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THE THERMOHALINE CIRCULATION AND THE CONTROL OF ICE AGES

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ABSTRACT

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It is found that the global salinity variations associated with the thermohaline circulation may have a tendency to make the circulation increasingly asymmetric with respect to the equator. As a consequence the salinity difference between the Pacific and the Atlantic Ocean may be slowly increasing. Such a process could have a time scale long enough to be comparable with the time span between major glaciations. A speculative glaciation cycle is proposed which involves the above mentioned property of the thermohaline circulation. In this cycle the role of a Northern Hemisphere glaciation is to bring excess freshwater from the Pacific to the Atlantic.

INTRODUCTION

The oceans are evidently important for the thermal conditions on the earth. The heat capacity of the oceans tends to even out seasonal variations, and the thermal circulation similarly decreases temperature differences between polar and equatorial regions. We are thus led to the general first conclusion that the overall effect of the ocean is to stabilize climatic fluctuations rather than the opposite.

However, the ocean because of its size may provide long time lags which under suitable circumstances could be crucial for feedback mechanisms involved in the amplification of climatic changes. In this paper we look for such possibilities. In particular we ask whether the general thermohaline circulation of the world ocean could be actively involved in the climatic fluctuations associated with the major glaciations.

The time scale for the glaciation cycle is of the order of 50,000 yrs., which can hardly be matched by the turnover time of the ocean being of the order of 1000 yrs. However, the thermohaline circulation primarily controls the exchange of properties between surface and deep layers. The exchange between different oceans e.g. the Pacific and the Atlantic is a secondary effect of the general circulation. It is thus possible that the exchange processes

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between the northern parts of the Pacific and the Atlantic could be characterized by a timescale much larger than 1000 yrs. and perhaps large enough to be comparable with the time span between glaciations.

If we look at the hydrographic state of the World Ocean we find a conspicuous feature namely a pronounced skewness in the distribution of salinity. We thus find (see Fig.1a, b):

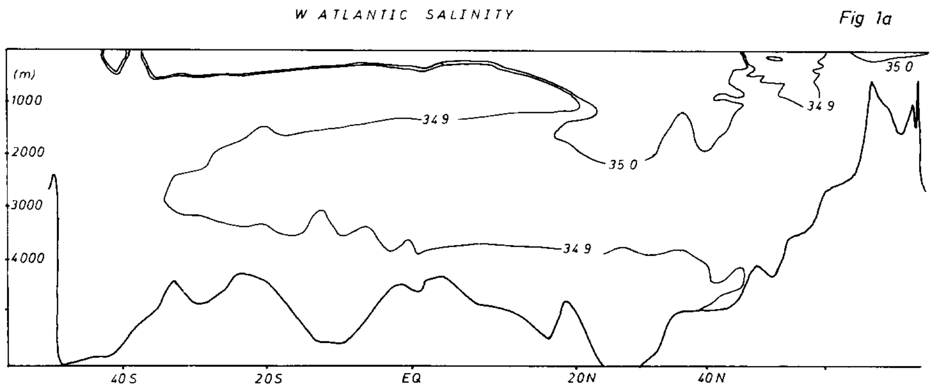
(1) The salinity distribution is markedly asymmetric with respect to the equator in the Pacific as well as in the Atlantic Ocean.

(2) The salinity is generally lower in the Pacific than in the Atlantic Ocean. Apparently this skewness has been persisting for as long as man has been observing the salinity distribution. It is however perfectly possible that this very large scale feature is slowly changing on very large time scales.

In this paper we discuss the following possibilities:

(1) The observed skewness of the salinity field is the result of (or at least amplified by) a spontaneous instability of the thermohaline circulation.

(2) The overall salinity discrepancy between the Pacific and the Atlantic Ocean is still in a state of slow increase.



