A monthly analysis of the global wind stress and the ocean transports predicted from a numerical model

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SUMMARY

The monthly wind stress fields over the oceans have been calculated from marine climatic data incorporating the effect of thermal stability. The resulting angular momentum balance due to frictional torques indicates that the drag coefficients used in the study are essentially correct. A half-yearly oscillation of zonal wind stress in high temperate latitudes in the southern hemisphere is clearly evident. The curl of the wind stress computed from the wind stress data has been used as the driving for a numerical model of the transports of a homogeneous flat-bottomed world ocean. The variations of the ocean transports obtained are comparable with the rather sparse data except in the case of the Circumpolar Current and the Gulf Stream.

1. INTRODUCTION

The calculation of the wind stress over the ocean has been the subject of many investigations (Priestley 1951; Hidaka 1958; Hellerman 1967). It is important for two reasons. Firstly, through considerations of the relative angular momentum of the atmosphere it allows a global check to be made on the values of the drag coefficient over the sea deduced from micrometeorological measurements (Priestley 1951; Newton 1972). Secondly, the calculation is essential for an investigation of oceanic transports.

This study is based on monthly averaged wind stresses computed by Eyre (1972) by methods outlined in section 2. These stresses which are found to be satisfactory from angular momentum considerations are used in a numerical model to predict the variability of oceanic transports.

2. METHOD OF WIND STRESS ANALYSIS

The wind stress fields used in this study were computed by Eyre from wind roses, gale roses, surface atmospheric pressure and air temperature, and air–sea temperature differences given in U.S. Navy Hydrographic Office Publications (1955–1963, 1957) for each month of the year.

For each wind rose, and for each month, the components of surface wind stress in the \( x \)-direction (towards the east) and in the \( y \)-direction (towards the north) are computed from the relations

\[
\tau_x = \sum_j \sum_i f_{ij} \rho c_i u_i^2 \sin \theta_j
\]

\[
\tau_y = \sum_j \sum_i f_{ij} \rho c_i u_i^2 \cos \theta_j
\]

where \( f_{ij} \) is the frequency at which the wind in the Beaufort Scale \( i \), characterized by speed \( u_i \) (nominally at 10m) and in direction \( \theta_j \), is observed, and where

\[
\theta_j = 45j, \quad j = 0, 1, \ldots, 7;
\]

\( c_i \) is the surface drag coefficient for Beaufort Scale \( i \); and \( \rho \) is the air density.
TABLE 1. THE NEUTRAL DRAG COEFFICIENT

<table>
<thead>
<tr>
<th>Beaufort Scale</th>
<th>Characteristic speed ( u_i ) (m s(^{-1}))</th>
<th>( c_{i0} \times 10^3 )</th>
<th>Eyre (1972)</th>
<th>Hellerman (1967)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.77</td>
<td>1.20</td>
<td>0.85</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>3.91</td>
<td>1.25</td>
<td>0.85</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>8.55</td>
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<td>0.90</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>14.25</td>
<td>1.70</td>
<td>1.28</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>19.1</td>
<td>1.94</td>
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<td></td>
</tr>
<tr>
<td>6</td>
<td>22.0*</td>
<td>2.12*</td>
<td></td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>22.7</td>
<td>2.14</td>
<td>2.39</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>26.6</td>
<td>2.50</td>
<td>2.43</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>30.7</td>
<td>2.94</td>
<td>2.43</td>
<td></td>
</tr>
</tbody>
</table>

* If gale roses, i.e., the distribution of frequencies of winds in Beaufort Scales \( \geq 5 \), are not available, this value is used for scales \( \geq 5 \).

