Landfast sea ice breakouts: Stabilizing ice features, oceanic and atmospheric forcing at Barrow, Alaska

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Abstract

Landfast sea ice is an important seasonal feature along most Arctic coastlines, such as that of the Chukchi Sea near Barrow, Alaska. Its stability throughout the ice season is determined by many factors but grounded pressure ridges are the primary stabilizing component. Landfast ice breakouts occur when these grounded ridges fail or unground, and previously stationary ice detaches from the coast and drifts away. Using ground-based radar imagery from a coastal ice and ocean observatory at Barrow, we have developed a method to estimate the extent of grounded ridges by tracking ice motion and deformation over the course of winter and have derived

ice keel depth and potential for grounding from cumulative convergent ice motion. Estimates of landfast ice grounding strength have been compared to the atmospheric and oceanic stresses acting on the landfast ice before and during breakout events to determine prevailing causes for the failure of stabilizing features. Applying this approach to two case studies in 2008 and 2010, we conclude that a combination of atmospheric and oceanic stresses may have caused the breakouts analyzed in this study, with the latter as the dominant force. Preconditioning (as weakening) of grounded ridges by sea level variations may facilitate failure of the ice sheet leading to breakout events.

Keywords: landfast sea ice, breakout, grounded ridges, coastal radar, ice deformation, Arctic

1 Introduction

Landfast sea ice is an important seasonal feature along circumpolar coasts around the Arctic. Depending on a region's oceanic and atmospheric process regimes, landfast ice can form in place with sea water freezing to form a smooth apron of ice extending from the land, or drifting ice can become stationary in near shore waters and sea water can freeze around it. Landfast sea ice in the Laptev, Kara, and Eastern Siberian Seas can extend smoothly for hundreds of kilometers offshore (Zubov, 1945; Divine *et al.*, 2004; Eicken *et al.*, 2005). In this region, islands help to stabilize the landfast ice from drifting away under offshore winds or currents. In the Alaskan Arctic, Beaufort Sea landfast ice is more extensive and of longer duration than landfast ice in the adjacent Chukchi Sea primarily due to atmospheric

conditions and pack ice drift regimes relative to bathymetry and coastline orientation (Mahoney *et al.*, 2014). In the Baltic Sea, landfast ice typically extends to the outer skerries, in waters 5-15 m deep (Granskog *et al.*, 2006).

The landfast ice season begins near Barrow, Alaska (Figure 1) when drifting ice becomes stationary near the shore. Forming in late fall or early winter (Mahoney *et al.*, 2007b; 2014), the stationary ice (henceforth landfast ice) that extends from the shore serves as a platform for subsistence activities by local residents (Gearheard *et al.*, 2006; George *et al.*, 2004; Sonnenfeld, 1957) and ice use by industry for ice roads in the context of natural resource development (Eicken *et al.*, 2009). It also has provided a platform for extensive scientific studies over several decades, including the 35-year time series of bowhead whale population abundance estimates (e.g., George *et al.*, 2004; Givens *et al.*, 2013). The ice also acts as a barrier, protecting the shore from winter storms (Reimnitz *et al.*, 1994; Reimnitz and Maurer, 1979). While the ice is used as a transportation corridor, it can also be hazardous to marine traffic later in the season (Eicken *et al.*, 2009; Wilson *et al.*, 2004). Biologically, the landfast ice cover is an essential habitat for larger vertebrates (seals and bears) and microorganisms which contribute to primary production in the Arctic Ocean (Gradinger *et al.*, 2009; Horner and Schrader, 1982).

With substantial changes to the sea ice extent and thickness over the past couple of decades and projected into the future (Serreze and Stroeve, 2015), the Arctic is experiencing a major increase in maritime shipping, off- and near-shore resource exploration, and socio-economic changes affecting the region (Brigham, 2010). Eicken *et al.* (2009) demonstrated the need for arctic observing networks

that both address key scientific research questions and provide information relevant to sea ice system stakeholders who need to adapt to changing arctic conditions. In this context, the monitoring of coastal sea ice dynamics near Barrow, Alaska (Figure 1) contributes not only to our fundamental understanding of landfast ice breakouts. The data also provides insight into the stability of landfast ice as a platform for various uses, the duration and extent of the landfast sea ice, and near-shore ice morphology. Such information can inform regulation of activities in ice covered near-shore regions around the Arctic (Lovecraft et al., 2013).

The current study uses a land-based marine radar at Barrow, Alaska to monitor near-shore sea ice and obtain an estimate of the extent of grounded

Figure 1. Map of study area with location of sensor sites. The black circle outlines the typical extent of sea ice radar coverage. Red dots mark locations of the radar, oceanographic mooring, mass balance sites, and airport where wind data was collected. The dots labeled MBS09 and MBS10 are the locations of the mass balance sites in 2009 and 2010.

pressure ridges that provide stability to a landfast ice cover. The marine radar also registers the detachment of landfast ice from the coast. Other studies have utilized coastal marine radars to map near-shore ice and calculate properties of the ice velocity field (Tabata *et al.*, 1980), as well as use optical flow fields for tracking sea ice in radar imagery (Karvonen, 2016).

The atmospheric and oceanic conditions are analyzed in terms of their forcing of mobilization and failure of stationary ice. While ice conditions at Barrow may only be representative of parts of the eastern Chukchi and western Beaufort Sea (Norton and Graves Gaylord, 2004), the insight gained into the interactions between the atmosphere, landfast ice, and ocean is applicable throughout the Arctic where dynamic landfast sea ice processes occur and may help in the evaluation of ice stability and possibly inform prediction of ice hazards.

2 Background and Regional Setting

Landfast ice in the Point Barrow region (Figure 1) exhibits characteristics of both Beaufort and Chukchi Sea ice. Due to the prevailing winds coming from the northeast and currents moving towards the northeast in the area, and the orientation of the landmasses, the ice conditions around Barrow are dynamic, with significant ice deformation that results in persistent patches of open water as well as strongly ridged ice (Eicken and Mahoney, 2015; Norton and Graves Gaylord, 2004). The dominant drift pattern in the southern Beaufort Sea is westward, with winds primarily from the north-northeast, routinely resulting in an open water lead beyond the landfast ice along the eastern Chukchi Coast. Native Iñupiat people have exploited the lead that often forms between stationary landfast ice and mobile drift ice for subsistence hunting for well over a thousand years (Gearheard *et al.*, 2006; George *et al.*, 2004). Similar conditions characterize the ice regime along the northeastern Chukchi Sea where most communities have been established near

coastline features that promote formation of open water, close access to lead systems, and ice deformation features (e.g., Nelson, 1969).

Ocean currents along the eastern Chukchi coast are generally towards the northeast, part of the Alaska Coastal Current (Woodgate *et al.*, 2005). Wind-driven reversal of currents can happen occasionally, with water moving towards the southwest along the coast and possible upwelling of warm water into the coastal area (Weingartner *et al.*, 1998). Tides are not very large in the eastern Chukchi Sea, with a range of 0.15 m near Barrow (Sprenke *et al.*, 2011). The composition of the seafloor is gravel-like sediment that decreases in depth with distance from the shore (Barnes *et al.*, 1983). The region is heavily scoured by ice ridge keels, though the seafloor is rapidly reworked by currents and wave action.

There are also instances during the ice season when the pack ice drifts shoreward from the west, closing the coastal polynya. Iñupiat hunters take note of such events during the early ice and stable ice periods, as they can result in deformation of the landfast sea ice, and form grounded ridges (Joe Leavitt, personal communication, 2012; Mahoney *et al.*, 2004, 2007a). In-situ deformation and ridge formation is an important process in the stabilization of the landfast ice cover near Barrow as a higher number of grounded ridges is observed than would be expected from the distribution of deep first year ridge keels (≥ 20 m) across the Arctic (Mahoney *et al.*, 2007a).

