Atlantic Ocean Upper Layer Salinity Budget

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ABSTRACT

Production of North Atlantic Deep Water (NADW) transfers upper-layer thermocline water into abyssal depths. Export of NADW across 35°S in the Atlantic Ocean into the Indian and Pacific Oceans by the Antarctic Circumpolar Current (ACC) requires a compensating flow of upper-layer water from the circumpolar zone of the Southern Ocean into the Atlantic. This water, en route to the NADW production regions, becomes saltier because evaporation exceeds precipitation and continental runoff. This process is responsible for a relatively salty Atlantic Ocean. Using estimates of the net freshwater flux, the increase of upper-layer salinity versus latitude is calculated for two NADW production rates: 15 × 10⁶ and 20 × 10⁶ m³ s⁻¹. The 20 × 10⁶ m³ s⁻¹ production rate provides the best relationship with the linear trend in salinity as determined from hydrographic data. It is suggested that a contributing factor to the establishment of a salty Atlantic Ocean, and possibly of NADW formation, is the removal of freshwater from the Atlantic Ocean by the ACC.

1. The problem and the model

The Atlantic Ocean north of the circumpolar ocean belt is significantly more saline than its Pacific and Indian counterparts. This generally well-known fact is clearly brought out in the world-ocean volumetric temperature–salinity relationship, the most recent being the detailed study of Worthington (1981). Worthington (1981, his table 2-1) shows the entire volume of the North Atlantic in the mean is 0.52% saltier than the North Pacific Ocean. The only stratum in which the Atlantic is less salty than the other ocean is within the Antarctic Intermediate Water and the cold arctic surface-water layers. Both the thermocline and abyssal water of the Atlantic are relatively salty. As the circumpolar zone is approached, inter-ocean differences diminish so that the South Atlantic is 0.21% and 0.05% saltier than the South Pacific and South Indian respectively. Within the circumpolar ocean the mean salinity varies little from sector to sector. The primary difference occurs in temperature (the Atlantic sector is significantly colder because of the large Weddell Sea region, see Plates 10-15 of Gordon and Baker, 1982).

A simple circulation model (as proposed by Stommel, 1957, and described in further detail by Reid, 1961) is used in an attempt to explain the relatively high salinity of the Atlantic thermocline. It can be thought of as an inverted estuary model (Fig. 1): in the northern North Atlantic, upper layer or thermocline water sinks into the abyssal layer during the production of North Atlantic Deep Water (NADW); NADW is exported to other oceans via the Antarctic Circumpolar Current (ACC), forcing a compensating northward flux of upper layer water from the South Atlantic across the equator and into the North Atlantic. Trans-equatorial flow of South Atlantic Central Water has been noted by Sverdrup et al. (1942, page 632). Worthington (1981, page 58) states: "Thus the formation of North Atlantic Deep Water follows a classic pattern—cold, dense water is formed at high latitudes, and warm surface water flows poleward to replace it. In this pattern, the process of water-mass formation changes the water characteristics radically." The objective of this study is to test this hypothesis by estimating the latitudinal salinity change within the upper layer due to ocean-atmospheric freshwater flux.

The southern limit of the model is taken at 35°S, which marks the northern boundary of the circumpolar zone: north of 35°S the Atlantic has both a western and eastern boundary. Across 35°S there is a southward flux of NADW in the abyssal layer, which is responsible for the circumpolar salinity-maximum (Deacon, 1937; Georgi, 1981). The density surface of \( \sigma_4 = 27.6 \) can be considered the boundary dividing the NADW outflow from the upper-layer inflow. This density surface separates the Antarctic Intermediate Water from NADW within the Atlantic Ocean. Georgi (1981) shows that the increased salinity in the Circumpolar Deep Water (CDW), due to

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1 Lamont-Doherty Geological Observatory Contribution No. 3507.
2 On leave from Servicio de Hidrografia Naval, Buenos Aires, Argentina.

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the input of NADW, is restricted to density values greater than 36.84 $\sigma_t$, which closely corresponds to 27.6 $\sigma_t$.

Northward flowing upper-layer water traverses two climatically dry subtropical zones en route to the formation regions of NADW near 65°N. Along its mean northward route the upper layer loses fresh water by means of evaporation ($E$), while it gains fresh water by precipitation ($P$), and continental runoff ($R$). In order to calculate the latitudinal salinity increase of the upper-layer water the atmospheric forcing must be known. In addition, the upper-layer salinity would be affected by vertical exchange with the NADW layer.

