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Natural Desalination and Equilibrium Salinity Profile of Old Sea Ice

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Abstract

Owing to the great local and temporal scatter of ice salinity, the shape of its steady-state profile in ice of equilibrium thickness is only approximately known. Hence the purpose of theorizing on the way in which it establishes itself is to suggest pertinent experiments rather than to explain physical causes. Four mechanisms of salt migration are discussed:

1) "Brine pocket diffusion", as explained by Whitman. It is too slow to be of significance here.

2) "Gravity drainage", as observed in the laboratory by Kingery and Goodnow. It is unlikely to occur in natural, thick floating ice. However, their basic concept may be applicable in a modified form as

3) "Flushing" or washing-out. A quantitative calculation which assumes the replacement of brine by meltwater from the surface to be a function of ice salinity and maximum temperature leads to a steady-state salinity profile similar to that suggested by observations.

4) "Brine expulsion" as a result of temperature changes and the separation of liquid and gaseous inclusions during the cooling cycle (Nakaya, Bennington) is also treated numerically and results, as does 3), in a steady-state salinity profile resemblant of observations.

I. Introduction

It is well known that sea ice whose salinity immediately after formation may be between 3 and 25 per mil, depending on growth-rate and other parameters (Weeks and Lofgren, 1967) becomes nearly salt-free after a few years of exposure to its natural environment. As the ice reaches an equilibrium thickness after several years of undisturbed growth in a given climatic environment it must also develop an equilibrium salinity distribution.

The following discussion is not meant to be a description or explanation of natural desalination of sea ice but rather an appraisal of possibilities which will have to be investigated by experiments.

In the state of equilibrium between surface ablation and bottom accretion, the velocity of upward ice displacement, w, results in an upward flux of brine with a flux density of $w \cdot s$. (s is the salinity of the ice in g/cm^3. This unit will be used for flux calculations. For general discussion, salinities will be given in the conventional units of per mil.) To maintain a steady-state salinity distribution there must be a downward flux of salt, F_s, in which the brine moves relative to the ice. Hence, for any whole number of years,

$$F_s(z) = w \cdot s(z)\,. \qquad (1)$$

While a direct observation of F_s would be difficult, w and $s(z)$ are, at least in principle, easy to measure in the field. The potential usefulness of approaching the problem of brine migration by way of the equilibrium salinity profile lies in the fact that, once $F_s(z)$ is known, one can attempt to relate it to other parameters such as temperature, rate of temperature change, brine volume, etc.

II. Field Data

Unfortunately, steady-state sea ice is an abstraction not found in nature because the environment changes from year to year and over the distance covered by the drift of the ice. Furthermore, our standard sampling procedures are not adequate to overcome the large local scatter of ice salinity, as shall be demonstrated by the following example.

A set of 12 cores in a grid of three by four meters was taken from ice of 200 cm thickness which was known to have grown without breakage or flooding. The cores were taken with a 7.5 cm SIPRE corer which has become the most widely used, more or less optimal hand-tool for that purpose. The cores were cut in sections of 10 cm length, melted, and the salinity of the melt was measured by hydrometers with special care to the best accuracy this instrument allows. The mean salinities and their standard deviations at each depth are given in Table 1.

Table 1. Ice salinities (‰) and their standard deviations, σ, in a sample of 12 cores taken at 1 m grid distance from sea ice of 200 cm thickness, at Drifting Station ARLIS II, 1964

Depth cm	0–10	10–20	20–30	30–40	40–50	50–60	60–70	70–80	80–90	90–100
‰	8.0	5.4	5.6	5.9	6.1	6.6	6.4	6.1	6.0	6.2
σ	2.86	0.91	0.36	0.51	0.59	0.74	0.81	0.78	0.80	0.61

Depth, cm	100–110	110–120	120–130	130–140	140–150	150–160	160–170	170–180	180–190	190–200
‰	6.1	5.9	6.0	6.1	5.8	5.8	5.4	4.9	5.1	5.8
σ	0.67	0.91	0.72	0.70	0.69	0.74	0.58	0.83	0.66	1.32

