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The shallow wind driven overturning cell in the Indian Ocean

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KNMI: Koninklijk Nederland Meteorologisch Instituut

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Introduction of the laboratory

The KNMI: Royal Netherlands Meteorological Institute is an agency of the Ministry of Transport and Public Works.
It was founded in 1854. The head office is located in De Bilt (near Utrecht) and there are also regional offices strategically located at airports and at points along the North Sea Coast.
It fulfills the following public tasks and responsibilities:

- -Weather forecasts and weather alert
- -Climate monitoring
- -Collecting and providing meteorological data and the related infrastructure
- -Model development
- -Aviation Meteorology
- -Scientific Research
- -Public information

The scientific research department is divided in three sections:

1-Applied research
2-Climate research
3-Seismology research

Climate research is carried out in different departments:

- -Oceanographic research
- -Atmospheric research
- -Climate Variability Research
- -Atmospheric Composition
- -Climate Analysis
- -Climatological Services

I worked in the Oceanographic research division supervised by Sybren Drijfhout and Wilco Hazeleger.

Introduction

In the Atlantic and Pacific Ocean, subtropical shallow overturning cells have been observed on both side of the equator. Equivalent cells exist in the Indian Ocean. But they must be quite different because of the asymmetry of the Indian Basin and the extreme seasonal variations. Therefore, we investigate here the detailed picture and mechanism of these surface cells.

The subtropical cells play an important role in the mean circulation of all three tropical oceans (Pacific, Atlantic and Indian Ocean), and may play an important role in climate variations.

The objective of this training period is to picture the subtropical, or cross equatorial cells in the Indian Ocean and to visualize in three dimensions the pathways of those cells.

To this end, we used as dataset output from a primitive equation numerical model of the global ocean. We applied a particle tracking algorithm to the model's velocity field.

In the first part, we will introduce some useful concepts to understand the problem. Then, we will describe the methodology of this study, and finally, we will present our results and discuss them.

1 Generals points

a-Atmospheric circulation

The atmosphere and the ocean influence one another. To understand the surface currents in the ocean, we need to know the surface winds in the atmosphere. Figure1 shows a schematic of the circulation cells in the atmosphere. The tropical Indian Ocean circulation is influenced by winds near the equator. Consequently, it is influenced by the lower branch of the Hadley cell. The associated winds are equatorward. The Coriolis force tends to deflect wind to the right in the northern hemisphere, and to the left in the southern hemisphere.

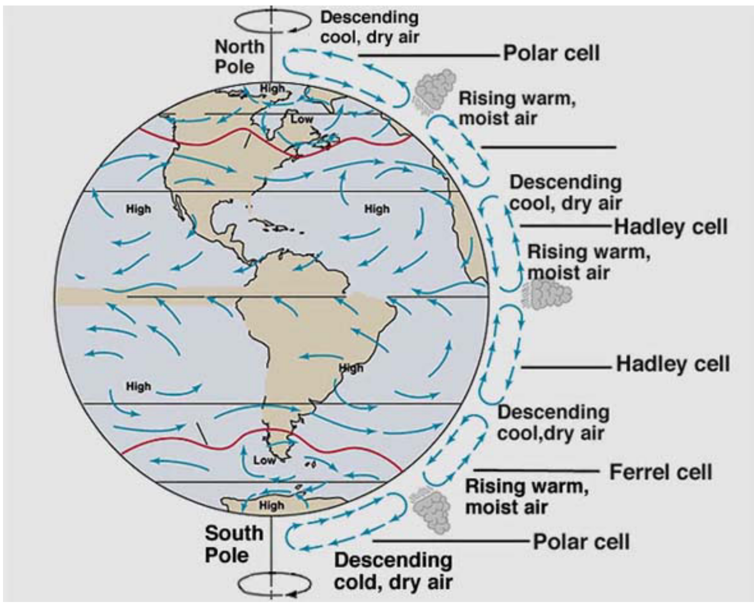


Figure.1: Circulation cells in the atmosphere

Figure2 shows the winds at the Earth's surface. We can see the winds of the lower branch of the Hadley cells that are called the Trades winds. The area where these winds strongly converge is called the ITCZ (InterTropical Convergence Zone).