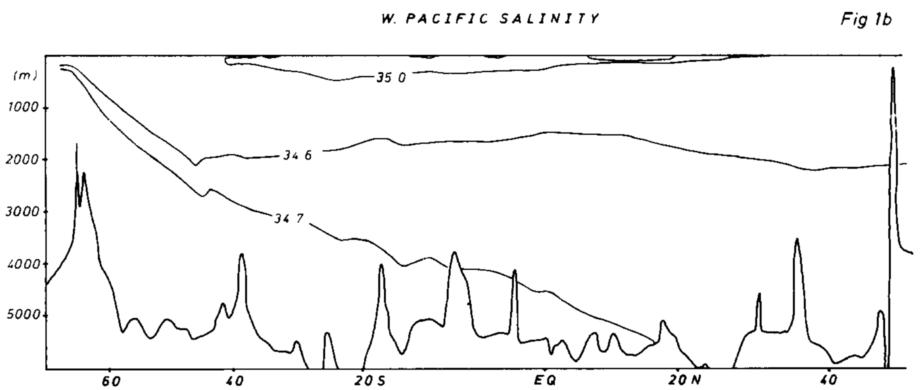


Fig.1. a. Salinity on section through Western Atlantic. b. Salinity on section through Western Pacific (from Geosecs Atlas, 1981).

(3) The slow build up of this salinity gradient may provide the watch controlling the time between major glaciations.

Against these hypotheses speaks an analysis presented by Stigebrandt (1984), in which the presently observed salinity distribution is shown to be compatible with a time-independent description. In Stigebrandts model the low salinity in the Pacific is caused by excess rainfall derived from overflow of water vapor from the Atlantic. A balance is achieved by the leakage of low saline water through The Bering Strait.

Of course the two processes recognized by Stigebrandt should be taken into account in any quantitative description of the system. In particular the "buffering" mechanism of the Bering Strait is vital since it provides a process capable of limiting the spontaneous build up of the salinity contrast between the two oceans proposed in this note. In what follows we will however disregard external forcing of the type discussed by Stigebrandt and focus on the inherent properties of the circulation system.

THE THERMOHALINE CIRCULATION

In this section we will discuss some properties of the thermohaline circulation in a very elementary and simplified manner.

A purely thermal circulation

To begin with let us consider the basic properties of a purely thermal circulation, i.e. assuming that the density variations caused by salinity may be ignored. This is intended as a background. As we will find salinity variations are crucial for the qualitative behaviour of the system.

An extremely simplified model is illustrated in Fig.2. The system works in the following way:

(1) The local heat balance essentially controls the temperature at the ocean surface, i.e. the two temperatures T_0 and T in the model.

(2) The mixing conditions (essentially wind mixing) determines the depth of the warm layer, while the rest of the system is filled up by the cold water with temperature T.

(3) The warm top layer is less dense than the cold water and is spread by gravitation. The associated volume flow m is determined basically by the thickness H and the density difference between the warm and the cold water (assumed proportional to $T_0 - T$).

(4) The warm water is gradually cooled down to the temperature T while moving along the surface of the system towards the polar regions.

The simplest possible relation between the quantities m, H, T_0 and T is:

$$m = kH^2(T_0 - T) \tag{1}$$

where k is a proportionality factor. Equation 1 may be motivated by assuming that the warm outflow is in geostrophic balance. However, essentially,