To test the representativeness of these mean values we require that, with a probability of 90%, the observed mean should lie within ± 0.2‰ of the true mean. Given a mean standard deviation of about 0.8 (and assuming random scatter) the test (Bochow *et al.*, 1958) indicates that we need about 13 cores. If we wish the observed mean to lie within ± 0.1‰ of the true mean, then we need over 40 cores. Thus it is obvious that the salinity profiles obtained from one or a few cores have little significance. A similar degree of small-scale local scatter was found in ice both thicker and thinner than the example given here. Adding to this the large-scale scatter in time and space mentioned earlier, it is evident that, in order to find the exact mean steady-state salinity profile, we need ice samples of unmanageable size or number. The best documented mean profile in perennial ice is the one given by Schwarzacher (1959) which is reproduced as curve S in Fig. 1. The probable error of this curve is not known. The available data on the average vertical ice velocity w for equilibrium thickness ice in the Central Arctic are somewhat more reliable (Hanson, 1964; Untersteiner, 1964). An average ablation of 40 cm of ice (30 cm at the surface, 10 cm at the bottom), leads to $w=30$ cm/year. On this basis, we shall investigate four mechanisms of brine transport and their possible role in producing a salinity profile of the shape shown in Fig. 1.

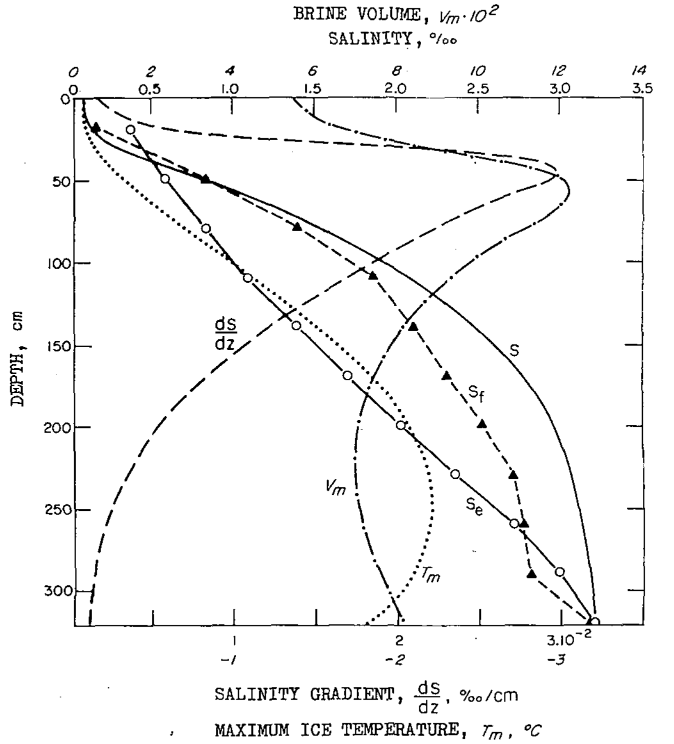

Fig. 1. S and dS/dz = salinity and salinity gradient in ice (after Schwarzacher, 1959)
T_m = highest ice temperature attained by 20 August, the average date of freeze-up (after Untersteiner, 1964)
V_m = highest brine volume attained by 20 August, calculated from T_m and S
S_f = salinity profile calculated for desalination by "flushing"
S_e = salinity profile calculated for desalination by "expulsion"

