Laboratory Experiments on Internal Solitary Waves in Ice-Covered Waters

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Key Points:

- Internal solitary waves in partially ice-covered water are generated in the laboratory.
- Internal solitary wave induced currents can transport the ice horizontally.
- In the laboratory experiments, dissipation of turbulent kinetic energy under the ice is comparable to that at the wave density interface.

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Abstract

Internal solitary waves (ISWs) propagating in a stably-stratified two-layer fluid in which the upper boundary condition changes from open water to ice are studied for cases of grease, level and nilas ice. The ISW-induced current at the surface is capable of transporting the ice in the horizontal direction. In the level ice case, the transport speed of, relatively long ice floes, non-dimensionalised by the wave speed is linearly dependent on the length of the ice floe non-dimensionalised by the wave length. Measures of turbulent kinetic energy dissipation under the ice are comparable to those at the wave density interface. Moreover, in cases where the ice floe protrudes into the pycnocline, interaction with the ice edge can cause the ISW to break or even be destroyed by the process. The results suggest that interaction between ISWs and sea ice may be an important mechanism for dissipation of ISW energy in the Arctic Ocean.

1 Introduction

Internal waves (IWs) are ubiquitous in stratified water. They propagate through the stratified water column and, in the case of a two-layer fluid, along the density interface. In the Arctic Ocean, IW dynamics (i) constitute an integral part of the circulation and thermodynamics (Levine, Paulson, & Morison, 1985; Sandven & Johannessen, 1987), and (ii) play an important role in the vertical mixing of the upper layers (D'Asaro & Morison, 1992; Fer, 2014; Kirillov, 2006) which, in turn, affects renewal of nutrients and sea ice evolution.

The majority of past observations have been of low-frequency IWs, based on sparse *in situ* measurements due to year-round sea ice cover. Recently, however, as a consequence of prolonged ice retreat, synthetic aperture radar (SAR) observations have facilitated the mapping of high-frequency internal solitary waves (ISWs) in open water areas of the Arctic Ocean and its marginal seas (Kozlov et al., 2015; Kozlov, Zubkova, & Kudryavstev, 2017; Zimin, Kozlov, Atadzhanova, & Chapron, 2016). The waves are typically non-linear, rank-ordered packets in which the leading wave has an amplitude that is comparable to the upper layer depth.

A dominant generator of IWs is tidal flow over topography, where the energy of the surface tide is converted to the internal (baroclinic) tide, typically at semidiurnal frequencies. For most of the major tidal constituents, the Arctic is above the critical latitude where the propagation of linear IWs is suppressed by the Earth's rotation. Yet, IWs have been observed in the Arctic Ocean, associated with increased turbulence (Padman & Dillon, 1991; Rippeth et al., 2017). The mechanisms of generation and energy pathways to turbulence are not yet fully understood. The frequency of the IW packets and their characteristic spatial scales suggest that the majority are linked to the barotropic tide and are a consequence of a lee wave generation process first proposed by Vlasenko, Stashchuk, Hutter, and Sabinin (2003). Recent idealized modelling and field observations (Rippeth et al., 2017) have supported the lee wave generation process and have linked such non-linear IWs to enhanced mixing through collocated velocity microstructure measurements.

Enhanced levels of dissipation of turbulent kinetic energy (TKE) are well documented over areas of sloping topography in the Arctic Ocean (D'Asaro & Morison, 1992; Fer, Skogseth, & Geyer, 2010; Padman & Dillon, 1991; Rainville & Winsor, 2008; Rippeth, Lincoln, Green, Sundfjord, & Bacon, 2015). This is in contrast to the central Arctic Ocean, which is remarkably quiescent (Fer, 2009; Lincoln et al., 2016) and has shown a lack of significant long-term trends despite changes in sea ice concentration (Guthrie, Morison, & Fer, 2013). By synthesizing a pan-Arctic data set Rippeth et al. (2015) suggest tidal conversion as the main source of enhanced dissipation rates and vertical heat fluxes, despite much of the Arctic Ocean lying poleward of the critical latitude. Short scale IWs (which are not affected by the Earth's rotation) have been proposed as the agency responsible for the transfer of energy from the tide to turbulent mixing at such latitudes (Rippeth et al., 2017).

IWs are known to cause flexure of sea ice (Czipott et al., 1991; Marchenko, Morozov, Muzylev, & Shestov, 2010) and theoretical studies (Muench, LeBlond, & Hachmesiter, 1983; Saiki & Mitsudera, 2016) suggest they are responsible for the formation of ice bands in the marginal ice zone (MIZ). The annual variation of the Arctic ice edge is monitored carefully (i) to assess climate change and (ii) for a variety of practical reasons involving sea traffic, fisheries, offshore operations and military marine activities. There is clear interplay between IWs and sea ice and motivation to study the topic is wide ranging, yet very few dedicated investigations exist.

