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Seasonal variation in meridional overturning and poleward heat transport in the Atlantic and Pacific Oceans: a model study

by Kirk Bryan

ABSTRACT

Numerical solutions for a model of the World Ocean are analyzed to illustrate the important changes in the meridional circulation of the ocean induced by seasonal changes in wind stress. Seasonal variations of meridional circulation and associated changes in poleward heat transport predicted by the model are most important in the Pacific where the background thermohaline circulation is relatively weak. Numerical experiments show that seasonal variations of cross-equatorial heat transport basically depend on the variations of the zonal component of the wind stress. Seasonal variations of the meridional wind stress exert an influence on cross-equatorial heat transport which is 180° out of phase in the seasonal cycle from the more dominant influence of seasonal variations of the zonal wind stress.

1. Introduction

One of the author’s first exposures to oceanography was a cruise aboard the RV Crawford with L. V. Worthington. A naive prediction of the local surface currents based on Ekman drift was greeted by Worthington with jovial, but profound skepticism. The skeptical view was confirmed when we attempted a few brief measurements with drifters in the vicinity of the mid-Atlantic Ridge. Worthington was aware of the dismal history of attempts by Ekman and Helland-Hansen (1931) to confirm the Ekman spiral by direct measurements at sea. It is only recently that enough measurements have accumulated to allow us to conclude that Ekman and Helland-Hansen’s measurements were a classic failure in experimental design. The Ekman drift was simply too weak a signal to be detected easily in a noisy background.

A parallel problem now exists on a planetary scale. In a few regions the response of the ocean to seasonal changes in the wind is easy to see in standard data. For example, the dramatic reversal of currents along the coast of Somali in response to the monsoon was well known to early navigators. In other areas the response to winds on a large scale is more subtle. It is difficult to establish cause and effect

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without rather elaborate monitoring networks, which are still in the planning stages. On the other hand, computer models allow rather detailed predictions of the response of the equatorial ocean circulation to large-scale variations in the wind. With predictions based on reasonably sound physical principles, but without the means to make a field verification, modelers share the frustrations of Professor Ekman.

In this paper we describe predictions for wind-driven changes in the ocean circulation which illustrate some of the difficulties of verification noted above. The predictions are based on a numerical model of the World Ocean (Bryan, 1979; Bryan and Lewis, 1979; hereafter referred to as BL). The original papers only describe the zonally averaged, global, meridional circulation and heat transport. Here we will discuss the meridional overturning and heat transport in individual ocean basins. Although heat storage data suggest that important redistributions of heat take place in the upper ocean in tropical areas, the data are too sparse in most areas for detailed verification. The motivation for presenting the results are plans that are now being made to monitor heat content of the ocean over wide areas, and possibly to measure poleward heat transport directly in several ocean basins. Since the physical basis of the model is given in the original BL paper, a very brief description of the model will suffice.

The World Ocean is divided into a large number of cells. In the low resolution case the cells were approximately 500 km on a side. In the high resolution case the cell size was reduced to approximately 250 km. In the deeper parts of the ocean the cells have a thickness of nearly 1 km, but the thickness is gradually decreased upward to resolve the main thermocline. Cells at the upper surface of the ocean have a thickness of only 50 m. The equations of motion, the continuity equation and appropriate conservation equations are used to predict potential temperature and salinity in each rectangular cell, and velocity at the corners of each cell. Transfer of water mass properties between cells is a function of the explicitly calculated flow and diffusion. The diffusion parameters are specified in Table 1. The vertical diffusion coefficients given in Table 1 refer to the stably stratified case only. Where the stratification is predicted to be locally unstable, vertical mixing is assumed to be effectively infinite and the temperature and salinity fields are adjusted accordingly. The solutions of the model are determined solely by upper boundary conditions specified at the surface from observations. Wind stress is taken from tables prepared by Hellerman (1967). Rayleigh damping with a time constant of 10-20 days forced the temperature and salinity of the surface cells toward observed values given by data from Levitus and Oort (1977). The boundary condition for both wind stress and water mass properties included only the first annual harmonic of the seasonal variation. In retrospect, a more detailed representation of the seasonal cycle in boundary conditions would have been desirable. This has been done in a more recent study of a similar model by Meehl et al. (1982).
Table 1. Important constants used in a global ocean model by Bryan and Lewis (1979). Where upper and lower values are indicated, a smooth analytical function of depth connects values in the upper thermocline with those indicated for deep water. The vertical viscosity, $A_{VM}$, has a uniform value of $10^{-2}\text{m}^2\text{s}^{-1}$.

