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Fluxes Associated with Brine Motion in Growing Sea Ice

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Summary. Laboratory and field studies have demonstrated that fluid motion occurs at two locations in growing sea ice: in a network of brine channels and within the skeletal layer at the ice-water interface. Brine channel fluxes estimated using brine channel areal density from natural sea ice and channel velocities from laboratory studies are compared with recent measurements reported in the literature. Fluxes into the porous skeletal layer of sea ice may be estimated using rates of nutrient uptake by ice algae and adjacent seawater nutrient concentrations. Both approaches indicate fluxes of the order of 10^{-6} cc cm^{-2} s^{-1} (l m^{-2} h^{-1}), which are about equal to fluxes reported in bioirrigated sediments. Fluxes of this magnitude indicate a very short residence time for the liquid phase in the skeletal layer, suggesting that this fluid motion may be important in maintaining the ice algae community.

Introduction

Interfaces, such as the air-sea interface and the sediment-water interface, are well known as the sites of mass transport processes and enhanced biological and chemical activity. The sea ice-water interface is also recognized as a biologically active interface (Alexander and Chapman 1981) and serves as substrate for a population of epontic ice algae. These algae, consisting largely of diatoms, are responsible for the first significant primary production in ice-covered seas each spring. Physical processes that permit flushing at the sea ice-water interface are known, but until recently, no direct flushing rate measurements have been available. This paper summarizes recent field and laboratory work on the sea ice interface and uses the results to place bounds on volume fluxes across this interface. These fluxes are then compared with fluxes across a more familiar interface where biological irrigation rates have been measured

recently, the sediment-water interface. The sea ice-water interface, with circulation driven by purely physical processes, appears to be as actively flushed as intensively bioirrigated sediments.

Sea Ice Structure

Sea ice has a distinctive structure, due to the presence of brine during freezing and to brine drainage features that form following initial freezing. Brine channels consist of vertical channels connected with smaller radially-oriented, downward-sloping tributary drainage channels. Lake and Lewis (1970) examined features of brine drainage channels by excavating a large specimen of natural sea ice and dissecting it in slabs normal to the vertical axis. Freezing of surface water obscured the ice-water interface, limiting studies of drainage features in this area. Channels with diameters greater than 5 mm occured at a density of 1 per 180 cm^2. The vertical channels approached 10 mm in diameter near the ice-water interface and decreased vertically to a vanishing point at about 90 cm; the tributary channels were 2 – 8 mm in diameter and 2 – 3 cm long. The channels frequently have constrictions at the ice-water interface (Eide and Martin 1975). The average diameter for the overall pattern was about 4 cm. Martin (1979) indicates that eutectic melting and drainage of high salinity surface brines may lead to top-to-bottom brine channels in multi-year sea ice.

The high porosity (15% brine volume) skeletal layer occupies the bottom 10 to 30 mm of growing sea ice. The sea ice algae responsible for the so-called brown zones inhabit the skeletal layer. Ice platelets are separated by brine in this region, forming grooves and slots.

Ice bridges the volume between platelets at the top of the skeletal layer, isolating brine in pockets and tubes located along crystal boundaries. Lake and Lewis (1970) reported that brine tubes (0.4 ± 0.13 mm diameter) occurred at the 1 cm level of natural sea ice with a density

of 42 tubes cm^{-2}. These brine tubes are 2−3 cm long and are typically closed at the top with a bulbous cap.

Water Motion in Sea Ice

Brine Channels. Eide and Martin (1975) and Niedrauer and Martin (1979) produced laboratory sea ice using a plexiglas freezing cell designed to limit lateral heat fluxes. Eide and Martin (1975) observed brine drainage channels similar to those in natural sea ice in their laboratory ice and demonstrated oscillatory motion in the brine channels using dyes introduced to the adjacent seawater and generated electrolytically within the brine drainage channels. They reported oscillations with a period of roughly one hour, the outflow being about three times longer than the inflow. The oscillatory flushing was explained by generation of internal pressures and a "salt oscillator" effect (Martin 1970) caused by density differences and formation of a constricting neck at the bottom of the channels. Dye experiments (Niedrauer and Martin 1979) permitted estimates of outflow velocities of 0.25 mm s^{-1} at the mouth of a 1.4 mm diameter channel. Another series of observations on a dye particle in a channel of unspecified diameter yielded an average velocity of 0.044 mm s^{-1}. Brine velocities down to 6×10^{-3} mm s^{-1} were observed. Niedrauer and Martin (1979) concluded flow in the channels was convective and observed that the flow was bi-directional. Incoming sea water occupied the top portion and outflowing brine occupied the bottom portion in sloping channels.

