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FRAZIL ICE IN RIVERS AND OCEANS

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INTRODUCTION

Frazil ice is the terminology for small discs of ice measuring 1–4 mm in diameter and 1–100 μm in thickness that form in turbulent, slightly supercooled water (Kivisild 1970). Because of the surface properties of these discs, once they form, they rapidly aggregate together and adhere to foreign material in the water. In rivers these crystals, which form at rapids and other areas of open water where their production rate can be as large as 10^6 m^3 day^{-1}, cause serious problems with hydroelectric facilities. These problems include the reduction of available head by 25%, the blocking of turbine intakes, the blockage of hydroelectric reservoirs, and the freezing open of gates.

Because of the economic importance of these problems, river frazil ice is the subject of many papers. Recent reviews of river frazil ice include Michel's (1971) comprehensive survey, Osterkamp (1978), and Ashton (1978). Also, the *Proceedings on Ice Problems of the International Association for Hydraulic Research* (Int. Assoc. Hydraul. Res. 1970, 1975, 1978) contain many excellent papers on frazil ice. Finally, Hobbs (1974) gives a comprehensive review of the literature on the physical properties of ice, much of which is relevant to the frazil crystal.

In the ocean, frazil ice forms during winter both at the ice edge and in the large regions of open water within the pack ice called polynyas, in particular, in near-shore regions where the predominant winds blow the ice away from shore so that a large expanse of seawater at its freezing point is exposed to the cold air. Ice production proceeds rapidly in these polynyas and is accompanied by an outflow of saline water into the ocean. Because ocean frazil ice has become a source of economic concern only in the last decade with the onset of oil and gas development in the Arctic, the literature on ocean frazil ice is sparse and the work on frazil ice has been almost entirely observational. Previous to the past decade, Zubov (1943) provides the only historical review.

The present review article primarily concentrates on the work of the last fifteen years. Specifically, the next or second section discusses the nucleation and multiplication of the basic frazil disc, and the surface properties of the discs that cause them to sinter together into clusters. The third section then discusses the generation of frazil ice in rivers, the flow of a suspension of the discs, the increase in size of the suspended material through sintering, and the subsequent deposition of the frazil ice under a solid cover. The fourth section is a parallel discussion for ice in the ocean.

THE CRYSTAL PROPERTIES

Supercooling and the Crystal Shape

In rivers frazil-ice production is a transient phenomenon that occurs in turbulent supercooled water. According to Carstens (1966) frazil ice grows in open water in the following way: first, because of wind and

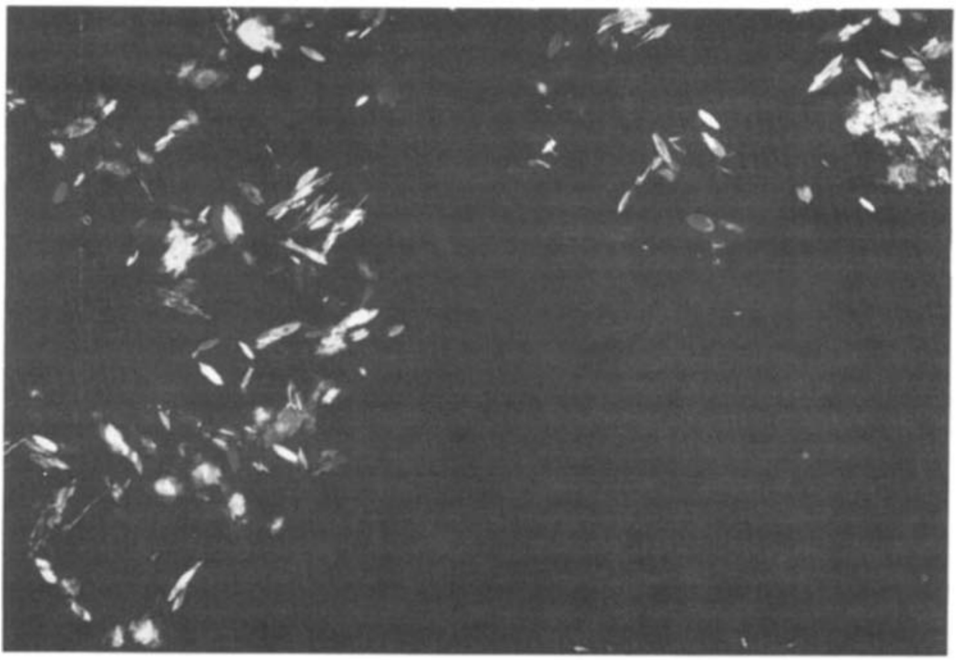

Figure 1 Photograph through crossed polaroids of a suspended solution of frazil-ice crystals; the photograph covers 25 mm in the vertical (from Martin & Kauffman 1981).

radiative cooling, the water temperature drops about 0.01–0.1 degrees below the freezing point; then the water abruptly fills with numerous frazil discs; finally, the water temperature following the mass of crystals increases on the order of minutes toward the freezing point.