The drag coefficient for neutral conditions \( (c_{i0}) \) which was used in the computations (see Table 1) is based on an extensive survey of the literature (Kitaygorodskiy 1969; Deacon and Webb 1962; Smith 1970), and differs slightly from that used by Hellerman (1967). In non-neutral conditions the drag coefficient \( c_i \) was obtained from a series of graphs of \( c_i/c_{i0} \) against air-sea temperature difference \( (\Delta T) \) for various wind speeds (Webb 1970 and private communication; Dyer and Hicks 1970). This enabled some account to be taken of thermal stability, which has not been done in previous studies (Priestley 1951; Hidaka 1958; Hellerman 1967). However, no correlation of air-sea temperature difference with wind speed was available from the data.

Approximately 240 wind roses, fairly well distributed through all the oceans, although the borders of individual maps sometimes overlapped with an apparent duplication of data, were analysed for each month. Annual mean wind stresses based on this analysis are shown in Fig. 5. For convenience the wind stress components were interpolated on to a grid of 12 Gaussian latitudes and 24 equally spaced longitudes for analysis into a spherical harmonic series representation with 13 waves in the north-south direction and 12 waves in the east-west direction. The stress components were represented as follows

\[
\tau_x = \sum_{M=0}^{n} \sum_{N=0}^{12} (A_{MN} \cos M\lambda + B_{MN} \sin M\lambda) P_N^M N + M(\mu) \quad (3)
\]

\[
\tau_y = \sum_{M=0}^{n} \sum_{N=0}^{12} (C_{MN} \cos M\lambda + D_{MN} \sin M\lambda) P_N^M N + M(\mu) \quad (4)
\]

where \( \mu = \sin \phi \). \( \phi \) is latitude, \( \lambda \) is longitude and \( P_N^M N + M(\mu) \) is the normalized associated Legendre function

\[
P_N^M N + M(\mu) = \frac{1}{2^n + M(N + M)!} \left( \frac{(2(N + M) + 1)!}{2(N + 2M)!} \right)^{1/2} (1 - \mu^2)^{M/2} \frac{d^{N + 2M}}{d\mu^{N + 2M}} (\mu^2 - 1)^M.
\]

A full discussion of the analysis, and tables of \( A_{MN}, B_{MN}, C_{MN}, \) and \( D_{MN} \) are given in Eyre.

The zonal wind stress \( (\tau_{zz}) \) obtained from Eq. (3) is given by

\[
\tau_{zz} = \sum_{N=0}^{12} A_{0N} P_N^0(\mu) \quad (5)
\]

For the zonal stress it was found that the coefficients for \( N \leq 8 \) were significant, the inclusion of the higher coefficients did not improve the representation of the field. The fine structure of the data, especially near the equator, was not well represented by the spherical
harmonics but this was not considered to be a serious shortcoming as much of this fine structure can be attributed to noise in the original data.

3. The Annual Zonal Wind Stress

The annual zonal wind stress obtained by Eyre differs from that of Hellerman principally in that its maximum in the southern Westerlies is about 0.04 N m\(^{-2}\) smaller. There is also an interesting symmetrical structure about the equator in the trade winds, with local stress minima at 20°S and 20°N. Eyre’s results extend further south than Hellerman’s and indicate a well-developed easterly stress south of 62°S. The inclusion of the effect of thermal stability does not appear to have caused any appreciable changes in the distribution,
although in the equatorial regions typical values of \( \Delta T = -3 \) degC and \( u_t = 3.91 \text{ m s}^{-1} \) would yield an enhancement of the drag coefficient above its neutral value by a factor of about 1.2 (Dyer and Hicks 1970).

The credibility of the deduced wind stress fields can be investigated by calculating the wind stress torque per radian around a latitude circle \( T = -2na^3 \bar{\tau}_{xx} \cos^2 \phi \), where \( \phi \) is latitude, \( a \) is the radius of the earth, and \( \bar{\tau}_{xx} \) is the annual mean zonal stress. A comparison with the frictional torque of Newton (Fig. 2) shows good agreement in that the respective estimates show no systematic variation.