If we compare Figure2a and Figure2b, we see an extreme seasonal variation over the Indian Ocean. This variation is due to asymmetry of the Indian basin and the importance of land in the northern hemisphere. In the northern winter, the Asia's land mass is much colder than the ocean. The air above the land is cooler and denser than the air over the ocean. It creates a large high pressure zone over southern Asia. The resulting pressure gradient leads to a northerly or north-easterly flow of air from Asia to the equator. This is the North-East monsoon. During the northern summer, the land is warmer than the ocean. It creates a low pressure zone over Tibet generating a south-westerly flow. This is the south-west monsoon that is much stronger than the north-east monsoon. As a result, the ITCZ position is very different in both seasons.

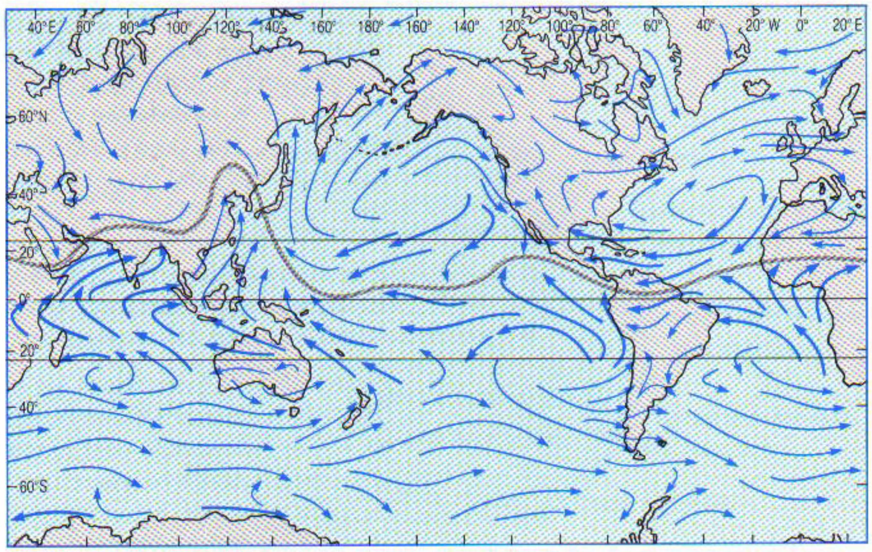


Figure2a: Prevailing winds at the Earth surface during northern summer

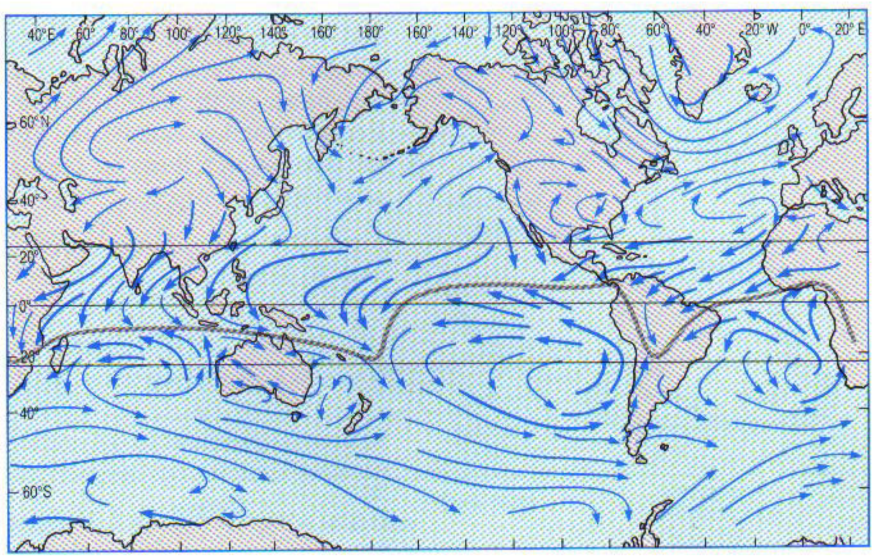


Figure2b: Prevailing winds at the Earth surface during northern winter

b-Ekman transport and surface currents

When wind blows over the ocean, the frictional force acting on the sea surface is known as **the wind stress**. It depends on the wind speed, and the roughness of the sea surface. In the equations of the motions, this force has only importance in a surface layer called the **Ekman Layer**. In this layer of a few tens of meters, motion is due to the balance between the wind stress and the Coriolis force. The effect of the wind stress at the surface is transmitted downward as a result of internal friction within the Ekman layer. It decreases exponentially with depth. The current vectors therefore form a theoretical spiral pattern known as the **Ekman spiral**. The spiral is shown in figure3. This figure is valid for the northern hemisphere. In the southern hemisphere, the sign of the Coriolis force changes, therefore the spiral turns in the opposite way. The average motion of the Ekman layer, gives a transport that is 90 degrees on the right of the wind direction in the northern hemisphere, on the left in the southern hemisphere.

