Title:

Formation processes of sea ice floe size distribution in the interior pack and its relationship to the marginal ice zone off East Antarctica

Authors:

Takenobu Toyota[1*], Alison Kohout[2], and Alexander D. Fraser[1.3]

Submitted to Deep Sea Research II special volume on SIPEX-2

Affiliation

1*: Institute of Low Temperature Science, Hokkaido University

N19W8, Kita-ku, Sapporo, 060-0819, Japan

*corresponding author (toyota@lowtem.hokudai.ac.jp)

Tel: +81-11-706-7431   Fax: +81-11-706-7142

2: National Institute of Water and Atmospheric Resources,

10 Kyle St Riccarton, Christchurch, New Zealand 8011

(Alison.Kohout@niwa.co.nz)

3: Antarctic Climate & Ecosystems Cooperative Research Centre,

University of Tasmania, Private Bag 80, Hobart 7001, Tasmania, Australia

(adfraser@utas.edu.au)

Abstract

To understand the behavior of the Seasonal Ice Zone (SIZ), which is composed of sea ice floes of various sizes, knowledge of the floe size distribution (FSD) is important. In particular FSD in the Marginal Ice Zone (MIZ), controlled by wave-ice interaction, plays an important role in determining the retreating rates of sea ice extent on a global scale because the cumulative perimeter of floes enhances melting. To improve the understanding of wave-ice interaction and subsequent effects on FSD in the MIZ, FSD measurements were conducted off East Antarctica during the second Sea Ice Physics and Ecosystems eXperiment (SIPEX-2) in late winter 2012. Since logistical reasons limited helicopter operations to two interior ice regions, FSD in the interior ice region was determined using a combination of heli-photos and MODIS satellite visible images. The possible effect of wave-ice interaction in the MIZ was examined by comparison with past results obtained in the same MIZ, with our analysis showing: 1) FSD in the interior ice region is basically scale invariant for both small- (< 100 m) and large- (> 1 km) scale regimes; 2) although fractal dimensions are quite different between these two regimes, they are both rather close to that in the MIZ; and 3) for floes < 100 m in diameter, a regime shift which appeared at 20-40 m in the MIZ is absent. These results indicate that one role of wave-ice interaction is to modulate the FSD that already exists in the interior ice region, rather than directly determine it. The possibilities of floe-floe collisions and storm-induced lead formation are considered as possible formation processes of FSD in the interior pack.

Key words: Sea ice; Floe size distribution; Ice melting; Scale invariance

1. Introduction

Sea ice plays an important role in the polar climate system, due largely to its reduction of heat transfer from ocean to atmosphere and its high reflectance of solar radiation. Therefore the behavior of the sea ice extent particularly in the SIZ has a significant impact on the climate variability in the surrounding regions. Since the SIZ is composed of numerous ice floes with various sizes, FSD is an important parameter which controls the behavior of the SIZ. From a dynamical standpoint it is closely related to the deformation process of the ice cover, while thermodynamically it affects the melting rates of sea ice because smaller floes absorbs heat more efficiently from the surrounding seawater than larger floes (Rothrock and Thorndike, 1984). As for melting effects, it is suggested that FSD also contributes to the rapid decreasing trend in the Arctic summer sea ice extent (Asplin et al., 2012). According to their results, large expanses of open water introduce long fetch in the Arctic Ocean, leading to the storm-induced ice breakup, which accelerates the melting process. The effect of FSD on melting rate was shown to be significant for ice floes smaller than about 30 m (Steele, 1992).

To predict the retreat rates of the extent of the SIZ on a global scale, it is important to understand the melting processes in the MIZ, which is an outer fringe of the interior ice pack area. The MIZ is characterized by individual ice floes at typically lower ice concentration and vigorous wave-ice interaction that plays an important role in determining the FSD due to wave-induced flexural failure of ice (Squire, 2007; Squire and Moore, 1980; Wadhams et al., 1988). As a storm can induce wave-ice interaction even in the interior ice pack region in the Antarctic seas (Kohout et al., 2014), in this study we refer to the MIZ and the interior ice region as regions with comparatively lower and higher ice concentration, respectively (Fig.4). Since relatively small ice floes are dominant in the MIZ, FSD is a controlling factor of the melting processes. Given that FSD in the MIZ is determined by the interplay of penetrating waves with the preexisting sea ice, it is an important issue to clarify the FSD in the interior ice region and the effect of wave-ice interaction on the formation processes of FSD in the MIZ.

Recent studies revealed that FSD in the MIZ has a different regime for floes smaller than a few tens of meters (d_t m) compared with larger floes (Lu et al., 2008; Toyota et al., 2006, 2011). The cumulative number distribution, *N(d)*, defined as the number of floes per unit area with diameters no smaller than *d*, was found in both regimes to follow the power law, $N(d) \propto d^{-\alpha}$, indicating that FSD for both regimes is basically scale invariant. Yet the exponent α was shown to be quite different between these regimes. Whereas for $d > d_t$ α often exceeded 2, for $d < d_t$ α took significantly lower values ranging from 0.7 to 1.5 depending on the distance from the ice edge (Lu et al., 2008; Matsushita, 1985; Toyota and Enomoto, 2002; Toyota et al., 2006, 2011). This indicates that wave-ice interaction plays an important role in determining the FSD in the MIZ. It follows that understanding wave-ice interaction is requisite for the prediction of the retreating rate of sea ice extent on a global scale.

On the other hand, it was shown in earlier studies focusing on FSD in the interior ice region that *N(d)* follows a power law for floes larger than about 100 m and α often exceeds 2, similar to the case of $d > d_t$ in the MIZ (e.g., Holt and Martin, 2001; Rothrock and Thorndike, 1984; Weeks et al., 1980). However, since measurements of FSD for floes smaller than about 100 m are sparse, the properties of FSD covering a wide range of floe sizes in the interior ice region is not yet fully understood. Although

Steer et al. (2008) showed for floes in the interior ice region of the Weddell Sea in the melting season that FSD for $d < 20$ m had a different regime from that for $d > 20$ m, it is likely that FSD for smaller floes was much more affected by melting than by dynamical processes. Besides, field observations of wave activities in the MIZ have been very limited (Liu et al., 1991; Squire and Moore, 1980; Wadhams et al., 1988), with no concurrent observation of FSD made so far. Therefore, it still remains unclear how waves produce FSD in the MIZ, and how this differs from that in the interior ice region through wave-ice interaction, which may be one of the possible factors that has hampered the accurate prediction of sea ice extent retreat in numerical sea ice models (Holland et al., 2006).