The annual meridional northward flux of relative angular momentum relative to the south pole across any latitude circle due to the frictional torque, \( F(\phi) = -\int_{\pi/2}^{\phi} T d\phi \), is probably the major part of the total flux. The residual flux at the north pole, \( F(\pi/2) \), must be balanced by mountain torques, time changes of zonal angular momentum and latitudinal redistributions of atmospheric mass (Newton 1972). Our value of \(-0.6\text{H.U.} \) is small compared with the maximum magnitude of \( F(\phi) \) and is smaller than Newton’s value of 2.3H.U.

This leads to the conclusion that the drag coefficients used by Eyre are essentially correct. It should be noted that an exact balance of angular momentum could be achieved by an increase of westerly zonal stress over the whole earth of only about \( 2 \times 10^{-4} \text{ N m}^{-2} \).

4. THE MONTHLY VARIATION OF ZONAL WIND STRESS

The monthly variation of wind stress differs markedly with latitude (Fig. 3). A half-yearly oscillation of stress dominates the variance for 50° and 60°S and is apparent also at

![Figure 3. The monthly variation of zonal wind stress at latitude deciles.](image)

* One Hadley unit (H.U.) = 10^4 kg m^2 s^-1.
Figure 4. The flux of relative angular momentum due to the frictional torque across selected latitudes in the southern hemisphere. Negative values denote southward flux.

**TABLE 2. THE SEASONAL FRICTIONAL TORQUE**

<table>
<thead>
<tr>
<th>Latitude</th>
<th>December–February</th>
<th>Seasonal frictional torque (H.U/10° latitude belt)</th>
<th>March–May</th>
<th>June–August</th>
<th>September–November</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$T$</td>
<td>$T_p$</td>
<td>$T$</td>
<td>$T_p$</td>
<td>$T$</td>
</tr>
<tr>
<td>90 N</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>80</td>
<td>-0.5</td>
<td>0.6</td>
<td>-0.4</td>
<td>0.3</td>
<td>-0.1</td>
</tr>
<tr>
<td>70</td>
<td>-2.7</td>
<td>0</td>
<td>-1.4</td>
<td>0</td>
<td>-0.2</td>
</tr>
<tr>
<td>60</td>
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<td>-3.4</td>
<td>-2.0</td>
<td>-3.2</td>
</tr>
<tr>
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<td>-17.7</td>
<td>-9.8</td>
<td>-11.0</td>
<td>-3.4</td>
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<tr>
<td>40</td>
<td>-14.3</td>
<td>-17.4</td>
<td>-7.0</td>
<td>-8.7</td>
<td>-0.4</td>
</tr>
<tr>
<td>30</td>
<td>6.4</td>
<td>6.2</td>
<td>8.0</td>
<td>3.0</td>
<td>6.8</td>
</tr>
<tr>
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<td>16.0</td>
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<td>11.9</td>
<td>9.1</td>
<td>4.1</td>
</tr>
<tr>
<td>10</td>
<td>10.7</td>
<td>10.4</td>
<td>7.7</td>
<td>4.5</td>
<td>2.1</td>
</tr>
<tr>
<td>0</td>
<td>5.8</td>
<td>5.3</td>
<td>8.9</td>
<td>8.2</td>
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<tr>
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<td>14.0</td>
<td>19.4</td>
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</tr>
<tr>
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<td>11.2</td>
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<td>9.3</td>
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</tr>
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<td>-3.1</td>
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</tr>
<tr>
<td>50</td>
<td>-10.9</td>
<td>-9.6</td>
<td>-12.6</td>
<td>-13.2</td>
<td>-10.1</td>
</tr>
<tr>
<td>60</td>
<td>1.0</td>
<td>-1.8</td>
<td>1.1</td>
<td>-1.3</td>
<td>2.9</td>
</tr>
<tr>
<td>70</td>
<td>0.9</td>
<td>2.8</td>
<td>1.4</td>
<td>1.4</td>
<td>1.5</td>
</tr>
<tr>
<td>80</td>
<td>0.1</td>
<td>0.4</td>
<td>0.1</td>
<td>0.2</td>
<td>0.1</td>
</tr>
</tbody>
</table>

