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► **To cite this version:**

Fabrice Ardhuin, Mark Otero, Sophia Merrifield, Antoine Grouazel, Eric Terrill. Ice Breakup Controls Dissipation of Wind Waves Across Southern Ocean Sea Ice. *Geophysical Research Letters*, 2020, 47 (13), 10.1029/2020GL087699 . hal-03094089

HAL Id: hal-03094089

<https://hal.science/hal-03094089>

Submitted on 4 Jan 2021

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1 **Ice break-up controls dissipation of wind-waves across**
2 **Southern Ocean sea ice**

3 **Fabrice Ardhuin¹, Mark Otero¹, Sophia Merrifield¹, Antoine Grouazel², and**
4 **Eric Terrill¹**

5 ¹Scripps Institution of Oceanography, La Jolla, California
6 ²Univ. Brest, CNRS, IRD, Ifremer, Laboratoire d'Océanographie Physique et Spatiale (LOPS), IUEM,
7 Brest, France

8 **Key Points:**

- 9 • Wind-waves attenuate across the Antarctic sea ice with a narrow directional
10 distribution
11 • Scattering of waves by ice floes plays a negligible role in wave attenuation
12 • Observed wave attenuation is consistent with ice break-up modulating the dis-
13 sipation strength

Corresponding author: Fabrice Ardhuin, ardhuin@ifremer.fr

Abstract

Sea ice inhibits the development of wind-generated surface gravity waves which are the dominant factor in upper ocean mixing and air-sea fluxes. In turn, sea ice properties are modified by wave action. Understanding the interaction of ice and waves is important for characterizing both air-sea interactions and sea ice dynamics. Current leading theory attributes wave attenuation primarily to scattering by ice floes. Here we use new *in situ* wave measurements to show that attenuation is dominated by dissipation with negligible effect by scattering. Time series of wave height in ice exhibit an “on/off” behavior that is consistent with switching between two states of sea ice; a relatively unbroken state associated with strong damping (off), possibly caused by ice flexure, and very weak attenuation (on) across sea ice that has been broken up by wave action.

Plain Language Summary

Waves created by wind at the ocean surface are strongly attenuated when they travel across ice-covered regions. Until now, this effect was thought to be the result of waves reflection off pieces of ice. Using new measurements of wave directions, we show that waves do not come from a broad range of directions, and scattering must be weak. Instead we find that attenuation is highly variable and related to the size of ice floes. We hypothesize that attenuation may be caused by cyclic deformation of the ice. When the waves are large enough to break the ice up, this deformation stops and the attenuation is much less. This finding is important for forecasting waves in ice-infested waters as well as predicting seasonal sea ice extent.

1 Introduction

Recent studies have revealed that ocean waves play a significant role in the ice momentum balance [Stopa *et al.*, 2018], ice extent [Kohout *et al.*, 2014], and the rapid thickening of ice when the ocean freezes [Sutherland and Dumont, 2018]. Previous wave measurements in ice-covered waters [Doble and Bidlot, 2013; Kohout *et al.*, 2014] have shown that waves in ice attenuate over longer distances when the open water wave height increases, often due to longer corresponding wave periods [Meylan *et al.*, 2014]. This result was used to define an *ad hoc* parameterization of wave attenuation [Kohout *et al.*, 2016] that fails to reproduce the large range of observed attenuations for any given off-ice wave height and wave period [Stopa *et al.*, 2018]. This poor performance calls for a detailed physical understanding of ice-wave interaction processes.

Following early works [Wadhams, 1988], the attenuation of waves that propagate under the ice is generally attributed to scattering [Squire, 2020]. Scattering is a partial reflection of waves at the boundaries of ice floes, broadening the distribution of wave directions. Other processes dissipate wave energy into heat and narrow the wave direction distribution around the shortest propagation path. Measured wave directions within the ice can thus reveal the importance of scattering. For example, the narrow directional spread of waves with periods 19 to 23 s measured with directional tilt-meters in the Arctic showed that scattering could not be a significant source of attenuation for these very long waves [Ardhuin *et al.*, 2016]. Here we generalize these observations to more typical wave periods from 10 to 20 s using wave buoy measurements. We evaluate two different dissipation processes: under-ice friction [Stopa *et al.*, 2016], and the anelastic dissipation associated with ice flexure [Cole *et al.*, 1998]. The latter process yields large dissipation rates when the ice is flexing. However, flexing only occurs for waves shorter than about twice the ice floe diameter as floes tilt over longer waves [Boutin *et al.*, 2018]. This varying behavior as a function of wavelength can explain the observed sudden drop in wave attenuation when ice is broken by waves [Collins *et al.*, 2015]. Here the floe sizes are constrained by Synthetic Aperture Radar (SAR) imagery.

64 2 Measured wave properties and sea ice conditions

65 Freely drifting wave buoys [Drazen *et al.*, 2016] were deployed in the Southern
 66 Ocean in 2018, and advected into advancing sea ice. Fig. 1.a shows the positions of
 67 three buoys numbered 623, 624 and 625. Successive positions of the ice edge, defined
 68 as the 15% ice concentration contour, are estimates from the Advanced Microwave
 Scanning Radiometer 2 (AMSR2) radiometer [Spren *et al.*, 2008].

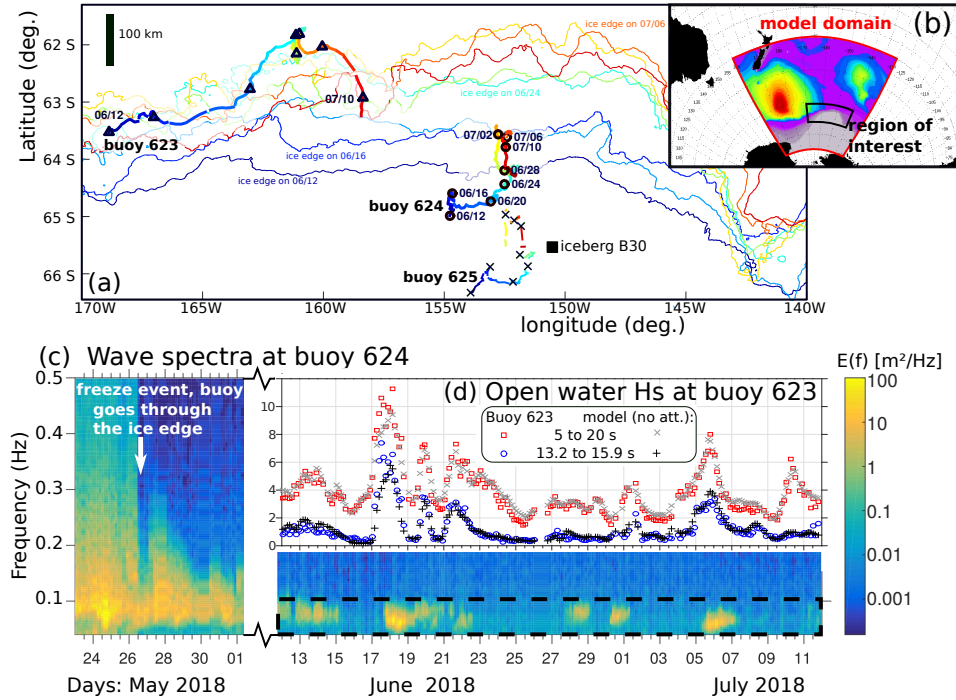


Figure 1. (a) Positions of the ice edge, defined as the 15% ice concentration contour, every 4 days from June 12 to July 10, 2018, and positions of three mini-buoys, number 623 (triangles in open water), 624 (circles in the ice) 625 (crosses, further in the ice). (b) Geographical context and domain of the numerical wave model used in section 3. The black rectangle corresponds to the region shown in (a), the colors show simulated significant wave heights from 0 to 10 m on July 4th at 12:00 UTC. (c) Wave energy spectra at buoy 624. (d) Time series of measured and modeled wave height at the open water buoy 623 for period ranges 5 to 20 s, and 13.2 to 15.9 s. See also supplementary video S1.

69

70 Buoy 624 measured a typical wave signal as shown in Fig. 1.c. High wave
 71 frequencies (above 0.3 Hz) vanish once the buoys is a few kilometers into the ice on May
 72 26. This attenuation is similar to other measurements obtained during the formation of
 73 meter-scale rafts of ice known as pancakes [Thomson *et al.*, 2018]. Frequencies between
 74 0.10 and 0.30 Hz gradually disappear as the buoy gets deeper into the ice, 80km from
 75 the ice edge on June 12. This evolution is presumably associated with the progressive
 76 welding of pancakes into much larger floes forming solid plates that can extend over
 77 kilometers. After June 12, only frequencies under 0.1 Hz are episodically recorded
 78 as the ice margin continues to extend further offshore reaching a distance of 150 km
 79 from 624 by June 24. Wave energy exceeds the instrument noise floor during seven
 80 events observed by 624; each of which is associated with a storm (Fig. 2.a). Buoy
 81 625 kept a nearly constant distance south of 624, 120 km further into the ice field,

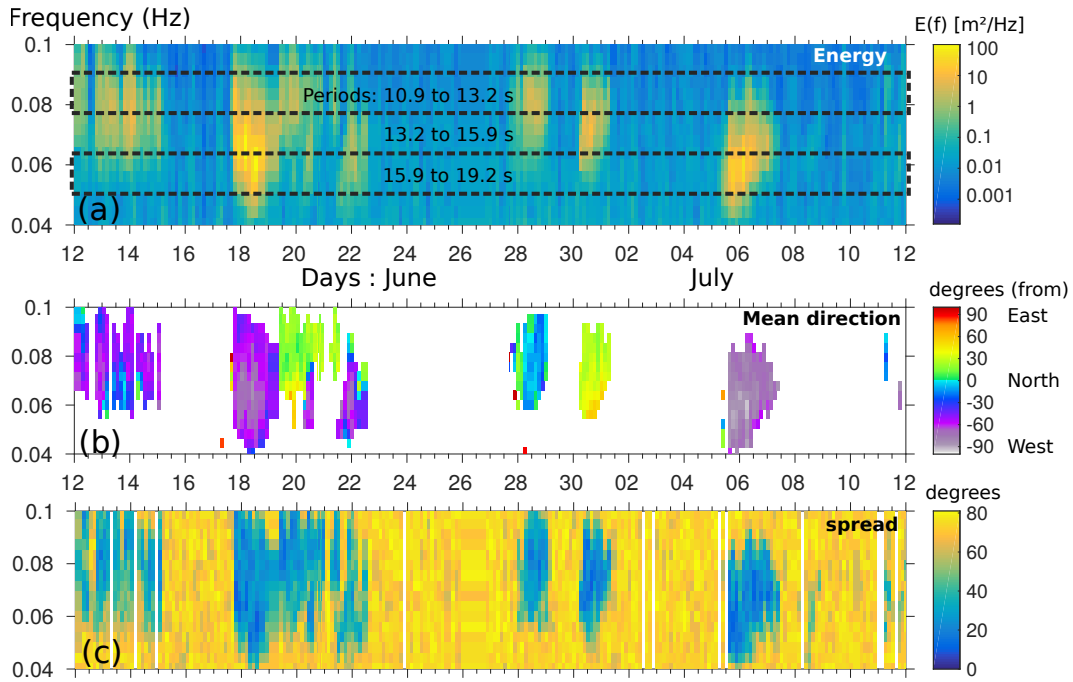


Figure 2. Wave properties measured by buoy 624 as a function of time and frequency. (a) Energy spectrum, with dotted boxes around the short and long period ranges used in Fig. 4, (b) Mean directions, shown when the energy is above -10 dB, avoiding random noise for low energies. (c) directional spread.

82 until its trajectory crosses iceberg B30 on July 14. This 30 km long iceberg is clearly
 83 visible in all the AMSR2 images as it cuts a wake of open water in the sea ice. Before
 84 July 13, buoy 623 was out of the ice, providing measurements representative of open
 85 water. A numerical model simulation without ice attenuation shows good agreement
 86 with these measurements (Fig. 1.d). This model is used to quantify the time delay
 87 and dispersion due to propagation from open water (buoy 623) into the ice (buoys 624
 88 and 625). The difference between the model and observations can then be attributed
 89 to wave-ice interaction processes.

