Frazil Ice Formation during the Spring Flood and its Role in Transport of Sediments to the Ice Cover

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Abstract - The article describes full-scale experimental studies performed in the Transdrift-IV expedition from fast ice in the near-delta of the Lena river in the Laptev Sea during the period immediately preceding the flood (late May) and during the peak of the flood (early and mid-June). Processing of data has revealed the presence of supercooled water layers 5 to 150 cm thick in the zone of river-sea water contact (in the upper part of the seasonal pycnocline). The supercooling value was observed to be -0.8 ^{0}C. Together with the thickness of supercooled fluid it depended upon both the time (before or during the flood) and the site of measurements (in the zone of main branches or beyond). At one of the stations a conglomerate of frazil ice was found attached to the cable of a bottom temperature meter at the depth of the pycnocline. Using the known conditions, the probability for supercooling and further frazil ice formation at all stations was determined. The results of observations have allowed the local Richardson numbers to be calculated for the river-sea water contact zone - the layer of supercooled fluid. Based on the theory of entrainment at the flat turbulent jet margin and a semi-empirical turbulence theory, it was possible to correctly relate the mean current velocity U in the upper freshened layer to the dynamic velocity U_* (root-mean-square velocity of turbulent variations) and present the entrainment at the river-sea water boundary as a kind of entrainment at the flat turbulent jet margin. Using ratios from laboratory studies of frazil ice formation, the actual rates of frazil ice formation in the river/sea water contact zone were estimated. They were calculated for the different mean motion velocities in the freshened layer during the different periods of the flood development in the near-mouth region of the Lena River. Based on the known concentrations of suspended sediments in the layer freshened by river water, the fluxes of suspended matter to the bottom ice surface, governed by the process of frazil ice formation, were estimated.

Introduction

Many large rivers that carry a vast amount of freshwater and sediments, especially during the spring floods, empty onto the Siberian Shelf Seas (and, in particular into the Laptev Sea). The temperature of fresh river water during the winter-spring season is close to its freezing point. Entering the ice-covered sea, freshwater spreads over the surface of colder, saline sea water. Nansen (1956) found that at this contact some portion of fresh river water becames supercooled and forms frazil ice, especially in the near-mouth sea regions. It should be noted that formation of frazil ice is a phenomenon typical of the entire Arctic Basin, including the summer season (Golovin et al., 1993; Golovin et al., 1996; Timokhov, 1989). Formation of frazil ice at the interface of fresh and saline water leads to the growth of essentially freshwater ice crystals (Golovin et al., 1996). As a result, ice with a chaotic crystals orientation ice formed. This allows us to identify the interlayers of this ice 2 cm to 2 m thick within the general structure and texture of multiyear drifting and fast ice both in the Arctic and Antarctic (Petrov, 1971; Cherepanov, 1972; Cherepanov and Kozlovsky, 1972; Weeks and Ackley, 1982; Martin, 1971).

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The main cause for supercooling of water freshened by river runoff and the subsequent formation of frazil ice is a more effective loss of heat to the underlying cold and saline sea water, than gain of salt via the countering salt flux. The different efficiencies of heat and salt exchanges are governed by the fact that the molecular temperature conductivity coefficient Kt_m exceeds the molecular salt diffusion coefficient Ks_m by two orders of magnitude ($Kt_m/Ks_m=10^2$). This allows us to call the mechanism of supercooling and frazil ice formation under consideration „double-diffusion" (Martin and Kauffman, 1974; Krylov and Zatsepin, 1992) or contact mechanism (Timokhov, 1989).

The article is devoted to the study of possibility of frazil ice formation at the river sea water interface during flood in the near-mouth region of the Lena River, the Laptev Sea. The rate of frazil ice formation in natural environments has been estimated on the basis of laboratory experimental evidence. This has allowed for estimating the intensity of riverine suspended matter supply towards the ice cover due to the double-diffusion mechanism of frazil ice formation.

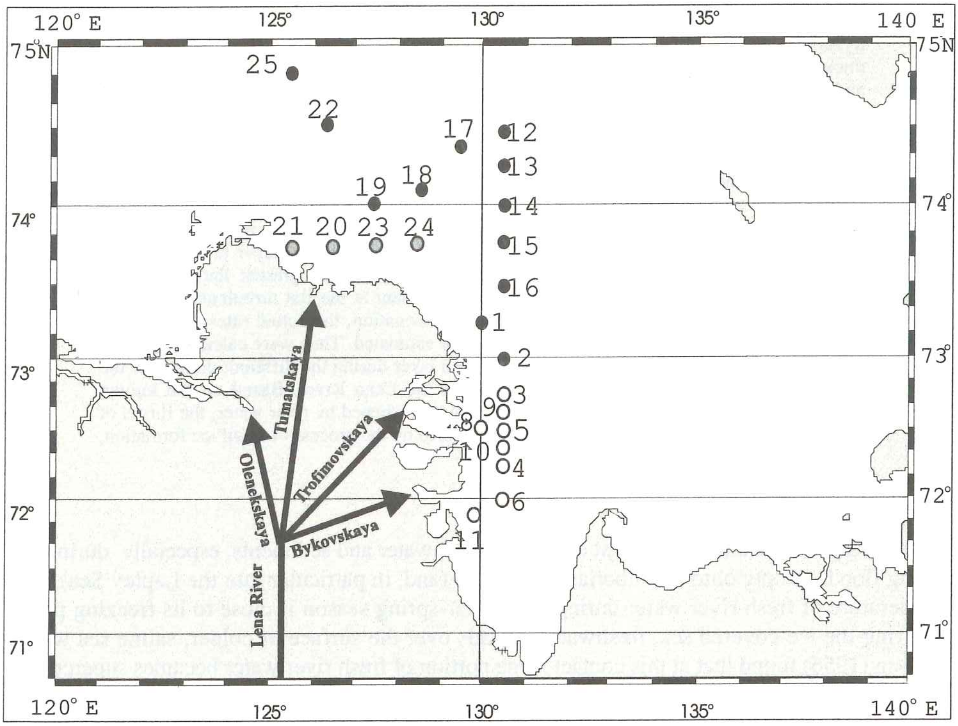

Figure 1: An array of comprehensive oceanographic stations in the Transdrift IV expedition (LN96..); ● - stations taken only during the period before the flood (for example LN9610);. ● - stations taken twice: before the flood and after its onset (for example LN9610, LN9610a); O - stations taken three times: before the flood and twice after its onset (for example LN9610, LN9610a, LN9610b).

