complete set of zonal lines to be collected during the World Ocean Circulation Experiment (WOCE) program will permit even more points of comparison between the surface estimates and ocean flux divergences. Indeed, preliminary results of J. Toole [*personal communication,* 1994] indicate that the mid-latitude Southern hemisphere oceans carry half the heat flux and twice the water flux of the Northern hemisphere oceans. This striking asymmetry suggests that the atmosphere/ocean system has multiple ways of satisfying the global radiation budget (i.e., an enhanced latent heat flux in the atmosphere/ocean system can make up for reduced sensible heat transport in the ocean). It is quite possible that the enhanced water flux over the southern ocean helps to decrease the heat-flux-carrying thermohaline circulation there by freshening the surface waters and limiting deep convection. In a stronger CO_2 greenhouse climate it is hypothesized that the hydrologic cycle will intensify. Could it intensify sufficiently to limit the thermohaline circulation in the North Atlantic?

This question is of obvious first order importance, yet we will have trouble addressing it because our knowledge of precipitation and evaporation over the ocean is so rudimentary. The difficulties of making precipitation measurements at sea, the paucity of data for calculating evaporation, and lingering uncertainty about the bulk formula combine to make the net air–sea water exchange a very poorly known quantity. As can be seen in Figure 5, there are significant discrepancies between available climatologies for the North Atlantic which serve to illustrate the problem. When summed over the North Atlantic the *Schmitt et al.*, [1989], numbers yield 10^5 m^3/s more evaporation than the Baumgartner and Reichel (1975) summation. This is over half the flow of the Amazon. Even greater discrepancies arise when more recent climatologies are considered for the evaporation field. For instance, *Isemer and*

Hasse, [1987] have revised the Bunker estimates for the North Atlantic. While they adjust the exchange coefficients in the bulk formulae downward, their revision of the Beaufort wind scale leads to stronger trade winds and increased evaporation in the tropics and subtropics. Their estimate of 20 to 40 cm/year more evaporation south of 40°N suggests that an additional freshwater loss of nearly 4×10^5 m^3/s may occur over the North Atlantic. Thus, current estimates of net E-P over the basin differ by 2.5 times the flow of the Amazon (1.9 $\times$ 10^5 m^3/s). Since the North Atlantic is a small, relatively well sampled basin, we can only infer that the uncertainties for the other ocean basins are even larger. *Zauker and Broecker* [1993] find that atmospheric general circulation models at present also do a poor job of representing the vapor transport out of the Atlantic basin.

Application to Ocean Dynamics

Barotropic Response: The Goldsbrough-Stommel Circulation

The exchange of water between ocean and atmosphere has consequences for ocean flows beyond the requirement of mass conservation. The simplest (yet largely neglected) response is that governed by vorticity conservation requirements which derive from the conservation of angular momentum on our rotating planet. *Goldsbrough*[1933] first showed how surface mass fluxes drive steady flows in the ocean, in what may be viewed as a precursor to the well-known Sverdrup relation. That is, it is readily shown [*Huang and Schmitt,* 1993] that the steady vorticity balance in the ocean interior is given by:

$$\beta_f V = \mathbf{k} \cdot \nabla \times \tau + f(E - P), \qquad (1)$$

where f is the Coriolis parameter, β_f its meridional gradient, τ the wind stress and V the meridional velocity. Goldsbrough (who considered only the (E-P) term above) suggested that a subtropical-gyre-like circulation would result from an interior precipitation regime and an evaporative western boundary region. This is of course quite unrealistic; *Stommel* [1957] suggested that the interior Goldsbrough circulation could be closed by adding western boundary currents. The complete gyres can be refered to as the Goldsbrough-Stommel circulation.

Huang and Schmitt [1993] have presented a preliminary calculation of the Goldsbrough-Stommel circulation for the world ocean (Figure 6). While E-P is smaller than Ekman convergence (1 m/yr vs. 20 m/yr), the flows generated are generally opposite to the wind driven currents. In the North Atlantic, a western boundary current is required which reaches 2 Sverdrups southward at 35°N, in direct opposition to the Gulf Stream. Other basins have similarly strong boundary currents. We suggest that these adverse flows can affect the separation latitude of the western boundary currents. For the Gulf Stream, the effect may be as large as 75 km. As there are strong horizontal temperature gradients associated with the western boundary currents, any small shifts in separation latitude could have large impact on SST distributions in the western portions of the subtropical gyres.

