I. Introduction

As the boundary condition on sea surface temperature (SST) alone is sufficient to compute the state of the atmosphere, it is tempting to ignore the ocean below the air-sea interface. Indeed the daily observations of SST allow routine weather forecasts on weekly time scales. For predictions on time scales longer than the horizon of synoptic weather systems, the ocean comes into play for its capacity to store and transport heat. With the mass of the ocean exceeding that of the atmosphere by a factor of 250, and the specific heat capacity of water exceeding that of air by a factor of 4, the thermal inertia of the ocean is 3 orders of magnitude larger than that of the atmosphere. Subject to a heat loss of 10 Wm\(^{-2}\), (an uncertainty on the low side in air-sea heat fluxes) the atmosphere cools by 1\(^\circ\) in about two weeks while it will take the ocean 50 years to cool by the same amount. Storage means delay: on seasonal time scales, the ocean stores heat in a mixed layer of order 100 m which is responsible for the roughly two months delay between the summer (winter) solstices and the SST maximum (minimum) observed at mid-latitudes. On longer time scales, the heat transport by the ocean comes into play. The solar differential heating creates temperature gradients, hence density and pressure gradients that drive the ocean currents. A second cause of oceanic motions is simply the mechanical friction exerted by the wind at the sea surface. Since both forcings are surface forcings, it is expected that the currents will be surface intensified. Indeed they are with the vertical depth scale controlled in part by the depth reached by convective instabilities during the winter season. Typical large scale velocities are of the order of a few cm s\(^{-1}\) at the surface becoming 0.1 cm s\(^{-1}\) in the abyss. Although smaller than winds by nearly three orders of magnitude, these velocities are sufficient for oceanic meridional heat transport to compete with atmospheric heat transport on account of the huge afore mentionned oceanic thermal inertia.

2. Dynamics

Because of the symmetry of the forcings (wind stress and insolation), the ocean circulation would be expected to preserve those symmetries but this is not the case. Neither zonal nor meridional symmetries are currently observed, in distinctive contrast with atmospheric circulations patterns. Besides inertia another obvious difference between the ocean and atmosphere is the fact that the oceans are enclosed (or semi enclosed) in bowl type basins and Stommel (1948) was able to show that the western intensification of ocean currents (Gulf Stream, Kuro-Shio, Brazil current, Somali current, East Australian current) resulted from the spherical shape of the rotating earth and the presence of meridional boundaries. This was shown in a homogeneous ocean subject to wind driving alone. In Stommel's solution, the interior is geostrophic with turbulent friction relegated in a western boundary layer. When a container rotates very fast, the Taylor-Proudman-Poincaré’s theorem says that fluid velocities must be independent of depth measured along the axis of rotation and consequently free flows are possible in the zonal direction only. Zonal flows are indeed truly dominant in the atmosphere for precisely this reason. A similar "atmospheric" situation exists in the ocean at southern latitudes where the fluid can move around the Antarctic
continent unimpeded by meridional coast lines, but this is an exception (albeit an important one). In subtropical basins, the ocean interior forced by a negative wind stress curl, responds in moving equatorward according to the Sverdrup relation (Sverdrup, 1947). At any given latitude however, a return poleward flow must exist to close the mass balance. This return flow cannot be geostrophic and new frictional effects appear in a boundary layer. If such frictional effects are to oppose the wind effects which induce a tendency for clockwise circulation, the response must also be clockwise and the boundary layer must be on the western side of an oceanic basin. Of course molecular viscosity (10^6 m^2 s^-1) is unable utterly to produce the 50 km widths of the observed western boundary currents. This is where the oceanic mesoscale eddies come in …

The first direct, ship observations of the ocean beginning in the early 20th century were vertical profiles of temperature and salinity (hence density) carried out every few hundred nautical miles. If the basin scale was resolved by such early observations, it missed completely the eddies with diameters of 50 to 200 km. The latter were discovered in the 70’s when the first autonomous oceanic measurements were made with current meter moorings and freely drifting floats. Although smaller by a factor of 10 than the synoptic perturbations of the atmosphere, the oceanic eddies have identical dynamical origins, their energy coming from the release of large scale potential energy through baroclinic (and to a lesser extent barotropic) instability processes. They scale as the internal Rossby radius of deformation NH/f and their small size compared to atmospheric equivalents bears witness to the stronger stratification of the atmosphere (N_{atmosphere} \sim 10^4 \text{ rad s}^{-1}; N_{ocean} \sim 10^2 \text{ rad s}^{-1}). Such eddies stir the fluid transporting heat, momentum and vorticity to smaller scales where molecular dissipation sets in (typically the cm scale in the ocean). From floats tracing the motions of fluid parcels, the eddy diffusivity can be measured and can reach up to 10^8 m^2 s^-1 values in energetic western boundary current regions. With a mixing length of the order of the eddy scale (50 km), such diffusivities require velocities of order 10/50 x 10^3 or 20 cm s^{-1}, in the range of observed values. In such regions, the eddy velocities are as large as the permanent mean boundary current velocities. The boundary layer width (diffusivity/\beta)^{1/3} required by this vorticity dissipation is then (10^2/2 \times 10^{-11})^{1/3} = 80 \text{ km} in agreement with observations (\beta = 2\Omega \cos \text{ (latitude)/ Earth radius}).