Numerous other field observations of the formation of river frazil ice such as those of Michel (1971), Arden & Wigle (1973), and Osterkamp (1978) all report that frazil formation begins for supercoolings on the order of 0.1 degrees with the preferred crystal shape being a 1–5 mm diameter circular disc. As an example, Figure 1 shows a photograph taken by Martin & Kauffman (1981) of frazil discs of various orientations suspended in a 10 mm thick water layer, where the basic disc has a 1 mm diameter and a 1–10 μm thickness.

In his review of the preferred shapes of the ice crystals that form in supercooled water, Hobbs (1974, pp. 572–77) confirms the above observations. He shows that in fresh water that is supercooled by 0.1 to 0.3 degrees, then is nucleated by the introduction of either a very cold metal rod or snow crystals, the preferred growth habit of the resultant crystals is the circular disc. Lesser amounts of needles and semicircular discs also form. Arakawa (1955, also shown in Hobbs 1974, Figure 9.1) shows that the disc develops into a six-pointed stellar dendrite as it grows in radius. This dendritic form also affects the sintering process discussed below.

For a specific example of frazil ice formation, Wigle (1970) and Arden & Wigle (1973) describe a study in the Niagara River, which is 29 km long, has a mean depth of 4 m, and lacks a seasonal ice cover. Here frazil production on cold, clear nights causes an abrupt reduction in the river flow that can be as large as 25 percent. Through use of instrumentation such as underwater lights, collection trays, and thermistor arrays, the authors observed on cold, clear nights that frazil formation began at supercoolings of 0.05 degrees. Underwater observations showed that the frazil ice resembled "a driving snowstorm, as seen through the headlight beams of a moving automobile at night" (Arden & Wigle 1973, p. 1299). As time progressed, the entire river depth became supercooled and the frazil discs began to adhere to the rocks, gravel, and sediment of the river bottom, as well as to each other. When the crystals adhered to sands and sediments, they picked up the adhered material and moved with the river. When the crystals adhered to rocks and gravel, the crystals remained on the bottom then grew in size both from the adherence of additional crystals and from heat transfer to the supercooled water to form what is called "anchor ice." If the anchor ice grew large enough, the rocks and gravel also rose off the bottom and moved with the flow. In the morning following a period of frazil growth, the authors found that ice floes, discolored by entrained material, covered the river surface. The authors feel that the combination of the anchor ice growth, the suspended frazil ice, and ice floating on the surface caused the abrupt reduction in the river flow rate.

Nucleation and Multiplication of Frazil Ice

The way in which frazil nucleates and multiplies in turbulent water has been a source of controversy; namely, do the ice crystals form spontaneously in the water column, or can new crystals only grow from ice already present in the water? Hobbs (1974) shows that for liquid water to freeze or to overcome the free-energy barrier associated with a change of phase to ice, the water, even if contaminated with clays or biological material, requires a supercooling on the order of 10 degrees. Since the observed supercooling in rivers is 0.01 to 0.1 degrees, homogeneous or spontaneous nucleation probably rarely occurs in nature. Therefore, the formation of frazil crystals must proceed from heterogeneous nucleation, or from the introduction of an ice crystal from outside the water.

Osterkamp (1977) reviews various natural seeding mechanisms for the initiation of frazil ice. These include crystals introduced into rivers as snow or frost crystals, water droplets on a μm scale splashing into the air then freezing and re-entering the water, and growth from natural ice at the riverbanks. As evidence of natural seeding, Osterkamp reports finding numerous ice crystals in air samples taken at night above arctic streams.

Once an ice crystal is introduced to the supercooled water, two experiments on supercooling and turbulence show that there is a rapid multiplication of ice nucleii through what is called "collision breeding." Garabedian & Strickland-Constable (1974), working with 0.5 liters of distilled and de-ionized water, which was supercooled in a cooling bath and stirred at 800 rpm, find that for supercooling of 0.1 degrees 44 crystals formed from the original unbroken seed in five seconds. Müller & Calkins (1978) find similar results for an oscillating grid in a one-liter tank, and use their data to estimate heat transfer to the nucleii. They also show that for experiments beginning with the same initial supercoolings the rate of frazil-ice production increases with the turbulence levels. In both cases, because the multiplication process is microscopic, the actual mechanism is unclear, but apparently the collision of the crystal with both the walls of the tank and other nucleii leads to the rapid crystal multiplication.

