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Wave Attenuation by Sea Ice Turbulence

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Key Points:

- Turbulence may be an important dissipative mechanism of wave energy in the MIZ
- Turbulence-induced attenuation coefficient can be determined from measurements of wave energy and turbulence at a single location
- Turbulence dissipation is parameterized through characteristic wave and ice properties

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Abstract

The dissipation of wave energy in the Marginal Ice Zone (MIZ) is often attributed to wave scattering and the dissipative mechanisms associated with the ice layer. In this study we present observations indicating that turbulence generated by the differential velocity between the sea ice cover and the orbital wave motion may be an important dissipative mechanism of wave energy. Through field measurements of under-ice turbulence dissipation rates in pancake and frazil ice, it is shown that turbulence induced wave attenuation coefficients are in agreement with observed wave attenuation in the MIZ. The results suggest that the turbulence-induced attenuation rates can be parameterized by the characteristic wave properties and a coefficient. The coefficient is determined by the ice layer properties.

1 Introduction

Ocean waves can penetrate hundreds of kilometers into vast sea ice covers before the ice fully attenuates their energy (Kohout, Williams, Dean, & Meylan, 2014; Wadhams, Squire, Goodman, Cowan, & Moore, 1988). Along the way, these waves impose stresses on the sea ice, enabling the waves to shape the region between the ice pack (referred to as the Marginal Ice Zone, MIZ) by breaking-up, moving and melting the ice (e.g. Squire, Dugan, Wadhams, Rottier, & Liu, 1995). Ocean surface waves are therefore expected to accelerate Arctic ice retreat in the near future due to seasonal opening of the Arctic seas (Q. Liu, Babanin, Zieger, Young, & Guan, 2016; Thomson & Rogers, 2014) and, with loss of the seasonal MIZ around Antarctica, ocean swell may contribute to the disintegration of the Antarctic ice shelves (Massom et al., 2018). To include these effects in weather and climate forecasting models, spectral wave models require process-based parameterizations of wave attenuation to predict the wave field transformation in the MIZ. The physical processes that dictate the dissipation of wave energy by sea ice are, however, still under debate.

Various processes are currently identified as contributors to the observed wave attenuation in the MIZ and are parameterized in spectral models, including wave scattering (e.g. Montiel, Squire, & Bennetts, 2016; Wadhams et al., 1988); ice layer interactions, where the ice is often parameterized as a viscoelastic layer (Mosig, Montiel, & Squire, 2015; Wang & Shen, 2010); and under-ice turbulence (e.g. A. Liu & Mollo-Christensen, 1988). The first is not a dissipative process as the reflection of waves by solitary ice floes merely alters the direction of wave energy, such that part of the energy is scattered back into the open ocean while the rest is transmitted further into the MIZ. Complex dissipation processes associated with the ice layer are typically parameterized through viscoelastic theories, where dissipation can be regulated through the mechanical properties of the ice layer model. Models for wave attenuation based on parameterization of wave scattering or dissipation by representing the sea ice cover as a viscoelastic layer have been able to explain, in part, wave observations in the MIZ. However, as they often require careful calibration of the modeled ice layer characteristics, the properties of which vary greatly in both space and time, our predictive abilities are severely restricted without advancement in the physical description of processes that actually drive wave attenuation in the MIZ. Additionally, recent experimental observations in the laboratory, where measured material properties of the ice layer were used as input for the viscoelastic model, revealed large discrepancies between the measured wave attenuation and those predicted by parameterizing the ice cover as a viscoelastic layer (Sree, Law, & Shen, 2018).

Under-ice turbulence has remained a relatively unexplored dissipative mechanism since the study of A. Liu and Mollo-Christensen (1988) (with some exceptions, such as Ardhuin, Sutherland, Doble, and Wadhams (2016); Kohout, Meylan, and Plew (2011); Shen and Squire (1998)). Waves receive energy by wind, while they lose energy through the production of turbulence. Turbulence is generated through boundary layer develop-

66 ment under the ice, wake formation within the ice-layer around solitary ice floes (or form
 67 drag, e.g. Kohout et al., 2011) and ice-floe collisions (Rabault, Sutherland, Jensen, Chris-
 68 tensen, & Marchenko, 2019). Through order of magnitude estimates (A. Liu & Mollo-
 69 Christensen, 1988) and spectral model calibration (Ardhuin et al., 2016), the generation
 70 of under-ice turbulence shows reasonable agreement with in-situ wave attenuation ob-
 71 servations, or is shown to be much larger than estimates of energy dissipation by other
 72 attenuation processes, including scattering (Shen & Squire, 1998). The few turbulence
 73 estimates presently available are, however, not substantiated by any turbulence measure-
 74 ments in the field as measuring turbulent properties in these extreme environments is
 75 a major challenge. Hence, it remains uncertain whether turbulence is an important con-
 76 tributor to the attenuation of wave energy and should be included in spectral wave mod-
 77 els.

78 The objective of this study is to assess the importance of under-ice turbulence in
 79 attenuating wave energy in the MIZ. Here, in-situ measurements of the turbulence dis-
 80 sipation rates from the Arctic Sea State Program are used to estimate turbulence-induced
 81 wave attenuation under mixtures of pancake and frazil ice and are compared against mea-
 82 surements of wave attenuation.