III. Solute Diffusion

Whitman (1926) has shown that a temperature gradient in the ice establishes a concentration gradient within the brine pocket if phase equilibrium is to be maintained at the brine-ice interface. The resulting diffusion of solute dissolves ice at the warm end and causes freezing at the cold end, and the entire pocket migrates toward the warm side. The results of laboratory experiments as well as a basic mathematical formulation of the velocity of migration were given by Kingery and Goodnow (1963, p. 237). A more thorough theoretical treatment was given by Tiller (1963) in the general context of zone melting, and further laboratory observations were reported by Hoekstra *et al.* (1965). Both Kingery and Hoekstra find a certain disagreement between theory and observation which seems to be due partly to simplifications in formulating the theory and, to a larger degree, to some uncertainty about the diffusivity of sea salts in water. In any case, theory and observation give migration velocities which make pure brine diffusion irrelevant in the present context. The velocity of migration is roughly $(-D/T)\cdot(\mathrm{d}T/\mathrm{d}z)$ where D is the diffusivity of sodium chloride and T the temperature in °C. If we insert representative values for T (Untersteiner, 1964) it turns out that, for instance, between 10 and 20 cm below the surface, the downward migration between August and April is about 2 cm, and it is compensated by an almost exactly equal upward migration between May and July.

IV. Gravity Drainage

Kingery and Goodnow (1963) also performed experiments on the brine drainage by gravity and concluded that this "…is the primary mechanism by which salt is eliminated from new sea ice…" This is doubtless true for high salinity and high brine volume ice, and particularly for ice which is, as in their experiments, not floating on water. It is unlikely, however, that gravity drainage is important in removing salt from natural floating sea ice of more than a few decimeters thickness as, in this case, the only force acting on the brine is due to the small difference in specific gravity between brine and sea water.

V. Flushing

A process similar to gravity drainage may be called "flushing." We assume that, as in the case of gravity drainage, the brine moves in interconnected tubes and cavities but not by its own gravity. The force to overcome capillary retention is the hydrostatic head which exists when snow or ice melt at the surface. Hence, this process should be limited in time to the melting period and to the period when the ice above the freeboard level (usually 20 to 40 cm in old ice) is permeable and allows the meltwater to penetrate. Most of the meltwater formed during the summer drains over the edges of the floes. Whether appreciable amounts seep through ice of 3 to 4 m thickness is not certain. Some meltwater ponds retain a water surface higher than the sea level throughout the summer. The variable rate of melting and lateral water movement through the highly porous surface layer would make it difficult to ascertain the actual amount of vertical seepage. Dye experiments performed by Bennington (personal communication) indicate that some seepage should occur.

For the sake of a quantitative formulation we assume that each year a certain part of the brine in the ice is replaced by fresh water percolating down from the surface. A steady-state salinity profile would be established when the downward flux of salt, F_s, is equal to the upward flux, $w \cdot s$, mentioned earlier. To further specify F_s we make the following tentative assumption: when the brine volume is extremely large (say 20 or 30%) then all cavities are interconnected and all the brine can be flushed out. On the other hand, there must be a lower limit to the brine volume at which none of the cavities are interconnected and no brine can be flushed out. In other words, we postulate that $\partial F_s/\partial z$ is some function, f, of the greatest brine volume, V_m, attained during the time when meltwater is available at the surface. In the Central Arctic the average melt season lasts from about 1 June to about 20 August. If z denotes the vertical coordinate, positive downward, then the stationary case is described by

$$\frac{\partial s}{\partial t} = -w\frac{\partial s}{\partial z} + f(V_m) \equiv 0\,. \tag{2}$$

Since (Assur, 1958)

$$V_m = c\frac{s}{T_m}\,, \tag{3}$$

with c=const. $\cong$55, and T_m=highest temperature attained during the year, we have

$$w\frac{\partial s}{\partial z} = f\left(c\frac{s}{T_m}\right). \tag{4}$$

V_m is given in cm^3 of brine per cm^3 of ice.

