On the Seasonal Transports of Freshwater and Salt in the Tropical Atlantic Ocean

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열대 대서양에서의 계절별 담수 및 염분의 수송

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요 약

열대 대서양의 해표면으로부터 500m 깊이까지의 계절별 담수 및 염분의 수송량과 방향을 조사하기 위하여, 먼저 연속방정식과 염분보존방정식에서 각각 이론적인 계산방법을 유도하고 관측자료를 사용하여 수송값들을 얻었다.

담수(강수량, 증발량, 그리고 강물방출량)의 기후자료에서 계산된 표면의 담수발산장을 북쪽에서 남쪽방향으로 적분하여 담수수송값을 얻었다. 연간 담수수송량의 방향은 북향으로 수송량은 적도부근에서 가장 적었고 (0 Sv), 12°N와 20°N부근에서 가장 많았다(0.3 Sv). 계절별 수송량의 크기는 북향 1.35 Sv과 남향 0.45 Sv의 사이의 값을 보였고 강한 북향 수송량이 여름과 가을사이의 전이기간에서 나타났다. 이러한 계절변화는 지역적인 담수저장의 변화 뿐만 아니라 열대 수렴대의 이동과 관련이 있었다.

1900~86 기간의 염분관측값을 객관분석(objective analysis)한 후, 연간 및 계절별 염분수송값을 계산하는데 사용하였다. 수평이류에 의한 염분수송량이 수평확산에 의한 수송량과 균형을 이룸으로 해서 연간 수송량은 없었다. 단지 수평확산에 의한 염분수송량은 북향이었고 15° N에서 최대값(5×10^6kg/s)을 보였다. 계절별 수송량의 크기는 북향 30×10^6kg/s과 남향 35×10^6kg/s사이의 범위내에 있었다. 계절별 수송방향은 여름과 가을사이의 기간을 제외하고 다른 기간에서 북향이었다.

Abstract

The transports of the seasonal freshwater and salt from surface to 500 m depth in the tropical Atlantic Ocean are derived from the equations of the continuity and saltconservation, respectively.

The freshwater transport is obtained by southward integration of the divergence of surface freshwater flux, using climatological freshwater (i. e. precipitation, evaporation, and river discharge) data. The annual freshwater transport is northward, ranging from 0 Sv near the equator to 0.3 Sv at 12° N and 20° S. The seasonal meridional transport amounts of freshwater range from 1.35 Sv to-0.45 Sv. The strong northward freshwater transports prevail for the intraseasonal period summer to fall. This seasonal cycle is caused by the shifts of the ITCZ as well as the changes in the local freshwater storage.

Annual and seasonal salt transports are calculated from objectively analyzed historical (1900-86) salinity observations. The annual salt flux in the ocean is zero, showing that the salt flux by horizontal advection balances the flux by horizontal diffusion. The salt flux due to the diffusion is northward, and has a maximum of 5 x 10^6kg/s at 15°N. Seasonal transport amounts of salt range from 30 x 10^6kg/s to -35 x 10^6kg/s. The direction of the seasonal salt transports is northward except for the intraseasonal period summer to fall.

1. Introduction

One of the important factors determining the oceanic water regime is the distribution mechanism of salts in the water. The salinity change at the sea surface affects the penetrative convection through its influence on the buoyancy flux. Additionally, the most significant effects in the mixer layer produced by the salinity budgets are the reduction of the deepening rate of the corresponding changes in the temperature of the mixed layer (Miller, 1976), and thus are important in the climate.

Although the mass of the salt in a basin is cnserved, it would be if there were a loss (gain) of the salt mass due to the input (removal) of freshwater at the surface. In other words, in spite of the conservation of salt mass, there is the change of salinity depending upon the freshwater input. Thus, the freshwater transports can be estimated from the salinity differences with time on the zonal section of the tropical Atlantic.

The first attempt to determine one monthly estimate of the meridional salt flux in the Atlantic Ocean was made by Bennett (1978) across 24°S with magnitudes ranging from -3 $\times 10^6$ kg/s to 2×10^6 kg/s during October 1958. Gordon and Piola (1983) suggest an annual mean northward salt flux of 700×10^6 kg/s over the 35° S-65°N region in their 'box' model with the annual freshwater forcing of Baumgartner and Reichel (1975), assuming a value for the North Atlantic Deep Water (NADW) production rate.

This paper presents the methods of computing the transports of seasonal freshwater and salt for the tropical Atlantic, including their transport estimates.

After showing how to compute the transport of seasonal freshwater and salt in section 2, we describe data and analysis in section 3. In sections 4 and 5, we describe the zonally averaged freshwater and salt transports, respectively. Here we would like to determine the relative importance of the seasonal local storage and horizonal divergence in balancing the freshwater and salt budgets.

2. Theory for the transports of freshwater and salt

We first derive the equations of the annual-mean and seasonal freshwater budgets and then the equations of the annual-mean and seasonal salt budgets, used for the calculations of their flux of freshwater meridional transports.

If we assume that the flux of freshwater (F) contributes an fractional amount of freshwater dF to a water column, the excess freshwater will reduce the column salinity with time. With the notation of vertical average,

$$\overline{\cdot} = \frac{1}{H}\int_{H}^{0} \cdot \; dz \tag{1}$$

applying salinity conservation principle within the column gives

$$\overline{\rho^{n} s'^{n}}\, dx\, dy\, H = \overline{\rho^{n+1} s'^{n+1}}\, dx\, dy\, (H + dz) \qquad (2)$$

where ρ_n is the density of sea water at time n, ρ_n^{+1} the density at time n+1, S'^n the salinity [$^0/_{00}$] at time n, S_n^{+1} the salinity at time n+1, the depth H = 500m, and $z=0$ the equilibrium position at the sea surface. We can rewrite (2) approximately as

$$\frac{1}{\overline{\rho s'}}\frac{d(\overline{\rho s'})}{dt} = -\frac{F}{H} \qquad (3)$$

where the flux of freshwater is $F=\frac{dz}{dt}$. Rearranging (3), we get the freshwater conservation equation as

$$-\frac{H}{\overline{\rho s'}}\frac{d(\overline{\rho s'})}{dt} = -\frac{H}{\overline{\rho s'}}\left(\frac{\partial(\overline{\rho s'})}{\partial t} + \nabla_h \cdot (\overline{\rho s' \vec{u}})\right) = F \qquad (4)$$