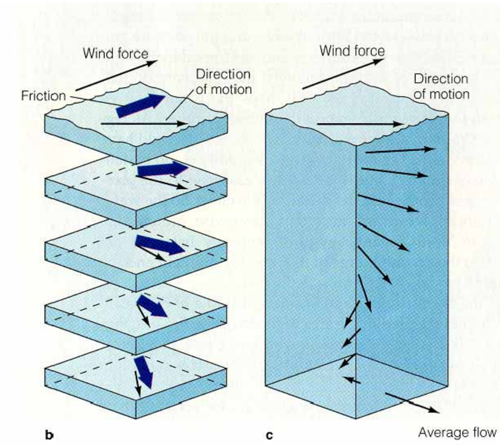
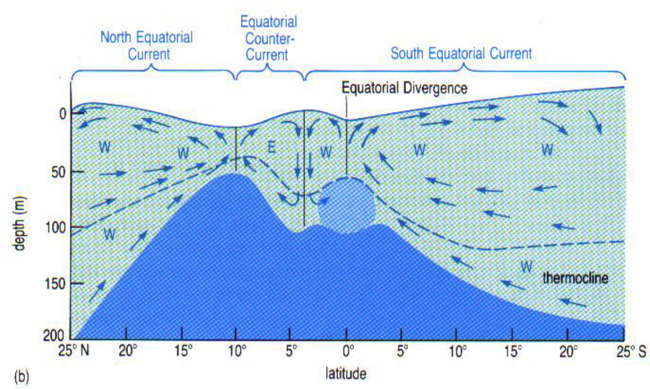


Figure3: Ekman Spiral

c-Subtropical cells

As we saw from the Ekman transport, it is clear now that the wind influences surface currents. In the Atlantic and Pacific Ocean, the situation is quite clear. Trade winds generate an Ekman transport northward in the northern hemisphere and southward in the southern hemisphere. Those transports are the upper branches of the subtropical cells. Consequently, there is a divergence near the equator. This divergence creates an upwelling in this region (which is the vertical branch of the cells). There are downwelling sites in the subtropical gyres. Finally, an equatorward subsurface flow in the thermocline closes the cells.

Figure4 summarizes those phenomena and shows subtropical cells (that are on the right and left side of the figure). Between those cells, we can see two little cells north of the equator. This is due to the position of the ITCZ that is often north of the equator. The subtropical cells have been observed in the Pacific and Atlantic Ocean. In both basins, there are two subtropical cells, one north of the equator, and one south of it.



Figule4: Schematic of wind driven overturning cells

In the Indian Ocean, the situation is very different due to the presence of monsoons and land-sea distribution.

d-The Indian Ocean

The surface circulation in the northern part of the Indian Ocean changes seasonally, in response to the monsoon.

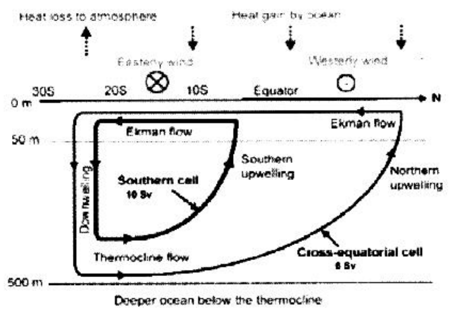


Figure5: Schematics of the shallow meridional circulation of the Indian Ocean, extract from T. Lee *et al.* poster.

During the northern summer, South-east Trades dominate over the Indian ocean. After crossing the equator, they turn to the right because of the Coriolis force. Therefore north of the equator, the winds have an eastward component, and south of it, they have a westward component. As a result, they drive southward Ekman transport on both sides of the equator. Thus, the southern subtropical cell is much stronger, and becomes a cross equatorial cell. You can see in figure6 a hypothetic scheme of this cell. The northern subtropical cell is very weak, or even may not exist.