83 2 Turbulence-Induced Wave Attenuation Coefficient

84 Wave energy in the MIZ is known to decrease exponentially with distance into the
 85 ice cover (Wadhams et al., 1988):

$$86 \quad S(f, x + \Delta x) = S(f, x) \exp(-\alpha \Delta x) \quad (1)$$

87 where $x + \Delta x$ (m) is the distance in the direction of wave propagation relative to an ar-
 88 bitrary point x within the MIZ, $S(f, x)$ is the spectral energy density at the position x
 89 and α (1/m) is the frequency dependent wave attenuation coefficient. When the verti-
 90 cal structure of the turbulent kinetic energy (TKE) dissipation rates ε is known, the to-
 91 tal dissipation of TKE D (W/m²) is determined by:

$$92 \quad D = \int_{-\delta}^0 \rho \varepsilon dz \quad (2)$$

93 where ε is the TKE dissipation rate of turbulence generated by wave-ice interactions,
 94 δ is the thickness of the turbulent boundary layer, ρ is the density of the ocean water
 95 and z is the distance from the sea ice interface. If the wave attenuation is dominated by
 96 the turbulent shear stress between the sea ice interface and the orbital motion of the fluid,
 97 the change of wave energy in the direction of wave propagation is:

$$98 \quad E(x + dx) - E(x) = -Ddt = -Ddx/c_g \quad (3)$$

99 where c_g is the group velocity of a characteristic wave period, $E(x) = \rho g \int S(f, x) df$
 100 (Ws/m²) is the total wave energy at a distance x . Note that as a first order approxima-
 101 tion, wind input is ignored as source of wave energy in Eq. 3. Total wave energy (instead
 102 of the spectral energy) is considered here as the wave frequency dependence of ε cannot
 103 be determined from the turbulence measurements by itself, as all turbulence generated
 104 below the sea ice interface is subjected to the turbulence energy cascade.

105 Under the assumption that most of the measured TKE dissipation rate originates
 106 from a narrow range of frequencies, Eqs 1 and 3 can be combined to yield:

$$107 \quad -D/c_g = -\alpha_t E(x) \exp(-\alpha_t \Delta x) \quad (4)$$

108 where the subscript ‘t’ in α_t refers to the turbulence-induced attenuation of wave energy.
 109 As α_t corresponds to the attenuation of the total wave energy, α_t is here simply a func-
 110 tion of the characteristic frequency of the spectral energy density $S(f, x)$, for example,

111 the mean or peak frequency. If both turbulence and waves are then measured at the same
 112 location (i.e. $\Delta x = 0$), it follows from Eq. 4 that the turbulence-induced wave attenu-
 113 ation coefficient can be determined by:

$$114 \quad \alpha_t = \frac{D}{c_g E(x)} \quad (5)$$

115 It follows from Eq. 5, that concurrent measurements of turbulence and waves are required
 116 to determine the turbulence-induced attenuation coefficient. It should be stressed that
 117 ε (and consequently α_t) is critically defined here as the TKE dissipation rate of turbu-
 118 lence generated through wave-ice interactions only.

119 Hereafter, the mean wave period $T_{m01} = m_0/m_1$ (where m_0 and m_1 are the ze-
 120 roth and first-order moment of the energy spectrum) is the characteristic wave period
 121 used to determine c_g . As the open water dispersion relation is likely to hold for waves
 122 in the MIZ with periods in the range of 3–10 s (Collins, Doble, Lund, & Smith, 2018),
 123 a linear paradigm is used to model the wave propagation speed.

124 3 Methods

125 The data used in this study were obtained during the Arctic Sea State Program
 126 in the Beaufort and Chukchi Seas. The measurement campaign included arrays of wave
 127 buoys deployed in frazil and/or pancake ice during 7 Wave Experiments (WE) from Oc-
 128 tober to November 2015. A summary of the Sea State Program and measurement cam-
 129 paign can be found in Thomson (2015) and Thomson et al. (2018). Here, we focus on
 130 a brief description of the SWIFT drifters and the processing of data.

131 SWIFT is a wave following drifting buoy which simultaneously measures the ocean
 132 surface motion and the under-ice turbulent velocity components. Part of the drifters were
 133 mounted with an acoustic Doppler current profiler (ADCP, Nortek Aquadopp HR) in
 134 upward direction to measure profiles of the turbulent velocity fluctuations within 1 m
 135 from the surface. After quality control measures, including ice-masking, the measured
 136 turbulent velocity components of the drifters were used to estimate profiles of the tur-
 137 bulance dissipation rate through a second-order structure function, as per Smith and Thom-
 138 son (2019). This method has been validated against independent dissipation rate mea-
 139 surements from a second instrument (single-point acoustic Doppler velocimeter, ADV)
 140 mounted on a SWIFT drifter, where dissipation rates were estimated by fitting Kolmogorov's
 141 $-5/3$ law to the inertial subrange of the measured velocity spectrum (Thomson, 2012).
 142 The reader is referred to Smith and Thomson (2019) for a detailed description on the
 143 processing of the data obtained by the SWIFT buoys.

144 The SWIFT buoys equipped with upward looking ADCP's were deployed in tan-
 145 dem with SWIFT buoys with a downward looking ADCP, measuring velocity profiles from
 146 1.5 to 21 m below the ocean surface. By collocating the upward and downward looking
 147 ADCP's, the mean relative velocity between the ice and the ocean ΔU is taken as the
 148 velocity in the upper bin of the downward looking ADCP (Smith & Thomson, 2019) and
 149 will be used to exclude mean shear between the ice and the upper ocean (drift) as a sig-
 150 nificant source of turbulence.