Skeletal Layer and Brine Tubes. Lake and Lewis (1970) reported temperature time-series measurements at the freezing interface of natural sea ice. These measurements were obtained by mounting a thermistor array vertically below the ice and allowing the growing ice to envelop the thermistors. The time-series showed roughly periodic temperature fluctuations that could not be explained by diffusive propagation of surface temperature changes. These temperature fluctuations were taken as evidence of brine movement in the skeletal layer.

Niedrauer and Martin (1979) equipped their plexiglas freezing cell with an array of thermocouples to study the advancing ice-water interface in detail. Dye experiments showed that seawater advanced up into the porous skeletal layer on a series of broad fronts and exited the ice in discrete plumes. These plumes were separated by about 10 mm and extended down to the ice interface. Temperature fluctuations of the same magnitude (0.05 °C) as those reported by Lake and Lewis (1970) were observed as the freezing front passed the thermocouple array. These temperature variations were apparently associated with the brine plumes resulting from convection in the skeletal layer. Both Lake and Lewis (1970) and Niedrauer and Martin (1979) showed that brine tubes with diameters less than about 1 mm cannot permit convection, but that convection does occur in the skeletal layer.

Results

Flux Estimates and Measurements

Brine Channels. The fluxes associated with convective motion in brine channels may be estimated using the reported velocities for individual channels and areal density of the brine channels. This approach uses limited laboratory data combined with field data on channel densities and neglects the effects of shear between seawater and sea ice, but it can be used to establish lower limits and orders of magnitude for these fluxes.

Niedrauer and Martin's (1979) measured velocity of 0.25 mm s^{-1} in a 1.4 mm diameter channel gives a flow of 0.38 mm^{-3} s^{-1}. The lowest velocity reported (6×10^{-3} mm s^{-1}) applied to the same size channel gives a flow of 9.2×10^{-3} mm^{-3} s^{-1}. Assuming brine channels of this small diameter occur at the same density as larger diameter brine channels (1 per 180 cm^2 or 55.5 m^{-2}), fluxes are 5×10^{-8} to 2.1×10^{-6} cc cm^{-2} s^{-1}. If these velocities apply to 5 mm diameter channels at the above density, the calculated flux increases by a factor of 12.7. If the observed velocities are adjusted with a r^2 correction (Schlichting 1968) to velocities in 5 mm diameter channels, the velocities and fluxes increase by an additional factor of 12.7 ($(2.5/0.7)^2$). The water motion in brine channels is either oscillatory (Eide and Martin 1975) or bi-directional (Niedrauer and Martin 1979), so the outward flux should be about half of those calculated here, or 4×10^{-6} to 1.7×10^{-4} cc cm^{-2} s^{-1}. Depending on values chosen for channel velocity, channel diameter, channel density, velocity scaling and the outflow: inflow ratio, calculated outward fluxes range between 10^{-4} and 10^{-8} cc cm^{-2} s^{-1}.

The only direct measurements of brine fluxes from sea ice are those of Wakatsuchi (1977) and Wakatsuchi and Ono (1983). Wakatsuchi (1977) used a schlieren optical system and estimated volume fluxes by analysis of motion pictures. This work showed a slight increase in volume flux with growth rate and reported volume fluxes ranging between $3.18 \pm 0.67 \times 10^{-5}$ and $3.74 \pm 0.38 \times 10^{-5}$ cc cm^{-2} s^{-1}. Wakatsuchi and Ono (1983) collected brine streamers excluded by growing sea ice using a funnel attached to a rubber balloon. They showed an increase in the salinity and a decrease in the volume flux as the growth rate of sea ice decreased. A salt and mass balance model and the measured volume and salinity of brine excluded by growing sea ice yielded a volume flux ranging between 6.3×10^{-6} and 3.4×10^{-5} cc cm^{-2} s^{-1}, agreeing reasonably with the fluxes calculated from laboratory velocities and field brine tube densities. Table 1 summarizes the brine channel flux estimates and measurements discussed above.