Due to the importance of landfast sea ice, an understanding of the processes that form a stable ice cover and the forces that work to destabilize ice would be beneficial. Ridges ground by either drifting into or being built through ice

convergence in shallow water (Mahoney *et al.*, 2007a). In either case, such grounded pressure ridges and the atmospheric and oceanic forces controlling their formation and decay ultimately affect the stability of a landfast ice cover at Barrow and other coastal sites.

The same forces, although often coming from different directions, that give rise to stabilizing grounded ridges can work to dislodge a stable landfast ice cover, resulting in what is known as a breakout of seemingly stable landfast ice (Druckenmiller *et al.*, 2009; Mahoney *et al.*, 2007a). Currents and/or winds can exert enough force to break ridge keels or dislodge the keels from the seafloor gouge created during the grounding process. Drifting ice can come into contact with landfast ice and knock it loose in some instances (Joe Leavitt, personal communication, 2012). Other factors can also precondition the landfast ice for a breakout. For example, a rise in sea level can lift the ridge keels out of their gouges, while a drop in sea level could cause the ridge keel to shorten under the added weight. Warm water intruding into the coastal region can weaken the cohesiveness of grounded ridge keels, which can then be compromised under certain circumstances. The exact conditions that lead up to and contribute to a breakout are not well known, and difficult to measure, as are the extent and strength of grounded ridges opposing breakout forces.

3 Data and Methods

The Barrow Sea Ice Observatory provided most of the data for the breakout case studies. The Observatory is an on-going program aimed at observing near-shore sea

ice characteristics and dynamics, and providing useful information to the local community at Barrow. A full description of all components of the observatory is provided by Druckenmiller *et al.* (2009). In the following sections, we provide the relevant details of the sea ice mass balance site, land-based marine radar, satellite data of the local region, offshore moorings and local Iñupiat ice experts' observations.

3.1 *Sea ice mass balance site*

The Barrow sea ice mass balance site (MBS) is a seasonally-deployed suite of sensors frozen into the landfast sea ice near Point Barrow (Figure 1). Down-looking Campbell's Scientific SR50A acoustic altimeters above the ice measure changes in the position of the air/snow interface or the air/ice interface in the absence of snow. An upward-looking Teledyne Benthos PSA-916 underwater acoustic altimeter below the ice measures changes in the position of the ice/water interface. Together, these observations allow us to derive changes in snow depth and ice thickness from early winter until late spring, capturing the annual maximum ice thickness and the onset of top and bottom ablation. Each year, the MBS is located within a large, level stretch of landfast ice so that these observations can be assumed to be representative of the thermal processes governing growth and decay of undeformed ice in the Barrow region. These ice thickness data are used in the ridge volume calculations later in this work.

3.2 *Marine radar*

A marine radar is employed in monitoring near-shore ice development and movement at high spatial and temporal scales. The radar is a Furuno FR7112 10kW,

X-band (3 cm, 10GHz) system that is situated 22.5 m above sea level on one of the largest buildings in Barrow (Figure 1). The operational effective range of the radar is 11 km, although this range varies with atmospheric conditions. The 1.65 m slotted waveguide antenna provides a horizontal beamwidth of 1.2° with the vertical beamwidth at 22°. With a range of 10 km, the minimum radial distance between reflectors must be 209 m for the radar to discern them as two separate objects. At a pulse length of 0.3 μ m, two objects need to be approximately 45 m apart in range from the radar to be resolved. At a range of 5 km, the surface relief needs to be about 20 cm to be rough enough for detection. This means wind-roughened open water can be detected by the radar, especially rough water closer to the radar unit itself. Radar return signals from rough water during the ice season, however, are not persistent and filtered out in further processing of the radar imagery.

With the aid of a Xenex XC2000 digital controller, the radar is operated remotely from the Geophysical Institute at the University of Alaska Fairbanks (UAF). The controller captures a radar image every 120 rotations (approximately 4 minutes) and stores it on a local hard drive. The radar imagery is then sent to UAF every 10 minutes where it is processed and archived. The processing results in an orthorectified, georeferenced image with an effective pixel size of 21.5 ± 0.5 m (Mahoney *et al.*, 2015).

The radar receives backscatter from sea ice features within its range of about 11 km, but the backscatter signal changes depending on atmospheric and ice conditions. Ridges, the edges of ice floes, and other rough surface features not in the

shadow of other such elements of the ice cover are discernible in the radar imagery. In general, areas of flat ice and calm open water do not generate enough backscatter to be detected by the radar (Mahoney *et al.*, 2007a). Areas devoid of return signals do not necessarily imply flat ice or open water in that area. Ice topography, if sufficiently high, can occlude ice features at a greater range. The strength of a radar signal is also dependent upon the orientation of the reflection surface to the impinging radar signal, as well as physical characteristics of the ice.

Animations of the radar imagery are a valuable source of qualitative information on the development and evolution of the landfast ice cover and near-shore sea ice dynamics near Barrow (Druckenmiller *et al.*, 2009; Mahoney *et al.*, 2007; Shapiro and Barnes, 1991). Some types of quantitative data can be extracted from the radar animations through simple observation; examples are the length of time the pack ice is moving to the southwest or northeast, the dates of first and last stationary ice near the coast, and the number of deformation events at the seaward landfast ice edge (SLIE) and the SLIE location relative to the near-shore bathymetry. Timeseries of radar imagery can also be used to generate quantitative ice velocity data, as described in section 3.7.

3.3 *Satellite products*

There are many satellite data products available to the Barrow Sea Ice Observatory via the Alaska Ocean Observing System and the Geographic Information Network of Alaska. In this study, advanced synthetic aperture radar (ASAR) imagery acquired by Envisat and provided by the Barrow Area Information Database is used to identify landfast ice edges beyond the footprint of the land-

based radar, and possible pack ice interactions with landfast ice.

3.4 Offshore moorings

Starting with the 2009-2010 ice season, the Barrow Sea Ice Observatory has included two offshore moorings deployed in collaboration with Hokkaido University in Japan (Ito et al., 2015; Mahoney et al., 2015; Petrich *et al.*, 2012). The moorings are equipped with a Teledyne RDI Workhorse Sentinel acoustic Doppler current profiler (ADCP), which determines the water current velocity profile by measuring the Doppler shift of acoustic signals returned from the water column. During the 2009-2010 ice season, one mooring was located approximately 4 km offshore from the radar's location (Figure 1), and was underneath a landfast ice cover or landfast ice extensions for much of the season. For the case studies, the water current in the uppermost bins unaffected by ice or waves (8-18m water depth) and water temperature was analyzed. Considering no irregular topography, the current data from the ADCP can be assumed to be representative of the currents in the area of the landfast ice near Barrow to some extent. The current data does generally agree with the observations of native ice experts from Barrow (Mahoney *et al.*, 2011).

3.5 Local ice observations

Joe 'Joe Mello' Leavitt, an Iñupiaq hunter and sea ice expert from Barrow, has been making daily sea ice observations since 2006 (Apangalook et al., 2013; Eicken *et al.*, 2014). His observations provide insight into ice dynamics and conditions that can only be made by physically observing the ice, as well as providing an alternative context in which to describe and determine the stability of the landfast ice cover. Mr. Leavitt's ice expertise is grounded in indigenous and traditional environmental

knowledge and has developed from instruction by Iñupiaq elders and many years of subsistence activities on the ice, requiring him to constantly evaluate the condition of the ice, weather, and currents in order to be successful and safe.