Figure 1 shows T_m as a function of z as calculated in a previous investigation (Untersteiner, 1964). It is the highest temperature reached at each depth by 20 August. Hence, the points composing the curve pertain to dates which progress from the surface down. Also shown in Fig. 1 are the profiles of V_m according to eq. (3), and of dS/dz. The similarity of the two curves indicates that indeed the greatest loss of salt occurs at the level where the brine volume reaches its highest value. Figure 2 is a plot of $w \cdot ds/dz$ versus V_m (eq. 4) for each 10 cm layer. Despite the large scatter, some relationship is discernible. A linear function would make the salt advection zero at V_m=0.05 indicating that at a brine volume of 5% the ice is no longer permeable. This is not in accordance with laboratory observations. Therefore, a simple, non-linear function was chosen (dashed line in Fig. 2) because it gives a reasonable approximation in the range of the data and allows for some permeability of the ice even at an extremely small brine volume. In practice, the rate of salt loss at those brine volumes is unimportant because the present

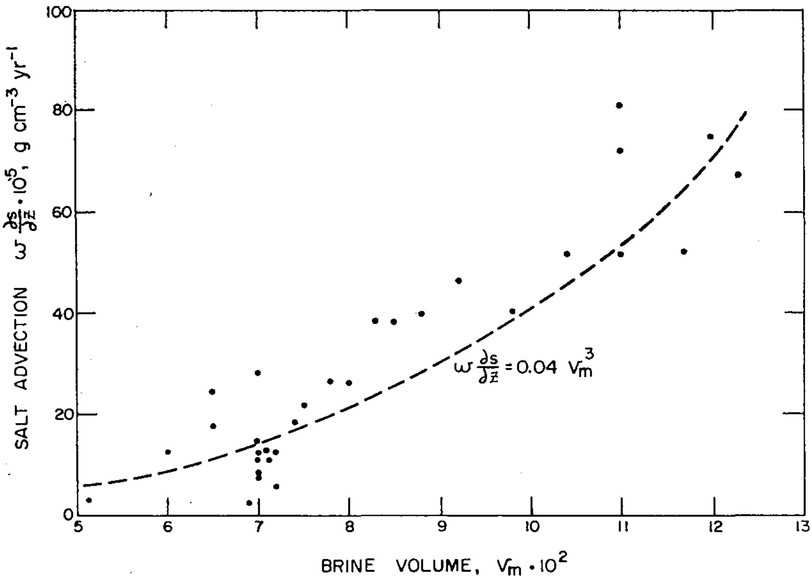

Fig. 2. Relationship of maximum brine volume and salt advection required to maintain the steady-state salinity profile shown as curve S in Fig. 1

argument is predicated on the availability of meltwater from the surface and only pertains to a time when the ice is warm and the brine volume large. As can be seen in Fig. 1, V_m is always greater than 5%.

Our steady-state case is then

$$w\frac{\partial s}{\partial z} \equiv w\frac{\mathrm{d}s}{\mathrm{d}z} \equiv \frac{\mathrm{d}s}{\mathrm{d}t} = 0.04\ V_m^3\ , \qquad (5)$$

which we can use to re-calculate the salinity profile, beginning at the bottom line of Fig. 1 ($z=320$ cm, $S=3.2‰$), proceeding at 30 cm or 1 year increments. The result, curve S_f in Fig. 1, resembles curve S well within the range of its uncertainty.

What has been shown is merely that meltwater flushing can explain a gradual desalination of an individual piece of ice during its ascent from the bottom of the ice sheet to the surface.*

It must be pointed out here that the preceding argument may be nothing more than a proof of its own internal consistency: We began with a given salinity distribution and certain boundary conditions at the surface and the bottom of the ice and calculated (Untersteiner, 1964) the temperature field. Owing to the dependence of the specific heat of sea ice on salinity, which is particularly strong near 0°C, the maximum temperature at each level is influenced by the salinity distribution. The latter two parameters determine the maximum brine volume any, in our model, the rate of desalination which, in turn, led us back to the salinity profile introduced originally. It may well be that the reasoning applied here could "explain" and number of salinity distributions. To investigate this, one would have to re-compute the temperature field for each case, which seems hardly warranted without knowing the real mechanism of desalination.