Skeletal Layer and Brine Tubes. The results from temperature measurements and dye studies show that convective motion occurs in the skeletal layer. Fluxes into the skeletal layer may be estimated using ice algae to inte-

Table 1. Sea ice brine channel volume fluxes

Study	Method	Remarks	Flux (cc cm^{-2} s^{-1})
Lake and Lewis (1970)	Brine channel densities (field) (1 per 180 cm^2)		2.5×10^{-8} to 1.05×10^{-6}
			3.2×10^{-7} to 1.33×10^{-5}
Eide and Martin (1975)	Oscillatory flow (laboratory)	1.4 mm diam. channels	
		5 mm diam. channels	
Niedrauer and Martin (1979)	Channel velocities (laboratory) (0.25 – 0.006 mm s^{-1} in 1.4 mm diam. channels)	5 mm diam. channels w/r^2 correction	4.06×10^{-6} to 1.7×10^{-4}
Wakatsuchi (1977)	Schlieren movie analysis		$3.18 \pm .67 \times 10^{-5}$ to $3.74 \pm .0.38 \times 10^{-6}$
Wakatsuchi and Ono (1983)	Collected brine streamers, mass and salt budget model		6.3×10^{-6} to 3.4×10^{-5}

grate the effects of diffusion and convection. The input flux of seawater to the skeletal layer may be estimated using measured rates of nutrient uptake by ice algae if the following conditions hold: (1) the sea ice must be growing and convection must be occurring, (2) the adjacent seawater nutrient concentrations are invariant with depth and time and (3) the nutrients are completely removed by the ice algae as seawater moves up into the skeletal layer. This approach estimates the rate at which nutrients must enter growing sea ice to sustain the observed ice algae growth rates and does not distinguish between diffusion and convection.

The ice algal community, which consists predominantly of motile diatoms, is confined to the skeletal layer of sea ice. These algae are largely protected from grazing and grow to large standing stocks. The question of their nutrient supply was addressed by Meguro et al. (1967), who considered three possibilities: (a) bacterial conversion of organic compounds in the ice, (b) penetration of seawater from under the ice, and (c) supply of nutrients during spring melt. These authors calculated that sufficient nitrogen could be supplied by brine drainage and favored the last process.

Clasby et al. (1973) developed an *in situ* technique for measuring primary productivity in sea ice and in later work (Clasby et al. 1976) reported NH$_4^+$ and NO$_3^-$ uptake rates determined using ^{15}N-labeled tracers. Since they avoid cell lysis that accompanies salinity decreases during melting, these techniques provide the most reliable rate data available for sea ice algae. Results from the nitrogen tracer studies agree with Redfield-ratio corrected ^{14}C primary productivity measurements conducted at the same time. Although the ^{15}N-labeled tracers were added in concentrations above ambient levels, no stimulation was apparent in experiments performed using varying tracer concentrations.

Alexander et al. (1974) reported air temperatures of −15 °C and a porous ice-water interface with an intact ice algal zone, so the ice in their study was growing and the first condition appears to be met. Nutrient data reported by Alexander et al. (1974) and included in Table 2 indicate that variations in NH$_4^+$ and NO$_3^-$ concentrations with depth and time in the adjacent seawater were small, so the second condition is met. The third condition is most difficult to assess, as little is known about the growth kinetics of these algae. The fact that the algae experienced no stimulation at varying tracer concentrations suggests that they are growing at saturation. R. J. Barsdate (personal communication) observed that ice algae can migrate in a matter of hours and re-establish themselves at the interface in ice cores whose bottom portions were inverted. This observation also supports the notion that the ice algae have the capacity to effectively remove nutrients.

Table 2, using data from Clasby et al. (1976) (Table 20.1), presents seawater input fluxes to the skeletal layer calculated from measured ^{15}NH$_4^+$ and ^{15}NO$_3^-$ uptake rates and adjacent seawater concentrations. The fluxes range between 2 and 70 $\times$ 10^{-5} cc cm^{-2} s^{-1}. If nutrient uptake is less efficient than the 100% assumed here, the actual fluxes may be higher. If regenerated nutrients are important in maintenance of the algal population, this method overestimates the actual fluxes.