During the northern winter, the situation is quite similar to the other ocean basins. Trade winds tend to generate a divergence under the ITCZ. However, the ITCZ is situated in the southern hemisphere, near 10S. Because of the Coriolis force, winds have a westward component north of the equator and an eastward component south of it. Thus, the Ekman transport is directed northward on both sides of the equator. The cross equatorial cell may turn in the other direction, and the northern subtropical cell may be more important than during summer.

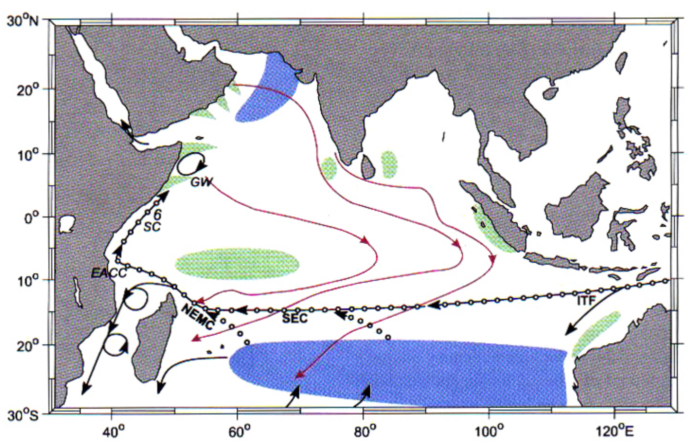


Figure6: Schematic representation of the Indian Ocean Cross Equatorial Cell with subduction (blue) and upwelling (green) zones that participate in the CEC. It has been extracted from Schott *et al.*(2004). It also shows model surface trajectories (magenta) of southward CEC return flow originating from upwelling sites off Somalia, Oman and west of India.

Several studies have been made on this topic. Some of them are referenced in the bibliography at the end of this report.

According to those studies, there is no annual mean equatorial upwelling as in the other ocean basins. Instead, there are several upwelling regions in the northern hemisphere. Those upwelling regions occur near Somalia, Oman, India, and Sumatra (green in figure6). Satellite color images suggest the existence of an open-ocean upwelling zone from 5S to 12S, but direct evidence of this upwelling is still lacking.

This hypothetic upwelling region points toward the existence of a second overturning circulation that remains confined to the southern Indian Ocean. According to Miyama *et al.* (2003), sources of water for the subsurface branch of the cross equatorial cell are subduction in the southeastern Indian Ocean, the Indonesian Throughflow, and flow into the basin across the southern boundary.

In Figure6, subduction seems to occur predominantly in the southern subtropical Indian Ocean (blue) which agrees with Miyama *et al.* (2003). A small subduction site exists also in the Arabian Sea.

e-Objectives of the study

Objectives of this study are to picture the cross equatorial cell and the subtropical cell in the Indian Ocean with the aim to understand mechanism that generate these cells and the role of those cells in the mean circulation.

2 Methodology

a-Global Model description

This study uses results from a calculation within the Ocean Circulation and Climate Modeling Project (OCCAM) as data to which the trajectory algorithm is applied. This OCCAM project is being carried out by researchers at the Southampton Oceanography Centre in collaboration with universities of East Anglia and Edinburgh.

The model is based on the ocean “primitive equations” (Bryan *et al.,* 1969), including a free surface. To obtain those equations, three important approximations are made:

1-Ocean water is incompressible

2-In the vertical momentum equation, the vertical velocity is small and the terms involving it can be neglected.

3-Small changes in density can be neglected except when they affect the horizontal pressure gradient.

The OCCAM model assumes that the velocity and the gradient of potential temperature and salinity normal to solid boundaries are zero on all solid boundaries. The model depths were calculated from the US Navy DBDB5 dataset.

The model has 36 levels in the vertical spanning 5500m, increasing from 20m at the surface to 255m at the bottom. In the horizontal, it uses 2 grids, each with 0.25 degree resolution. The first is the standard latitude-longitude grid that covers the Indian, Pacific and South Atlantic Ocean. The second is a rotated grid which covers the North Atlantic and Arctic Ocean. It is used to overcome the singularity at the North Pole due to the convergence of meridians.