<table>
<thead>
<tr>
<th>Name</th>
<th>Symbol</th>
<th>Low resolution $(\text{m}^2\text{s}^{-1})$</th>
<th>High resolution $(\text{m}^2\text{s}^{-1})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>E-W spacing</td>
<td>$\Delta\lambda$</td>
<td>$\pi/32$</td>
<td>$\pi/64$</td>
</tr>
<tr>
<td>N-S spacing</td>
<td>$\Delta\phi$</td>
<td>$\pi/38$</td>
<td>$\pi/76$</td>
</tr>
<tr>
<td>Lateral diffusion</td>
<td>$A_{HH}(\text{upper})$</td>
<td>$2.5 \times 10^3\text{m}^2\text{s}^{-1}$</td>
<td>$10^3\text{m}^2\text{s}^{-1}$</td>
</tr>
<tr>
<td>coefficient of heat</td>
<td>$A_{HH}(\text{lower})$</td>
<td>$0.5 \times 10^8$</td>
<td>$0.5 \times 10^8$</td>
</tr>
<tr>
<td>and salt</td>
<td>$A_{HM}$</td>
<td>$8.0 \times 10^6$</td>
<td>$10^6$</td>
</tr>
<tr>
<td>Lateral viscosity</td>
<td>$A_{LM}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vertical diffusion</td>
<td>$A_{VH}(\text{upper})$</td>
<td>$0.3 \times 10^{-4}$</td>
<td>$0.3 \times 10^{-4}$</td>
</tr>
<tr>
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</tr>
<tr>
<td>and salt</td>
<td></td>
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</tr>
</tbody>
</table>

A straightforward way to determine the equilibrium solution of such a model is an extended numerical integration with respect to time until a statistically steady state is reached. However, it is possible to reach a steady state much more rapidly by a relaxation procedure using an asynchronous numerical integration for different levels and different variables in the model. Details are given in BL and Bryan et al. (1975).

2. Seasonal changes in meridional circulation

The zonally averaged winds from 90N to 10S taken from Oort and Rasmusson (1971) are shown in Figure 1. The upper panels show the zonally averaged seasonal east-west wind component and the lower panels the meridional component. From the standpoint of ocean response, we are interested in winds at the surface. Note that in winter both the northern hemisphere easterlies and westerlies intensify, while the southern hemisphere easterlies weaken. Thus seasonal changes in the zonal component of the wind are antisymmetric across the equator. On the other hand, seasonal changes in the meridional component of the wind are symmetric across the equator. The difference between zonally averaged surface stress between January and July from data by Hellerman (1967) is shown in Figure 2a. The seasonal changes in zonal stress are asymmetric across the equator, but the implied seasonal change in total meridional Ekman mass transport is,

$$\delta T_{BK} = -\int_0^{2\pi} (\delta r^2/2\Omega \sin \phi) a \cos \phi d\lambda$$

(2.1)
Figure 1. Zonally averaged winds from Oort and Rasmusson (1971). Upper panels are the eastward component in m/s. Lower panels are the northward component.

Figure 2. a) The difference between January and July zonally averaged east-west wind stress over the ocean from Hellerman (1967), (1 Pa = 1 pascal = 1 newton/m² = 10 dynes/cm²).
The zonal average of the argument of the integral on the right-hand side of (2.1) is shown in Figure 2b. \( \delta \tau^x \) is the seasonal change in the zonal component of the wind stress over the oceans, \( \lambda \) and \( \phi \) are longitude and latitude respectively, and \( \Omega \) is the angular velocity of the earth. \( a \) is the radius of the earth. Note that \( \delta T_{RK} \) is symmetric due to the \( \sin \phi \) factor in the denominator of the integral.