Materials and methods

For our studies the materials of hydrological observations collected during the Russian-German expedition Transdrift IV were used. The work was carried out on fast and drifting ice in the

Laptev Sea in the near-delta region of the Lena river from mid-May to mid-June during the period that directly precedes the spring flood, and during the period of its development. The array of comprehensive oceanographic stations is presented in Figure 1. Observations of water temperature and salinity were conducted by means of a fine-structure CTD sonde. We used an OTS-PROBE Serie 3 sonde (Meerestechnik Electronic CmBH, Germany). The main technical characteristics of an OTS-PROBE are the following:

- water temperature: accuracy 0.01°C ; time costant 160 ms;
- electric conductivity: accuracy 0.02 mS/cm; time constant 100 ms;
- hydrostatic pressure: accuracy 0.1%, time constant 40 ms;
- information exchange rate 1200 bit/s.

Temperature and salinity sensors of the sonde are situated close to each other. They are of the same size. Multi-channel sonde is able to perform a parallel cross-examination of sensors. The rate of vertical sounding is about 0.5-1.0 m/s, i.e. about 12-24 parameter values per 1 meter of sea water body (average 15-17). All values of temperature and salinity parameters have been used for analyzing the principal possibility and the value of supercooling in the extremely sharp halo-pycnocline during flood period. Data filtration with depth has not been performed.

Observation results

Processing of temperature and salinity observational data has shown the existence of zones of supercooled water 5 cm to 1.5 m thick in the pycnocline (the interface between river and sea water). Supercooling in some thin layers of these zones $\Delta T_f = T - \tau_s$, where T is the measured (actual) water temperature and τ_s is the water freezing temperature at the given salinity, was calculated using standard algorithms (Fofonoff and Millard, 1983). At a first glance, it was reaching the improbable values of -0.6 to -0.8°C (Figure 2c). Mean super-cooling values varied from -0.01°C up to -0.1°C. The persistence of such supercooling under natural conditions has not been earlier observed. At least the authors are not aware of cases where similar values were directly determined in such an extensive region.

The depth of the supercooled layer, its thickness characterizing the extent of penetration of cooling to the pycnocline and the value of supercooling itself depended on the time and place of measurements. Prior to the spring flood (end of May), supercooling was only observed at stations located at the traverse of the main discharge branches Trofimovskaya and Bykovskaya. The upper boundary of supercooled water is confined to the lower boundary of freshened surface water (Figure 2). Similar position of the location of the upper boundary of the supercooled layer was observed at all stations where this layer was recorded. The supercooled layer itself at the weak inflow of river water occupies a small upper part of the pycnocline (Figure 2a). At the time of maximum discharge from early to mid-June it occupied either much of the pycnocline or the entire pycnocline (Figures 2b, c). Prior to the flood, the value of supercooling in the upper part of the pycnocline was small and rarely exceeded 0.05°C (Figure 2a). With the rising flood the intensity of freshwater inflow sharply increased, causing the lense of fresh water to greatly expand its area below the fast ice. The thickness of the fresh river water layer gradually increased with entrainment and mixing of the underlying layer of saline and cold water (Figure 3). This is especially well seen at the stations located along the traverse crossing the outflow from the main branches (Figure 3). Finally, river water at some comparatively shallow stations occupied the entire water column between the fast ice and the seafloor. With increasing thickness of the river water layer (Figure 3), the upper boundary of the supercooled layer is lowered (Figure 2). The increased rate of river water inflow, associated