Garabedian & Strickland-Constable (1974) also found that for a supercooling of 0.5 degrees, a stirring rate of 900 rpm, and the addition of two 3-mm glass beads to the tank, no nucleation occurred even after stirring for one hour. Müller & Calkins (1978) also observed in their experiment that even for supercoolings of one degree, spontaneous nucleation never occurred; crystal growth took place only when an ice crystal was added to the agitated solution. The conclusion of these authors and Hanley (1978) is that the supercooled water must be seeded for frazil production to begin. In summary, the laboratory studies suggest that for both rivers and oceans, frazil-ice production begins with the addition to turbulent supercooled water of outside ice crystals which rapidly multiply into frazil ice through collision breeding.

Sintering

Unfortunately, once the disc-shaped crystals appear in the supercooled water, they do not remain as separate crystals. Rather they rapidly sinter together into larger groups on the scale of 3–10 mm across called "flocs." Osterkamp (1978, Figure *1b*) shows from field observations a photograph of these "flocs," and the crystal clusters in Figure 1 show a laboratory example. Once the flocs form, they too sinter together into larger chunks so that the size of the suspended material in the water increases with time. Thus, sintering causes both a rapid increase in the size of the suspended material and, as Martin & Kauffman (1981) show for the ocean case at least, a viscosity that increases as the rate of shear decreases for water-frazil suspensions.

To understand the physics of sintering, Hobbs (1974, Chapter 6) both reviews earlier work and summarizes current work on the surface properties of ice. This work shows that ice crystals, both alone and in contact with each other, adjust their shapes such that their surface free energy tends toward a minimum. For example, experiments show that the point of contact between two ice spheres is thermodynamically unstable. Because a chemical-potential gradient exists between the point of contact and the unstressed ice surface, there is a transfer of material to the contact point so that a neck forms between the two spheres. Extrapolation of the data in Hobbs' Figure *6b* shows, for two spheres in air with radii of curvature of 10 μm, the time for the neck to grow to 1/4 of the sphere diameter is 6 s; for 1 μm the time is 10^{-2} s, even though Hobbs states that plastic flow of the ice will affect the results for the 1 μm spheres. This work implies that for discs with radii of order 1 mm and thicknesses of 1–10 μm, in contact in the manner shown in Figure 2, bonds between the discs form quickly. Further, because of the

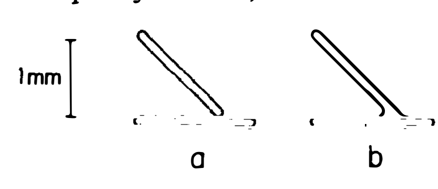

Figure 2 The sintering together of two frazil discs: (*a*) at the moment of contact; (*b*) a short time later. See text for additional description.

radius-of-curvature decrease at the edge, these sintering times may be even shorter for discs that have evolved into stellar dendrites.

On a longer time scale, the frazil disc is by itself unstable. To minimize the surface free energy, the crystal shape over a period of days tends toward a sphere. Hobbs (1974, Chapter 6.10.1) discusses the analogous evolution of snow crystals, and Michel (1966) shows photographs of evolved frazil ice. This process, which tends to strengthen snow, may also strengthen deposits of frazil ice.

RIVER FRAZIL ICE

River frazil ice occurs in two kinds of rivers. First, it occurs in those that remain open throughout the year such as the Niagara (Arden & Wigle 1973) and the rivers of southern Sweden (Larsen & Billfalk 1978). Second, in very cold climates frazil ice also forms in rivers where only sections with high velocities and strong rapids remain open, such as the Thjorsa river system in Iceland which feeds the Burfell power project (discussed in Rist 1970, Carstens 1968, and Carstens 1970b) and the La Grande River in Canada (reviewed in Michel 1978). For hydroelectric plants on the Swedish rivers Larsen & Billfalk (1978, page 235) observe that "the most common cause of problems was the occurrence of frazil," which clogged intake trash racks with ice and ice-carried rocks and boulders, one of which weighed 30 kg, and caused in one instance a gate to freeze open, resulting in a runaway turbine.

For both kinds of rivers, hydrologists concerned about power production use a simple classification of frazil ice to predict when it will be hazardous, and an empirical condition to predict as a function of river velocity when frazil ice will form. For the first, Devik (cited in Carstens 1966) divides frazil ice into "active" and "passive" ice. Active frazil, which occurs in supercooled water, adheres to almost all foreign material and causes the serious problems in hydroelectric plants. Passive frazil occurs when the surrounding water warms up to the freezing point; passive frazil is not sticky and, as Carstens (1968) shows, is handled hydraulically as an inverse sediment problem by the proper design of spillways and settling basins.

