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AN ATTEMPT TO ESTIMATE THE DEEP WATER PRODUCTION FROM MOLECULAR EFFECTS.

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ABSTRACT.

The density anomaly induced in the upper layers of the ocean by the surface heat loss is determined. A two stage process is assumed, where in the first stage the density increase creates an instability and a small scale laminar convection, which redistributes the excess density over a larger depth. The larger, now dense volume, then sinks into the deeper layers as a turbulent thermal, which interacts with the underlaying waters as in a filling box.

It is found that freezing creates larger density anomalies and thus is capable of driving a deeper convection than cooling.

The results are used to estimate the deep and bottom water production in the Greenland Sea. The obtained value 0.5 . 10^6 m^3 s^{-1} is small but not unrealistic.

1. Introduction.

The renewal of deep and bottom waters in the world oceans occurs mainly at high latitudes due to freezing and cooling. To create densities high enough to make the transformed waters penetrate deep into the water column either a weak stratification or a lower boundary capable of trapping the water and allow its density to increase sufficiently, are needed (Killworth 1981).

The latter mechanism is operating over the shallow shelf areas in the Arctic Ocean, where freezing and brine release create dense, saline waters, which eventually sink into the deep basins. Similar but more sophisticated mechanisms are present in the Antarctica (Foster & Carmack 1976, Gill 1973, Killworth 1977).

The first process works in deep basins, where the surface density increases and the water column overturns, bringing the surface water directly down into the deeper layers. The most studied cases of open ocean convection are those in the Mediterranean and in the Labrador Sea (Clarke & Gascard 1983, Gascard 1978, Gascard & Clarke 1983, Killworth 1976, 1979).

The density increase is in both cases due to cooling and a deep homogenous water column is observed at the convection area. The location of the convection depends upon a preconditioning of the large scale density field and has been descibed by the authors cited above.

In the Greenland Sea, another area where deep convection is expected, no homogenous water column has, so far, been observed. In spite of the weak stratification the θ-S diagrams show a characteristic "hook" like structure with a cold salinity maximum at about 2000 m above a fresher less saline bottom water. Throughout the water column intrusions of water masses with different θ-S properties are observed (Fig. 1).

These differences from e.g. the Labradaor Sea I believe are due to the colder temperatures in the Greenland Sea, which permit freezing and a density increase due to brine release rather than to cooling. The freezing will create larger density anomalies and the intrusions and the fresh, cold bottom water could be the result of small dense "blobs" penetrating down to and spreading at their density levels. The convection would then take place in a "filling box" mode (Baines and Turner 1969) rather than manifest itself as a massive gradual overturning.

For a filling box convection to run, it is necessary that an excess density anomaly is accumulated at the surface, before the particles sinks down, so that they can by-pass the intermediate layers. Below I shall suggest a mechanism, which, starting from the molecular effects induced by the heat loss on a water particle at the sea surface, will allow such density anomaly to form (section 2). In section 3 the results are applied to the central Greenland Sea Gyre. Finally in section 4 some remarks on the effects of mechanical mixing at the surface and on the horizontal variations in the density field are made and the larger scale convection is briefly commented upon.

2. The unstable surface layer.

When no mechanical mixing is present, a perhaps unattainable ideal, the heat loss at the sea surface will penetrate down into the water mass by molecular diffusion. Following Howard (1964) the penetration depth may be written as

$$\delta = (\pi \kappa_T t)^{1/2} \qquad (1)$$

where κ_T is the coefficient of heat conduction.

If the water initially is at the freezing point the removal of heat from the layer δ will give rise to a super cooling, which leads to the formation of tiny ice crystals. These crystals then rise and accumulate at the surface, the salinity in the layer increases and the stratification becomes unstable. The instability may be described by a Rayleigh number

$$Ra = \frac{g \beta \Delta S \delta^3}{\nu \kappa_s} \qquad (2)$$

based upon the penetration depth. ν is the coefficient of viscosity and β the coefficient of salt contraction, κ_s is the diffusion coefficient of salt and ΔS is the change in salinity in the layer δ resulting from the removal of fresh water as ice. In contrast with equation 1, κ_s rather than κ_T has to be used in the denominator, since it is the diffusion of salt, not heat, which may prevent the instability from becoming critical.

For a given heat loss Q the amount of ice formed per unit area in t seconds is $QL^{-1}t$, where L is the heat of fusion. The density anomaly then becomes

$$\Delta\rho = \beta\left(S - \frac{S(\pi\kappa_T t)^{1/2}}{((\pi\kappa_T t)^{1/2} - QL^{-1}t)}\right) \qquad (3)$$

The ice crystals need a finite time to fractionate out of the layer and accumulate at the surface, and to compensate for this time lag a larger value (10^4) than costumary has been used for the critical Rayleigh number. This is an rather arbitrary choice, but it should act in the right direction. However, any Rayleigh number will be uncertain, since the profile is neither linear nor stationary.

Solving for t^*, the time required for the instability to become critical, equations 2 & 3 gives

$$t^* = \left(\frac{Ra^* \rho\nu\kappa_s((\pi\kappa_T)^{1/2} - Qt^{*1/2})}{g\beta SQ(\pi\kappa_T)^{3/2}}\right)^{1/2} \qquad (4)$$

which may be solved by iteration.