While field observations provide insight into IW dynamics in the Arctic Ocean, none to date isolate the effects of individual ice, ocean and wind parameters. Moreover, field measurements in the MIZ are particularly challenging and hazardous due to variable ice thickness, and high potential for ice floe breakup. In addition, the harshness of the environment and the remoteness of the location can limit observations, especially in periods of ice-cover. SAR imagery has provided valuable new insight but its use is restricted to areas of open water and to climatic conditions being favourable for the observations to be made. There is a clear need, therefore, to supplement field work with modelling studies. In this study, a laboratory investigation of ISWs in a two-layer stratified flow propagating from open water to under ice is presented. The ice type is varied and the interaction between the ISW and ice investigated. The aim of the paper is to model physically the interaction of an ISW with different types of ice and hence demonstrate that this interaction can lead to dissipation of ISW energy through different pathways including (i) boundary friction at the underside of the ice, (ii) transportation of the ice and (iii) deformation of the wave form. The study is primarily illustrative and qualitative; a full sweep of parameter space being beyond the scope of the current paper. This is the first time that ISWs have been generated under ice in a laboratory setting.

2 Experimental Set Up and Procedure

2.1 Physical Arrangement and Wave Generation

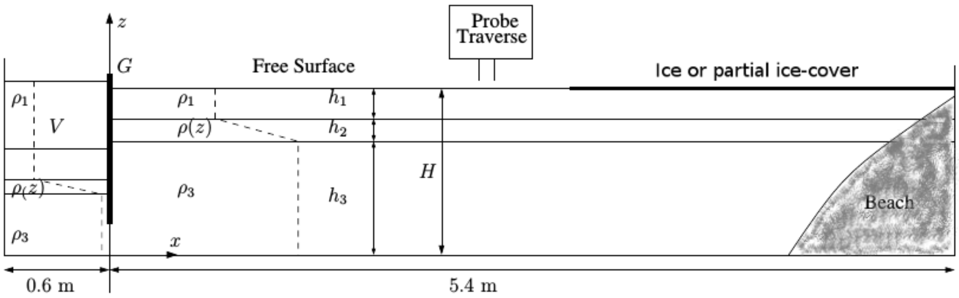

Figure 1. Schematic diagram (not to scale) of the laboratory arrangement.

The experiments took place in a wave flume 6m in length, 0.47m in width and 0.6m in depth. The flume was built within the HSVA Arctic Environmental Test Basin facility in Hamburg. The flume was of plexiglass construction to allow visualisation from the side and illumination from below. It was filled with homogeneous salt water of prescribed density $\rho_3 = 1045$ kg/m^3. Less dense brine solution of density $\rho_1 = 1025$ kg/m^3 was then slowly added via an array of floating surface sponges. Consequently, an interface (pycnocline) between the two fluids formed in which the density varied as a linear function of depth z. The density range was chosen such that neutrally-buoyant particles could be suspended throughout the depth of the water column enabling flow visualisation and measurement (see §2.4). A gate G, was inserted at the upstream end of the tank (see fig 1) and lowered to approximately 1 cm above the bed of the flume. A fixed volume, V, of water of density ρ_1 was then added behind the gate. Due to hydrostatic balance, fluid of density ρ_3 flowed under the gate into the main section of the flume. Once the volume, V, had been added, the total fluid depth H, in the main section of the flume, was measured using a pre-set tape on the plexiglass window. The laboratory air temperature was kept just above 0°C while the flume was filled.

2.2 Ice Formation

After the flume had been stratified, different ice cover types were applied. A vast array of ice types can be found in the Arctic Ocean hence the ice types investigated were chosen to cover as broad a range as practically possible. The laboratory was cooled to approximately -2°C or -15°C in cases where ice was added or made respectively. The temperature and thickness of the ice was monitored throughout the duration of a given experiment. No change of form or melting was observed.

(i) **Model level ice** was formed from a 0.7% sodium chloride solution following the procedure outline in Evers (2015). During the freezing process tiny air bubbles of diameter 200−500 μm were embedded into growing ice crystals. Due to the embedded air, the mechanical properties of the model ice (strength, elasticity, fracture behaviour and density) were made as close as possible to the mechanical properties of ice prevailing in nature. In particular, the model ice had a density range of 782−803 kg/m^3 which is within the natural range for sea ice of 720−940 kg/m^3. Sections of level ice were cut and removed from the laboratory and kept in a cold storage unit. When required, the floes of level ice were carefully lowered into the stratified flume. The level ice floes had a length, $l_f \approx 1$ m, and two different thickenesses were considered namely $d_f = 0.013$ m and $d_f = 0.058$ m.

(ii) **Grease ice** was made by crushing stored model level ice and adding it carefully to the surface of the stratified water while taking care to disturb neither the pycnocline nor the upper layer. Individual pieces of grease ice had complex geometries and the typical length scale of a piece ranged from 0.001 m to 0.05 m. These pieces were arranged such that a surface length of 2.65 m was covered by the ice and the total thickeness of the grease ice layer was approximately 0.02 m.