To improve the understanding of the formation processes of FSD in the SIZ through wave-ice interaction, we planned the concurrent observations of wave activity and FSD from the Australian R/V "Aurora Australis" off Wilkes Land, East Antarctica during SIPEX-2 in late winter 2012. In this experiment, wave activity was observed using five buoys equipped with accelerometers on stable ice floes in the MIZ (see Kohout et al., 2014 for details). Since logistical reasons limited helicopter operations to only two interior ice regions about 250 km from the ice edge due to weather conditions (Fig.1), however, in this study we focus on FSD in the interior ice region by combining heli-photo data with MODIS channel 1 visible, 250 m resolution satellite images. Instead of direct measurements, we examine the effect of wave-ice interaction on FSD by comparing this study with previous results obtained in the MIZ off Wilkes Land in 2007 (Toyota et al., 2011) on the assumption that FSD is almost the same in the same region and in the same season. The result obtained from the buoys is used to interpret our analytical result of FSD. Ice thickness data were also obtained along the ship track with a video system (Toyota et al., 2004) to test theoretical studies that show ice thickness is by far the most important factor in determining the scattering and break-up of sea ice (Kohout and Meylan, 2008; Meylan, 2002).

The major purpose of this study is to i) detail the properties of floe size distribution in the interior ice region, ii) speculate on the effects of wave-ice interaction on FSD in the MIZ by comparing the results with those obtained previously in the MIZ of the same region, and iii) improve the understanding of the formation process of FSD in the MIZ. In all analyses, the property of scale invariance will be emphasized. The formation processes of FSD in the interior ice region will also be discussed based on the data obtained and the meteorological reanalysis dataset (ERA-Interim). To support our discussion, additional observational evidence from the expedition will be documented.

2. Data

During the SIPEX-2 expedition, FSD was produced from heli-borne camera photos and MODIS satellite images for the interior ice region. Ice thickness along the ship track was also monitored with a video system. Here the heli-borne photos, ice thickness video system, and the analytical procedure to obtain FSD from the MODIS satellite images will be outlined.

2.1 Helicopter observation

The SIPEX-2 expedition was conducted from the Australian icebreaker R/V “Aurora Australis” for the period from September 15 to November 16, 2012 off East Antarctica. The expedition was an interdisciplinary project, including physical oceanography, sea ice physics, chemistry and biology (Meiners et al., this issue). The ice concentration in the study region from AMSR-E is shown in Fig 2. During this expedition, the ship navigated within the sea ice zone from September 23 to November 10. Floe size observations were conducted with a heli-borne digital camera (GoPro) in the two interior ice regions, both located about 250 km inward from the ice edge: around 63.74^{o}S 119.70^{o}E on September 25 and around 63.86^{o}S 115.69^{o}E on November 5. The tracks of the ship and helicopter and ice concentrations on those days are shown in Fig.1 and Fig.2, respectively. During the observations, the weather was clear and there was only a small amount of cloud. Around these areas the dominant floe size was larger than a few km and floes smaller than 100 m were only seen between large floes. In addition to a heli-borne camera, an approximate FSD, unsuitable for quantitative analysis, was recorded every minute with a forward-looking camera installed on the upper deck of the ship. According to this measurement, the dominant floe sizes in the MIZ were about 2-3 m, 5-6 m, and 10-20 m in the zones of 0-70 km, 70-100 km, 100-190 km from the ice edge (61.0^{o}S 122.0^{o}E), respectively.

A heli-borne digital camera, installed on the step of the helicopter, took the photos of the ice conditions directly below the helicopter every five seconds along each flight track with a fish-eye lens (view angle: 170 degrees) to cover a broad area. During the flights, the position and altitude were recorded every 10 seconds with GPS (Garmin, GPSMAP196) with a nominal accuracy of < 15 m. The helicopter flew at several stable altitudes around 400 m and 600 m on September 25 and around 800 m and 1100 m on November 5. The fish-eye lens distortion was corrected using PC software (Adobe Photoshop Elements 11). To determine the scale of each image, the ship’s hull was embedded into an image at an altitude of 789 m on November 5. The pixel scale was then determined for each image. Since the dominant floe size was much larger (> 1 km) than the camera view area (~1 km) in this region, suitable images were limited. Therefore, two representative images, which contain a sufficient number of individually distinguishable floes, were selected for each flight. Since the number of the images is limited, we should keep it in mind that the result obtained is a case study. The width, length, and altitude of each image are summarized in Table 1. From this table, the horizontal resolution is estimated to be between 0.6 m and 0.8 m for each image. The total area of the four images amounted to 12.4 km^2.

2.2 Ice thickness

Ice thickness measurements were conducted with a downward-looking video camera installed on the ship's rail that continuously recorded the ice conditions along the ship's hull. Post-cruise, the video images were downloaded to PC and ice thickness was measured with the PC software 'Micro Analyzer' (Japan Pola Digital Co.) for each ice floe that was overturned alongside of the hull. The scale was determined by lowering a measuring stick onto the ice surface while the ship was stationary, following Toyota et al. (2004). The measurement error is less than a few centimetres. In this way, ice thickness data were obtained for three hours per day while the ship was navigating. The hourly mean ice thickness distribution is shown in Fig.3. However, it should be noted that this method is designed basically for the measurement of undeformed ice thickness. The most deformed ice, which is hard to overturn, is beyond the measuring capability of this method, and snow depth is also included in ice thickness because it is sometimes hard to determine the boundary between snow and sea ice. Even so, since the thickness of undeformed ice is related to the strength of sea ice, the obtained data provide useful information to interpret FSD. A total of 1784 ice thickness measurements along the ship track were made and the average thickness was 0.59 ± 0.25(s.d.) m.

2.3 MODIS satellite imagery

Daily MODIS/ Aqua or Terra satellite images with a nominal horizontal resolution of 250 m were used to analyze large ice floes in the interior ice region near the observation area. MODIS Level 1B Channel 1 visible imagery was projected to a polar stereographic projection of 250 m resolution, covering the region 60.00°S to 66.66°S, 110.00°E to 127.90°E (750 km x 875 km) (Fig.4). Since MODIS images are subject to the presence of cloud, we selected three images (September 24, October 4, and November 5) where the observation area was mostly cloud free and suitable for analysis. Then, to examine regional properties of floe size distribution, four sectors (A, B, C, and D in Fig.4) were extracted from each image for the analysis of the properties of individual ice floes in each sector. The position of each sector was semi-flexible so that it could contain as many floes as possible and avoid clouds. Therefore, the area of each sector changes somewhat for each day (Table 2). The rationale for selecting each sector is as follows: sectors B and C were selected to view the spatial variation of FSD while moving westward across the sea ice area relatively close to the MIZ, and sectors A and D were selected to examine the sea ice properties further south in the deep inner pack. Fortunately, the selected dates of September 24 and November 5 nearly coincided with the days when the heli-borne measurements were conducted, allowing the combination of both datasets for each day which produced a wider range of floe sizes: from about 4 m to about 10 km. Additional sampling from three repeats over the period provides evidence of the stability of FSD in the interior region at almost the same distance from the ice edge.