$T_p$ is the seasonal frictional torque taken from Newton (1972 - Table A1) \( 1 \text{ H.U} = 10^{18}\text{kg m}^2\text{s}^{-2} \)
40°S; and possibly at 20°N, 30°N, and 50°N. The (westerly) maxima of the half-yearly oscillations occur in spring and autumn in the southern hemisphere, with greater magnitude in spring. The southern hemisphere results agree with analyses of the tropospheric geostrophic wind (Van Loon and Jenne 1970; Van Loon 1971). Our analysis shows a sharp decrease in relative importance of the half-yearly oscillation between 60°S and 50°S, and a change in phase between 50°S and 40°S. This indicates a neutral latitude somewhere between 40°S and 50°S which compares with a neutral latitude of about 50°S from Van Loon's analysis.

The flux of angular momentum due to the frictional torque $F(\phi)$ at 30°S, 40°S, 50°S, and 60°S indicates a dominant annual cycle at 30°S, a half-yearly cycle at 40°S and 50°S and almost no variation at 60°S (Fig. 4). This may be interpreted as being due to a sudden increase in the southward flow of angular momentum below 30°S in the Austral autumn to almost the same magnitude as the spring peak which is present in all the southerly temperate latitudes. The mean value of the flux of angular momentum across 30°S is −35H.U. which is close to the value obtained by Newton. However, the range of variation is somewhat larger than that obtained by Newton. The seasonal fluxes of angular momentum due to frictional torques at all latitudes may be obtained from Table 2.

5. The annual global wind stress

The global distribution of data (Fig. 5) in general has a coherent pattern of wind stress vectors on scales of order 1000km. The region of weakest annual wind stress is in the East Indies, with variable light wind stresses in the central South Pacific Ocean, and in the Sargasso Sea. A tendency for stronger winds almost parallel to coasts is apparent.

Near the antarctic continent the rather sparse data field (many wind roses are only

![Figure 5. The annual global wind stress. Stress vectors are shown for locations for which data is available for 10 or more months.](image-url)
given for some summer months in these regions) reveals the presence of two large wind gyres, with southerly outflows from the two great seas of the Antarctic, the Weddell and Ross Seas. The maximum magnitude of the annual wind stress, 0-29N m$^{-2}$ occurs at 67°S 62°E at which position very high wind stresses apparently occur throughout the year.

6. THE MONTHLY VARIATION OF WIND STRESS CURL

The wind stress curl $C$ which has the form

$$C = \frac{1}{a \cos \phi} \left( \frac{\partial \tau_x}{\partial \phi} \cos \phi - \frac{\partial \tau_y}{\partial \lambda} \right)$$

has been calculated from the wind stress data specifically for use in the modelling of large scale ocean circulations. Fig. 6 shows the curl for the months of January, April, July and October and illustrates the main features of the yearly and half-yearly wind stress cycles. The highest values, approximately 0.3 $\mu$N m$^{-2}$, occurs in the Amundsen Sea and are due to the high zonal wind stress shear in this region (Fig. 5). This feature, together with the Ross and Weddell Sea wind gyres described in section 5, gives rise to the characteristic wind stress curl patterns in the Antarctic.

In the southern portions of the ocean basins typical magnitudes are about 0.15 $\mu$N m$^{-2}$, and about 0.1 $\mu$N m$^{-2}$ in the northern portions. The central regions of the Southern Oceans show a tendency for the formation of multiple gyres which change from month to month.

