

# Salinity and the Global Water Cycle



BY RAYMOND W. SCHMITT

Climate change is in the news frequently these days, and better-informed commentators note that changes in the water cycle pose the greatest challenge to society. We may be able to deal with an increase in temperature with a bit more air conditioning, but significant intensification in droughts and floods will be a more severe test. Civilization thrives at a range of temperatures at different latitudes, but it cannot cope without water. Thus, a number of national and international programs have been organized to assess and understand the “global water cycle.”

For the oceanographer, these programs have fallen short. The focus is invariably on the water cycle over land; generally, the much larger oceanic water cycle is ignored. This bias is understandable, because society’s water needs are land based. But, to gauge the water cycle, it is impossible to ignore the ocean. The numbers speak for themselves: the ocean contains 97% of the free water on our planet, and 86% of the evaporation and 78% of the precipitation occurs over the ocean (Schmitt, 1995). The ocean is the primary return conduit for water transported by the atmosphere. It is the dominant element of the global water cycle, and clearly one of the most important components

of the climate system, with more than 1100 times the heat capacity of the atmosphere. There is ever-increasing evidence that the origins of flood and drought cycles on land are closely linked to variations in the much larger oceanic water cycle. It takes only 1% of the rain falling on the area of the Atlantic Ocean to match the volume discharge of the Mississippi River. Sea-surface temperatures are well correlated with rainfall in adjacent land areas. In addition to these direct effects, water-cycle variations over the ocean have yet-to-be-understood impacts on the ocean’s ability to sequester and transport heat. No program can pretend to address the global water cycle without a substantial oceanic component.

Articles in this salinity-themed *Oceanography* issue articulate the potential of our rapidly expanding ability to measure salinity to enhance understanding of the global water cycle. A discussion of the articles appears at the end of this introduction, and the last of the articles (“What’s Next for Salinity?”) provides a short summary of the work of the CLIVAR Salinity Working Group. This article briefly discusses some basic aspects of the water cycle; more complete treatments can be found in Schmitt (1995) and Wijffels (2001).

## WHERE'S THE WATER?

The volume of water in the ocean is  $\sim 1.4 \times 10^9 \text{ km}^3$ , about 24 times as much as resides in lakes and glaciers on land and about 100,000 times as much as the atmosphere holds. Fluxes between these water-cycle reservoirs are highly concentrated over the ocean: the global total of evaporation from the ocean is estimated to be about 13 Sverdrups (Sv,  $1 \text{ Sv} = 1 \times 10^6 \text{ m}^3/\text{s}$ ), and the total rainfall on the ocean is about 12 Sv. The difference is made up by terrestrial runoff of 1 Sv. This 1 Sv difference is the same difference as that between terrestrial precipitation of about 3 Sv and evapotranspiration of about 2 Sv. Further perspective is gained by contrasting the water fluxes for individual ocean basins with particular land sources such as river discharges and glacial melt. Table 1 shows some of these quantities, in both cubic kilometers per year and Sv.

Comparison of these fluxes clearly shows that the water cycle is predominantly an ocean-atmosphere phenomenon. Even the mighty Amazon River appears insignificant alongside the total evaporation or precipitation of an ocean basin. Much-discussed issues, such as increasing glacial-melt rates, turn out to be small contributors to the water cycle when compared with the basic exchanges of water between ocean and atmosphere within ocean basins (Schmitt, 1998; Peterson et al., 2006). It is important to retain this perspective when considering the impact of a changing climate on salinities and ocean circulation. It is what goes on over the vast and poorly monitored open ocean that really matters!

