The Goldsborough–Stommel Circulation of the World Oceans *

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ABSTRACT

Goldsborough first showed how the mass flux at the ocean surface due to the difference between evaporation and precipitation could induce barotropic flow in the ocean interior through the requirement of vorticity conservation. Stommel proposed to close this circulation by adding the western boundary currents. Here, a first-order description of the Goldsborough–Stommel circulation for the world oceans is presented, using available climatologies. While such flows are an order of magnitude smaller than the wind-driven circulation, the interaction between the Goldsborough–Stommel gyres and the wind-driven and thermally driven circulation determines the salinity distribution of the world oceans. Therefore, it is important to study the Goldsborough–Stommel circulation and its interaction with motions driven by other forcings. In addition, the western boundary currents required to close the Goldsborough interior and to satisfy interbasin mass transport can be substantial. In the Atlantic the southward western boundary current reaches two Sverdrups \( (Sv = 10^5 \text{ m}^2 \text{ s}^{-1}) \) at 35°N. It is suggested that this adverse current causes a southward shift in the separation point of the Gulf Stream; a simple model indicates that the displacement is about 75 km.

1. Introduction

Goldsborough (1933) was the first to understand how surface mass fluxes could generate steady barotropic flows in the ocean interior on a rotating planet. He attempted to model the subtropical gyre of the North Atlantic with an unrealistic distribution of precipitation in the ocean interior and evaporation at the western edge. The unlikely forcing required and the weakness of the circulation caused the work to be largely ignored. It does, however, presage the contributions of Sverdrup (1947) and Stommel (1948), who appreciated that convergence of Ekman transport was the driving force for the geostrophic interior. Stommel (1957) pointed out that the Goldsborough circulation driven by an arbitrary evaporation minus precipitation pattern in the ocean interior can be closed by adding western boundary currents. Thus, we will refer to the coupled interior and boundary currents generated by surface water fluxes as the Goldsborough–Stommel circulation. Stommel (1984) acknowledges an intellectual debt to Goldsborough, but as far as we know, no one has actually calculated the Goldsborough–Stommel circulation with realistic evaporation and precipitation distributions.

It is timely to examine this issue, given increasing interest in the thermohaline circulation, for while the Goldsborough–Stommel gyres are relatively weak, the interaction of the Goldsborough–Stommel circulation with the wind-driven and thermally driven circulations acts to control the thermohaline circulation in the world oceans. All existing numerical models of oceanic general circulation have treated only the buoyancy flux due to surface water exchange. Usually, a zero vertical velocity is assumed at the upper surface and a fictitious salt flux across the air–sea interface is applied to generate the baroclinic circulations due to the evaporation minus precipitation. Thus, the Goldsborough–Stommel circulation is excluded for all these models. We hope this work will stimulate the development of models that use the surface water fluxes as the real boundary conditions for the model, so that the barotropic Goldsborough–Stommel circulations and the baroclinic circulation associated with the freshwater forcing can be reproduced accurately.

An obvious impediment to this calculation is the lack of definitive data on evaporation and precipitation \( (E - P) \) over the oceans. For this first-order look at the global Goldsborough–Stommel circulation we have utilized the climatology of zonally averaged \( (E - P) \) of Baumgartner and Reichel (1975, hereafter BR). In addition, we contrast the BR estimates with those of Schmitt et al. (1989, hereafter SBD) and the combination of Dorman and Bourke (1981) for precipitation and Ismer and Hasse (1987) for evaporation in the North Atlantic. While differences exist between these datasets, the patterns and amplitudes are sufficiently similar that most major features, such as the western boundary transport, are well defined for this basin.

Western boundary transports are of particular in-
We require that the sum of the western boundary and interior flows equal the mass flux determined by interbasin transport and net $E - P$ forcing in order to be consistent with the global water transport scheme recently presented by Wijffels et al. (1992). This is necessary because the meridional velocity in the ocean interior is determined by the wind stress curl and $E - P$, and excess transport must be confined to the western boundary, as in Stommel and Arons (1960).