If the density increase is due to cooling instead of freezing the heat accumulated in the layer δ is

$$\rho C_p \Delta T(\pi\kappa_T t)^{1/2} = Qt \qquad (5)$$

and the density change becomes

$$\Delta\rho = \alpha\Delta T = \alpha Qt/(\pi\kappa_T t)^{1/2}\rho C_p \qquad (6)$$

The Rayleigh number is then given by

$$Ra = \frac{g\alpha Q\pi t^2}{c_p\rho^2\nu} \qquad (7)$$

which leads to the following expression for t^*

$$t^* = \left(\frac{Ra^* \rho^2 C_p \nu}{T g \alpha Q} \right)^{1/2} \quad (8)$$

In equations 6, 7 and 8 α is the coefficient of heat expansion, C_p the heat capacity, and ΔT the change in temperature.

In this case no ice crystals have to be removed from the volume for the instability to become critical and the value 10^3 is used for the critical Rayleigh number. t^* , δ^* and $\Delta\rho^*$ are shown as functions of Q for both freezing and cooling in Fig. 2. It is seen that t^* and δ^* are larger and $\Delta\rho^*$ smaller for temperature than for salt induced instability.

When the instability becomes critical, the unstable layer detaches from the surface and sinks into the waters below (Howard 1964). As the dense particles sink they will loose their excess density to the ambient water by diffusion and the descent will gradually become slower and finally stop.

To estimate the vertical velocity and the depth over which the initial density anomaly is redistributed by this small scale convection the sinking water particles are approximated by spheres with radius $(\delta^*/2)$. Using Stoke's resistance law (Batchelor 1967, Turner 1973) the velocity may be written as

$$W_0 = \frac{g \Delta\rho \left(\delta^*/2\right)^2}{3 \rho \nu} \quad (9)$$

The depth Z_0 reached by the particles depends upon how fast the density anomaly is removed. The rate of diffusion is propotional to the surface area of the particles and the density changes as (Turner 1973)

$$\Delta\rho = \Delta\rho^* \exp(-\kappa/\pi\delta^{*2}) \quad (10)$$

where κ is either κ_T or κ_S . The maximum depth then becomes

$$Z_0 = \frac{g \pi \Delta\rho^* \delta^{*4}}{12 \rho \kappa \nu} \quad (11)$$

The velocity and depth of the small scale convection driven by cooling or freezing are shown as functions of Q in Fig. 3.

It should perhaps be pointed out that for the case of freezing the water particles are larger and their densities and sinking velocities less than those found by Wakatsuchi (1983) in his study of convection beneath a growing ice sheet. The particles also become more saline with rapid cooling contrary to Wakatsuchi's results. One reason for these discrepancies could be the asumption of convecting thermals rather than viscous plumes. Another reason might be the neglect of the trapping and later release of brine in the growing ice sheet.

As long as the dense particles retain their identity during sinking the surrounding water mass will not appear denser than the underlaying waters. Not until the density anomaly has diffused from the particles to the ambient waters will the effect of the surface heat loss be redistributed over the layer Z_0 and the stratification in the upper layers becomes unstable. The dense surface layer will then start to sink into the underlaying water mass (Fig. 4).

To estimate the time required for this "larger" instability to be actualized one may take the e-folding time

$$t_0 = \frac{\pi \delta^{*2}}{\kappa} \qquad (12)$$

and the density anomaly in the layer Z_0 becomes

$$\Delta\rho_0 = \frac{\Delta\rho^*(\pi\kappa_T t^*)^{1/2}}{z_0} \cdot \frac{t_0}{t^*} = \Delta\rho^* \left(\frac{\pi\kappa}{t^*}\right)^{1/2} \frac{t_0}{z_0} \qquad (13)$$

The time required for the density to diffuse into the layer Z_0 and the density anomalies reached for cooling and freezing are shown in Fig. 5 as functions of Q .

If the entire unstable layer Z_0 leaves the surface as an isolated thermal with radius $\sim Z_0$ its vertical velocity becomes (Scorer 1978).

$$W = (g\Delta\rho_0 Z_0)^{1/2} \qquad (14)$$

Not all heat or salt will have diffused from the particles to the surrounding water after t_0 seconds, but the particles will be trapped inside the sinking thermal and eventually add their excess density. Again the velocity is shown as a function of Q in Fig. 6.

According to this picture the dense water will leave the sea surface intermittently as thermals or plumes, which sink into the interior, entraining ambient water during their descent, until they reach their density levels. An upward return flow is necessary due to mass continuity. The intensity of this flow will be weak close to the surface. Not until the diameter of the thermals has increased substantially will the upward and downward motions be equally vigorous. This would result in a more thorough mixing and homogenizition of the water column in its deeper parts than closer to the surface in spite of the larger distance from the energy source. The upper layer may then retain a rather distinct signature throughout the convection.

The thermals formed by freezing have a larger density anomaly than those formed by cooling. Freezing will then permit a deeper penetration of the thermals and a filling box model of deep convection driven by freezing might then be of some use. The resulting water column will display intrusions of convective elements spreading at their density levels and an overall θ-S structure may be retained in the presence of convection if the stability is weak enough to allow the denser parcels to by-pass the intermediate layers.