For the second, Carstens (1970a) gives the empirical condition used to predict frazil formation in Norwegian rivers. For cold air temperatures and a river surface velocity V, the condition is as follows: For $1.2 > V > 0.6$ m s^{-1}, frazil ice forms and accumulates on the surface; for $V > 1.2$ m s^{-1}, the frazil forms and is suspended in the water column. Carstens further adds that this criterion applies over depth variations of a factor of 5, so that it cannot be scaled as a Froude-number criterion. This criterion is in rough agreement with the laboratory experiments of Hanley & Michel (1977), who set 0.24 m s^{-1} as the threshold velocity for frazil formation.

Evolution of Frazil Ice

For both active and passive conditions, sintering causes the frazil discs to stick to each other. This leads to a rapid growth in the size of the suspended ice. Michel (1966, 1971) and Osterkamp (1978) use the following classification with photographic illustrations for the evolution of river frazil ice: 1. the previously described discs; 2. "flocs," which are collections of discs with scales of 5–100 mm; 3. "pans," which accumulate on the surface with diameters of order 1 m and thicknesses of 0.1–0.5 m; and 4. floes, with diameters of 1–30 m and thicknesses of 0.5–5 m. As Carstens (1968) observes, this rapid growth in the size of the suspended particles greatly complicates any modeling of frazil ice.

In an open river the ice is swept downriver where it accumulates against a boom or solid ice cover. As Michel (1966) states, this accumulation can lead to a 10–40 km-per-day growth of the solid ice sheet upstream. This growth continues until the ice sheet reaches the foot of a high-velocity rapid. As reviewed in Michel (1971), rapids serve as "ice factories." For the Russian Angara River described in Michel (1971, p. 66), ice-production rates during winter in regions of open water are of order $1-100\times10^6$ m^3 day^{-1}. The currents at the foot of the rapids carry this enormous mass of ice under the solid ice cover, where it accumulates in large, irregularly shaped billows. Dean (1977) found from observations in the St. Lawrence River that these billows are a mixture of frazil ice and larger pieces of solid ice. Figures 3 and 4 show examples of these accumulations; Figure 3 shows the frazil distribution in the intake pond of the Gamlebrofoss power plant in Kongsberg, Norway (Tesaker 1975), and Figure 4 is from the La Grande River in northern Quebec (Michel 1978).

The Gamlebrofoss observations in Figure 3 show the frazil-ice deposition pattern in both plan and cross-sectional views for 13 January 1972, where the frazil ice forms in the rapids upstream of the intake pond. Tesaker estimates that the volume of the under-ice deposition is 4×10^4 m^3, or about 20% of the volume of the pond, and, from the two cross-sectional profiles, that the frazil ice blocked about 60% of the normally ice-free area. Tesaker also observed for both 1971 and 1972 that the frazil ice appeared to be deposited in an inverted sediment fan. Second, the La Grande River observations from the winter of 1972–1973 by Michel & Drouin (cited in Michel 1978) show a 16-km-long under-ice

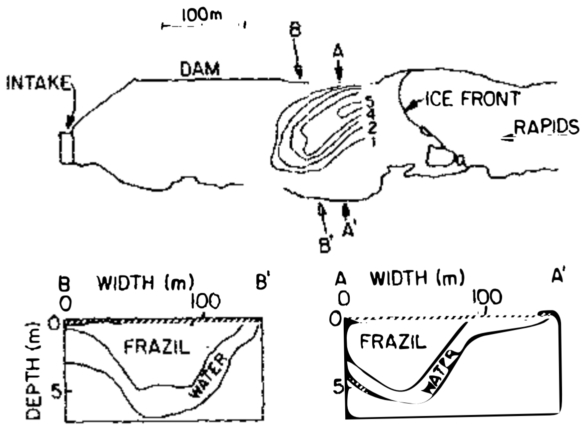

Figure 3 Plan and cross-sectional views of the frazil-ice distribution in the Gamlebrofoss intake pond on 4 February 1971. Contours on plan view show frazil depth in m; the lines *AA'* and *BB'* indicated on the plan view show location of cross sections (redrawn from Tesaker 1975).