Baroclinic Response: The Thermohaline Circulation and Climate Models

Evaporation or precipitation serves to concentrate or dilute the dissolved salt content of surface sea water, and thus directly affects its density. The haline density flux ($\beta_s F_S$) is given by [*Stern,* 1975]:

$$\beta_s F_S = \frac{\rho_o \beta_s (E-P) S}{\rho_S (1-S)} \approx \beta\ (E-P)\ S \qquad (2)$$

where $\beta_s = 1/\partial\rho(\partial\rho/\partial S)$, is the haline contraction coefficient, and ρ_0 and ρ_S are the densities of pure and sea water. The haline contribution to the buoyancy flux is often comparable to that of the thermal buoyancy flux ($= \alpha Q$, where Q is the net heat flux and $\alpha(= -1/\rho(\partial\rho/\partial T))$ is the thermal expansion coefficient). *Schmitt, et al.* [1989] compared the annual mean thermal and haline buoyancy flux estimates for the North Atlantic by contouring the absolute value of the ratio of the two components. They find that in the subtropical gyre the haline density flux is often larger than the thermal density flux, though the general pattern is one of thermal flux dominance in high and low latitudes. However, this is based entirely on open ocean estimates of surface fluxes. Continental runoff and ice freezing and melting processes contribute to significant concentration of the haline density flux at the coasts, where it can easily dominate the buoyancy change locally.

The role of the haline buoyancy flux in ocean dynamics was first examined by *Stommel* [1961] with a simple box model. He found multiple stable states of the system, which was forced by both heating/cooling and evaporation/precipitation. Other box models were discussed by *Rooth* [1982] and *Walin* [1985]. *Bryan* [1986] showed how a general circulation model would display instability of the thermohaline circulation when restoring boundary conditions were replaced by flux boundary conditions for the salinity. He coined the phrase "haline catastrophe" to refer to the freshening of high latitude surface waters and shutdown of the thermohaline circulation. Since then a great variety of models of varying complexity have focussed on the sensitivity of the thermohaline circulation to freshwater forcing [*Marotske, et al.* 1988; *Huang, et al.*, 1992; *Huang and Chou,* 1994; *Quon and Ghil,* 1992; *Zhang, et al.*, 1993; *Weaver and Sarachik,* 1991; *Weaver et al.*, 1991; *Marotske and Willebrand,* 1991; *Stocker and Wright,* 1991; *Wang and Birchfield,* 1992; *Weaver and Hughes,* 1994; *Tziperman, et al.*, 1994; *Mikolajewicz and Maier-Reimer,* 1994]. *Holland, et al.* [1995] discuss such models in their review. The essential point is that the "conveyor belt" of the thermohaline circulation can be very sensitive to the

