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Wave–ice interactions in the marginal ice zone. Part 1: Theoretical foundations

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ABSTRACT

A wave-ice interaction model for the marginal ice zone (MIZ) is reported that calculates the attenuation of ocean surface waves by sea ice and the concomitant breaking of the ice into smaller floes by the waves. Physical issues are highlighted that must be considered when ice breakage and wave attenuation are embedded in a numerical wave model or an ice/ocean model.

The theoretical foundations of the model are introduced in this paper, forming the first of a two-part series. The wave spectrum is transported through the ice-covered ocean according to the wave energy balance equation, which includes a term to parameterize the wave dissipation that arises from the presence of the ice cover. The rate of attenuation is calculated using a thin-elastic-plate scattering model and a probabilistic approach is used to derive a breaking criterion in terms of the significant strain. This determines if the local wave field is sufficient to break the ice cover. An estimate of the maximum allowable floe size when ice breakage occurs is used as a parameter in a floe size distribution model, and the MIZ is defined in the model as the area of broken ice cover. Key uncertainties in the model are discussed.

1. Introduction

Access to the seasonally ice-covered seas is increasing due to the impact of climate change (see, e.g., Stephenson et al., 2011) and commercial activities there are proliferating as a result. High precision forecasts of these regions are therefore in great demand. This paper and its companion (referred to as Part 2, Williams et al., submitted for publication) is a step towards making those forecasts as accurate as practicable, by including additional physics that is currently absent in today's ice/ocean models.

Improved spatial resolution has significantly enhanced how models represent the mean sea state and its variability, but it has also highlighted a number of problems that have previously remained hidden. One of them concerns the role of surface gravity waves in shaping the so-called marginal ice zone (MIZ), an important region between the open ocean and the interior pack ice where intense coupling between waves, sea ice, ocean and atmosphere occurs. The MIZ is identified visually as a collection of relatively small floes. Surface waves are the main agent responsible for ice fragmentation and, depending upon wave and sea ice properties, they can propagate long distances into the ice field and still contribute to breakage. Indeed, Prinsenberg and Peterson (2011) recorded flexural failure induced by swell propagating within multiyear pack ice during the summer of 2009, even at very large distances from the ice edge in the Beaufort Sea. (Asplin et al., 2012, further analyzed this event.) While the local sea ice there qualified as being heavily decayed by melting (Barber et al., 2009), and thus more fragile, these observations suggest that such events could occur more frequently deep within the ice pack in a warmer Arctic that is no longer protected by a durable, extensive shield of sea ice.

Interactions between ocean waves and sea ice occur on small to medium scales, but they have a profound effect on the large-scale dynamics and thermodynamics of the sea ice. On a large scale the ice cover deforms in response to stresses imposed by winds and currents. It is customary to model pack ice as a uniform viscous-plastic (VP) material (Hibler, 1979; Hunke and Dukowicz, 1997), but alternatives such as the elasto-brittle rheology of Girard et al. (2010) have been proposed to account for the discrepancies in spatial and temporal scalings of ice deformations between VP model predictions and observations (Rampal et al., 2008; Girard et al., 2009). These models, however, function best when the sea ice is highly compact and sustains large internal stresses with deformation primarily along failure lines.

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In contrast, floe sizes in the MIZ are generally smaller due to wave-induced ice breakage and the ice cover is therefore normally less compact, internal stresses are less important than other forcing because the ice floes are freer to move laterally, and deformations occur more fluently compared to the plastic-like, discontinuous deformation of the compact central ice pack. In this regime, internal stresses arise more from floe-floe contact forces than from any connate constitutive relation that embodies the behaviour of sea ice at large scales. Evidently, a model of the MIZ requires knowledge of how waves control the floe size distribution (FSD). Recognizing this, Shen et al. (1986) and Feltham (2005) have proposed granular-type rheologies for the MIZ that contain an explicit dependence on floe size, while others have presented direct numerical simulations of the MIZ using granular models with either a single floe diameter (e.g. Shen and Sankaran, 2004; Herman, 2011), or with floe diameters sampled from a power-law type FSD (Herman, 2013). Parameterizations for floe size-dependent thermodynamical processes have also been developed (Steele et al., 1989; Steele, 1992).

The distance over which waves induce the sea ice to break, i.e. the width of the MIZ, is controlled by exponential attenuation of the waves imposed by the presence of ice-cover. The rate of wave attenuation depends on wave period and the properties of the ice cover (Squire and Moore, 1980; Wadhams et al., 1988). Wave attenuation is modeled using multiple wave scattering theory or by models in which the ice cover is a viscous fluid or a visco-elastic material. In scattering models, wave energy is reduced with distance traveled into the ice-covered ocean by an accumulation of the partial reflections that occur when a wave encounters a floe edge (Bennetts and Squire, 2012b). Scattering models are hence strongly dependent on the FSD. In viscous models (e.g. Weber, 1987; Keller, 1998; Wang and Shen, 2011a) wave energy is lost to viscous dissipation, so these models are essentially independent of the FSD. We will use an attenuation model that includes both multiple wave scattering and viscous dissipation of wave energy. This means that there is a feedback between the FSD and wave attenuation, since the amount of breaking depends on how much incoming waves are attenuated, and the amount of scattering depends on how much breaking there is.