3. Image processing

Analysis was essentially the same for the two datasets of photography and satellite imagery, with the image processing technique developed by Toyota et al. (2006). Each ice floe was extracted according to its brightness, and then its area (*A*), perimeter (*P*), and maximum/ minimum caliper diameters ($d_{\max}/d_{\min}$) were measured using the PC software Image-Pro Plus ver.4.0 (Media Cybernetics Co.). In this study, floe size (*d*) is evaluated as the diameter of a circle that has the same area as that of the floe: $d=\sqrt{4A/\pi}$. We adopted our definition because of its simplicity in calculation. While other definitions were used in past studies, such as mean caliper diameter (d_{mc}: the average of caliper diameters in all orientations) following Rothrock and Thorndike (1984) and Lu et al. (2008) and the side of the square that has the same area as that of the floe (Steer et al., 2008), it was proved that these definitions of floe size are highly correlated (Rothrock and Thorndike, 1984).

In this analysis, the key is to precisely determine the edge of individual ice floes. The details are described by Toyota et al. (2006, 2011). Grey areas caused by nilas rafting which sometimes appeared between floes in the interior ice region was carefully excluded because such sea ice cannot be regarded as an independent ice floe. Excluded from the analysis were those floes which:

1) are intersected by the boundary of the image;

2) have an area less than 30 pixels; or

3) have aspect ratios (d_{max}/d_{min}) exceeding 5.

Criterion (2) was included to examine the shape property of ice floes. According to this, the lower limits of the floe size are estimated about 4 m for photography and 1545 m for satellite imagery (Tables 1 and 2). Criterion (3) was included because extremely distorted ice floes are unsuitable for the definition of floe size. The fraction of excluded floes by this criterion is only 0.1% and 0.6% for heli-photos and MODIS images, respectively, and thus does not affect the result significantly. An example demonstrating this analytical process for photography and satellite imagery is shown in Fig. 5. Floes that appear to be identifiable but left unanalyzed in Fig.5 are mostly those which we judged not to be independent or have unclear outlines when we magnified them. If failure in identifying floes may occur, this effect is considered to be biased to smaller floes due to the horizontal resolution of the images. To reduce subjectivity, we repeated the analysis twice for all the images. Consequently, the total number of ice floes analyzed amount to 4,247 for photography and 8,994 for satellite imagery.

4. Results

The extracted floes shown in Fig.5 demonstrate that FSD appears to be significantly different between MODIS images and heli-borne photos. Whereas several large floes and a number of relatively small floes are coexisting with a spacing of a few kilometers between long linear leads in the MODIS images, a broader size distribution is present more tightly in a smaller area in the photos taken from the helicopter. The shape of individual floes also appears to be different between these two datasets. Floes in the heli-photo look somewhat more rounded than those in the MODIS images. These features suggest a difference in formation processes between these scales.

To show these different properties from statistics, the FSD was expressed as the cumulative number distribution *N(d)*, defined by the number of floes per unit area with size no smaller than *d*, following past studies (e.g. Rothrock and Thorndike, 1984). The results are shown in Fig. 6 for heli-photos, where *N(d)* obtained individually at two different times are averaged together for each day, and in Fig. 7 for the MODIS images, where the results obtained at the four sectors are all averaged together for each day. In both cases, the graphs are drawn only for the range where *d* is larger than the lower limit and *N(d)*>5. The latter condition for the upper limit was introduced because the upper few samples tend to have extremely large sizes. It is found in both figures that while a slight deviation from a straight line is found especially for MODIS images, *N(d)* basically behaves like $d^{-\alpha}$. This indicates that floe size distribution is basically scale invariant over a wide range of 4 m to 10 km. However, the exponent takes significantly different values between two datasets. The exponent α for the heli-photos (hereafter, referred to as R_S, where S denotes small scale) is estimated by the least squares method for 10 < *d* < 60 m, where the effect of upper truncation (shown later) is small, to be 1.41 ± 0.09 for September 25 and 1.27 ± 0.10 for November 5, with a significance level of 95%. For MODIS images (hereafter, referred to as R_L, where L denotes large scale) α is estimated to be 3.10 ± 0.46 for September 24, 2.93 ± 0.35 for October 4, and 2.90 ± 0.34 for November 5. A notable feature, observed in the past results in the MIZ, is that for floes less than 100 m (R_S) a clear transition size exists at which α

changes significantly (Toyota et al., 2006, 2011). Such a feature cannot be seen in Fig. 6. Although α has a decreasing trend for floes larger than about 70 m in one of the two lines in Fig. 6a, it is likely that this comes from the upper floe size truncation caused by the limited area (Burroughs and Tebbens, 2001).

To see this effect more clearly, the lines of the upper truncated power law, $M(d) = A \cdot \left(d^{-\alpha} - d_{top}^{-\alpha}\right)$, fitted to the mean data following the General Fitting Function (GFF) method of Burroughs and Tebbens (2001), and the underlying power law, $N(d) = A \cdot d^{-\alpha}$, are also drawn in Figure 6, where d_{top} is the floe size of upper truncation and was given as 175 m for Fig.6a and 333 m for Fig.6b from observation. The estimated exponents α are 1.40 for Fig.6a and 1.27 for Fig.6b, close to the values obtained from the least square method above. It is shown from this figure that both lines for the underlying power law fit with the observed lines well, indicating that the decreasing trends in the observed cumulative number distribution can be explained well by the truncation effect. We applied this method also for R_L. Figure 7 shows that overall upper truncated power law fits with the observed cumulative number distribution and the estimated exponents α, 3.09 for Fig.7a, 2.92 for Fig.7b, and 2.89 for Fig.7c, are close to the values obtained from the least square method. These results indicate that FSD is basically scale invariant over both ranges of R_S and R_L.