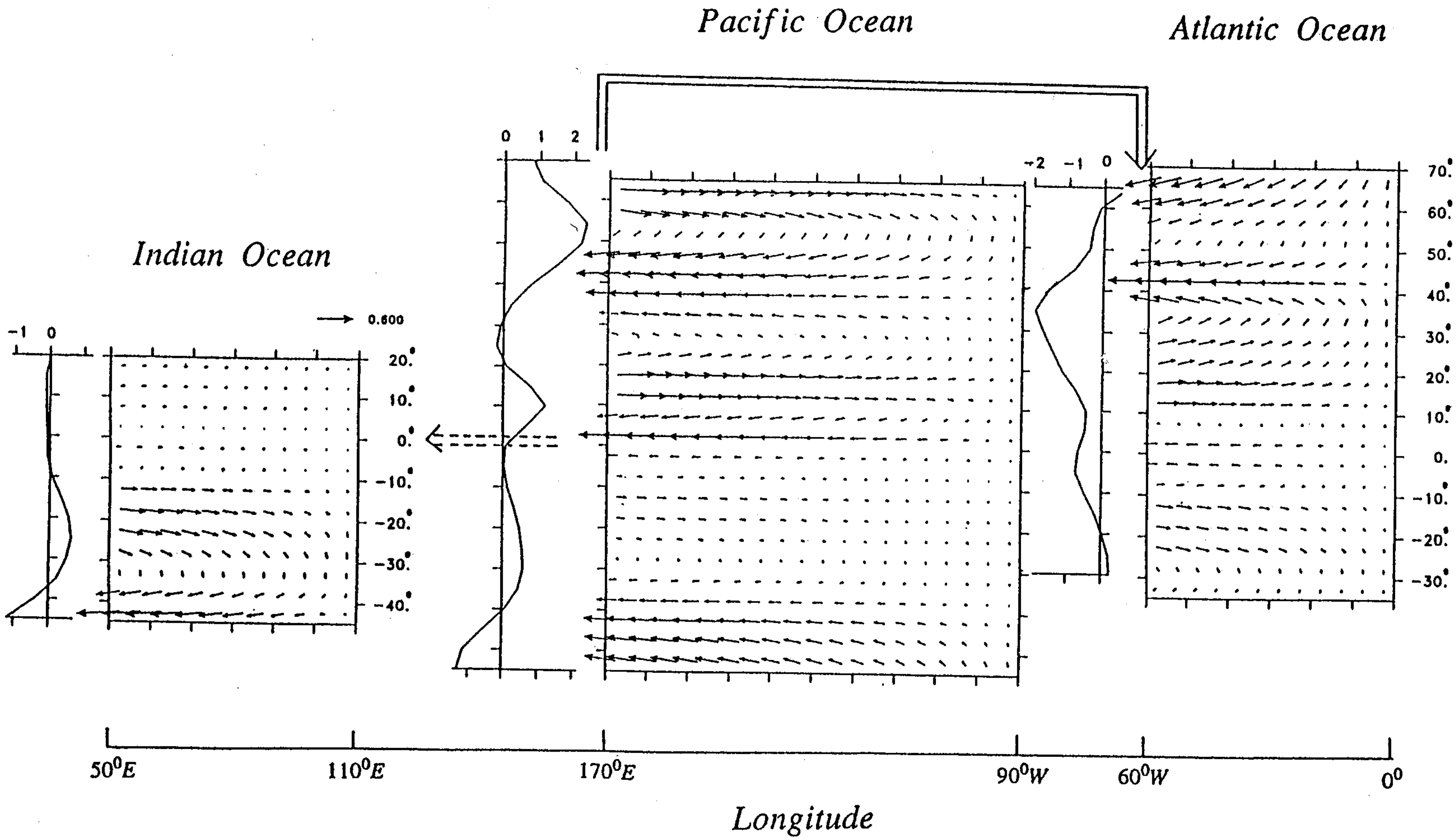

Figure 6. The Goldsbrough-Stommel circulation in the global ocean, according to *Huang and Schmitt* [1993]. Each arrow shows the horizontal mass flux integrated over a 5° square, in Sverdrups. Along the left edge of each basin, there is a curve indicating the northward transport within the western boundary (in Sv) which is required to close the circulation. The Pacific to Atlantic transport via the Arctic is shown, and is included in the boundary current transports. The Pacific to Indian transport is only depicted schematically.

freshwater flux at high latitudes. Many of these models display fluctuations on millenial time scales. However, *Delworth, et al.* [1993] have carefully examined the behavior of a realistic circulation model, and found significant variability on 40–60 year time scales. Advection of heat and salt play an important role in the variability, which appears to be governed by the strength of a baroclinic gyre set up in mid-basin. The persistence of sea-surface temperature anomalies is an important element, so the common relaxation boundary conditions, which couple the ocean too tightly to the atmosphere, are not appropriate. *Huang* [1993] has pointed out the importance of real freshwater flux as a boundary condition for models, in part because of the Goldsbrough-Stommel circulation, which is missing when a psuedo salt flux is applied. Also of interest is the modelling study of *Shaffer and Bendtsen* [1994], who find a great sensitivity of the thermohaline circulation to the transport through Bering Strait. Because this supplies freshwater to the Atlantic, it acts as a negative feedback on the thermohaline conveyor belt. They propose that a more variable climate would be realized with a higher sea level and thus greater Bering Strait transport. Such conditions may have been achieved during the last interglacial period, when the Greenland Ice Cores indicate that the climate had less stability than at present *Dansgaard, et al.*, 1993]. Other paleoclimate records suggest that sudden changes in the North Atlantic thermohaline circulation occured many times during the last deglaciation [*Lehman and Keigwin,* 1992], when substantial discharges of meltwater are thought to have occurred at high latitudes. We also note that strong haloclines in the tropics, formed by the rainfall under the ITCZ, may also have an impact on upper ocean mixing and heat exchange processes there [*Lukas and Lindstrom,* 1991; *Carton,* 1991; *Delcroix and Hénin,* 1991].