If the convection is driven by cooling, however, the density anomalies are smaller and the renewal rate higher. This implies that even if the convection occurs as a filling box the upward return flow will be intense enough to make the convection appear more like a massive overturning, gradually becoming deeper and deeper homogenizing the water column.

3. Deep convection in the Greenland Sea.

To find, if the view of the convection presented above gives sensible results, the Greenland Sea will now be considered in some detail.

In spite of its weak stratification the Greenland Sea does not exhibit a deep homogenous water column. The θ-S diagrams show a characteristic structure with a deep, cold salinity maximum at 2000 above the colder, fresher bottom water (Fig. 1). Throughout the water column intrusions of anomalous water masses are observed.

One possible interpretation of this structure is that the Greenland Sea Deep Water is formed from two basic water masses, one saline water mass with temperature ranging from -1.0 to -0.5 advecting into the Greenland Sea from the Arctic Ocean, and one convective contribution, formed locally in the Greenland Sea by freezing with temperatures at the freezing point and a narrow salinity range with a maximum sufficient to bring the convection intermittently down into the bottom layer (Rudels 1986). The intrusions would be remnants of convective events not reaching deep enough into the water column, while the fresh signature of the bottom water indicates that the local convection on the average is capable of not only freshening the advective inflow but also to by pass the Arctic Ocean Deep Water layer and renew the Greenland Sea Bottom Water (Fig. 7). The temperature of the bottom water while low is above the freeezing point, which indicates entrainment of intermediate water masses into the descending plumes or thermals.

The stability is weakest in the central Greenland Sea gyre and the Hudson station 58 (Clarke et al. 1984) may be taken as representative for the area. The upper 60 m are homogenous with the temperature at the freezing point and a salinity of 34.80. The potential density is 28.01. From 60 to 125 m there is a weak pycnocline, where the density increases to 28.05. Below 125 m the density increases slowly until it reaches the 28.08 of the bottom water at 400 m.

Disregarding the effects of wind mixing a density of 28.10 would be reached in the surface layer Z_0 with a heat loss of 300 W m^{-2}. This is probably not enough to bring a small (5-10 m) thermal down into the deeper layers in view of entrainment. However, if the upper hundred meters are first recycled as a filling box the density of the surface layer will reach 28.05 after 24-48 hours. The thermals would then attain a density of 28.14, which might be sufficient to bring even small "blobs" and thermals down to the bottom.

To estimate the amount of Deep Water produced in the Greenland Sea the net heat loss needs to be known. Bunker & Worthington 1976 (Fig. 6) put the net annual heat loss to $0.8 - 1.7 \cdot 10^9$ J m^{-2}. Assuming that this heat loss occurs in four winter months after the seasonal heat storage has been removed 40 days with a heat loss of 300 W m^{-2} is needed. During such periods the deep and bottom waters may be formed.

The convection cannot run continously, but will be cut off, when the produced sea ice becomes thick enough to inhibit the heat loss. This will probably happen, when the ice thickness has reached 50 cm or after about one week. The ice then somehow has to be removed before the Deep Water formation can start again. Six such weekly event would account for the observed heat loss and each event would in the five days the deep convection is running tranform about 150 m of water per unit area into deep and bottom water. The total production during the winter would then be 900 m per unit area.

The convection in the central Greenland Sea is expected to reach down to more than 3000 m This implies that the entire water column will not be ventilated and the concentrations of dissolved gases such as oxygen and CO_2 would be less than the saturation values.

Carmack & Aagaard (1973) estimate the area of the Greenland Sea gyre to $0.18 \cdot 10^{12}\ m^2$ and if 10% of that area has a water column as weakly stratified as the one observed from the Hudson a formation rate of $0.5 \cdot 10^6\ m^3 s^{-1}$ is obtained. The convection should be intermittent in space as well as time but 1/3 of the area would be actively convecting throughout the winter.

This production of Greenland Sea Deep Water is small but not unrealistic and clearly depends upon the assumed area capable of forming Deep Water. The entrainment of ambient waters as the thermals descend will increase the volume, but since the entire water column in the central gyre consists of Deep Water by definition ($\theta < 0$) this will not increase the production but only lead to a redistribution of the deeper water masses. Due to different cooling conditions at the sea surface the density of the sinking waters will vary and the renewal will be distributed throughout the water column.

The most serious objection to the assumption of ice formation driving the Deep Water convection in the Greenland Sea is that too much ice is produced. To supply the necessary heat 3 m of ice must form over the entire gyre, which is not realistic. However, up to now advective heat fluxes have been ignored. This may be correct for the central part of the gyre but not at its boundary, where Atlantic Water is present in a subsurface layer. If the Atlantic Water could be brought to the surface by the convection it would be capable of supplying the necessary heat, inhibit further ice production and perhaps melt already formed ice. This is what Killworth (1979) assumes for the Weddell Sea polynya. If Atlantic Water is brought to the surface over 2/3 of the gyre the net ice production is reduced to 1 m which is more plausible.

It should be noted that the removal of ice by the bringing up of Atlantic Water also will stop the deep water formation, since the density anomaly, which may be accumulated at the surface now will be smaller than during freezing. The Atlantic Water will by cooling and freshening be transformed into upper Artctic Intermediate Water (Swift & Aagaard 1981). With a cooling of 1^0C of the Atlantic Water the production of uAIW may be estimated from energy balances to $1 \cdot 10^6$ m^3 s^{-1}. This requires that the residence time for the Atlantic Water in the Greenland Sea is more than a year.