The notion and importance of integrating wave-ice interactions into an ice/ocean model is not new; indeed it was broached by the third author (VAS) more than two decades ago. Since then, several authors have presented numerical models for transporting wave energy into ice-covered fluids. Masson and LeBlond (1989) were the first to incorporate the effects of ice into the wave energy transport/balance equation that had previously been only used to model waves in open water (Gelci et al., 1957; Hasselmann, 1960; WAMDI Group, 1988; Ardhuin et al., 2010). Masson and LeBlond (1989) studied the evolution of the wave spectrum with time and distance into the ice and their theory was used subsequently by Perrie and Hu (1996) to compare the attenuation occurring in the ice field with experimental data. Meylan et al. (1997) derived a similar transport equation to that of Masson and LeBlond (1989) using the work of Howells (1960), and concentrated on the evolution of the directional spectrum. While, like us, they neglected non-linearity and the effects of wind and dissipation due to wave breaking, they improved the floe model by representing the ice as a thin elastic plate rather than as a rigid body. Doble and Bidlot (in press) have also recently extended the operational wave model WAM into the ice in the Weddell Sea, Antarctica, using the attenuation model of Kohout and Meylan (2008). While this model does not allow for directional scattering, it does include the usual open-water sources of wave generation and dissipation in the same way that Masson and LeBlond (1989) and Perrie and Hu (1996) did.

The above papers give the framework and demonstrate some implementations of wave energy transport into the sea ice, but all neglect ice breakage. In fact, it is only recently that this effect was included by Dumont et al. (2011) (hereafter referred to as DKB) in a wave transport problem. Previous papers modeling ice fracture are those by Langhorne et al. (2001) and Vaughan and Squire (2011). However, those authors only looked at general properties of the ice cover, such as the lifetimes of ice sheets and the width of the MIZ. The method used involved modeling the attenuation of an incident wave spectrum and defining probabilistic breaking criteria to decide when the strains in the ice would exceed a breaking strain. The model of DKB provides a fuller description of the resulting ice cover: it estimates the spatial variation of floe sizes throughout the entire region where breaking occurs and also allows the temporal evolution to be investigated. In addition, it considers the coupling between the breaking and the transport of wave energy.

Although the DKB model is one-dimensional, i.e. it only considers a transect of the ocean, it is theoretically generalizable to include the second horizontal dimension. Before this geometrical restriction is tackled, however, important themes have been identified for discussion and investigation, which is the purpose of this paper. Firstly, we put the work of DKB into the context of previous work on modeling wave energy in ice (Masson and LeBlond (1989); Perrie and Hu (1996); Meylan and Masson (2006)) and we correct their interpretation of the spectral density function. Secondly, we revise the floe-breaking criteria based on monochromatic wave amplitudes employed by DKB, and propose one that is based on wave statistics instead. Numerical issues, sensitivity analyses and model results are reserved until Part 2.

2. Description of the waves-in-ice model

2.1. Overview

Fig. 1 shows the flow of information into and out of the waves-in-ice model (WIM), whose three components, namely advection, attenuation and ice breakage, are discussed in more detail in Section 3. We briefly describe their relationship to the inputs and outputs here.

The advection and attenuation steps depend on the group velocity, c_g, and the attenuation coefficient, $\hat{\alpha}$. Both c_g and $\hat{\alpha}$ depend on frequency in addition to the ice properties. The advection and attenuation steps describe how the wave energy is transported into the ice-covered ocean. The WIM therefore extends contemporary external wave models (EWMs, e.g. WAM, WAVEWATCH III), which typically do not operate in ice-covered oceans. The presence

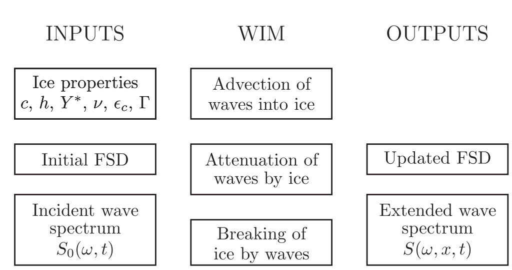

Fig. 1. The information flow in and out of the waves-in-ice model (WIM). An incident wave spectrum with density function $S_0(\omega, t)$ is prescribed at $x = 0$, where ω is the radial frequency (2π multiplied by the frequency), t is time, and x is the spatial variable. The ice properties shown as inputs—respectively the concentration, thickness, effective Young's modulus, Poisson's ratio and breaking strain of the ice, and the viscous damping parameter—combine with the initial floe size distribution (FSD) to affect the three components of the WIM itself: advection, attenuation and ice breakage. This results in the wave spectral density function $S(\omega, x, t)$ being extended into the ice (i.e. into the $x > 0$ region), and in the FSD changing.