There is also historical evidence that perturbations of the hydrologic cycle do influence the climate system. In particular, fresh salinity anomalies in the subpolar basin have been observed to survive for the order of a decade while being advected by the surface gyre. The Great Salinity Anomaly (GSA) of the 60's–80's was observed for 15 years as it traveled around the entire subpolar North Atlantic gyre [*Dickson,et al.*, 1988]. There is strong evidence that the GSA had substantially supressed deep water formation in the Greenland and Norwegian Sea and flux in the western boundary currents [*Schlosser, et al.*, 1991]. The GSA is thought to have arisen from excess ice discharge from the Arctic through Fram Strait. An increase in the annual export of ice of only 20% [*Aagaard and Carmack,* 1989] could have provided the necessary freshwater anomaly.

A potentially related process has been identified by *Deser and Blackmun* [1993]. They found a mode of Atlantic surface and air temperature variability with a dipole structure concentrated in the western North Atlantic, by examining 90 years of temperature data. Winter temperatures east of Newfoundland are out-of-phase with those off the US mid-Atlantic states. The mode has an amplitude of about 1°C and a 10–12 year period. Interestingly, it appears to be related to the ice extent in the Labrador sea, with Labrador sea ice leading low temperatures off Newfoundland by one-two years. Advection of a low salinity cap formed by the melting ice would be consistent with such a relation.

Thus, there is substantial evidence that the thermohaline circulation has intimate ties to climate, especially on the decadal time scale. It is therefore important to further define its sensitivity to perturbations in the freshwater flux, which is only hinted-at in the currently meager data sets. Both improved models and an expanded observational base are required.

Implementation of an Observational System

Having discussed some of the fundamental aspects of the ocean hydrologic cycle and its importance to ocean dynamics, it is now useful to identify approaches with which we will be able to improve open ocean estimates of water exchange with the atmosphere. As there is no "reference standard" against which to calibrate such estimates, we will be best served by a multitude of approaches. Both the mean state and its temporal variability are of interest; the different techniques possess strengths and weaknesses for these applications.

Development of climatologies

As noted earlier, presently available global climatologies leave much to be desired in terms of the spatial and temporal resolution of net water flux over the ocean. While separate climatologies for precipitation [*Legates and Wilmott,* 1990] and evaporation [*Ebensen and Kushnir,* 1981; *Hsiung,* 1986] have been developed, there have been no attempts to compute a global E-P climatology since *Baumgartner and Reichel* [1975]. In part this is due to the distrust of the process of differencing two numbers with sizeable uncertainties. However, failure to attempt such a climatology leaves even the most fundamental features of the system unspecified. There is a great need for definition of at least the general amplitude and patterns of surface water fluxes as input for ocean models and for comparison with evolving observational techniques.

Precipitation estimates over the sea have been developed by compiling statistics on the frequency, type and intensity of rainfall from the "present weather" observations of mariners. *Tucker* [1961] originated this approach, in which coast and island rainfall data are used to develop empirical relations with the local marine observations. These regressions are then applied over the open ocean. *Reed and Elliot* [1979] prepared precipitation maps for the North Atlantic based on rain frequency with an additional seasonal correction. *Dorman and Bourke* [1979, 1981] use rainfall frequency, type and intensity as well as an additional correction for temperature, in order to achieve adequate dynamic range in the tropics. As millions of ship observations are available over the open ocean, this technique eliminates the need to extrapolate across the ocean basins from coast and island data alone. Thus, much more detail can be seen in their charts compared to those of Baumgartner and Reichel.