VI. Expulsion

An alternative mechanism of natural desalination may be a phenomenon originally described by Nakaya (1956, p. 15). He observed that the vapour bubble in a Tyndall figure became separated by a bridge of ice when the sample was cooled, and that the mounting pressure in the liquid caused the ice to fail along the crystallographic basal plane, allowing some liquid to escape and migrate toward the warm end of the sample. Bennington (1963) has corroborated Nakaya's observations and proposed that, whenever sea ice is cooled, brine is expelled.

The following calculations are based on an extremely schematized model of this process: While the temperature of the ice rises, some of the ice surrounding the brine pocket melts and a vapour bubble with a volume of about 1/10 of the volume of the melted ice forms. When the temperature of the ice drops again, the vapour bubble is separated immediately, a certain volume of ice accretes at the walls of the pocket and expels an equal volume of brine which is assumed to leave the ice entirely.

If s denotes the bulk salinity of the ice and s_b that of the brine, then the loss of salt, $-\mathrm{d}s$, by the change of brine volume during cooling, $-\mathrm{d}V$, is

* A simple calculation shows that a cylindrical hole with a diameter of the order of millimeters, and filled with fresh water, would heal in a matter of hours, even at the time of maximum ice temperatures.

$$-\mathrm{d}s = -\Delta\rho\, s_b\, \mathrm{d}V\,, \tag{6}$$

where $\Delta\rho \cong 0.1$ is the difference in density between water and ice. No change of s occurs when the ice is warming and V is increasing. The salinity of brine is a function of temperature alone; in the range of temperature which pertains here we can write approximately

$$s_b = T/c\,, \tag{7}$$

with $c \cong 55$, and T expressed in negative degrees Celsius. Combining eqs. (6) and (7) we have

$$\frac{\mathrm{d}V}{\mathrm{d}s} = \frac{1}{\Delta\rho}\frac{c}{T}\,. \tag{8}$$

Ice salinity, brine volume, and temperature are approximately related by

$$V = c\frac{s}{T}\,. \tag{9}$$

Differentiating eq. (9) and combining it with eq. (8) gives

$$\frac{1}{\Delta\rho}\frac{c}{T} = -c\frac{s}{T^2}\frac{\mathrm{d}T}{\mathrm{d}s} + \frac{c}{T}\,, \tag{10}$$

which is readily integrated between the boundaries T_0 and T and s_0 and s:

$$s = s_0\left(\frac{T_0}{T}\right)^{\frac{\Delta\rho}{1-\Delta\rho}}\,. \tag{11}$$

This equation holds only while the temperature is continuously dropping from T_0 to T and it describes the concomitant decrease of the bulk ice salinity from its initial value s_0. Hence the salinity of the ice at any time $t>0$ should depend on its initial value at $t=0$ and on all the periods of cooling that it has undergone in its history.

For a crude test of this theory we use the thermal history of a particle of ice starting at the bottom of an ice sheet as computed in a previous study (Untersteiner, 1964) and shown in Fig. 3. If we apply eq. (11) successively between the peaks (T_0) and valleys (T) of curve A, beginning at a salinity of 3.2‰ at the bottom, we obtain the