where $\nabla_h = \vec{i}\frac{\partial}{\partial x} + \vec{j}\frac{\partial}{\partial y}$ is the horizontal operator and $u = iu + jv$ is vertically averaged velocity vector. Assuming steady-state condition over times longer than a year, (4) reduces to

$$(F) = \left\{ -\frac{H}{\overline{\rho s'}} \nabla_h \cdot (\overline{\rho s' \vec{u}}) \right\} \qquad (5)$$

where braces are used to indicate annual-mean values. Defining the zonal and meridional integration as

$$<\cdot> = \int_{25^\circ N}^{\varphi} \int_{\lambda_1}^{\lambda_2} \cdot \; a \cos(\varphi)\, d\lambda\, d\varphi \qquad (6)$$

the annual northward freshwater transport (AFT) in the volume flux [m^3/s] is

$$AFT = AFT\,\big|_{25^\circ N} + <(F)> \qquad (7)$$

where a is the earth's radius, λ_1 and λ_2 are the longitudinal boundaries of the sides of the basin. The zonal and meridional integration of (7) is performed in one degree intervals. However, in Table 1 we show the values in 2.5 degree intervals for the comparison with the freshwater transports described in earlier studies. We choose 25° N to be the northern boundary of the integration because of the weak freshwater transport at this latitude (discussed later). As shown in (7), in short, annual meridional transport estimates of freshwater fluxes in this study are obtained by integrating the net exchange of freshwater across the air-sea interface of the oceans, Based on (5), the freshwater flux can be obtained in terms of the flux of salt mass, which has an opposite direction to that of freshwater.

Now we begin to examine seasonal freshwater budget by calculating the storage term of freshwater in (4). By evaluating the time integral of (4) from season 1 (one season) to season 2 (next season), we obtain approximately

$$H(\overline{\ln(\rho s')_1} - \overline{\ln(\rho s')_2}) + (\overline{\nabla_h \cdot (\ln(\rho s' \vec{u})_1)} - \overline{\nabla_h \cdot (\ln(\rho s' \vec{u})_2)})\, H\, \Delta T = (F_2 - F_1)\, \Delta T \qquad (8)$$

where $(\rho s')_1$ and $(\rho s')_2$ are the salt mass [10^{-3} kg/m^3] of season 1 and season 2, repectively, vertically integrated between the surface ($z=0$) and the depth ($z=-H$). F_1 [m/s] is the freshwater of season 1, and F_2 [m/s is the freshwater of season 2. ΔT is one season (i. e. three months). Dividing both sides in (8) by ΔT and rearranging the equation, we get

$$\Delta W = H(\overline{\nabla_h \cdot (\ln(\rho s' \vec{u})_1)} - \overline{\nabla_h \cdot (\ln(\rho s' \vec{u})_2)}) = \underbrace{\frac{H}{\Delta T}(\ln\overline{(\rho s')_2} - \ln\overline{(\rho s')_1})}_{A} + \underbrace{F_2 - F_1}_{B} \qquad (9)$$

where ΔW is the freshwater divergence in the units of volume flux per unit area. The term A is the seasonal change of freshwater storage, shich we can calculate from the difference between salt masses of one and the next season. This seasonal density is obtained from the Knudsen formula (Bryan and Cox, 1972) as a function of the seasonal temperature, salinity and depth (or pressure). The term B is the mean surface freshwater flux from season 1 to season 2. When integrated zonally and meridionally, the seasonal northward freshwater transport (SFT) volume flux [m^3/s] is :

$$SFT = SFT \mid_{25^\circ N} + < \Delta W > \qquad (10)$$

The use of the boundary value of 25° N in (10) is not likely to produce a large error for our calculation of the transport because the freshwater transport at this latitude is generally an order of magnitude less than that in other latitudes of the tropical Atlantic according to Hall and Bryden (1982), Baumgarther and Reichel (1975) and Schmitt *et al.* (1989) (Table 1). The good agreement of these independent estimates suggests that the seasonal cycle is small here as well.

Table 1. Annual zonal freshwater budget (P-E+R) in the tropical Atlantic Ocean and annual meridional freshwater transport using the boundary value at 25° degree latitude band. Units are $10^4 m^3 s^{-1}$.

Latitude	P-E+R	Freshwater Transport: This Study	Baumgartner & Reiches (1975)	Schmitt *et al.* (1989)	Hall & Bryden (1982)	Bennett (1978)
25° N	-2.9	-2.9	3.0	3.0	-2.9	
23.8~25	-2.4	-0.5				
22.5	-3.0	2.5	17.2	19.8		
20	-4.8	7.3				
17.5	-5.6	12.9	27.4	35.1		
15	-6.1	19.0				
12.5	-4.3	23.3	30.9	44.4		
10	-0.2	23.5				
7.5	3.4	20.1	21.0	33.2		
5	6.2	13.9				
2.5	8.3	5.6	10.7	10.1		
0	4.7	0.9				
2.5	1.2	-0.3	1.7			
5	-0.1	-0.2				
7.5	-2.7	2.5	8.2			
10	-5.3	7.8				
12.5	-6.2	14.0	21.0			
15	-5.6	19.6				
17.5	-4.5	24.1	33.4			
20	-3.4	27.5				
22.5	-4.5	32.0	44.4			
25° S						-6~10(24S)

Two alternative methods exist for determining terms in the freshwater balance. The first alternative is the direct method which balances weatern boundary volume transports and interior Sverdrup and Ekman transports through zonal section (e.g., Bennett, 1978; Hall and Bryden, 1982). Thus, a direct calculation of the freshwater fl-ux from oceanographic sections can be attempted on the basis of the salinity profiles together with the volume transport data. Estimates obtained by this direct method, can be used as an independent check of the freshwater transports deduced from integration of the net exchange of freshwater across the air-sea interface between lateral boundaries. How-ever, since Sverdrup mass transport is weak in the tropics (e.g., Leetmaa and Bunker, 1978), the direct method has never been applied to the freshwater transport in the oceans between 20° N and 20° S.

The second alternative, which has been tried by Fu (1986) for the South Indian Ocean, is to apply an inverse method to existing hydrographic sections. The results obtained from the inverse method, which has not been attempted in our study area, are constrained to be consistent with the conservation of mass and salt.

Now we derive the equations for the annual-mean and seasonal salinity budgets, showing the relationship between the salinity and previously calculated freshwater budgets.