7. THE OCEAN TRANSPORT MODEL

The model used is of a linear homogeneous flat-bottomed ocean, driven by a prescribed wind stress field, with constant atmospheric pressure and is fully described in Bye and Sag (1972). For a body friction law as used by Stommel (1948) one obtains a vorticity equation for the model which, on the Mercator mapping, takes the form

$$RM^2 \nabla^2 \psi + \frac{2\Omega}{a} \frac{\partial \psi}{\partial X} = M^2 \left[ \frac{\partial}{\partial Y} \left( \frac{\tau_x}{M} \right) - \frac{\partial}{\partial X} \left( \frac{\tau_y}{M} \right) \right]$$

where $M = \sec \phi$ is the map magnification factor, $R$ is the friction parameter, $\Omega$ is the angular speed of rotation of the earth, $X$ and $Y$ are the Mercator coordinates towards the east, and the north, $\psi$ is the transport stream function for transports $U$ and $V$ in the $X$ and $Y$ directions such that $U = M \frac{\partial \psi}{\partial Y}$, $V = -M \frac{\partial \psi}{\partial X}$ and $\nabla^2 \psi = \frac{\partial^2 \psi}{\partial X^2} + \frac{\partial^2 \psi}{\partial Y^2}$.

Eq. (7) has been solved for the annual mean ocean transports in several studies (Sag 1969; Ilyin et al. 1969; Takano 1969; Bye and Sag 1972). For this study ocean transports for each month were computed using a finite different equivalent of Eq. (7) with driving supplied by the wind stress curls computed from the wind stress field for each month. In the following sections the response of the homogeneous flat bottomed ocean to seasonal changes in the wind is discussed and compared with the behaviour of the real ocean.

8. THE VARIABILITY OF THE OCEAN TRANSPORTS

Examples of transport fields obtained from the solution of Eq. (7) for a friction parameter $R$ of $10^{-6}$ s$^{-1}$, and using finite difference net 2 (Bye and Sag 1972) for the months of January, April, July and October are shown in Fig. 1. A summary of the transports of the main ocean currents is given in Table 3.

The average magnitude of the circumpolar transport is 276 Mt s$^{-1}$ (megatonnes per
second). This value is somewhat lower than the value of 326Mt s⁻¹ obtained with driving by Hellerman's wind stress in Bye and Sag. The absolute values of the transports are however not our main concern in this paper, being in fact critically dependent on the choice of \( R \), and to a small extent the finite difference network. For a fixed value of \( R \) and finite difference network, we compare the variability of the derived transport with that of observations, and note the source of variability in the transport fields.

Questions of the response of the ocean to changes in wind stress cannot be answered using this steady-state model. However, it should be mentioned that the time constant for the model is \((1/2R)s\), which is about 6 days, and this causes almost complete adjustment to changes in forcing of period two months, the Nyquist frequency of the wind data.

Table 3 indicates that the variability of the ocean gyres falls into four main groups. Firstly the highest variability occurs in the Somali Current; secondly high variability occurs for the sub-polar gyres. The Brazil and Agulhas Currents show the lowest variations and the sub-tropical gyres, the Circumpolar Current, and the Flinders Current have moderate variability.

![Fig. 6 (a)](image)