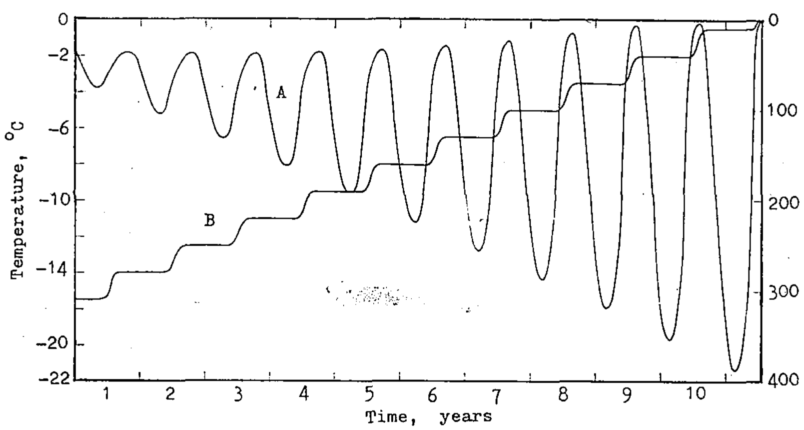

Fig. 3. Thermal history of a particle of ice (curve A) starting at the bottom of the ice sheet and rising (curve B) to the surface in 11 years (after Untersteiner, 1964)

salinity profile designated S_e in Fig. 1. Like S and S_f, it arrives at a salinity of about 0.3‰ a few decimeters below the surface, but it does not show the curvature of the other profiles. In evaluating eq. (11) only the mean annual course of temperature was considered. Allowing for some additional superimposed fluctuations caused by changes of the weather would shift the upper portion of curve S_e toward somewhat lower salinities.

VII. Conclusions

Two possible mechanisms of natural sea ice desalination are the flushing-out of brine by surficial meltwater and the expulsion of brine by changes of volume during periods of cooling. Both were treated quantitatively by crude models, and both rendered salinity profiles within the range of certainty with which the true equilibrium profile is known. In both models, starting with a salinity of several parts per thousand in the first year, the ice reaches the surface nearly fresh, with a salinity of a few tenths of a part per thousand. Flushing as the predominant mechanism implies that the ice can lose salt only during the summer. Schwarzacher (1959) states, however, that he did not detect a systematic decrease of salinity during the melt season. As stated at the outset, the purpose of this investigation was to point out existing possibilities and to show the need for more experimental research.

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References

1) ASSUR, A. 1958 Composition of sea ice and its tensile strength. *In* Arctic Sea Ice, *Nat. Acad. Sci.-Nat. Res. Council U.S.A., Publ.* **598**, 106–138.

2) BENNINGTON, K. O. 1963 Some crystal growth features of sea ice. *J. Glaciol.*, **36**, 669–688.

3) BOCHOW, W., HOEHNE, W. and RAEUBER, A. 1958 Über die Bestimmung der für einen repräsentativen Mittelwert notwendigen Anzahl von Einzelwerten. *Angew. Meteorol.*, **3**, 170–173.

4) HANSON, A. 1965 Studies of the mass budget of arctic ice floes. *J. Glaciol.*, **5**, 701–709.

5) HOEKSTRA, P., OSTERKAMP, T. E. and WEEKS, W. F. 1965 The migration of liquid inclusions in single ice crystals. *J. Geophys. Res.*, **70**, 5035–5041.

6) KINGERY, W. D. and GOODNOW, W. H. 1962 Brine migration in salt ice. *In* Ice and Snow (W. D. KINGERY, *ed.*), M. I. T. Press, Cambridge, Mass., 237–247.

7) NAKAYA, U. 1956 Properties of single crystals of ice, revealed by internal melting. *CRREL Res. Paper*, **13**, 1–80.

8) SCHWARZACHER, W. 1959 Pack-ice studies in the arctic ocean. *J. Geophys. Res.*, **64**, 2357–2367.

9) TILLER, W. A. 1963 Migration of a liquid zone through a solid. *J. Appl. Phys.*, **34**, 2757–2762.

10) UNTERSTEINER, N. 1964 Calculations of temperature regime and heat budget of sea ice in

the central arctic. *J. Geophys. Res.*, **69**, 4755–4766.

11) WEEKS, W. F. and LOFGREN, G. 1967 The Effective solute distribution coefficient during freezing of salt solutions. *In* Physics of Snow and Ice, Part 1 (H. ÔURA, *ed.*), Inst. Low Temp. Sci., Sapporo, 579–597.

12) WHITMAN, W. G. 1926 Elimination of salt from sea water ice. *Amer. J. Sci.* 5, **11** (62), 126–132.