3.6 Other data

Data used in this work that was not collected from the Barrow Ice Observatory was all obtained from the National Oceanic and Atmospheric Administration (NOAA). Bathymetry data were downloaded from the National Centers for Environmental Information website at

<https://www.ngdc.noaa.gov/mgg/bathymetry/maps/area7.html>. Weather data, including wind data, were downloaded from the National Climate Data Center at <http://www.ncdc.noaa.gov/qclcd/QCLCD?prior=N>. NOAA's National Ocean Service Center for Operational Oceanographic Products and Services operated water level gauges near Barrow from 2008 to 2010 (Sprenke *et al.*, 2011). Water level data from the NOAA gauges was downloaded from <http://tidesandcurrents.noaa.gov/inventory.html?id=9494935>.

3.7 Detection of breakout events

Defining the SLIE is a difficult task, since it is not necessarily marked by the presence of open water against an actual edge of the ice cover; moreover, its location and shape may vary rapidly in time. Previous analyses of landfast ice (Fraser *et al.*, 2010; 2012; Mahoney *et al.* 2005; 2007b; 2015) employed a 20-day time period to define the edge of the landfast ice; if ice extending from the coast remained stationary for 20 days, it was considered part of the landfast ice and the edge was seaward of the stationary ice. This 20-day time period was chosen based

Figure 2. Cross section of typical ice zones present along Alaska's Arctic coast. A rough vertical scale is provided.

on typical synoptic time scales and the repeat interval of Radarsat SAR scenes of the eastern Chukchi and western Beaufort Sea regions. In the current case studies, a time period of seven days of stationary ice was chosen to determine whether or not ice in the near-shore region is considered part of the stable landfast ice or an extension that will drift away once the conditions holding it adjacent to the true SLIE change. Using a criterion based on such a shorter time scale is made possible by the much higher temporal resolution and smaller spatial coverage of the coastal sea ice radar compared to SAR imagery. This seven-day stability criterion also better reflects local use of the ice cover on shorter time scales.

The seven-day landfast criterion is important when determining if ice divergence seen in the radar animations corresponds to a breakout event or the drifting away of a landfast ice extension (Figure 2). Landfast sea ice shoreward of grounded ridges is considered to be stable, as it is the grounded ridges that prevent the ice from becoming mobile. Without actually determining whether or not ridges are grounded, some other measures must be employed in order to determine if the landfast ice cover is stable. One way to evaluate the stability of the landfast ice is to

determine whether the ice cover remains immobile under strong easterly winds or offshore currents that would be able to dislodge unanchored ice. Also, the longer the ice remains motionless, the more likely it is stably attached to the coast by grounded ridges.

In order to identify breakout events we visually inspected radar animations for the 2006-2007 and 2009-2010 ice seasons. The boundary between visibly moving ice and stationary ice (i.e., the SLIE) was noted on a daily basis. A breakout is considered to have occurred if the SLIE moved shoreward having previously remained in the same position for at least seven days. If the radar footprint was completely filled with stationary ice before a possible breakout, the resulting SLIE was compared to the most recent prior SLIE to determine the extent of the breakout.

Based on the overall extent of landfast ice, and the area of formerly landfast ice involved in the breakout, an index for the severity of the breakout was developed, assuming the landfast ice extent has reached the 20 m isobath (~2 km from shore) indicating the beginning of the stable ice period (Mahoney *et al.*, 2007b). A breakout of Severity Level 1 corresponds to an event in which the landfast ice width (distance of SLIE from shore) is over 2 km and reduced by at least 0.5 km but less than half of the pre-breakout width, measured by the perpendicular distance from the coast to the SLIE. For example, landfast ice width of 3 km from shore that is reduced to 2 km would be a breakout of Severity Level 1. If the landfast ice width is reduced by at least half, but remains at least 1 km, a breakout of Severity Level 2 occurred. Breakout events of Severity Level 3 occur when the landfast ice width is reduced by at least half, and the SLIE comes to within 1 km or less of the shore. Since the

deployment of this radar system for the 2006-2007 ice season, through the 2009-2010 season (four ice seasons total), there have been three breakouts at Severity Level 1, five at Severity Level 2, and one at Severity Level 3, for a total of nine breakouts observed with the radar.

3.8 *Tracking sea ice in the radar imagery*

The problem of tracking ice features through radar imagery is challenging due to the characteristics of the data gathered. The area within the range of the radar may not be filled with ice, or may appear to not be filled with ice due to occlusions that also pose a problem to traditional tracking algorithms for radars (MV *et al.*, 2013). Due to occlusions and non-rigid body deformation, traditional feature tracking methods are not always effective in tracking sea ice from marine radar imagery. Instead, a combination of existing and newly developed methods is used (MV *et al.*, 2013). For areas of high ice concentration, a normalized cross correlation algorithm is employed to estimate the motion field. In other areas, a combination of tracking methods are used to track mobile radar reflectors (MV *et al.*, 2013). To reduce computation time, the radar images are downsampled by a factor of two, resulting in a pixel size of 43 m. Ice velocities are calculated for 10x10 pixels subsets, such that there are 430 m between velocity grid points.

To minimize the potential impacts of noise in radar backscatter signals, filters are applied to the velocity fields. Mahoney *et al.* (2015) showed that the application of a running 1-hour median filter to a year-long timeseries of radar-derived ice velocity produced a close match (RMS difference of 0.12 m s⁻¹) with a comparable timeseries of ice bottom-track velocity measured with an under-ice

ADCP located within the radar footprint. We apply the same 1-hour temporal median filter to each grid point in the velocity fields used here, with the requirement that there are at least five valid measurements over that hour. This reduces most of the noise-impacted vectors in any given velocity field, although wayward vectors are sometimes not filtered in areas of sparse radar reflectors or high noise to signal ratio.

To remove most of the remaining spurious vectors from the velocity field, a similarity filter is applied. The similarity filter compares each vector's zonal and meridional components to (i) the temporal average of that vector determined for the present, the previous and subsequent velocity field, (ii) the spatial average of its eight immediate neighbors both for the same time period and for the velocity fields for the immediately preceding and subsequent scenes. If either of the considered vector's components deviate by more than 1 m s^{-1} from any of the average values determined for (i) and (ii) then the vector is excluded from the calculation of a final velocity field.

The velocity fields were also processed to remove spatial data gaps. These gaps tend to correspond to center sections of large floes associated with occlusion or specular reflection of the radar signal, while the surrounding reflectors behave as a rigid body. Such regions can result in divergence artifacts and are removed by interpolating to the nearest valid velocity neighbor.

3.9 Estimation of grounded ridge extent

With the gridded velocity fields provided by the ice tracking program, an estimate of the number, and length, of grounded ridges is made possible by deriving

and analyzing ice deformation fields. A principal orthogonal component of the two dimensional strain field is given by (Thorndike, 1986)

$$E_I = \frac{\partial u_x}{\partial x} + \frac{\partial u_y}{\partial y} \quad \text{Eq. 1}$$

where E_I is the divergence/convergence component, and u_x and u_y are the velocity components in the zonal and meridional directions, respectively. Since the velocity components are determined on a regular grid, the derivatives in the equation above can be approximated with a centered difference scheme (Mahoney *et al.*, 2004).

Divergence represents a change in area, with positive values representing an increase in area (i.e., open water production) and negative values indicating a reduction in area (i.e., lead closing and ridge building), and is thus dimensionless.

The velocity output of the ice-tracking program is provided in terms of the displacement in pixels over the span of three images (approximately 8 minutes). A displacement of one velocity grid cell (10 pixels) between images therefore corresponds to an ice velocity of approximately 0.9 m s^{-1} , though due to the combined uncertainties in the spatial scale and timing of the imagery, there is an estimated 5 % error in these values (Mahoney *et al.*, 2015). Divergence, as defined by Equation 1, has dimensions of s^{-1} . However, for a physical interpretation of the quantity the spatial dimensions of the velocity grid have to be considered. With a grid spacing of 430 m, a divergence of -1 therefore corresponds to a shortening of 430 m in the direction of ice motion.