90 The wave height observed within the ice is not a simple function of the open
 91 water measurements, as shown in Fig. 1.d. The attenuation from open water to buoy
 92 624 depends on the wave frequency and direction. Attenuation is stronger for higher
 93 frequencies. Attenuation is also stronger when waves travel a longer distance under the
 94 ice, namely when their directions are more oblique relative to the ice edge (e.g. from
 95 the West, in grey-purple colors on July 4–7, Fig. 2.b). The buoy measures narrow
 96 directional distributions with spreads under 20 degrees (blues in Fig. 2.c) as soon as
 97 the energy level exceeds $0.1 \text{ m}^2/\text{Hz}$. The measured waves are thus similar to open
 98 ocean swell with long crests and a narrow range of propagation directions.

99 We now turn to Sentinel-1 radar imagery for information on ice properties. These
 100 SAR images were analyzed for two types of features (Fig. 3).

101 First we determined the presence (blue symbols) or absence (red symbols) of
 102 leads. Leads are straight features at the boundaries of large scale floes that may
 103 correspond to an ice-free surface, which appear as dark bands, or that can re-freeze and
 104 appear bright due to the presence of frost flowers. When leads are visible, the diameters
 105 of floes exceed 1 km and waves have not been able to break the ice. In the absence of

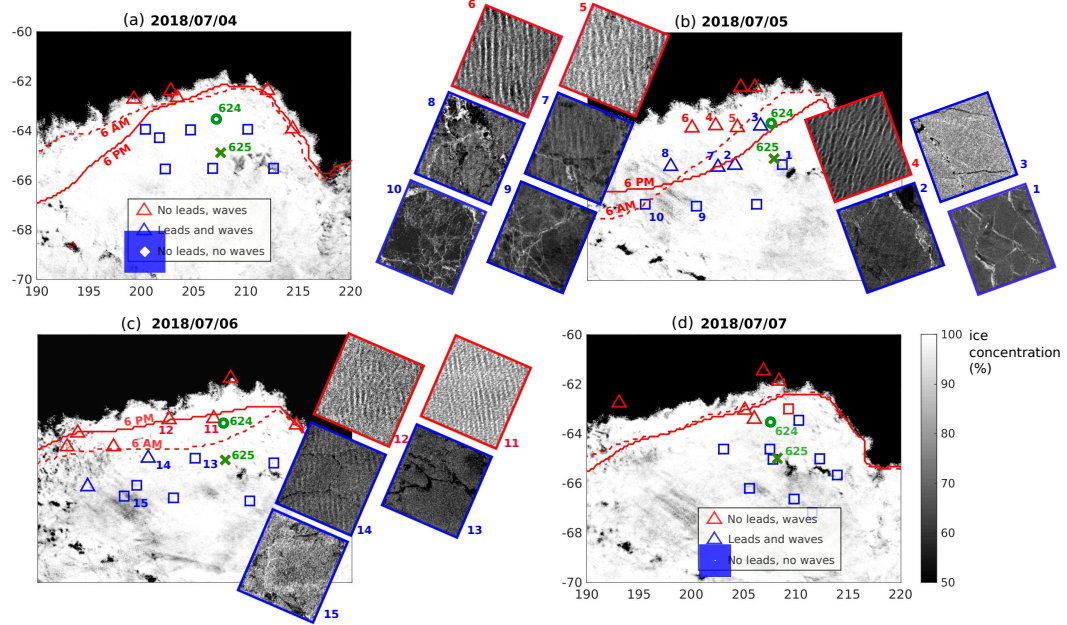


Figure 3. Daily maps of ice concentrations from AMSR2-ASI as provided by University of Bremen), here positions and locations of Sentinel 1 Wave Mode SAR imagery classified by the presence of leads and waves, through a storm event from (a) July 4th to (d) July 7th. Each SAR image is represented by a blue symbol if leads are detected, red if no lead is detected, triangles if waves are visible, squares otherwise. For some of the images, numbered 1 to 15, sub-panels of the SAR images are also shown. Each sub-panel is 5 km by 5 km and full images are shown in Fig. S1-S15. Finally, the contour of maximum floe diameter $D_{max} = 300$ m, as estimated from the model run using aerologic dissipation, is shown with a dashed line at 6 AM and solid line at 6 PM marking the extent of the region where ice is broken by waves in the model.

106 leads, it is expected that ice is broken into small floes and their differential advection
 107 yields a horizontally uniform brightness. The second type of feature is wave patterns.
 108 Images numbered 4–6 acquired on July 5 clearly show a wave field propagating from
 109 the West, with a dominant wavelength around 400 m that is reduced to 270 m in
 110 images 11, 12 and 14 acquired on July 6. Estimated wave heights [Ardhuin *et al.*,
 111 2017] range from 0.7 to 2.5 m. Much fainter wave patterns are also visible in images
 112 2, 3 and 7. All these wave features are long-crested except in image 8 where ring
 113 waves are observed in probable association with reflections off an iceberg. The radar
 114 imagery acquired around the trajectories of the buoys thus confirms the penetration,
 115 200 km inside the ice, of long-crested waves with heights on the order of 1 m. Also,
 116 we interpret the disappearance of leads towards the ice edge on July 5 and 6 as the
 117 result of ice break up by wave action.

118 3 Interpretation using a wave attenuation and ice break-up mode

119 The magnitude and a plausible cause of wave attenuation will now be examined
 120 with a numerical wave model, based on version 6.05 of the WAVEWATCH III modelling
 121 framework [The WAVEWATCH III[®] Development Group, 2019, hereinafter WW3],
 122 that is constrained by these ice observations. We consider three model simulations
 123 that differ only in their parameterizations of wave-ice interactions:

- 124 • A simple "no attenuation" simulation in which the effect of the ice is limited to
 125 shutting off the generation of waves by wind forcing and dissipation associated
 126 with wave breaking. The difference between the measurements and the "no
 127 attenuation" run provides an estimate of the ice-induced attenuation.
- 128 • A scattering + viscous basal friction ("S1V1" for short) in which the scatter-
 129 ing coefficient is determined by the ice thickness and maximum floe size fol-
 130 lowing *Williams et al.* [2013] with the addition of back-scattering (so that the
 131 scattering term conserves energy) and viscous dissipation taken from *Liu and*
 132 *Mollo-Christensen* [1988] with the viscosity taken to be the molecular viscosity
 133 of sea water at the freezing point. Our scattering term conserves wave energy
 134 which is isotropically redistributed as discussed in *Boutin et al.* [2018]. Scatter-
 135 ing strength is based on the normal reflection of waves travelling from the open
 136 ocean under a semi-infinite ice floe with a straight boundary [*Williams et al.*,
 137 2013].
- 138 • An anelastic dissipation based on known microscopic rheological properties of
 139 dislocations in ice crystals [*Cole et al.*, 1998; *Cole*, 2020], adapted by *Boutin*
 140 *et al.* [2018] to represent the dissipation of waves when floes with diameters
 141 larger than half the wavelength are flexing. This parameterization is combined
 142 with viscous basal friction and scattering, with the scattering coefficient reduced
 143 by a factor 5.

144 Apart from the differences in parameterizations, the common features of the three
 145 model simulations reported here are a spatial resolution of 0.1 degree in latitude and
 146 0.2 degree in longitude forced at the boundaries (red line in Fig. 1.b) by a 0.5 degree
 147 global model. Wind forcing is provided by the European Center for Medium Range
 148 Weather Forecasts (ECMWF) operational analyses and forecasts. Ice concentration
 149 was provided by AMSR2 with a correction designed to limit the errors caused by
 150 atmospheric water vapor [*Gloersen and Cavalieri*, 1986]: all concentrations above 70%
 151 was set to 100%. Based on climatology data, the ice thickness was set to a constant
 152 0.77 m. Given our focus on wave periods longer than 10 s, we neglected the effect of
 153 ice on wave dispersion [*Collins et al.*, 2018]. This is discussed in the Supplementary
 154 Information. The model uses 24 directions and an exponential frequency grid from
 155 0.037 to 1 Hz.

156 The ice flexural strength σ_c was adjusted to obtain a position of the break-up
157 front consistent with SAR imagery, as shown in Fig. 3, giving $\sigma_c = 0.6$ MPa.

158 In all cases the maximum floe size diameter is estimated based on the expected
159 local maximum flexural strength over the model time step of 600 s [Boutin *et al.*, 2018].
160 This floe size only has an impact on wave dissipation rate in the case of the anelastic
161 simulation.

162 We also note that other wave-ice interaction effects have been investigated in
163 other studies [Squire, 2020], including ice floe collisions [Herman *et al.*, 2019] or the
164 breaking of steep waves over ice floes [Toffoli *et al.*, 2015]. Although these processes
165 are reasonably important right at the ice edge, we expect them to play only a minor
166 role overall because they are typically non-linear with higher relative dissipation for
167 higher waves, hence leading to a shorter attenuation distance for steeper ocean waves.
168 This is not consistent with available large scale observations that show a constant or
169 weaker attenuation for steeper waves [Meylan *et al.*, 2014; Stopa *et al.*, 2018], and we
170 thus have not considered these processes here.

171 Modeled wave parameters are compared to the buoy measurements in Fig. 4.
172 A detailed analysis of the shape of wave spectra is performed in the Supporting In-
173 formation, see Figures S16-S21. Starting with the directional spread in Fig. 4.a,
174 measurements during wave events are only slightly larger than those modeled without
175 ice attenuation. In contrast, the modeled spread is much larger than observed values
176 when scattering is a dominant term in the attenuation (run S1V1). Reducing the
177 spread to levels comparable to measured values requires a reduction of the scatter-
178 ing coefficient by at least a factor 5, as in the "Anelastic" simulation. This reduced
179 scattering requires stronger dissipation processes to explain the measured wave height
180 attenuation. This attenuation is estimated as the ratio between the "no attenuation"
181 simulation and the measured height, and it varies dramatically in time and across
182 frequencies. Some of these variations are caused by different offshore directions (Fig.
183 2.b). Still, for similar wave directions, the observed attenuation varies from 80% to
184 20% between June 20 and June 30 (purple arrows in Fig. 4.c). The gradual variation
185 of wave attenuation in the S1V1 simulation is markedly different from observations
186 which exhibit a sharp cut-off after each wave event such as on July 2 or 8 when no
187 wave energy is measured at buoy 624 (Fig. 4.b-d). The anelastic simulation repro-
188 duces that effect which, in the model, is due to displacement of the "break up front"
189 shown with red contours on Fig. 3. When buoy 624 is far from the "break up front",
190 the distance traveled by waves across unbroken ice is large and the attenuation is very
191 strong. In contrast, when the ice is broken up closer to the buoy, the attenuation is
192 much less with an overall wave height attenuation by ice as low as 20%.

193 The anelastic model scenario presented above is not a unique solution. We as-
194 sumed an ice thickness $h = 0.77$ m and adjusted the ice flexural strength to $\sigma_c =$
195 0.6 MPa in order to obtain a spatial extent of broken ice consistent with SAR imagery.
196 Because the ice break-up condition in the model is set by a maximum stress that is
197 proportional to $\sigma_c h^2$, the same results may be obtained by increasing h by a factor a
198 and reducing σ_c by a factor a^2 . Likewise, the magnitude of the anelastic dissipation
199 is controlled by the compliance of the relaxation of dislocations δ , and the number of
200 dislocations per unit area [Cole, 2020]. Since the wave dissipation rate is proportional
201 to $h^3 \delta$, any change in h can be compensated with a change of δ to give the same wave
202 evolution. Conversely, assuming that δ and σ_c are well constrained, it is possible to
203 estimate the ice thickness from SAR imagery using either wave heights and periods at
204 the break-up front or the attenuation of waves in unbroken ice.