There is also hope that the variability in oceanic precipitation can be determined by use of the present weather codes. D. Cayan [*(personal communication,* 1994] reports that many 5° squares in the North Atlantic and Pacific have more than 200 observations per month. While systematic trends in the data and changes in reporting procedures raise some difficulties, he finds significant correlation of marine precipitation patterns with nearly land data.

Evaporation is computed from the standard "bulk formula" relating air–sea exchange of moisture to boundary-layer humidity gradients and the 10 meter wind speed. A number of basin scale [*Bunker,* 1976; *Isemer and Hasse,* 1987] and global scale [*Hsiung,* 1986; *Ebensen and Kushnir,* 1981] climatologies have been produced. Calculations have also been made of variability in the latent heat flux [*Cayan,* 1992].

Atmospheric Water Flux Divergence

Meteorologists are able to describe the large-scale patterns of atmospheric water vapor transport by the use of humidity and velocity profiles from radio-sondes. The scales of variability in the atmosphere are large compared to the ocean, so widely spaced station data may be sufficient. By taking the divergence of estimated horizontal fluxes, the transport into or out of the surface can be calculated. *Rasmusson* [1968, 1971] has used this technique to examine the hydrology of North America. *Peixoto and Oort* [1983] have developed a global climatology of surface water flux using the divergence calculations. While such data can provide general patterns, the lack of data over the oceans prevents the detection of features like the E-P maximum in the vicinity of the Gulf Stream, which is observed in the ship-based climatologies [*Schmitt, et al.,* 1989]. However, *Walsh, Zhou and Portis* [1994] have utilized radio-sondes from Ocean Weather Ships along with land stations from the high latitude Atlantic to examine the trends of surface water flux within a few regional boxes. The results appear to be significant and to agree with the seasonal cycle as given by climatology. However, a rather small number of weather ships remain in operation, so this technique can be applied in only a few areas.

Numerical Weather Prediction Models

Another approach to the problem of evaluating the water cycle over the ocean is to examine the output of numerical weather prediction models (NWP). There is optimism that the surface fluxes of the hydrologic cycle might be more accurately modeled than the heat flux, because the constraint of water conservation and lack of complex radiation terms makes for a potentially simpler problem. Recent improvements in precipitation algorithms have made the models more accurate, at least over land where comparisons can be made [*Environment Canada, 1993*]. Preliminary comparisons of the monthly average E-P for two years from the European Center for Medium-range Weather Forecasts (ECMWF) model and the *Schmitt, et al.* [1989] 30 year climatology for the Atlantic are promising in the position and amplitude of major features. Both ECMWF and National Meteorological Center are planning "reanalysis" programs, in which the newer, more accurate models are run with data for the past one or two decades. If run with sufficient resolution, the time series of surface water flux over the ocean could be of great utility for climate studies.

Satellite Detection of Precipitation

The measurement of precipitation at sea remains one of the most challenging observational issues. Wind flow over ships influences the collection efficiency of gauges, and the short space and time scales of precipitation make point measurements difficult to interpret. Thus, remote sensing techniques must be used. Several satellite based systems, shipboard radar, and ocean acoustic noise measurements show potential for monitoring on the necessary space and time scales. *Rasmusson and Arkin* [1993] have reviewed the climate scale estimates of global precipitation derived from satellites. However, calibration of such techniques remains an area requiring new approaches and much work.

There are several indicators of rainfall which can be derived from satellite observations [*Arkin and Ardanuy,* 1989]:

Highly Reflective Cloud (HRC) — This technique uses visible light to identify those clouds which are most likely to be rain sources. The frequency of HRC can be related to rainfall measurements from atoll stations. It is somewhat subjective, suffers from poor diurnal sampling, and is restricted to the tropics. It does have the advantages of a long record, beginning in the early 1970s, and high spatial resolution. This approach is one of those used in the WCRP Global Precipitation Climatology Project (GPCP).

Outgoing Longwave Radiation (OLR) — Values of OLR in the tropics respond strongly to variations in cloudiness, and is thus a useful index of convective activity. It has been used as both a qualitative and quantitative indicator of precipitation. Its sampling of the diurnal cycle, at twice per day, is better than HRC, though still suboptimum. It is restricted to low latitudes where precipitation is primarily from deep cumulonimbus convection (+−40 °). *Janowiak and Arkin* [1991] and *Arkin and Xie* [1994] describe some result of its use within the GPCP. OLR data are also being used as indirect forcing of the heating in NWP systems; these systems in turn are giving improved estimates of tropical precipitation.