151 As the SWIFT buoys were not continuously drifting in the ice covers, only parts
 152 of the wave experiments are considered here. Based on the images captured by the cam-
 153 era mounted on the SWIFT buoys, the buoys were within the ice covers on: WE3, 10-
 154 13 Oct. (SWIFT 9, 11, 14 and 15); WE4, 17-18 Oct. (SWIFT 11, 14 and 15); WE6, 23-
 155 24 Oct. (SWIFT 9, 11 and 12); and WE7, 31 Oct-1 Nov (SWIFT 9, 11, 13 and 15). A
 156 summary of the wave and ice conditions during these deployments is provided in Table
 157 S1.

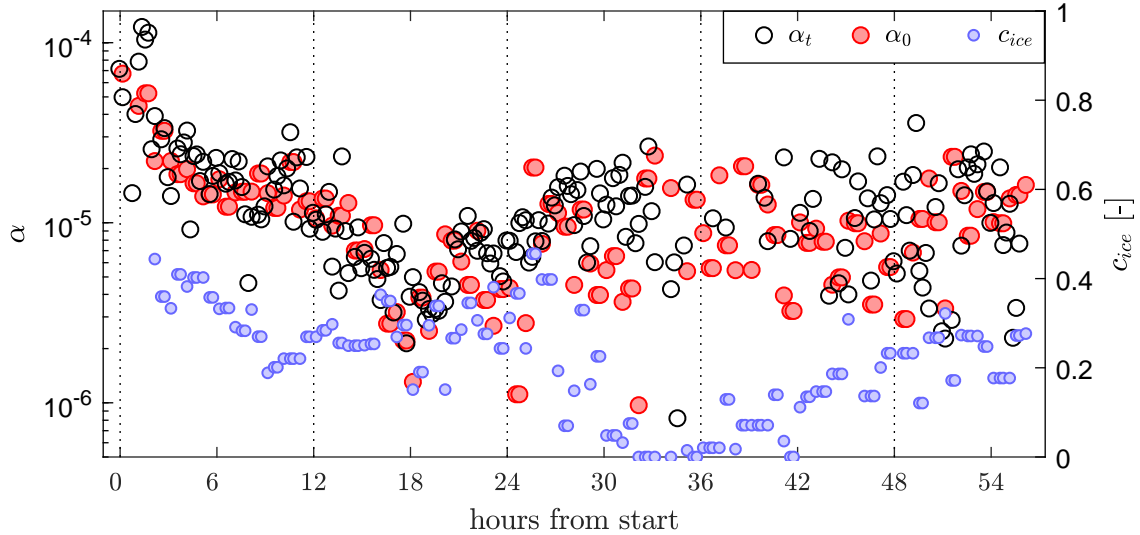


Figure 1. Time series of the turbulence-induced wave attenuation α_t (black), the observed total wave attenuation α_0 (red) and ice concentration c_{ice} (blue) during the deployment of SWIFT 14, WE3, on 11-13 October. Note that c_{ice} (obtained from AMSR2) does not always agree to visual observations of ice concentrations by camera images.

By deploying multiple SWIFT buoys at different positions within the MIZ, the total wave attenuation α_0 is determined by evaluating the total wave energy decay between buoy pairs:

$$\alpha_0 = \frac{1}{\Delta x} \ln \left(\frac{\int S(f, x) df}{\int S(f, x + \Delta x) df} \right) \quad (6)$$

where $\Delta x = \Delta x_\theta \cos(\theta_m - \theta)$ is the distance between the buoy pair along the mean direction of the wave field, Δx_θ is the great-circle distance between the buoys, θ_m is the mean wave direction and θ is the bearing angle of the buoy pair. Quality control criteria for α_0 from Cheng et al. (2017) were adopted, though in this study the maximum allowable angle between θ_m and θ is set to 70° . Additionally, a minimum buoy pair distance of $\Delta x_\theta = 250$ m is enforced to ensure energy dissipation across the full spectrum can be reasonably measured. Only buoy pairs containing two SWIFT drifters and at least one SWIFT drifter with an upward looking ADCP were used to determine α_0 . When multiple observations of α_0 were available at one instance, α_0 is taken as the mean of the logarithms of these observations, as equally valid measurements can be more than one order of magnitude apart. Note that, similarly as in Eq. 3, we neglect wind input as possible source of wave energy in Eq. 6. The ice concentration c_{ice} during the wave experiments are approximated using AMSR2 (Spren, Kaleschke, & Heygster, 2008).

4 Results and Discussion

4.1 Example time series of α_t and α_0

Figure 1 shows a comparison of observed total wave attenuation α_0 and estimated turbulence-induced wave attenuation α_t for a deployment during WE3 on 11 October. Close agreement between α_t and α_0 is observed up to $t = 24$ hours from the start ($r^2 = 0.80$, in contrast, $r^2 = 0.54$ for the full time series). Beyond $t = 24$ h, both α_t and α_0 vary considerably in time but remain, nevertheless, similar in order of magnitude. Although correlation between α_t and c_{ice} is weak for the time series, comparable trends can be ob-