For the North Atlantic, the western boundary transport induced by freshwater forcing is largely southward, due to the flux required by the Bering Strait throughflow and high-latitude precipitation and runoff. It reaches 2 Sv ($Sv = 10^6$ m$^3$ s$^{-1}$) at 35°N. We suggest that this adverse flow may shift the separation latitude of the Gulf Stream southward, thus explaining a discrepancy between the observed and modeled Gulf Stream. A simple model of a wind-driven gyre and an adverse boundary current is presented; it indicates that a displacement of about 75 km is to be expected for the Gulf Stream separation point.

The global Goldsbrough–Stommel circulation is presented in the next section. The impact of interbasin transports is discussed in section 3. Variations to be expected from alternative datasets for the North Atlantic are presented in section 4. Finally, a simple model of western boundary current separation is given in section 5.

2. Global Goldsbrough–Stommel circulation

In the interior ocean the motion is governed by the geostrophic relation plus the wind stress on the surface, so that the basic equations are

\[ -fV = -p_x + r^z, \]  
\[ fU = -p_y + r^z, \]  
\[ U_x + V_y = -(E - P), \]

where $U$ and $V$ are the vertically integrated mass fluxes, and $(E - P)$ is evaporation minus precipitation. By cross differentiating (1) and (2), one obtains the vorticity equation

\[ \beta V = \mathbf{k} \cdot \nabla \times \tau + f(E - P). \]  

This equation can be called the Goldsbrough–Sverdrup relation, which represents the lowest-order vorticity balance in the ocean interior. The first term on the right-hand side is the well-known wind stress curl. The second term indicates the contribution due to evaporation and precipitation. In order to calculate the zonal velocity in a closed basin, a meridional boundary must be chosen as the starting point. According to the well-known vorticity argument, there is no eastern boundary current; instead there should be a western boundary current that serves to close the loop of mass flux and the global vorticity balance. Thus, we have adapted the boundary condition of no zonal flow along the eastern boundary for starting the zonal velocity calculation.

Since the wind-driven circulation has been discussed in many papers, we will concentrate on the horizontal velocity pattern forced by the evaporation and precipitation alone. To develop a global picture we use the zonal averages for 5° latitude bands of Baumgartner and Reichel (1975) in each basin (Fig. 1). Using the eastern boundary condition, one can calculate the horizontal velocity in the ocean interior forced by evap-

![Fig. 1. The distribution of zonally averaged water flux into the oceans as a result of precipitation, evaporation, and river discharge.](image-url)
oration and precipitation as well as the western boundary current transport required by mass conservation within the basin. The results of the calculation are presented in Fig. 2, which is a composite map of the Goldsborough–Stommel circulation in the World Ocean, including the interior flow and the western boundary currents in the three main basins. Minor deviations would be expected from this scheme if zonal variations were introduced. However, we lack appropriate data for most of the globe and since the interior flows are small, such detail would be of little practical value. The western boundary currents are properly estimated from the zonal averages alone.

In this calculation, the river runoff is assumed to be uniformly distributed at the latitude of the river mouth. In addition, the water exchange through the Strait of Gibraltar is treated as if it was evaporation over a 5° latitude band between 35° and 40°N. Strictly speaking, the river runoff and the Mediterranean flow exchange should be treated as real water flux at the lateral boundaries of the basin. This is a significant error only for the Amazon, but since it discharges at the equator, it has no impact on the Goldsborough interior circulation.

In the Atlantic Ocean strong evaporation drives a poleward flow in the subtropical basin, and precipitation at high latitude drives an equatorward flow in the subpolar basin (Fig. 2a). The interior meridional transport is proportional to $f/\beta$; thus, it is zero near the equator and relatively large at high latitude. Compared with the mass flux due to the wind-driven circulation, the barotropic velocity due to the evaporation and precipitation is fairly small. For most cases, the mass flux through a $5^\circ \times 5^\circ$ box is no more than 0.5 Sv. Nevertheless, the freshwater flux across the air–sea interface controls the salinity balance in the world oceans. Considering the fundamental role of the thermohaline circulation in the climate system, it is very important to calculate the freshwater flux accurately.