5. Discussion.

In the preceding sections the wind generated mechanical mixing has been ignored and the convection has also been considered apart from larger scale variations in the density field and the stratification has been assumed horizontally homogenous.

The turbulence will distribute the cooling and freezing over a deeper layer. Ice crystals will form out of water particles, which then become saline. The water particles and the crystals subsequently separate. The crystals rise and the saline particles sink down just as vizualized in the description given above. The turbulent eddies will probably not disrupt the actual convection due to its small scale, but the thermals are likely to become larger and less dense by incorporation of ambient water. How much is just a guess.

Large and meso scale horizontal motions induce variations in the thickness of the stable surface layer. Areas with thinner upper layers could be sites, where the "blobs" and thermals initially break through and then merge to form more massive sinking plumes. The necessary return flow would then be horizontal rather than vertical and the convecting plumes may drive motions similar to tornadoes and dust devils. The convection would in this case create and/or reinforce a horizontal circulation.

If such a coupling exists between the small scale density build up and the larger scale density field, the problem of how the vertical return flow may be maintained in the presence of a large number of sinking thermals, growing as they descend and gradually blocking the return flow, could be avoided. In a horizontally uniform situation the descending plumes and the upward return flow probably have to be considered together below the upper few hundred meters.

These remarks are, however, just speculations on how one might proceed to integrate the approach suggested here with the large scale motions.

From the result presented above it should be possible to estimate the volume, which is ventilated in a certain area, from the surface heat loss and the stratification.

When the stratification is strong, the same water will recirculate several times as a filling box, gradually increasing its density before it penetrates further. The amount of ventilated water is then determined by the vertical density structure and can be found by observing the stratification in the winter months.

On the other hand, if the stratification is weak the convection reaches deeper but not all the intermediate water is brought to the surface and ventilated in one winter season. The volume of ventilated water is determined by the heat loss, not by the stratification.

These features will not change if larger scale circulations are taken into account and the information could perhaps be used not only when convection, water mass productions and the deep circulation are studied but also in climatological problem such as the rate of CO_2 uptake by the oceans.

Table I

Used parameter values

$\alpha = 0.8 \cdot 10^{-4}\ {}^{0}C^{-1}$	$\nu = 0.015\ cm^2/s$	$\kappa = 10^{-5} \cdot cm^2/s$
$\beta = 8 \cdot 10^{-4}$	$\kappa = 0.001\ cm^2/s$	$S = 34.8$
$g = 1000\ cms^{-2}$	$L = 80\ cal\ gram^{-1}$	$C_p = 1\ cal\ gram^{-1}\ {}^{\circ}C^{-1}$

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Figure captions:

Figure 1: θ-S diagrams for 7 stations taken by m/s Polarsirkel in March 1980 in the Greenland Sea along $74^0 15'$N.

Figure 2: a) Time required for the instability at the surface to become critical as function of Q.

b) Penetration depth δ^* as a function of Q.

c) Density anomaly $\Delta\rho^*$ in the layer as a function of Q.

Solid line: freezing.
Broken line: cooling.

Figure 3: a) Vertical velocity W_0 of the sinking laminar termals as function of Q.

b) Depth Z_0 reached by the laminar convection as function of Q.

Solid line: freezing.
Broken line: cooling.

Figure 4: The two stage convection process.

a) The lower temperature penetrates down into the upper, ice forms and accumulate at the surface.

b) The unstable layer sinks down as laminar plumes or thermals in the interior of neutrally stratified surroundings.

c) The density anomaly diffuse into the ambient water which becomes dense with respect to the layers beneath.

d) The thicker unstable layer sinks as a turbulent mass convection. The return flow increases in intensity in the deep due to the entrainment into the thermals.

Figure 5: a) Time t_0 rewuired for the density anomaly to diffuse into the ambient water as function of Q.

b) Density anomaly $\Delta\rho_0$ of the layer Z_0 after t_0 as function of Q.

Solid line: freezing.
Broken line: cooling.

Figure 6: Vertical velocity of the turbulent thermals as they leave the sea surface as function of Q.

Solid line: freezing.
Broken line: cooling.

Figure 7: The formation of Greenland Sea Deep Water. Water made dense by freezing at the surface sinks through the intermediate layers. The surface and entrained waters together with advected Arctic Ocean Deep Water maintain the θ-S structure of the Greenland Sea. The ratio of advective to convective contribution is estimated to 2:1. (Rudels 1986.)

FIGURE 1.