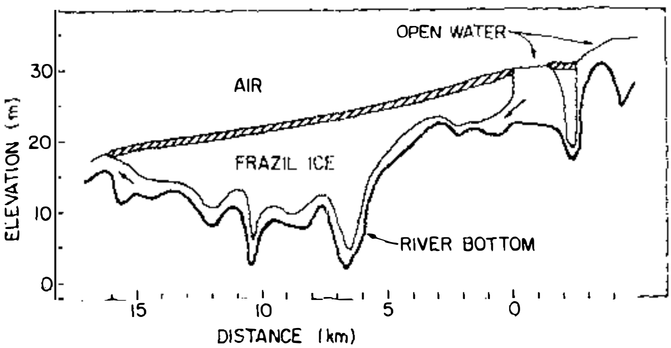

Figure 4 Frazil-ice distribution in the La Grande River during winter 1972–1973 (redrawn from Michel 1978).

frazil deposition with an approximate volume of 5×10^7 m^3. This deposition began with the formation of a solid ice cover away from the rapids, followed by the accumulation of the frazil ice beneath the solid cover, beginning downstream and progressing upstream. As Figure 3 shows, the frazil ice filled most of the river depth. Michel (1978) further states, although not from the La Grande River observations, that frazil accumulations in low-velocity zones gradually increase the head loss so that in some cases the frazil-producing rapids are flooded out, a solid ice cover forms over the rapids, and the process advances upstream.

When the frazil ice generated in a rapid encounters a solid ice sheet, many investigators (as reviewed in Michel 1978 and Ashton 1978) use a Froude-number criterion to define the conditions under which this ice will flow beneath the solid cover. This criterion depends on the ice particle size and porosity, as well as water depth and velocity, and thickness and porosity of the ice edge. Michel (1978) and Tesaker (1975) have also had some success with the use of a Froude-number condition to predict the maximum thickness of the ice deposition beneath the cover. Finally, Michel (1978) and Tesaker (1975) briefly describe how the combination of solar radiation and increased water flow in the spring leads to the erosion of the deposited billows; they also point out that further observational and theoretical work is needed on frazil-ice decay.

A Model Experiment

Carstens (1968) carried out the only model experiment on frazil-ice flow that is available in the literature. In his study, he built a hydraulic model of the Burfell, Iceland, power project, in which he assumed frazil ice could be treated as a suspension of buoyant sediments. He modeled the frazil behavior in spillways and inverse settling ponds through use of polyethylene shavings with a specific gravity of 0.92, even though, as he states, these shavings do not sinter together. Later, Carstens (1970b) briefly compares the flow in his model studies with that in the completed Burfell project. His photographs of the frazil-ice flow in the actual project show the existence of abrupt transition lines from fluid to solid behavior which did not occur in his model. In their experiments on wave absorption by ocean frazil ice discussed in the next section, Martin & Kauffman (1981) observed similar transitions, which were caused by the sintering-induced nonlinear viscosity of frazil ice. This nonlinear viscosity may also be the cause of the transition lines in the river-ice case, which suggests that a viscous model of frazil-ice flow will be of value to hydrologists.

Heat-Transfer Models

For the calculation of frazil-ice production in rivers, Freysteinsson (1970) used the Rymsha-Donchenko model as described and elaborated on by Dingman et al (1968). Dingman et al (1968) use this model for calculations of the length of ice-free water produced in a river by the discharge from a thermal power plant, and obtain a fair agreement between the model and observation. Freysteinsson (1970) finds that this model gives good agreement with heat-flux measurements on the Thjorsa River System. Larsen (1978), however, carries out a term-by-term critical review of the different models that exist for radiative, evaporative, and turbulent atmospheric heat transfer. From observations made in rapids, which suggest that the frazil production rate is several times greater than predicted, he concludes that the effect of turbulent water flow must also be included in these heat-transfer models. In summary, the field appears to lack a satisfactory theoretical model for the prediction of frazil ice production in a turbulent water flow.

OCEAN FRAZIL ICE

Where It Occurs

Historically, the terminology for ice-disc formation is slightly different for the ocean than for rivers. For example, Armstrong et al (1966) classify oceanic ice discs into "frazil ice," or a light suspension of the individual platelets, and "grease ice," which is the official World Meteorological Organization terminology for a dense slurry of ice platelets on the ocean surface. The term grease ice is old whaling terminology which refers to the greasy appearance that the slurry gives the surface by damping the capillary waves. In the polar ocean frazil- or grease-ice formation takes place in at least four different situations: in regions of open water called leads and polynyas; at the interface between two fluid layers, each at their freezing point and with different salinities; adjacent to ice shelves and icebergs; and from the drainage of cold dense brine from sea ice into the underlying water.