![Fig. 6 (b)](image)
### TABLE 3. The Variability of the Ocean Transports

<table>
<thead>
<tr>
<th>Current</th>
<th>Average transport (1) Mt s(^{-1})</th>
<th>Maximum transport (2) Mt s(^{-1})</th>
<th>Minimum transport (3) Mt s(^{-1})</th>
<th>Variability index (2) − (3) (1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Circumpolar</td>
<td>276</td>
<td>396</td>
<td>208</td>
<td>0.67</td>
</tr>
<tr>
<td>E. Australia</td>
<td>44</td>
<td>55</td>
<td>31</td>
<td>0.55</td>
</tr>
<tr>
<td>Brazil</td>
<td>23</td>
<td>28</td>
<td>21</td>
<td>0.30</td>
</tr>
<tr>
<td>Agulhas</td>
<td>30</td>
<td>37</td>
<td>27</td>
<td>0.33</td>
</tr>
<tr>
<td>Somali</td>
<td>6</td>
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<td>1</td>
<td>3.2</td>
</tr>
<tr>
<td>Kuroshio</td>
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<td>46</td>
<td>20</td>
<td>0.90</td>
</tr>
<tr>
<td>Gyasohio</td>
<td>7</td>
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<td>Gulf Stream</td>
<td>19</td>
<td>27</td>
<td>13</td>
<td>0.74</td>
</tr>
<tr>
<td>Labrador</td>
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<td>14</td>
<td>4</td>
<td>1.4</td>
</tr>
<tr>
<td>Weddell Sea</td>
<td>27</td>
<td>49</td>
<td>11</td>
<td>1.4</td>
</tr>
<tr>
<td>Amundsen Sea</td>
<td>14</td>
<td>29</td>
<td>7</td>
<td>1.6</td>
</tr>
<tr>
<td>Flinders</td>
<td>5</td>
<td>7</td>
<td>3</td>
<td>0.80</td>
</tr>
</tbody>
</table>

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**Figure 6.** The wind stress curl. (a) January (b) April (c) July (d) October. Contour units are µN m\(^{-2}\) × 10\(^{-2}\).
From the monthly changes in the transports (Fig. 8) it is clear that the half-yearly oscillation dominates the variation of the circumpolar flow and the East Australian Current. The Agulhas Current appears to be influenced equally by the monsoonal wind system in the Indian Ocean and the sub-tropical half-yearly oscillation. The gyres in the Weddell and Amundsen Seas have an annual maximum in the Austral Winter. In the northern hemisphere the gyres all have dominant annual oscillations with possibly a trace of the half-yearly oscillation in the Gulf Stream.

9. COMPARISON OF NUMERICAL AND OBSERVED OCEAN TRANSPORTS

Data on total transports, which are necessary for meaningful comparisons, are sparse, however some comments can be made.

For the Circumpolar Current the numerical model exhibits a strong correlation between zonal wind stress in the latitude belt 50° to 60°S and the transport (Figs. 3 and 8). This implies a correlation between zonal wind stress and the north–south pressure gradient across the Drake Passage. However, observations of mean sea level differences (McKee 1971; Van Loon
1972) and density differences (Gordon and Bye 1972) in the region of the Drake Passage indicate that the pressure gradient is about 180° out of phase with the zonal wind although both exhibit a half-yearly oscillation. Furthermore, the maximum possible variability index (Table 3) based on adding the effects of observed mean sea level and density differences is found to be about half that obtained from the numerical solution. The phase and variability index differences are probably due in part to the neglect of bathymetry in the numerical model.

The maximum of the Weddell Sea gyre, which coincides with a minimum in the circumpolar transport, is impressive. It is associated with a northward transgression of the northern boundary of the gyre (The Bellinghausen Front, Gordon 1967) from about 65°S and 57°S (Fig. 7). The northward movement of this boundary may be of fundamental importance in Antarctic bottom water formation because of the following. If the stream

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Figure 7. The ocean transport stream function for $R = 10^{-6}$ s$^{-1}$. (a) January circumpolar transport 218 (Northern boundary 500, Antarctic boundary 282); (b) April circumpolar transport 325 (northern boundary 500, Antarctic boundary 175); (c) July circumpolar transport 241 (northern boundary 500, Antarctic boundary 259); (d) October circumpolar transport 396 (northern boundary 500, Antarctic boundary 104). Contour and boundary value units are $\text{Mts}^{-1} \times 10^{-1}$. 
function value of a particle of water passing through the Drake Passage is less than that for the boundary of the Antarctic continent the particle will be entrained in the Weddell Sea gyre. Otherwise it continues its circumpolar ambulation. Thus the rapid increase in the intensity of the Weddell Sea gyre between April and May and to a lesser extent between November and December is due to entrainment of water mass (Fig. 7). This water mass would be subsequently modified within the gyre and detrained as Antarctic bottom water.