Part of the challenge in understanding the water cycle stems from the difficulty in measuring fluxes over the ocean; the

Table 1. Some components of the global water cycle in  $\text{km}^3/\text{yr}$  and Sv ( $= 10^6 \text{ m}^3/\text{s}$ )

	$\text{km}^3/\text{yr}$	Sv
Pacific Evaporation	212,655	6.74
Pacific Precipitation	228,529	7.24
Atlantic Evaporation	111,085	3.52
Atlantic Precipitation	74,626	2.36
African Precipitation	20,743	0.66
N. American Precipitation	15,561	0.49
European Precipitation	6,587	0.21
Amazon River Discharge	6,000	0.19
Mississippi River Discharge	560	0.017
Greenland Glacial Melt Discharge*	225	0.007

\*According to Baumgartner and Reichel (1975) and Rignot and Kanagaratnam (2006)

numbers in Table 1 are rough estimates, with the largest uncertainties in the larger fluxes. We have only a basic description of the mean state of the water cycle and are very ignorant of its variations on all time scales. Evaporation is estimated from wind speed, air and water temperatures, and humidity. Rainfall estimates over the ocean using anything but statistical inferences have been rare until recently when satellite-based estimates became available. We have endeavored to address this lack by assembling some of the most recent climatologies for precipitation (the Global Precipitation Climatology Project, GPCP, at <http://precip.gsfc.nasa.gov/>) and evaporation using bulk formulae (Yu and Weller, 2007). After differencing, the resulting maps of annual average evaporation minus precipitation (E-P) reveal the general patterns of water exchange over the global ocean (Figure 1a).

The subtropical ocean stands out

as the main source of water for the atmosphere. The dry, subsiding air of the subtropical high-pressure domes feeds the trade-wind systems that evaporate water vapor from the sea surface and carry it equatorward. There, the rising air of the Intertropical Convergence Zone (ITCZ) releases the water vapor to rain out in the high-precipitation bands of the tropics. On the poleward side of the subtropical highs, the westerlies carry water toward high-latitude precipitation zones, with storms providing much of the moisture transport in contrast to the steady trade winds of the tropics.

These patterns of evaporation and precipitation are reflected in the distribution of surface salinity in the global

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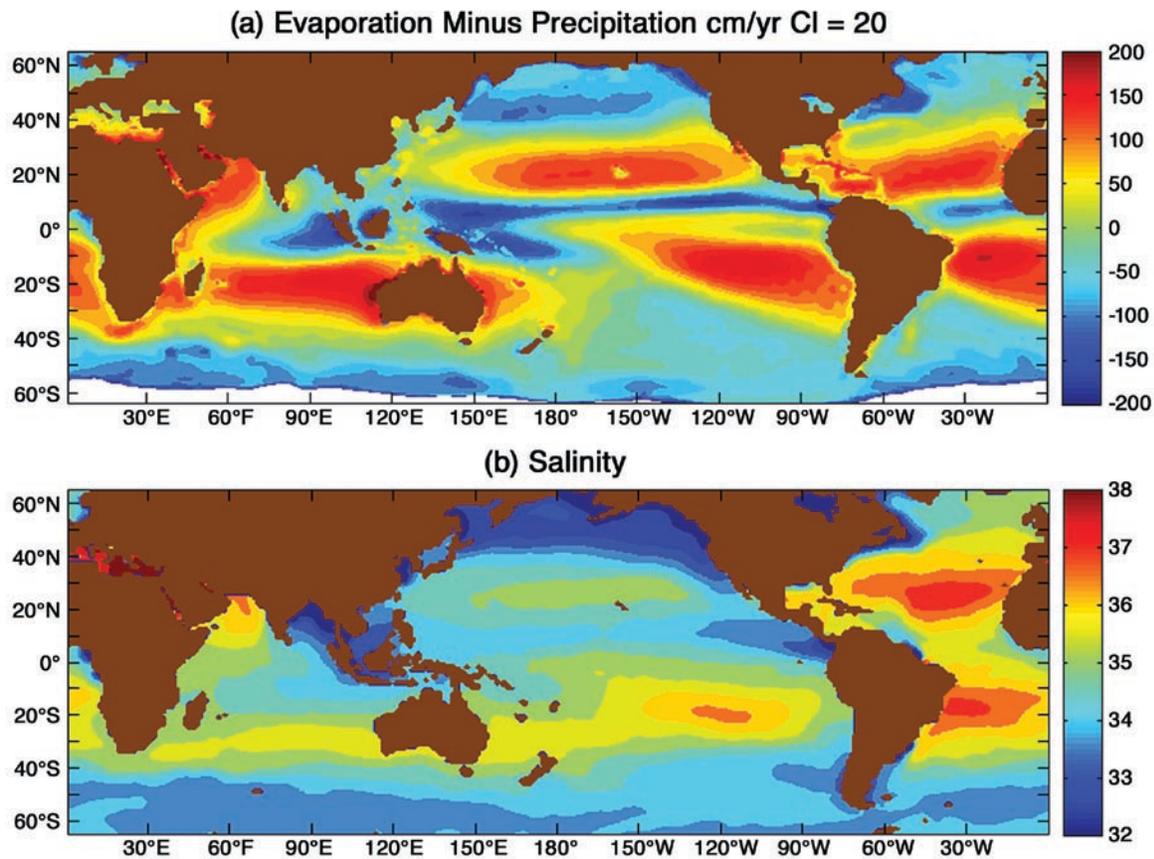