LEADS AND POLYNYAS Frazil ice forms when cold winds blow across regions of open water that is at its freezing point. Here the cold air and wind waves combine to cool and agitate the surface water such that frazil ice forms. Dunbar & Weeks (1975), Martin et al (1978), and Martin & Kauffman (1981) review frazil-ice formation in regions of open water, where regions with width scales less than 100 m are called "leads" and larger regions are called "polynyas." In leads, the above observations show that the frazil crystals form throughout the open water, then a combination of wind and waves herds the crystals downwind to pile them up to depths of 0.1–0.3 m at the end of the lead. As this process continues, the ice cover advances laterally in the upwind direction until the open-water area is reduced to the point that frazil formation ceases. In large polynyas, a Langmuir circulation herds the frazil ice into streaks parallel to the wind and piles it up downwind to depths of order 1 m. Figure 5 shows a photograph of these Langmuir streaks in a large polynya south of Nome, Alaska, on 5 March 1978. At this time, the air temperature was $-20°C$ and the wind velocity was 15 m s^{-1}, parallel to the streaks, and blowing toward the large floes. The predominant wavelength in the photograph is 6 m; the floes are about 100 m across. The photograph clearly shows the Langmuir streaks and the piling up of the grease ice to measured depths of order 1 m against the edges of the large floes. Again, the observations suggest that the

Figure 5 Oblique aerial photograph from 150 m of grease-ice formation off Nome on 5 March 1978; see text for further description (from Martin & Kauffman 1981).

frazil-ice crystals form in the open water, then are herded both into the Langmuir streaks and downwind.

BETWEEN WATERS OF DIFFERENT SALINITIES A second well-documented frazil formation mechanism occurs in the polar ocean when a layer of fresh melt water at its freezing point lies over a layer of seawater at its freezing point. This occurs in the mouths of arctic rivers (McClimans et al 1978), and in the polar summer when fresh melt water accumulates under the pack ice to form inverted melt ponds (Martin & Kauffman 1974). In both cases the temperature difference between the 0°C fresh water and the -1.6°C seawater leads to the production of both supercooled water and frazil ice. For the second case, Martin & Kauffman (1974) show that a 0.2–0.3-m-thick fresh-water ice layer grows during the polar summer under the melting pack ice. Lyons et al (1971) show that this process also contributes to the ice growth *from the bottom* of about 10^2 km of the Ward Hunt Ice Shelf, which serves as a partial dam across Disraeli Fjord. In this case, fresh glacial melt water accumulates behind the dam, then flows out beneath it over the colder arctic seawater. The heat transfer between the two layers leads to the production of fresh-water ice. As McClimans et al (1978) state, the process is understood for the case of no relative velocities between the two layers, while the nature of this process in a velocity shear remains to be investigated.

ICE SHELVES AND ICEBERGS Another source of oceanic frazil ice occurs in the seawater adjacent to ice shelves and icebergs. Here, frazil ice may form in two ways: first, by direct cooling of the seawater from the cold ice; second, as Foldvik & Kvinge (1974, 1977) show both theoretically and observationally, by the upward movement of a seawater parcel that is at its freezing point. Because of the freezing-point depression with depth, as the parcel rises it becomes supercooled; if nucleation occurs in the rising water then under certain conditions the ice-water buoyancy increases, so that the parcel continues rising. To estimate the amount of ice generated by this process, Robin (1979) shows from Foldvik & Kvinge's data that if a 10-m-thick water layer at its freezing point moves upward at 10 mm s^{-1} over a 100 m elevation change, 10^3 m^3 yr^{-1} of ice is generated per unit-meter width of ice shelf. These calculations suggest that ice shelves contribute to the frazil ice observed around Antarctica.

BRINE DRAINAGE The possibility also exists that the frazil ice observed beneath a solid ice cover forms locally. Lewis & Lake (1971) and Lewis & Milne (1977) suggest that the slow drainage of cold dense brine from the sea ice, which occurs over its growing season, leads to growth of the frazil-ice platelet layers. Observations summarized in Dayton & Martin (1971) and Lewis & Milne (1977) show that this brine leads to the growth under a solid ice sheet of ice stalactites which form around the brine outflow; Martin's (1974) laboratory study of stalactites shows that as much as half of the ice represented by the cold brine flows into the underlying water as frazil crystals, instead of going into stalactite growth. The sparse number of stalactites observed in the field, however, suggests that brine drainage cannot account for the large mass of observed frazil ice.

Underwater Observations

The underwater observations in the Arctic and Antarctic are very similar to the river observations. Dayton et al (1969) observed in McMurdo Sound large billows of frazil platelets measuring 1–4 m in thickness under the solid ice cover (Figure 6). They also observed that on some days in winter the water filled with small ice spicules, and that anchor ice formed down to depths of 33 m both on the sea bottom and on lines hanging in the water. Both the ice spicules and the anchor ice

Figure 6 Frazil-ice accumulation under the pack ice of McMurdo Sound (photograph courtesy Paul Dayton).

appeared suddenly, after the passage of days with no frazil formation. The formation of this frazil layer may be caused by the nearby McMurdo Ice Shelf.