The meridional flux of moisture across a latitude circle occurs partly in the atmosphere, partly in the form of land surface processes such as ground water seepage and river flow, and partly by transport in the ocean. The local storage of water between two latitude bands can be calculated either from mass or salt conservation. Let us begin with the mass conservation equation, but include horizontal and vertical diffusion.

$$\frac{\partial \rho}{\partial t} = -\nabla \cdot (\rho \vec{u}) + \nabla_h k_h \cdot \nabla_h \rho + \frac{\partial}{\partial z}\left(k_v \frac{\partial \rho}{\partial z}\right) \qquad (11)$$

where k_h is the horizontal diffusivity of density in water and k_v the vertical diffusivity of density in water. The surface represents a source and sink of mass by exchange with the land or atmosphere. Integrating (11) over the ocean volume gives us the surface boundary condition that

$$W(z=0) = -(P - E + \Sigma R_i\ \delta(x-x_i, y-y_i)) \quad (12)$$

where P, E and R are the rates of precipitation, evaporation and runoff in m s^{-1}. We assume for simplicity that river runoff enters the ocean at the surface at discrete entry points (X_i, y_i) along the boundary. The law of conservation of salt per unit volume is similar to the law for conservation of mass :

$$\frac{\partial(\rho s')}{\partial t} = -\nabla\cdot(\rho s'\vec{u}) + \nabla_h k_h \cdot \nabla_h(\rho s') + \frac{\partial}{\partial z}\left(k_v \frac{\partial(\rho s')}{\partial z}\right) \quad (13)$$

where s' is salinity per unit mass [‰] for salt. In (13), a source function for $\rho s'$ is not required because processes affecting salinity occur only at boundaries, e. g. river runoff, the difference between evaporation and precipitation. Such effects would be included as boundary conditions. We have assumed that density and salinity have the same coefficients of vertical diffusion. Rearranging this equation gives us an alternative equation for mass storage

$$\frac{\partial\rho}{\partial t} = \frac{1}{s'}\left[-\rho\frac{\partial s'}{\partial t} - \nabla\cdot(\rho s'\vec{u}) + \nabla_h k_h\cdot\nabla_h(\rho s') + \frac{\partial}{\partial z}\left(k_v\frac{\partial(\rho s')}{\partial z}\right)\right] \quad (14)$$

This equation can be rewritten as

$$\frac{\partial\rho}{\partial t} = -\rho\frac{\partial \ln(s')}{\partial t} - \nabla\cdot(\rho\vec{u}) - \rho\nabla\cdot(\ln(s')\vec{u}) + \frac{\nabla_h k_h\cdot\nabla_h(\rho s')}{s'} + \frac{\partial}{s'\partial z}\left(k_v\frac{\partial(\rho s')}{\partial z}\right) \quad (15)$$

where we have assumed $\nabla s'/s' \gg \nabla\rho/\rho$. Using (11), (13) can be reduced to

$$\frac{\partial s'}{\partial t} + \nabla\cdot(s'\vec{u}) = \nabla_h k_h \cdot \nabla_h s' + \frac{\partial}{\partial z}\left(k_v\frac{\partial s'}{\partial z}\right) \quad (16)$$

Back to (13) and from its vertical integration between a depth z = -H and z = 0, a salt conservation equation can be obtained as

$$H\frac{\partial \bar{s}}{\partial t} + H\,\overline{\nabla_h\cdot(s(\vec{u}))} + s_o W_o - s_{-H} W_{-H} = H\,(\overline{\nabla_h k_h\cdot\nabla_h s}) + \left(k_v\frac{\partial s}{\partial z}\right)_o - \left(k_v\frac{\partial s}{\partial z}\right)_{-H} \quad (17)$$

where s = ρs', and s means the mass of salt per unit volume of seawater [kg/m^3]. We also assume that the vertical velocity at the interface is

$$W_o = -(P-E+R) = -F \quad (18)$$

This means physically that the mass loss due to evaporation in a layer near the surface is compensated by upward mass movement, while the mass surplus due to precipitation and river runoff leads to downward mass movement through this layer. In addition, this assumption implicitly required $\frac{\rho_{fw}}{\rho}\approx 1$, where ρ_{fw} is the (salt-free) freshwater density. When the surface elevation η is small for large-scale processes and river runoff is regarded as a quantity similar to precipitation, the assumption of (18) is consistent with the kinematic boundary condition at the ocean surface described by Monin (1986) as

$$W = \frac{\partial\eta}{\partial t} + \frac{v}{a}\frac{\partial\eta}{\partial\varphi} + \frac{u}{a\cos(\varphi)}\frac{\partial\eta}{\partial\lambda} - \frac{1}{\rho}(\rho_{fw}P - \rho_{fw}E) \quad (19)$$

where a is the earth's radius, λ is longitude, φ is latitude, and the free surface, whose equilibrium position is $z=0$, has perturbed position $z=\eta$. Combining the surface bounary conditions (W_0=-F) and the bottom boundary condition (W_{-H} 0) gives

$$H\frac{\partial\bar{s}}{\partial t} + H\overline{\nabla_h\cdot(s\vec{u})} - s_o F = H\overline{\nabla_h k_h\cdot\nabla_h s} + (k_v\frac{\partial s}{\partial z})_o - (k_v\frac{\partial s}{\partial z})_{-H} \tag{20}$$

We have treated river runoff as a quantity similar to precipitation, assuming that the runoff is concentrated near surface. Rearranging (20), we get

$$-H\frac{\partial\bar{s}}{\partial t} + s_o F + H\overline{\nabla_h k_h\cdot\nabla_h s} + (k_v\frac{\partial s}{\partial z})_o - (k_v\frac{\partial s}{\partial z})_{-H} = H\overline{\nabla_h\cdot(s\vec{u})} \tag{21}$$

When a sufficiently thin layer (i.e. a small vertical integration integration interval) is considered, the horizontal advective-turbulent salt flow can be neglected (Bortkowski, 1967). The vertical mixing term at z=-H, $(k_v\frac{\partial s}{\partial z})_{-H}$ may be also neglected except for strong upwelling regions. These assumptions lead to an additional surface boundary condition that $(k_v\frac{\partial s}{\partial z})_o + s_o F = 0$ (see also Neumann, 1972). This surface boundary condition is analogous to the surface heat flux condition on temperature. Substituting it into (21) gives an equation which when integrated over the ocean volume says that the total salt will never change. For instance, with a calm sea surface and no horizontal motion, evaporation produces a skin layer of high surface salinity and strong vertical salt gradient in the upper few millimeters or meters is reduced and spread to lower layers by a turbulent vertical density-coupled net flux of mass. Thus, assuming that the annual average of the first term in (21) is negligable, and integrating zonally and meridionally

$$AST = AST|_{25^\circ N} + H\overline{<(\nabla_h\cdot(s\vec{u}))>} = AST|_{25^\circ N} + \underbrace{H\overline{<(\nabla_h k_h\cdot\nabla_h s)>}}_{C} - \underbrace{<\{(k_v\frac{\partial s}{\partial z})_{-H}\}>}_{D} \tag{22}$$