In order to examine the changes in meridional circulation associated with seasonal variations of wind stress in the BL model we must first define a transport stream function for the zonally averaged flow. Let \( v \) and \( w \) be the meridional and vertical components of velocity, respectively.

\[
\int_{\lambda_2}^{\lambda_1} \rho \upsilon \ a \cos \phi \ d\lambda = -\partial_z \psi
\]

\[
\int_{\lambda_2}^{1} \rho \omega \ a \cos \phi \ d\lambda = a^{-1} \partial \psi
\]

\( \lambda_1 \) and \( \lambda_2 \) are the longitudes of east and western boundaries of the basin. If the basin is enclosed by meridional walls, the zonally integrated flow is nondivergent and a transport stream function is well defined.
Figure 3. Transport stream function for the zonally integrated flow from the BL model for the Pacific. Isolines are drawn for intervals of 10^9 kg/s.

The January and July patterns of Pacific meridional transport for the high resolution experiment III of BL are shown in Figure 3. A comparison of the patterns shows that seasonal effects are very strong, relative to the Pacific mean annual circulation. For the real ocean, interannual variations of the wind stress can be as strong as seasonal variations, and this is particularly true in the Pacific. Thus, if the meridional circulation shown in Figure 3 could actually be measured, we would expect to see response to large-scale interannual variability of wind patterns associated with the southern oscillation (Horel and Wallace, 1981) as well as seasonal variations. Figure 3 shows that in July, the dominant surface flow in the tropics is from the northern to the southern hemisphere, balanced by a deep recirculation. Note that the deep recirculation is not evenly distributed with depth indicating some baroclinic adjustment near the equator. Theory suggests that far from the equator, the return flow would be uniform with depth because the annual period is too short to allow baroclinic adjustment.

According to the model the seasonal changes in the meridional circulation of the Atlantic are much less. The patterns of meridional circulation shown in Figure 4 indicate that the thermohaline circulation plays a much larger role in the Atlantic
relative to the Pacific. It is somewhat surprising to find that the strength of sinking motion in the subarctic gyre of the North Atlantic is actually weaker in January than in July. One would think that stronger water mass formation in winter would cause stronger sinking. The model, however, shows that the extra upwelling in the North Atlantic caused by stronger westerlies, more than offset the extra cooling in winter. At lower latitudes the meridional circulation is dominated by southward flow at the depth of 1 km compensated by northward flows at the surface and near the bottom. This pattern agrees with the classical picture derived from water mass analysis (Worthington, 1976).

In an analysis of IGY data and direct transport measurements in the Florida Current at 24N, Bryden and Hall (1980), Hall and Bryden (1982) and Roemmich (1980) estimate that the strength of meridional overturning is approximately 18 megatons/s ($18 \times 10^9$ kg/s). Since the 10 megaton/s contour in Figure 4 does not extend south of 30N, the model indicates a much weaker meridional circulation in the North Atlantic. If the measurements at 24N turn out to be representative, this is a very serious discrepancy. We will return to this point in connection with poleward heat flux in the next section.
Figure 5. Same as Figure 3 for the World Ocean.

For completeness the January and July patterns for the entire World Ocean are shown in Figure 5. These patterns include the circulations shown in Figures 3 and 4 plus the contribution by the Indian Ocean. Since annual variations in zonal wind stress are not very strong in the Southern Ocean, and semi-annual variations in wind stress are not included in BL, the patterns of meridional circulation are nearly the same for January and July near the Antarctic continent. Sinking predominates adjacent to the continent with a broad upwelling further north supporting the Antarctic halocline. The powerful cell centered at 60S consists of surface Ekman transport toward the equator under the southern hemisphere westerlies, compensated by deep return flow. This feature remains about the same in the January and July patterns. As pointed out by BL, the existence of this cell is a powerful influence on the climate of the southern hemisphere. It has the effect of holding back the poleward flow of warmer surface waters, helping to maintain the strong thermal gradient of the Antarctic Circumpolar Current.