Figure 1. (a) A new estimate of annual average evaporation minus precipitation, based on the evaporation climatology of Yu and Weller (2007) and satellite-based precipitation estimates from the Global Precipitation Climatology Program (GPCP; available at: <http://precip.gsfc.nasa.gov/>). (b) Average surface salinity of the world ocean, contoured from the World Ocean Database of NODC (<http://www.nodc.noaa.gov/OC5/SELECT/dbsearch/dbsearch.html>).

ocean (Figure 1b). Higher salinities are found where water vapor leaves the surface, and lower salinities are noted where precipitation and continental runoff dilute the surface ocean. Significant differences in ocean-basin salinities are also apparent, in particular, the higher salinities of the Atlantic relative to the Pacific. In part, this difference is due to the narrowness of the Atlantic; a greater fraction of its surface area is under the influence of dry continental air. But, the arrangement of the continents also plays a role: moisture is easily transported to the Pacific across the Central American

isthmus, but there is little transport from the Indian Ocean to the Atlantic, and no significant moisture transport from the Sahara Desert. A central goal of expanding salinity measurements is to develop a better understanding of what these salinity patterns are telling us about the water cycle.

#### THE OCEAN IS THE MAIN RETURN PATH FOR THE WATER CYCLE

In addition to being the main reservoir for heat and water in the climate system, the ocean is the primary return path for

water transported by the atmosphere. That is, water evaporated from mid-gyre must be returned from the high-precipitation zones, or we would observe changes in regional sea level of several meters per year. River flows are too small and too limited in geographical range to contribute much to water cycling on the global scale (Schmitt, 1995). Integration of surface-water flux estimates provides estimates of freshwater flux divergences within the ocean that can be compared with fluxes derived from oceanic sections (Wijffels, 2001). With the large differences in net water balance in the

oceans (the Pacific Ocean gaining 0.5 Sv, the Atlantic Ocean losing over 1 Sv), it is clear that significant volumes of water must also be redistributed between the various basins. This water movement is accomplished by the interbasin connections of the Bering Strait, the Indonesian Throughflow, and the Southern Ocean. The water cycle plays a role in driving these flows; the North Pacific is fresher and stands ~ 50-cm higher than the North Atlantic, thereby providing a pressure gradient to drive the Bering Strait's flow. Associated with this water flow is salt transport around the Americas, a fact that is not often recognized. Wijffels et al. (1992) discuss the caveats in using locally referenced salinities to define freshwater anomalies. This approach can be valid for regional budgets but cannot be used to describe the global transports of water and salt.

### THE IMPACT OF THE WATER CYCLE ON OCEAN CIRCULATION

In addition to the mass-conserving flows induced by evaporation and precipitation patterns, the distribution of surface fluxes shown in Figure 1 also dictates a barotropic circulation response in the ocean. Just as surface Ekman convergences drive vorticity-conserving Sverdrup circulations on the spherical Earth (the classical wind-driven gyres of Stommel, 1948, and Munk, 1950), the loss or gain of water at the ocean surface drives equivalent "Goldsbrough" circulations that can be ~ 10% of the wind-driven flows (Huang and Schmitt, 1993). For the North Atlantic, the mass-conserving and dynamically driven flows lead to a *southward* western boundary current of about 2 Sv in mid-latitude. Huang

and Schmitt suggest that such counter flows can affect the separation latitude of wind-driven western boundary currents.