Since the total mass of salt between z=-H and z=0 must be constant over a long time period (i.e. no salt accumulation), AST should balance the mass of salt which gained by meridional diffusion (term C) and lost by vertical mixing (term D) at ocean bottom. However, in the tropical Atlantic the vertical mixing term at the deqth of 500 m is an order of magnitude less than the horizontal diffusion term, which leads to annual meridional salt transport (discussed later). Thus, (22) requires that meridional diffusion balance meridional advection.

Now we investigate the equations for the seasonal salt budget by computing the storge value in (21). Using the surface boundary condition of $(k_v\frac{\partial s}{\partial z})_o + s_o F = 0$, (21) is rewritten as

$$-H\frac{\partial s}{\partial t} + H\overline{\nabla_h k_h\cdot\nabla_h s} - (k_v\frac{\partial s}{\partial z})_{-H} = H\overline{\nabla_h\cdot(s\vec{u})} \tag{23}$$

By integrating (23) from season 1 (one season) to season 2 (next season), we have

$$-H\int_1^2\frac{\partial\bar{s}}{\partial t}dt + H\int_1^2\overline{\nabla_h k_h\cdot\nabla_h s}\,dt - \int_1^2(k_v\frac{\partial s}{\partial z})_{-H}dt = H\int_1^2\overline{\nabla_h\cdot(s\vec{u})}\,dt \tag{24}$$

Diving both sides of (24) by ΔT (i.e. one season period), this equation can be rewritten as

$$\Delta Y = H(\overline{\nabla_h\cdot(s\vec{u})_2} - \overline{\nabla_h\cdot(s\vec{u})_1}) = \underbrace{\frac{H}{\Delta T}(s_1 - s_2)}_{E} + \underbrace{H(\overline{\nabla_h k_h\cdot\nabla_h s_2} - \overline{\nabla_h k_h\cdot\nabla_h s_1})}_{F} - \underbrace{((k_v\frac{\partial s_2}{\partial z})_{-H} - (k_v\frac{\partial s_1}{\partial z})_{-H})}_{G} \tag{25}$$

where ΔY is the difference between salt mass divergences of the season 2 and season 1. The term E is the seasonal change of salt storage, calculated from the differences between salt masses of one season to the next.

The term F is the contribution of the mean horizontal salt diffusion from season 1 to season 2, while the them G is the contribution of the mean vertical mixing at ocean bottom from season 1 to season 2. In practical calculations, we use the values $k_h=2\times10^3$ m^2/s and $k_v=4\times10^{-3}$m^2/s (e. g., Neumann, 1972; Haney and Davis, 176). We also assume that

$$\nabla_h k_h \cdot \nabla_h s \simeq k_h \nabla_h^2 s = \frac{k_h}{a^2}\left(\frac{\partial^2 s}{\partial \varphi^2} - \frac{\partial s}{\partial \varphi}\tan(\varphi)\right) \tag{26}$$

S_1 and S_2 are the salinities in kg m^{-3} of season 1 and season 2, respectively. When integrated zonally and meridionally, the seasonal northward salt transport (SST) mass flux (kg/s) is :

$$SST = SST\,|_{25^\circ N} + \langle \Delta Y \rangle \tag{27}$$

The meridional integration is performed in one degree intervals.

On the calculation in (27), we have used following four assumptions : a) River runoff is regarded as a quantity similar to precipitation. b) $W_o = -(P-E+R) = -F$, surface boundary condition. c) $(k_v \frac{\partial s}{\partial z})_o + s_o F = 0$, surface boundary condition.

d) In the Laplacian calcualtion, $\nabla_h k_h \cdot \nabla_h s = k_h \nabla_h^2 s$. In the zonal average of the freshwater or salt, the error from (a) is negligible. The validity of the surface boundary conditions, (b) and (c), has been proved to be reasonable in earlier studies (Bortkowski, 1967 : Neumann, 1972 : Monin, 1986). Since the horizontal diffusivity, k_v, is almost constant in the tropical ocean, the error from the assumption in (d) can be ignored. In particular, the term in (27) which includes the horizontal diffusion is smaller by a factor of 2-3 than the storage term. Thus, total error which comes from four assumptions in (27) is not significant

Calculations of seasonal meridional salt transports in (27) requires the boundary values of the transports at 25° N. Combining Hall and Bryden's (1982) freshwater transport during October 1957 at 25° N with annual salinity value at the same latitude provides only an approximate annual value of the salt transport at 25° N. Since the Gulf stream flow through the Florida Straits in October is approximately close to its annual mean flow (Niiler and Richardson, 1973), the October freshwater transport of Hall and Bryden is assumed to be the annual freshwater transport at 25° N in this study. Thus, the seasonal salt transports are inferred from both the annual salt transport and the seasonal Florida Current transports of Niiler and Richardson (1973). Niiler and Richardson show that the seasonal variability of the total transport of the Current across the Florida Current is ±14 % of the average value. The annual meridional freshwater transport at 25° N is significantly less than those in other latitudes of the tropical Atlantic, as mentioned earlier, resulting in negligable meridional salt transport. Thus, we assume the boundary values of the annual and seasonal salt transports at 25° N are zero.

3. Data and analysis

Seasonal surface freshwater flux for the tropical Atlantic has been available in Yoo and Carton (1990), who obtain the flux by combining estimates of evaporation (Oberhuber, 1989), precipitation (Yoo and Carton, 1988), and river discharge (UNESCO, 1969 : 1971a, b : 1979) (Fig. 1a). Since the InterTropical Convergence Zone (ITCZ) in which much of the freshwater surplus due to heavy rainfall takes place, undergoes strong seasonal shifts, one can expect similarly strong seasonal changes in the sea surface salinity (Yoo, 1992). In this region the seasonal dounward moisture flux (i. e. E-P $<$ 0) is about 9-27 cm /month, while in the northern and southern subtropics the upward moisture flux (i. e. E-P $>$ 0) is about 9-15 cm/month. Inclusion of river runoff enhances the zonally averaged freshwater flux by up to 8 cm/ month. We can also anticipate the strong seasonal variations and lowest values in the sea surface salinity near the months of large rivers.