Passive Microwave — a. SSM/I (Special Sensor Microwave/Imager) This technique works over the oceans, where a uniform background of microwave emission allows rainfall to be detected by the absorption of microwave energy by raindrops or the scattering of higher-frequency microwave energy by ice particles. There are polar orbiting satellites carrying the appropriate sensors, so it shows promise for the important problem of the high latitude, open ocean detection of rainfall [*Wilheit* et al., 1991]. However, the technique cannot be used over land, snow or ice. The 30-km footprint is larger than most rain cells; the partial occupation of the footprint and a nonlinear response to rain degrade its utility in the tropics. Larger scale rain systems at higher latitudes may be adequately resolved by this technique. Because of spotty space and time coverage, accurate climate scale estimates of accumulations ($5^{\circ} \times 1$ month) may be problematic, though less so outside the tropics. In the tropics, the passive microwave satellites may be used to "calibrate" the OLR technique, which has better space/time sampling. *Chang, Chiu and Wilheit* [1993] provide a global precipitation map derived from the SSM/I.

b. MSU (Microwave Sounding Unit). The microwave brightness temperature measured from satellites can be corrected for air mass temperature; when it exceeds a certain threshold precipitation is inferred because of cloud-water and rainwater induced warming. *Spencer* [1993] has developed a global ocean precipitation climatology for 1979-1991 based on the MSU. It compares favorably to the *Legates and Wilmott* [1990] compilation (based largely on the Dorman and Bourke application of the Tucker method), though it indicates higher precipitation in the eastern tropical Pacific (to 5.6 m/yr).

Tropical Rainfall Measuring Mission (TRMM) - This is a rain radar satellite to be launched in 1997 as a joint Japan/U. S. mission (*Simpson et al.*, 1998). It will fly in a low-altitude, low inclination orbit to provide good resolution at latitudes less than $\pm$ 35 and will be non-sun-synchronous in order to sample the diurnal cycle over monthly periods. Rain radars must use additional information such as the reflectivity as a function of dropsize distribution, or direct rain gauge measurements, in order to estimate a rain rate. Comparisons with point measurements are good to only a factor of 2; however, averaging over larger space and time scales improves the accuracy.

In-Situ Rain Measurements

The foregoing remote sensing techniques all require sea truth to generate credible precipitation products. This is a formidable challenge because the measurement of precipitation at sea is anything but routine. Rain gauges on small, low atolls (with no orographic or "warm lagoon" effects) are extremely valuable for calibration, but are not available in most areas. Rain gauge measurements from ships are generally regarded as futile. The most promising approaches are:

Instrumented Surface Buoys: Two rain gauge types are in use:

Capacitive filling bucket gauges - These are being used on the improved meteorological measurement (IMET) buoys being deployed in upper ocean experiments by R. Weller.

Optical rain gauges - Assumptions about drop size distribution are required to convert the output to a rain rate. These are being used on TOGA TAO moorings in the equatorial Pacific.

Subsurface acoustic — The ambient noise created by rainfall may be distinguished from that caused by wind by the peak near 15 kHz [*Scrimger, et al.*, 1987]. This technique has the advantage of being an integral measure rather than a point sample. However, much work remains to be done (in particular, calibration against other approaches) before it can be considered a routine measurement. It may be possible to deploy acoustic rain (and wind) gauges on drifting surface buoys.

Surface radar — Rainfall is estimated from meteorological radars operated from ships during special experiments. While only good to 50 to 100% in individual comparisons, data on rain drop size distributions and averaging on larger space and time scales can improve accuracy to 10–25%. This is likely to remain a research tool rather than a routine monitoring technique.