The geometry of ice floes is also an important part of wave-ice interaction process, as shown by Meylan (2002), and provides useful information on formation processes of ice floes. Here the floe geometry is examined from the ratio of maximum (d_{max}) and minimum (d_{min}) caliper diameters. The results are plotted for individual floes in Fig. 8. It is shown that while they are correlated well for both R_S and R_L, the correlation is much more remarkable for R_S (correlation coefficient = 0.98) than for R_L (0.83). On average the aspect ratio (d_{max}/d_{min}) is estimated as 1.84 ± 0.52 (sd) for R_S and 1.93 ± 0.64 for R_L. It is interesting to note in Table 3 that the aspect ratio for R_S takes almost the same values as that for the MIZ of other seasonal ice zones. This suggests that the floe formation process may be common among these regions. A somewhat smaller value (~1.63) for the MIZ off Wilkes Land, which is rather close to 1.5-1.6 for multi-year ice (Hudson, 1987), might be explained by the higher wave activity off Wilkes Land, induced by stronger intensity of the cyclone system in winter, compared with the other regions (Jones and Simmonds, 1993). It is plausible that higher wave activity increases the roundness of floes through collision processes, but not as much as expected for multi-year ice which experiences significant amount of collision between floes in the interior ice pack.

Next, we examine the temporal variation of the FSD for R_L during the observation period based on MODIS images on Sep 24, Oct 04, and Nov 05. As shown earlier, the exponent α averaged for all the sectors was almost constant during the period. However, the pattern of FSD is highly variable within each sector. As an example, the variation of ice conditions within sector B is shown in Fig.9. It is seen that the pattern of FSD changed drastically and decreased somewhat in size with time. The mean floe size of sector B decreased from 3169 m on Sep 24 to 2759 m on Oct 04 and 2406 m on Nov 05. Corresponding to this temporal change, the slope for sector B and D became steeper on Nov 05 (Fig.7). On the other hand, the slopes for sector A and C became gentler, resulting in almost the same slope when averaged over all four sectors. The mean floe size averaged for all four sectors decreases from 3235 m on Sep 24, to 2782 m on Oct

04, and remains at 2780 m on Nov 05. Storm events in which the wave significant height exceeded two meters occurred once between Sep 24 and Oct 04 and six times between Oct 04 and Nov 05 (Kohout et al., this issue), which does not necessarily correspond to the change of mean floe size. This is possibly because this result does not correspond to the exact temporal evolution of FSD due to the advection of the sea ice area. Even so, it is interesting to note that although the pattern and mean size of FSD changed drastically on a local scale (< 100 km), it was kept almost constant on a larger scale (~400 km) in the interior ice region. This is consistent with the result of Holt and Martin (2001) which showed that the exponent α of FSD in the interior ice region of the Arctic Ocean (horizontal scale > 300 km) was not affected by the passage of storms, although the mean floe size decreased.

The above results are summarized in Table 3, including past results obtained from the MIZ of the Sea of Okhotsk, the Weddell Sea, and off Wilkes Land for comparison. In Table 3, for convenience the results obtained for the regimes of $d < d_t$ and $d > d_t$ in the past studies are listed in the column R_S and R_L, respectively.

The characteristics are summarized as follows:

1) For both R_S and R_L, FSD is basically scale invariant.
2) The exponent α is much less than 2 for R_S, while around 3 for R_L. Both values are rather close to those found in the MIZ in the past observation;
3) For floes less than about 100 m (R_S) a regime shift which appeared in the MIZ from the past observations does not occur in the interior ice region;
4) On average the aspect ratio of individual ice floes is not significantly different between R_S (1.84) and R_L (1.93);

5) For R_L, the exponent α of FSD averaged for all sectors was nearly stable during the period although it varied significantly within each sector.

Points 1 and 2 are important because as Rothrock and Thorndike (1984) pointed out, if α for R_S is larger than 2, total area of floes would become infinite. Points 2 suggests that the formation processes of FSD are different for R_L and R_S, which will be discussed in the next section. There may be a possible effect of failure to identify all floes especially for MODIS images (Fig.5b). However, considering that this effect seems to be biased to smaller floes, the real α for R_L would have rather larger values than our estimates. Therefore, we do not consider that this effect can alter our result essentially. Although Perovich and Jones (2014) pointed out that a constant decrease in floe size due to lateral melting can cause a decrease in α for smaller floes, we consider that this effect is small because our observation was conducted in late winter with the air temperature ranging mostly from -20 to -5^{o}C before significant melting began.

It is interesting to note in Table 3 that α for R_S takes a somewhat smaller value for thicker sea ice, suggesting that the strength of sea ice is related to α. Point 3 indicates that the regime shift which appeared for d < 100 m in the MIZ is closely related to wave activities and that wave-ice interaction plays an important role in determining the transition size d_t. The detail will be discussed in the next section. Point 4 means that floes are not of a circular shape but usually distorted. This seems reasonable when we consider the obvious effects of swell break-up in the interior tend to have some anisotropy in the aspect ratio (e.g. Worby et al., 1998). Point 5 suggests that the statistics of FSD in the interior ice region may be maintained on a large scale.

5. Discussion

5-1. Formation processes

We now examine the formation processes for the individual scales of R_S and R_L in the interior ice region. Firstly, we discuss it for R_S from geometric properties (Table 3). It is noticeable in Fig.5a that a number of ice floes are closely packed and some floes are fractured into halves or quarters, apparently due to collision with neighboring floes. Since storm-induced waves can penetrate into the interior ice region (Kohout et al., 2014), it might be possible that the fracturing was induced by waves or swells. However, considering that the surrounding large ice floe was not broken, fracturing due to collisions is more probable. Of interest is that the fractal dimension of this regime (1.3-1.4) is close to the value of 1.31 for a typical fractal geometry known as the Apollonian gasket. A key geometric feature of the Apollonian gasket is that each circle is in contact with the three surrounding circles at any scale, which is similar to the appearance of Fig. 5a. This suggests that the major formation process of FSD for R_S relate to the collisions between floes.

If the given probability of a break-up process by collision is independent of scale, it is natural that the floe size distribution produced becomes scale invariant. Toyota et al. (2011) attempted to explain how the fractal dimension α for R_S in the MIZ is determined through wave-ice interaction by introducing a "fragility" parameter, *f*, which represents the likelihood of break-up as a function of ice strength relative to wave activity, and correlated α with *f*. Our result suggests that a similar concept can be applied to R_S in the interior ice region. In this case, fragility is considered to represent the likelihood of break-up due to collision between floes. If this is true, there is no reason for producing a regime shift for R_S in the interior ice region where wave activity is usually quite small. This explains point 3 presented in the previous section. The slight difference in α between Sep 25 (1.41) and Nov 05 (1.27) may be explained by the difference in mean ice thickness (0.37 m and 0.79 m, respectively). It is plausible that thicker ice has tougher strength and tends to reduce the break-up of ice floes, resulting in a lower fractal dimension.