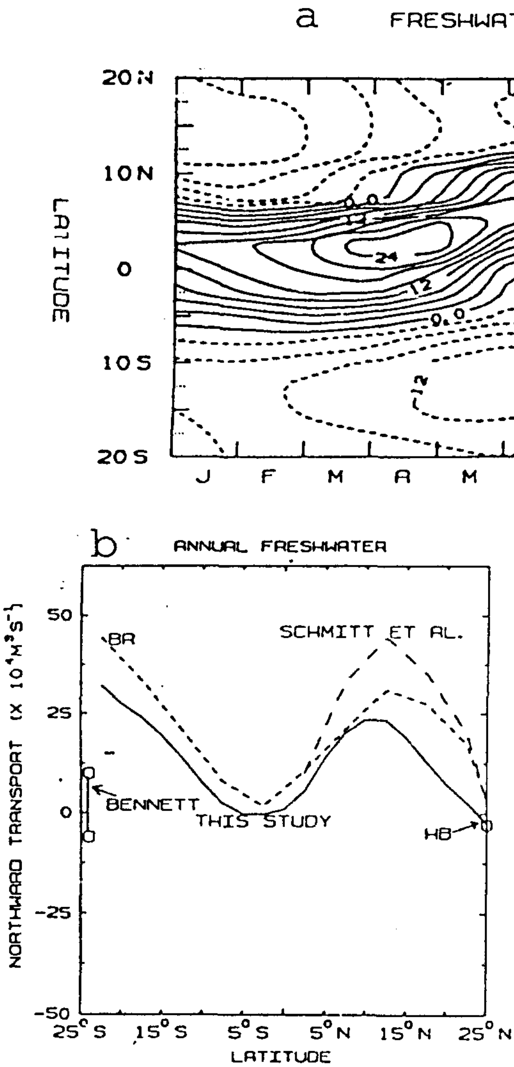

Fig. 1. Annual freshwater budget and transport for the tropical Atlantic.
(a) Freshwater budget as a function of latitude and time. Contour interval is 3 cm/month. Negative values indicate the excess amounts of evaporation over the summation of precipitation and river runoff.
(b) Northward transport of freshwater in the tropical Atlantic from Baumgartner and Reichel (1975, BR), Schmitt et al. (1989), Bennett (1978), Hall and Bryden (1982, HB), and this study. The value of Hall and Bryden (1982) at 25°N is used as a northern boundary in this study. Direct estimates are indicated by 'o'. Units are $10^4 m^3 s^{-1}$.

Salinity data 20° S-25° N has been obtained from the National Oceanographic Data Center (NODC) for the period 1900 to 1986. The number of full vertical profiles range from 9400 in winter to 19100 in summer, and the fewest observations have been collected in boreal winter. The results must be viewed with caution in data-sparse regions, for instance, the South Atlantic in winter. These data counts are about 30% larger than the data in the widely used anlaysis of Levitus (1982). The data in that analysis was restricted to the period prior to 1977.

The procedures for the construction of seasonal salinity in this study are as follows : first, we interpolate the salinity data to the standard NODC depths from the surface to 500 m depth (14 levels;see Yoo, 1992). Here, if a salinity profile has any level observations missing, the whole profile is discarded. The data are then composited into seasonal mean values in 1°×1° boxes. Secondly, the salinity differences between seasonal NODC observations and Levitus' annual average salinity are computed. This means that Levitus' annual salinity is used as a first-guess. We analyze the salinity differences using statistical objective analysis. In the analysis, a correction to the first-guess at each gridpoint is computed based on an optimal weighting of data surrounding the gridpoint. The weighting depends inversely on the distance between the gridpoint and the surrounding gridpoints (see Haltiner and Williams, 1980) with a correction scale of 400 km. Thus, the analyzed value at the gridpoint is the sum of the first guess values and the difference. Lastly, by adding Leitus' annual estimate back to the objectively analyzed Salinity differences, We recomstruct our estimated seasonal salinity fields on a 1° ×1° grid at the 14 standard levels.

It is well-known that the determinarion of the vertical salinity (S) distribution on the basis of the available observations is mord difficult than in the case of temperature (T), and that temperature is well correlated with salinity below the mixed layer. We have tried to improve the analysis of salinity by using the local T-S relationship in combination with Levitus' (1982) seasonal temperature to provide an improved first guess.

Comparison of our analysis of sea surface salinity with Levitus (1982) shows that in our analysis salinities for winter are systematically higher by 0.1-0.5 ‰ in the region of the ITCZ (not shown). The two annual-mean climatologies at the sea surface agree to within ± 0.1% except in the areas offshore of the major rivers, and agree to within ±0.05 ‰ in the vertical cross-section of zonally averaged salinity as a function of latitude and depth. In the region of strong seasonal variation associated with the North Equatorial Countercurrent, the vertical movement of salinity isohalines is larger in our analysis, more consistent with the motion of isotherms.

4. Meridional freshwater transport

4.1. Annual freshwater transport

Here we estimate the meridional fluxes based on the freshwater data (Table 1 and Figs. 1a-b). The freshwater budget is computed by adding P-E to the river runoff (Fig. 1a). The freshwater distribution as a function of latitude and time is similar to that of P-E, showing a semiannual variation (Yoo and Carton, 1990). However, the amount of freshwater flux in spring in the latitude area of 0-3° N is intensified due to the Amazon water, compared to other seasons.

By integrating the hydrological data of Baumgartner and Reichel (1975), Stommel (1980) computed the meridional annual mean freshwater transports of the global oceans including the Atlantic Ocean as a function of latitude (Table 1 and Fig. 1b). Stommel has shown that the North and South Atlantic Oceans have equatorward direction of freshwater flux. Following the same method, Schmitt *et al.* (1989) computed the annual freshwater transport in the North Atlantic using the precipitation estimates of Dorman and Bourke (1981), the evaporation estimates of Bunker (1976), and the runoff and boundary (70°N) freshwater estimates of Baumgartner and Reichel (1975). A comparison between the present freshwater transport and that of Stommel (1980) and Schmitt *et al.* (1989) in the tropical Atlantic shows a good agreement on direction, of which maximum is about 0.3-0.4 Sv northward (Table 1 and Fig. 1b). Maximum transports in both studies occur at 12°N or 22°S. However, our results are systematically lower by 30-50%. The reason is because of the higher freshwater flux in this study. The difference due to the boundary value makes our values systematically slightly lower (0.059 Sv) than the previous two studies.