3. Seasonal changes in Atlantic and Pacific heat transport

Let \{ \} denote an integral over an entire latitudinal wall across an ocean basin, and [ ] indicate a zonal average across the same basin. Let \( H \) be the total depth.
\[ \{v\} = \int_{\lambda_1}^{\lambda_2} \cos \phi \, d\lambda \, dz \]  

(3.1)

and

\[ \{\theta\} = -\frac{1}{\lambda_2 - \lambda_1} \int_{\lambda_1}^{\lambda_2} \cos \phi \, d\lambda \]  

(3.2)

In cases where the basin is divided by a mid-ocean ridge, (3.2) becomes more complicated. For a basin closed at one end, net flow across a latitude is approximately zero, so that

\[ \{v\} = 0 \]  

(3.3)

In that case the poleward heat transport may be written

\[ HT = \{\rho_c c_p \, v \theta\} \]  

(3.4)

where \(c_p\) is the specific heat at constant pressure and \(\theta\) is the potential temperature in °C (Bryan, 1962). If we write,

\[ v = [v] + v^* \]

\[ \theta = [\theta] + \theta^* \]  

(3.5)

The poleward heat transport in the BL model at any latitude may be written,

\[ HT = \rho_c c_p \{[v] [\theta] + [v^* \theta^*] + A_{HH} \rho \theta \} \]  

(3.6)

The three terms on the right-hand side of (3.6) may be interpreted as the effect of overturning in the meridional plane, the effect of correlations in the horizontal plane, and a horizontal diffusion term. The last term is a crude representation of mesoscale eddies in the BL model.

It is shown in BL that the second and third terms on the right-hand side of (3.6) change very little with season. Almost the entire seasonal variation is due to the Ekman transport changes associated with the meridional overturning term. The second and third term depend on large-scale features of the density structure, which can only adjust on a very long time scale at middle and high latitudes. On the other hand, the Ekman transport can respond within one inertial period.

The seasonal changes in poleward heat transport in the Pacific (Fig. 6a) are much larger than those shown for the Atlantic in Figure 6b. In the equatorial Pacific the poleward heat transport varies from about +1 PW in January to −1 PW (10^{13} W) in July. At higher latitudes, the seasonal changes are just the opposite of that in the tropics. The difference is what one would expect from the meridional circulation patterns shown in Figure 3. In winter the increased strength of the westerlies forces surface waters to the south, while the increased strength of the easterlies forces warm surface waters to the north. The result is a strengthening of the tendency for
convergence in the subtropical gyre region in winter relative to summer in each hemisphere.

The same tendency is shown in Figure 6b in the Atlantic, but the poleward heat transport is dominated by the thermohaline circulation which is relatively constant
Table 2. The sensitivity of global heat transport at 24N in the Bryan and Lewis (1979) low resolution model. \( HT_0 \) for the standard low resolution experiment is 0.6 PW.

<table>
<thead>
<tr>
<th>Name</th>
<th>Symbol</th>
<th>( \frac{\partial HT}{\partial \ln(A_i/A_{i0})} ) (10^{15}W)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lateral diffusion coefficient of heat and salt</td>
<td>( A_{HH} )</td>
<td>0.031</td>
</tr>
<tr>
<td>Lateral viscosity</td>
<td>( A_{HM} )</td>
<td>-0.028</td>
</tr>
<tr>
<td>Vertical viscosity</td>
<td>( A_{VM} )</td>
<td>0.056</td>
</tr>
<tr>
<td>Wind stress amplitude</td>
<td>(</td>
<td>\tau</td>
</tr>
</tbody>
</table>

throughout the year. As shown in Figure 4, the thermohaline circulation is dominated by a net northward movement of surface waters compensated by a return flow at the base of the thermocline. As mentioned previously, direct estimates from a single IGY section at 24N by Bryden and Hall (1980), and Roemmich (1980) indicate a meridional circulation almost twice as intense as that in the model. It is not surprising that the model heat transport is only about 50% of the 1.1 PW estimated by Bryden and Hall (1980), Hall and Bryden (1982), and Roemmich (1980).