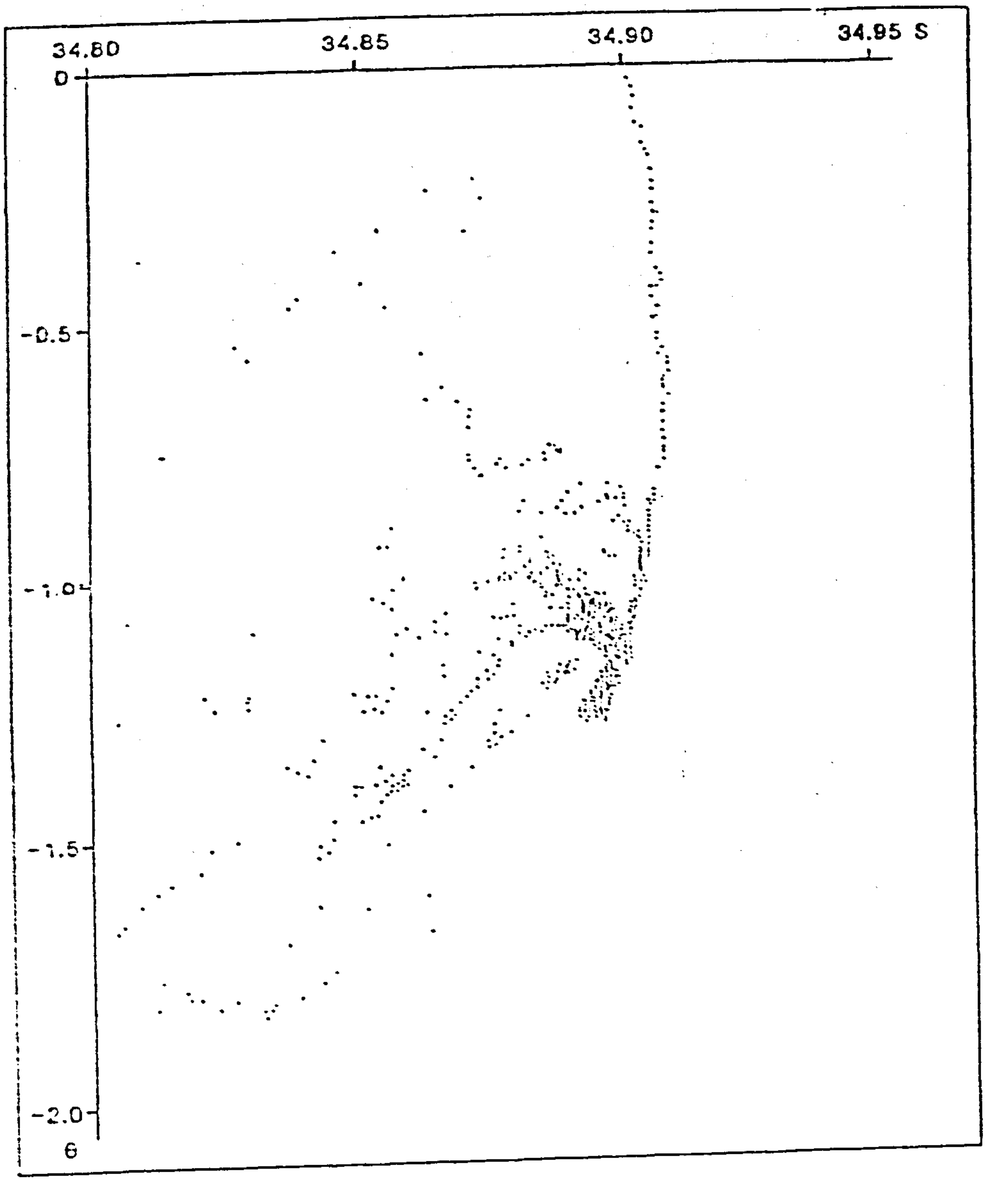
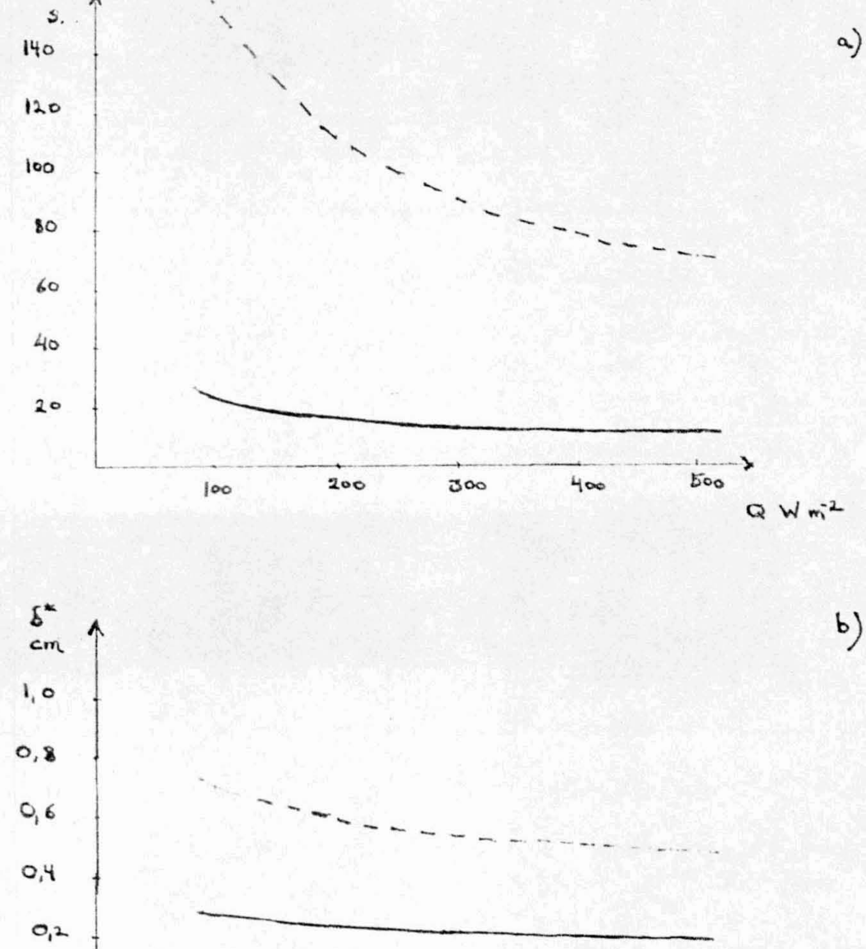