Comparisons of freshwater transport show that the differences are larger in the Sourh Atlanstic. The freshwater flux estimates of both this study and Baumgartner and Reichel (1975) are also substantially larger by 3-4 times than those of Bennett (1978), calculated from hydrographic sections along 24°S.

4.2. Seasonal freshwater transport

Now we begin to examine seasonal meridional freshwater transports vertically averaged from surface to 500 m based on (10) using quarterly freshwater storage, and freshwater flux and divergence (Fig. 2).

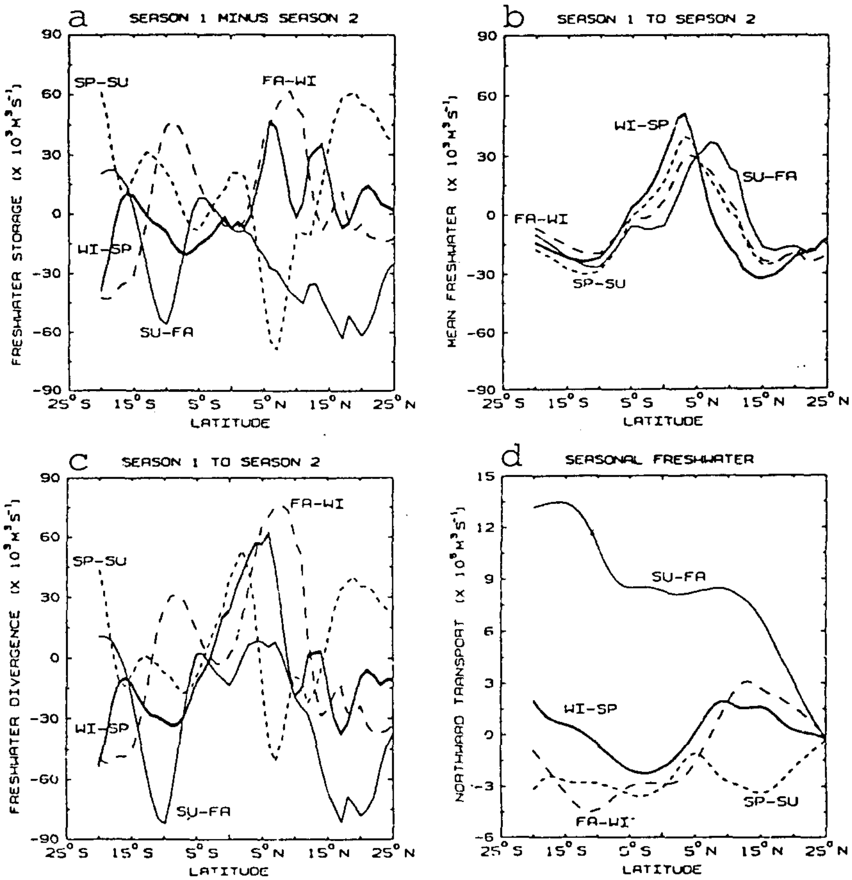

Fig. 2. Terms from the intraseasonal (season 1 to season 2) and zonal-average freshwater transport, vertically averaged from surface to 500m deqth for the tropical Atlantic.
(a) Freshwater storage, which corresponds to the freshwater difference between season 1 and season 2.
(b) Mean freshwater contributions from season 1 to season 2 to freshwater divergence,
(c) freshwater divergence, and (d) seasonal northward freshwater transport.
Units are $10^3 m^3 s^{-1}$ in (a), (b) and (c), and $10^4 m^3 s^{-1}$ in (d).

A large increase in freshwater storage ($6 \times 10^4 m^3 s^{-1}$) occurs at 19°N between spring and summer, while a large decrease ($-6 \times 10^4 m^3 s^{-1}$) occurs at 17°N-20°N between summer and fall (Fig. 2a).

Mean freshwater maxima move with seasonal mean position of the ITCZ (Fig. 2b). This figure reflects well the annual mean position of the ITCZ at 5°N. Freshwater surplus in the ITCZ results in freshwater divergence, while freshwater deficit in the subtropics results in freshwater convergence (Figs. 2b-c). The highest contributions ($3.5-5 \times 10^4 m^3 s^{-1}$) of freshwater to its divergence occur at 2°N-8°N, while the highest contributions ($-3 \times 10^4 m^3 s^{-1}$) of freshwater to its convergence occur at 15°N. The contributions ($-3 \times 10^4 m^3 s^{-1}$) of freshwater to its convergence occur at 15°N. The contribution of mean freshwater to its divergence is extremely large in the estuaries of large rivers. However, the procedure of zonal average reduces extrema spatially. Maximum freshwater at 2-3°N in the winter to spring curve is partially due to the large Amazon discharge. The spatial variations of the freshwater divergence field due to mean freshwater is lower by about 20 % than that due to the storage (Figs. 2a-c).

When we add the mean freshwater to the freshwater storage, we obtain seasonal freshwater divergence (Figs. 2c). For all four curves, it is common that there are freshwater divergence fields in the region between 5°S and 10°N and the convergence fields in the subtropics. Thus, the meridional distriburions of seasonal freshwater divergence in the ITCZ are generally in phase with those of mean freshwater contribution, suggesting more importance of the mean surface freshwater flux in spatial variation of freshwater divergence than the freshwater storage. However, there are substantial modifications of the freshwater divergence due to the strong seasonal changes of freshwater storage in the region between 5°N and 10°N and during the period spring-to-summer.

Based on the definition in (10), we integrate the divergence fields southward from 25°N in order to calculate the northward freshwater flux (Fig. 2d). The seasonal meridional transport amounts of freshwater from surface to 500 m depth in the tropical Atlantic Ocean range from 1.35 Sv to-0.45 Sv. We find that the freshwater transports have a maximum northward flux of 1.35 Sv at 15° S during summer-to-fall and a maximum southward flux of 0.45 Sv at 12°S during fall-to-winter. Northward freshwater transports prevail during summer-to-fall, while southward freshwater transports prevail during spring-to-summer.