In order to determine the cross-sectional size of a ridge created through convergence, we need to know the parent ice sheet thickness and the fraction of the

ridge volume occupied by voids. Taking a keel porosity of 0.3 (Nuber *et al.*, 2013), the sail porosity would be 0.22 according to

$$\rho_s = \frac{\rho_k^{-0.14}}{0.73} \quad \text{Eq. 2}$$

where ρ_s is the porosity of the sail and ρ_k is the keel porosity, and a keel area to sail area ratio of 8, following Timco and Burden (1997). The area of ice in a cross-section of the ridge is given by

$$A_{ri} = S_{vg} H_i E_I \quad \text{Eq. 3}$$

with S_{vg} as the shortening length of a velocity grid cell, H_i is the thickness of the parent ice sheet, and E_I is the divergence calculated from the velocity fields. In terms of the area of the ridge keel and sail, ice area in the ridge becomes

$$A_{ri} = (1 - \rho_k) A_k + (1 - \rho_s) A_s \quad \text{Eq. 4}$$

where A_k is the total cross-sectional area of the ridge keel, and A_s is the total cross-sectional area of the ridge sail. Rearranging Equation 4 to solve for keel area and combining with Equation 3 gives

$$A_k = \frac{S_{vg} H_i E_I - (1 - \rho_s) A_s}{(1 - \rho_k)} \quad \text{Eq. 5}$$

The depth of a ridge keel H_k can then be found using

$$H_k = \sqrt{A_k \tan(\alpha_k)} \quad \text{Eq. 6}$$

with α_k as the ridge keel angle of repose.

Figure 3. Cross-section of idealized first year sea ice ridge. Relevant parameters are identified. Adapted from Timco and Burden (1997).

For a parent ice thickness of 1 m, a radar-derived divergence of -1 would result in a ridge with a keel cross section of 540 m^2 . Using Timco and Burden's (1997) ratio of cross-sectional keel area to sail area of eight, a keel angle of 26.6° (Timco and Burden, 1997) and assuming a triangular ridge shape, the depth of such a keel would be 17 m. Figure 3 displays an idealized grounded ridge cross-section with relevant parameters, adapted from Timco and Burden (1997).

In a case in which the ridge keel reaches the sea floor and convergent deformation continues, we assume the ridge will grow seaward and take on a trapezoidal shape. The sail can also cease to gain height and grow wider instead of

taller. Here it is assumed that macroscopic porosity of the ridge does not change with the change in direction of primary ridge growth from vertically to laterally, and that the ridge keel does not significantly gouge into the sea floor. It is important to note that with the same parameters as the triangular ridge case (1 m thick parent ice and 30% keel porosity), the total cross section of the ridge keel will not change. Therefore, if the depth is taken constant at 12 m in the location of ridge formation, a divergence of -1 would create a ridge that is 69 m wide (perpendicular to the direction of ice motion) with 21 m of the ridge's keel in contact with the sea floor.

The frictional coupling of the ridge with the sea floor is found with

$$F_{sb} = \frac{c_f n_g \rho_w g \bar{D}}{\tan \alpha_k} \quad \text{Eq. 7}$$

(based on Druckenmiller, 2011) where F_{sb} is the estimated frictional stress of the ridge with the sea floor, c_f is the coefficient of static friction with the sea floor (0.5; Shapiro and Metzner, 1987), n_g is the number of grounded ridges (one in this hypothetical case), ρ_w is the density of sea water (1027 kg m^{-3}), g is gravitational acceleration (9.81 m s^{-2}), $\bar{D}$ is the degree of grounding, and α_k is the angle of the ridge keel. The degree of grounding, $\bar{D}$, is given as (Druckenmiller, 2011)

$$\bar{D} = \frac{1}{n_g} \sum_{i=1}^{n_g} (H_k - w_g)^2 \quad \text{Eq. 8}$$

where w_g is the water depth. The frictional coupling of the hypothetical ridge described above, with an estimated keel depth of 17 m in 12 m of water, would be 220 kN. For a full derivation of the above equations, see Druckenmiller (2011).

Hypothesized ridge geometry plays an important role in estimating the strength of a grounded ridge's coupling to the sea floor. The ridge geometry used in

this work is based on the ridge characteristics described by Timco and Burden (1997), who only considered floating ridges. Since the current work focuses on grounded first-year pressure ridges, ridge geometry may differ significantly from that of an idealized floating first-year pressure ridge. One of the characteristics of ridge geometry implicit in the equation for determining a ridge keel's coupling to the sea floor is the angle of the ridge keel, α_k .

Atmospheric and oceanic forces act on landfast sea ice, thereby contributing to the occurrence of a breakout event. In terms of effective stresses, atmospheric stress F_a exerted by the wind on stationary landfast ice is given by

$$F_a = L_{LFI} \rho_a c_a U_a^2 \quad \text{Eq. 9}$$

where L_{LFI} is the length of landfast ice over which the wind acts, ρ_a is the density of air, c_a is the coefficient of drag for air moving over sea ice, and U_a is the wind velocity. Similarly, effective oceanic stress on the underside of the ice is

$$F_w = L_{LFI} \rho_w c_w U_w^2 \quad \text{Eq. 10}$$

where L_{LFI} is the length of landfast ice the current is acting on, ρ_w is the density of sea water, c_w is the drag coefficient at the ice-water interface, and U_w is the current velocity.

Balancing Eq. 7 with Eq. 9 and Eq. 10 provides ratios of the grounded ridges frictional coupling to the seafloor (F_{sb}) to atmospheric stress from winds (F_a) and oceanic stress from currents (F_w)

$$B_a = \frac{F_{sb}}{F_a} = \frac{c_f n_g \rho_w g \bar{D}_g}{L_{LFI} \rho_a c_a U_a^2 \tan \alpha_k} \quad \text{Eq. 11}$$

and

$$B_w = \frac{F_{sb}}{F_w} = \frac{c_f n_g \rho_w g \bar{D}_g}{L_{LFI} \rho_w c_w U_w^2 \tan \alpha_k} \quad \text{Eq. 12}$$

If B_a or B_w is less than 1, the landfast ice cover can be destabilized due to atmospheric or oceanic stresses, respectively.

4 Results: Analysis of Landfast Ice Breakout Events

Nine breakout events were observed in the radar animation data between the 2006-2007 ice season and the 2009-2010 ice season. For the in-depth analysis that follows, we selected two severity level 2 breakout events that occurred after a stable landfast ice cover had formed and for which a complete, or nearly complete, set of concurrent environmental data and local observations was available. In addition, both events were deemed to be significant for the local community with regard to landfast ice stability. The breakouts, which we will refer to as breakout events 1 and 2, occurred on February 27, 2009 and March 24, 2010, respectively.