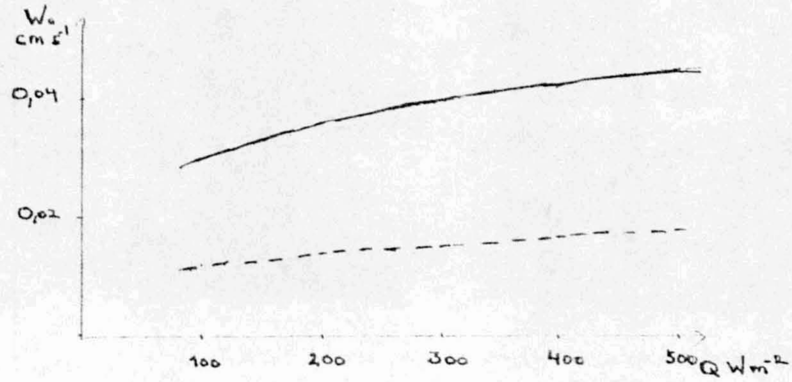
FIGURE 2



a)
t^*
s.
140
120
100
80
60
40
20
100
200
300
400
500
Q W m^{-2}
b)
δ^*
cm
1,0
0,8
0,6
0,4
0,2
100
200
300
400
500
Q W m^{-2}
c)
$\Delta\rho^*$
g cm^{-3} 10^6
2,4
2,0
1,6
1,2
0,8
0,4
100
200
300
400
500
Q W m^{-2}

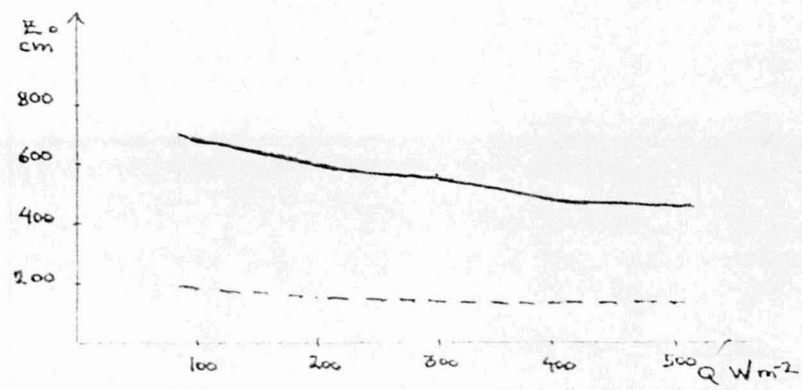
FIGURE 3

a)