An attempt is made in BL to determine the sensitivity of poleward heat transport in the model to specified parameters. To illustrate the results consider a Taylor expansion in parameter space representing heat flux at a particular latitude. Dropping higher order terms,

\[
HT = HT_0 + \sum_i \frac{\partial HT}{\partial \ln(A_i/A_{i0})} \frac{\partial \ln(A_i/A_{i0})}{\partial \ln(A_i/A_{i0})} (3.7)
\]

\( HT_0 \) is the poleward heat flux for a standard experiment, and \( A_i/A_{i0} \) is the ratio of a given parameter \( A_i \) to its value in the standard experiment. In BL a series of numerical experiments were carried out with the low resolution model. Values of the parameters are shown in Table 1. Sensitivity values estimated from the different experiments are shown in Table 2. The numbers in Table 2 represent the increase or decrease in heat transport in PW (10^{15}W) corresponding to an e-fold increase or decrease of a parameter relative to its standard value.

It is obvious from Table 2 that heat transport at 24N is most sensitive to the amplitude of the wind stress. An e-fold increase in wind stress will double the heat transport, while an e-fold change in the other parameters has a negligible effect. However, the results for the high resolution case are not consistent with (3.7) and Table 2. In the high resolution case (see Table 1) the horizontal viscosity was decreased by a factor of 8, and the horizontal diffusivity by a factor of 2. Equation (3.7) would predict an increase of poleward heat flux of only 0.04 PW at 24N. In
fact heat flux increased from 0.6 PW to 0.9 PW, suggesting the sensitivity results in Table 2 cannot be reliably extrapolated to regimes of lower mixing levels and more work is needed to determine the full potential of the model to simulate ocean heat transport.

The higher northward heat transport in summer in the North Atlantic subtropical gyre indicated in Figure 6b implies that the seasonal heat storage in the North Atlantic should be higher than can be accounted for by local seasonal heating during the summer season. This effect has already been noted from examination of the North Atlantic heat storage data by Bryan and Schroeder (1960).

4. Seasonal wind-driven currents in an ocean covered globe

In the preceding sections there has been no real explanation of the wind-induced seasonal changes in meridional circulation aside from an appeal to Ekman transport. In this section we will analyze the effect of seasonal wind changes in a little more detail extending the work of Schopf (1980). Schopf illustrated with a very simple linear model how seasonally varying meridional Ekman transport regime could be joined across the equator by a forced gravity wave.

To derive the model two drastic assumptions are made. In the first place, we assume an ocean covered globe. This implies that boundary currents do not play a primary role in the seasonal changes of the meridional circulation. The second assumption is that nonlinear effects can be neglected. Neither assumption can be rigorously defended. Let us assume a uniformly stratified ocean covering the entire globe, driven by regular seasonally varying wind stresses. Writing the velocity and pressure in vertical modes (Moore and Philander, 1977) and the first annual harmonic,

\[ u, v, p/\rho_o = \Sigma (U, V, P)^n Z^n \exp(i\omega t) \]  

(4.1)

where \( Z^n \) is the shape function of the \( n^{th} \) vertical mode and \( \omega \) is equal one cycle per year. We will also assume that the combined effects of friction and diffusion act like Rayleigh damping with a time constant much less than the annual period. If the inverse time constant of Rayleigh damping is designated as \( \kappa \), \( \omega \ll \kappa \). The linearized equations of motion for each vertical mode may then be written

\[
\begin{pmatrix}
\kappa - f \\
f & \kappa
\end{pmatrix}
\begin{pmatrix}
U^n \\
V^n
\end{pmatrix}
= 
\begin{pmatrix}
X^n \\
Y^n
\end{pmatrix}
- \frac{a^{-1}}{\partial \phi} P^n + Y^n
\]

(4.2)

\( f \) is equal to \( 2\Omega \sin \phi \), and \( X^n \) and \( Y^n \) are the projections of the east-west and the meridional component of the wind stress on the \( n^{th} \) vertical mode. Both \( X^n \) and \( Y^n \) are assumed to be independent of longitude. The pressure equation is,

\[
kP^n + \frac{gh^n}{a \cos \phi} \partial \phi (V^n \cos \phi) = 0
\]

(4.3)
Figure 7. January minus July poleward heat transport by ocean currents in the BL model. Upper curve is for the case in which the seasonal variation of the $\tau^x$ stress is suppressed. The lower curve is the case for which the seasonal variation of the $\tau^x$ stress is suppressed.