2. Atmospheric forcing

Estimates of the freshwater volume flux across the sea surface are available for 5° latitude bands in Baumgartner and Reiche (1975, hereafter referred to as BR), Fig. 2 and Table 1.

The best known component of the oceanic freshwater balance is the continental runoff. The major contribution to $R$, approximately one-third of the total continental runoff into the Atlantic Ocean, is introduced near the equator by the Amazon River. Estimates of $P$ and $E$ over the oceans in the BR study are based on existing distribution maps (e.g., Albrecht, 1960).

An independent estimate of the meridional distribution of precipitation over the Atlantic Ocean in the latitude range 30°S to 65°N is given by Dorman and Bourke (1981), also shown in Fig. 2. Using their precipitation values averaged over 2° latitude bands, the total precipitation volume is $\sim 10\%$ higher than BR for the same latitude range (Table 2). About 40% of the difference occurs in the South Atlantic where observations are scarce and precipitation estimates less certain.

The meridional distribution of evaporation in 10° latitude bands can be calculated from Bunker’s analysis of the air–sea energy fluxes in the latitude range 30°S to 65°N over the Atlantic Ocean (Bunker, 1980; Fig. 2). Differences between these estimates and those of BR arise primarily in the Northern Hemisphere where Bunker’s values are $\sim 15\%$ higher.

From the available data it is therefore possible to obtain two estimates of the net freshwater volume flux, one from the BR study and another by combining BR continental runoff, Dorman and Bourke’s (1981) precipitation and our estimates of evaporation derived from Bunker’s study. These net fluxes are shown in Fig. 2. It should be noted that there are larger discrepancies among other estimates than those discussed above. Total precipitation volume over the North Atlantic from Reed and Elliott (1979) is $\sim 20\%$ lower than BR. Similarly, the total evaporation volume derived from Bunker’s data using Budyko’s (1963) method is higher by $\sim 25\%$. 

Fig. 1. Schematic diagram of the meridional circulation model for the Atlantic Ocean. The upper layer is required to flow northward from 35°S across the equator and into the northern North Atlantic in order to conserve the mass lost by formation of North Atlantic Deep Water. The salinity of the upper layer is altered by freshwater exchange with the atmosphere. The Atlantic abyssal layer exports the relatively salty NADW into the Antarctic Circumpolar Current and atmospheric water vapor is exported southward across 35°S and eventually contributes in decreasing the salinity of the upper layer south of Africa. The parameters $M_a$, $M_o$, $M_s$ are described in the text.
Since Baumgartner and Reichel provide the total $F$ values, their estimates are used to test the inverted estuary model. For a steady state condition the excess amount of the fresh water balance must be removed by the northward flux of upper-layer water involved in the production of NADW. The $35^\circ$S to $65^\circ$N excess is $-22,982$ km$^3$ year$^{-1}$, (the negative sign denotes net freshwater loss). A decision is required as to how this net should be distributed as a function of latitude. The complexity of the three-dimensional ocean circulation does not allow for a local distribution of $F$: one in which the excess freshwater value for each $5^\circ$ latitude belt is used to determine the salinity change of the northward flux across the $5^\circ$ latitude belt.

A simple approach is followed: The total excess fresh water for the ocean area between $35^\circ$S and $5^\circ$N ($-12,186$ km$^3$ year$^{-1}$, Table 1) and from $5^\circ$ to $65^\circ$N ($-10,796$ km$^3$ year$^{-1}$, Table 1) is distributed uniformly within each area. Subdivision of freshwater flux at $5^\circ$N is chosen as it marks the mean position of the Inter-Tropical Convergence Zone in the Atlantic (Newton, 1972; Fig. 9.14b) which marks the separation of the hemispheric wind regimes. A more refined distribution of the net fresh water requires detailed knowledge of the three-dimensional advective–diffusion pattern; specifically, the actual path and depth of the northward flow component which balances NADW formation.

3. Thermocline to NADW conversion rate

The rate of salinity increase of the northward flowing upper-layer water depends not only on the atmospheric forcing, but also on the magnitude of the northward flux. This is dependent on the conversion rate of thermocline water into NADW.