The gain or loss of water from the sea surface has a direct influence on upper-ocean salinity and thus density. The density flux due to E-P can be expressed as:

$$\beta F_s = \frac{\beta S \rho_0 (E - P)}{\rho_s (1 - S)} \equiv \beta S (E - P), \quad (1)$$

where  $\beta = 1 / \rho (\partial \rho / \partial S)$  is the haline contraction coefficient and  $\rho_0$  and  $\rho_s$  are the densities of pure water and seawater, respectively. Schmitt et al. (1989) computed this density flux and that due to heat exchange with the atmosphere, and found them comparable in many regions. The general pattern is one of thermal buoyancy-flux dominance in high and low latitudes, but nearly equal thermal and haline buoyancy forcing in mid-latitudes. Except in the tropics, the thermal and haline buoyancy fluxes are opposing one another in their effects on surface density. This open-ocean buoyancy pattern is modified by continental runoff so that the highest and lowest latitudes can have regional areas of haline buoyancy-flux dominance

(Schmitt, 1998). Because some of these high-latitude coastal regions are the sites of deep-water formation, thermohaline circulation may be particularly sensitive to such freshwater inputs. Indeed, a vast array of paleoclimate studies are quite

dependent on this relationship, with a comparably large paleo-modeling effort devoted to scenarios in which large pulses of freshwater from melting ice flood the North Atlantic and disrupt deep water formation (Alley et al., 2003). The details of how such water entered the ocean, whether it was confined to coastal regions as dynamics would suggest, and how it actually mixes, are crucial points that are not well addressed in the current generation of models.

### SALINITY AND MIXING

When surface salinity increases due to evaporation, the increased surface density contributes to a destabilization of upper-ocean stratification. Though the total buoyancy flux also depends on the sign and magnitude of the heat flux, surface "salinification" will generally contribute to deeper convective mixing of surface waters. Conversely, freshwater inputs due to rainfall or river runoff will decrease surface salinity, make a positive contribution to the stratification, and generally inhibit mixing.

These effects, of course, dominate in different geographical regimes. High

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evaporation rates, deep winter mixed layers, and high salinity are found in the eastern and central subtropical gyres. The winter surface-salinity maximum is at the surface in such regions. Thus, the gradients are favorable for double-

diffusive salt-finger mixing in the upper ocean. Potential energy in the unstable distribution of salt is released by faster heat diffusion at the centimeter scale, which generates vertically advecting salt fingers. These salt fingers enhance vertical salt flux over heat flux and serve to directly modify the temperature and salinity structure (St. Laurent and Schmitt, 1999; Johnson, 2006). Schmitt (1999) proposes that the action of salt fingers on upper-ocean temperature and salinity anomalies generated by varying freshwater exchanges with the atmosphere is an essential cause of the tight temperature-salinity relationship in the Central Waters.

Where an excess of freshwater input is decreasing salinity and stabilizing the upper ocean, mixing by mechanical turbulence is inhibited. Sufficiently large freshwater inputs can lead to the formation of “barrier layers” in which a strong halocline inhibits vertical exchange between the mixed layer and the ther-

Pacific halocline resists deep convection in the winter. The deeper surface mixed layers of the salty Atlantic also provide a different response to inertial-period wind forcing; the thinner mixed layers of the North Pacific are more easily accelerated to higher speeds, but the North Atlantic displays a more energetic inertial response due to the thicker mixed layers in winter (Park et al., 2005). Such salinity-controlled differences in mixed-layer depths due to the patterns of surface water fluxes may be providing an unexpected regional modulation of inertial wave energy propagation, dissipation, and mixing in the thermocline.