205 For such applications a more realistic dissipation model may be needed. Indeed,
206 the present parameterization tends to overestimate the dissipation of frequencies above
207 0.085 Hz, as shown in Supporting Figures S16-S20. One possible reason for this model

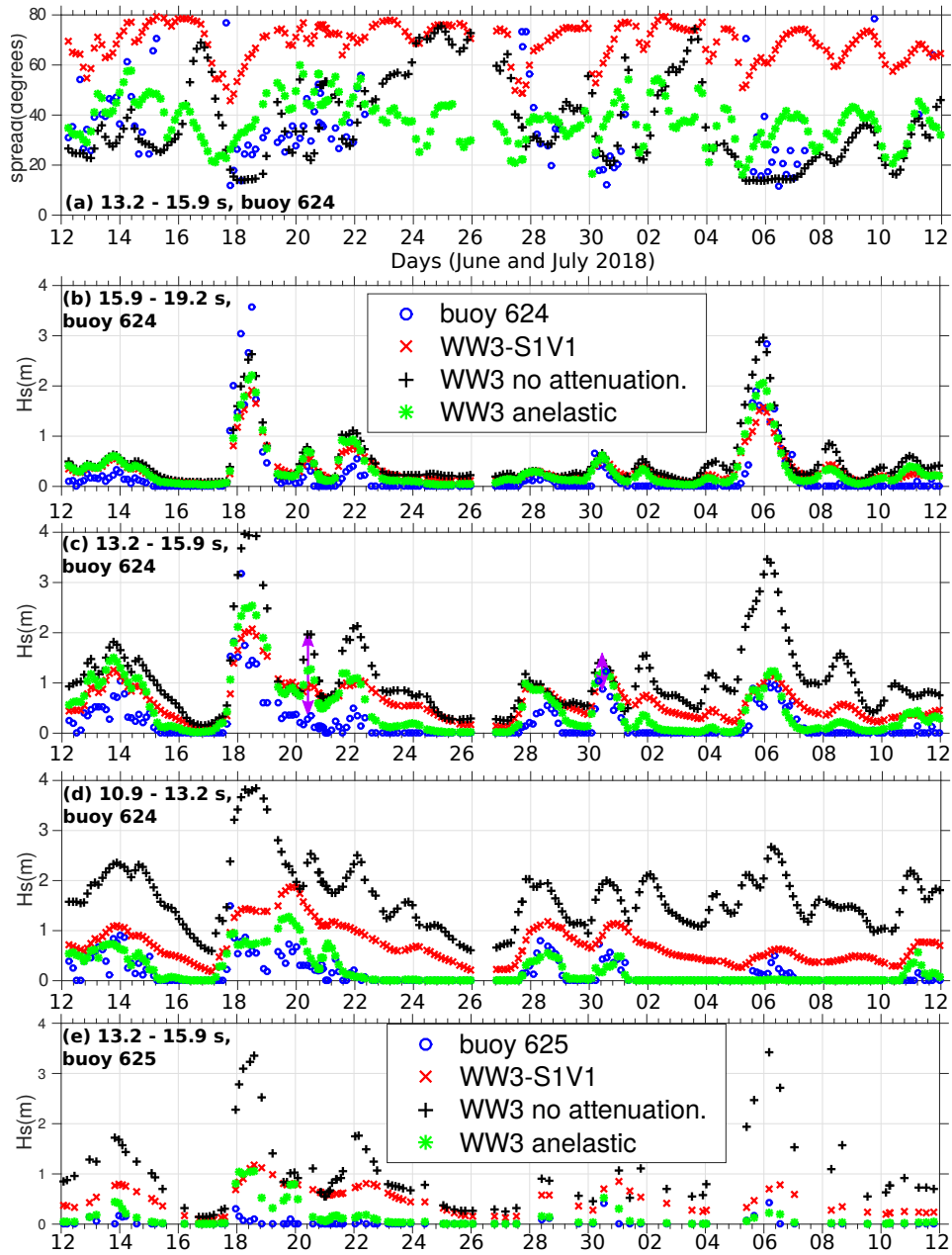


Figure 4. Time series of measured and modeled (a) directional spread and (b-e) wave heights for the ranges of wave periods shown in Fig. 2.a. The model results are shown for the 3 different parameterizations of wave-ice interactions. Purple arrows in (c) point to markedly different attenuation rates, 80% on June 20 and 20% on June 30.

208 error is the neglected effect of ice in the dispersion relation [Robinson and Palmer,
 209 1990], but other dissipative processes may be involved, including flow in brine pockets
 210 during ice deformation, and viscous effects in ice deformation [Meylan *et al.*, 2018;
 211 Marchenko and Cole, 2017; Cole, 2020].

212 4 Conclusions

213 As storms passed over the open ocean, we measured waves in the ice with narrow
 214 directional spreads that are not compatible with existing parameterizations of wave
 215 scattering by ice floes. Instead, dissipation associated with ice flexure is a plausible
 216 explanation for the previously observed non-linear wave attenuation [Stopa *et al.*, 2018]
 217 with weaker attenuation for higher waves. This dissipation provides a transition from
 218 a non-broken and highly dissipative flexing ice field to one which is fragmented into
 219 small floes with weak attenuation, consistent with our wave and ice observations. This
 220 new paradigm also has consequences on how the waves impact the ice. In particular,
 221 a dissipation-dominated attenuation concentrates the wave-induced push to the ice
 222 around the ice break-up front. In the melting season, the waves may have a more
 223 lasting impact as large floes may not reform after each storm, allowing break-up to
 224 progress further with the next storm.

225 Acknowledgments

226 We thank the New Zealand Navy for buoy deployments, and the EU/ESA Copernicus
 227 program for the acquisition of SAR imagery. The analysis was supported by
 228 ONR. Buoy data has been uploaded to the UC San Diego Library Digital Collections
 229 (<https://doi.org/10.6075/J0V40SM5>), Level-1 SAR wave mode data was provided by
 230 the European Space Agency and their processing is supported by Sentinel-1 A Mis-
 231 sion Performance Center (Contract No. 4000107360/12/I-LG), and can be viewed at
 232 <http://www.ifremer.fr/datavore/exp/dvor/#/s1quicklook>. Finally model simulations
 233 are available at <http://tiny.cc/wavesinice>. Comments from anonymous reviewers and
 234 discussions with David Cole are gratefully acknowledged.

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Supporting Information for "Ice break-up controls dissipation of wind-waves across Southern Ocean sea ice"

Fabrice Ardhuin¹, Mark Otero¹, Sophia Merrifield¹, Antoine Grouazel², and Eric Terrill¹

¹Scripps Institution of Oceanography, La Jolla, California

²Univ. Brest, CNRS, IRD, Ifremer, Laboratoire d'Océanographie Physique et Spatiale (LOPS), IUEM, Brest, France

Contents of the Supplementary information

1. Introduction
2. Movie S1
3. Figures S1 to S21. Note that S16 to S21 are included in the following discussion.
4. Discussion of wave spectral shapes and possible linear dissipations proportional to the frequency to a power n .

1. Introduction

This Supporting information contains a movie that complements Figure 1 and helps visualize buoy locations relative to the ice edge in the context of the storms observed. The supplementary figures S1-S15 provide a more detailed view of the ice and wave features observed on July 5 and 6, complementing figure 4. Figures S16 to S21 are associated with the discussion of spectral shapes and the possible identification of a linear dissipation process proportional to f^n , following the type of analysis performed by Meylan et al. (2018).

2. Movie S1

Location of buoys 623, 624, and 625 in the context of the evolving sea ice concentration field estimated from the Advanced Microwave Scanning Radiometer 2 and processed by the University of Bremen (Spreen et al. 2008). Maximum floe size contour $D_{\max} = 300$ m, wave heights, and mean directions are taken from the the "Anelastic" model run.

3. Figures S1-S15

Quicklooks of SAR images labeled 1-15 in Fig. 3.

These can be obtained from <http://www.ifremer.fr/datavore/exp/dvor/#/s1quicklook>

Images are displayed in acquisition geometry with range (satellite cross-track direction) on the x -axis and azimuth (satellite along-track direction) on the y -axis. The original Level 1 product, with a resolution of 4 m, is subsampled to a 600 x 600 pixel image of 35 m resolution. The grey scale representing radar backscatter after SAR processing is automatically adjusted in each image to linearly cover the 1st to 99th percentiles of the image intensity in order to maximize contrast. The mean sea surface roughness value (μ) is indicated on each figure. The samples shown in Fig. 3 were rotated to orient North upward.