The mass flux of the Stommel western boundary current is determined by three requirements: 1) to balance the interior flow given by the Goldsborough relation, 2) to balance the evaporation and precipitation within each basin, and 3) to balance the water flux between basins (Wijffels et al. 1992). However, in order to isolate the effects of local freshwater forcing we defer discussion of interbasin transports to a later section and here assume no mass flux through the northern boundary of each basin. In the Atlantic, the western boundary current forced by the interior Goldsborough circulation is northward at high latitudes (Fig. 3a), due to southward flow in the interior forced by precipitation excess. In the subtropical gyre the western boundary current flows southward in order to balance the poleward flow in the interior driven by net evaporation associated with formation of 18°C water near the Gulf Stream and salinity maximum water in midgyre as well as water lost to the Mediterranean. At 35°N, the southward flux of the western boundary current reaches 1 Sv. This adverse current may have an impact on the Gulf Stream separation latitude, which will be discussed in section 5.

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**Fig. 2.** The Goldsborough–Stommel circulation of the world oceans. Each arrow indicates the horizontal mass flux integrated over a $5^\circ \times 5^\circ$ box, in Sverdrups. Along the western boundary of each basin, there is a curve indicating the northward mass flux ($10^8$ m$^3$ s$^{-1}$) within the western boundary, which is required to close the circulation. In this calculation, the interbasin transports are neglected.
In the Pacific Ocean (Fig. 3b) strong precipitation drives equatorward flow in the high-latitude interior, except the subtropical basin where evaporation drives relatively weak poleward flow. The Stommel western boundary current is northward north of 40°N. It exceeds 1 Sv between 47° and 60°N. The Stommel western boundary current flows southward south of 40°N and reaches 1 Sv near 25°N. It is important to note that we have not included any transport in the South Pacific to accommodate the flow from the Pacific to the Indian Ocean through the Indonesian passages. As noted by Wijffels et al. (1992), this transport, possibly as large as 20 Sv, will dominate mass fluxes in the South Pacific and Indian basins. Once the size of this transport is more clearly established, it can be added to the net and western boundary transports; the Goldsbrough interior flows of course depend only on $E - P$.

The Indian Ocean north of 10°S has very weak net Goldsbrough–Stommel circulation, largely because of offsetting zonal variations in $E - P$ (Fig. 3c). In the southern subtropical gyre there is a strong (1 Sv) poleward interior flow forced by evaporation, and a strong equatorward boundary current. Again, we have neglected the Pacific–Indian throughflow. If present this transport would have to be accommodated as a westward flow on the equator, since cross-equatorial flow is not permitted except at the western boundary in linear Sverdrup–Goldsbrough dynamics. The western boundary current would then carry the throughflow transport to the south.

3. Interbasin transports

As pointed out by Wijffels et al. (1992), the transport between basins must be considered when developing global water and salt budgets. The impact of the poorly known Pacific–Indian throughflow on the western boundary currents of the South Pacific and Indian oceans has already been discussed. For the North Pacific and Atlantic oceans we have better information on the proper northern boundary conditions for the net water transport because of the Coachman and Aagaard (1988) estimate of the flow through Bering Strait. Driven in part by the geopotential height difference between the fresher North Pacific and the saltier Atlantic, the Bering Strait flow is an important component of the global hydrologic cycle. In fact, nearly all of the precipitation excess of the North Pacific is ex-
ported through Bering Strait, carrying with it a significant salt flux as well. The northern boundary condition for the Pacific is a net northward transport of 0.8 Sv. The boundary condition for the Atlantic at 65°N is a net southward flow of 0.92 Sv, representing the sum of Bering Strait throughflow and the net mass added in the Arctic and high latitude. While the precipitation distribution in the Arctic will induce meridional motion there, only the net meridional transport is necessary for computation of flow patterns to the south. The corrected Atlantic and Pacific transports are shown in Figs. 4a and 4b. We now find that the western boundary current in the Atlantic is southward at most latitudes, reaching 2 Sv at 35°N. It only becomes northward south of 20°S. In the Pacific, the western boundary current is nearly everywhere northward, exceeding 2 Sv between 50° and 60°N and exceeding 1 Sv at 10°N.