(iii) **Nilas ice** was made by reducing the air temperature in the laboratory (to -15°C) so that the surface of the water column froze. Open water sections were maintained by placing styrofoam at the surface during the freezing process. The styrofoam lids were removed just prior to an experiment commencing. The thickeness of the nilas ice layer was approximately 0.006 m and it covered a surface length of 2.69 m.

2.3 Wave Generation

ISWs were generated by the swift, smooth removal of the gate in the vertical direction. After a sorting distance of approximately 1 m, an ISW of depression propagated horizontally along the pycnocline into the main section of the tank. A beach (of polyether filter foam) was located at the downstream end of the flume to absorb some of the ISW energy but a reflected ISW signal was still seen in all but one of the experimental runs. Once an experiment was finished and the water column stationary, the gate was re-inserted and a fixed volume of fluid of density ρ_1 was again added behind the gate so that a second run could be performed.

2.4 Flow Measurement and Flow Visualisation

With the ice in place and the water column stationary, high precision micro-conductivity sensors (Munro & Davies, 2009) were used to measure the form of the stratification. The sensors were mounted on a rigid rack and pinion traverse system fitted with a potentiometer. The sensors were moved vertically through the water column and density profiles were obtained by calibrating the potentiometer output and conductivity data against known values of height and fluid density respectively.

Analysis was restricted to two dimensions, (x, z), where x denotes the horizontal direction and z the vertical direction, with the origin $x = 0$ corresponding to the horizontal location of the gate and $z = 0$ to the bed of the flume (see fig 1). The vertical extent of the upper, middle and lower layers of the stratification were denoted by h_1, h_2 and h_3 respectively. The ratio of the layer thicknesses, and in particular the upper two layer thicknesses, were chosen to be similar to those observed in the Eurasian Arctic shelf seas and slopes (Fer et al., 2010; Kozlov, Romanenkov, Zimin, & Chapron, 2014; Padman & Dillon, 1991). The formation of the stratification was difficult to control precisely due to (i) disturbances during filling, (ii) disturbances when adding the ice and (iii) differences in filling times. As a result the layer depths varied such that, $h_3 = 0.32{\pm}0.01$ m, $h_2 = 0.04 \pm 0.01$ m and $h_1 = 0.045 \pm 0.002$ m for the first run of a given experiment, and $h_3 = 0.32 \pm 0.01$ m, $h_2 = 0.04 \pm 0.01$ m and $h_1 = 0.055 \pm 0.002$ m for the second run.

A light source (intense LED strip passed through a double slit), was placed beneath the transparent base of the tank. It generated a thin, vertical column of light which was arranged to illuminate a two dimensional slice of the flow field in the mid-plane of the flume aligned with the x axis. The water column was seeded with neutrally-buoyant, light-reflecting tracer particles of "Pliolite" having diameters in the range $150-300$ μm (Fructus, Carr, Grue, Jensen, & Davies, 2009). Motions within the vertical light sheet were viewed and recorded from the side using three fixed digital video cameras set up outside the flume. The cameras (UNIQ UP-1830-CL-12B) had a spatial resolution of $1372{\times}1372$ pixels and a capture rate of 30 frames per second. The three cameras were synchronised in time and two were positioned to have overlapping fields of view.

The software package *DigiFlow* (Dalziel, Carr, Sveen, & Davies, 2007) was used to process the digital video records. The time series function of *DigiFlow* was used to measure wave phase speed c, wave amplitude a, wave length λ and wave-induced ice floe speed c_f. The Particle Image Velocimetry (PIV) function of *DigiFlow* was used to calculate continuous synoptic velocity and vorticity field data along the illuminated cross-section in the middle of the flume. The average ice thickness d_f and ice length l_f were measured with calipers and a rule respectively. The width of the ice was as close as practically possible to the width of the tank.

3 Results

3.1 Ice Motion

Provided the ice was free to move, the ISW-induced current at the surface transported the ice horizontally in the same direction as that of wave propagation. This is illustrated in figure 2 (a) in which still images from an experiment with level ice of dimensions $d_f = 0.013$m and $l_f = 0.95$m are presented. The images show light scattered from the neutrally-buoyant tracer particles and are displayed in a false colour scheme to optimise the visualisation. The ice at the surface can be seen by light reflected off its underside and the pycnocline can be identified by the clustering of tracer particles at the density interface. The ISW propagated from left to right.