A further landmark in ocean circulation theory was marked when Anderson and Gill (1975) studied the spin up of a wind driven ocean with a given stratification. The previous interior depth independent circulation is adjusted rapidly in a week or so by barotropic Rossby waves crossing the basin from east to west. Slower interior Rossby modes are then observed to progressively destroy deep circulation until, as time goes to \infty, the circulation is entirely confined to the surface. This singularity, the so-called Gill's catastrophe, is a direct consequence of the linearized, non diffusive buoyancy (or temperature) equation \nabla \cdot \mathbf{v} = 0, implying \mathbf{v} = 0. With geostrophy, the vorticity balance reduces to \beta \mathbf{v} = f \mathbf{w}, so that if \mathbf{w} = 0 then \mathbf{v} = 0 and continuity gives finally \mathbf{u} = 0 ! By relaxing the hypotheses leading to this non physical behavior, namely non linear buoyancy advection by the large scale flow and eddy mixing, two theories by Rhines – Young (1982) and Luyten-Pedlosky-Stommel (1983) invented two very different ways to prevent Gill's catastrophe. The first theory relies on vertical eddy transport of momentum while the second process is a ventilation from the "sides" when deep isopycnals intersect the sea surface at subpolar latitudes.
While such conceptual theories exist for the wind driven circulation, nothing of the sort exist for the thermohaline circulation (THC), that is, the circulation forced by heat and fresh water (evaporation – precipitation) fluxes. In reality the two are intrinsically coupled by nonlinear terms, but if remains useful nevertheless to concentrate on the sole THC. The difficulty when dealing with the THC is that an eventual theory must take into account the calculation of both the velocities and the stratification (= the temperature, salinity and density distributions) and this has not been done so far. There is no lack of incentive though, given the role that the THC is thought to play in climate. Given warming (cooling) at low (high) latitudes in a basin limited by meridional walls, poleward (equatorward) flow is expected at the surface (at depth) .i.e. in the sense of a direct thermally driven circulation. It is the combination of a poleward warm branch and an equatorward cold branch that produces the overall meridional oceanic heat transport. In the Atlantic for instance a volume transport of 20 Sv (1 Sv = 10^9 m³ s⁻¹) associated with a 10° C (2°C) temperature in the upper (lower) layer generates a heat transport \(\rho C_p \Delta T \sim 6 \times 10^{14}\) W.

Although symmetry about the equator of such patterns of heat transport is expected, such symmetry is again broken ! **In the Atlantic, the heat transport is directed everywhere towards the North pole.** (More symmetric but weaker patterns are found, however, in the Pacific). What happens near the North pole ? Of course the warm upper branch must be given flow continuity with the cold lower branch. In fact due to extreme heat losses in the Norwegian Sea, the fluid overturns in a very high Rayleigh number convection. It cannot do so on the large scale because of the rotation constraint of the Taylor-Proudman-Poincaré theorem and bottom reaching plumes are observed on "small scales" (100 m or so). Note that In the atmosphere, the most extensive convective pattern (the ITCZ) is found in the vicinity of the equator where the horizontal Coriolis force vanishes. North Atlantic deep water is produced, one of the most important water mass in the ocean. It tends to flow equatorwards again on the western side of the ocean and invades the world ocean. The upward moving branch of the THC is more elusive and is one of the major difficulty in physical oceanography today. Since the ocean is nearly everywhere stably stratified, (except in the deep water formation region just mentioned) any upward motion requires (turbulent) mixing across isopycnals. Such mixing in a stably stratified fluid increase the potential energy and therefore a mechanical energy source is necessary. Diverse energy sources include external factors such as winds and tides (Munk, 1966 and Munk, Wunsch, 1998) but it is difficult to rule out the possibilities of the THC to generate its own turbulent mixing with cascades from the eddy scales, through internal gravity waves to dissipative scales. Whatever the energy sources, the magnitude of the mixing required by deepwater formation is fairly well determined. Suppose that the 20 Sv of water formed at the pole rise uniformly across the thermocline between 30° S and 30° N, roughly half the surface area of the ocean \(\sim 1.5 \times 10^{14}\). This gives a velocity of 1.2 \(10^3\) m s⁻¹ or 3.6 m yr⁻¹ (by contrast the Ekman convergences in the subtropics are an order of magnitude larger). Near the base of the thermocline, the temperature (or salinity) vertical profiles decay with a vertical scale of about 700 m and vertical advective-diffusive balance then leads to a mixing coefficient K of order \(10^{-4}\) m² s⁻¹ (1.2 \(10^{-2}\) x 700 m). Although this large scale estimate is difficult to doubt, direct measurements of the mixing have systematically shown lower values \(\sim 10^{-5}\) m² s⁻¹). Recently however, much higher values have been found above rough topography and in regions of overflows from one oceanic basin to another (deeper) one. The slowly emerging picture is then that of a very inhomogeneous distribution of the values of K. By contrast Stommel-Arons (1961) "solved" for the deep branch of the THC by assuming a uniform upwelling
at intermediate depths and a homogeneous deep ocean. With such a forcing, the fluid moves polewards towards the deep water source at the poles! To balance mass, deep western boundary currents are invoked whose transports can then exceed the volume rate of the sources. Although a remarkable step ahead, this abyssal theory does not relate the THC to the surface forcing and eludes the central point of heat (or fresh water) transport. It is essentially a circulation based on vorticity concepts and not on thermodynamic ones.