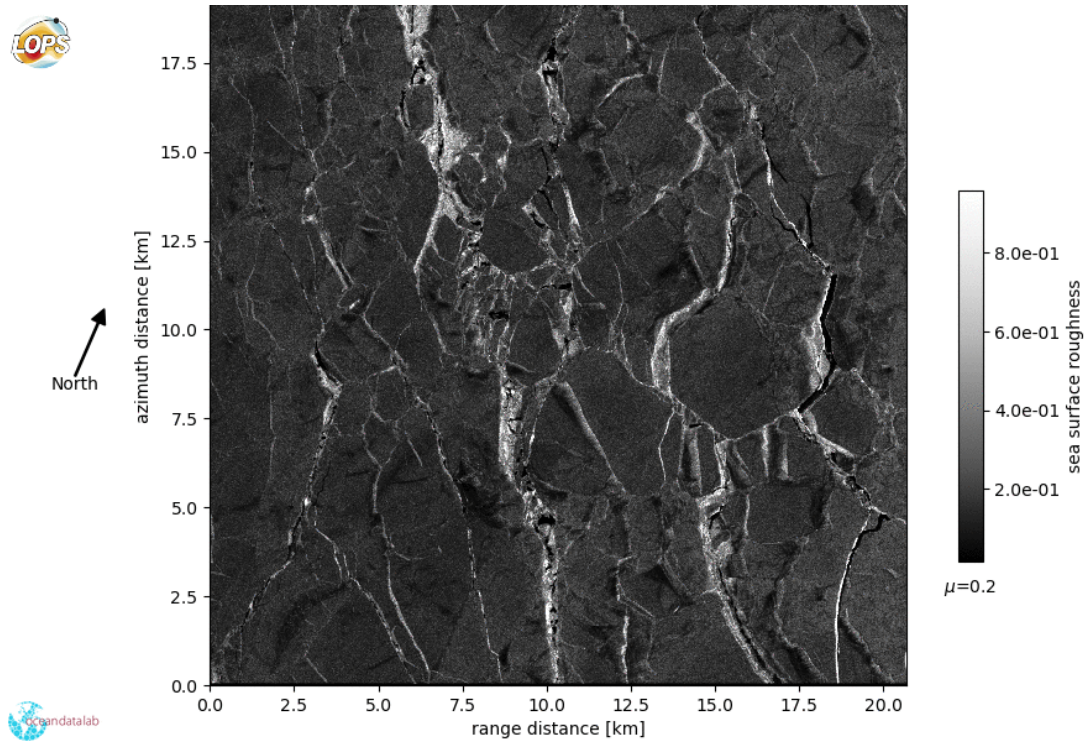


Figure S1. SAR image acquired 2018/07/05 at 06:01:08

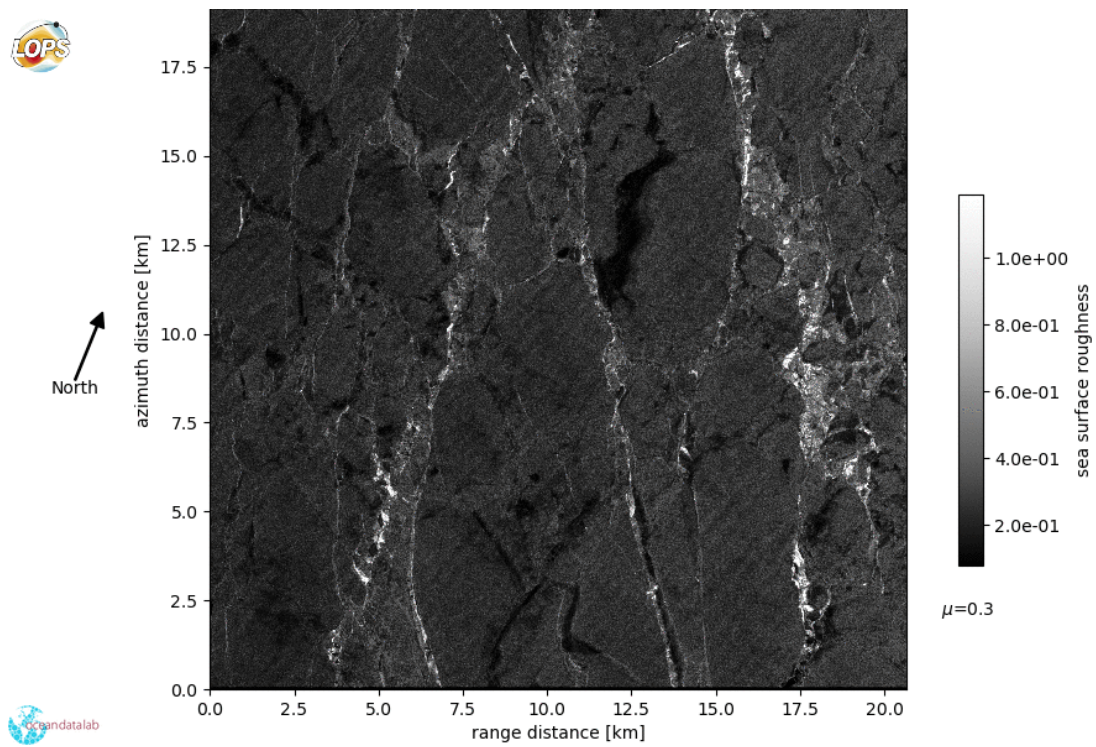


Figure S2. SAR image acquired by Sentinel 1B on 2018/07/05 at 06:01:38

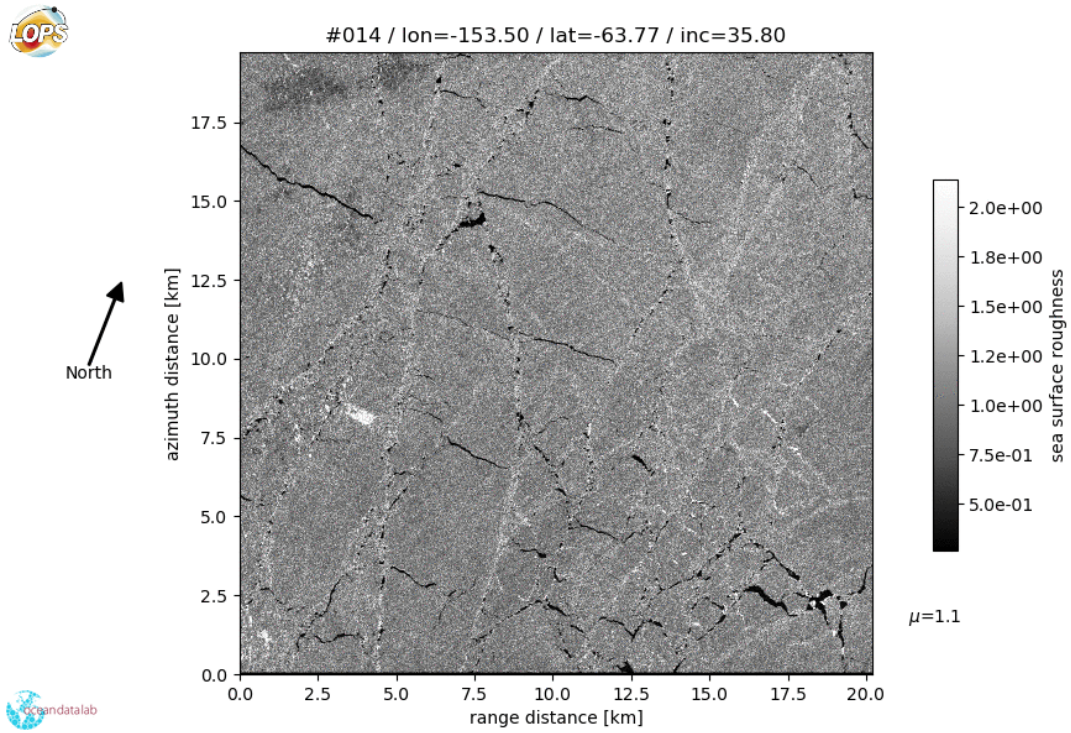


Figure S3. SAR image acquired by Sentinel 1B on 2018/07/05 at 06:01:52

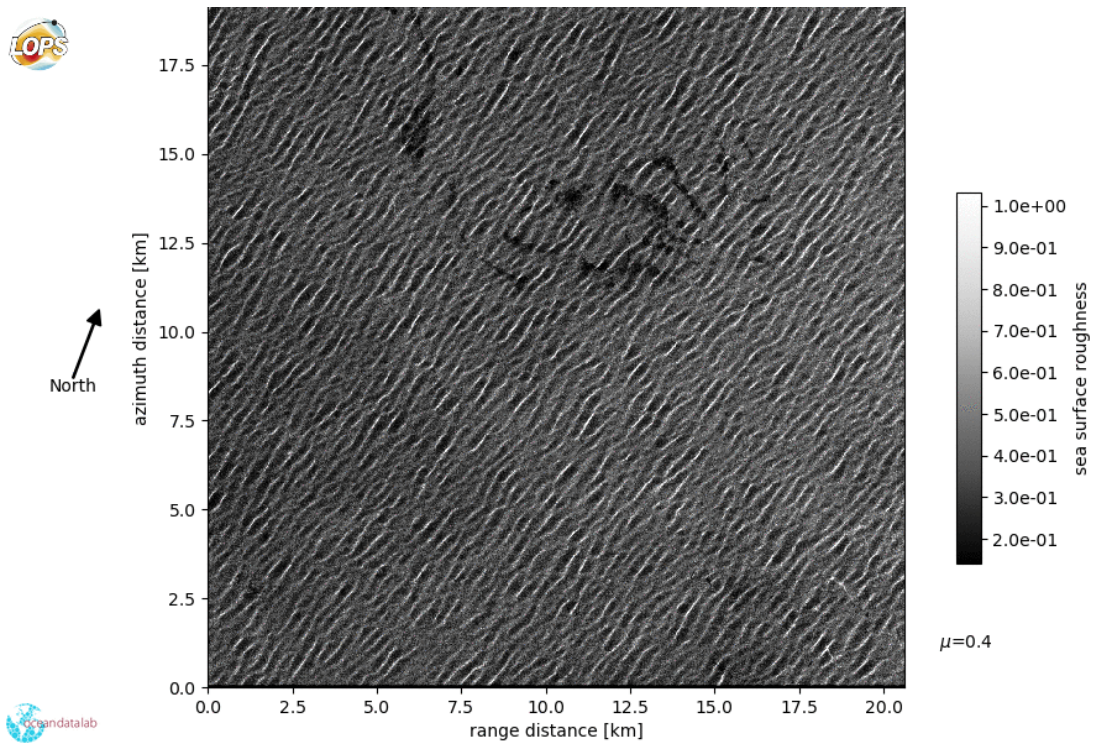


Figure S4. SAR image acquired by Sentinel 1B on 2018/07/05 at 06:02:07

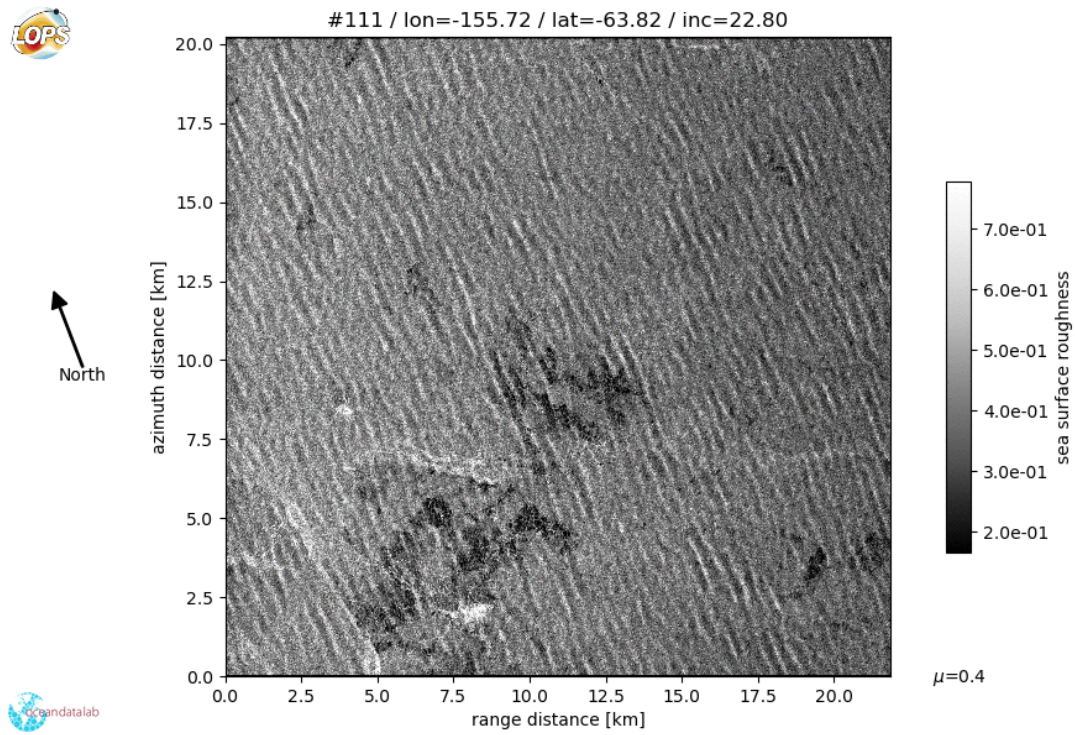


Figure S5. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:52:37

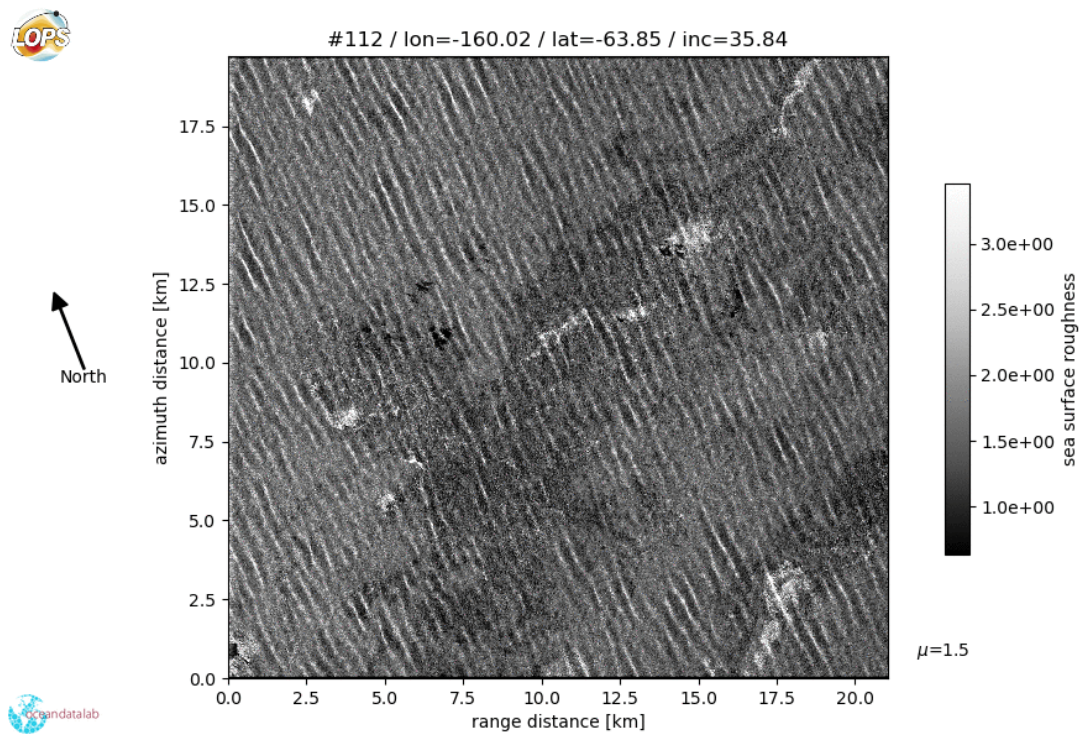


Figure S6. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:52:52

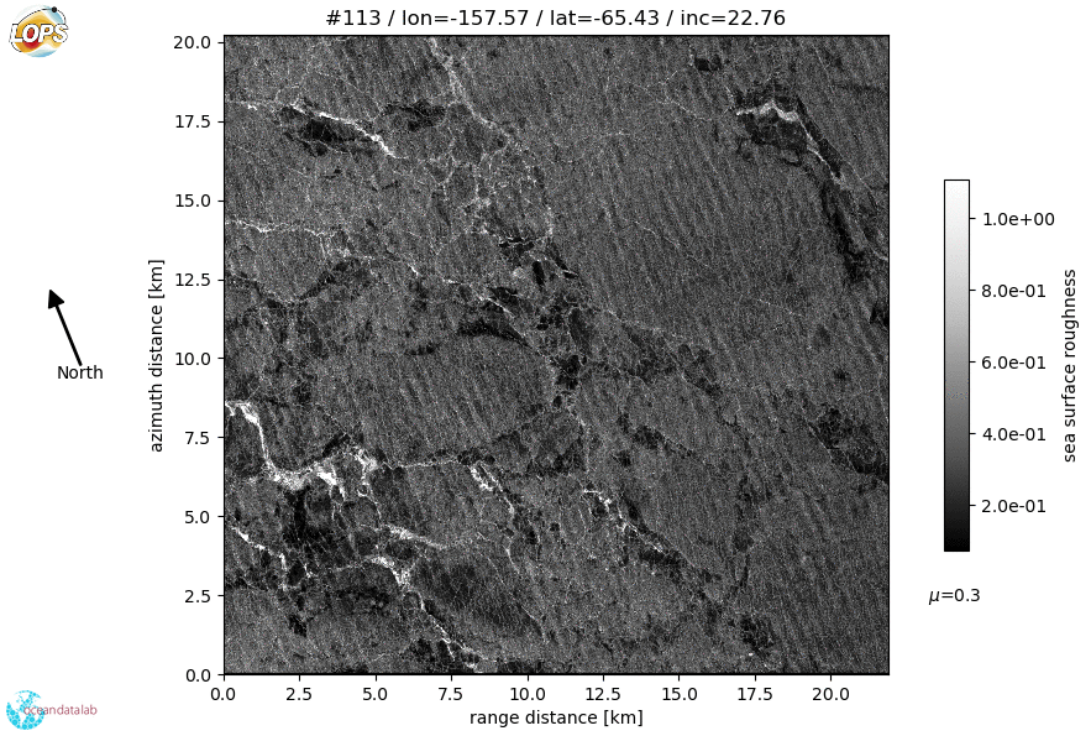


Figure S7. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:53:06

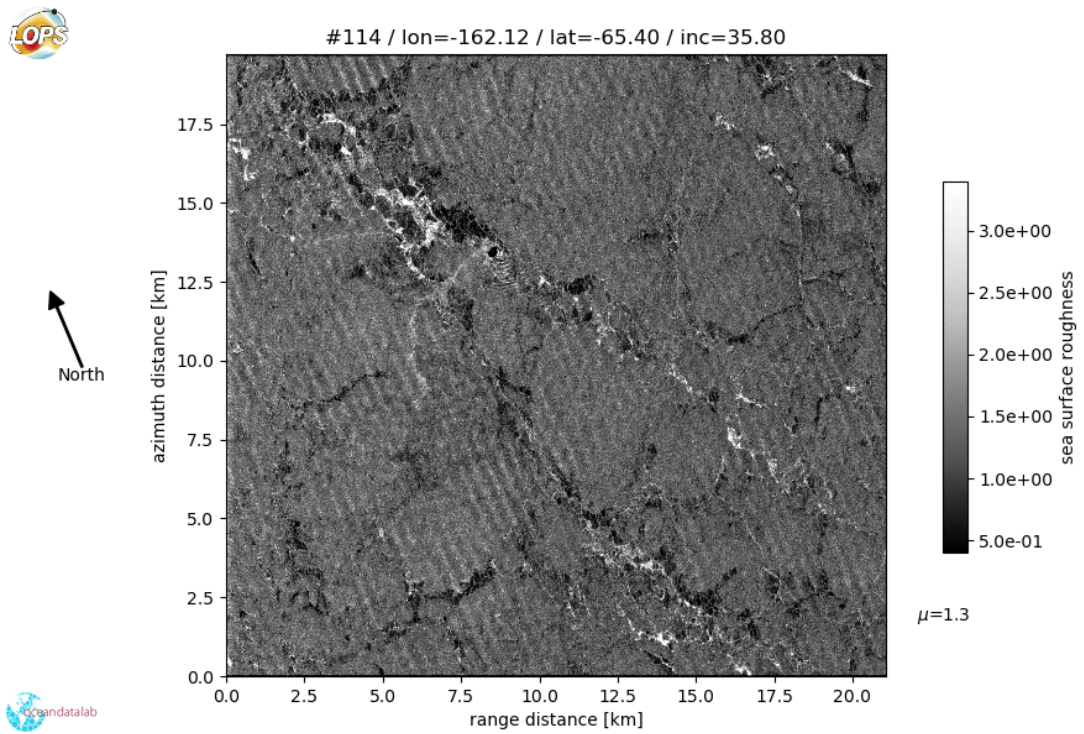


Figure S8. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:53:21

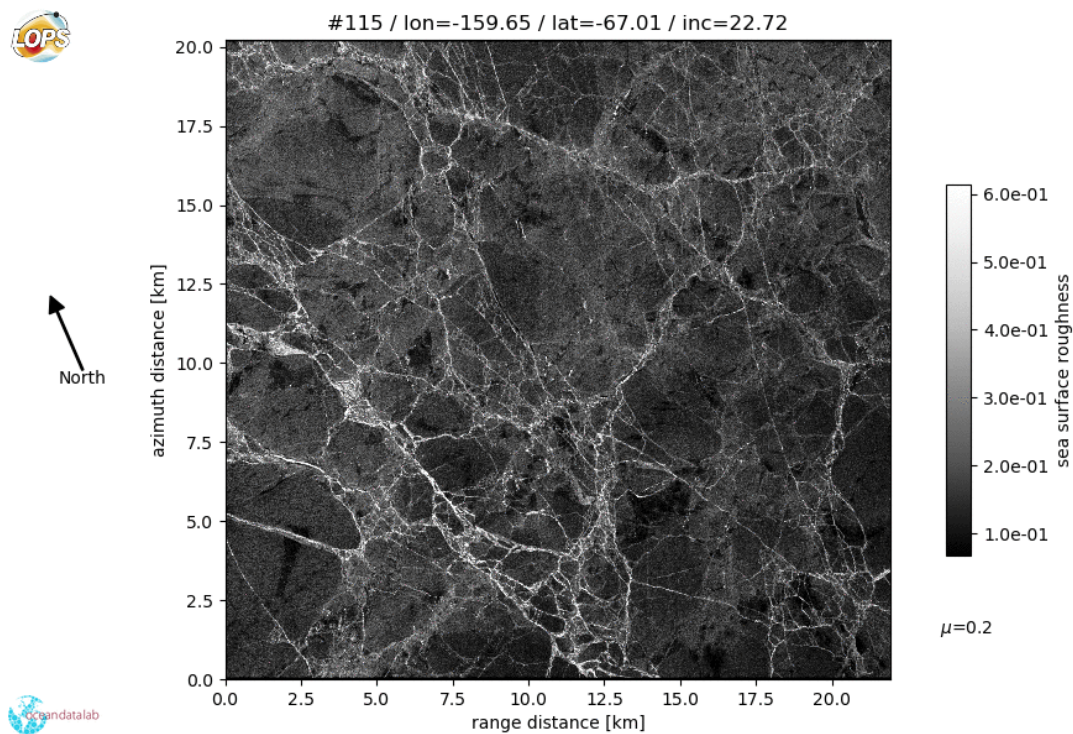


Figure S9. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:53:3

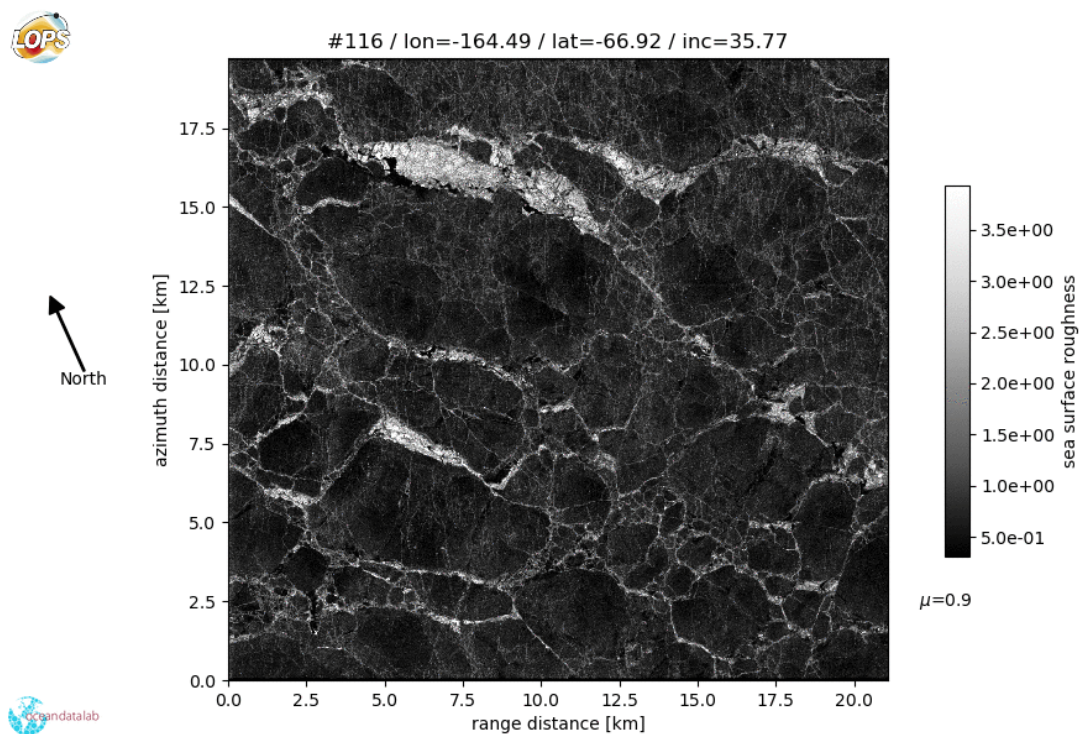


Figure S10. SAR image acquired by Sentinel 1A on 2018/07/05 at 14:53:50

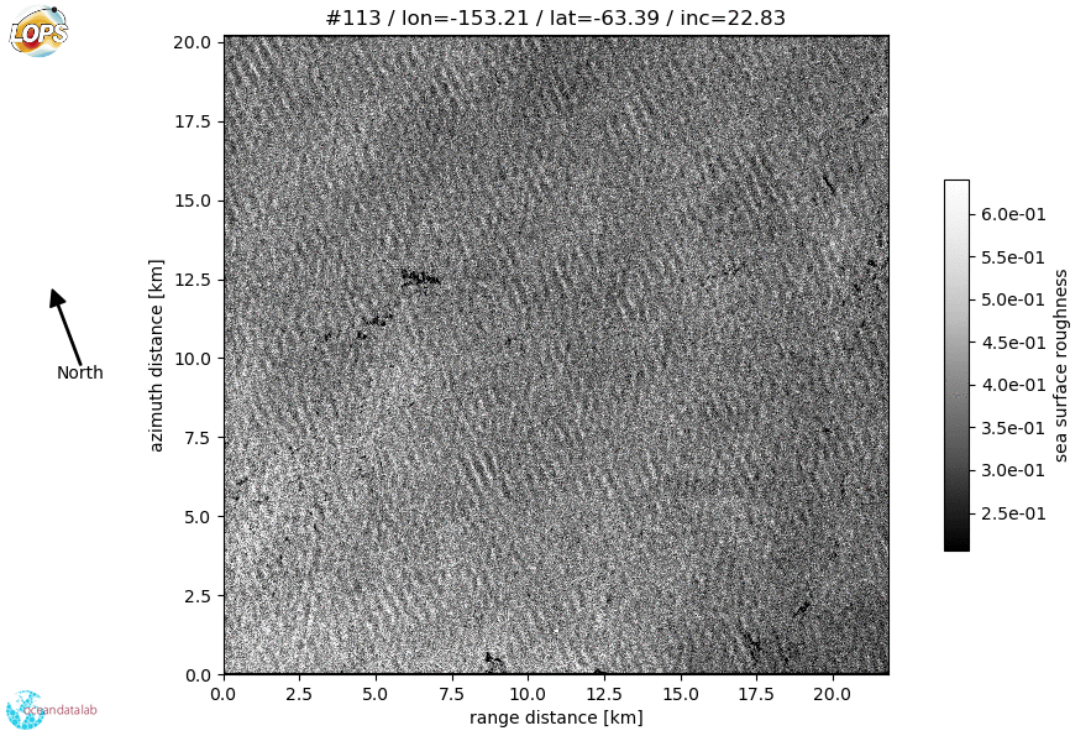


Figure S11. SAR image acquired by Sentinel 1B on 2018/07/06 at 14:43:34

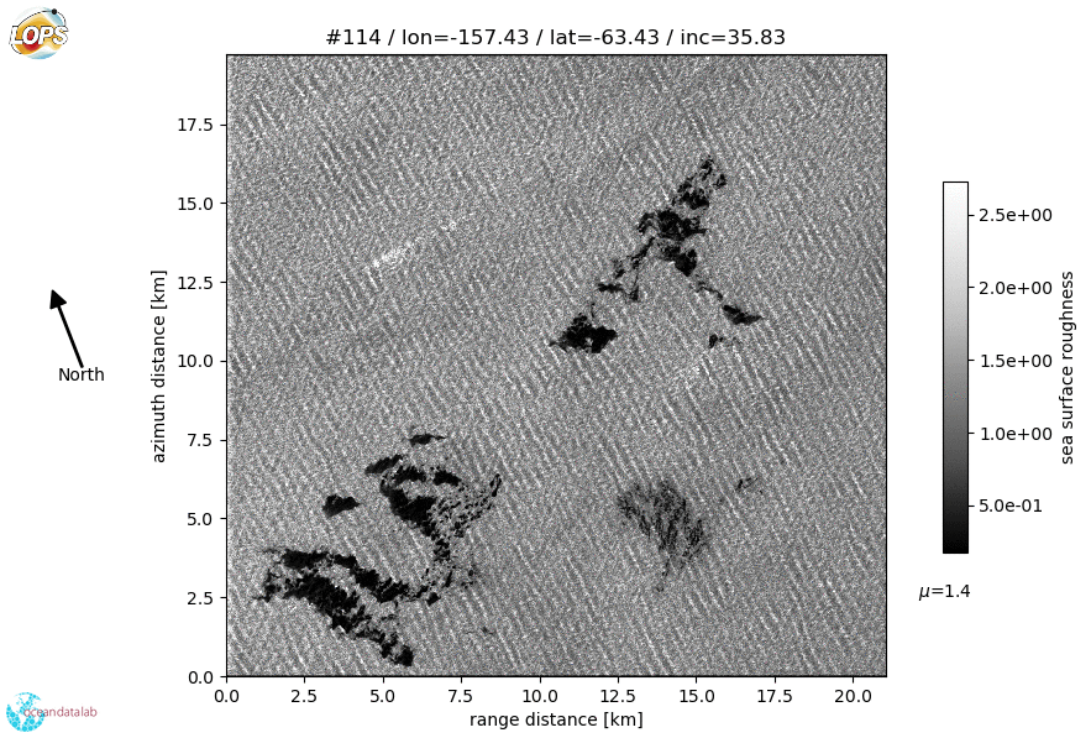