In order to understand the deformation mechanisms involved in any landfast ice breakout event, it is important to review the formation and growth history of both the ice involved in the breakout and the ice that remains attached to the coast. Below, we provide descriptions of ice motion and environmental conditions leading up to and during both breakout events mentioned above. This analysis is based on data from the coastal radar, the MBS, meteorological observations from Barrow airport as well as Leavitt's near-daily observations (Apangalook et al., 2013; see also Eicken *et al.*, 2014)

4.1 Breakout event 1

4.1.1 Pre-breakout ice conditions

477 All events described in this section can be viewed in radar animations
478 included in the supplemental material under BreakoutEvent1 available online, or at
479 http://seaice.alaska.edu/gi/data/barrow_radar/recent-radar-animations. Table 1
480 gives the dates of the emplacement of the ice involved in the breakout, and the date
481 of the breakout event. In early February, the radar footprint was filled with
482 stationary ice beginning on the 6th. Figure 4 displays radar images with pre-
483 emplacement and post-emplacement SLIEs outlined before and after the
484 deformation event described below, as well as the post-breakout SLIE. When some
485 of the ice that filled the radar footprint resumed movement towards the southwest
486 on February 13, the pre-emplacement SLIE was a little over 1.5 km offshore, 0.5 km
487 further out than the prior SLIE position. This accretion of landfast ice shifted the
488 SLIE into waters approximately 18-20 m deep. Advance of the SLIE beyond the 15 m
489 isobath has served as an indicator of the beginning of the stable landfast ice period
490 (Barry *et al.*, 1979; Mahoney *et al.*, 2007b). Indeed, the ice accreted during this
491 episode persisted through the end of the ice season. Leavitt's observations indicate
492 that it was foggy and cloudy during this period, although he saw open water ~1.5
493 km out on February 14 (Apangalook *et al.*, 2013).

494 On February 18, the SLIE was discernable 2 km from shore when the ice
495 beyond it became mobile, but on February 19-20 (Figure 4a) the pack ice advected

Table 1. Dates of important events during the 2008-2009 ice season.

Event	Date
Emplacement of ice that breaks out	19-Feb-09
Breakout event	27-Feb-09

496 from the northwest toward the SLIE with new ridges appearing in what we assume
497 was newly-formed (i.e. undeformed) ice due to the appearance of new reflectors. On
498 February 21, the ice became mobile once more, with post-emplacement SLIE at
499 about 5.5 km offshore (Figure 4B). The landfast ice that attached during this
500 sequence of events was subsequently involved in the breakout on February 27. After
501 the ice drifted away on February 21, there was no ice movement in the radar
502 footprint until late in the day of February 23 when ice drifting to the northeast
503 became visible. Northeastward ice drift persisted until late on February 25, when
504 the radar footprint filled with immobile ice. There was no movement of radar
505 reflectors until between 7:29 and 7:33 on February 27. Shortly before the breakout
506 event, 'flickering' of radar targets in the ice that was to breakout occurred, as
507 described in more detail by Mahoney et al. (2007a). When the breakout occurred,
508 ice began to move almost due north, leaving the post-breakout SLIE about 2 km
509 from shore adjacent to Barrow (Figure 4C).

4.1.2 Conditions during the breakout event

The environmental conditions during the breakout events were used to estimate the atmospheric and oceanic stresses acting on the landfast ice, as well as to elucidate other processes, which affect landfast ice stability. Estimates of the degree of grounding of in-situ formed pressure ridges, based on the convergent portion of divergence calculations on the ice drift velocity fields, are also presented.

The atmospheric conditions (air temperature, wind direction, and wind speed) and water level (relative to mean higher high water, MHHW) leading up to and during breakout event 1 on February 27, 2009 are shown in Figure 5. The effective stress on the landfast ice was found using Equation 9 with $\rho_a=1.3 \text{ kg m}^{-3}$, $c_a=0.0042$ (extremely rough first year ice, Guest *et al.*, 1994), and $L_{LFI}=4 \text{ km}$ from the radar imagery.

Figure 4. Radar scenes annotated with ice motion direction (green arrow) and location of SLIE (red line). A) Pre-emplacement SLIE, before February 19-20, 2009 deformation that resulted in building of grounded ridge keels. B) Post-emplacement SLIE, after deformation event when SLIE could be determined on February 22. C) Post-breakout SLIE on February 27.

Figure 5. Atmospheric conditions (A-C) and water level (D) for the month of February 2009, with estimated induced stress from winds (E) on the landfast ice. The estimated grounding strength of grounded ridges relative to water level is also shown in E. Vertical gray line indicates time of emplacement of ice that broke out, and the red line the time of the breakout.

522   At the time of the breakout, winds were from the SSE (nearly directly
523 offshore) at 7 m s^{-1} , resulting in an effective stress of $\sim 1.5 \text{ kN m}^{-1}$ acting across the 4
524 km long stretch of ice that broke out. However, some of the ice that broke out was
525 accreted on February 19-20 and therefore experienced similar wind conditions on
526 February 25 (SSE at about $7\text{-}8 \text{ m s}^{-1}$) and even stronger offshore winds on February
527 21-22 ($12\text{-}14 \text{ m s}^{-1}$) without detaching from the coast.

528 An estimation of the extent of grounded ridges is given in Figure 6. In order
 529 to estimate the degree of grounding, the parent ice sheet thickness used in the
 530 calculations is 0.95 m, obtained from the mass balance site. The ice divergence fields
 531 derived from the sea ice velocity fields were summed over the time period from
 532 February 17 to the time of the breakout on February 24. The blue scale grid cells in
 533 the offshore portion of the image represent the difference between estimated keel
 534 depth and water depth, with darker cells being more extensively grounded than
 535 lighter grid cells. The number of grounded grid cells and the difference between

Figure 6. Estimation of grounded ridge extent
 resulting from February 19-20, 2009
 deformation. The scale shows the difference
 between calculated keel depth and water
 depth. The thick red line is the pre-
 emplacement SLIE on February 19. The pink
 line is the post-emplacement SLIE from
 February 22. The black line is the post-
 breakout SLIE on February 27. Gray cross is
 location of ADCP.

average water depth for that grid
 cell and keel depth are used in the
 calculation of the degree of
 grounding of the landfast ice that
 ultimately broke out.

Prior to breakout of ice on
 February 24, there were 72
 grounded ice grid cells within the
 radar footprint. Based on
 Equation 7 in combination with
 the equations used to estimate
 keel depth (Equations 5 and 6),
 with an ice thickness of 0.95 m,
 this corresponded to a grounding
 strength on the order of 17×10^4

kN from ridge keel-sea floor coupling. However, only 50 of these cells were in ice involved in the breakout (Figure 6). Therefore a grounding strength of around 8.0×10^4 kN had to be overcome in order to detach the section of ice that broke away. The grounded keels in the landfast ice that remained attached served to strengthen the grounding force of that ice, and probably resisted the stresses that contributed to the breakout.

4.2 Breakout event 2

4.2.1 Pre-breakout ice conditions

Table 2 gives the dates of important events in the 2009-2010 ice season.

Below is a more detailed description of events immediately before breakout event 2 on March 24, 2010. All events described in this section can be viewed in radar animations included in the supplemental material under BreakOutEvent2 available online, or at http://seaice.alaska.edu/gi/data/barrow_radar/recent-radar-animations.

The radar footprint was filled with stationary ice from March 12 until March 16, when the ice moved to the north leaving the pre-emplacement SLIE 1.5 km from shore, in agreement with Leavitt's notes (Apangalook et al., 2013). Figure 7 displays radar images with pre-emplacement and post-emplacement SLIEs outlined before and after the deformation event described below, and the post-breakout SLIE. The pack ice continued to drift slowly offshore to the north until the morning of March

Table 2. Dates of important events during the 2009-2010 ice season.

Event	Date
Emplacement of ice that breaks out	17-Mar-10
Breakout event	24-Mar-10

17 (Figure 7A) when it began to take a more eastward drift direction. While the ice was drifting shoreward, new ice between the mobile pack ice and the SLIE was apparently being deformed. Throughout March 17 and early on March 18, intermittent shoreward movement resulted in ridging and deformation of the ice at various distances from the SLIE, assuming 100% ice cover in the radar footprint. By ~06:00 local time on March 18, no movement occurred shoreward of 6.5 km from the coast adjacent to Barrow, though the SLIE increased from the southwest to the northeast. On March 19, the post-emplacement SLIE became apparent when ice further from shore began drifting to the north/northwest (Figure 7B). The ice that became landfast on March 17 and 18 was the ice involved in the breakout on March 24.