4. North Atlantic variations

Since surface freshwater fluxes are a very poorly known ocean forcing function, one is justified in questioning the representativeness of the foregoing calculations. Here we compare results from three different climatologies for the North Atlantic, where a better database enables a closer examination of the hydrologic forcing. In addition to BR, we utilize data from SBD, and the combination of precipitation from Dorman and Bourke (1981) and evaporation from Ismer and Hasse (1987, IH1). Schmitt et al. used the evaporation estimates of Bunker (1976); Ismer and Hasse revised those estimates to account for a decreased exchange coefficient in the bulk formula and increased wind speeds due to a recalibration of the Beaufort wind scale. The IH data have 30–40 cm yr⁻¹ greater evaporation than Bunker south of 40°N. The zonal averaged freshwater input, including runoff from BR, for the three cases is shown in Fig. 5a. Differences between the datasets can reach 0.07 Sv for 5° bands in low to mid-latitudes, about 30% of the signal at 20°N. The interior Goldsborough transports (Fig. 5b) show comparable variations, with increasing sensitivity at high latitudes due to the \( f/\beta \) dependence. Poleward flow of around 1 Sv is predicted between 20°N and 40°N; southward flow of around 0.7 Sv is estimated north of 50°N. While there are differences between the transports, there is a general agreement in overall patterns and amplitudes.

The total meridional water transport (Fig. 5c) is proportionally most sensitive to the differences in the datasets. The extra evaporation in the IH data leads to substantial differences when integrated from the north. We note also that these curves are all started at 65°N at a flux value determined from the Bering Strait throughflow (Coachman and Aagaard 1988) and BR’s estimates of the net hydrologic forcing in the Arctic and high latitude Atlantic. These carry an uncertainty of at least 0.1 Sv.

In contrast, the western boundary currents estimated from the three datasets are rather less sensitive to their differences (Fig. 5d). This is because the extra Stommel boundary flow forced by increased evaporation is somewhat offset by the decreased net southward mass transport required. The boundary current is everywhere southward, though mostly less than 0.5 Sv north of 50°N and south of 10°S. It is between 1.7 and 2.0 Sv at 35°N. Thus, while there remain significant uncertainties in our knowledge of the hydrologic cycle over the ocean, the general agreement among these datasets for the North Atlantic suggests that the global Goldsborough–Stommel circulation presented here is a reasonabably first-order description.

We have also included a case of no flow through the Bering Strait, depicted as the marked thin line in Fig. 5d. The northward mass flux in the western boundary
current increases 0.9 Sv, so it is poleward North of 46°N; meanwhile, the maximum southward flux at 35°N is reduced to 1 Sv. Because the sill of the Bering Strait is only about 50 m deep, the water flux through the strait could fluctuate greatly with changing sea level. The climatic impact of variable mass flux through Bering Strait such as occurred during the last glaciation is an issue worth close examination.

