Cold Regions Science and Technology 124 (2016) 1–10

Contents lists available at ScienceDirect

Cold Regions Science and Technology

journal homepage: www.elsevier.com/locate/coldregions

Influence of melt-pond depth and ice thickness on Arctic sea-ice albedo and light transmittance

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ARTICLE INFO

Article history:
Received 17 November 2014
Received in revised form 2 December 2015
Accepted 23 December 2015
Available online 6 January 2016

Keywords:
Albedo
Melt pond
Transmittance
Radiative transfer
Arctic sea ice

ABSTRACT

Solar radiation drives the melting of Arctic sea ice in summer, but its parameterization in thermodynamic modeling is difficult due to the large variability of the optical properties of sea ice in space and time. Here, a two-stream radiative transfer model was developed for the propagation of solar radiation in ponded sea ice to investigate the dependence of apparent optical properties (AOPs), particularly albedo and transmittance, on sky conditions, pond depth, ice thickness, and the inherent optical properties (IOPs) of ice and water. The results of numerical experiments revealed that decrease in melt-pond albedo during melting results not only from increase in pond depth but also from decrease in underlying ice thickness, and the latter is more important for thin ice with thickness less than 1.5 m. Hence, a parameterized pond albedo as a function of both pond depth and ice thickness is more suitable for thinning Arctic sea ice than the previously used exponential function of pond depth, which is valid for thicker ice. The increase in broadband transmittance during melting can be explained by the decrease in underlying ice thickness, because its dependence on ice thickness is nearly three times stronger than on pond depth. The spectral dependence of the pond albedo on depth is significant only in the 600–900-nm band, while it depends clearly on ice thickness in the 350–600-nm band. The uncertainty resulting from the absorption coefficient of ice is limited, while the effect of scattering in ice is more important, as determined by a sensitivity study on the influence of the IOPs on the AOPs of sea ice. The two-stream model provides a time-efficient parameterization of the AOPs for ponded sea ice, accounting for both absorption and scattering, and has potential for implementation into sea-ice thermodynamic models to explain the role of melt ponds in the summer decay of Arctic sea ice.

1. Introduction

Melt ponds are characteristic features of Arctic sea ice in summer. They have smaller albedo than dry ice, leading to more shortwave radiation flux into ice and the water beneath the ice (Katlein et al., 2015; Nicolaus et al., 2012) and to further melting and then to reduction in albedo. This nonlinear interaction is called the ice–albedo feedback (e.g. Curry et al., 1995), which is one of the reasons for the rapid decline of Arctic sea ice in summer (Hall, 2004; Pinker et al., 2014). Extensive research on melt-pond evolution and its effect on ice mass balance has been carried out in recent years (e.g. Landy et al., 2014; Schröder et al., 2014; Webster et al., 2015). However, the influence of melt ponds on sea-ice albedo, particularly in climate models, is still inadequately described. In most parameterization schemes, the albedo depends mainly on the surface type and temperature, while a few have also accounted for snow depth and ice thickness (Pedersen and Winther, 2005). Only recently new schemes have been examined for the melt-pond parameterization (Holland et al., 2012; Hunke et al., 2013; Pedersen et al., 2009). A deeper understanding of the fundamental nature of melt ponds and their influence on albedo is required, due to their significance and large spatial and temporal variability.

Many field campaigns have been carried out to investigate melt ponds on sea ice (e.g. Polashenski et al., 2012; Webster et al., 2015). The size of these ponds varied from 1 to 10^5 m^2 (Hohenegger et al., 2012). Their color ranged from bright blue to almost black (Lu et al., 2010), depending mostly on the thickness and optical properties of the underlying ice (Perovich et al., 2009). The pond coverage varied from 0% to nearly 80% in undeformed first-year ice (FYI) and from 0% to 40% in multiyear ice (MYI), due to the undulating MYI surface topography (Nicolaus et al., 2012). Pond depth was mostly less than 50 cm on FYI and less than 70 cm on MYI, varying as ice melting progressed (Morassutti and Ledrew, 1996). Consequently, the melt-pond albedo varied during the melting season from 0.5 in the initial stage to 0.1 for mature ponds, i.e. much lower than the albedo of bare ice and snow (Perovich and Polashenski, 2012). In FYI, melt-pond albedo was typically lower than in MYI (Fetterer and Untersteiner, 1998), and for shallow