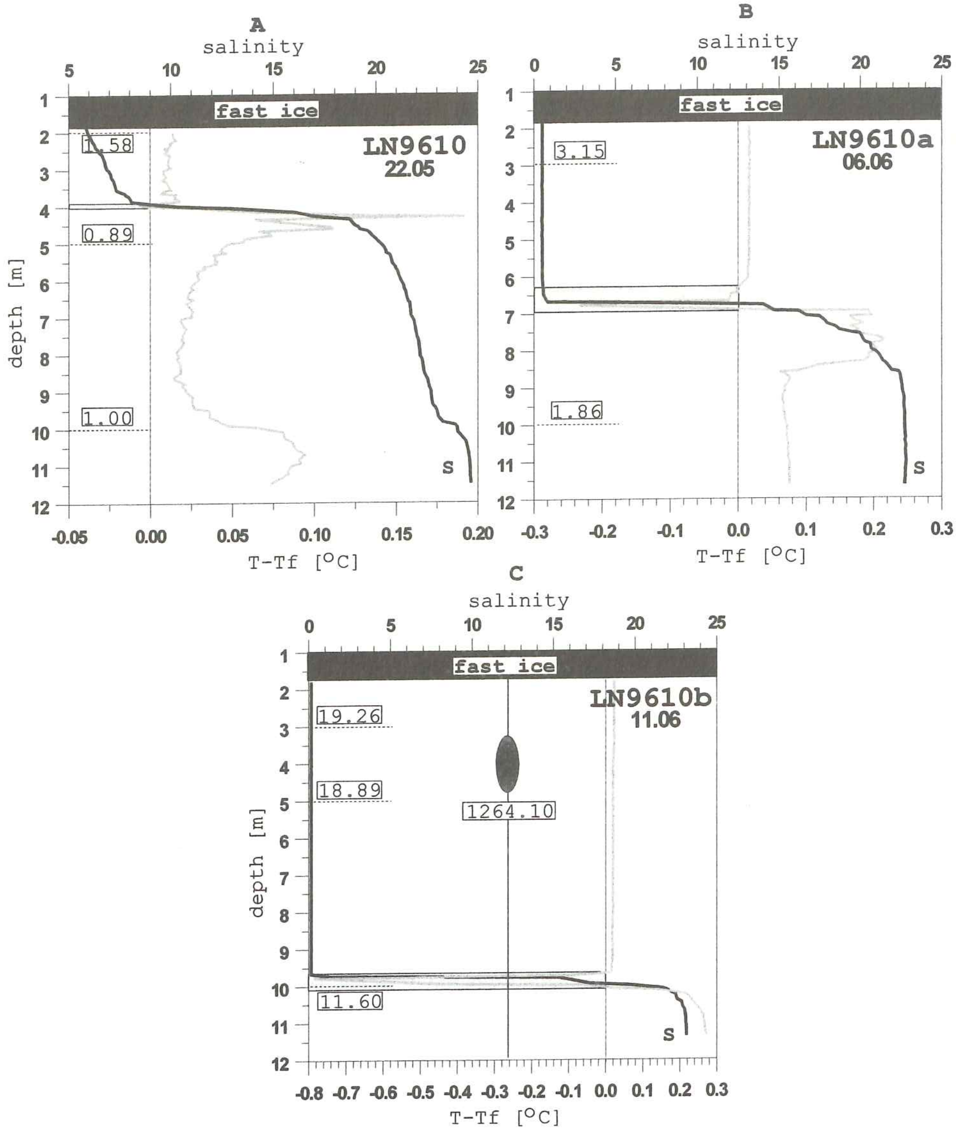

Figure 2: Vertical salinity distribution and differences between the in situ temperature and the freezing temperature at a given salinity for stations LN9610(A), LN9610a(B), LN9610b(C). 2.25 - contamination of suspended particulate matter into the water and frazil ice samples (mg/l); - zone of supercooling; - an agglomerate of frazil ice.

with increased horizontal velocity and turbulence results in a thickening supercooled layer and a greater absolute value of supercooling. This is due to a more intense turbulent entrainment at the river-sea water boundary (Figure 2).

It should be especially stressed that maximum supercooling rather occurs within pycnocline (Figures 2b, c) than at the freshwater/pycnocline boundary, as has been, for instance,

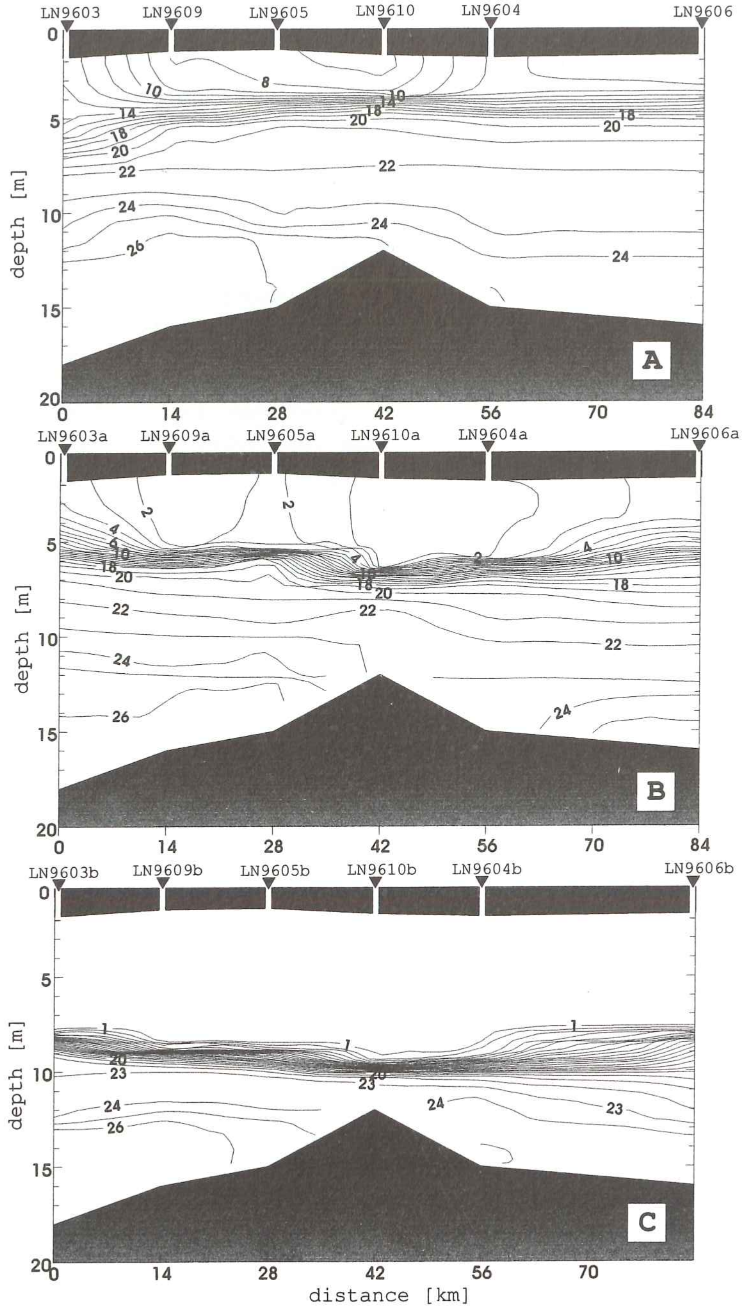

Figure 3: Evolution of vertical water salinity distribution at the onset of the spring flood at the transect along meridian 130 30'E from 72 to 72 45'N; (A) - May 18-22, (B) - June 3-6, (C) - June 10-11.

suggested by Martin and Kauffman (1974). This is determined by a specific process of turbulent entrainment at the pycnocline/freshened layer boundary. In course of laboratory experiments Kan and Tamai (1994) described the mechanism of inner waves initiation at the pycnocline boundary due to velocity shift together with their subsequent collapse due to Kelvin-Helmgolz instability. This results in formation of vortical structure at the pycnocline/fresh water boundary (Kan and Tamai, 1994). The latter favours more effective local penetration of fresh and warm water from the upper layer to pycnocline represented by more saline and cold water. In this case maximum supercooling occurs at that level within pycnocline where non-turbulent freshened water has already penetrated. This happens if loss of heat is more effective than heat exchange at molecular and molecular-turbulent levels (Krylov and Zatsepin, 1992; Voropayev et al., 1995). Buoyancy force hinders fast mixing within pycnocline, whereas part of more saline and cold water from pycnocline rapidly gets mixed when penetrates into the strongly turbulent upper layer (through this mechanism of turbulent entrainment). Hence the supercooling at the pycnocline/upper water layer boundary is considerbly weaker than that observed within pycnocline. If during the period preceding flood intensive turbulent entrainment at this boundary is absent, weak supercooling is observed only at the very boundary between the freshened water layer and pycnocline, and not within pycnocline (Figure 2a). This is in a good accordance with observational data (Martin and Kauffman, 1974).