A review of the estimates for NADW formation is provided by Warren (1981). The full Norwegian Sea overflow transport is believed to be $\sim 10 \times 10^6$ m$^3$ s$^{-1}$; Labrador Sea input to the NADW is $3.5 \times 10^6$ m$^3$ s$^{-1}$, and the outflow from the Mediterranean Sea is $1 \times 10^6$ m$^3$ s$^{-1}$, though the direct contribution of Mediterranean overflow to the abyssal volume is not known. It is possible that the bulk of Mediterranean outflow enters the Atlantic upper layer, eventually contributing to the NADW formation. Thus the total NADW production seems to be $\sim 14 \times 10^6$ m$^3$ s$^{-1}$. Broecker (1979), using radiocarbon data, suggests a NADW production rate of $\geq 20 \times 10^6$ m$^3$ s$^{-1}$, which is the value used by Stommel (1958) and Stommel and Arons (1960).

For this model, northward fluxes of $15 \times 10^6$ m$^3$ s$^{-1}$ and $20 \times 10^6$ m$^3$ s$^{-1}$ are used.

4. Model versus observed upper-layer salinity values

The observed salinity is determined from the IGY Atlantic data (Fuglister, 1960). The area-weighted
TABLE 1. Net freshwater volume flux $35°$S–$65°$N.

<table>
<thead>
<tr>
<th>Latitude belt</th>
<th>$F^a$</th>
<th>Area $^b$ $(\text{km}^2 \times 10^8)$</th>
<th>$F^c$</th>
<th>Arctic input $^d$</th>
</tr>
</thead>
<tbody>
<tr>
<td>35–30°S</td>
<td>-1707</td>
<td>3728</td>
<td>-1716</td>
<td>0</td>
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<tr>
<td>30–25°</td>
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<td>3034</td>
<td>-1397</td>
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<td>4602</td>
<td>-1152</td>
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</tr>
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<td>-158</td>
<td>3909</td>
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<td>2303</td>
<td>-576</td>
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<td>+1470</td>
<td>2380</td>
<td>-596</td>
<td>0.3</td>
</tr>
</tbody>
</table>

$^a$ $F$ is the freshwater input, defined as Precipitation – Evaporation + Runoff, in units of km$^3$ per year within 5° latitudinal belts (Baumgartner and Reicheil, 1975). The total value for $F$, 35°S–65°N, is $-22.982$ km$^3$ per year (0.732 $\times 10^8$ m$^3$ s$^{-1}$). The total value (excess evaporative water loss) 35°S–5°N (South Atlantic Central Water) is $-12.186$ km$^3$ per year (0.388 $\times 10^8$ m$^3$ s$^{-1}$). The total value (excess evaporative water loss) 5°S–65°N (North Atlantic Central Water) is $-10.796$ km$^3$ per year (0.344 $\times 10^8$ m$^3$ s$^{-1}$).

$^b$ Ocean area for each 5° latitude belt. The total ocean area from 35°S–65°N is 69.585 $\times 10^8$ km$^2$, 26.468 $\times 10^8$ km$^2$ from 35°S–5°N; 43.117 $\times 10^8$ km$^2$, 5°S–65°N (areas from Baumgartner and Reicheil, 1975).

$^c$ The net freshwater flux (F, negative value denotes excess evaporation) spread uniformly within each 5° latitude belt. The 35°S–5°N and 5°N–65°N excess evaporative water-loss values are treated separately. Values are in km$^3$ per year. The F' values north of 5°N are smaller than those south of 5°N because the net evaporative water loss north of 5°N is less and there are more 5° belts.

$^d$ Arctic input (see Section 5) is expressed as the ratio of the input of excess freshwater derived from the Arctic region north of 65°N introduced into each 5° latitudinal belt, determined from the approximate area ratio within each belt with surface salinity below 35%. The Arctic input of freshwater across 65°N is composed of the runoff and excess precipitation over evaporation into the polar sea (Baumgartner and Reicheil, 1975) of +4179 km$^3$ per year (0.133 $\times 10^8$ m$^3$ s$^{-1}$) plus Pacific-to-Atlantic transfer of low salinity water (33%) across the Bering Straits of 1.5 $\times 10^8$ m$^3$ s$^{-1}$ (Coachman et al., 1975).

average salinity for each IGY trans-Atlantic section from the sea surface to the surface where $\sigma_o = 27.6$ is presented in Fig. 3. The best linear fit to the data is shown in this figure. This is this linear trend of salinity that is a response to the net evaporative water flux. The superimposed variability is interpreted as a product of the full circulation and freshwater balance.

The salinity increase as a function of latitude is determined using the inverted estuary model and by applying salinity and mass conservative principles within each 5° latitude box. The salinity of each box is determined by

\[
M_n + M_i + F = 0, \quad (1a)
\]

\[
M_nS_n + M_iS_i = 0, \quad (1b)
\]

where $M_n$ is volume transport across the northern boundary of each box, with a salinity of $S_n$, $M_i$ the volume transport across the southern boundary of each box, with a salinity of $S_i$ and $F$ the freshwater volume flux into each box.