3. THC and climate

In the 80's computer power increased significantly to allow numerical spin up of the THC. Thermodynamic equilibrium is achieved over a diffusive time scale (ocean depth = 4000 m)²/(K_V = 10⁻⁴) in the range of several thousand years, very much larger than the Rossby wave basin crossing time which governs the wind driven response. Over such time scales, several authors simplified considerably the primitive equation models by neglecting inertia i.e. reducing the circulation to a Stokes flow but with rotation added. These so-called planetary geostrophic models allowed very fast calculations of the THC, still sufficiently accurately when the grid scale did not resolve the mesoscale eddies (i.e. typically several degrees of latitude). In these calculations it was found that the value of the K_V determined entirely the strength of the response and the depth of the thermocline. When the two components T and S were considered with their respective surface fluxes, new phenomena occurred. THC calculations are run now with surface values of T and S restored to climatological T,S values because the errors in surface fluxes are too important. Bryan (1986) was the first to change the boundary condition on S. After a first spin up with restoring boundary conditions, he diagnosed the surface salt flux that was kept constant thereafter. A remarkable flow change occurred under such mixed boundary conditions. The salinity of the Northern latitudes decreased, a halocline formed with deep convection unable to reach the deep ocean. Because the meridional density gradient then decrease, the overturning mass transport decreases as well. Under such a weaker circulation, anomalous salinity distributions are built up that oppose the THC with evaporation (precipitation) creating positive (negative) salinity anomalies. In the meantime the slowing of the western boundary current decreases the surface heat losses that also diminish northern densities. The net result is that very rapidly (in a few decades), the THC collapses entirely in a so called polar halocline catastrophe (PHC). This absence of circulation (and density gradients) persists until the deep waters warm sufficiently by diffusion. At some point the stratification is statically unstable, polar convection is triggered and the positive advective feedbacks used for destruction of the THC will be just as active for its construction. Poleward advection of strong S anomalies of low latitude origin will increase the density of polar waters leading to stronger convection etc… Under certain conditions Winton and Sarachik, 1993 showed that the THC had the potential to oscillate between these two advective and diffusive states over the diffusive time scale (thousand years). At about that time an overlooked 1961 paper by Stommel came back to the surface… In a simple two box model that mimicks "mixed boundary conditions", Stommel showed that the THC could exhibit multiple states under identical freshwater fluxes. When the freshwater fluxes in the 2 boxes are below a critical value, three steady states are possible, a strong (stable) thermal state, a weak (unstable) thermal state and a salinity (stable) state. When the fluxes are above critical only the latter persists. In the language of dynamical system theory a saddle node bifurcation occurs and a catastrophe sets in at the critical central point; a catastrophe in the sense that small perturbations
in the state of the system or in the freshwater) forcing can lead to very different flow fields.

Very early on this sensitivity of the THC to freshwater fluxes has been assigned an important role in paleoclimates variability by Broecker (see Broecker and Denton, 1989).

Paleo-archives in sediment and ice cores reveal a startling record of the last 100,000 y. Evidence of North Atlantic cooling associated with Increased Rafting Detritus (IRD) North of 40° N appears several times in the record. In these so called Heinrich events, one speculates that massive iceberg releases in the North Atlantic create sufficient freshwater perturbations for the THC to collapse. With the vanishing of oceanic heat transport to high latitudes, a massive reorganisation of the Ocean-Atmosphere system occurs. Other important features are the so called Dansgaard-Oeschger events whose signatures are abrupt warmings followed by slower coolings. Because the duration of the events and their irregular spacing is in the millenial time scale, it is widely believed that oscillations of the THC are involved (see Broecker-Denton, 1989). Any proper explanation of the phenomena will have to come to grips with the contrast between the "millenial climate turbulence" of the last glacial cycle and the present very stable "Holocene" period ... stable, yes, yes ... but what about our anthropogenic increase of Greenhouse gases ? Under a warmer climate, a larger moisture transport to high latitudes in the atmosphere is expected and indeed many coupled ocean-atmosphere GCMs favor a decline of the THC in doubled CO₂ experiments reminding of the oceanic story under mixed boundary conditions (Manabe, Stouffer 1993).

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