Figure S12. SAR image acquired by Sentinel 1B 2018/07/06 at 14:43:49.

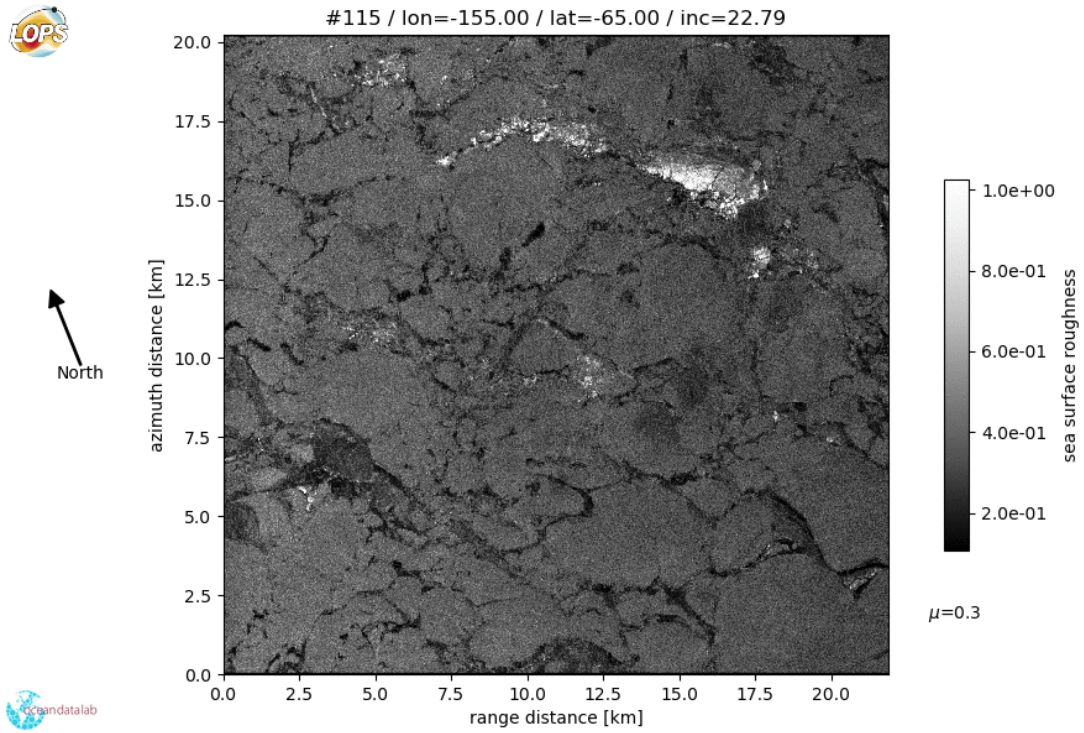


Figure S13. SAR image acquired by Sentinel 1B on 2018/07/06 at 14:43:03.

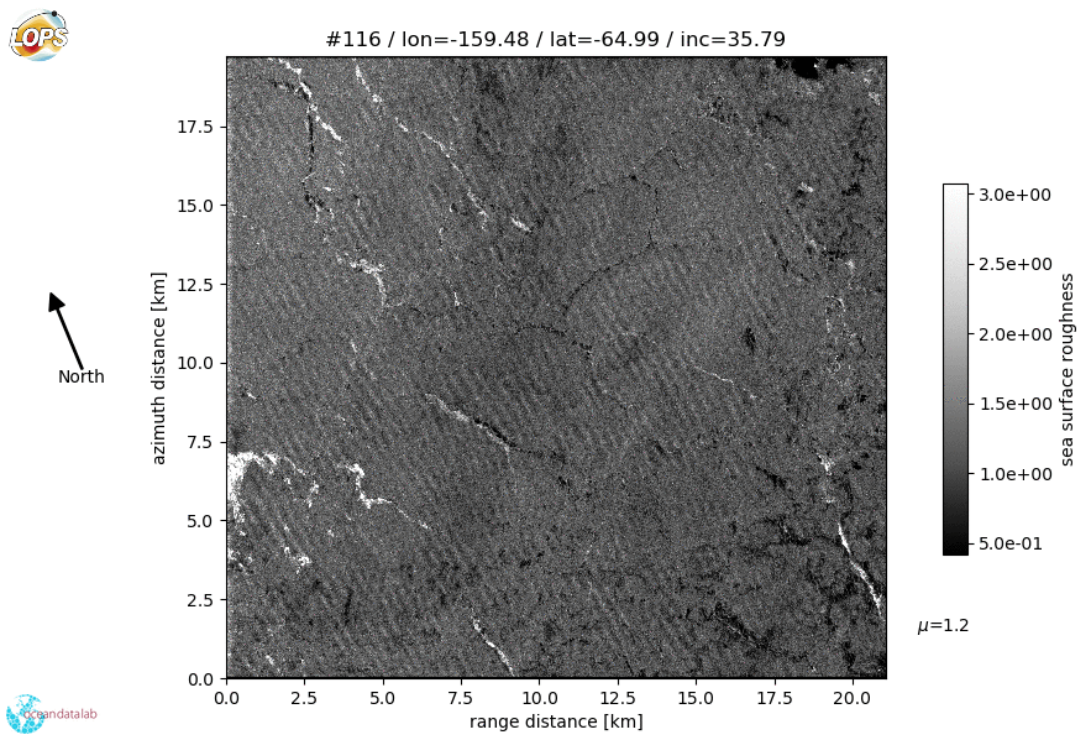


Figure S14. SAR image acquired by Sentinel 1B on 2018/07/06 at 14:44:18.

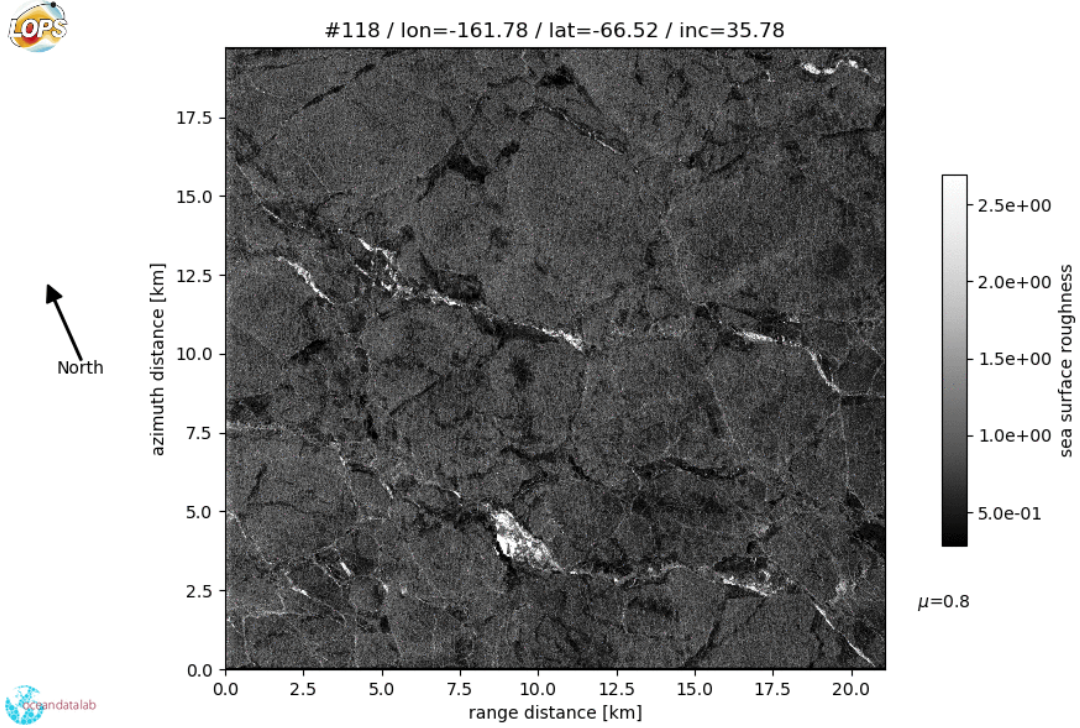


Figure S15. SAR image acquired by Sentinel 1B on 2018/07/06 at 14:44:50.

4. Discussion and Figures S16-S21

Whereas the main body of the paper focuses on the temporal evolution of wave properties, here we examine the spectral distribution of wave energy which provides a complementary perspective of possible dissipation processes. Indeed, Meylan et al. (2018) have argued that different dissipation processes have distinct spectral signatures. This idea has been used to characterize the effect of frazil and pancake ice on spectral shape.

This approach is particularly well suited for linear processes. Considering a 1-dimensional propagation along the x axis, the wave spectrum $E(x, t, f)$ is attenuated by a source function $S_{ice}(x, t, f) = a(x)f^m E(x, t, f)$, namely

$$\partial E(x, t, f) / \partial t + \partial [C_g(f)E(x, t, f)] / \partial x = S_{ice}(x, t, f). \quad (1)$$

Using the method of characteristics this gives,