Of interest is that these values of α are close to that for R_S obtained in the MIZ (Table 3). The floe geometry (aspect ratio) in the interior ice region was also shown to be close to that in the MIZ, although the floes in the MIZ off Wilkes Land have a somewhat more rounded shape. This means that the original form of the FSD for R_S in the MIZ was already created in the interior ice region before wave-ice interaction influences the MIZ significantly. It is consistent with the idea that the more rounded floes in the MIZ off Wilkes Land might be attributed to the more vigorous wave activity compared with other MIZs and the interior ice region.

Next we discuss the formation process for R_L in the interior ice region. A notable feature in Fig.9 is a number of long linear leads are running between floes with a spacing of 1 to > 10 km in various directions. Since the spacing of leads almost coincides with the floe size analyzed, it is natural to think that occurrence of such leads is relevant to the formation of FSD for this regime. If the deep water approximation is applied, a wave length of 10 km corresponds to a period of 125 seconds (= $\sqrt{g/2\pi \cdot 10^4}$), where g is acceleration due to gravity. Given that wave activity is usually quite small at such a long period in the region more than 200 km inward from the ice edge (Squire and Moore, 1980), it is unlikely that wave activity is responsible for the lead formation on this scale unless storm-induced waves are involved. Kohout et al. (this issue) showed that storm-induced waves can penetrate into the interior ice region and create ice breakup even when the wave height becomes quite small. But as a major factor we consider that dynamic failure due to deformation processes of sea ice should work efficiently for initially formed cracks and lead formation, as shown by Erlingsson (1988) and Schulson and Hibler (1991). Schulson and Hibler (1991) suggested that the enhancement mechanism of initially formed cracks due to compressive forcing can produce a scale invariant pattern of leads on a scale larger than tens of kilometers, which may explain the scale invariant property of FSD for R_L to some extent.

Besides these effects, here we point out the effect of wind for the lead formation. As discussed by Coon and Evans (1977), the elastic property of sea ice is insufficient to induce ice cracking through wind forcing. Even so, there is a possibility that wind is involved in the development of leads. As an example, Fig.10 shows a MODIS image on October 23, 2012, with the wind pattern obtained from the European Centre for Medium-Range Weather Forecasts Interim Re-analysis (ERA-Interim) dataset (1.5^o x 1.5^o, four times per day). This figure demonstrates that major linear leads are aligned in parallel with a spacing of a few to 30 km normal to the wind direction, showing the relevance between wind and lead formation. In the magnified figure we find a number of cracks are running in various directions, possibly caused by wind with various directions in the past.

The possible wind effect is considered as follows. Initially, we start with one large ice floe (> a few km) which is formed by the aggregation of a number of relatively small ice floes (< 1 km) with various thickness and/or sizes. When a strong wind blows over this region associated with the passage of a cyclonic system, breakup may occur between small floes due to the penetration of storm-induced waves and/or the dynamic effect of deformation. Suppose that the ice thickness is significantly different between neighboring broken floes (Fig.11). Since the wind forcing at the upper surface would make little difference between aggregating small floes, a different ocean drag forcing acts on the bottom surface between these floes and consequently differential ice velocity is produced, which works to increase the lead width between these small floes (Fig.11). Since the spatial variation of strong winds accompanying a cyclonic system usually occurs at a scale larger than 100 km, the leads would appear linearly with a scale of a few tens of kilometers. It is likely that the change of wind direction associated with the passing of cyclonic systems induces various multi-directional cracks and leads. A lead development event induced by wind that occurred during the expedition will be documented in the next section. Thus, when wind works efficiently to develop leads, it is possible that wind also plays a role in forming FSD in R_L. In this case, the scale invariant property may be produced by a combination of the flexural fracture due to the penetration of storm-induced waves (Kohout, 2014), the reconnection of separated ice floes, and the break-up due to floe-floe collisions. The fact that the aspect ratio of this regime is not significantly different from that in the R_S regime (Table 3) suggests that break up due to collision works effectively as well.

Finally, we discuss the role of wave-ice interaction in forming FSD in the MIZ. To see the difference between the two datasets more clearly, the combined figures for Sep 24/25 and November 05 are shown in Fig.12. In the figures thick solid lines denote the averaged data for each dataset. It should be kept in mind that the difference in the study area between the two datasets is not taken into account. If this effect is included, the line for R_S will be shifted somewhat downward in Fig.12. Even so, the meeting point of two extended lines with different slopes lies at around 1 km. Thus it is found that the major formation process and properties of floe size distribution in the interior ice region differs for floe sizes above and below 1 km. The value of α for R_S in the interior ice region is almost unchanged in the MIZ and the major difference between the MIZ and the interior ice region is the presence of a regime shift at d = 20-40 m. These facts suggest that the major role of wave-ice interaction is in creating a transition size (d_t) of 20-40 m in the MIZ by modulating the floe size distribution for $d > d_t$ in the interior ice region. As a modulating process, the break-up of floes due to flexural forcing by waves would work effectively. Considering that α for $d > d_t$ in the MIZ is rather close to that for R_L in the interior ice region, it is possible that there may be a similarity in the formation processes between these regimes and that the major role of wave-ice interaction is to enhance them.

So why is the modulating process limited to larger floes ($d > d_t$) and what determines d_t? Concerning the response of sea ice to swell, it was shown from theoretical studies that when the ice floe size is smaller than 100 m, flexural failure becomes difficult for any period or amplitude of swell (Fox and Squire, 1991; Higashi et al., 1982; Meylan and Squire, 1994). And Mellor (1986) theoretically derived the minimum ice length at which flexural failure will occur as a function of Young's modulus, Poisson's ratio, and ice thickness of sea ice. Here the minimum ice length for wind-induced fracture is estimated as 20-40 m, corresponding approximately to d_t, and is independent of the degree of wave activity. This explains why α for R_S in the interior ice region remains unchanged even if the MIZ boundary expands poleward due to melting in the MIZ and that d_t is commonly seen in the MIZ of the SIZ. In theoretical studies, Toyota et al. (2006 and 2011) hypothesized that flexural failure by ocean swell plays an essential role in producing a regime shift. Our observational result that a regime shift was absent for floes less than 100 m in the interior ice region seems to support their hypothesis implicitly.