The presence of supercooled water layers in the pycnocline caused the active formation of frazil ice in these layers. A bottom - temperature and pressure meter was suspended from cable at st. LN9610 at the traverse of the Trofimovskaya branch (Figure 1). It remained there from May 22 to June 11. Thus the period from before through the flood was covered. When the instrument was recovered, an agglomeration of frazil ice was found on the cable at the depth of the pycnocline. Its weight was about 1.5-2 kg and its height was about 50-60 cm. It consisted of transparent chaotically oriented crystals of frazil ice wither which inclusions of sediments transported by river water were clearly seen (Figure 4). The cable in this case served both as a nucleus and a place where crystals of frazil ice were trapped.

Thus as a result of the experimental studies in the Lena delta during spring flooding breakup, the existence of supercooling in the river-sea water contact zone, as well as that of frazil ice formation in this zone were established.

Discussion

Let us estimate the probability of supercooling in the river/sea water contact zone and the rate of possible frazil ice formation at the different regimes of the heat-mass exchange through the pycnocline using the results of laboratory studies.

Features of supercooling in the zone of river-sea water contact

As is known, supercooling of fresh river water from the underlying saline and cold sea water is the necessary physical condition for frazil ice formation. Stable formation of frazil ice is possible only when the freshened layer is close to its freezing point (Golovin et al., 1996; Krylov and Zatsepin, 1992; Voropayev et al., 1995,), i.e. when the following condition is fulfilled:

$$T_1 \sim -a \cdot S_1 , \qquad \text{or } T_1 / S_1 \sim -a \qquad (1)$$

Here T_1 and S_1 are temperature and salinity of the layer, freshened by river runoff; a=0.055 ^{o}C/ppt is the linear relation coefficient between the freezing temperature and salinity. The

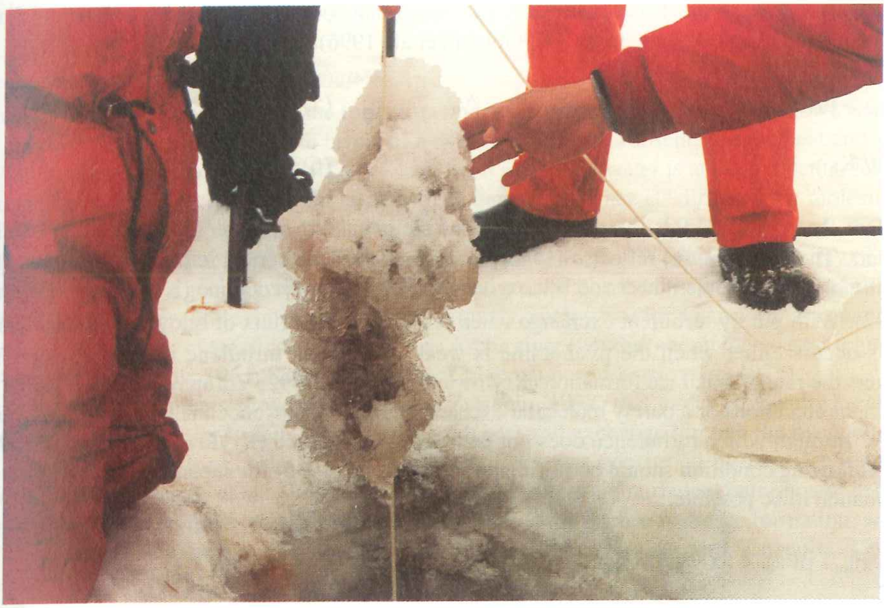

Figure 4: A agglomerate of frazil ice observed at st. LN9610b (Photo by Dr.J.Hölemann).

most favorable conditions for frazil ice formation at the lower boundary of the freshened layer exist when the temperature is also close to its freezing point in both the lower and the upper layers (Krylov and Zatsepin, 1992). Thus supercooling at the lower boundary of freshened water (the upper part of the pycnocline) (Figure 2) is possible at the condition:

$$\Delta T < a \cdot \Delta S, \qquad \text{or } \Delta T/\Delta S < a \qquad (2),$$

where ΔT and ΔS are the temperature and salinity differences through the density interface between the upper layer freshened by the river runoff and the underlying layer of cold and saline water (Figures 2, 3) ($\Delta T > 0$, $\Delta S > 0$). Since supercooling and formation of frazil ice are possible only at a more intense heat exchange through the density interface, as compared to salt (double-diffusion) (Krylov and Zatsepin, 1992; Voropayev et al. 1995; Golovin et al., 1996), the following condition should be fulfilled:

$$Ft > a \cdot Fs \qquad (3)$$

or in the density expression:

$$\alpha\, Ft / \beta Fs = (Kt/Ks) \cdot R\rho^{-1} > a \cdot (\alpha/\beta) \qquad (4)$$

where Ft and Fs are the heat and salt fluxes through the pycnocline, respectively. They have different signs, since Ft is directed downward from the warm freshened layer and Fs is directed upward from the saline underlying layer. Here $\alpha = -1/\rho \cdot (\partial\rho/\partial T) = 7 \cdot 10^{-5}\,(^{o}C)^{-1}$ is the thermal expansion coefficient, $\beta = 1/\rho \cdot (\partial\rho/\partial S) = 8 \cdot 10^{-4}$ is the salinity compression coefficient whose dimension is inverse to salinity and Kt and Ks are the effective heat and salt exchange coefficients through the density interface. Two important conditions for supercooling and frazil ice formation were obtained from (4) (Golovin et al., 1996):

$$R\rho > \beta/(\alpha \cdot a) \sim 200 \qquad (5)$$

$$Kt/ Ks > 1 \qquad (6).$$

Here $R\rho = \beta\Delta S/\alpha\Delta T$ is the density ratio for the pycnocline between freshened and saline waters. The condition (6) reflects the fact that the most efficient frazil ice formation is possible with molecular-turbulent heat and salt exchange through the pycnocline (Krylov and Zatsepin, 1992). With purely turbulent exchange when $Kt/Ks \approx 1$, the effect of buoyancy is negligible. This occurs either when the pycnocline is weak or external turbulenc is very strong. This lowers the rate of frazil ice formation (Krylov and Zatsepin, 1992; Voropayev et al., 1995).