$$E(x, t, f) = E(x_0, t - C_g(f) \times (x - x_0), f) \times \exp \left[f^m / C_g(f) \int_{x_0}^x a(x') dx \right], \quad (2)$$

which can be generalized to any polynomial function of the frequency. Also, for a simple dispersion relation, $C_g(f)$ is also a power law of f and than can be combined with S_{ice} to give,

$$E(x, t, f) = E(x_0, t - C_g(f) \times (x - x_0), f) \times \exp \left[f^n \int_{x_0}^x b(x') dx \right]. \quad (3)$$

Without ice effect, the spectrum would have been

$$E_{noice}(x, t, f) = E(x_0, t - C_g(f) \times (x - x_0), f), \quad (4)$$

such that the power n can be obtained by taking the log of the ratio $r(x, t, f) = \log [E(x, t, f) / E_{noice}(x, t, f)]$. From values of r at two frequencies, f_i and f_j , we get,

$$n(x, t, f_i, f_j) = \log (r(x, t, f_i) / r(x, t, f_j)) / \log (f_i / f_j) \quad (5)$$

For Airy waves, $n = m + 1$ and the wavenumber k is proportional to f^2 . For viscous friction S_{ice} is proportional to $k\sqrt{f}$ and thus $m = 2.5$ and $n = 3.5$. For anelastic dissipation, we expect a much steeper variation with frequency, as illustrated in Figure S16, because the stress and strain are both proportional to the surface curvature, giving a k^4 contribution, that is reduced a little by the $f^{-0.5}$ variation of the phase shift between stress and strain Cole et al. (1998), leading to $n = 9.5$.