Discussion

Fluxes across the ice-water interface to the skeletal layer, estimated using nutrient uptake rates by ice algae, appear to be quite important to mass transport in the skeletal layer. Considering a skeletal layer porosity or liquid water content of 15% and a thickness of 3 cm, water residence time of 0.2 to 6 h result from the fluxes presented in Table 2. Since convective flushing in the skeletal layer provides a relatively continuous supply of nutrients to the ice algae, it is probably more important to their maintenance than the brine drainage mechanism favored by Meguro et al. (1967). Convective flushing in brine channels accounts for a minor portion of the heat flux in sea ice, though. Niedrauer and Martin (1979) calculate that the heat flux due to cyclic convection in brine channels is only 1% of the conductive heat flux.

Table 2. Seawater flux into sea ice skeletal layer

Experiment	Date	Uptake rate (μmole m^{-2} hr^{-1})	Seawater concentration (μmole l^{-1})			Seawater flux (cc cm^{-2} s^{-1})
			0 m	1 m	3 m	
^{15}NH$_4^+$	1 June 1972	9.29 $\pm$ 1.42	–	1.81[a]	–	14.2 $\times$ 10^{-5}
^{15}NH$_4^+$	18 April 1972	6.90 $\pm$ 1.49	3.1	2.9	2.9	6.19 – 6.61 $\times$ 10^{-5}
^{15}NH$_4^+$	16 May 1973	25.00 $\pm$ 3.03	2.0	1.1	1.0	34.7 – 69.4 $\times$ 10^{-5}
^{15}NO$_3^-$	18 April 1973	5.24 $\pm$ 1.80	6.5	6.4	6.6	2.25 – 2.19 $\times$ 10^{-5}
^{15}NO$_3^-$	23 May 1973	21.43 $\pm$ 0	8.1	7.0	4.4	7.36 – 13.5 $\times$ 10^{-5}

[a] Sampled 29 May 1972

Table 3. Sediment irrigation fluxes

Study/location	Method	Irrigation flux	
		(reported)	(cc cm^{-2} s^{-1})
Hammond and Fuller (1979) San Francisco Bay	^{222}Rn deficiencies	3 $\times$ 10^{-7} m s^{-1}	3.0 $\times$ 10^{-5}
McCaffrey et al. (1980) Narragansett Bay	^{22}Na added to water overlying box core	0.7 $\pm$ .4 cc cm^{-2} d^{-1}	.81 $\times$ 10^{-5}
Smethie et al. (1981) Washington shelf Columbia R.	^{222}Rn deficiencies	0.2 – 4.45 cm d^{-1}	.23 – 5.15 $\times$ 10^{-5}

These results may be placed in perspective by comparing them with a similar flushing process that occurs at the sediment-water interface. Burrowing organisms are known to enhance benthic fluxes by two processes: bioturbation, the mixing of sediment particles, and irrigation, the pumping of water into burrows in the sediment. Irrigation rates of individual organisms have been reported (see Aller 1977, Tables 1 and 2, pp 24–25), and volume fluxes can be obtained by using population densities and irrigation rates for individual species. These calculations, however, neglect patchiness and produce very large volume fluxes (Smethie et al. 1981). Several recent studies have taken advantage of the natural tracer ^{222}Rn (Hammond and Fuller 1979; Smethie et al. 1981) or added ^{22}Na (McCaffrey et al. 1980) to obtain estimates of irrigation fluxes in marine sediments. The results of these studies, which represent lower limits, integrate the volume flux of water across the sediment-water interface over several days and are summarized in Table 3.

Comparison of Tables 1 and 2 with Table 3 shows that volume fluxes across the sea ice interface are of about the same magnitude as those across the sediment-water interface. Although the fluxes are similar at these two interfaces, the processes driving the motion are different. The sea ice interface is driven entirely by physical processes, while the activities of organisms drive flushing at the sediment-water interface. Flushing at the ice-water interface should be active any time the sea ice is growing.

It is important to ice algae from the point when light increases to levels that can support photosynthesis until the ice melts. The distribution of snow cover on sea ice appears to control ice algae distributions (Alexander and Chapman 1981). Biological irrigation in sediments also varies in time and space and seems to be most active in late summer (Martens and Berner 1977; Elderfield et al. 1981).

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