Under conditions of a purely molecular exchange through the pycnocline when the "molecular core" through which turbulence does not penetrate is preserved (Krylov and Zatsepin, 1992), the following condition should be fulfilled (Golovin et al., 1996) for supercooling and frazil ice formation to be possible:

$$\alpha\, Ft_m / \beta Fs_m < (\alpha Kt_m/\beta Ks_m) \cdot a = 0.48 \qquad (7).$$

Here $\alpha\, Ft_m$ and βFs_m are buoyancy flux associated with the molecular heat and salt fluxes through the pycnocline, and Kt_m and Ks_m are the molecular coefficients of temperature conductivity and salt diffusion respectively. The condition (7) can be presented in the form:

$$R\rho^{-1} \cdot (Kt_m/Ks_m) < 0.48 \qquad (8).$$

Let us check the conditions for supercooling and frazil ice formation (1), (2), (5) and (8) at the upper boundary of the pycnocline. For this purpose we will use the results of measurements at station LN9610 where measurements were conducted prior to the flood on May 22, on June 6, and on 11 - the flood period. Table 1 presents the corresponding estimates.

Table 1: Results of checking the conditions for supercooling and frazil ice formation in the zone of river-sea water contact at st. LN9610, LN9610a and LN9610b.

Station (date)	LN9610 (May 22)	LN9610a (June 6)	LN9610b (June 11).
$T_1 / S_1 \sim -a$	0.051	0.058	0.045
$DT / DS < a$	0.042	0.035	0.045
$Rr > 200$	272	303	255
$Rr^{-1}(Kt_m/Ks_m) < 0.48$	0.37	0.30	0.39

As is seen from the table, all conditions for the occurrence of supercooling and frazil ice formation at the lower boundary of the freshened layer and in the upper part of the pycnocline in the region of the Trofimovskaya branch are fulfilled. These conditions are also fulfilled in the other near-delta regions where an intense inflow of river water was observed. Thus, even if there were no fine temperature and salinity measurements allowing us to directly determine the layers with supercooled water (Figure 2), then based on ratios (1), (2), (5) and (8), we would still be able to estimate the probability of supercooling in the river-sea water contact zone.

The dynamic aspect of frazil ice formation

Studies of double-diffusion frazil ice formation under laboratory conditions have shown that if the external turbulent effect in both layers of a two-layer system is absent, and the heat and salt exchange through the density interface between fresh and saline water is at the molecular level, then the value of supercooling is not large and the density interface is "diffused" by molecular diffusion (Krylov and Zatsepin, 1992; Martin and Kauffman, 1974; McClimans et al., 1978; Stigebrandt, 1981). Here less than several millimeters of frazil ice can form in a day. This important result is in good agreement with frazil ice growth observed in a summer lead filled with melt freshwater (Golovin et al., 1996). Obviously, such small ice formation rates cannot produce meter-layers of frazil ice observed near Arctic river mouths and in the coastal regions of Antarctica (Bulatov, 1963; Kozlovsky, 1971; Cherepanov and Kozlovsky, 1972).

Laboratory experiments showed (Krylov and Zatsepin, 1992) that under artificially created turbulence of a two-layer system and the increased heat-salt exchange through the pycnocline, the rate of frazil ice formation strongly increases. This was confirmed by later laboratory experiments (Voropayev et al., 1995). Turbulence in the layers and turbulent entrainment in the contact zone between freshened and saline waters provide for intense frazil ice formation, with ice crystals subsequently surfacing and sticking together (Krylov and Zatsepin, 1992, Voropayev et al., 1995).

Both-scale studies and laboratory experiments show that supercooling and frazil ice occur in a comparatively thin contact zone at the upper boundary of the pycnocline (Golovin et al., 1996, Krylov and Zatsepin, 1992, Voropayev et al., 1995). Hence, for investigating the rate of frazil ice formation depending on the level of the heat-mass exchange, the parameters and structure of turbulence near the density interface should be known.

In some laboratory experiments turbulence was generated by velocity shear (Turner, 1973; Kantha and Phillips, 1977; Kato and Phillips, 1969; Kan and Tamai, 1994). Such experiments were carried out in circulate flumes. Shear flow was either generated by a rotating plastic screen (Turner, 1973) or by belts installed (Kan and Tamai, 1994) on the bottom and surface of on of the flume straight parts. Roughness elements were attached on the belts. In other laboratory experiments turbulence was generated by oscillating grids (Turner, 1973; Krylov and Zatsepin, 1992; Voropayev et al., 1995). The grids attached to a single rod were mechanically activated. Turbulence induced by these means is called "grid" turbulence. Its structure near pycnocline differs from that of the turbulence generated by velocity shear (the latter predominates in natural environments). Hence, as noted by Turner (1973), this makes comparison of the results of these experiments difficult. However, since "grid" turbulence is more easily parametrizated, this way of turbulence generation is more frequently used in laboratory experiments studying the processes of properties transmission through pycnocline.

Studies of the rate of frazil ice formation (Krylov and Zatsepin, 1992; Voropayev et al., 1995) were performed by means of "grid" turbulence. In these experiments, parameterization of the non-dimensional entrainment velocity U_e/U_* through the pycnocline at turbulent mixing between the layers and interpretation of the results, including the rate of frazil ice formation, were performed by means of the local Richardson number $Ri_* = g \cdot (\Delta\rho/\rho) \cdot L/U_*^2$. The local Richardson number is an analogue of the global number, but the external scale of mean horizontal velocity U is replaced by the root-mean-square velocity of turbulent variations U_* near the density interface. Also, the thickness of the mixed layer Z is substituted to the integral scale of turbulence L. The latter characterizes the mean scale of the most energy-carrying eddies near the density interface which then penetrate the pycnocline and participate in turbulent entrainment (Prandtl, 1949; Turner, 1973; Shlikhting, 1974).