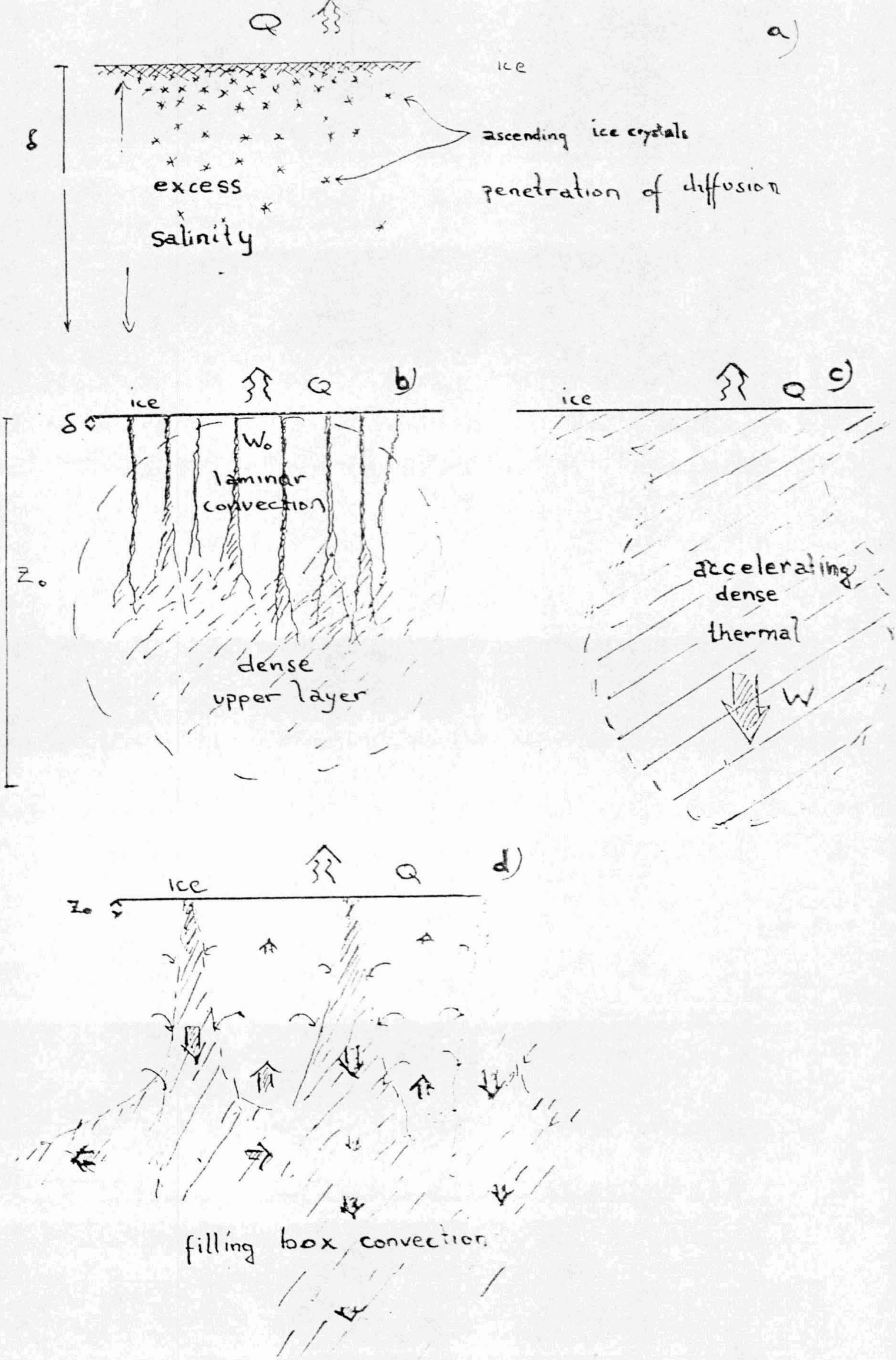
W₀ cm s⁻¹
0,04
0,02
100
200
300
400
500
Q W m⁻²

b)



z₀ cm
800
600
400
200
100
200
300
400
500
Q W m⁻²

FIGURE 4



a)
Q
ice
δ
ascending ice crystals
penetration of diffusion
excess
salinity
b)
Q
ice
δ
Wo
laminar
convection
Zo
dense
upper layer
c)
ice
Q
accelerating
dense
thermal
W
d)
ice
Q
Zo
filling box convection

FIGURE 5

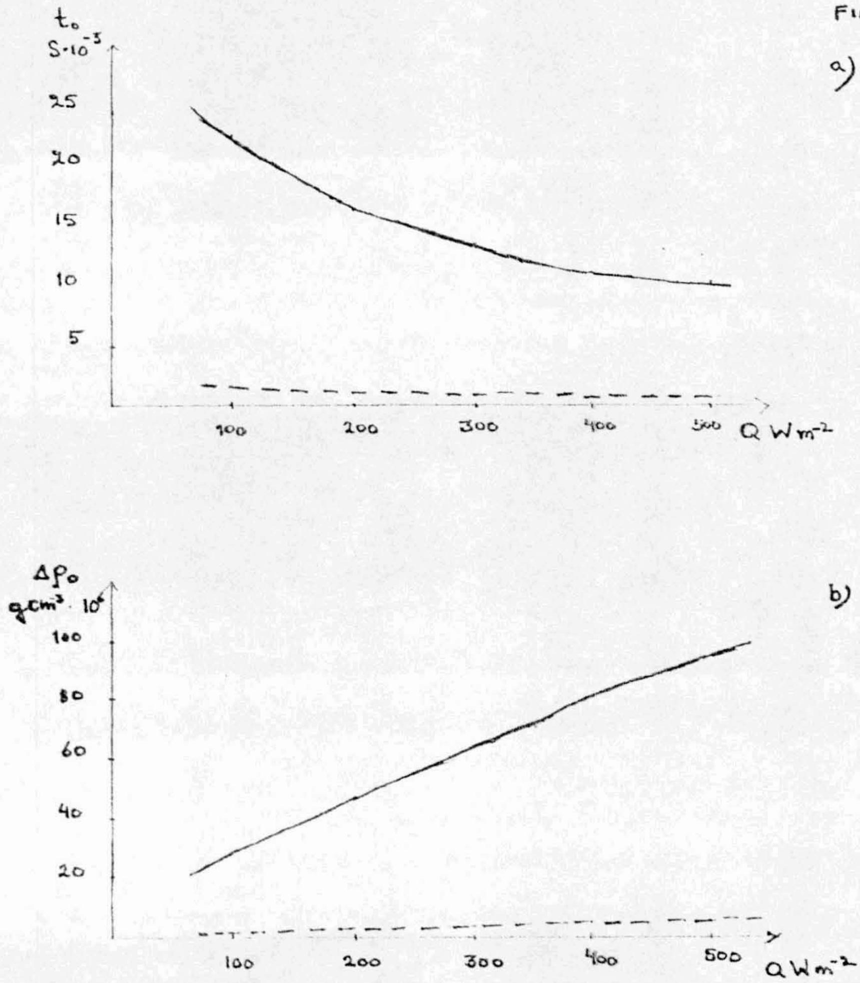


FIGURE 6.