The East Australian Current is known to be a very variable stream (Boland and Hamon 1970). Transports computed from pairs of stations relative to 1300db (approximately 1300m) near 33°S off the Australian coast range from 27 to 51Mt s⁻¹. The seasonal variation of the five reported values (Fig. 8) is not inconsistent with a half-yearly oscillation. The magnitudes of the observed and numerical transports are in fair agreement. The somewhat lower observed values may be due to the shallow reference level. The maximum of the East Australian Current shifts latitudinally with season between 24° and 34°S. (Fig. 7).

The Agulhas current was estimated to have a magnitude of 45Mt s⁻¹ in August/September 1964 (Harris 1972). The numerical maximum value 37Mt s⁻¹ is similar. An important feature of the Indian Ocean circulation is the division of the South Equatorial water between the Somali and Agulhas Currents (Fig. 7). It is clear that in the Austral winter the South Equatorial Current is at a maximum giving rise to maxima in both the tributary streams (Fig. 8). It is observed that the Somali Current responds rapidly to the onset of the south-west monsoon which usually occurs in the central Indian Ocean in mid or late May (Warren, Stommel and Swallow 1966). The response is clearly indicated by the numerical solutions (Fig. 7) and has been studied elsewhere in a time-dependent model of the Indian Ocean (Cox 1970).

South of Australia, the Flinders Current (Bye 1968; Veronis 1973) responds to the

![Diagram](image)

Figure 8. The monthly variation of transports of the major ocean currents. Circles indicate transports, based on observations.
local wind stress curl, and interacts with the current gyres in the Indian and Pacific Oceans, and with the Circumpolar Current. The monthly transport fields indicate two basic configurations (Fig. 7). Firstly the outflow to the west, south of Western Australia, consists solely of water recycled from the circumpolar flow (May–December). Secondly, the outflow is maintained by a flow from the Pacific Ocean south of Tasmania (January–April). The maximum southerly excursion of the East Australian Current occurs during these months. In the first case, a closed eddy also exists south of Australia from May to September. The outflow, which has a half-yearly period, and the closed eddy if it exists, are identified with the Flinders Current (Table 3). The transports of the numerical model for this current are about half of those observed relative to 2000db (Fig. 8).

The variation of the transport of the Gulf Stream through the Straits of Florida has perhaps been the most extensively observed of all ocean transports. Points on the curve of best fit to the annual harmonic obtained from fitting total transports observed between 1964 and 1970 (Niiler and Richardson 1973) are shown for each month in Fig. 8. There is lack of agreement in seasonal variation between the best fit curve and the numerical results but the annual mean of the numerical transports is close to that of the observations.

There appears to be no observational data on the total transports of the other northern subtropical gyres with which to compare the numerical estimates.

10. Conclusion

The results of this paper seem to be as far as the analysis of existing climatic wind data over the ocean can be taken. It is encouraging that the assumed drag law appears to yield a reasonable global atmospheric angular momentum balance. Further advances await the results of the programmes of direct measurement of wind stress in the deep sea from meteorological buoys.

On the oceanographic front, the problem of total transport determination along a number of key sections requires very extensive (and expensive) deep-sea programmes. It would appear that the monitoring of the Circumpolar Current through the Drake Passage should have a high priority in such measurements. Of the southern hemisphere gyres the Agulhas Current appears to be best suited for a transport section, and in the northern hemisphere, a section across one of the polar gyres should yield some interesting results.

The thermohaline circulation, associated with the density structure of the ocean, can be computed from more sophisticated ocean models, using in addition to the wind stress, forcing due to the exchange of heat and mass across the sea surface. Such computations, which would be amenable to comparison with existing hydrographic data, are at present being carried out by the authors and it is hoped to report on them separately in the future.

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