In order to use the ratios obtained by Krylov and Zatsepin (1992) for estimating the rate of frazil ice formation in the river/sea water contact zone based on full-scale studies, we should correctly relate the mean horizontal velocity in the freshened layer U to the root-mean-square velocity of turbulent variations U_*, as well as determine the integral turbulence scale L near the pycnocline. Turner (1973) suggested considering entrainments at the upper pycnocline boundary due to turbulent motion of the upper layer to be entrainment at flat turbulent jet margin. In this case the momentum and mass balance between layers would be probably reached, and the rate of turbulent entrainment would be quasisteady.Current laboratory studies of the transfer processes through the density interface confirm this interpretation (Kan and Tamai, 1994). This allows us to use the experimentally well-tested theory of turbulent entrainment at the expanding margin of a flat turbulent jet (Prandtl, 1949; Shlikhting, 1974) for relating the external mean velocity of the layer U to the mean-root-square velocity of turbulent variations U_* near the pycnocline.

Where a turbulent flat jet mixes with the ambient unmoving fluid the length of the mixing distance corresponding to the integral scale of turbulence L, according to the theoretical and laboratory studies (Prandtl, 1949; Shlikhting, 1974), is proportional in each cross-section to the width of the jet b in this section:

$$L = \alpha_v \cdot b \tag{9}$$

where α_v is the entrainment constant, varying from 0.08 to 0.12 in different experiments (Turner, 1973). Prandtl (1949) and Shlikhting (1974) determined its value to be 0.125. In our case, where entrainment at the boundary of river and sea water is quasi-steady, the width of the jet b will represent the thickness of the freshened layer (Figure 3). The mean velocity shear and eddies at the upper pycnocline boundary produce the internal waves that become unstable thus generating vorticity in the pycnocline (Turner, 1973; Kan and Tamai, 1994). As a result, part of the fresh water penetrates the more saline and cold water and part of the more saline water is entrained into the turbulent, swiftly moving river water, where it is rapidly mixed (Kan and Tamai, 1994). Thus, in the vorticity area of the pycnocline, there is active contact of saline and freshwater leading to supercooling. The thickness of this layer and the rate of supercooling depend on the characteristic scale of the energy carrying eddies which move in the mean flow generating vorticity in the pycnocline. In turn, the scale of these eddies depends on the external scale of the horizontal velocity in the freshened layer (Phillips, 1977). It follows from the above that the thickness of the supercooled layer in the pycnocline is actually the integral scale of turbulence L penetrating the pycnocline. We can determine it directly from the observational materials by analyzing the profiles $\Delta T_f = T - \tau_s$ (Figure 2).

According to the theory of Prandtl (1949), an additional tangential stress is created for free turbulence at the entrainment boundary of the parallel turbulent jet with unmoving fluid:

$$\tau = \rho \cdot \alpha_v{}^2 \cdot U_{max}{}^2 \tag{10}$$

where U_{max} is the maximum speed in a flat turbulent jet. In the case under consideration it is a typical scale of the horizontal velocity ($U_{max} = U$) in the upper freshened layer (Figures 2, 3). Based on the semi-empirical turbulence theory (Prandtl, 1949), let us determine the friction velocity (or the dynamic velocity) in the contact zone of freshened and saline waters from (10) as:

$$U_* = (\tau / \rho)^{1/2} = \alpha_v \cdot U \tag{11}$$

Prandtl (1949) showed that Reynolds stresses $U_* \cong (\overline{u \cdot v})^{1/2}$ characterize the intensity of turbulent (vertical - u' and horizontal - v') velocity variations and depend on the external velocity scale. Hence U_*, represented by the expression (11) determines the required velocity scale for the local Richardson number expressed through the external velocity scale U. As is seen from (11), U_* corresponding to $(\overline{u \cdot v})^{1/2}$, is approximately 10 times as small as the characteristic external velocity scale in the turbulent layer U, which is fully consistent with direct measurements of $(\overline{u})^{1/2}$, $(\overline{v})^{1/2}$ and Reynolds stresses $(\overline{u \cdot v})^{1/2}$ obtained from them (Kan and Tamai, 1994). The maximum value of turbulent velocity variations is observed in the thin upper and lower parts of the pycnocline where vorticity is generated and turbulent entrainment takes place (Kan and Tamai, 1994). Hence the expression for the local Richardson number for the pycnocline between river and sea water under conditions when the salinity field is governed by the density field ($\Delta\rho/\rho \cong \beta\Delta S$), will have the following form:

$$Ri_* = g \cdot \beta\Delta S \cdot L / \alpha_v{}^2 \cdot U^2 \qquad (12).$$

The above considerations concerning the interpretation of entrainment at the river-sea water boundary as entrainment at the boundary of a flat turbulent jet can be tested by a completely different method. Observations of the vertical hydrological structure at LN9610a and LN9610b stations performed during the time span between June 6 and 11 have been strictly bound to the same point on the fast ice. During the period of observation the upper pycnocline boundary displayed considerable downward shift (Figure 2). If considering the turbulent entrainment at the pycnocline/upper layer boundary to be responsible for such changes in the vertical hydrological structure, then its rate U_e can be directly determined. Intensive supply of flood water into the near-deltaic sea area during this time period (especially via the main branches) resulted in gradual entrainment of the underlying saline water into the riverine one together with their subsequent mixing. The U_e value was about $6 \cdot 10^{-4}$ cm/s. It should be noted that entrainment process is usually thought to be related to temperature and salinity variations in either the upper layer (if it is more turbulent than the lower one) or in both layers if their turbulence characteristics are similar (Turner, 1973). However, it should be remembered that indirect identification of the turbulent entrainment process at the density interface has been applied in an enclosed volume of water in course of the laboratory experiments (Turner, 1973). It is apparent that it is impossible to apply it for natural conditions since the boundaries are not closed. Moreover, in our case, when the upper layer remains fresh due to continuous flood water inflow to the sea, identification of such kind is inapplicable.