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http://dx.doi.org/10.1016/j.coldregions.2015.12.010
0165-232X/

ponds, it was generally larger than for deep ponds (Hanesiak et al., 2001). Ponds on thick ice have larger albedo than ponds on thin ice (Morassutti and Ledrew, 1996), and ponds with a sediment-covered bottom or with cryoconite holes have lower albedo than clean ponds (Eicken et al., 1994).

To account for melt ponds properly in climate models, the dependence of pond albedo on the thicknesses and properties of the pond water and underlying ice needs thorough investigation. Podgorny (1995) introduced a method for calculating the reflectance of a plane-parallel pond and further estimated the pond albedo and radiation distribution in the pond and underlying ice (Podgorny, 1997). Skyllingstad et al. (2009) also developed a method for determining pond albedo by considering solar radiation transfer in a pond and found variations with pond depth and underlying ice albedo. These models prescribed the reflectance of the pond bottom, producing a mathematically too limited relationship between pond albedo and the properties of the underlying ice. Taylor and Feltham (2004) adopted the two-stream radiation scheme of Perovich (1990) and applied it in a three-layer model comprising a melt pond, underlying sea ice, and a refrozen ice layer on the pond surface. The relationship between the albedo and transmittance of ponded ice and the properties of the meltwater and ice was still lacking. Briegleb and Light (2007) developed a radiative transfer scheme to calculate the multiple scattering of solar radiation in sea ice, using the delta-Eddington approach. The apparent optical properties (AOPs), e.g. albedo, transmittance, and absorption, were computed, using the inherent optical properties (IOPs) that specify the scattering-absorption properties of snow, sea ice, and ponds, and included absorbing impurities. This model was sophisticated and feasible for ice melting or growth, but it is computationally expensive (Holland et al., 2012; Hunke et al., 2013).

We aim here to develop a rapid model for the AOPs of ponded sea ice that is applicable for long-term investigations. It can be used to examine the radiation transfer through sea ice and can eventually be embedded in sea-ice thermodynamic models. Our model is based on the principle of Taylor and Feltham (2004), but there are differences in the treatment of the vertical three-layer structure. In Taylor and Feltham (2004), these layers were surface ice, melt pond, and ice below, while in our case, they are melt pond, ice below, and ocean water beneath the ice. The boundary conditions are specified for each layer boundary to finally obtain the analytical solutions for the AOPs. The model description is given in Section 2, and the boundary conditions and analytical solution are presented in Section 3. The results of the numerical experiments are presented in Section 4, followed by a discussion in Section 5. The final conclusions are drawn in Section 6.

2. Model description

2.1. Background

Transfer of solar radiation through sea ice has been treated with several levels of model complexity. Although the theory becomes better level-by-level, the applicability does not improve much due to the strong space–time variability of the structure and resulting optical properties of sea-ice fields in summer (Perovich, 1996). The simplest approach has been to take albedo as a step function for a few categories and use the Beer's model for attenuation of radiation in snow, ice, and water (Maykut and Untersteiner, 1971). This model has an attenuation coefficient (κ_B) accounting for the absorption and scattering. It ignores scattering from the ice bottom and is formally applicable only for infinitely thick layers (Flocco et al., 2015). Two-stream models provide a more realistic family. They are based on the different theories of Dunkle and Bevans (1956) and Chandrasekhar (1960). One class of two-stream model considers downwelling and upwelling irradiance and has been used for snow (Dunkle and Bevans, 1956), glacier ice (Hoffman et al., 2014), and sea ice (Taylor and Feltham, 2004) applications. Another class is based on diffuse and direct radiance and approximating the scattering phase function by a delta function (Joseph et al., 1976) and mainly used for snowpack (Flanner and Zender, 2006), sea ice (Briegleb and Light, 2007), and atmospheric investigations (e.g. Räisänen, 2002). These two classes of two-stream models are physically different, although they come to a similar pair of ordinary differential equations. Sophisticated multistream models based on the discrete ordinates method of Chandrasekhar (1960) have been also developed to treat scattering in more detail and examine the angular distribution of radiance (e.g. Marks and King, 2014).