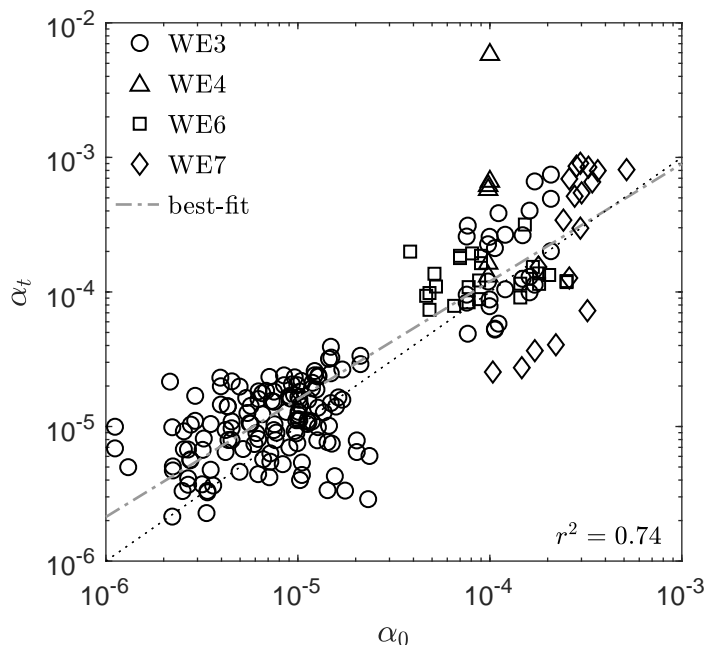


Figure 2. Comparison of the turbulence-induced wave attenuation α_t (Eq. 5) against the total wave attenuation α_0 (based on the mean of buoy-pair observations). Markers identify the different wave experiments. Best-fit to the data is given by the dash-dotted line.

183 served during the first 12 hours, where an initial decrease in wave attenuation corresponds
 184 to a decrease in ice concentration, after which both α_t and c_{ice} seem to increase till around
 185 24 hours from the start. Estimates of ice concentrations after $t = 24$ h do not follow
 186 trends of α_t , however, such estimates should be interpreted with caution, as ice cover het-
 187 erogeneity can occur at scales smaller than the resolution of AMSR2. For instance, while
 188 no ice is present around $t = 36$ h according to AMSR2 approximations (see also Fig-
 189 ure S1), images captured by the SWIFT buoy suggest a mixture of frazil and pancake
 190 ice to be present from $t = 29$ h till $t = 40$ h (see Figure S2 for images captured dur-
 191 ing the deployment). Although daylight limits observations of conditions throughout the
 192 entire deployment, based on daytime images, it is hypothesized that ice conditions at the
 193 air-ice interface remain relatively constant after $t = 24$ h. This would be consistent with
 194 the trend of α_t during this time. The correlation between α_t and α_0 seen for this deploy-
 195 ment demonstrates that turbulence generated through wave-ice interactions can explain
 196 wave energy dissipation in the MIZ.

197 4.2 Overall comparison of α_t and α_0

198 The total wave attenuation is then compared against the turbulence-induced wave
 199 attenuation for all wave experiments over half-hour periods (Figure 4). Good agreement
 200 is observed across the wave experiments and over a wide range of wave attenuation co-
 201 efficients ($r^2 = 0.74$). The results include a correlation associated with scaling both axes
 202 by wave energy; however, this correlation only explains 19% of the variance of the sig-
 203 nal. Thus, it suggest that turbulence-induced wave attenuation may be an important dis-
 204 sipation mechanism of wave energy in the MIZ.

205 Central in this study is the assumption that the dominant source of measured tur-
 206 bulence under the ice is from wave-ice interactions. A few physical processes could be

responsible for the production of TKE below the ice layer, including turbulence generation by ice floe collisions (which can lead to jet-like injections of fluid into the ice layer), overwash, wake flow around ice formations, keel ridges, TKE production by the mean shear (drift) at the sea ice interface and by wind related processes. Note that wind can only impact the production of turbulence indirectly through drift currents and wind generated waves. As the TKE dissipation rate D scales with the cube of the characteristic velocity scale of the dominant physical process, turbulence input through wind-induced wave generation are expected to scale with the wind speed $D \propto U_{10}^3$, while turbulence production by the mean shear (including wind-induced drift) scales with the differential velocity between the ice and the upper ocean, i.e. $D \propto \Delta U^3$. For the current field experiments, there is no correlation between D and ΔU ($r^2 = 0.05$), and limited correlation between D and wind speed ($r^2 = 0.32$). Note that this correlation could be spurious, as the wind field is intrinsically linked to the waves and, as a result, correlated to wave-ice interactions as well, implying that wave-ice interactions dominate turbulence production in this study. This is consistent with the strong correlation observed between α_0 and α_t (i.e. Figures 1 and 4), as only turbulence generated through wave-ice interaction processes can contribute to wave attenuation in the MIZ. Hence, turbulence induced wave attenuation could be an important dissipative process of wave energy in the MIZ.

Note that observations of α_t can exceed the measured wave attenuation (i.e. by up to a factor of two for very low wave attenuation rates, see best-fit to the data in Figure 4), which implies that other processes and/or energy sources are present. Due to the lack of correlation between D and ΔU , the most likely source of wave energy or turbulence is by local wind input (Smith & Thomson, 2019; Zippel & Thomson, 2016), a source term that is neglected in this study to a first order approximation in both α_t and α_0 (i.e. Eqs. 3 and 6). In particular, Li et al. (2017) determined that wave energy input through wind in the MIZ should be considered and observed that the rate of energy transfer is dependent on the strength of the wind field.