5-2. Ice cracking induced by wind

For the period October 26 to November 4, the R/V "Aurora Australis" was completely stuck in a thick (5-6 m), large (> 1 km) sea ice floe at around 65^{o}S 117^{o}E (Fig.1). Then with the passing of a cyclonic system near this area from northwest to southeast, a persistent southerly wind increased in strength from 5 m/s at midnight to 20 m/s at 15:00 (local ship time) on November 4, according to the wind data recorded on the ship. With the ship pointing north (340 degrees), the wind was blowing from stern to bow. At 14:30, when the southerly wind speed reached nearly 18 m/s, a linear crack running in the east-west direction, normal to the wind direction suddenly appeared in sight about 800 m ahead of the ship. Images taken from the top of the ship showed that the lead became prominent at 15:30. On the following day (Nov 5), the width of the lead was estimated as 770 m at 13:38 from heli-borne imagery. Taking into account that the wind direction turned from southerly to westerly at 03:30 (local time) on Nov 5, the opening rate of the lead can be estimated to be 1.8 cm/s (= 770 m / 12 hours). During the opening of the lead, the average speed of the southerly wind was 10.4 m/s. While stuck in the ice, the ocean condition was calm. Although due to lack of the observational data we cannot say assuredly what caused the ice crack initially, it might be possible that associated with the passage of a storm over the open sea area north of the region on November 3 to 4, dynamic failure due to the deformation processes of sea ice, as shown by Erlingsson (1988) and Schulson and Hibler (1991), or breakup due to the penetration of storm-induced waves, as shown by Kohout et al. (this issue), were involved in the crack event. After the lead developed, our ship repeatedly rammed the ice for about 1.5 days and eventually escaped from the thick floe at 11:04 on November 6.

Based on the above data, we attempt to estimate the ice conditions. According to the discussion in the previous section, ice thickness is expected to be significantly different between the ice floe on the other side of the lead (F_1) and the ice floe where R/V "Aurora Australis" became stuck (F_2). Now we suppose a steady condition in which the drag forcings of air-ice (τ_a) and ice-ocean (τ_w) are balanced (Fig.11). Although the internal stress between ice floes usually plays an important role in the balance equation in the ice pack region, we set this assumption because wind with almost uniform speed and direction was blowing over this region on a scale of a few hundreds of kilometers during the event, and so the differential velocity between floes which causes the internal stress is considered to be relatively small. From the force balance equation ($\tau_a = \tau_w$), Eq.1 is derived based on the assumption of no ocean current and U_a (wind speed) $\gg V_i$ (ice velocity):

$$V_i = \sqrt{\frac{\rho_a \cdot C_{Da}}{\rho_w \cdot C_{Dw}}} \times U_a \qquad (1)$$

where ρ_a and ρ_w are densities of air and seawater, respectively. C_{Da} and C_{Dw} are drag coefficients between air-ice and ice-ocean, respectively. Here we assume that C_{Da}

is common between F_1 and F_2 and relative ice velocity is produced by the difference in C_{Dw} between F_1 and F_2. Ice velocity for F_1 and F_2 can be described as follows:

$$V_{i1} = \sqrt{\frac{\rho_a \cdot C_{Da}}{\rho_w \cdot C_{Dw1}}} \times U_a, \quad V_{i2} = \sqrt{\frac{\rho_a \cdot C_{Da}}{\rho_w \cdot C_{Dw2}}} \times U_a \qquad (2)$$

In Eq.2 we set $C_{Dw2} = k\, C_{Dw1}$, and 3×10^{-2} was given to $\sqrt{\frac{\rho_a \cdot C_{Da}}{\rho_w \cdot C_{Dw1}}}$ as a typical Nansen number for Antarctic sea ice (Lepparanta, 2005). Then Eq. 3 can be derived:

$$1 - 1/\sqrt{k} = (V_{i1} - V_{i2})/U_a \times 1/(3 \times 10^{-2}). \qquad (3)$$

By substituting observed values of 10.4 m/s and 1.8 cm/s for U_a and $V_{i1} - V_{i2}$ in Eq.3, k=1.13, i.e. $C_{Dw2} = 1.13 \times C_{Dw1}$ is obtained. According to Lu et al. (2011), the ice-water drag coefficient C_{Dw} for ridged ice is a function of ice concentration and the spacing and depth of ridge keels. According to their results, for IC= 90 % and the same keel spacing, k = 1.13 is achieved when the keel depth for F_2 is about 1.5 times greater than that for F_1. Considering that in reality the internal stress among ice floes may work additionally to reduce the ice motion, it is deduced that this estimation provides the minimum variation of the ice conditions. Given the considerable variation of mean ice thickness depending on ice floes in this region, as shown to be 1.4 to 3.6 m obtained from autonomous underwater vehicles during this expedition by Williams et al. (2015) and 0.6 to 2.2 m obtained from drilling during the SIPEX expedition in 2007 by Worby et al. (2011), this estimate seems plausible and suggests that lead development induced by wind may contribute to the formation process of floe size distribution for R_L. Paradoxically, we might have been able to escape from the thick ice because the floe was significantly thicker than the surrounding ice floes.

6. Conclusion

To elucidate the properties of FSD in the interior ice region and its relationship to FSD in the MIZ, the observation of FSD was conducted in the interior ice region off East Antarctica in late winter 2012, using a helicopter-borne digital camera. Heli-photos were used for the analysis of floes smaller than 100 m and MODIS images were also used for floes larger than 1 km. By combining these two datasets, we obtained the properties of FSD in the interior ice region over a wide range in this area. Ice thickness data were obtained along the ship's track with a video system. These data were used to interpret the properties of FSD. The likely impact of wave-ice interaction on FSD was examined by comparing the result of this study with the past result obtained in the MIZ of the same region. As a result, it was revealed that:

(a) For both floe size regimes (< 100 m and > 1 km) FSD is shown to follow a power law, $N(d)=\beta \cdot d^{-\alpha}$, indicating that both are basically scale invariant.

(b) However, the values of the exponent α, corresponding to fractal dimension, are quite different between these two regimes: 1.3-1.4 for floes < 100 m and 2.9-3.1 for floes > 1 km. These values are both rather close to those of two regimes obtained for sea ice floes in the MIZ in the past studies.

(c) The regime shift which was found at a floe size of 20-40 m in the MIZ is absent in the interior ice regions.

(d) Based on the observational evidence, the major formation process of FSD in the interior ice region is deduced to be the break-up due to collision between floes for d < 100 m and lead formation possibly induced by ice deformation, penetrating waves and wind for d > 1 km.

Among these results, point (b) indicates that the original form of FSD in the MIZ is already created in the interior ice region. Therefore it is deduced that the role of wave-ice interaction is to modulate the FSD that already exists in the interior ice region until the boundary size of the two regimes decreases down to 20-40 m rather than to create a new FSD in the MIZ. If this is the case, point (c) supports the hypothesis proposed by Toyota et al. (2006 and 2011) that the transition size is closely related to the minimum length of sea ice that can cause flexural failure due to ocean wave (Mellor, 1986). Point (d) indicates that the behavior of ice floes is dynamic even in the interior ice region where wave-ice interaction is usually quite small.