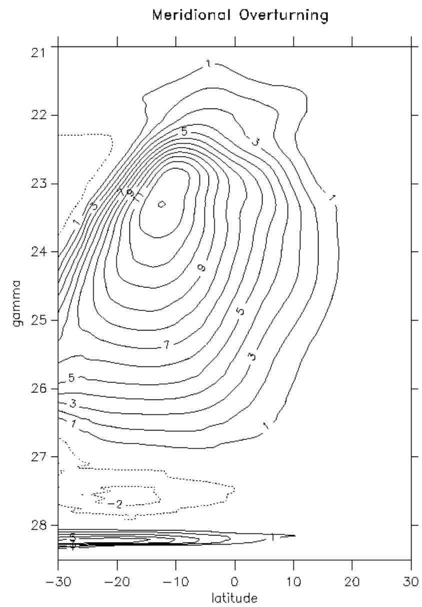


Figure7: Representation of the streamfunction (Sv) as a function of neutral density.

The monthly-averaged wind stresses from European Center for Medium-range Weather Forecast (ECMF calculated from years 1986-1988, Gibson *et al.*,1997) are applied for the first 8 years simulation. In the following three years, the six-hourly winds and wind stresses for 1993-1995 are applied.

Figure7 shows the vertical circulation in the Indian Ocean as simulated by OCCAM. It presents stream lines as a function of neutral density. These are calculated by integrating the zonally integrated transport across a given latitude between the ocean bottom and a certain depth. The figure shows a two dimensional picture of the cross equatorial cell, with northward subsurface flow and a southward surface flow. This cell is associated with downwelling south of 15S, and upwelling north of 10S extending into the northern hemisphere. Our objective is to use a particle tracking algorithm to get a three dimensional picture of this cell.

<u>b-The trajectory Algorithm</u>

The time dependent trajectory algorithm employed in this study is based on the version developed by De Vies and Döös (2001). Water masses are represented by numerous small water parcels seeded on chosen geographical sections (starting criteria). Each particle’s trajectory is then individually calculated from one grid box to the next until it reaches the ending criteria. Each of them carries an elementary transport. The algorithm analyzes data in Lagrangian terms by following individual particle trajectories. Each particle is volume conserving and follows true streamlines through the simulated velocity field. The algorithm can trace the particles forward or backward in time. Pathways are visualized as horizontal streamfunctions obtained by the vertical integration of the 3D transport field represented by the particles displacement. A streamfunction is defined only for a non divergent flux. Indeed, for a non divergent flux:

$$\frac{du}{dx} + \frac{dv}{dy} = 0$$

u and v are respectively the zonal and meridional component of the velocity.

Therefore, we can define a streamfunction ψ as:

$$u = -\frac{d\Psi}{dy} \text{ and } v = \frac{d\Psi}{dx}$$

We can write this in the vectorial form: **v = k$\times$grad(ψ).** Because the isolines of Ψ correspond to streamlines for the nondivergent velocity, the spatial distribution of **v** can be easily pictured by plotting lines of constant Ψ on a map. Thus, the streamfunction is a good parameter to understand the circulation in the ocean. The mean horizontal transport follows the streamfunction contours, with strongest flows in regions where the contours are close together. The streamfunction is plotted in units of Sverdrup (=10^6 $m^3.s^{-1}$).

In fact, the flow is never strictly nondivergent in the oceans. Consequently, we can calculate this streamfunction only in particular cases that allow to neglect the divergence. In this case we consider a vertical integral of the flow of a water mass. This vertical integral is non divergent as the flow has to be continuous. When it is not possible to work with a streamfunction, we calculate footprints. These also show the vertically integrated mass transport, but do not give any information about the direction of the transport.

We modified the algorithm to memorize the upwelling and downwelling site for each trajectory that is calculated. The difficulty was to find a good criterion to define upwelling and downwelling sites. We decided to allow for an up or downwelling when the particle crosses the bottom of the mixed layer. This criterion appeared not sufficient. Indeed, some particles pace up and down in the mixed layer many times before they make a real loop. To relieve this inconvenience, we choose to keep the up and downwelling sites separated by a maximum period of time.