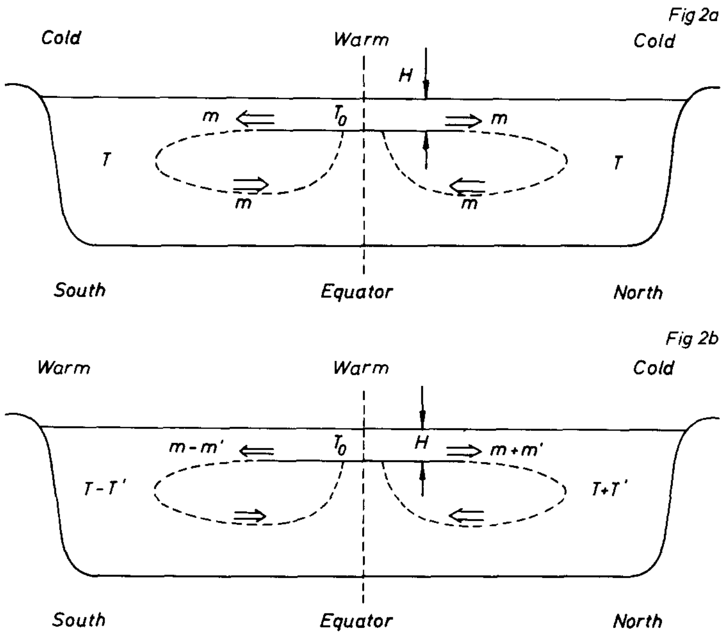


Fig.2. a. Model of purely thermal circulation. The warm layer is spread by gravitation in a symmetric manner. b. Thermal circulation with asymmetric perturbation. Perturbation generally decays.

Eq. 1 simply reflects our belief that the warm water spreads because of its lower density.

The heat balance for e.g. the right-hand half of the cold water mass may be written:

$$m(T_0 - T) - Q(T) = 0 \qquad (2)$$

where $Q(T)$ is the net heat loss resulting from radiation, sensible and latent heat flow through the sea surface. It should be recognized that the term $m(T_0 - T)$ is small compared with the different contributions to $Q(T)$.

Let us now consider the stability of an anti-symmetric perturbation on T and m as illustrated in Fig.2b. The counterparts of Eqs. 1 and 2 become:

$$m + m' = kH^2(T_0 - T - T')$$

$$\frac{\partial}{\partial t} T' \sim (m + m')(T_0 - T - T') - Q(T + T')$$

or after making use of Eqs. 1 and 2:

$$\frac{\partial}{\partial t} T' \sim -mT' + m'(T_0 - T) - Q' \qquad (3)$$

where:

$m' = -kH^2T'$; $Q' = Q(T + T') - Q(T)$

It is easily seen that all three terms on the right-hand side of Eq. 3 contribute to the stability of the system (i.e. to the elimination of the perturbation T'). The main effect will most likely come from the term Q' corresponding to increased local heat loss through the ocean surface (if $T' > 0$). It should be noted however that the advective terms also contribute to the stability of the system.

The above given discussion of the purely thermal circulation reveals nothing unexpected. The reason for the presentation is that we need it as a background for the analysis of how the system is modified by the presence of salinity variations. (In the above given analysis — and in what follows — we have treated H as a constant. This is not generally permissible since H depends on the flow of water from the warm layer and also on the density contrast in the region where the warm water is mixed up into the surface layer. The justification depends on our choice of an anti-symmetric disturbance in which case H is unaffected to a first approximation.)

The influence of salinity on the circulation

Let us now consider in the simplest possible way what effect salinity variations might have on the circulation. The model is illustrated in Fig.3a. The model conforms as far as possible with our model for the purely thermal case. In this case we have assumed that the top layer is not only warmer but also more saline than the deep water. The salinity contrast is forced by a net evaporation $2R$ from the warm layer which returns as net rainfall to the cold parts of the system. This implies that the deep return flow $2(m + R)$ is somewhat stronger than the outflow from the top layer.

In a steady state the system works essentially in the same way as the purely thermal case. The difference shows up in the stability analysis which depends on two circumstances:

(1) The increased salinity of the top layer reduces the overall density contrast and thereby counteracts the circulation. We may think of the haline circulation as being parasitic on the thermal circulation.

(2) The local fluxes (evaporation and rainfall) which controls the salinity field are almost completely independent of the salinity. This means that the most important feed back mechanism which keeps the temperature field under control (i.e. $Q(T)$) has no counterpart for the salinity field.

In analogy with Eqs. 1 and 2 describing the purely thermal case we have:

$$m = kH^2[\alpha(T_0 - T) - \beta(S_0 - S)] \tag{4}$$

$$mS_0 - (m + R)S = 0 \tag{5}$$

where α and β are such that $\alpha(T_0 - T) - \beta(S_0 - S)$ is the density difference between the top layer and the deep water. We note that the volume flow m is reduced by the influence of the salinity contrast $(S_0 - S)$. Equation 5 represents the salt balance for one half of the deep water.