Salinity

Sea surface salinity (SSS) is a good indicator of net surface water exchange [*Wüst*, 1936; *Jacobs*, 1948]. SSS is controlled by E-P, advection and mixing with underlying water. With a sufficiently well resolved SSS, surface velocities and mixed layer depth, one could infer E-P from ocean data alone. P. Niiler [*Personal communication*, 1994] has reported progress with this technique using climatological data. With data of sufficient spatial resolution, the detailed patterns of the mean water flux should be discernible. However, temporal variability will be difficult to establish, given the paucity of SSS data. Features such as the climatically significant Great Salinity Anomaly [*Dickson, et al.*, 1988] are at present detected only by serendipity. Thus, an important priority for future ocean monitoring is the improvement of surface and upper-ocean salinity measurements. Ultimately, the incorporation of upper ocean salinity data into assimilative ocean models will provide an additional method of constraining the surface water fluxes.

Fortunately, the measurement of salinity has become more routine due to the development of compact, stable conductivity sensors. It is now possible to obtain reliable salinity records from shipboard thermosalinographs and long term records from moorings or drifting buoys. In addition, expendable and autonomous profilers are now available. Finally, remote sensing techniques have been proposed. These approaches are discussed below.

Moored salinity measurements. *McPhaden et al.* [1990] have successfully deployed conductivity- temperature recorders on upper ocean moorings in the Equatorial Pacific for periods of 6–7 months.

Drifting salinity measurements. *Swenson et al.* [1991] have explored the feasibility of deploying C/T sensors on surface drifters, with good results.

Thermosalinographs. Somewhat trouble-prone in the past, there is reason to believe that new instrumentation and a systematic approach to deployment on the Volunteer Observing Ships (VOS) will yield a valuable enhancement of surface salinity data. Several pilot projects are currently underway.

XCTD. While long in development, it appears that some success has been achieved in using the eXpendable Conductivity-Temperature-Depth probe (XCTD) as an operational instrument [*Roemmich, personal communication* 1992]. If deployed in quantity from the VOS, it would greatly enhance the subsurface data sets.

Retrievable CTD. Small conductivity-temperature-depth (CTD) packages with internal recording could be easily deployed on light line and retrieved, much like the mechanical bathythermograph. This would allow inexpensive CTD profiles to be obtained from a moving ship.

Autonomous Profilers. R. Davis (Scripps Institution of Oceanography) has deployed profiling floats (Autonomous Lagrangian Circulation Explorer, ALACE) that can be fitted with conductivity sensors. Promising results have been obtained to date. Other autonomous profilers under development, such as the self powered Slocum [*Stommel,* 1989], will also provide salinity profiles.

Remote Sensing. The Electonically Scanned Thinned Array Radiometer (ESTAR) is a potential remote sensing technique for monitoring the large scale distribution of surface salinity [*Swift,* 1993]. It depends on the influence of salinity on microwave emissions; this effect is strongest at 1.4 GHz. However, temperature has a larger effect and high accuracy temperature measurements (to .02° K) must also be made. This can be done using another frequency (2.65 or 5.0 GHz) or temperature data from another source. This is a very demanding requirement, and yields a salinity accuracy of 0.05 parts per thousand (ppt). It can only be achieved by long time and space averaging, i.e. 30 days by 100 km. This could be useful for climate scale monitoring. It is estimated that the Great Salinity Anomaly could have been detected by an ESTAR, though the sensitivity is reduced to 0.1 ppt at low temperatures. Ideally, long term service by a series of satellites would be continued for decades to be of utility for ocean climate. If improved resolution of 10 km is desired in coastal areas then the accuracy degrades to 2 ppt. Another motivation for ESTAR is its primary application over land: soil moisture measurements. However, at the present time there are no plans for deployment of an ESTAR.

Ocean Flux Divergence

As discussed in Section 1, oceanographic flux divergences can be used to constrain the net water exchange between ocean, atmosphere and land. Some aspects of making such estimates are discussed here. Depending on the volume in question, river flows can be an important part of the water budget. These are generally monitored to sufficient accuracy and are available from the Global Runoff Data Center in Koblenz, Germany. However, some rivers, such as the Amazon, are gauged well upstream of the mouth, and the discharge to the ocean may not be adequately known.