After the mobile ice drifted away on March 19, no movement was detected until March 22, when pack ice was visible moving to the southwest beyond the SLIE. These radar observations correspond closely to observations made by Mr. Leavitt from the shore (Apangalook et al., 2013). Radar targets in the ice that was to breakout flickered in the hours leading up to the event. On March 24, between 12:06 and 12:22 local time, the landfast ice that had been present since March 17 began to drift southwest, breaking into 4 or 5 large pieces. The post-breakout SLIE adjacent to Barrow was moved shoreward and reoccupied its previous position (as of March 16) ~1.5 km distant from shore, increasing to about 3 km from the shore farther to the north (Figure 7C).

4.2.2 Conditions during the breakout event

In addition to estimates of wind stress, water level, and grounded ridge strength, our analysis of event 2 draws on the oceanographic moorings with data for direction and speed of ocean currents. Using Eq.10, the effective current stress on the underside of the landfast ice can be estimated. The atmospheric conditions for March 2010 are displayed in Figure 8, along with water level. Current direction and velocity, and water temperature, are also displayed in Figure 8. Principal component analysis was performed on the current direction and magnitude, and the displayed data are the two orthogonal components that explain the greatest fraction of the variance in the data, corresponding to along-shore and cross-shore currents. The effective stress calculations for both winds and currents are also displayed in Figure 8. The effective stresses were estimated with $\rho_a=1.3 \text{ kg m}^{-3}$, $c_a=0.0042$, $\rho_w=1027 \text{ kg m}^{-3}$, $c_w=0.0175$ (based on $c_a/c_w=0.24$; Leppäranta, 2005), and $L_{LFI}=23 \text{ km}$ from the radar and satellite imagery.

Figure 7. Radar scenes annotated with ice motion direction (green arrow) and location of SLIE (red line). A) Pre-emplacement SLIE, before March 17-18, 2010 deformation that resulted in building of grounded ridge keels. B) Post-emplacement SLIE, after deformation event when SLIE could be determined on March 19. C) Post-breakout SLIE on March 24.

Figure 8. Atmospheric conditions (A-C) and oceanic conditions (D-G) for the month of March 2010, with wind (H) and current (I) induced stresses on landfast ice. The estimated grounding strength of grounded ridges relative to water level is also shown in H and I. Vertical gray line indicates time of emplacement of ice that broke out, and the red line the time of the breakout.

Figure 9. Estimation of grounded ridge extent resulting from March 17-18, 2010 deformation. The scale is the difference between calculated keel depth and water depth. The thick red line is the pre-emplacement SLIE from March 17. The pink line is the post-emplacement SLIE from March 19. The black line is the post-breakout SLIE on March 24. Gray cross is location of ADCP. Note that the scale is different than that of Figure 7.

At the time of the breakout and for 7 days prior, winds were from the northeast. The wind speed at the time of the breakout was $\sim 11 \text{ m s}^{-1}$, for an effective stress of $\sim 15 \text{ kN m}^{-2}$ on the 23 km of ice that broke out. The water level was dropping to a month-long low, although the breakout occurred shortly before the minimum. Water temperature was slightly below -1.5°C , although it increased to $\sim 0.5^\circ\text{C}$ immediately after the breakout occurred. Ocean currents were

to the southwest and slightly offshore at the time of the breakout. The currents moving to the southwest imparted an effective stress of $\sim 50 \text{ kN m}^{-2}$, higher than that of wind stress on the ice. An estimation of the extent of grounded ridges in the ice that broke out is displayed in Figure 9. The same procedure for calculating the grounded ridge estimate as in Figure 7 is used, with a parent ice sheet thickness of 1.1 m, based on data from the MBS. The summed divergence calculation ranges from March 17 to the time of the breakout on March 24. There are 26 grid cells in the ice

that broke out that are heavily grounded, providing a grounding strength on the order of $\sim 6.6 \times 10^4$ kN from ice ridge keel coupling with the sea floor.

5 Discussion

5.1 Discussion of breakout events

The following section discusses the breakout events from the case studies separately. These events were chosen for a couple of reasons. Many of the other breakout events observed with the radar between 2006 and 2010 lacked radar data for the emplacement of the ice, or there was a short data gap during which the breakout itself occurred such that a precise time could not be determined. Other case study candidates did not have clear post-emplacement SLIEs as the breakouts chosen for this study. Both the February 27, 2009 and March 24, 2010 breakouts occurred with the ice having been in place seven to eight days, just longer than the definition of landfast ice that can be involved in a breakout. Due to the present day conditions of frequent emplacements and breakout sequences near Barrow, the breakouts chosen for this study are representative of this more dynamic regime. Also, this 'new' ice involved in the emplacement process is thick and representative of ordinary first year ice. There is significant and durable accretion during the emplacement process, such that high current speeds themselves are not sufficient to dislodge the ice.

5.1.1 Breakout event 1

Based on the wind stress calculations presented above, the derived wind stress of ~ 1.5 kN m⁻² acting on the landfast ice cover would not have been sufficient to overcome the estimated grounding strength, unless the shorefast ice was wider

than ~50 km. That is, B_a from the force balance (Equation 11) would not have been less than one if the landfast ice width was greater than 50 km. While the ice that broke away was wider than the footprint of the radar, it is highly unlikely that the landfast ice width locally was much greater than 15 km, based on analysis of satellite data for the time period 1996-2008 (Mahoney *et al.*, 2014). Furthermore, since the water depth increases beyond the radar footprint, we assume that the only grounded ridges were those observed by the radar. Therefore, it appears that the ice that broke free on February 27, 2009 was pre-conditioned for the breakout.

Any grounded ridges created during the shoreward deformation on February 19 and 20, would have provided anchoring strength to the landfast ice cover. On February 21, the ice cover experienced offshore wind stress that was greater than that which occurred during the breakout event on February 27 (Figure 5). However, before the February 21 wind event, local sea level was dropping (Figure 5), likely increasing both the grounding area of any grounded ridges and the degree of compaction of their keels. At the time of the breakout on February 27, sea level had just started rising after reaching its lowest point since the beginning of January. We therefore speculate that the drops in sea level that occurred prior to the breakout preconditioned the landfast ice by compacting and effectively shortening the keels of grounded ridges and potentially causing tensional failure of the ice along a system of tide cracks.

We note that on February 21 the water level began rising again while the offshore winds remained stronger than they were at the time of the actual breakout event (Figures 5). The fact that the landfast ice did not break-out at this time

indicates that while the drop in sea level that occurred on February 21 helped stabilize the landfast ice, it either did not lead to significant ridge keel compaction or it was insufficient to result in the tensional failure of the ice at the line where it subsequently detached. Just prior to the actual breakout event, on February 26, the water level dropped to a point almost 0.1 m lower than the minimum on February 21. We therefore speculate that either this drop in sea level exceeded some critical threshold that led to greater keel compaction and flexural failure, or that it was the combination of the two drops in sea level that destabilized the landfast ice prior to its detachment on February 27.

Lacking ocean current data for 2008-2009, it is not known what role currents could have played in the breakout. For instance, onshore currents could have been opposing the offshore winds on February 21, holding the landfast ice in place. Then on February 27, the currents could have been offshore, or not strong enough to fully oppose the wind stress acting on the ice, contributing to or allowing the breakout to take place. As will be seen in the discussion of the March 2010 breakout, ocean currents do not necessarily follow the wind direction in this area. They can respond to the movement of low pressure systems locally or at great distance away.