The model neglects horizontal mixing of salt. Using a value of $10^7$ cm$^2$ s$^{-1}$ for the coefficient of horizontal mixing with the mean meridional salinity gradient (Fig. 3) shows a southward salt flux of $2 \times 10^8$ g s$^{-1}$. This is much less than the northward salt flux of $7 \times 10^{11}$ g s$^{-1}$ associated with a mean northward transport of $20 \times 10^8$ m$^3$ s$^{-1}$ within the upper-layer, as discussed below, justifying the purely advective box model.

The initial value for $S_i$ (the salinity of the water entering the 35°–30°S box) is taken as 34.5%. This is determined by a linear interpolation between the mean salinity at 32°S (Fig. 3) and from a METEOR section along 41°S (not shown in Fig. 3).

The model results, for both $15 \times 10^6$ m$^3$ s$^{-1}$ and $20 \times 10^6$ m$^3$ s$^{-1}$ for the northward flow of upper layer water and the $F$ (Table 1) are given in Fig. 3. The model results for $20 \times 10^6$ m$^3$ s$^{-1}$ yield the best agreement with the observed linear trend.

5. Arctic and upwelling effects

Further refinement of the model is carried out by including the southward freshwater flux across 65°N supplied from the Arctic and the upwelling of abyssal water into the upper layers.

The addition of Pacific water to the Arctic Sea by transport through the shallow Bering Straits and excess fresh water primarily due to continental runoff introduced directly into the Arctic Sea result in a net southward flux of low-salinity water into the open North Atlantic Ocean. Coachman et al. (1975) give a mean northward flux through the Bering Strait of $1.5 \times 10^8$ m$^3$ s$^{-1}$ with a mean salinity of 33%. Baumgartner and Reicheil (1975) give the total yearly accumulation of fresh water into the Arctic Sea north of 65°N as 4179 km$^3$, or 0.133 $\times 10^8$ m$^3$ s$^{-1}$.

The total Arctic accumulation of fresh water and Bering Straits input must ultimately spread into the open Atlantic south of 65°N. This is accomplished by the southward flowing current in the western part

TABLE 2. Net freshwater volume flux into the Atlantic Ocean (30°S–65°N) in km$^3$ per year.

<table>
<thead>
<tr>
<th></th>
<th>$R$</th>
<th>$P$</th>
<th>$E$</th>
<th>$F$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baumgartner and Reicheil</td>
<td>17 135</td>
<td>51 421</td>
<td>-89 831</td>
<td>-21 275</td>
</tr>
<tr>
<td>Combined</td>
<td>17 135</td>
<td>57 166</td>
<td>-96 419</td>
<td>-22 118</td>
</tr>
</tbody>
</table>
of the North Atlantic. In a steady-state situation the Arctic water must mix into the upper layer which is migrating towards the NADW formation site, hence depressing its salinity. Integration of the Arctic water is assumed to occur in the latitude 40°–65°N in approximate proportion to the ratio of surface water of salinity less than 35% to that greater than 35% within each 5° latitude belt across the Atlantic.

The role of upwelling between the two layers must also be considered. Mass conservation requires the return of NADW into the upper layer. A mechanism for this return is direct upwelling into the thermocline. If the global thermocline upwelling rate is constant, it is expected that the Atlantic thermocline would receive a portion consistent with its percentage of the total thermocline area of the world ocean, or \( \sim 25\% \) of the total. Hence if \( 20 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) of NADW are formed, \( 5 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) would upwell before reaching the Antarctic Circumpolar Current. Thus, only \( \sim 13.4 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) need be transferred into the upper layer across the southern boundary, with \( 1.63 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) (as derived from the Bering Strait and freshwater flux into the Arctic) added across the northern boundary at 65°.

Using a vertical diffusivity of 10 cm² s⁻¹ and the average vertical salinity gradient across the 27.6σθ surface, the upward vertical diffusive salinity flux is approximately one order of magnitude less than the salinity flux accomplished by upwelling. Hence diffusive exchange between the two layers is neglected.