Much has been made of the barrier-layer effect on the North Atlantic’s deep-convection regions where bottom water is formed. Alley et al. (2003) hypothesize that large freshwater discharges from the breaking of glacial melt-water dams at the end of the last ice age capped off deep-water-formation sites and disrupted North Atlantic thermohaline

inferred as a mechanism for abrupt climate change (Stommel 1961; Broecker, 1987), but ocean-mixing processes and freshwater-discharge mechanisms remain poorly understood. There are compelling reasons to better understand these processes, as global warming may be enhancing the water cycle and contributing to high-latitude freshening.

## GLOBAL WARMING AND THE WATER CYCLE

A strong, nonlinear dependence of water’s vapor pressure on temperature (Clausius-Claperyon relation) drives concern over the water cycle in a warming climate (Figure 2). At the current global average temperature of about 14°C, the vapor pressure increases by about 7% per degree C. This dependence drives the positive evaporation/sea-surface temperature correlation that provides the oceanic source of rain for many coastal areas, and it has provided a precipitation-forecasting tool in some regions. But, a 7% increase in the intensity of the water cycle per degree of warming is of serious concern for global change. It portends changes in water supplies that would not be easily accommodated by society.

Until ocean salinity changes began to reveal trends, evidence for a changing global water cycle was decidedly mixed. Terrestrial pan-evaporation measurements show an unexpected decrease in evaporation with warming since the 1970s. This paradox is apparently explained by decreasing solar radiation due to manmade aerosols (Roderick and Farquhar, 2002). Both industrial pollution and high-altitude jet contrails may play a role in decreasing the amount of energy reaching the surface. New evidence suggests this effect may be

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mocline. Lukas and Lindstrom (1991) describe the barrier-layer phenomena in the tropical Pacific where the ITCZ is the freshwater source. The tropical Atlantic also has barrier layers, where waters from the Amazon River outflow play a prominent role (Hu et al., 2003). There is more rainfall in the North Pacific than in the North Atlantic; thus, the North

circulation. Evidence for such events is found in the paleorecord in sediments from ocean cores; the type and isotopic composition of foraminifera varies with water-mass temperature and salinity. Apparent disruptions in ocean circulation are associated with the timing of freshwater discharges. Thermohaline circulation instability has long been

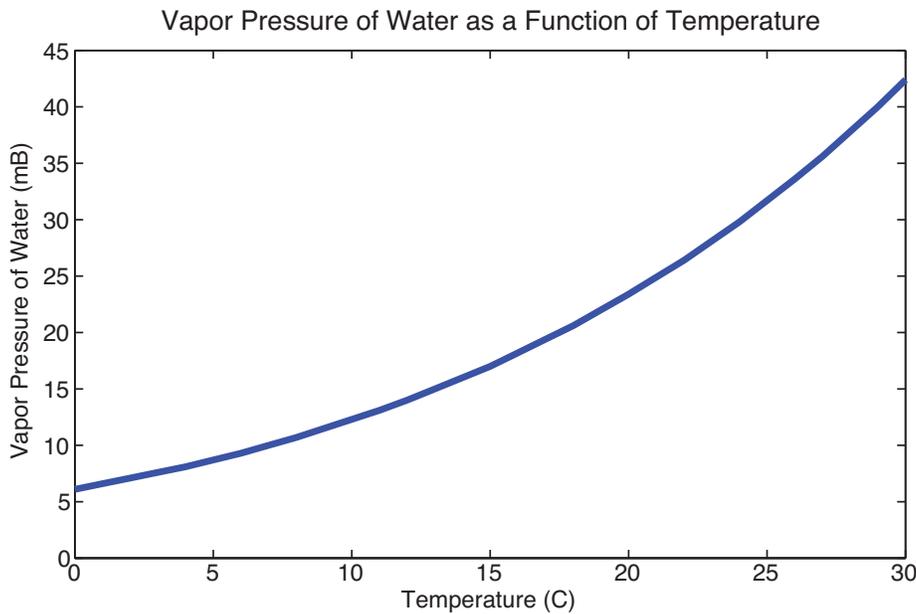


Figure 2. The vapor pressure of water as a function of temperature. The vapor pressure (in millibars) is a strong nonlinear function of temperature. This means that warm air can hold much more water than cold air. At the global mean surface temperature of about 14°C, the vapor pressure increases about 7% per degree C. Thus, strong changes in the water cycle are anticipated with global warming.