4.3 Modeling turbulence dissipation under sea ice

To estimate the dissipation of turbulence within the wave boundary layer (WBL), the presence of a balance between TKE production (P) and TKE dissipation is adopted (e.g. Tennekes & Lumley, 1972):

$$\varepsilon \approx P \approx -\overline{u'w'} \frac{d\bar{u}}{dz} \quad (7)$$

where $\overline{u'w'}$ is the ensemble averaged Reynolds stress and $d\bar{u}/dz$ the vertical mean velocity gradient. The Reynolds stress is approximated by $-\overline{u'w'} \approx u_*^2$ with u_* being the shear velocity, while the velocity profile in the WBL is simplified as linear such that $d\bar{u}/dz \approx u_{orb}/\delta$, where $u_{orb} = \pi H_{m0}/T_{m01}$ is the representative wave orbital velocity and δ is the thickness of the WBL. By substitution of the preceding into Eq. 7 and integration of the turbulence dissipation rate over the WBL thickness, the dissipation rate of TKE per square meter surface area can then be estimated by:

$$D_{WBL} \approx b_1 \int_{-\delta}^0 -\rho u_*^2 \frac{u_{orb}}{\delta} dz = \rho u_*^2 u_{orb} \quad (8)$$

where b_1 is a constant of proportionality. Using the analogy between a sea ice interface and a wave bottom boundary layer, the wave friction velocity is considered to be proportional to the wave orbital velocity (e.g. Madsen, Poon, & Graber, 1989). Thus, the dissipation of TKE (Eq. 8) can be interpreted as the product of the interfacial stress ($u_*^2 = C_D u_{orb}^2$) and the characteristic velocity scale u_{orb} (e.g. Smith & Thomson, 2019). The generation of turbulence by the differential velocity of the orbital wave motion and the ice layer is then proportional to u_{orb}^3 . Combining the drag coefficient C_D and constant

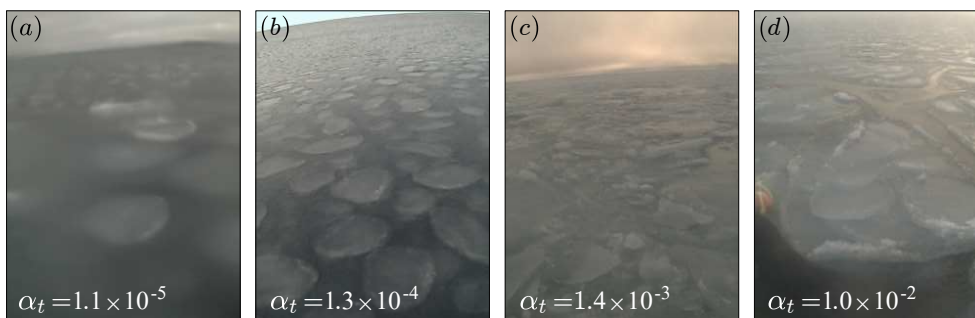


Figure 3. Images taken by the SWIFT drifters during deployment for varying ice conditions and measured turbulence induced attenuation rates: (a) WE3 SWIFT 14; (b) WE3 SWIFT 11; (c) WE6 SWIFT 12; (d) WE4 SWIFT 11.

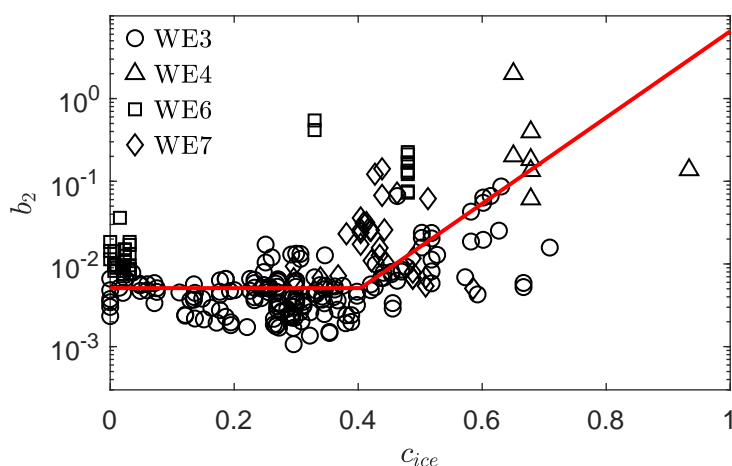


Figure 4. Variation of coefficient b_2 (see Eq. 9) with ice concentration c_{ice} for all wave experiments. For $c_{ice} < 0.4$, $b_2 = 3.6 \times 10^{-4}$, and for $c_{ice} \geq 0.4$, $b_2 = 1.0 \times 10^{-7} \exp(20c_{ice})$.

b_1 into a new coefficient b_2 , b_2 can be defined as:

$$D_{WBL} = \rho b_2 u_{orb}^3 = \rho b_2 \left(\frac{\pi H_{m0}}{T_{m01}} \right)^3 \quad (9)$$

The coefficient b_2 can be interpreted as the ratio of TKE dissipation rate to the kinetic energy of the local wave state and, hence, represents the relative dissipation rate of TKE.

While the orbital velocity is a reasonable approximation of the velocity scale that characterizes under-ice turbulence, the fundamental velocity scale that defines D_{WBL} is the differential velocity between the ice and the wave orbital velocity. Any deviation of the orbital wave motion by wave-following sea ice (for instance, as is often the case with pancake ice) is therefore embedded in the coefficient b_2 . Thus, b_2 is not only expected to be a function of ice roughness, but also of wave and other ice properties. In the case of loose ice, this includes the draft of the ice floes (or ice thickness), their diameter and the ice type. In particular, Rogers et al. (2016) found that wave dissipation rate α_0 sorted well by ice type. Similarly, we observed a higher turbulence-induced attenuation rate α_t in more consolidated pancake ice (Figure 3). As ice concentration is the most easily measurable characteristic of sea ice, the variation of b_2 is compared against the ice concentration in Figure 4. The results suggest that for $c_{ice} < 0.4$ the dissipation rates of TKE

271 become nearly independent of c_{ice} . Above $c_{ice} = 0.4$, however, dissipation rates tend
 272 to increase with ice concentration. Estimates of the coefficient b_2 in this study correspond
 273 well to equivalent coefficients observed by others. For instance, Lu, Li, Cheng, and Leppäranta
 274 (2011) report ice-ocean drag coefficients ranging from 1×10^{-4} to 5×10^{-2} for a broad
 275 variety of sea ice conditions. Additionally, Gallaher et al. (2016) found ice drag coeffi-
 276 cients in the range of 3×10^{-3} to 6×10^{-3} for first-year ice in the Arctic MIZ.