We seed within the "winter" mixed layer, that is the deepest mixed layer of the year, because in that case, downwelling means that the particle "really" escapes the mixed layer and will not be captured within a seasonal cycle of the mixed layer depth. As a result, we miss all particles the up and downwell above the deepest mixed layer and also flow back above this depth.

c-Starting and ending criteria

We choose to seed particles at 17.5S in the mixed layer where southward velocity occurs, and we traced the particles backward in time until they leave the mixed layer, and then came back in the mixed layer. This ending criterion corresponds to at least one downwelling event. The choice of 17.5S was somewhat arbitrary. The only criterion was to select a section south of the upwelling sites already known from previous studies. We used seasonal mean velocity fields.

We also tried to visualize the impact of the export from other oceans in the overturning circulation. Indeed, part of the overturning may be associated with inflow from the Pacific Ocean or Southern Ocean, and outflow into the Southern Ocean, with subduction sites outside the Indian Ocean.

To this end, we seeded along the southern boundary and traced particles backward in time until they reached the eastern boundary (inflow from the Pacific Ocean) or came back to the southern boundary (inflow from the Southern

Ocean). This calculation gives us the upwelling and downwelling sites due to those imports and exports.

3 Results and discussion

The pathways of the Cross Equatorial Cell in the Indian Ocean have been explored by using a tracking particles algorithm.

Backtracking the particles from 17.5S in the mixed layer until they leave the mixed layer, and then come back to the mixed layer, gives the upwelling and downwelling sites involved in the cross equatorial cell. According to our results, around 5Sv participate in the Cross Equatorial Cell.

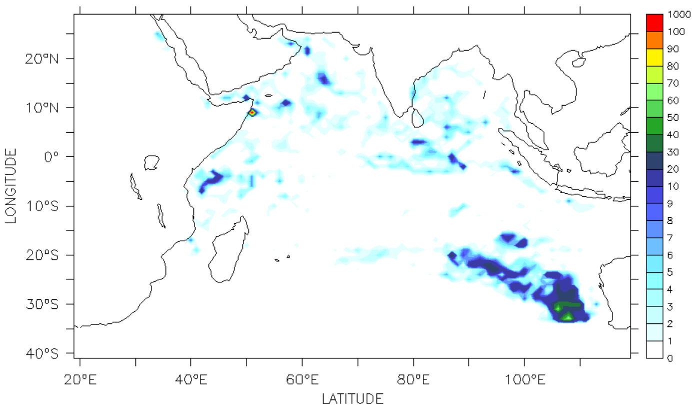


Figure8a: Downwelling sites of the CEC given by the particle tracking algorithm (in Sv)

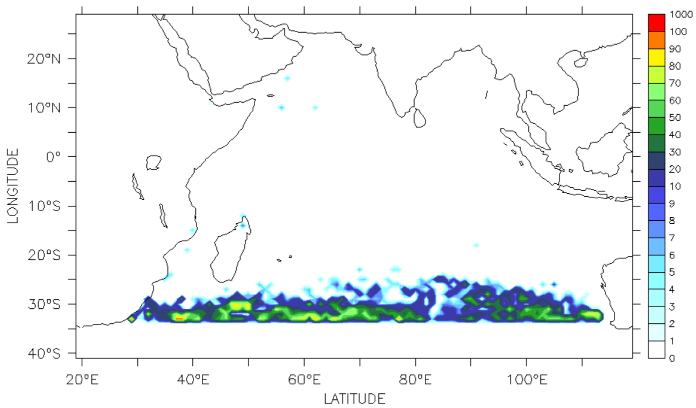


Figure8b: Downwelling sites involved in the inflows and outflows given by the particle tracking algorithm (in Sv)

These regions agree with the Schott *et al.* (2004) (cf. figure4). The downwelling region west of the Australia coast is smaller than in the figure4. The downwelling picture from the export study (figure8b) can explain this divergence. It seems that an important part of the downwelling observed by Schott *et al*. flows into the Southern Ocean without any participation to the overturning circulation. According to our study, this region is split in two parts by a separatrix at about 20S bending between 90E and 120E to 35S. North of this separatrix, subducted water flows northward in the Indian Ocean South of it, subducted water flows southward into the Southern Ocean.