5.1.2 Breakout event 2

Based on our analysis of atmospheric and oceanographic data available for this event, the landfast ice that broke out on March 24, 2010 does not appear to have been preconditioned in a manner similar to breakout event 1. The water level was dropping to a minimum value for the month just prior to the breakout (Figure 8D), and continued to fall after the breakout. This indicates that the uplift of

shortened ridge keels was not a factor in destabilizing the landfast ice. While the drop in sea level would have increased keel anchoring strength, it is also expected to have compromised flexural and tensile strength of the ice seaward of grounded ridges. For an estimated grounding strength of $\sim 6.6 \times 10^4$ kN, the combined wind and current stress, ~ 65 kN m⁻² was enough to overcome the grounding strength of the ice cover, over the 23 km alongshore extent of landfast ice. The force balance between frictional coupling with the sea floor and the ridge, and wind or current stresses shows that B_a was 1.91, and B_w was 0.57. In this case, the current strength by itself could possibly have caused the breakout provided the wind stress was not opposing the current stress, but the combination of the two in the same direction resulted in a 30% higher stress on the landfast ice over the current stress alone.

For nearly the entire week before the breakout, the local winds were blowing briskly from the east/northeast between 5 and 10 m s⁻¹, peaking at the time of the breakout (Figure 8B and 8C). At the same time, the currents under the ice involved in the breakout were flowing toward the northeast (opposing the winds), except for two short periods of relatively weak southwestward currents (Figure 8F and 8G). Shortly before the breakout on March 24, the currents reversed direction setting toward the southwest and were gaining strength at the time of the breakout. Large sheer stress in areas of relatively steep bathymetry gradients have been associated with current direction reversals, and may have contributed to weakening of the ice seaward of grounded ridges that remained after the breakout (Ohshima, 2000). Water temperature also increased to about -1.4°C before the breakout, and continued to do so afterwards. This inflow of warmer water may have weakened

frozen contacts between blocks in the ridge keel despite increasing grounding strength caused by the drop in sea level.

In an alternative scenario, assuming the increasing grounding strength due to sea level drop was enough to resist the wind and current stresses, rapidly moving pack ice may have come into contact with a prominent landfast ice feature, knocking the ice loose in an event well known by the Iñupiat and termed “tuuq” (Iñupiaq: ice chisel) due to the chiseling action of the floe coming in from the offshore (George et al., 2004). However, inspection of Envisat ASAR imagery acquired before and after the breakout does not support this scenario.

5.2 Discussion of Errors

Assumptions with respect to grounded ridge geometry, the uncertainty and spatial detail resolved by the ice tracking algorithm and other factors all contribute to potential sources of error in this study. Below, we discuss key sources of error and how these errors could affect the conclusions concerning the cause of the breakout events under investigation.

The ridge geometry used in this study is based on the findings of Timco and Burden (1997) who characterized the shape of first-year ridges all of which were floating. Strub-Klein and Sudom (2012) completed a more comprehensive study of floating first-year sea ice ridges, though the parameters used in the current work were not included (i.e. ratio of cross-sectional keel area to sail area and keel angle of repose). The results for ridge geometry characteristics in common between the Timco and Burden (1997) and Strub-Klein and Sudom (2012), such as the ratio of keel depth to sail height, were similar for the Beaufort Sea and Chukchi Sea regions

Figure 10. Plot of strength of frictional coupling of a ridge keel to the sea floor for different ridge keel angles. Black line represents the ridges from February 2009, the gray line is for the March 2010 ridges.

with the results of the former study being well within one standard deviation of the results of the latter. While the results of those studies provide a good baseline for first-year ridge characteristics, grounded ridges are the

focus of this work and so may have drastically different shape characteristics. We are unaware of any specific study of the shape of grounded first-year ice ridges performed to date. One characteristic of a grounded ridge, for instance, that may differ from the Timco and Burden model is the angle of the ridge keel, α_k , taken as 26.6°. For ridges that form in coastal waters and become grounded, this angle may be different. Løset *et al.* (1998) studied a grounded ridge with a keel angle of 10-13°. It may be that the ridge keel angle of repose for a grounded ridge may be steeper than that of a floating ridge, since increased pressure from the compression of the ridge increases the strength of individual freeze-bonds between blocks making of the keel (Repetto-Llamazares *et al.*, 2011). Since the angle of the keel dictates the maximum keel depth given a specific volume of ice, the grounding strength of the keel will be affected. Figure 10 displays the strength of the frictional coupling of the hypothetical ridge keels, formed before the February 2009 and March 2010 breakouts, with the sea floor. The strength of coupling increases rapidly with

770 increasing steepness of the ridge keel angle. There would be no grounding strength
 771 at an angle of zero, with all of the ice mass being incorporated into the ridge sail.
 772 However, this calculation assumes that the geometry of the ridge sail scales with
 773 that of the keel, which may not hold true for grounded keels.

774 Directly related to the ridge keel geometry is the porosity of the ridge keel. In
 775 this study, we assumed a porosity of 0.3, also taken from Nuber *et al.* (2013). Figure
 776 11 displays the strength of the ridge keels' frictional coupling to the sea floor for the
 777 grounded ridges in February 2009 and March 2010 for different ridge keel
 778 porosities. Frictional coupling becomes stronger as porosity increases, due to a
 779 higher number of ridge keels that may become grounded assuming higher ridge keel
 780 porosity. Porosity of the ridge keel is also expected to change with time, and
 781 computer simulations have shown that the shear strength of the ridge itself is
 782 dependent upon porosity of the rubble making up the keel under the ridge's
 783 consolidated layer (Polojärvi and Tuhkuri, 2009). Løset *et al.* (1998) report

Figure 11. Plot of grounding strength for different keel porosities. Black line is from February 2009 ridges, the gray line is from March 2010 ridges.

porosities of .1 to .25 for a grounded first year ridge keel near Spitsbergen using a thermistor string drilled through the keel. The average macro-porosity of ridge keels found in a multi-year study in the Barents Sea was 0.37, ranging from

Figure 12. Plot of grounding strength for different ice thicknesses. Black line is from February 2009 ridges, the gray line is from March 2010 ridges.

0.33 to 0.38 (Høyland, 2007). So a wide range of ridge keel porosities have been reported, though traditional methods for determining keel porosity have been shown to underestimate the water content within

the ridge keel volume, and a different approach using nuclear magnetic resonance showed that the range of keel porosities in first-year sea ice ridges may range between ~0.23 to ~0.50 near Barrow, Alaska (Nuber *et al.*, 2013). The range of frictional coupling strengths between ridge keels and the sea floor for the 2009 landfast ice was 12×10^4 kN at a keel porosity of 0.2 and 40×10^4 kN at a porosity of 0.5. For 2010, this range was 4.5×10^4 kN and 18×10^4 kN over the same range of porosities, indicating the significant contribution to total error by uncertainties in ridge porosity. However, such potential errors may be compensated by related factors. Thus, while higher porosity does increase the contact area of the modeled ridge with the sea floor, it is also likely to result in a reduction in the shear strength of the keel itself or reduced frictional coupling with seafloor since a lower number of ice blocks might contact the seafloor.

Due to the non-uniform nature of the current profile below the ice, the calculated stress imparted by the ocean varies according to the depth range over

which the current magnitude is averaged. In the analysis presented here, we excluded measurements at less than 8 m water depth, which were influenced by the roughness of the bottom surface of the ice, and used a depth range of 8-18 m. Averaging the current magnitude over greater depth ranges tended to result in greater ocean stresses and thus our calculations can be considered conservative.