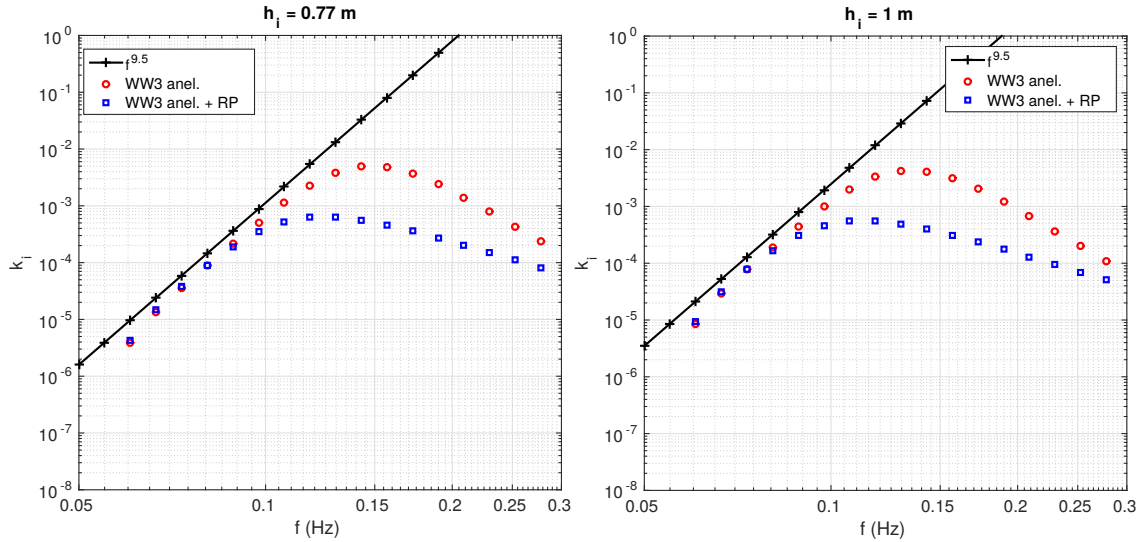


Figure S16. Amplitude decay (imaginary wavenumber) as a function of wave frequency for the anelastic model of Cole et al. (1998) when using a Robinson-Palmer (blue) or Airy (red) dispersion relation.

However, this is only valid as long as the elastic contribution to the wave energy can be neglected, which is a function of the ice thickness. Indeed, the conservation of the energy flux of waves propagating from the open water under the ice leads to a smaller wave amplitude.

In order to be fully consistent (which was not the case in the main body of the paper, for the sake of simplicity), the ice thickness should also be taken into account in the wave dispersion relation. For this we may use the dispersion relation by Robinson and Palmer (1990). This gives a transition of the wavenumber from $k \propto f^2$ at low frequencies to $k \propto f^{0.5}$ at high frequencies, and $C_g \propto f^{-1}$ to $C_g \propto f^{0.5}$. That effect of dispersion depends on the ice thickness, and for the value $h_i = 0.77$ m used in the paper, it is significant for wave frequencies above 0.1 Hz.

With all this in mind, we can now look at frequency spectra from the buoy measurements and model results. The average spectra at the off-ice buoy 623 and in-ice buoy 624 are shown in Figure S17.

From the monthly-averaged spectra, it is clear that below 0.08 Hz the model underestimates the off-ice wave energy. Above 0.08 Hz, the shape of the modeled spectra is too steep for the anelastic dissipation, and with about the right shape but too high by a factor 10 or more for the viscous dissipation. As shown in Figure S18, part of the discrepancy in the anelastic model could be due to the use of the Airy dispersion relation, with the Robinson and Palmer (1990) dispersion working better especially if we use an ice thicknesses larger than the 0.77 m used all the model results shown here.

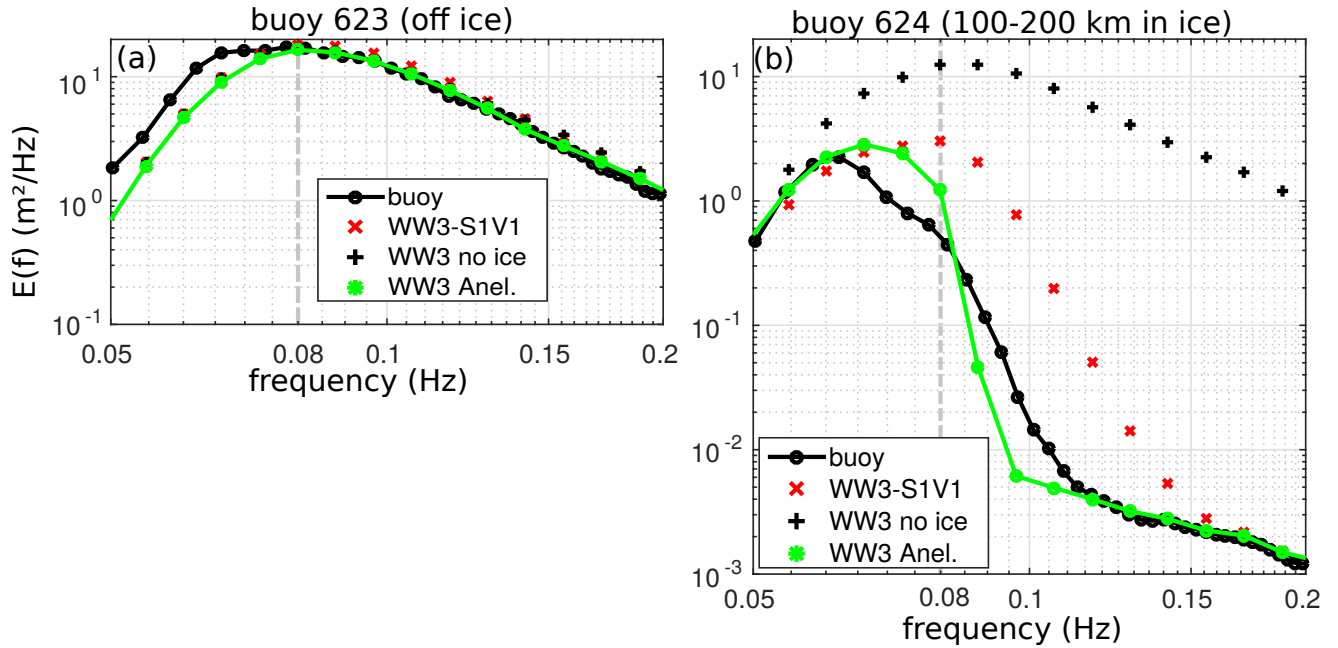


Figure S17. Mean spectra at buoys 623 and 624, from June 12 to July 12. A f^{-2} noise floor was added to the model to make it comparable to the measurements.

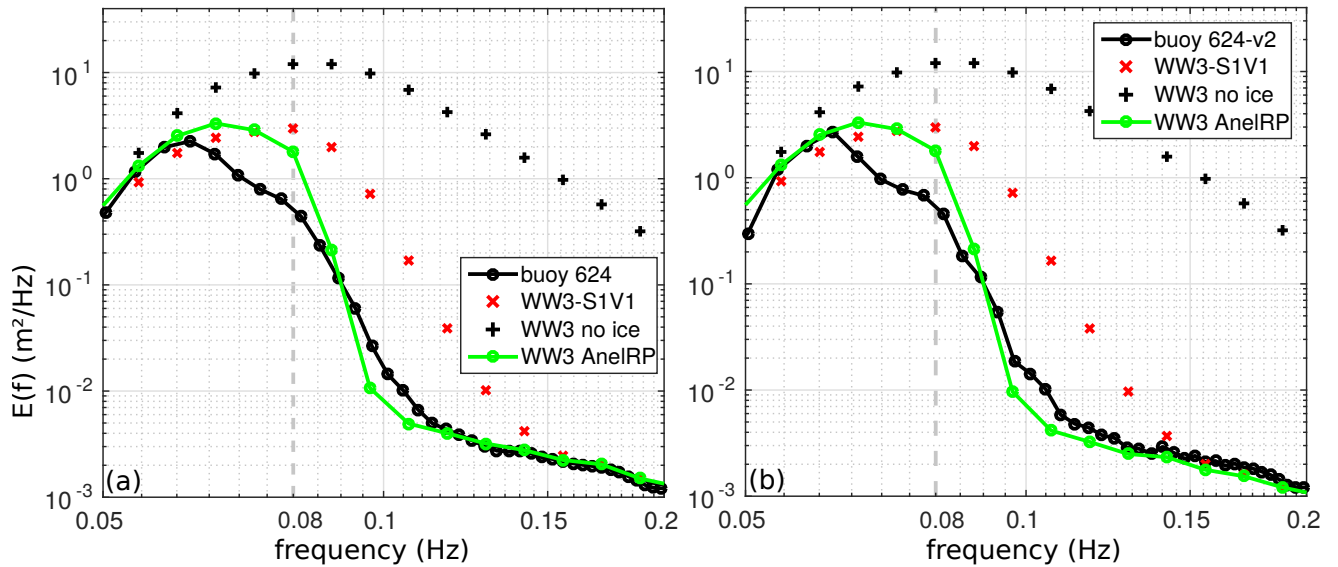


Figure S18. Mean spectra at buoy 624, from June 12 to July 12 with (a) the Anelastic WW3 model run now including the Robinson and Palmer dispersion relation, and (b) without the 3-point smoothing included in the data processing, giving a 30% lower spectral density at $f=0.97$ Hz.

Possibly the frequency smoothing that is used in the buoy data processing also contributes to that difference. Indeed, the default data processing has each spectral component obtained with a 3-point frequency (Hann filter) smoothing with weights $[0.25 \ 0.5 \ 0.25]$. That processing step tends to reduce the spectral slope at both 0.08 Hz (near the typical spectral peak) and at 0.097 Hz (where the spectrum approaches the noise level). Also, we note that when averaged over 1 day, the spectral estimate uses 128 degrees of freedom, hence a $\pm 24\%$ error at the 95% confidence level, which may be used to estimate an error on the fitted power law.

We may now take a look at the values of n estimated from the data. In practice we have computed used the discrete frequencies $f_{13} = 0.0815$, $f_{14} = 0.0854$, $f_{15} = 0.0894$, $f_{16} = 0.0933$, $f_{17} = 0.0972$ to compute $n_{14,15}$, $n_{15,16}$, $n_{16,17}$ and $n_{14,17}$, using daily-averaged spectra. The same n parameters were computed from the modeled spectra after their interpolation

on the observed frequency grid. The first result is that, even in the model runs, there is a wide variability of n . Also, on most of the days the energy level is below the noise floor and n cannot be estimated, so that any estimate of n will be biased towards the strong energy event.