5. Western boundary current separation

The Stommel western boundary currents may affect the separation latitude of the western boundary currents of the wind-driven circulation. In many numerical models simulating the North Atlantic, the Gulf Stream tends to separate from the coast north of the latitude determined from observations. We speculate that the missing Goldsbrough–Stommel gyres and the flow through the Bering Strait may be contributing to this separation problem. The first point is quite simple, and can be seen by reference to Eq. (4). In linear Sverdrup dynamics the boundary current separation is realized at the latitude of zero wind stress curl. Reasonable agreement between the observed Gulf Stream separation and the estimated zero of the wind stress curl can be seen in the maps of Lectma and Bunker (1978). Thus, inclusion of the term $f(E - P)$ in the full Goldsbrough–Sverdrup relation will shift the position of the zero of Eq. (4). In the case of the Gulf Stream, the midlatitude evaporation will tend to shift the separation point southward. However, depending on the scales and intensities chosen for the wind stress curl field, this effect may be only a few tens of kilometers.
Perhaps more important is the total adverse western boundary current realized because of mass conservation requirements determined by the Bering Strait throughflow and high-latitude precipitation in addition to the Goldsborough–Stommel flow. To illustrate this effect we construct a simple model using linear Sverdrup dynamics for the wind-driven circulation and add an adverse western boundary current. The two-gyre wind-driven circulation is forced by a simple sine profile of Ekman pumping velocity

\[ \nu_c = -1.0 \times 10^{-4} \sin(2\pi y/y_l), \]  

where \( y_l = 6000 \) km is the north–south width of the basin. The vertically integrated streamfunction consists of an interior solution matched with a Stommel boundary layer along the western boundary

\[ \Psi_w = \frac{f \nu_c}{\beta}(x_e - x)(1 - e^{-x/\epsilon x_e}), \]  

where \( x_e = 6000 \) km is the eastern boundary of the basin, \( \epsilon = 0.05 \) is the scale width of the western boundary current. In the case of purely wind-driven circulation, the western boundary currents separate from the coast at the latitude of zero Ekman pumping (Fig. 6a).

In the North Atlantic the Goldsborough–Stommel gyre and mass conservation requires a southward western boundary current of 2 Sv at 35°N. When this interacts with the boundary current of the wind-driven circulation, the separation latitude is pushed southward. We have assumed a similar form for the western boundary current of the Goldsborough–Stommel gyres

\[ \Psi_G = \Psi_0 e^{-x/\epsilon x_e}, \]  

where \( \Psi_0 = 2 \) Sv is the maximum mass flux of the western boundary current associated with the Golds-

brough–Stommel gyre and interbasin mass flux. Due to the interaction between the two boundary currents, the separation streamline is moved 75 km southward (Fig. 6b). Significant adverse boundary currents are also predicted for the subtropical gyre of the Indian Ocean and the subpolar gyre of the North Pacific, so we can expect similar effects in those oceans.

Western boundary current separation is a complex phenomenon, and effects of wind stress curl, coastline shape, topography, isopycnal outcropping, inertial terms, and diabatic processes should all be considered. The foregoing simple model is meant to be merely illustrative of how the southward western boundary current due to the Goldsborough–Stommel circulation and global mass conservation may contribute to the separation. This issue should be verified using new numerical models, in which the surface mass fluxes and interbasin transports are correctly implemented.

6. Summary

Given the sensitivity of the thermohaline circulation to freshwater forcing, it is important to improve estimates of ocean water flux and models of its effects. First, a more complete calculation of the Goldsborough–Stommel circulation would include zonal variations in evaporation minus precipitation, realistic basin geometry, and individual treatment of river and marginal sea transports at the boundaries, and we are working to develop the database for such calculations. Second, instead of a fictitious salt flux, the real water flux should be used in numerical models in order to simulate the hydrologic cycle more accurately. Third, while small compared with the Sverdrup circulation, the Goldsborough–Stommel gyres are often in opposition to wind-driven flows and may have detectable consequences.

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**Fig. 6.** Interaction between the western boundary currents of the Sverdrup gyres and the Goldsborough–Stommel gyres, streamfunction in Sverdrups. (a) The western boundary current separation of the purely wind-driven two-gyre circulation. (b) The western boundary separation is displaced to the south by the southward western boundary current required for the Goldsborough–Stommel circulation and mass conservation.
for key issues such as the latitude of western boundary current separation.

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