Finally, our results suggest that the FSD in the MIZ of the SIZ is closely related to that of the interior ice region via wave-ice interaction. It should also be kept in mind that this is speculative and that continuous observations of the evolution of the FSD should be undertaken to determine if this is true. This means that to understand the formation process of FSD in the MIZ, we need expanded studies to clarify the behavior of sea ice floes across the whole seasonal ice zone, including wind-ice interaction, wave-ice interaction, and ice thickness distribution. Since the FSD in the MIZ is potentially one of the controlling factors of retreating rates of sea ice extent on a global scale, further investigation is required to understand the polar climate system.

Acknowledgments

The authors deeply thank all the crew and scientists of R/V "Aurora Australis" for their support during SIPEX-2. Special thanks are given to the chief scientist, Dr. K. Meiners for his dedicated leadership. Discussion with Prof. K. Golden, Dr. R. Massom and Prof. H. Mitsudera and proof reading by Dr. Guy Williams are also acknowledged. Comments by two anonymous reviewers were very helpful to improve the manuscript. Image processing was carried out using Image Pro Plus ver.4.0. This work was supported partly by FY 2011 Researcher Exchange Program between Japan Society for the Promotion of Science and the Royal Society of New Zealand, partly by JSPS KAKENHI 24·510001 and 25·03748 [Grant-in-Aid for Scientific Research], and partly by a New Zealand Foundation of Research Science and Technology Postdoctoral award to A.K., the Marsden Fund Council, administered by the Royal Society of New Zealand, NIWA through core funding under the National Climate Centre Climate Dynamics programme, the Antarctic Climate and Ecosystems Cooperative Research Centre and the Australian Antarctic Science project 4073.

References

Asplin, M.G., R. Galley, D.G. Barber, S. Prinsenberg, 2012. Fracture of summer perennial sea ice by ocean swell as a result of Arctic storms. Journal of Geophysical Research, 117, C06025, doi:10.1029/2011JC007221.

Coon, M.D., Evans, R.J., 1977. On wind-induced cracking of sea-ice sheets. Journal of Glaciology, 18(78), 152-154.

Burroughs, S.M., Tebbens, S.F., 2001. Upper-truncated power laws in natural systems. Pure Applied Geophysics 158, 741-757.

Erlingsson, B., 1988. Two-dimensional deformation patterns in sea ice. Journal of Glaciology, 34(118), 301-308.

Fox, C., Squire, V.A., 1991. Strain in shore fast ice due to incoming ocean waves and swell. Journal of Geophysical Research 96 (C3), 4531-4547.

Higashi, A., Goodman, D.J., Kawaguchi, S., Mae, S., 1982. The cause of the breakup of fast ice on March 18, 1980 near Syowa station, East Antarctica. Proc. 4th Symposium on Polar Meteorology and Glaciology, National Institute of Polar Research, Tokyo, Japan, pp. 222-231.

Holland, M.M., Bitz, C.M., Tremblay, B., 2006. Future abrupt reductions in summer Arctic sea ice. Geophysical Research Letters, 33, L23503, doi:10.1029/2006GL028024.

Holt, B., Martin, S., 2001. The effect of a storm on the 1992 summer sea ice cover of the Beaufort, Chukchi, and East Siberian seas. Journal of Geophysical Research 106 (C1),1017-1032.

Hudson, R.D., 1987. Multiyear sea ice floe distribution in the Canadian Arctic Ocean.

Journal of Geophysical Research 92(C13), 14,663-14,669.

Jones, D.A., I. Simmonds, 1993. A climatology of Southern Hemisphere extratropical cyclones. Climate Dynamics, 9, 131-145.

Kohout, A.L., Meylan, M.H., 2008. An elastic plate model for wave attenuation and ice floe breaking in the marginal ice zone. Journal of Geophysical Research 113, C09016, doi:10.1029/2007JC004434.

Kohout, A., M.J.M. Williams, S.M. Dean, and M.H. Meylan, Storm-induced sea-ice breakup and the implications for ice extent, *Nature*, 509, 604-607, 2014.

Kohout, A., M.J.M. Williams, T. Toyota, J. Lieser, J. Hutchings. In situ observations of wave-induced ice floe breakup, Deep-Sea Research II, this issue.

Lepparanta, M., 2005. The Drift of Sea Ice, Praxis Publishing, Germany, pp.266.

Liu, A.K., Holt, B., Vachon, P.W., 1991. Wave propagation in the marginal ice zone: model predictions and comparisons with buoy and synthetic aperture radar data, Journal of Geophysical Research, 96(C3), 4605-4621.

Lu, P., Li, Z. J., Zhang, Z. H., Dong, X. L., 2008. Aerial observations of floe size distribution in the marginal ice zone of summer Prydz Bay. Journal of Geophysical Research 113, C02011, doi:10.1029/2006JC003965.

Lu, P., Li, Z., Cheng, B., Lepparanta, M., 2011. A parameterization of the ice-ocean drag coefficient, Journal of Geophysical Research, 116, C07019, doi:10.1029/2010JC006878.

Matsushita, M., 1985. Fractal viewpoint of fracture and accretion. Journal of the Physical Society of Japan 54 (3), 857-860.

Meiners, K. SIPEX-2, a study of sea ice physical, biogeochemical and ecosystem processes off East Antarctica during spring 2012. Deep Sea Research II, this issue.

Mellor, M., 1986. The mechanical behavior of sea ice. In: Untersteiner, N. (Eds.), Geophysics of Sea Ice, Springer, New York, USA, pp.165-281.

Meylan, M.H., 2002. Wave response of an ice floe of arbitrary geometry. Journal of Geophysical Research 107(C1), 10.1029/2000JC000713.

Meylan, M.H., Squire, V. A., 1994. The response of ice floes to ocean waves. Journal of Geophysical Research 99 (C1), 891-900.

Perovich, D.K., Jones, K.F., 2014. The seasonal evolution of sea ice floe size distribution. Journal of Geophysical Research Oceans, 119, 8767-8777, doi:10.1002/2014JC010136.

Rothrock, D. A., Thorndike, A.S., 1984. Measuring the sea ice floe size distribution. Journal of Geophysical Research 89 (C4), 6477-6486.

Schulson, E.M., W.D. Hibler, III, 1991. The fracture of ice on scales large and small: Arctic leads and wing cracks. Journal of Glaciology, 37(127), 319-322.

Squire, V.A., Moore, S.C., 1980. Direct measurement of the attenuation of ocean waves by pack ice. Nature 283, 365-368.

Squire, V.A., 2007. Of ocean waves and sea-ice revisited. Cold Regions science and Technology, 49, 110-133.