The Arctic and upwelling effects require modification of Eqs. (1a) and (1b):

\[
M_n + a(M_A + F_A) + M_n + M_n + F = 0, \quad (2a)
\]

\[
M_n S_n + aM_A S_A + M_n S_t + M_n S_u = 0, \quad (2b)
\]

where the additional terms are as follows:

- \( M_A \): The volume transport through the Bering Strait into the Arctic (1.5 \( \times 10^6 \text{ m}^3 \text{ s}^{-1} \)) where \( a \) is the ratio of this transport integrated into each 5° latitude box from 40° to 65°N (see Table 1) and \( S_n \) is the Bering Strait salinity (33%)

- \( F_A \): Freshwater input to the Arctic (north of 65°N) from the atmosphere and continental runoff (0.133 \( \times 10^6 \text{ m}^3 \text{ s}^{-1} \), see Table 1).

- \( M_u \): The volume transport of upwelling into each 5° latitude box, and \( S_u \) is the salinity of this upwelling water (taken as 34.85% characteristic of the lower layer; it is noted that the calculations are not sensitive to \( S_u \)).

The model is run for a \( 20 \times 10^6 \text{ m}^3 \text{ s}^{-1} \) NADW production rate with the Arctic and upwelling effects, the result given in Fig. 3. The model indicates the maximum salinity increase is attained by 45°N. The lowering of salinity to the north is a consequence of the Arctic effect. Hence if it were not for the Bering Strait transport and Arctic river runoff, the northern North Atlantic would be far more salty.

It is possible to examine the model sensitivity to changes in \( F \) and \( M_n \). For a single box extending from 35°S to the Bering Strait Eqs. (2a) and (2b) may be written as

\[
F = M_n k - \left( M_A + M_n \right) + \frac{M_n S_A + M_n S_u}{S_{35°S}}, \quad (3)
\]

where

\[
k = \frac{S_{35°S} - S_n}{S_{35°S}}
\]
and \( M_n \) is the volume transport with salinity of \( S_n \) from the upper layer across the 27.6\( \sigma_T \)-surface into the lower layer. The numerical value of the last two terms is \(-0.015\) and \( F \), which is now the total freshwater introduction from 35°S to the Bering Strait, is \(-0.599 \times 10^6\) m³ s⁻¹ (Table 1, notes a and d).

Lines for three \( k \) values are drawn in \( F \) versus \( M_n \) space (Fig. 4): \( k = -0.0203 \) corresponding to a \( S_n \) of 35.2% as given by Worthington (1970) for the inflow to the Norwegian Sea; \( k = -0.0292 \) and \(-0.0389\) corresponding to \( 20 \times 10^6 \) and \( 15 \times 10^6\) m³ s⁻¹ for \( M_n \) respectively, for \( F = -0.599 \times 10^6 \) m³ s⁻¹. The comparison of the model salinity versus latitude to the linear trend of the observed upper-layer salinity (Fig. 3) suggests a NADW production rate of \( 20 \times 10^6\) m³ s⁻¹. From Fig. 4 the salinity of the upper-layer water corresponding to \( M_n \) of \( 20 \times 10^6\) m³ s⁻¹ is 35.5% for \( F = -0.599 \times 10^6\) m³ s⁻¹, which is higher by 0.3% than the salinity of the Norwegian Sea inflow (Worthington, 1970). A net freshwater loss for 35°S to the Bering Strait of \(-0.42 \times 10^6\) m³ s⁻¹ (70% of the BR value) is required to yield an \( M_n \) of \( 20 \times 10^6\) m³ s⁻¹ with \( S_n \) of the Norwegian Sea inflow. Alternatively, an improved relation between the BR \( F \) values and acceptable NADW production rates is achieved if the mean salinity of the upper-layer water transferred to the lower layer is above 35.2%. In any case, the need for improved freshwater flux estimates for the world ocean is apparent.

6. Antarctic Circumpolar Current freshening

Prevailing winds carry water vapor poleward across 35°S (Newton, 1972; Fig. 9.14b), feeding the excess precipitation belt in the circumpolar region (BR). It is expected that this water freshens the surface layer of the ACC en route across the South Atlantic. The ACC then advects this fresh water into the Indian sector of the Southern Ocean.

Therefore, an independent test of the model is to inspect for freshening of the ACC surface water. The maximum amount of fresh water derived from excess evaporation between 35°S and 5°N is 12186 km³ per year, or \( 0.388 \times 10^6\) m³ s⁻¹ (Table 1). The northern boundary marks the approximate mean position of the Inter-Tropical Convergence Zone separating the hemispheric wind regimes, across which minimum water vapor flux is expected (Newton, 1972; Fig. 9.14b).