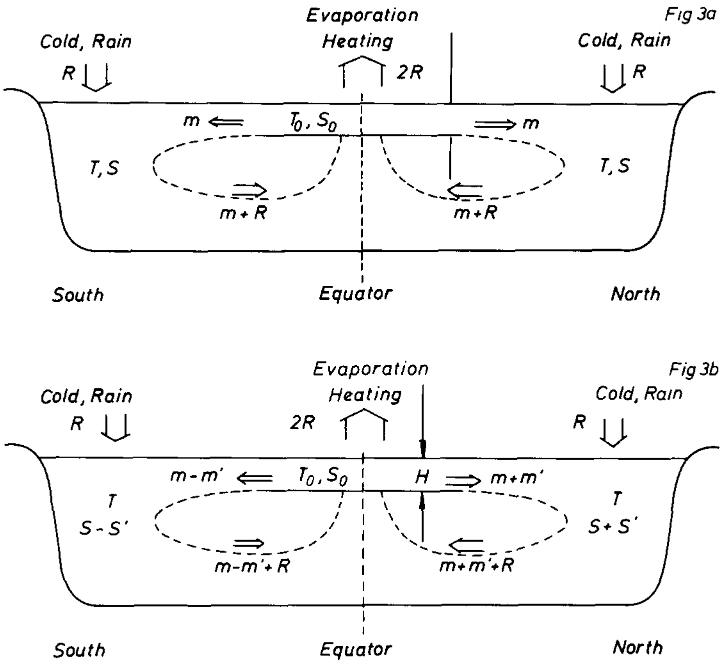


Fig.3. a. Model of thermohaline circulation. The top layer is warmer and more saline than the deep water. The salinity contrast is counteracting the thermally forced circulation. b. Thermohaline circulation with asymmetric perturbation on salinity and rate of circulation. The perturbation will grow if the salinity has sufficient influence on density between the top layer and deep water.

Let us now consider the stability of an anti-symmetric perturbation on the system (see Fig.3b). This time we will assume the temperature to remain unchanged which is justified by the strong local control of the temperature field as discussed in the previous section. The counterparts of Eqs. 4 and 5 become:

$$m + m' = kH^2[\alpha(T_0 - T) - \beta(S_0 - S - S')]$$

$$\frac{\partial}{\partial t} S' \sim (m + m')S_0 - (m + m' + R)(S + S')$$

After some manipulation and making use of Eqs. 4 and 5 we find:

$$\frac{\partial}{\partial t} S' \sim \{kH^2[-\alpha(T_0 - T) + 2\beta(S_0 - S)] - R\}S' \tag{6}$$

This result is essentially different from the corresponding result for the purely thermal case as described by Eq. 3. The important difference is of course that one of the terms on the right-hand side of Eq. 6 represents a positive feedback which can give rise to growing perturbations. In fact Eq. 6

shows that the anti-symmetric perturbation under consideration will grow spontaneously if:

$$kH^2 2\beta(S_0 - S) > kH^2\alpha(T_0 - T) - R$$

Since the net rain fall R is generally much smaller than $kH^2\alpha(T_0 - T)$ (which is always larger than m) we find with good approximation that the symmetric state is unstable if:

$$\gamma > \gamma_c \qquad \text{where:} \qquad \gamma = \frac{\beta(S_0 - S)}{\alpha(T_0 - T)}\,; \qquad \gamma_c \approx \frac{1}{2} \tag{7}$$

We note that the stability parameter γ is defined as the ratio between the density difference caused by salinity and the density difference caused by temperature.

Physically we can understand the instability phenomena in the following way. Increasing the salinity S by the amount S' decreases the breaking action of the imposed salinity contrast $S_0 - S$. The flow into the volume with raised salinity thus increases. This increased flow will bring more salt to the volume and thus increases the perturbation S'.