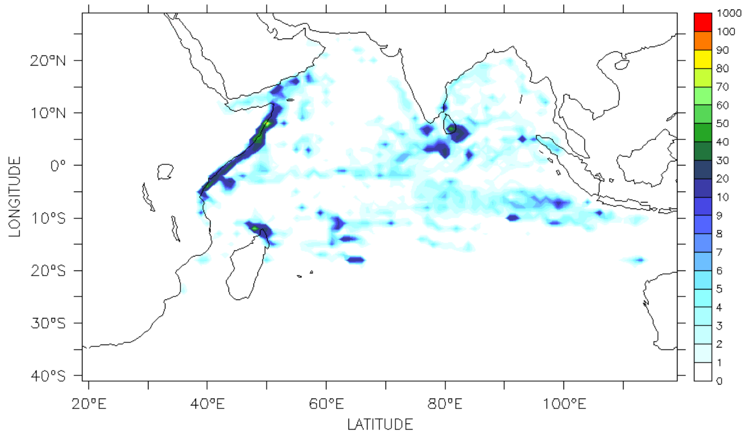


Figure9: Upwelling sites given by the particle tracking algorithm (in Sv).

Upwelling sites are situated in the Northern hemisphere, along Somalia, Oman, and the Indian coast. There is no upwelling site near Sumatra or in the open ocean as in Schott *et al*. (2004). These latter region may not be involved in the cross equatorial cell. They may participate in a Northern subtropical cell, much weaker than the cross equatorial cell. The upwelling occurs presumably in summer, when the southwesterly winds along Somalia coast induce an Ekman transport toward the open ocean.

To visualize pathways between upwelling and downwelling sites, we used footprints.

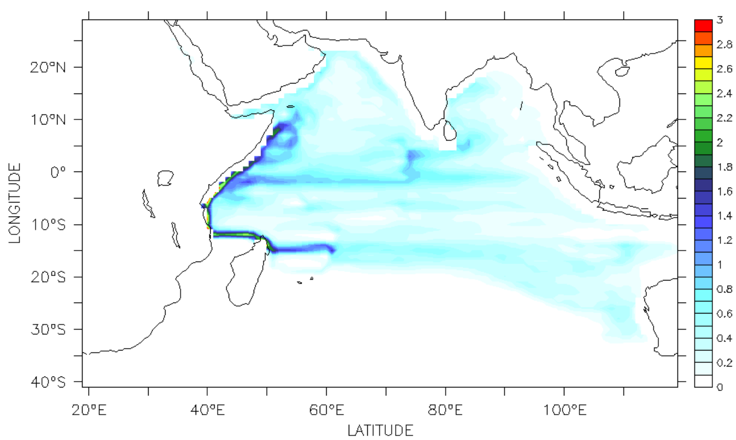


Figure10: Footprints of the lower branch of the overturning cell (in Sv)

Figure.10 shows the lower branch of the cross equatorial cell. It links the downwelling sites to the upwelling sites. We know that the most important downwelling region is west of the Australian coast. Notice that the separatrix is visible is this picture. The subsurface flow goes north-westward along this separatrix. It joints a strong current near Madagascar which flows north of this Island to join the Somalia Current. Along the African coast, part of the flow upwells. The other part is carried by the equatorial undercurrent to the Indian coast where it upwells.

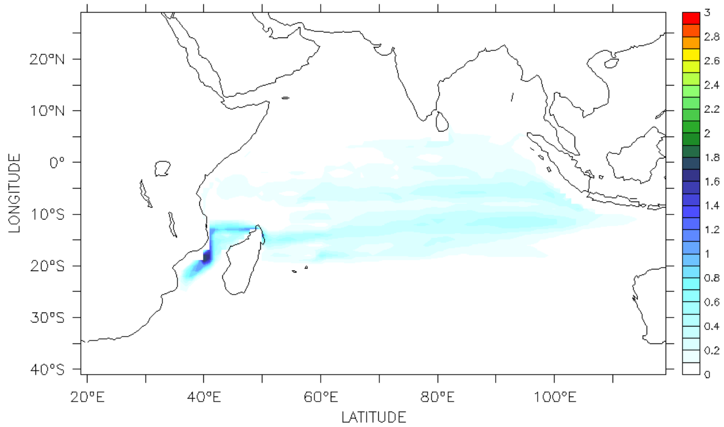


Figure11: Footprints of the upper branch of the overturning cell (in Sv)

The upper branch shows the pathways of the flow after the upwelling event. The water flows westward until it reaches the Agulhas current and flows into the Southern Ocean.

Because of our choice for the mixed layer depth, we miss a part of the cell that has its upper and lower branch above the subduction depth.