$h^x$ is the equivalent depth. Eqs. (4.2) and (4.3) may be combined to form a single equation in $V^x$,

$$-(\kappa^2 + f^2) V^x + \left(gh^x/a^2\right)\partial_\phi (\cos \phi - \partial_\phi \cos \phi V^x) = f X^x - \kappa Y^x$$  \hspace{1cm} (4.4)

As pointed out by Schopf (1980) the third term on the left-hand side is only important in a narrow region near the equator. Several degrees from the equator where $f \gg \kappa$, (4.4) becomes approximately,

$$f^2 V^x \approx -f X^x + \kappa Y^x$$  \hspace{1cm} (4.5)

A balance between the left-hand side and the first term on the right is simply the classical Ekman drift for an individual vertical mode.

Returning to the Oort and Rasmusson (1971) wind data in Figure 1 we see that the zonally averaged meridional wind at the surface tends to blow from the winter to the summer hemisphere. This is consistent with a positive correlation between $fX^x$ and $Y^x$. Strong easterlies in the northern hemisphere are associated with net southward flow and vice-versa. Thus the two terms on the right-hand side of (4.5) would tend to oppose one another in the annual cycle. To test this effect in a more complete model, the BL standard case with low resolution (Experiment II) was extended with a modification in the seasonal component of the surface wind stress. First the seasonal variation of the $\tau^x$ component, and then the seasonal variation of the $\tau^y$ component was suppressed. January minus July poleward heat transport for the BL model is shown in Figure 7. Suppressing the $\tau^y$ seasonal variation
enhances the summer to winter hemisphere heat transport. Suppressing the \( r^z \) component completely reverses the cross-equatorial seasonal transport, so that heat goes from the winter to the summer hemisphere. Examining (4.4) we can see that the effect of the \( r^z \) component depends fundamentally on the Earth's rotation. On the other hand, the effect of the \( r^v \) component on cross-equatorial transport could exist without rotation, although the amplitude would be somewhat modified.

The results of the BL model appear to be in qualitative agreement with the simple theory for an ocean covered earth. The \( r^v \) seasonal variations suppress the cross-equatorial seasonal heat transport, but the effect is too weak to overcome the fundamental driving effect of the seasonal variations of the \( r^z \) component of wind stress. Figure 7 also shows that \( r^z \) and \( r^v \) affect the heat transport in additive fashion in agreement with linear theory.

5. Discussion

Oort and Vonder Haar (1976) have made a pioneering contribution to the study of the ocean's role in climate by combining data from satellites, heat storage data from the ocean, and atmospheric radiosonde data. By combining all these different data sources the authors were able to compute the seasonal changes in ocean heat transport across latitude circles as a residual term. Oort and Vonder Haar (1976) found that a very large ocean heat transport from the summer to the winter hemisphere was required to maintain heat balance. The modeling studies of Webster and Lau (1977), BL, and more recently Meehl et al. (1981) all illustrate how the seasonal variations of the wind could produce such an effect. The original measurements, however, are so indirect, and the error bars are so large that no satisfactory quantitative comparison between observations and models can be made at this time. Confirmation must await more direct measurements through large-scale monitoring. The proposed satellite scatterometer for measuring surface wind stress and the altimeter for measuring sea surface elevation appear to be very promising instruments for providing data for testing model predictions.

The solutions shown in this study predict that the behavior of the Atlantic and Pacific in response to seasonal changes in the wind are quite different. In the Atlantic seasonal changes in the winds only slightly modify a vigorous overturning associated with the thermohaline circulation. In the Pacific, however, the thermohaline circulation is weaker. Therefore, the seasonal changes caused by the wind appear to be dominant. Since interannual changes in the wind pattern associated with El Niño are almost as strong as the seasonal changes in the Pacific, one might anticipate that there will be long term (1-10 yr) fluctuation in deep-sea circulation not very different from the seasonal changes predicted by the model. Thus many years of measurements may be required to obtain representative estimates of the deep-sea circulation in the Pacific.

Interest in the climate problem, an increasing amount of data from geochemical
tracers, and direct current measurements will motivate further modeling efforts which attempt to simulate water mass formation and the ocean circulation. The best data set exists for the North Atlantic Oceans, thanks in no small part to the efforts of L. V. Worthington. The North Atlantic is therefore a logical starting point for developing models.

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