Steele, M., 1992. Sea ice melting and floe geometry in a simple ice-ocean model. Journal of Geophysical Research 97(C11), 17,729-17,738.

Steer, A., Worby, A.P., Heil, P., 2008. Observed changes in sea-ice floe size distribution during early summer in the western Weddell Sea. Deep-Sea Research Part II, 55, 933-942.

Toyota, T., Enomoto, H., 2002. Analysis of sea ice floes in the Sea of Okhotsk using ADEOS/AVNIR images. Proc. 16th Int. Symposium on Ice, International Association for Hydro-Environment Engineering and Research, Dunedin, New Zealand, pp. 211-217.

Toyota, T., Kawamura, T., Ohshima, K.I., Shimoda, H., Wakatsuchi, M., 2004. Thickness distribution, texture and stratigraphy, and a simple probabilistic model for dynamical thickening of sea ice in the southern Sea of Okhotsk. Journal of Geophysical Research 109, C06001, doi: 10.1029/2003JC002090.

Toyota, T., Takatsuji, S., Nakayama, M., 2006. Characteristics of sea ice floe size distribution in the seasonal ice zone. Geophysical Research Letters 33, L02616, doi:10.1029/ 2005GL024556.

Toyota, T., Haas, C., Tamura, T., 2011. Size distribution and shape properties of relatively small sea-ice floes in the Antarctic marginal ice zone in late winter. Deep-Sea Research II, 58, 1182-1193.

Wadhams, P., Squire, V.A., Goodman, D.J., Cowan, A.M., Moore, S.C., 1988. The attenuation rates of ocean waves in the marginal ice zone. Journal of Geophysical Research 93 (C6), 6799-6818.

Weeks, W.F., Tucker, W.B., Frank, M., Fungcharoen, S., 1980. Characteristics of surface roughness and floe geometry of sea ice over the continental shelves of the Beaufort and Chukchi Seas. In: Prichard, R.S. (Ed.), Sea ice processes and models, University of Washington Press, Seattle, USA, pp. 300-312.

Williams, G., Maksym, T., Wilkinson, J., Kunz, C., Murphy, C., Kimball, P., Singh, H., 2015. Thick and deformed Antarctic sea ice mapped with autonomous underwater vehicles. Nature Geoscience, 8, 61-67.

Worby, A.P., R.A. Massom, I. Allison, V.I. Lytle, P. Heil, 1998. East Antarctic sea ice: A review of its structure, properties and drift. In East Antarctic Sea Ice: Physical processes, interactions and variability, *Antarctic Research Series*, 74, 41-67, AGU.

Worby, A.P., Steer, A., Lieser, J.L., Heil, P., Yi, D., Markus, T., Allison, I., Massom, R.A., Galin, N., Zwally, J., 2011. Regional-scale sea-ice and snow thickness distributions from in situ and satellite measurements over East Antarctica during SIPEX 2007. Deep-Sea Research II, 58, 1125-1136.

Figure captions

Figure 1. Map showing cruise track (thin line) and heli tracks (thick lines) for the SIPEX-2 expedition with the ice edge locations (broken lines) and the frame of MODIS images used for this study shown. Solid squares show the positions of heli-photos used for analysis and the star shows the position where R/V "Aurora Australis" became stuck (see Section 5.2 for details).

Figure 2. Ice concentration maps from AMSR-E off Wilkes Land, as of (a) September 25 and (b) November 5, 2012. The approximate areas of Fig. 1 are shown with thick white lines. (Data source: http://iup.physik.uni-bremen.de:8084/amsr2/)

Figure 3. Ice thickness distribution along the ship track obtained from the video system. Hourly averaged data are shown by color.

Figure 4. MODIS image showing the locations of four sectors (A-D) as of September 24, 2012. The frame of the image corresponds to the square in Fig.1.

Figure 5. An example showing the process to extract ice floes from (a) a camera photo image taken from the helicopter at 5:14 on Nov 5 and (b) MODIS image on Sep 24. For each case, upper figure shows original video image with each ice floe outlined in red after the process of determining ice edges; and lower figure shows extracted floes to be measured. For (a) the area is 1933 m x 2254 m and 630 ice floes are included for analysis. For (b) the area is 131 km x 126 km and 838 ice floes are included for analysis.

Figure 6. Cumulative number distribution *N(d)* for heli-photos on (a) Sep 25 and (b) Nov 5, respectively. In both figures, black broken lines denote the upper

truncated power law, $M(d) = A \cdot \left(d^{-\alpha} - d_{top}^{-\alpha}\right)$, fitted to the mean data, following the GFF method of Burrough and Tebbens (2001), where A = 1.49 × 10^{8}, α = 1.40, and d_{top} = 175 m for (a) and A = 2.02 × 10^{7}, α = 1.27, and d_{top} = 333 m for (b). Black solid lines denote the underlying power law, $N(d) = A \cdot d^{-\alpha}$. If the floe size measurement was not upper-truncated, these lines would be the cumulative number distribution.

Figure 7. Cumulative number distribution *N(d)* for MODIS images on (a) Sep 24, (b) Oct 4, and (c) Nov 5, respectively. In each figure the locations of A, B, C, and D are shown in Fig.4. In all figures, black broken lines denote the upper truncated power law, $M(d) = A \cdot \left(d^{-\alpha} - d_{top}^{-\alpha}\right)$, fitted to the mean data, following the GFF method of Burrough and Tebbens (2001), where A = 11.63 × 10^{12}, α = 3.09, d_{top} = 16.2 km for (a), A = 2.09 × 10^{12}, α = 2.92, d_{top} = 13.4 km for (b), and A = 1.39 × 10^{12}, α = 2.89, d_{top} = 12.0 km for (c). Black solid lines denote the underlying power law, $N(d) = A \cdot d^{-\alpha}$. If the floe size measurement was not upper-truncated, these lines would be the cumulative number function.

Figure 8. Scatter plot between d_{max} and d_{min} for (a) Heli-photo; (b) MODIS image with regression lines obtained from the least square method.

Figure 9. MODIS images of sector B on September 24, October 04, and November 05, showing temporal evolution of floe size distribution.

Figure 10. One example showing the relationship between crack alignments and wind patterns. (a) MODIS image on October 23. (b) Magnified figure of the square in (a). (c) Wind field obtained from ERA-Interim reanalysis on October 22.

The wind field shown is the daily mean (00, 06, 12, 18UTC) on this day. The area of MODIS image (a) is also shown by a square.

Figure 11. Schematic pictures illustrating the process of lead development induced by wind. Note that ice thickness is significantly different between F_1 and F_2.

Figure 12. Combined cumulative floe size distribution obtained from heli-photo and MODIS images on (a) September 24/25 and (b) November 05.