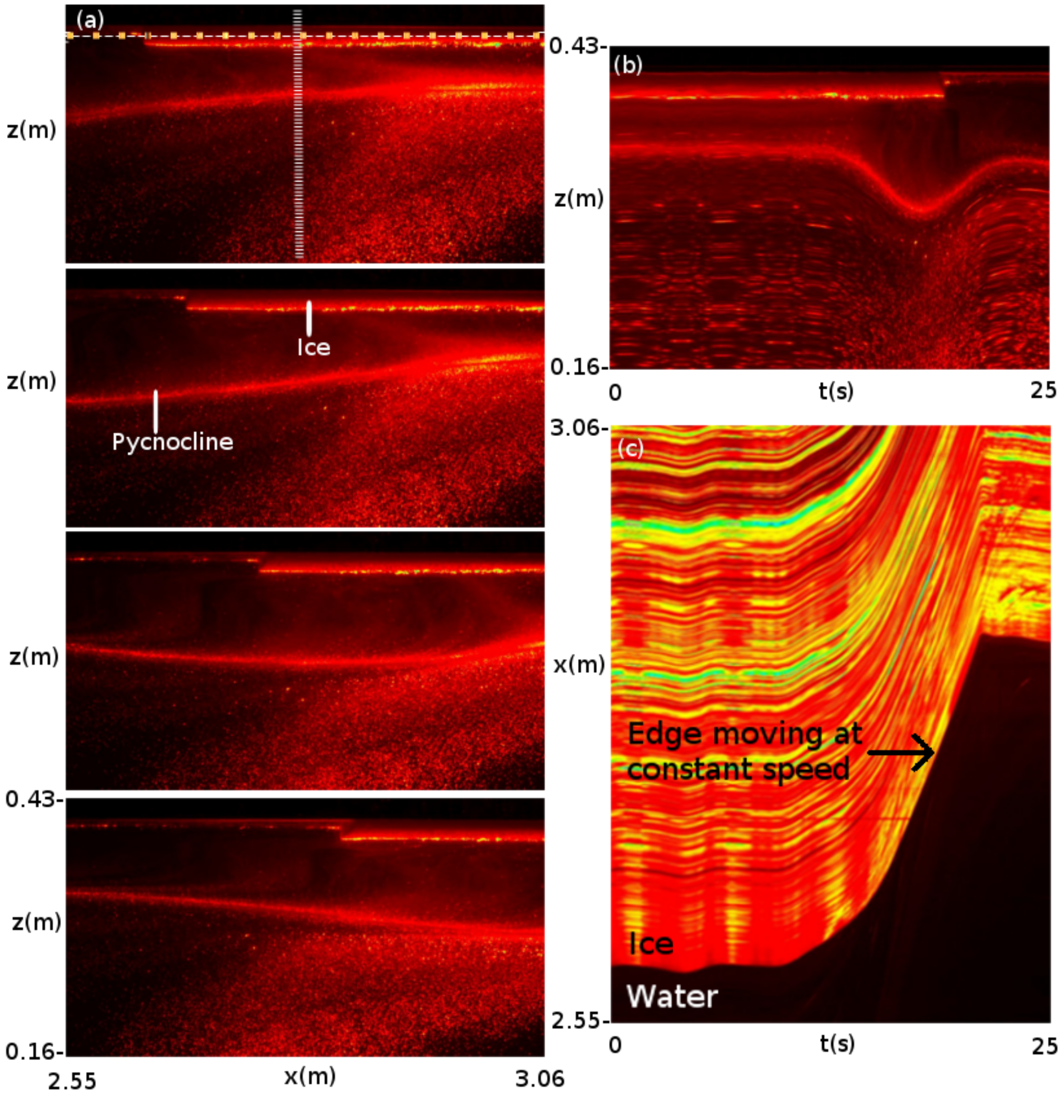

Figure 2. (a) series of still experimental images with level ice of dimensions $d_f = 0.013$m and $l_f = 0.95$m, showing light intensity, ordered sequentially in time from top to bottom at intervals of Δt=2.2 s, (b) vertical time series made from column marked by vertical dashed line in (a), and (c) horizontal time series made from row marked by horizontal dot-dashed line in (a) rotated through 90° anticlockwise.

Figure 2 (b) shows a time series constructed by measuring the temporal changes in image pixel values in a prescribed column at a fixed horizontal coordinate in each frame of the experimental movie over a time interval of 25 s. Such series were used to measure wave amplitude by tracing the vertical disturbance of a streamline at the top ($z = h_3 + h_2$) of the pycnocline from its undisturbed depth to maximal displacement. The time at which the pycnocline reached maximum displacement was also recorded. This process was repeated at three fixed locations $x_{1,2,3}$ over known horizontal distances $\delta x \approx 0.26$ m. The wave phase speed $\delta x/\delta t$ was obtained by noting the time δt between maximal interface displacement at the three fixed locations $x_{1,2,3}$. This process was repeated over all three cameras for waves in which there was no change of form. Average amplitude a and average wave phase speed c were then computed. The half wavelength, λ, was calculated from the time series by measuring the time taken for the pycnocline to be displaced from $z = h_3 + h_2 - a/2$ to $z = h_3 + h_2 - a$, to get a quarter wavelength and then multiplying the result by $2c$.