277 While ice concentration might be a good representative measure for the dominant
 278 roughness length scale for small ice concentrations, other roughness scales are likely to
 279 dominate friction when the sea ice cover turns more solid. Thus, the scatter in b_2 in Fig-
 280 ure 4 is likely a result of variation in ice type and thickness. It should be noted that the
 281 ice concentrations obtained by AMSR2 are spatially averaged estimates of the local ice
 282 conditions. In particular, due to the resolution of AMSR2, local ice conditions can de-
 283 viate significantly from those obtained of AMSR2. For instance, based on the estimates
 284 of AMSR2 for the two outliers during WE6 (see two squares in Figure 4), the ice con-
 285 centration was estimated to be 0.33, while images captured at these instant suggest con-
 286 siderably higher ice concentrations (see corresponding ice cover in Figure 3c).

287 Although the results suggest that turbulence induced by under-ice friction becomes
 288 relevant for ice concentration above 0.4 only, production of turbulence for lower ice con-
 289 centration cannot necessarily be ignored. Specifically, the observed value of $b_2 \approx 4 \times$
 290 10^{-4} corresponds well to the magnitude of turbulence in open water swell seas, i.e. $b_2 \approx$
 291 7×10^{-4} (Babanin, 2012). As b_2 represents the relative dissipation rate of TKE (i.e. see
 292 Eq. 9), the relative dissipation rate of TKE under sea ice covers for $c_{ice} > 0.4$ can be
 293 larger than that observed in open water. This also substantiates that there are no other
 294 significant sources of turbulence in this study, as they would have led to b_2 being larger
 295 than that observed for swell.

296 The results of this study, therefore, imply that turbulence generated by the differ-
 297 ential velocity between the orbital wave motion and the ice layer may be an important
 298 dissipative mechanism for wave energy in the MIZ. Thus, turbulence should be added
 299 to the list of wave-ice interaction processes in spectral wave models. A simple relation
 300 is suggested here where the dissipation rate of TKE per square meter surface area can
 301 be estimated using Eq. 9, by the following model for b_2 :

$$\begin{aligned}
 b_2 &= 3.6 \times 10^{-4} \quad \text{for } c_{ice} < 0.4 \\
 b_2 &= 1.0 \times 10^{-7} \exp(20c_{ice}) \quad \text{for } c_{ice} \geq 0.4
 \end{aligned}
 \tag{10}$$

304 However, implementation of the above model into spectral wave models is complicated
 305 by the integral approach that was inevitably adopted to measure and determine the TKE
 306 dissipation rate, rather than a spectral solution to D , i.e. varying with wave frequency.

307 While the TKE dissipation rate in this study is parameterized using wave and ice
 308 properties, alternative parameterizations based on wind speed can reproduce the observed
 309 turbulence as well (e.g. Smith & Thomson, 2019). This is not surprising as wind and waves
 310 are inevitably correlated and is particularly true when processes are averaged over large
 311 spatial scales or are observed at the edge of the MIZ. Additional studies are therefore
 312 required to further elucidate the complex interaction between wind, waves and the ice
 313 and connect these bottom-up (in terms of the waves) and top-down (in terms of the wind)
 314 approaches to a unified framework of momentum and energy conservation across the air-
 315 ice-sea interface. An experimental laboratory study in the absence or presence of wind
 316 may progress our understanding of the interactions across the ice layer, including the scal-
 317 ing of wind-input, ice drift and the impact of ice-layer properties on the coefficient b_2 .
 318 Further efforts are required to define the wave-induced attenuation coefficient α_t as a func-
 319 tion of wave frequency, as it is well known that wave attenuation decreases with increas-
 320 ing wave period (e.g. Cheng et al., 2017; Meylan et al., 2018; Rogers et al., 2016; Wad-
 321 hams et al., 1988).

5 Conclusion

Through field measurements of turbulence dissipation rates under pancake and frazil ice covers, we show that turbulence generated by the differential velocity between the ice and the orbital wave motion may explain the observed attenuation of wave energy in the MIZ. The large variability of the attenuation rates is argued to be the result of temporal and spatially varying ice conditions. Our results suggest that turbulence-induced wave attenuation rates can be parameterized through characteristic wave properties and a coefficient (b_2) where b_2 remains constant for ice concentrations c_{ice} below 0.4 ($b_2 = 3.6 \times 10^{-4}$) and increases for $c_{ice} > 0.4$ as $b_2 = 1.0 \times 10^{-7} \exp(20c_{ice})$. More experiments are, however, required to quantify the coefficient b_2 in terms of ice layer properties and determine the dependence of the turbulence-induced wave attenuation coefficient on wave frequency.

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Figure 1.