According to basic experiments by Turner (1973), at large values of the local Reynolds number $Re_* \cong \alpha_v \cdot U \cdot L/\nu$ (in our case it changes from 8400 to 84000 if U changes from 2 cm/s to 20 cm/s), at the mean thickness of the supercooled fluid in the pycnocline $\overline{L} \cong 35$ cm and at large Peclet numbers for salt $Pe_{*s} = Re_* \cdot (\nu/Ks_m)$ and heat $Pe_{*t} = Re_* \cdot (\nu/Kt_m)$, the non-dimensional ratio for the entrainment velocity U_e/U_* is function only of Ri_*. It approaches the constant value at $Ri_* \to 0$ (purely turbulent exchange regime). At the intermediate values of Ri_* (molecular-turbulent exchange regime), the change occurs according to the fundamental law determined by Turner (1973): $U_e/U_* \sim Ri_*^{-3/2}$. In experiments investigating frazil ice formation (Krylov and Zatsepin, 1992) the non-dimensional entrainment velocity for salt is well approximated by the same dependence:

$$U_e / U_* = A \cdot Ri_*^{-3/2} \qquad (13),$$

where A=0.56 is an empirical constant. From (13), knowing the mean thickness of the supercooled water layer $\overline{L}\cong 35$ cm during the June 6 to 11 period, mean salinity difference in the pycnocline $\overline{\Delta S}\cong 18.44$ and U_e (given above), and also considering (11), we can determine the scale of horizontal velocity in the upper freshened layer U. It equals 10-15 cm/s. Hence, according to (11), U* is about 1.3-1.9 cm/s. The estimated scale of horizontal velocity U in the upper layer seems to be high, but it has been obtained at the station situated at the traverse of the Bykovskaya branch (see Figure 1) which is the main channel for the Lena river fresh water runoff. Hence U value is quite realistic. Such independent evaluation of U and U* confirms our interpretation of the process of turbulent entrainment at the river/sea water boundary as an entrainment process at the flat turbulent jet margin.

The above given description of the approach used for determination of the local turbulence characteristics at the freshened layer-pycnocline boundary does not include evaluation of the Coriolis force influence, since it is insignificant and can be avoided. Similar approach to determination of the rate of entrainment at the upper pycnocline boundary (geophysical application) was used by Turner (1973). Besides this, according to Phillips (1977), at $Z< U_* / \Omega$, the influence of the Coriolis force can be neglected. Here Z is depth, Ω is angular speed of the Earth´s rotation. The U_* / Ω value for the above U_* estimates is about ≈260 m. Sea water depths in the region investigation (Figure 1) are only 6-20 m. So, it is obvious that during flood the Coriolis force does not influence the local turbulence patterns at the freshened layer/pycnocline boundary in the shallow sea area around the Lena river delta.

Estimate of the rate of frazil ice formation

As a result of laboratory experiments of frazil ice formation (Krylov and Zatsepin, 1992), the expression for estimating the rate of its formation was obtained as:

$$V_i = B\cdot U_*\cdot\Delta S\cdot Ri_*^{-1/2}\cdot\{(\Delta T/\Delta S - (7\cdot a/Ri_*)\} \qquad (14),$$

where $B = 3.3\cdot 10^{-3}$ $(°C)^{-1}$ at the density of frazil ice $\rho_i = 0.1$-0.3 g/cm^3 (Weeks and Ackley, 1982). An approximately similar, rough estimate of frazil ice density (≈0.2 g/cm^3) was obtained by measuring the water volume after melting a sample of frazil ice of known volume (at st. LN9610b). Here $\Delta T=T_1-T_2>0$ and $\Delta S=S_2-S_1>0$ are the temperature and salinity differences through the pycnocline. Based on the above interpretation of entrainment at the interface, let us determine from (12) the value of the local Richardson number Ri_* for the pycnocline between river and sea water at st. LN9610, LN9610a and LN9610b for various values of the external velocity scale in the freshened layer U at actually determined L and ΔS. The estimates are presented in Table 2. According to the classification by Krylov and Zatsepin (1992), at the local Richardson number $Ri_* >>10^2$ the exchange through the pycnocline occurs at the molecular level. At $Ri_*\approx 10^2$ with the decrease in the density difference in the pycnocline or the increase in the external velocity scale in the turbulent layers, the "molecular core" converges. This leads to the molecular-turbulent regime of mass- and heat exchange. At this regime the exchange of properties through the pycnocline mainly occurs due to sporadic outbursts of turbulence (eddy formation) in the area of a strongly sharpened density interface. However, turbulence in the interface area is significantly influenced by buoyancy and is not pronounced. This is probably the reason for the different effective exchange coefficients through the pycnocline at this regime. At $Ri_* <10$ there is a purely turbulent exchange regime. At $Ri_* < 2$ when stratification does not influence turbulence, the regime becomes unsteady resulting in a rapid mixing of layers.

Based on this classification, one see can from Table 2 that during the period prior to the flood, at mean velocity in the freshened layer of only about 5 cm/s, the regime of the molecular-turbulent mass-heat exchange already appears in the pycnocline, and at 15-20 cm/s there is a purely turbulent exchange regime. During the flood outflow of river water, when the upper layer becomes strongly freshened (Figures 2, 3), the molecular-turbulent exchange regime through the pycnocline contributing to intensification of frazil ice formation, occurs only at $U \approx 15$-20 cm/s and the turbulent regime at $U > 40$-50 cm/s.

Table 2: Estimates of the local Richardson number (Ri_*) for the pycnocline between river and sea water for stations LN9610, LN9610a and LN9610b at different velocities of river water transport in the upper layer

Station (date)	LN9610 (22.05)	LN9610a (06.06)	LN9610b (11.06)
U = 5 cm/s	106	1360	1167
U = 10 cm/s	26	340	292
U = 15 cm/s	12	151	130
U = 20 cm/s	6.6	85	73
U = 40 cm/s	-	21	18
U = 50 cm/s	-	13.6	11.7
U = 60 cm/s	-	9.4	8.1

Using Ri_* values from Table 2 and determining U_* from (11), let us estimate by means of (14) the real rate of frazil ice formation V_i for different values of the typical motion velocity of river water U in the upper freshened layer. Figure 5 presents the estimates of the rates of frazil ice formation in the contact zone between river and sea water at stations st. LN9610, LN 9610a and LN9610b.