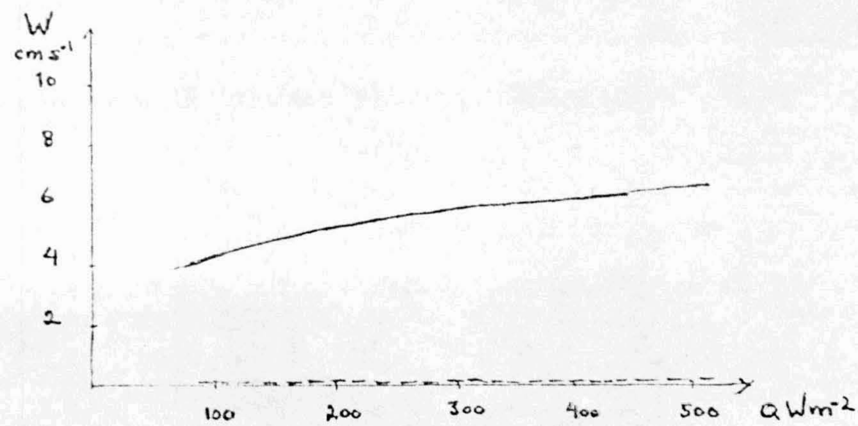
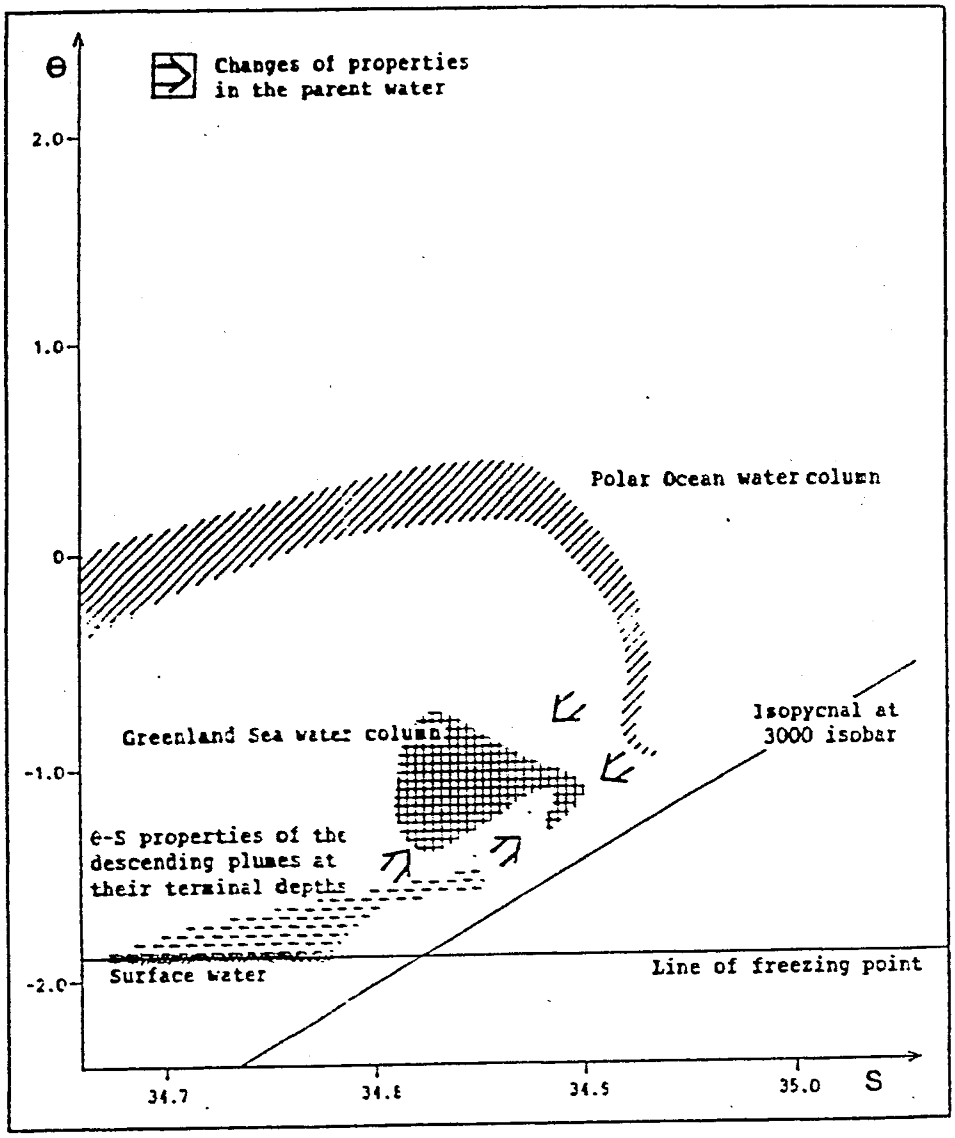


FIGURE 7.


Changes of properties in the parent water
Polar Ocean water column
Greenland Sea water column
Isopycnal at 3000 isobar
θ-S properties of the descending plumes at their terminal depths
Surface water
Line of freezing point
θ
2.0
1.0
0
-1.0
-2.0
34.7
34.8
34.9
35.0
S