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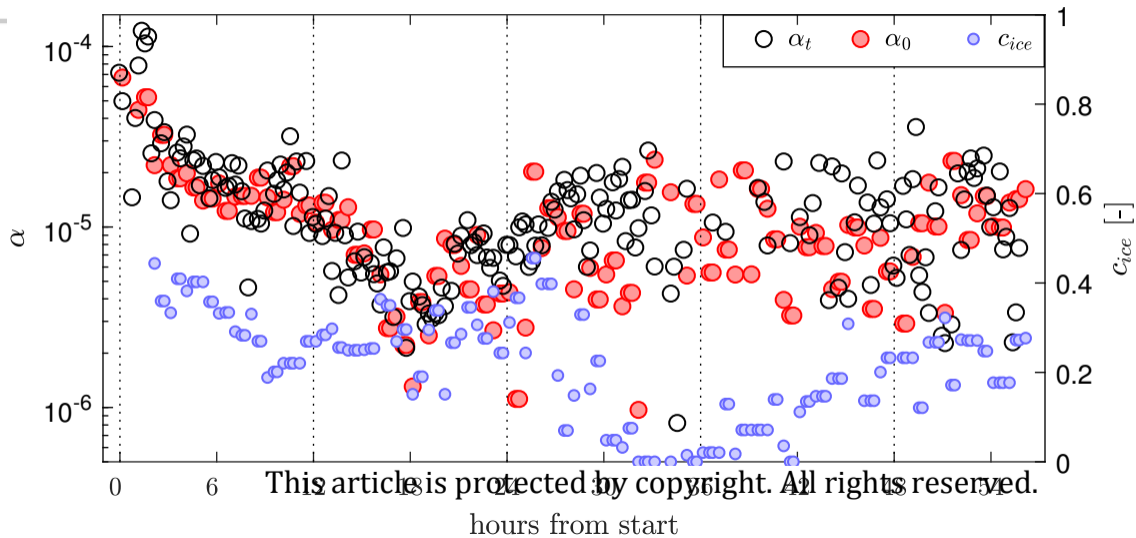


Figure 2.

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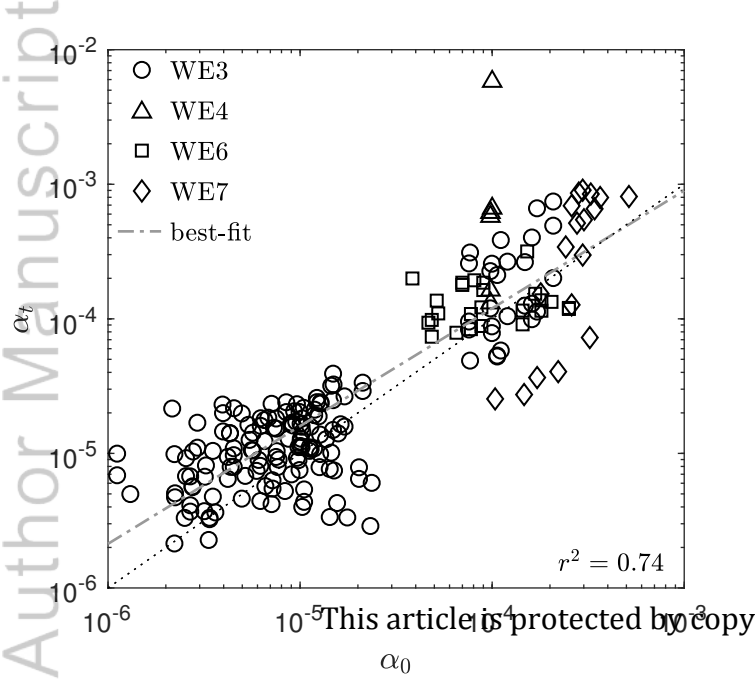


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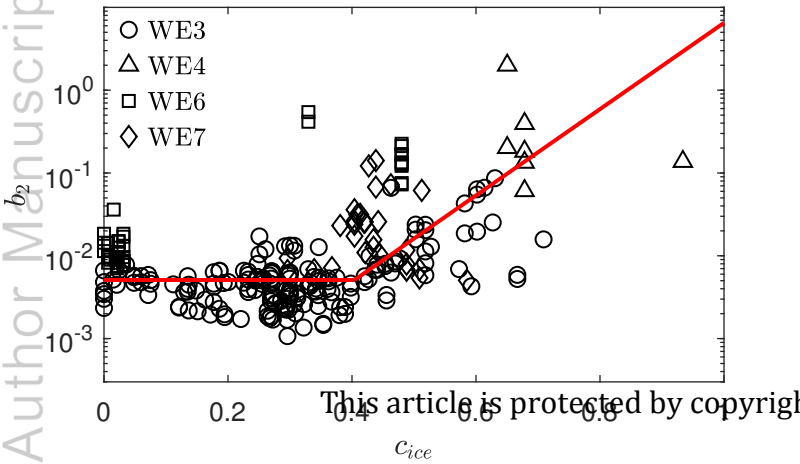
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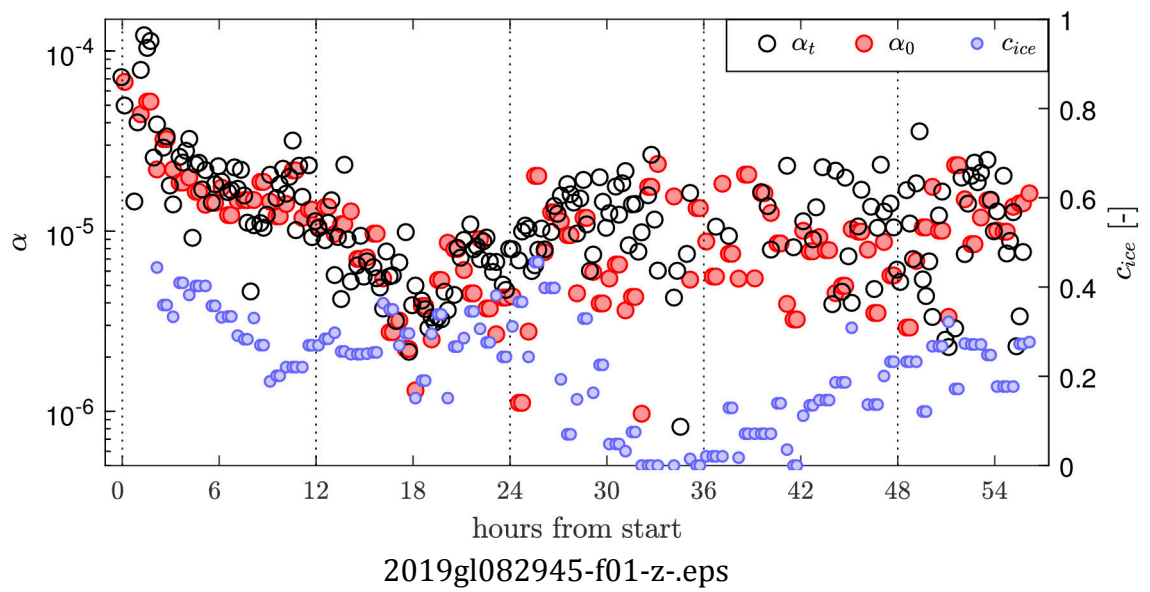