Figure 2 (c) shows a time series constructed by measuring the temporal changes in image pixel values in a prescribed row of lateral extent $x \in [2.55, 3.06]$ m at a fixed vertical coordinate in each frame of the experimental movie, over a time interval of 25 s. The horizontal slices (which coincide with the ice) are rotated through 90° anticlockwise and ordered sequentially from left to right. In this way the edge of the ice (marked by the interface between black and orange) can be traced. It can be seen that after an initial acceleration (due to the wave motion c.f. fig 2 (b)) the ice moves at a constant speed before it encounters the end of the wave flume and is arrested. The constant floe speed, c_f, was measured from the gradient of the straight section of the trace indicated by an arrow in fig 2 (c).

Figure 3 (a) summarises the range of wave phase speeds, wave amplitudes, ice types and dimensions investigated (see supplementary table for further details). In figure 3 (b) the non-dimensional floe speed c_f/c versus the non-dimensional ice floe length $l_f/2\lambda$ is presented for the level ice cases only. In these cases, the ice floe was relatively long, $l_f/d_f \in [17.2, 73.1]$ and completely free to move. In the grease ice case, the ice was not packed in tight and there were regions where it was possible to compress the ice but the ice initially extended to the end wall of the flume and hence was only partially free to move in the horizontal direction. Nevertheless, the wave-induced flow transported the grease ice horizontally (at a speed of $c_f/c \approx 0.25$). In the case of the nilas ice, the ice was fixed in the positive horizontal direction (by the end wall of the flume) and hence there was no wave-induced floe movement. Figure 3 (b) suggests that when the ice is free to move and the floe is relatively long, there is a linear relationship between the wave-induced floe speed and the floe length given by the equation of the plotted straight line, namely, $c_f/c = -0.61 l_f/2\lambda + 0.79$. The wave-induced floe speed was not found to have any dependence on the thickness of the ice floe d_f for the parameter range investigated ($d_f \in [0.013, 0.058]$ m).

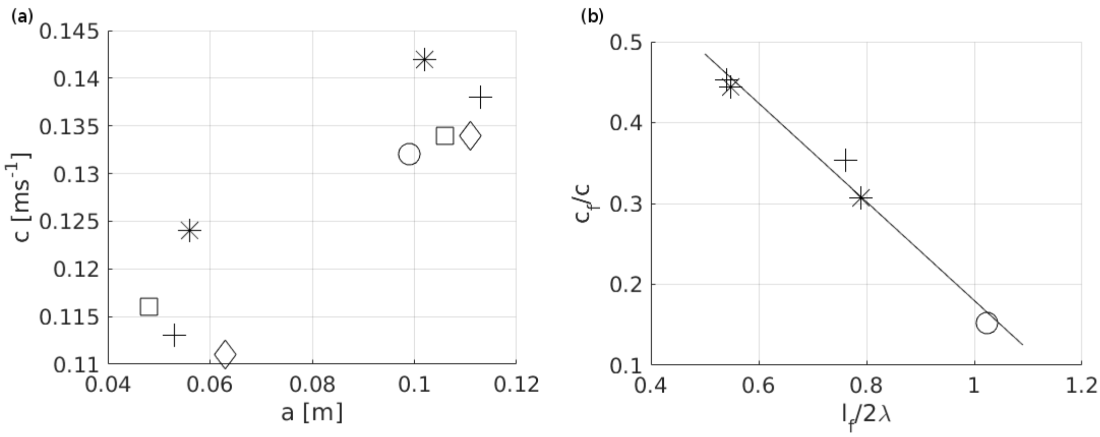

Figure 3. (a) Wave phase speed c versus wave amplitude a and (b) non-dimensional wave-induced level ice floe speed c_f/c versus non-dimensional level ice floe length $l_f/2\lambda$. Different symbols correspond to different ice types and dimensions as follows: $*$ level ice with $d_f = 0.013$ m and $l_f = 0.95$ m, $+$ level ice with $d_f = 0.058$ m and $l_f = 1.00$ m, $\circ$ level ice with $d_f = 0.058$ m and $l_f = 2.00$ m, $\diamond$ grease ice with $d_f = 0.021$ m and $l_f = 2.65$ m, $\square$ nilas ice with $d_f = 0.006$ m and $l_f = 2.69$ m. The solid straight line in (b) is a linear fit (with $R^2 = 0.982$) to the data points. Error bars are omitted from the data points as they were comparable with the marker size.

3.2 Wave/Ice Edge Interaction

In figure 4, still images from two experiments in which the ice protruded into the pycnocline are presented. The sub-panels are constructed from two overlapping camera outputs. In the smaller wave amplitude case (figure 4 (a)), the interaction of the wave with the ice edge caused deformation of the wave shape (figure 4 (a), (ii)-(iv)) and the wave amplitude was significantly reduced (figure 4 (a), (v)), so much so that in this case there was no reflected wave signal off the end wall of the flume. The wave was effectively destroyed by its interaction with the ice edge.