For the sea ice adjacent to the coast at the Russian Antarctic station Mirny, there are also many reports of frazil-ice formation both on the surface in the large polynyas that form behind islands and grounded icebergs by the katabatic winds, and under the ice in large billows. Morecki (1965) describes observations made in 1958 of under-ice frazil billows measuring 1–5 m thick, and the growth of anchor ice down to depths of 5–50 m on lines hanging in the water column under a solid ice cover on different days throughout the winter. He also reports supercooling of the water column on the order of 0.1 degrees and attributes the frazil growth to supercooled water generated at the nearby ice shelves.

Baranov et al (1968) describe similar observations made in 1963. They attribute the frazil growth to the heat sinks provided by the nearby ice shelves and grounded icebergs, and by the large offshore polynyas which opened and closed throughout the winter in response to the katabatic winds. Finally, Cherepanov & Kozlovskii (1973) review previous observations and attribute the growth of the under-ice billows to ice formed in the persistent offshore polynyas which is redistributed by the ocean currents. They state that the under-ice dispersal of frazil depends on local currents, and in regions such as the Molodeshnaya Road where the current velocity is zero, frazil-ice billows are not observed.

In the Arctic, there have also been several recent diving surveys under the ice offshore of Prudhoe Bay, Alaska, in Stefansson Sound. From such surveys, K. Dunton and E. Reimnitz (unpublished, 1980) observed large frazil-ice billows under the solid ice cover at three weeks after freeze-up in both 1978 and 1979. They also observed that the individual frazil crystals in the billows were covered with a fine sediment layer. Because both 1978 and 1979 freeze-up at Prudhoe Bay took place during severe storms, they attribute the billows to frazil ice that formed during the storms, after which the billows froze into place.

All of the frazil ice observations described above can be explained by heat losses to either adjacent polynyas or ice shelves, or by the formation of the billows during freeze-up. Since the field observations (Dayton & Martin 1971, Lewis & Milne 1977) show that the ice volume in the observed stalactites is several orders of magnitude smaller than the ice in the frazil billows, it seems unlikely that the billows form locally from brine drainage. Therefore, the literature suggests that the winter formation of frazil ice in the ocean takes place in the same way as frazil formation in rivers. Namely, the ocean supercools on the order of 0.1 degree by cooling from either a polynya or possibly an ice shelf, then nucleation occurs from introduction of outside ice crystals. The frazil then forms and either is herded by the wind and currents into billows that freeze into place, or is carried under the ice cover by currents.

Viscous Properties

To study the viscosity of frazil-ice slurry, Martin & Kauffman (1981) carried out in a cold room a wave-tank study of wave absorption by grease ice. Figure 7 shows a schematic diagram of the waves propagating into the ice; the surface waves herd the grease ice to the far end of the tank where the radiation stress creates the thickness increase with distance of the grease-ice layer.

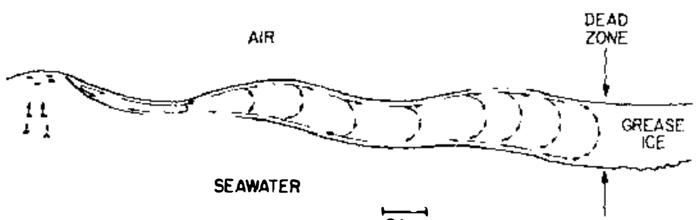

Figure 7 Schematic drawing of waves propagating into grease ice from the laboratory experiments of Martin & Kauffman (1981); see text for additional discussion.

The figure also shows that the grease ice divides into regions of liquid and solid behavior separated by an abrupt transition zone measuring 5–10 mm wide, which Martin & Kauffman term the "dead zone." Ahead of the dead zone, the waves propagate as heavily damped water waves, where the wave attenuation generates the mean velocities shown by the arrows within the grease ice; behind the dead zone, the waves propagate as elastic waves. Also ahead of the dead zone, when the grease-ice depth is greater than k^{-1} where k is the wavenumber of the incident wave, the wave amplitude decays linearly such that the decay slope α obeys

$$\alpha \approxeq \tfrac{1}{4}(a_0 k)^2, \tag{1}$$

where a_0 is the wave amplitude ahead of the grease ice. Related observations showed that the relative volume of the ice crystals varied from 18–22% at the leading edge to 32–44% at the dead zone, where the largest concentrations in the experiment occur at the dead zone for large values of a_0.

Martin & Kauffman explain the wave-decay observations in terms of a yield-stress-viscosity model,

$$\mu = \frac{b^2}{\gamma}, \tag{2}$$

where μ is the effective dynamic viscosity, γ is the scalar shear rate, and b^2 is the yield-stress coefficient. From their data,

$$b^2 = \tfrac{1}{4} kS, \tag{3}$$

where S is the radiation stress of the incident wave.