Analysis of the calculated rates of frazil ice formation V_i shows that before the flood the rates are insignificant though sometimes reach 20 cm/day.This rate is typical for the outflow regions of the Trofimovskaya and Bykovskaya branches. Approaching peak river discharge, the rate of frazil ice production increases considerably and can reach 1.7 m a day (Figure 5). This equivals to ≈34 cm of regular sea ice formation if porosity of frazil ice layer is 80 % (Weeks and Ackley, 1982). The maximum rate of frazil ice formation is observed at U ~ 40-50 cm/s. With the increase of U the value V_i is sharply reduced (Figure 5). This is governed by the fact that at certain actual density stability of the pycnocline, a large increase in U leads to the turbulent ($Ri_* < 10$, see Table 2) and then to the unsteady regime ($Ri_* < 2$). Buoyancy is no longer influence the character of the heat -mass exchange through the pycnocline. As a result, the rate of frazil ice formation decreases as the values of the effective salt and heat exchange coefficients become equal ($Kt \approx Ks$). This reduces the efficiency of double-diffusion supercooling or makes it impossible. The conditions (3) and (6) are not fulfilled. A similar conclusion also follows from laboratory experiments (Krylov and Zatsepin, 1992; Voropayev et al., 1995).

A frazil-ice production rate of 1.5 m/day seems to be high, but the observed supercooling in the pycnocline at this time is also very large (Figures 2b, c). Such high ice-production rates will lead to speculations that fast ice thickness could increase accordingly. However, much of this new ice probably also melts locally, since toward the conclusion of our field work, river water with above zero temperature was already spreading below fast ice (Figure 2). A portion of the frazil ice probably is exported with river water farther to the sea, where it may add to the bottom surface of fast or drifting ice.

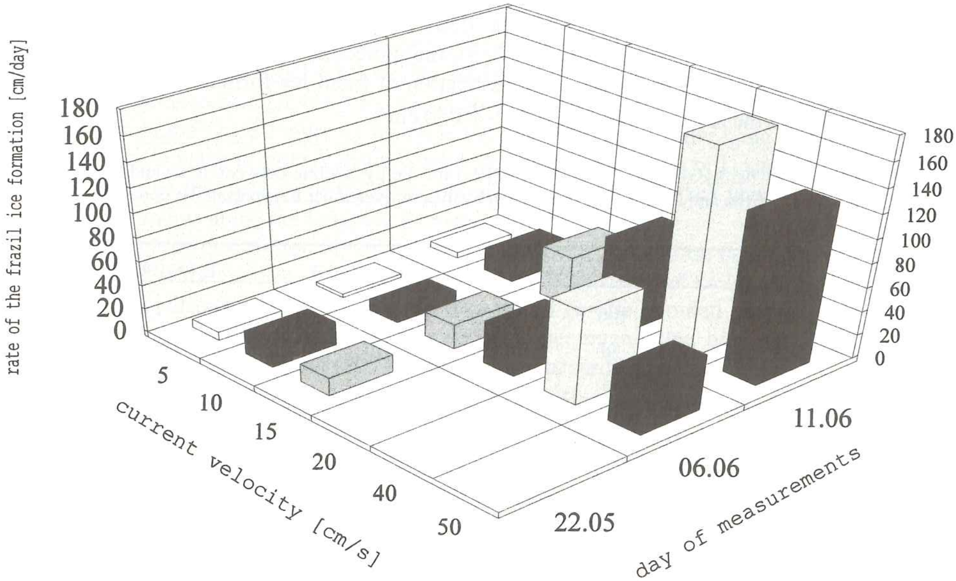

Figure 5: Estimates of the rate of frazil ice formation at st. LN9610 (22.05), LN9610a (06.06), LN9610b (11.06). at different current velocities in the upper freshened layer.

Along with the formation of frazil ice, sediments transported by river water are trapped. The mechanism of frazil ice formation plays quite a significant role for sediment transport to the surface of both fast and drifting ice. Thus at the rate of frazil ice formation of 170 cm a day at st. LN9610b (see Figure 5), the sediment flux to the fast ice can reach 7 g/m^2 a day at the actually measured sediment level in river water of 19.2 mg/l. In this case sediments were assumed to be entirely incorporated into frazil ice at its formation. Laboratory experiments have shown (Reimnitz et al., 1993) that for the actually observed granulometric composition of suspended sediments their significant increase in crystals of surfacing frazil ice can take place, as compared to the surrounding river water. Thus the rate of the sediment flux to the surface might be much higher. This fact was experimentally recorded during the Transdrift IV expedition. The level of suspended and dissolved matter in an agglomeration of frazil ice was 1264.1 mg/l, whereas in the upper freshened layer it was 18.89 mg/l and 19.26 mg/l at 5 m and 3 m depth, respectively.

Conclusions

Application of formula (14), obtained for a specific laboratory model for estimating the rates of frazil ice formation under natural conditions, can cause justified criticism. However, the results of studies in the Transdrift IV expedition made their interpretation possible on the basis of the laboratory experiments. This has permitted us to obtain quite important quantitative estimates of the rates of frazil ice formation and gain understanding of their spatial-temporal variability during the developing spring flood in the pro-delta region of the Lena river. These approaches can also be applied to other Arctic Seas subjected to high river discharge.

During the peak of its production frazil ice also entraps sediments transported by river water. When rising to the surface, frazil ice crystals can be incorporated into surface ice. The formation of frazil ice in the region where Arctic rivers discharge into the seas can be more or less intense during the whole year. This makes the mechanism of frazil ice formation and associated incorporation of sediments into the ice cover very important sediment transport and deposition in these regions. Future studies of the physical and sediment entrainment in pro-deltas also contribute to long-range transport in the Transpolar Drift.

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