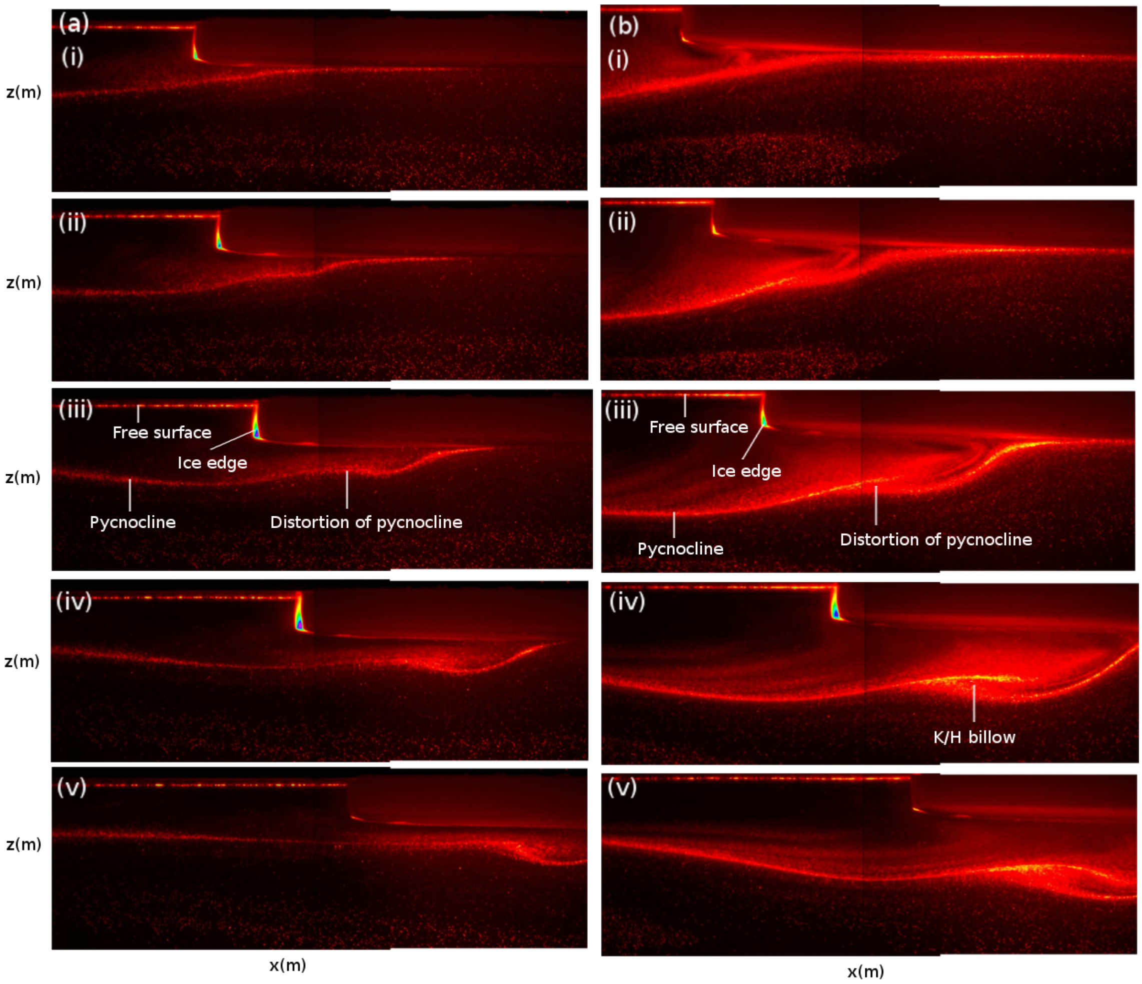

Figure 4. Series of experimental images for an experiment in which the ice was level ice of dimensions $d_f = 0.058$m and $l_f = 1.00$m, and the incident ISW had (a) $a = 0.053$ m and (b) $a = 0.113$m. The images show light intensity and ordered sequentially in time from top to bottom at intervals of $\Delta t = 2.14$ s; sub-panel dimensions are $(\Delta x, \Delta z) \approx (0.813\text{m}, 0.74\text{m})$. The ISW propagates from left to right.

In the larger wave amplitude case (figure 4 (b)), interaction of the wave with the ice edge caused similar deformation of the wave shape, to the smaller amplitude case but this time a billow resembling the overturning features associated with a Kelvin-Helmholtz instability (Fructus et al., 2009) occurred on the pycnocline (see figure 4 (b), (iv) & (v)). Note that the dynamics in the two cases were similar but the proximity of the ice to the pycnocline in case (a) impeded clear billow formation.