Another possible source of error in this study that affects the in-situ grounding of sea ice ridges is the thickness of the parent ice sheet. Here, we assume that the ice that contributes to ridge building is level and of the same thickness as ice that formed quiescently at the Mass Balance Site near Point Barrow. Due to the nature of the radar imagery, the thickness of the ice involved in the ridge building process is not known, although we can assume that it is already deformed as suggested by the radar backscatter signatures. In order to determine the effects of ice thickness on grounding strength of the landfast ice cover, the ridge building calculations were carried out across a range of ice thicknesses. Figure 12 displays the results of these calculations. In both instances, the grounding strength increases with increasing ice thickness. We assume that the increase in grounding strength results from increased contact of the ridge keel with the sea floor; such an increase in strength may vary with the number and strength of freeze-bonds between blocks making up the ridge keel, which would also vary with parent ice sheet thickness (Liferov and Bonnemaire, 2005). For instance, in 2009 a 10% decrease in ice thickness would have decreased the estimated grounding strength by about 23%, and an increase in ice thickness by 10% would have increased the estimated grounding strength by almost 26%.

A possible large source of error arises from tracking sea ice through the radar imagery. The average error in tracking sea ice is 1.8 pixels per eight minutes (MV *et al.*, 2013). This corresponds to a velocity of 0.079 m s^{-1} , although there are also uncertainties in the spatial scale of the radar pixel size and timing of the radar image capture. Following the procedure outlined above to determine a hypothetical keel depth, along with this velocity error, the range of keel depths is 15.8 m to 17.1 m. The resulting range of seafloor frictional coupling of such a ridge in 12 m of water would then be $\sim 145 \text{ kN m}^{-1}$ to $\sim 261 \text{ kN m}^{-1}$. Therefore the velocity error associated with ice tracking could greatly increase or decrease the estimated grounding strength of the landfast ice cover. Since the deformation cells (430 m x 430 m) are much larger than the initial bathymetry resolution (100 m), it is not known where in the deformation cell a ridge would ground, and it is thus difficult to estimate the influence of the ice velocity error on the calculated grounding strength. However, we expect the velocity errors entering into the calculations to be much smaller since the area over which the velocity is sampled is larger than the displacement of radar returns, and since the ice deformation is integrated over hours instead of merely eight minutes. The latter in particular reduces statistical errors to values well below the average short-term error.

The errors associated with tracking of ice features can be assessed by comparing the keel depth derived from the velocity fields to a keel depth calculated by other means. An alternate approach to derive keel depth is to determine the distance traveled by radar reflectors from the time of a mobile ice floe's contact with the edge of the landfast ice through visual analysis of radar animations of the

deformation events. The volume of ice involved in the ridging event is found from the total amount of shortening between ice reflectors in the radar imagery, and it can be treated the same as the volume of ice determined in the divergence calculation to arrive at an estimated keel depth. The difference between these different estimates of keel depth is on the order of 1 to 2 m, or about 3 to 5 % of the estimated keel depths. This difference is quite small compared to the keel depths needed to achieve grounding of a ridge, and the uncertainty in the exact grounding depth due to size of the deformation grid cell.

6. Conclusions

The goal of this study was to estimate the grounding strength of landfast sea ice near Barrow, Alaska prior to two breakout events and identify and quantify the process(es) that may have exceeded the anchoring strength and resulted in detachment of shorefast ice. The two case studies presented here illustrate the interplay of a number of factors involved in a breakout. Stresses on the upper and lower surfaces of the landfast ice from winds and currents may exceed (individually or in combination) the effective yield stress of the landfast ice as determined by the anchoring strength of grounded pressure ridges. The 2010 case study clearly demonstrated that currents (peaking at $>1 \text{ m s}^{-1}$) can be sufficiently strong to cause the ice to fail even under conditions where existing ridges are well anchored.

At the same time, the 2009 case study illustrated that preconditioning of the landfast sea ice can play an important role in the events leading up to a breakout, in particular under conditions where stresses acting on the shorefast ice are not as

large as under extreme current and wind forcing. In this specific event, a significant drop in sea level may have compacted and shortened grounded ridge keels, allowing the ridges to detach from the sea floor with a subsequent sea level rise. Intrusion of warm water, commonly associated with a reversal of current direction at the edge of the landfast ice (Pickart *et al.*, 2009), may further have weakened the frictional coupling of grounded ridges with the seabed, or reduced the shear strength of the ridge itself, enough to allow stresses to mobilize the landfast ice. In both cases, a combination of processes and stresses acting on the landfast ice contributed to the breakout events.

7. Future Work

The Iñupiat in northern Alaska have been reporting an increase in the occurrence of breakout events and reductions in shorefast ice stability in recent years (George *et al.*, 2004; Druckenmiller *et al.*, 2009). This study suggests that there may be different underlying causes for individual breakout events, which complicate our understanding of this development. More work is required to evaluate whether any of the stresses acting on the shorefast ice have increased in recent years, either due to changes in ice roughness or increased current or wind speed. Here, we highlight the importance of reduced anchoring strength as a major factor in reducing ice stability. As demonstrated in Figure 12 the anchoring strength is highly sensitive to the thickness of the parent ice sheet that is being deformed. While we have no insight into changes in calculated ice convergence based on the limited data available at this point, there is circumstantial evidence that the

thickness of parent ice sheets integrated into grounded ridges may have decreased in recent years (Kwok and Rothrock, 2009). Such reduction is not simply a result of reduced thickness of the shorefast ice, but would rather be driven by the combination of reductions in thick multiyear ice that can help build ridges through deformation of full-thickness first-year ice cover, as well as potential reductions in the generation of deep draft ridges in the offshore multi-year ice pack. Both of these reductions in deep ridge keel generation potentially reduce the chance of large ridges that could be advected into the nearshore area to become grounded.

These case studies highlight areas of research where more work is needed. The estimated grounding of ridge keels will be improved with better understanding of the macroscopic porosity of the ridge. The number and temperature-dependent strength of contact points between ice blocks making up the unconsolidated portion of the ridge keels is another property of grounded ridges that warrants further investigation to better determine the grounding strength of grounded keels within an ice cover and the effects of substantial short-term sea level drops on the depth these keels reach. Considering the atmospheric and oceanic stresses acting on the landfast ice, direction-dependent form drag based on surface and under ice relief would improve estimation of the force balance for the landfast ice cover.

The spatial resolution of the sea ice velocity fields could also be increased, but due to the amount of time and computing power required this was not performed at the highest possible resolution for this study. An increase in spatial resolution of the ice velocity fields would help improve assessment of areas of actual ridge grounding, and distinguish between convergent deformation of ice or

shoreward movement of ridges in the analysis of changes in ridge radar returns at the SLIE. Improvements in land-based radar detection of near-shore sea ice features could help further refine the sea ice tracking software as well. The flickering of radar targets before the breakout events is in agreement with observations by Mahoney *et al.* (2007a) and Shapiro (1987), and may indicate decoupling of grounded ridges with the sea floor. Detection of flickering of radar targets may provide a short term warning that a breakout event may occur. Greater availability of satellite imagery can also aid in the observation of ice interaction with landfast ice outside of the footprint of the radar.

These findings may not apply to all areas of the Arctic since wind, current, and ice conditions may greatly differ. For instance, ridges may not play a role in the stabilization of the landfast ice or currents may commonly be weaker and less variable than those at Barrow. Nevertheless, with the availability of atmospheric and oceanographic observations, a similar approach for estimating the stresses acting on any type of landfast ice cover is feasible. Comparing these observations with a measure of the ice cover's stabilizing features, whether grounded ridges, confining topographic features or simply the large-scale tensile strength of the ice in the region, allow for a better determination of breakout causes. Such analyses may also help validate and refine models of landfast sea ice strength whose performance strongly depends on the way landfast ice effective tensile strength is represented (Beatty and Holland, 2010). Sea ice observatories, similar to the Barrow Sea Ice Observatory (Druckenmiller et al., 2009), which provide near-real time observations of environmental conditions and landfast ice development, would allow for similar

characterizations of breakout events and derived strengths and stresses acting on the ice cover in other regions, thereby helping advance simulation and prediction of hazardous landfast ice failure events.

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