primarily terrestrial; satellite estimates of surface radiation suggest an opposite trend over the ocean (Pinker et al., 2005). There, the oceanic salinity record provides what is arguably the best evidence for an accelerating water cycle. Salinity contrasts have been increasing, as documented in several of the papers in this issue as well as a host of other analyses in the Atlantic (Brewer et al., 1983; Curry et al., 2003), Pacific (Freeland et al., 1997), and global oceans (Boyer et al., 2005).

New estimates of oceanic evaporation by Yu (2007) may explain these salinity trends. She finds that the globally averaged annual evaporation rate has had a consistent upward trend since the late 1970s. These improved latent heat-flux estimates show interannual variability and longer-term changes, such as the decline in evaporation up till the late

1970s and a stronger increase in evaporation since then. This trend was largely driven by changes in winds. The change over the last 30 years is approximately 10 cm/yr, which is equivalent to an additional 1.2 Sv of global evaporation. It is hardly surprising, then, that we see significant trends in salinity. Whether the available oceanic precipitation estimates are sufficient to address such trends remains to be seen. Confidence in such estimates is still weak. We must expect that evaporation and precipitation over the ocean is likely to remain poorly known quantities for some time to come. However, upper-ocean salinity is very sensitive to their difference, and this net exchange is the quantity of interest. Rather than relying on the difference between large and uncertain numbers, we would do better to monitor and use salinity as a first-order indicator of the

water cycle's strength. This technique is the oceanographer's way of using the ocean as a rain gauge. New satellite (Lagerloef et al., this issue) and in situ (Riser et al., this issue) technologies promise to make expanded salinity measurements possible, and we must fully exploit such data for improved understanding of the global water cycle.

### THE POTENTIAL OF SALINITY BUDGETING

If we assume that an improving salinity database will be obtained over the next few years, we can look to the oceanic conservation equations to see what can be learned. Key technical issues include representation of diapycnal mixing, and application of the conservation of mass with water entering and leaving the ocean surface. These processes are illustrated by examining the role of salinity within the mixed layer. Integrated vertically through a mixed layer of depth  $h(x, y, t)$ , the time rate of change of mixed layer salinity ( $[S]$ ) is determined by a combination of surface flux, horizontal advection, subsurface processes, and mixing:

$$\frac{\partial[S]}{\partial t} = \frac{(E - P)S}{h} - \bar{v} \cdot \nabla_h S + \text{subsurface} + [ML\text{mixing}] \quad (2)$$

$$\text{subsurface} = \alpha \{ [S] - S_{-h} \} \frac{\partial h}{\partial t} - [w \nabla_z S] + [\nabla_z (k \nabla_z S)]_{z=-h}$$

Here, the subsurface processes include mixed-layer deepening and entraining (if  $\alpha$  is positive and zero otherwise). The distribution of annually averaged surface water flux closely resembles the salinity (Figure 1), with high salinities in the subtropics and lower salinities in the tropics and subpolar regions. In