Physically, the cause of this shear-dependent viscosity, which increases as the shear rate decreases, is the combination of sintering and concentration changes. The sintering creates bonds between the crystals and the increase in concentration with radiation stress creates more bonds per unit volume. These bonds, which present resistance to the shear, create the nonlinear viscosity. A similar study by Uzuner & Kennedy (1974) on the response to stress of a field of ice chunks shows that bonds form between the chunks by both sintering and regelation. Their results give a qualitatively similar strength increase with increases in concentration and decreases in the shear rate.

From qualitative experiments with fresh-water frazil slurries, Martin & Kauffman also found that fresh-water crystals tended to form stronger bonds than crystals in salt water. The difference between the two cases is probably due to the absence of a salt-water coating on the fresh-water crystals. One effect of this difference is that the large ice flocs that form under water in fresh water do not appear in the salt-water experiments.

Thermal Properties

Because frazil ice forms in open water and is herded either downwind or into Langmuir streaks, the frazil growth rate is much greater than that of an undisturbed solid ice sheet. Although there are no systematic quantitative measurements of the growth rate, Martin & Kauffman (1981) found in the laboratory that for a large-amplitude wave field and an air temperature of $-20°C$ the equivalent thickness of 0.1 m of solid ice grew in one hour, while for an undisturbed ice cover where the thickness increased by conduction, a 0.1 m ice growth required 24 hours. In the wave field, ice growth took place both in the open water and in the grease ice ahead of the dead zone. In this part of the grease ice, the stirring induced by the wave oscillations and the mean circulation kept the temperature of the ice surface warm. For air temperatures of -20 to $-30°C$ with air blowing over the surface, the ice surface temperature ahead of the dead zone was only 10^{-1} to 10^{-2} degrees colder than the water temperature below the grease ice. Behind the dead zone, Martin et al (1978) show that the surface temperature decreased slowly with time, and grease ice solidified into the pancake ice. These observations ahead of the dead zone imply that the heat transfer in this region is of the same order as the open-water heat transfer.

Oceanic Importance

The above work implies as Martin & Kauffman (1981) show in the Bering Sea that polynyas where wind or currents periodically sweep away or break open the ice cover and replace it with frazil ice are important regions of ice production. In these polynyas there is the possibility of the growth of the equivalent thickness of 2 m of solid ice in 20 hours; whereas in the high Arctic, 2 m of undisturbed ice growth requires an entire season. Further, K. Aagaard (in preparation) shows that the rapid ice growth in the polynya that forms in the Chukchi Sea off Cape Lisburne and Point Hope, Alaska, is accompanied by an oceanic salt flux that generates a dense outflow along the sea bottom.

Frazil ice also plays a geological and biological role in the ocean. First, K. Dunton and E. Reimnitz (unpublished, 1980) observed in Stefansson Sound, Alaska, that a fine sediment layer coated each deposited frazil crystal. They speculate that frazil ice may serve as both a scouring and sediment-transfer agent. Second, Dayton et al (1969) describe the biological role of frazil ice in McMurdo Sound. They found that the frazil-ice billows serve to shelter biological colonies and that anchor ice both transports species from the sea bottom to the pack-ice bottom, and renders certain shallow-water regions uninhabitable. For example, they found that the sponge colonies do not exist above the maximum depth of anchor-ice formation. These diverse examples suggest that frazil ice is important in determination of the geology, biology, and physical oceanography of the shallow-water regions of the polar oceans.

SUMMARY

The present review discusses the problems of frazil-ice formation, flow, and deposition in rivers and oceans. In rivers, four research areas require attention. First, a model of frazil-ice production as a function of atmospheric and river-flow parameters needs to be constructed and confirmed in the laboratory and field. Second, the increase in the size and density of the suspended ice aggregates with time and the related viscosity of the suspension should be determined as a function of local shear and concentration. In particular, we need models of ice sintering and deposition under active and passive conditions, and related models for anchor-ice growth from both sintering of additional crystals to the deposited ice and from heat transfer to the surrounding supercooled water. Third, on a longer time scale we need models of the metamorphosis of single ice crystals and the effect of this shape change on the strength of the deposited frazil ice. Finally, we need additional studies on frazil-ice decay from flowing warm water and solar radiation.

Our needs in the ocean are similar. First, we need to clarify the mechanism of frazil-ice production; frazil ice is obviously produced in open water, while the literature suggests that frazil production also occurs at ice shelves and by brine drainage. The latter two mechanisms require field studies to determine their importance; the polynyas require a heat-transfer model (for calculation of the ice-production rate) as well as a related salt-flux model. Finally, the oceanic case also requires studies similar to those described for river ice on frazil-ice viscosity, deposition, and decay.

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