3.3 Dissipation of Turbulent Kinetic Energy Under the Ice

The three different ice types had different topographic features and associated roughness on their undersides. For example, the grease ice was lumpy in structure and the nilas ice grew columnar ice crystals down into the water, whereas the underside of the level ice was flat and relatively smooth. In figure 5, fields of horizontal velocity, u, vorticity, ω, dissipation of TKE, ϵ, and profiles of horizontally averaged TKE dissipation, $\bar{\epsilon}$, are presented for experiments with comparable wave amplitudes and wave shape but under varying ice types. The velocity and vorticity field were computed via PIV. TKE dissipation was computed from the PIV velocity fields following the direct gradient method given in Doron, Bertuccioli, Katz, and Osborn (2001), using 64×64 point, 75% overlapping, estimation windows. This estimate of dissipation requires the assumption that cross-plane velocity gradients have the same average magnitude as the measured in-plane velocity gradients. This estimate also assumes that the spatial resolution of the measured velocity fields is smaller than, or comparable to, the Kolmogorv scale, which was true in all cases presented here. The mean TKE dissipation, $\bar{\epsilon}$, was calculated by horizontally averaging the dissipation at every depth bin over the region plotted (minus a small section of width 0.1m at each side).

The approximate location of the underside of the ice and the top of the pycnocline were attained by analysing regions of light intensity in the experimental images. The horizontal velocity plots confirm the findings in §3.1 i.e. the grease ice and level ice were transported horizontally by the wave-induced flow. Enhanced vorticity is seen under the nilas ice when compared with the other cases. This is presumably because the nilas ice was stationary and hence the velocity gradient close to the ice, du/dz, is biggest in this case. Moreover, the vorticity plots show that the vertical extent of the vorticity layer directly under the ice is larger in the nilas and grease cases than the level case. This can be attributed to the relatively larger roughness of the grease ice and nilas ice compared to the level ice. Indeed in these cases, small vortices were seen to form at the underside of the ice downstream of any rough features. The TKE dissipation plots show (in all 3 cases) that dissipation levels under the ice are comparable with dissipation levels at the pycnocline, suggesting a mechanism for increased IW dissipation and consequently reduced propagation in the ice-covered Arctic Ocean. The mean TKE dissipation profiles confirm that there is a peak in the TKE dissipation below the ice and at the pycnocline. Integration of the TKE dissipation over these two regions of enhanced TKE dissipation give the near-ice dissipation as $77\%, 64\%$, and 54% of the pycnocline value in the grease, nilas, and level ice cases respectively. Or in other words, dissipation near the ice accounts for approximately $44\%, 39\%$, and 35%, of the total dissipation due to turbulence in the grease, nilas, and level ice cases respectively. The elevated TKE dissipation under the nilas and grease ice hypothesize that a rougher ice bottom and increased vertical shear of horizontal velocity lead to higher TKE dissipation in the upper boundary layer. Note that the TKE dissipation measures presented here are true for the experimental conditions tested, whether similar characteristics are seen in the field remains an open question.

4 Relevance to the Field

The interaction of ISWs with an ice edge are expected to be an effective mechanism of dissipating ISW energy and causing localised turbulence and mixing in the field. Similar dissipation is expected beneath any structures (e.g. ice ridge keels or rafted ice) that protrude into the pycnocline. In the field, vortex formation at the underside of the ice surface is expected to contribute further to dissipation of kinetic energy in a similar way to that found at the bottom boundary of the ocean.

The parameter space considered was limited by physical constraints. It differed from the ocean in both Reynolds number, $Re = ac/\nu$ (where ν is kinematic viscosity) and