Implications for the real ocean

The above given stability analysis is extremely simplified. Nevertheless it is believed that the parameter γ does control the stability of the system and that if γ is larger than some value γ_c a symmetric state is not possible, even though forcing and topography would suggest such a state. The precise value of γ_c and the definition of γ for a more realistic system will of course be a matter of further analysis.

What then happens if $\gamma > \gamma_c$? A most reasonable conclusion from our analysis is that the system "falls" into an asymmetric state characterized by:

(1) The circulation is stronger in one direction than in the other.

(2) The deep water salinity is larger (smaller) and possibly increasing (decreasing) on the side with the stronger (weaker) circulation.

Since the system may "fall" into either direction a topographically symmetric system may have two stable asymmetric configurations. However, it should be recognized that we do not know if the system has reached such a state when we observe it (or if a truly stable state exists at all). Let us now see what we find in the real ocean.

It has already been pointed out that the salinity fields of the Pacific as well as the Atlantic Ocean are markedly asymmetric (Fig.1). Studying the distributions further we find that observed values of γ typically are of order 0.5. These two observations strongly indicate that the ocean is in an overcritical state. (Note that we do not expect observed values of γ to exceed γ_c, just as we never observe an unstable density stratification.) Thus we have good reason to believe that the observed skewness of the salinity field is the result of a spontaneous process rather than a result of asymmetric forcing or topography.

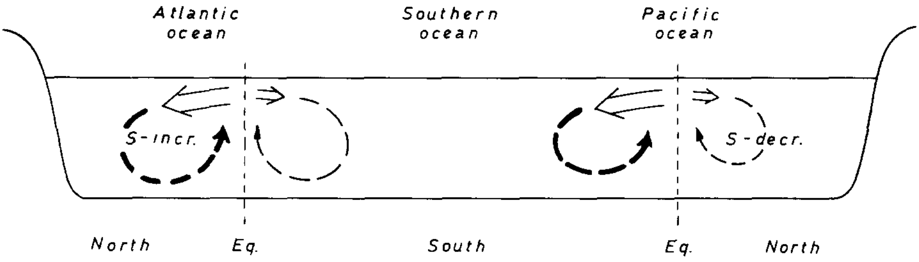


Fig.4. Thermohaline circulation in a coupled model of the Atlantic and the Pacific Ocean. The asymmetry resulting from spontaneous instability is fully developed. A finite salinity difference between the North Pacific and the North Atlantic is sustained and possibly further amplified.

We also see that the Atlantic and the Pacific Ocean are asymmetric with respect to the equator in opposite ways. In the Atlantic the preferred direction of the circulation is northwards while the Pacific behaves in an opposite way. Within the framework of our simplified model we thus arrive at a coupled Atlantic and Pacific model as illustrated in Fig.4. The thermohaline circulation systems in the Atlantic and the Pacific thus cooperate to build up a salinity contrast between their northern ends.

A GLACIATION CYCLE

In the previous section we found that the present asymmetric state of the ocean may be the result of a spontaneous process. This conclusion (if correct) opens two interesting possibilities:

(1) At other times the state of the oceans may have been entirely different, e.g. with a salinity maximum rather than a minimum in the North Pacific.

(2) The present state may still be transient and characterized by a salt flux towards the North Atlantic (i.e. a freshwater flux in the opposite direction).

Let us now briefly outline a hypothetical glaciation cycle based on the assumption that the second statement is true. A key process is provided by the flow of relatively low saline water through the Bering Strait from the Pacific into the Arctic Ocean. In the previously cited paper by Stigebrandt (1984) it is convincingly proposed that the flow through the Bering Strait is actually forced by the salinity difference between the North Atlantic and the North Pacific. (The lower salinity in the Pacific implies a higher water level which causes a spill-over through the shallow Bering Strait.) This means that an increasing salinity contrast will be followed by increased flow through the Bering Strait.