Here, we follow Taylor and Feltham (2004) and consider downwelling and upwelling irradiance as largely used in recent sea ice models (e.g. Flocco et al., 2015). The limitations of these models are that diffuse incident radiation is assumed and scattering must be taken as isotropic. The former assumption is not a major problem in summer Arctic due to the presence of low stratus cloud cover. The latter assumption may be inappropriate for sea ice with possibly more forward scattering than backscattering, but actually most studies of sea ice IOPs have still treated it as optically isotropic (Katlein et al., 2014). Moreover, internal melting makes sea ice more porous in summer, and then the geometric structure of ice becomes more irregular that can favor isotropic scattering in the ice (e.g. Leppäranta et al., 2003). Consequently, one may anticipate that the isotropic assumption is not badly biased for melting sea ice. The advantages of the present two-stream model are clear. It is mathematically straightforward, and an analytical solution can be obtained. This solution is especially desirable, since the impact of factors such as pond depth and ice thickness is explicitly obtained and computations can be conducted quickly and efficiently, avoiding the possible numerical difficulties of multistream models (Perovich, 1990).

The geometry of the present model is schematically illustrated in Fig. 1. For computational convenience, z denotes the depth in each layer rather than the depth in the entire medium, with subscripts p, i, w for pond, ice, and ocean, respectively. Across a finite vertical layer, irradiance is absorbed and reflected at the layer boundaries, with reflection at the lower boundary acting as gain.

2.2. Governing equations

In the isotropic approach, the IOPs of each layer are defined by the wavelength-dependent scattering coefficient σ_λ and absorption coefficient k_λ. The upwelling and downwelling irradiances in each layer are governed by two coupled first-order differential equations under the

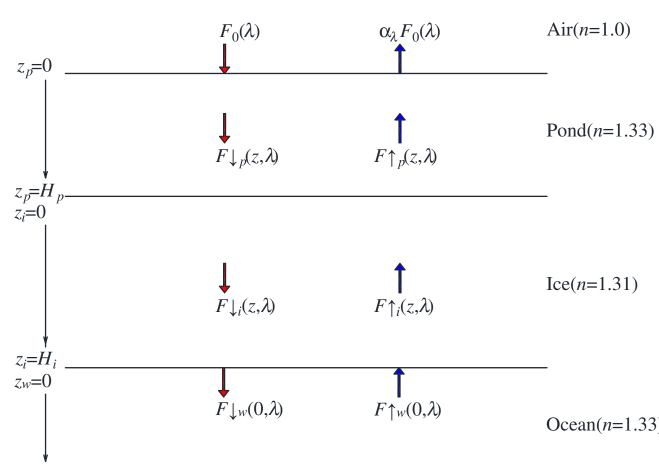

Fig. 1. Schematic graph of the radiative transfer model for the air–pond–ice–ocean system. $F_0(\lambda)$ is the incident solar irradiance; $F^{\uparrow}(z, \lambda)$ and $F^{\downarrow}(z, \lambda)$ are the upwelling and downwelling irradiances with subscripts p, i, w for pond water, underlying ice, and ocean, respectively; H_p is the pond depth; H_i is the thickness of the underlying ice; n is the refractive index with values of 1.0, 1.31, and 1.33 for air, ice, and water, respectively; α_λ is the spectral melt-pond albedo; and λ is the wavelength.