The seasonal cycle of the freshwater transports is caused by seasonal shifts of the ITCZ as well as seasonal change in the local freshwater storage. In particular, the intensification of convergence due to the strong negative freshwater storage during summer-to-fall in the same region (Figs. 2a, 2c, and 2d). As seen in two extreme curves of the turning-point season in the change of direction in seasonal meridional freshwater transports.

5. Meridional salt transport

5.1. Annual Salt transport

Based on the salinity data we estimate the annual meridional salt fluxes vertically

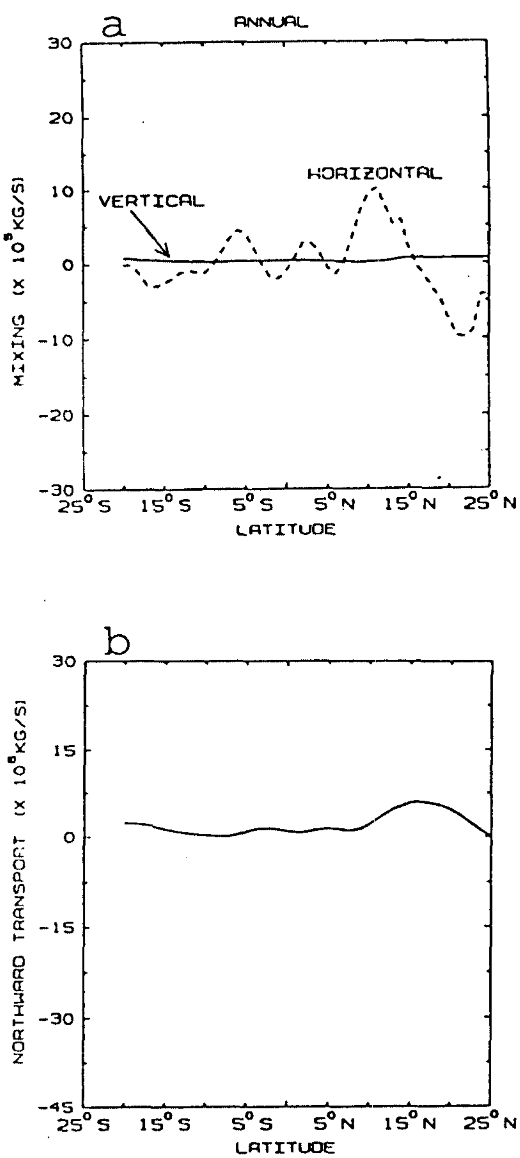

Fig. 3. Terms from the annual zonal-average salt transport, vertically averaged from surface to 500m depth for the tropical Atlantic.
(a) Horizontal mixing (dashed) and vertical mixing (solid). Since vertical mixing value is negligable, the amount of horizotal mixing is approximately equal to that of salt divergence.
(b) Annual northward salt transport due to horizontal diffusion. Units are 10^5kg/s in (a), and 10^6 kg/s in (b).

averaged from surface to 500 m depth in the tropical Atlantic by integrating the fields of salt divergence. As defined in (22), the salt divergence is the sum of the horizontal diffusion and negative the vertical mixing term at 500 m depth.

Since the vertical mixing term at the deqth is negligible, the horizontal diffusion value is almost identical to that of the salt divergence (Fig. 3a). Horizontal mixing curve which has a wave pattern shows maximum of positive horizontal diffusion (10^6kg/s) at 11°N, and of negative horizontal diffusion (-10^6kg/s) at 22° N. Here the positive diffusion leads to the salt divergence, while the negative diffusion leads to the salt convergence. In the steady steady state over a long time period, the annual salt flux in the tropical Atlantic is zero due to the conservation of salt mass. In other words, since the total mass of salt within a water column must be constant over such a periodc c. e. no salt accumulatirb the salt flux by horizatal advection balances the flux by horizontal diffusion.

5.2. Seasonal salt transport

As defined in (25), the calculation of seasonal salt transport requires for the information of the salt storage, the horizontal diffusion and the vertical mixing at the ocean bottom (Fig. 4). Since the vertical mixing term is 2-3 orders of magnitude less than the other two terms, the salt divergence is basically the sum of the salt storage and the horizontal diffusion.

The seasonal variation of the salt storage vertically integrated from surface to 500 m depth is weak in the region between 5° S and 4°N, and has an irregular pattern (Fig. 4a). Large positive storage estimates (2×10^6kg/s) occur in the summer minus fall at 10° S and in spring minus summer curve at 6°N, while large negative storage estimates (-2×10^6 kg/s) occur in the spring minus summer curve at 20°N and 20°S. Thus, the N-S gradient of the storage is largest during the period spring-to-

Horizontal diffusion which has nearly zero seasonal fluctuation is almost the same as the annual horizontal diffusion (Figs. 3a and 4b). In contrast to the freshwater case, the salt storage is the most important term in the calculation of the salt transport. The storage is 2-4 times as large as the horizontal diffusion (Figs. 4a-b).

By adding the horizontal diffusion to the salt storage, we obtain the salt divergence from one to the next season (Fig. 4c). Now the divergence field is generally similar to that of the salt storage, although the horizontal diffusion values in the region north of 10°N significantly modify the divergence. The salt divergence is predominant in the summer-to-fall curve, while in the spring-to-summer curve the salt convergence prevails. Maximum salt divergence (3×10^6kg/s) occurs in the summer-to-fall distribution at 10°N, while in the spring-to-summer distribution maximum salt convergence (3.2×10^6kg/s) occurs at 21° N.