The extent of the ice cover in the Arctic basin depends on the abundancy of brackish water to form an insulating top layer. Increased flow through the Bering Strait means expansion of the Arctic sea-ice in consequence with the model presented by Stigebrandt (1981).

The size of the Arctic ice-sheet interacts with climate through the albedo effect. The expanding ice-sheet may thus initiate a major Northern Hemisphere glaciation. Glaciation means that freshwater from the oceans is accumulated in the form of ice on the continents around the Arctic Basin and Northern Atlantic. Some unknown process probably involving carbon dioxide in the atmosphere ends the ice age. The ice melts and the accumulated freshwater is dumped into the North Atlantic (for topographical reasons). The net effect of the build up of land ice is thus to move freshwater from the Pacific to the North Atlantic. We thus end up with a process which reverses the spontaneous freshwater flux to the North Pacific which started the whole cycle. At this stage the process may start all over again ending up with a glaciation some 20,000 yrs. or so later as illustrated in Fig.5.

From the point of view of an oceanographer glaciations are thus merely a mechanism for the transportation of freshwater from the Pacific to the North Atlantic. We note that in the proposed glaciation cycle the time between glaciations is set by the ocean; more precisely by the time required for the ocean to spontaneously build up a sufficiently large salinity contrast between the North Pacific and the North Atlantic.

The above given discussion of the glaciation cycle is certainly very speculative and it is certainly not intended to represent the final answer. The purpose has rather been to discuss whether the ocean does at all provide a mechanism with a time-scale large enough to be of interest in this context. It is being hoped that the discussion might have given the reader some food for thought.

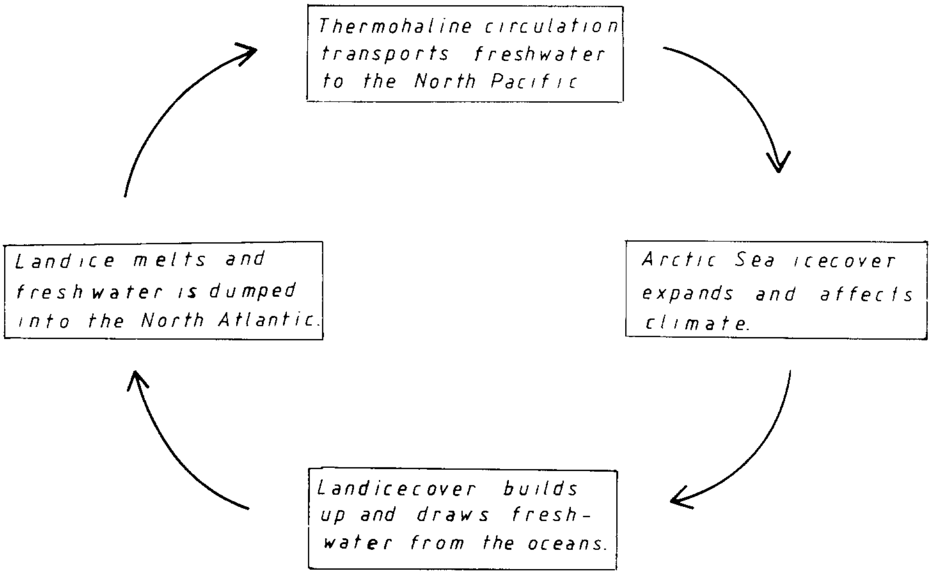


Fig.5. Illustration of the proposed glaciation cycle. The freezing and subsequent melting of land ice provides a mechanism for the transportation of freshwater from the North Pacific to the North Atlantic.

REFERENCES

Stigebrandt, A., 1981. A model for the thickness and salinity of the upper layer in the Arctic Ocean and the relationship between the ice thickness and some external parameters. J. Phys. Oceanogr., 11(10): 1407—1422.

Stigebrandt, A., 1984. The north Pacific: A global scale estuary. J. Phys. Oceanogr., 14(2): 464—470.

Geosecs Atlas, 1981. U.S. Government Printing Office, Washington, DC, Vol. 2, p. 7, Vol. 4, p. 7.