The latitudes of most interest for the global water cycle are distinct from those that best define the meridional heat flux. Ocean water fluxes reach extrema at around 10°and 45°N and S (Figure 2), reflecting evaporation in mid-latitudes and precipitation at high and low latitudes (Figure 3). In contrast, heat flux is maximal at 25–30°N and S. If we suppose that freshwater flux divergence could be estimated to about 0.1×10^6 m^3 /s from hydrographic data at 45°N and 10°N in the Atlantic, then this would correspond to an error in E-P of about 12 cm/yr, significantly less than the differences between present climatologies. With few zonal ocean sections having been occupied, and none on a seasonal basis, it is difficult to estimate the error of such calculations due to unsampled seasonal and interannual variability. *Friedrichs and Hall* [1993] estimate the error at 11°N to be twice that at 24°N in the Atlantic, based on only one section. The "direct" method of flux calculation appears to yield an accurate estimate of mean heat flux, which, in the Atlantic, is largely carried by the baroclinic overturning cell. Estimates of heat flux at 24°N from section data taken 3 different times over 35 years are found to be identical [*Hall and Bryden,* 1981; *Roemmich and Wunsch,* 1985; *Parilla et al.,* 1994]. Recent model runs [*Boning and Herrmann,* 1994] indicate that the North Atlantic response to wind stress variations is mainly barotropic, and thus does not substantially change the density field. While other oceans (and latitudes) may be more sensitive to seasonal variability, these results do suggest that further development of zonal hydrographic sections would be an important contribution toward understanding the global water and heat cycles.

The WOCE Hydrographic Programme will improve the global coverage of sections, however, long time series of repeat sections would be a very valuable follow-on to the first order picture developed from the WOCE data set. Occupation of sections with seasonal resolution may be justified at some latitudes (10°N), while others with a weak seasonal cycle (24°N) might be occupied every few years. In the Atlantic a zonal section can be completed in less than 3 weeks. As a example, we note the 27 occupations of a section at 165°E, from 20°S to 10°N, over a 7-year period by the French, U. S., and Chinese during TOGA. With the number of countries with strong oceanographic programs being quite high around the Atlantic, it should be relatively easy to focus resources on repeat occupations of several zonal sections in that ocean. The resulting improvement in ocean heat and freshwater flux estimates would provide valuable constraints on surface fluxes derived from satellites, climatologies, and weather forecasting models. They would also provide insight into the mechanisms of freshwater transport that is presently lacking.

Strait monitoring will also be necessary to define the water budget for certain regions. Bering Strait is the main mass flux conduit for water returning from the Pacific to the Atlantic, and an important source of buoyancy for the Arctic Ocean. *Coachman and Aagaard* [1988] estimate a standard deviation in interannual variability in Bering Strait transport of about 0.1×10^6 m^3 /s. This seems large enough to have climatic significance *Shaffer and Bendtsen*, 1994], especially for the Arctic ice budget. A more complete monitoring program, including salinity measurements, would be a very valuable climate indicator for the high latitude fresh water budget.

The two layer flow at Gibraltar is the result of a loss of water to evaporation in the Mediterranean basin that represents one of the larger terms in the North Atlantic mass budget. Fram Strait carries freshwater and ice from the Arctic to key deep convection areas of the North Atlantic. It is clear that time series measurements of transports in these straits would be very valuable in deciphering oceanic climate response to variations in freshwater forcing. Thus, establishment and long term maintenance of transport monitoring programs in these straits should be an immediate priority for new ocean monitoring programs. For the Mediterranean, the constraint on the water budget provides a link with the heat budget through the latent heat term [*Garrett et al.*, 1992]. Improved Gibraltar flux measurements could be instrumental in deciphering the ongoing controversy about exchange coefficients and radiation terms in the surface heat fluxes.

Conclusions

As the ocean hydrologic cycle remains one of the most poorly understood components of the climate system, all of the above approaches must be pursued. Large programs such as the Global Energy and Water Cycle Experiment (GEWEX), though focussed on the "fast" atmospheric and terrestrial processes, will help refine our picture of the water cycle. Intercomparison of the various techniques will help to identify the best methods for monitoring. However only long term records, such as those planned as part of the Global Ocean Observing System (GOOS), will allow us to describe and understand the functioning of the hydrologic cycle in decadal and longer time-scale climate processes. The next decade offers ideal opportunities to begin to build a much more complete picture of the ocean hydrologic cycle.

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