As discussed before, the salinity-enriched ACC water is restricted to density greater than \( \sigma_T = 27.6 \). The water less dense than \( \sigma_T = 27.6 \) can be inspected for a salinity decrease. The transport-weighted salinity for waters less dense than \( \sigma_T = 27.6 \) passing south of Africa is 0.198% less than that passing through the Drake Passage (Table 3).

The salinity south of Africa would be lowered by 0.203% if all of the net excess water vapor evaporated between 35°S and 5°N in the Atlantic Ocean fell as excess precipitation into the ACC passing across the South Atlantic. Thus the freshening of ACC surface water can account for freshwater export from the Atlantic Ocean drainage basin.

The required water vapor transfer out of the Atlantic basin north of 5°N of \( 10796\) km³ per year (0.344 \( \times 10^6\) m³ s⁻¹) must be accounted for in another way, most likely by atmospheric transport into the Pacific Ocean drainage basin. The route of this transport is

<table>
<thead>
<tr>
<th>Cruise/year</th>
<th>Geostrophic transport* (( \times 10^6 ) m³ s⁻¹)</th>
<th>Transport weighted salinities (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Drake Passage</td>
<td></td>
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<tr>
<td>FDRAKE 76</td>
<td>66.7</td>
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<td>FDRAKE 76</td>
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<td>FDRAKE 80</td>
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<td>South of Africa</td>
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<td>CONRAD 17</td>
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<td>ISLAS ORCadas 11 and 12</td>
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<tr>
<td>Mean values</td>
<td>54.8</td>
<td>( \Delta S = 0.198% )</td>
</tr>
</tbody>
</table>

* Relative to 3000 db.
fer is not clear, but usually the westward wind over Central America is considered the likely means of the transfer (Weyl, 1968).

7. Discussion

Comparison of salinity determined with an inverted estuary model for the Atlantic Ocean with observed salinity, particularly for a North Atlantic Deep Water production rate of $20 \times 10^6$ m$^3$ s$^{-1}$, is good. Thus we conclude that the Atlantic Ocean is relatively salty because NADW production forces northward movement of upper-layer water under climatically dry atmospheric conditions which force excess evaporation (however, the thermal field is not independent of evaporation; Warren, 1983). The NADW forms because of the high salinity of the North Atlantic which requires less heat removal to attain abyssal-layer density. As Gordon (1975) points out, NADW actually represents a heat input to the abyssal layer relative to the mean abyssal temperature in addition to a salinity input.

What came first: NADW formation or a salty Atlantic? We cannot rule out that an episodic event initiated a prototype NADW which then triggered inverted estuary circulation and a salty Atlantic. However, it is possible to envision a situation where a salty Atlantic preceded NADW formation. Such a condition would be established when plate-tectonic factors allowed a circumpolar current about 25 million years ago (Kennett et al., 1974). As discussed above, it is probable that freshening of the ACC within the Atlantic sector is derived from the South Atlantic subtropical net-evaporative zone, in which case the ACC extracts fresh water from the Atlantic and is hence a contributing factor to a relatively salty Atlantic. The increased salinity of the subtropical South Atlantic would mix into the North Atlantic circulation pattern, perhaps aided by the splitting of the South Equatorial Current by South America (5°S), and eventually NADW formation was made possible. Once NADW formation processes began, positive feedback occurred, resulting in the inverted estuary circulation.

As Reid (1961) states, a process which might contribute to the salinity difference between the Atlantic and Pacific Oceans is "the difference in the southern limits of South America (56°S), Africa (35°S) and Australia (44°S)." If the southern end of Africa extended to higher latitude, as does South America, it is possible that more low salinity water would return into the Atlantic rather than flow into the Indian and Pacific Oceans. In such a case the upper layer of the Atlantic would not be relatively salty and NADW would not form (by the present mechanism). Kennett (1978) suggests the thermal isolation of Antarctica by the establishment of the ACC led to deep-reaching convection and Antarctic bottom-water production.

It is possible the same event led to NADW production.

The establishment of the ACC by plate-tectonic factors should not be viewed in isolation from other geographic alteration. For example, the closing of the Atlantic–Pacific connection by shallowing of the Isthmus of Panama could be significant. The Panama uplift, estimated at 3–4 million years ago, strongly affected the circulation pattern in the Caribbean and eastern Pacific (Keigwin, 1982). It is expected that before the Panama uplift, South Atlantic water would escape into the Pacific and not continue to the western boundary current in the North Atlantic. Thus the establishment of the ACC may have been a necessary but not sufficient condition for NADW production.

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