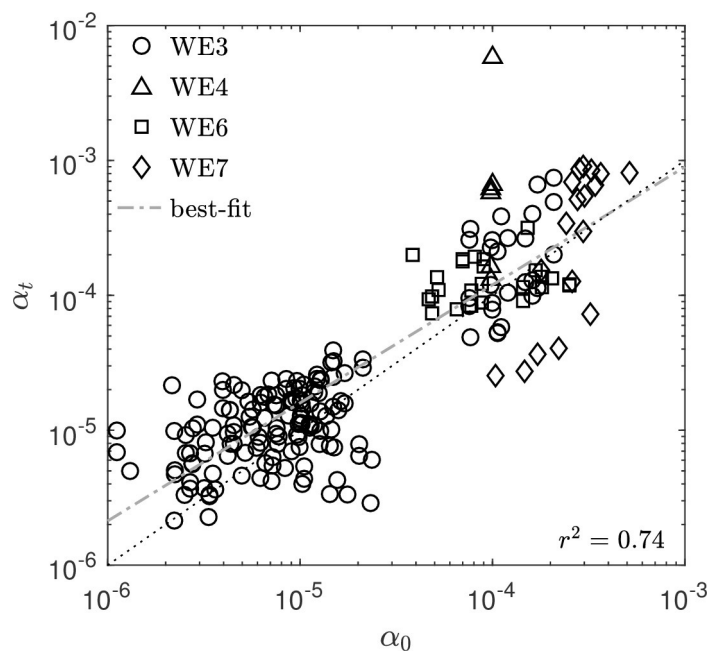
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Figure 4.

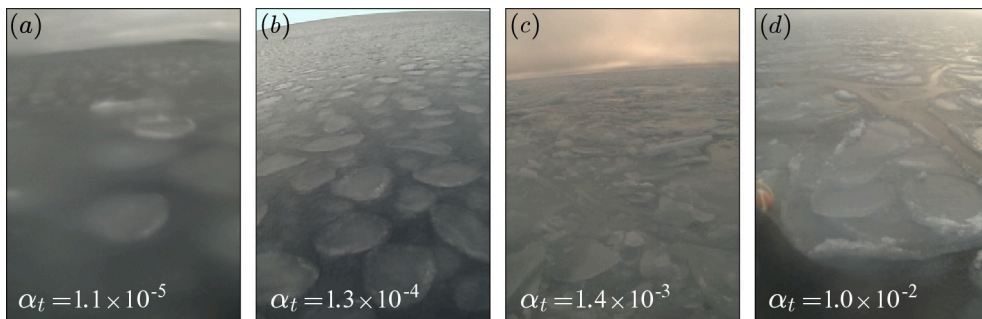
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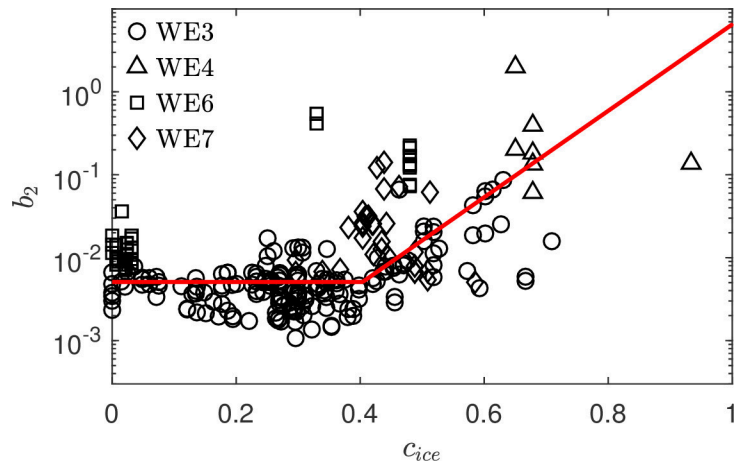




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