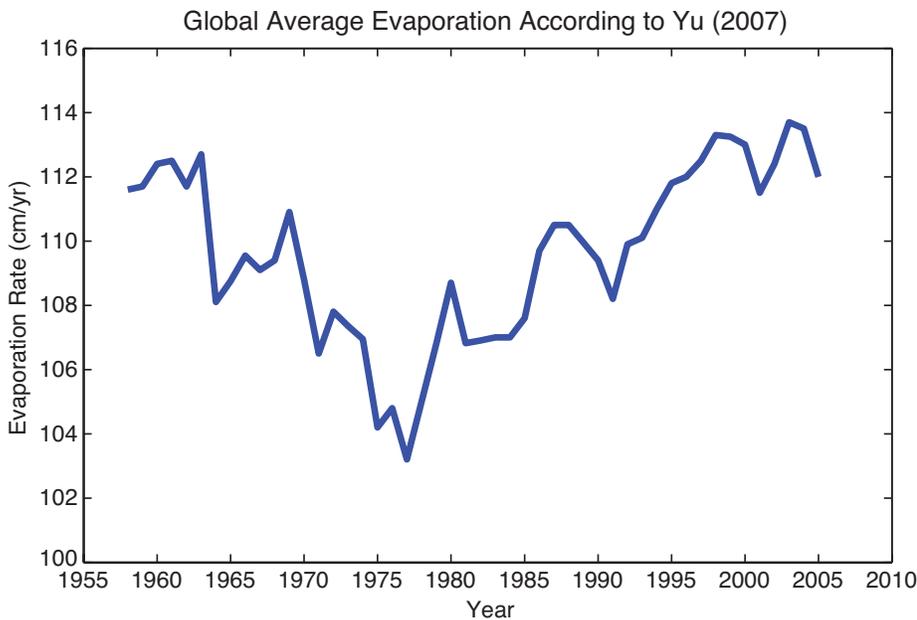


Figure 3. Global average annual evaporation rate from the ocean from 1958 to 2005 according to the Yu (2007) climatology, in cm/yr. A downward trend was operative in the 1960s, but the average evaporation has increased from 103 cm/yr to nearly 114 cm/yr since 1976.

some regions, evidence for advection is apparent, for instance, the surface salinity maximum in the North Atlantic is north of the maximum in surface-water loss, which results from poleward advection in the Ekman layer under the trade winds. Incorporation of salinity data into ocean-circulation models promises to provide an independent means of estimating evaporation minus precipitation over the ocean. These ideas are more fully developed in the final report of the US CLIVAR Salinity Working Group (2007).

#### THIS ISSUE OF OCEANOGRAPHY

Other articles in this issue of *Oceanography* provide an in-depth perspective on the value of salinity measurements. *Gordon and Giulivi* highlight the basic relationship between the water cycle and upper-ocean salinity, and contrast the North Atlantic and North Pacific. They

exploit the long-time-series station at Bermuda to examine salinity changes over the past 50 years, which reveal intriguing interannual and decadal signals suggestive of atmospheric teleconnections between the two basins. Oceanographers' traditional focus

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on regional water masses is used by *Yashayaev and Clarke* to show the strong salinity changes found throughout the Labrador Sea's water column in response

to changing surface forcing over the past six decades. Their work again highlights the value of long time series based on weather ships and repeat hydrographic lines and shows how quickly salinity can change in the deep waters that feed thermohaline circulation. *Lukas and Santiago-Mandujano* use the long-time-series station offshore Hawaii to examine the relationship between salinity changes and variations in surface freshwater fluxes in the North Pacific. They show how hydrographic stations can be used like paleostratigraphic cores because deeper salinity layers record the surface freshwater forcing extant when they were subducted into the thermocline. *Riser, Ren, and Wong* show the rapidly expanding ability of the Argo float program to document oceanic salinity variability on a global scale. The remarkable success of the Argo salinity program provides key evidence that our in situ sensor technology has matured to the point that we can now be much more ambitious in our planning for oceanographic studies of the water cycle. *Lagerloef et al.* describe the technology behind the coming Aquarius satellite that will, for

the first time, provide a remote-sensing capability for surface salinity. The launch of Aquarius in 2010 and the rapidly improving in situ salinity database will

provide an opportunity for oceanographers to finally obtain key measurements of the largest element of the global water cycle. The basic recommendations of the *CLIVAR Salinity Working Group* for advancing such approaches are given in the “What’s Next for Salinity?” article. We hope that these articles inspire a new appreciation for the salt in our favorite planetary fluid. The ocean represents the world’s largest possible rain gauge and it is time that we learned to understand what it is telling us.

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