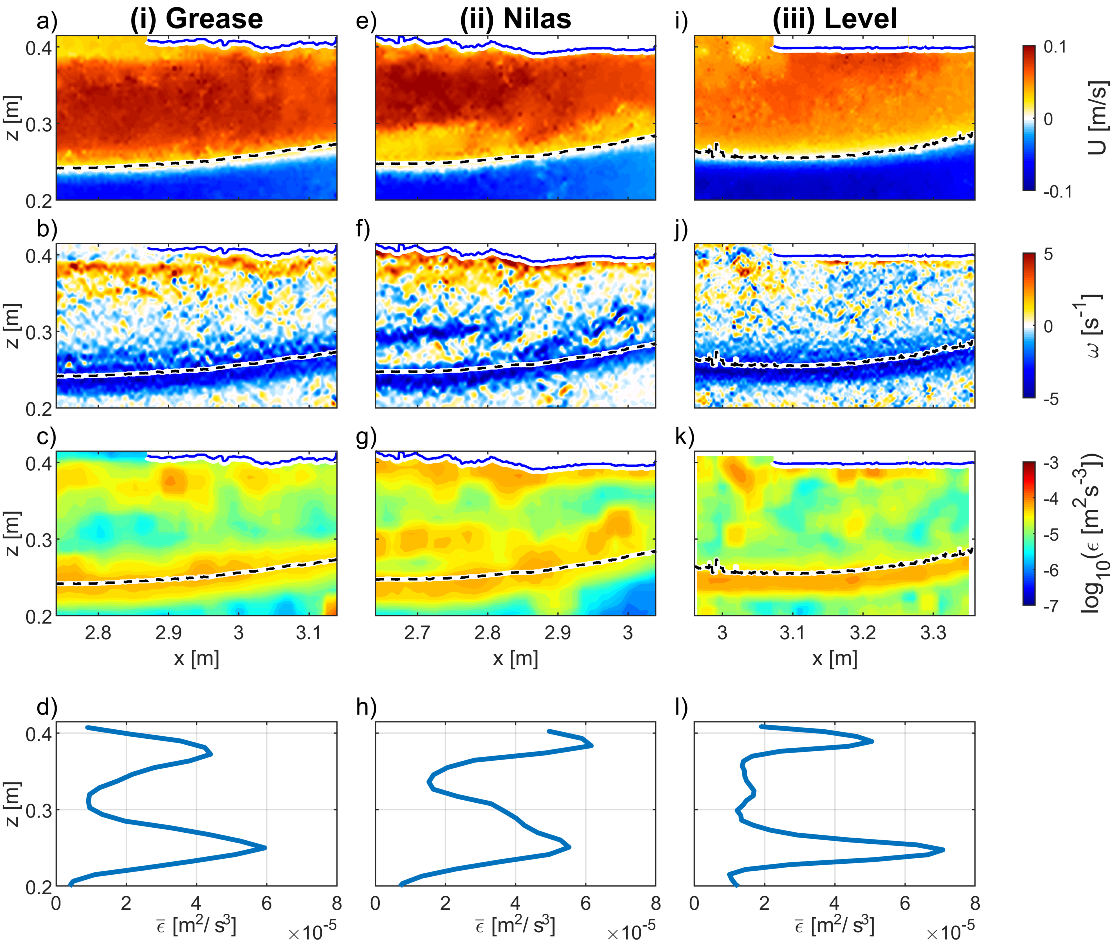

Figure 5. PIV measurements of (a, e, i) horizontal velocity u (ms^{-1}), (b, f, j) vorticity ω (s^{-1}), and corresponding calculations of (c, g, k) TKE dissipation $\log_{10}(\epsilon\ (\mathrm{m^2 s^{-3}}))$, and (d, h, l) mean TKE dissipation $\bar{\epsilon}$ (m^{2}s^{-3}) under (i) grease ice (a-d), (ii) nilas ice (e-h) and (iii) level ice (i-l). The black dashed line and the solid blue line depict the top of the pycnocline and the underside of the ice respectively.

buoyancy frequency

$$N = \sqrt{\frac{-g}{\rho_0}\frac{\partial \rho}{\partial z}}, \tag{1}$$

where g is acceleration due to gravity and ρ_0 is some reference density. Re and N in the laboratory were of the order 10^3 and 1 s^{-1} respectively. In the Arctic Ocean, when stratification is at its strongest, corresponding values are of the order 10^6 and 0.01 s^{-1}, respectively. Hence, to understand whether the laboratory results scale to the ocean requires variation of Re and N in the future. Moreover, study over a wider range of d_f/h_1 and $l_f/2\lambda$ would provide better insight into field scale dynamics.

The wave phase speed, c, in the experiments was $O(0.1)$ ms^{-1} and is comparable to field values which are typically 0.2−0.5 ms^{-1} in the Arctic Ocean. The experiments were conducted in a flume at rest, hence an inference regarding Froude number, Fr, (ratio of inertia to the external gravity field) is not relevant. The experiments can, however, be compared to nonlinear lee waves in the field (Rippeth et al., 2017) once the initially trapped and steepened wave (which has a $Fr \approx 1$) is released and allowed to propagate in sub-critical conditions ($Fr < 1$).

5 Conclusion

The ISW-induced current was shown to transport ice horizontally and the transport speed of relatively long ice floes was linearly dependent on the floe length for the parameter range considered. When the ice protruded into the pycnocline, interaction of the wave with the ice edge resulted in a Kelvin-Helmholtz billow forming on the pycnocline and the resulting wave signal being deformed or destroyed. If the underside of the ice was rough, boundary layer development due to the ISW-induced flow was observed and finally it was shown that levels of TKE dissipation under the ice were comparable to those at the pycnocline. The parameter range investigated was limited by the scale of the facility but nevertheless the observations show that the physical interactions of ISWs with ice could have important implications for dissipation of wave energy, and consequent mixing in the polar oceans and point to the need for further more detailed investigations. Moreover, the observation that an ISW can transport ice horizontally suggests that a series of ISWs may well constitute a mechanism by which sea ice banding can occur as proposed theoretically by Saiki and Mitsudera (2016).

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