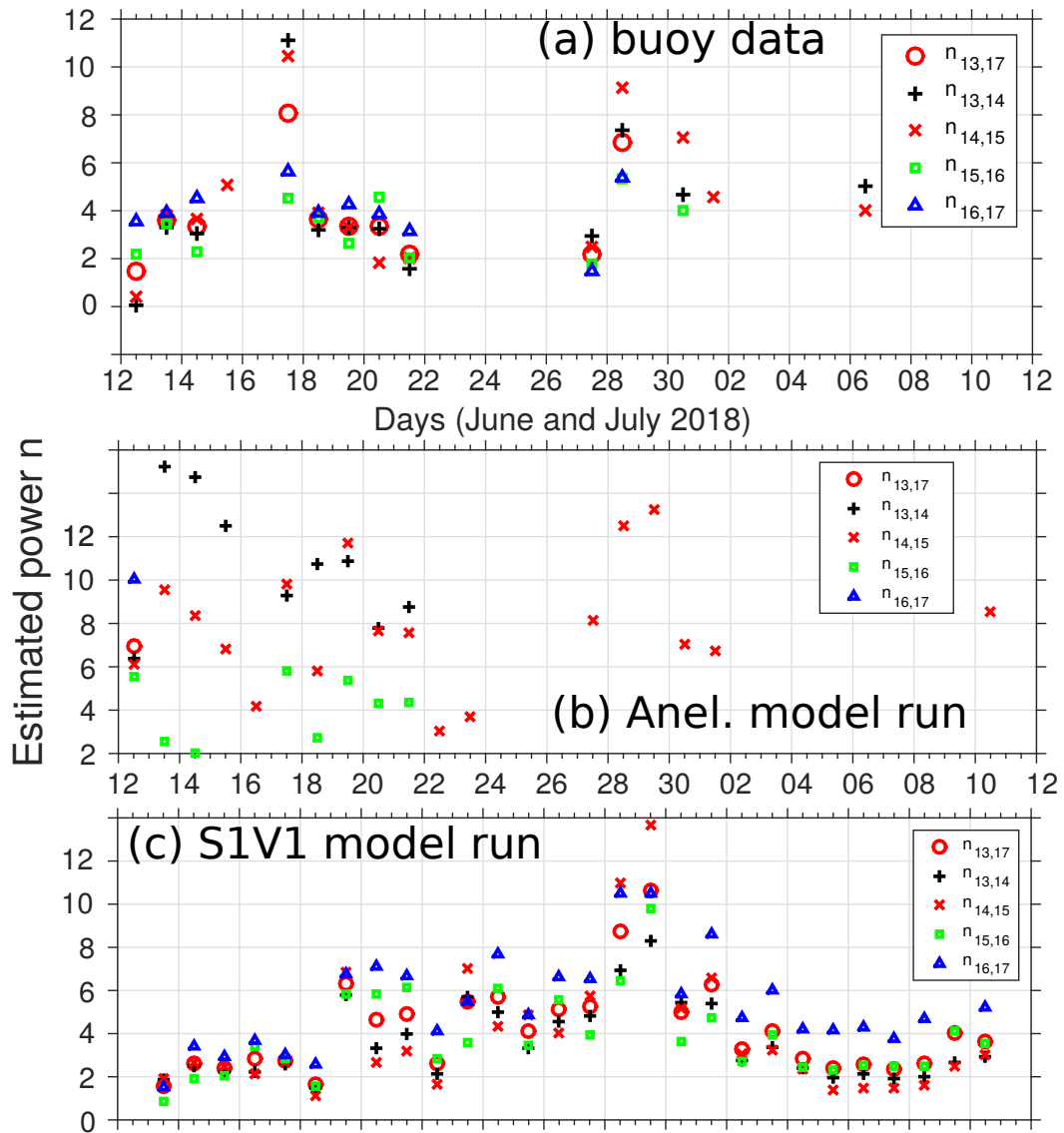


Figure S19. Dissipation power law parameters $n_{14,15}$, $n_{15,16}$, $n_{16,17}$ and $n_{14,17}$ estimated from daily-averaged (a) measured spectra, (b) modeled spectra with anelastic dissipation (c) modeled spectra with viscous dissipation. Estimates are only available when the spectrum is larger than 2 times the mean noise level.

Daily average estimates of n range from 0.2 to 11. The scatter is even larger for the anelastic model run, and the scattering-viscous model run (S1V1) has a large range of n values, showing that the simple linear attenuation of unidirectional waves does not represent well the more complex situation with waves from all directions. In order to understand a bit better this variability we now look at spectra for individual days, overlaying analytical examples of attenuations with powers $n = 9.5$, $n = 4$ and $n = 3$.

It is particularly interesting to contrast the spectral shape on June 17 at the start of a wave event, with a very steep high frequency decay consistent with $5 < n < 11$, and the next day with $n \approx 4$ (Figure S20). In the chosen frequency range (above 0.08 Hz) the model performs well off the ice at buoys 623.

On the wave event from 27 to 28 June, n estimated from the buoy data changes from 2 to 6 before the energy level disappears below the noise level on the 29th whereas a constant attenuation rate would have given a strong signal. Steep spectra can also be found such as on July 1st (bottom line, figure S21), but n cannot be estimated because only $E(f_{14})$ is well above the noise.

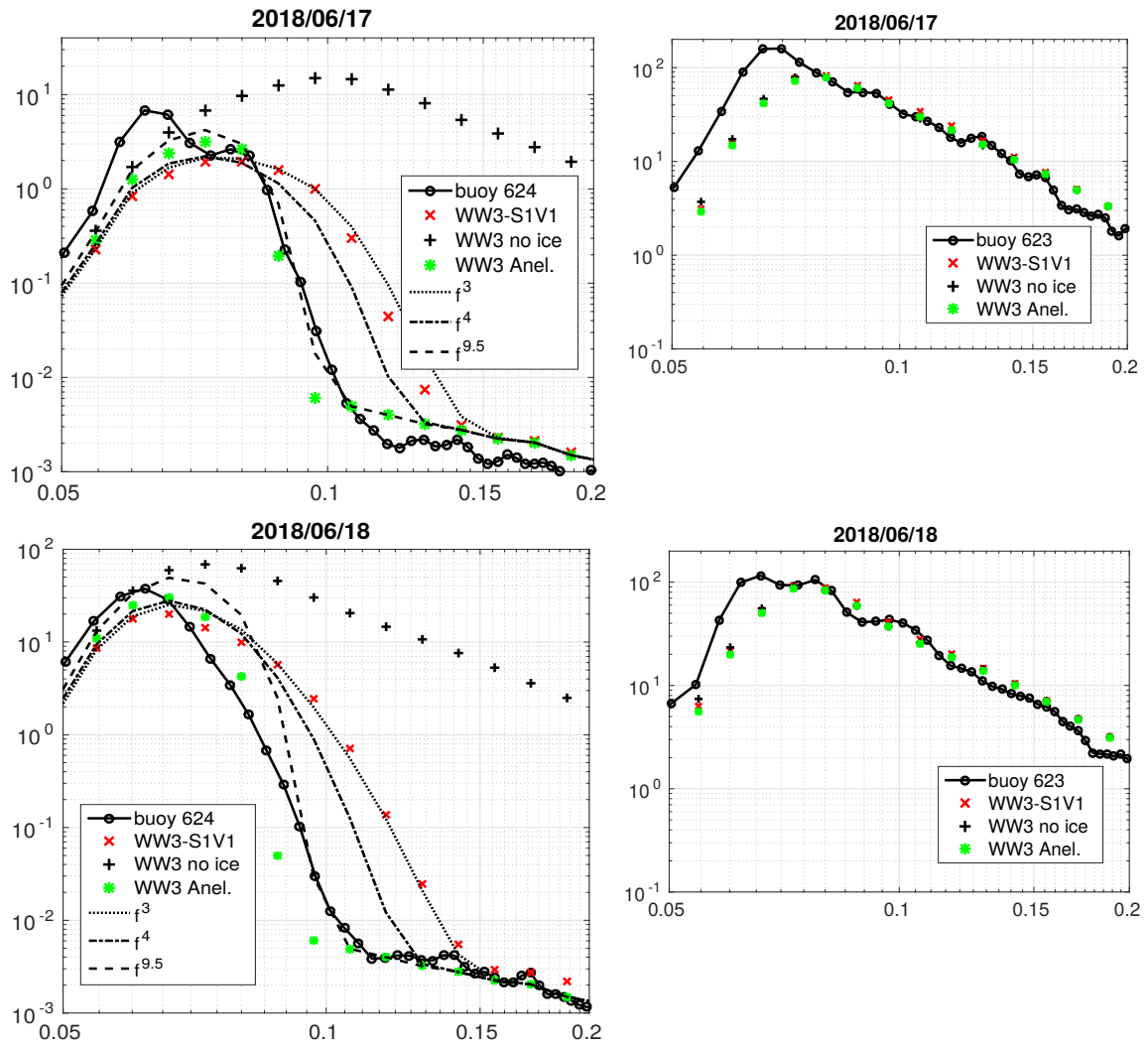


Figure S20. Measured and modelled spectra on June 17 (top) and June 18 (bottom), at buoys 624 (left) and 623 (right).

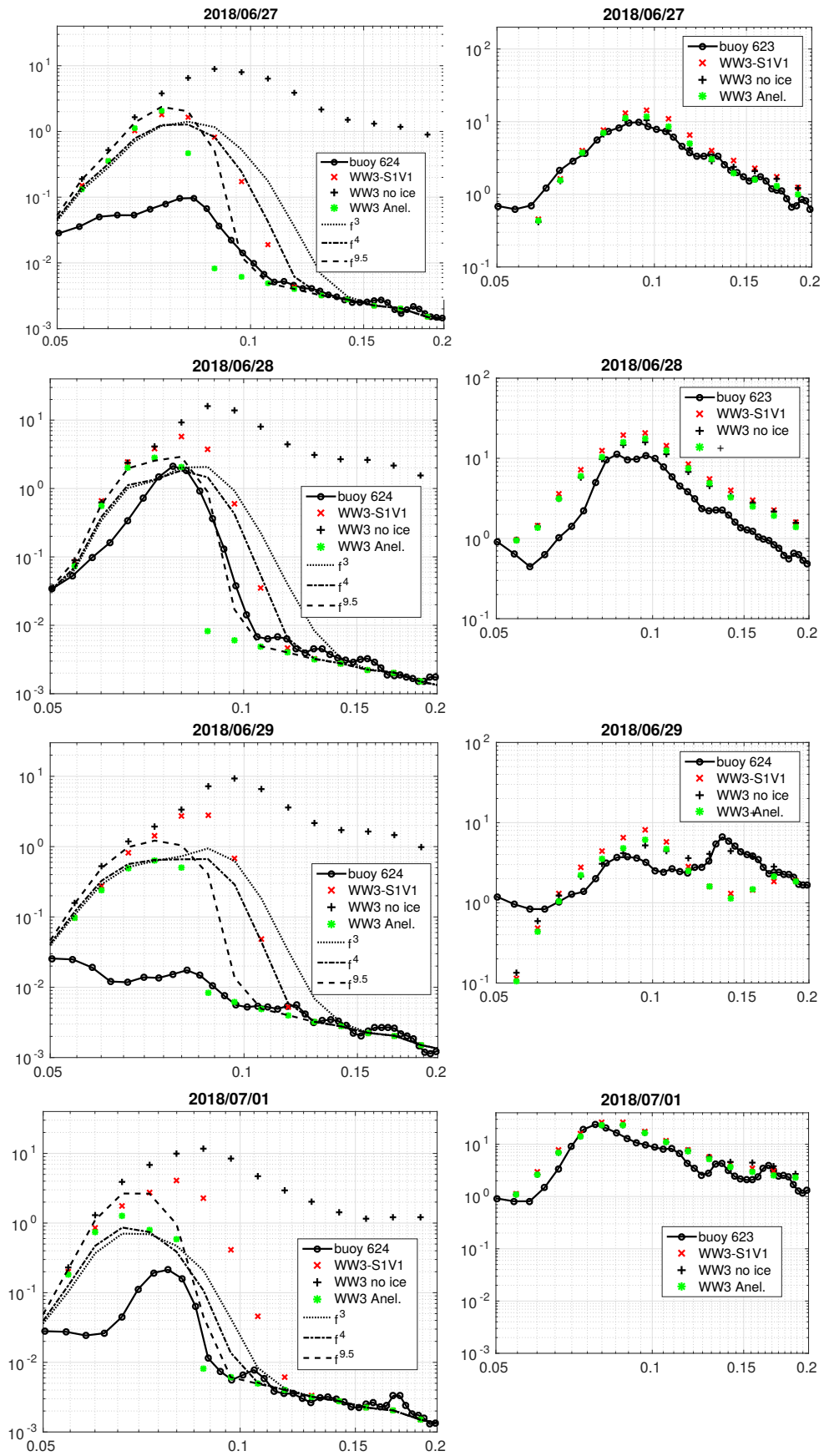


Figure S21. Measured and modelled spectra on (from top to bottom), June 27, June 28, June 29, and July 1st, at buoys 624 (left) and 623 (right).