Assuming that the seasonal salt fluxes at the boundary (25°N) are zero, we integrate the salt divergence field in order to compute seasonal meridional salt fluxes (Fig. 4d). The transport amounts of salt range from 30×10^6 kg/s at 20° S in the spring-to-summer curve to -37×10^6kg/s at 15° S in the summer-to-fall curve. As seen in the two extreme curves of the spring-to-summer and summer-to-fall, summer is a transition period when the direction of the transport converts from northward to southward. The direction of the seasonal salt transports in the tropical Atlantic is northward except for the period summer-to-fall. When the convergence of salt is particularly large at 21° N for the period spring-to-summer, the northward flux is strongest. The st-rong southward transport for the period summer-to-fall is due to the large positive salt storage and thus the large salt divergence in the region north of 10° N. The transports of high salinity water from subtropical regions of both hemispheres into equatorial region at the

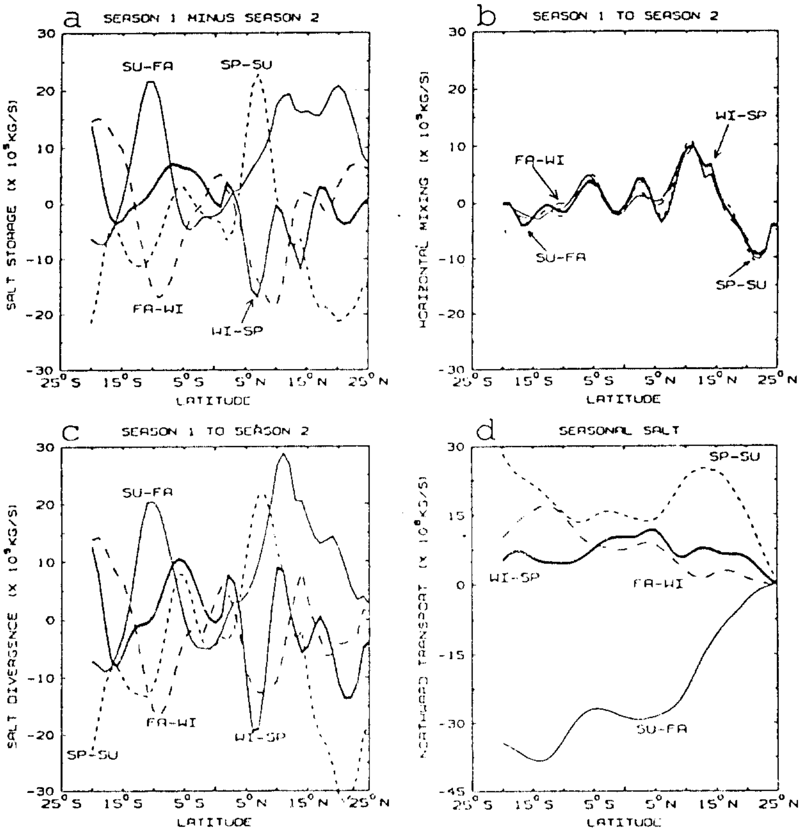

Fig. 4. Terms from the intraseasonal (season 1 to season 2) and zonal-average salt transport, vertically averaged from surface to 500m depth for the tropical atlantic.
(a) Salt storage, which corresponds to the salt difference between season 1 and season 2.
(b) The contribution of mean horizontal mixing from season 1 to season 2 to salt divergence,
(c) salt divergence, and (d) seasonal northward salt transport. Units are 10^5kg/s in (a), (b) and (c), and 10^6kg/s in (d).

thermocline depth may be also responsible for the seasonal salt transport (Neumann and Pierson, 1966).

Gordon and Piola (1983) obtained the northward salt flux of 700×10^6 kg/s from the box model of 35° S-65° N Atlantic at a NADW production rate of 20×10^6 m^3s^{-1}. In spite of the agreement in the transport direction, our value is one order of magnitude less than theirs. However, our estimate is one order of magnitude greater than that of bennett (1978), who has reported the salt meridional salt transport of -3×10^6 kg/s to 2×10^6 kg/s across 24° S during October 1958.

Uncertainties in the fluxes associated with the southward integration of the freshwater and salt transports must get larger as the calculation moves from north to south due to the accumulation of various types of errors. Thus, the cause for more seasonal variability in the transports south of the equator than north of the equator is unknown (Figs. 2d and 4d). For the verification, more observations in the southern ocean are required together with the direct and indirect methods.

6. Summary and Discussion

The problem addressed in this study concerns the seasonal freshwater and salt transports which are theoretically derived from the equations of the mass and salt conservation, respectively, showing local freshwater budget affects local salt budget.

The estimates of freshwater and salt transports from surface to 500 m depth in the tropical Atlantic Ocean are calculated from climatological freshwater data and objectively analyzed historical (1900-86) salinity observations, respectively.

The freshwater transport is obtained by southward integration of the divergence of surface freshwater flux. The annual freshwater transport in the tropical Atlantic is northward, ranging from 0 Sv near the equator to 0.3 Sv at 12° N and 20° S. This is consistent with earlier studies, obtained from the air-sea interaction method.

The seasonal meridional transport amounts of freshwater range from 1.35 Sv to -0.45 Sv. We find that the freshwater transports have a maximum northward flux of 1.35 Sv at 15° S during summer-to-fall, and a minimum southward flux of 0.45 Sv at 12° S during fall-to-winter. Northward freshwater transports prevail during spring-to-summer. This seasonal cycle is caused by the shifts of the ITCZ as well as the change in the local freshwater storage. However, the local storage has a more contribution by 20% to the transport than the surface freshwater flux.

We obtain annual salt transport from the salt divergence which is the sum of the horizontal diffusion and the vertical mixing at ocean bottom. Since the vertical mixing value is negligible, the horizontal diffusion is almost identical to the salt divergence. The annual salt flux is zero due to the conservation of salt mass. Here the salt flux by horizontal advection balances the flux by horizontal diffusion. The salt flux only due to the diffusion is northward, and has a maximum of 5×10^6kg/s ar 15° N.

Seasonal salt transports are calculated from the salt divergence, which is the sum of the salt storage, the horizontal diffusion and the vertical mixing at ocean bottom. The vertical mixing term 2-3 orders of magnitude less than other two terms, and the storage is 2-4 times as large as the horizontal diffusion. The transport amounts of salt range from 30×10^6 kg/s at 20°S in the spring-to-summer curve to -37×10^6kg/s at 15° S in the summer-to-fall curve. The direction of the seasonal salt transports is northward except for the period summer-to-fall. When the convergence of salt is particularly large at 21° N for the period spring-to-summer, the northward flux is strongest. The strong southward transport for the period summer-to-fall is due to the large positive salt storage and thus the large salt divergence in the region north of 10° N. The transports of high salinity water from subtropical regions of both hemispheres into equatorial region at the thermocline depth may be also responsible for the seasonal flux.

The seasonal variation in the freshwater and salt transports in the southern tropical Atlantic is, on average, 2-3 times as large as in the northern tropical Atlantic, although the cause is unknown. In addition to the lack of observations in the southern Atlantic, the present scheme of the souhward integration for the northward transports includes more uncertainties in the south due to the accumulation of errors. the southern ocean as well as the direct and indirect results, obtained from independent data sources. Another error source could be the transport by the deep water below 500 m depth, which has not been considered in this study.

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