When we calculate the streamfunction from inflow to outflow in the Indian Ocean, we find that most of the Cross Equatorial Cell is provided by the Indonesian Throughflow.

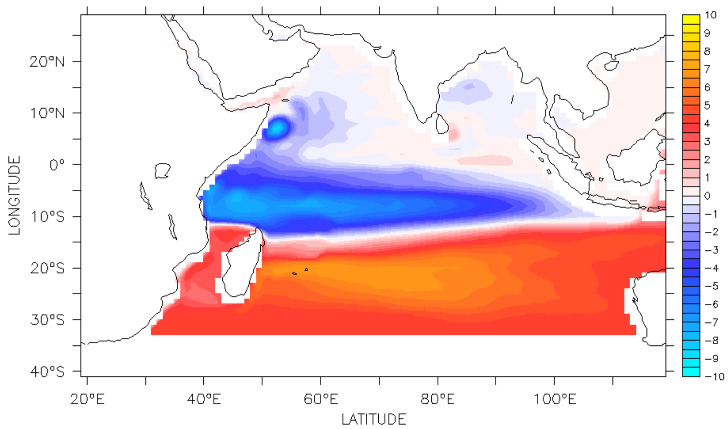


Figure12: Streamfunction that depict water mass transport from the Pacific Ocean to the Southern Ocean via the Indian Ocean. (in Sv)

Figure12 shows the streamfunction inflow from the Pacific (most of this inflow corresponds to the Indonesian Throughflow) to outflow to the Southern Ocean. As said before, the transport follows the isolines of the streamfunction with the strongest flow near the region where isolines are close together. As a convention, when the streamfunction is smaller than zero, the flow goes clockwise. In this picture, we can see an obvious clockwise circulation in the northern half of the basin provided by the Indonesian throughflow.

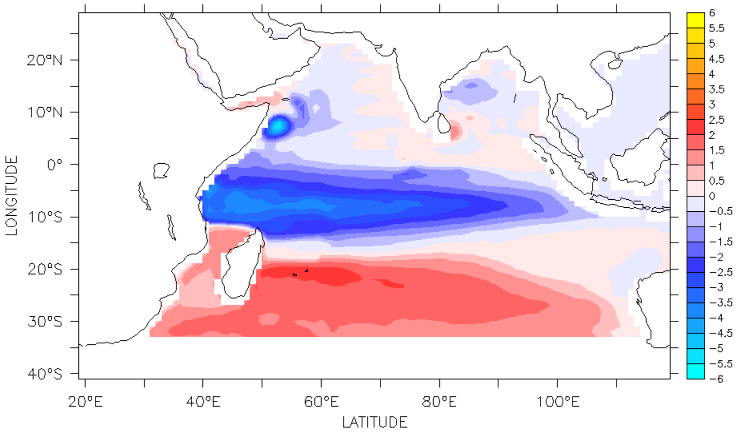


Figure13: Streamfunction that depict water mass transport from the Southern Ocean to the Southern Ocean via the Indian Ocean (in Sv)

Figure13 is the same picture from inflow from the Southern Ocean to outflow into the Southern Ocean. Water from the Southern Ocean involved in the Cross Equatorial cell follows the same clockwise circulation as the water that enters the Indian Ocean by the Indonesian Throughflow.

Conclusion

We have investigated the overturning circulation in the Indian Ocean using a Lagrangian tracking algorithm on seasonal-mean velocity fields from the OCCAM model. Our results are in agreement with previous study and support the theory of a cross equatorial cell. They give us a better idea of the pathways of the wind driven overturning circulation in the Indian Ocean. This circulation is mainly provided by the Indonesian throughflow. The water subducts near the Australian coast, then it flows into the southern Indian Ocean and goes northward in the Somalia current. It goes eastward into the equatorial undercurrent until the Indian coast. During this trip, part of the flow upwells along the African coast because of the summer monsoon; the other part upwells close to India. This water goes southwestward near the surface. It flows into the Southern Ocean by the Agulhas current.

To improve our understanding, we should reapply this methodology with a different choice of seduction depth. For instance, we can choose the annual mean mixed layer depth.

We also can repeat the protocol described here with perpetual seasonal means. The cell that we observed here may change with the season; the results we obtain are the net effect of a whole seasonal cycle which may obscure the temporal changes in the flow due to the seasons.

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