Cold Regions Science and Technology 125 (2016) 12–20

Contents lists available at ScienceDirect

Cold Regions Science and Technology

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

Investigating the thermophysical properties of the ice–snow interface under a controlled temperature gradient Part II: Analysis

Kevin Hammonds *, Ian Baker

Thayer School of Engineering at Dartmouth College, Hanover, NH, USA

ARTICLE INFO

Article history:
Received 23 July 2015
Received in revised form 30 November 2015
Accepted 23 January 2016
Available online 3 February 2016

Keywords:
Ice lens
Interface
Temperature gradient
Effective thermal conductivity
Thermal contact resistance
Kinetic snow metamorphism

ABSTRACT

In order to develop a more intuitive understanding of the physical mechanisms and processes responsible for enhanced kinetic snow metamorphism at the ice–snow interface, we have performed a detailed and quantitative analysis of the in situ micro-thermocouple data originally presented in Part I of this study. In our detailed analysis, we have focused primarily on the observed temperature gradients from within one millimeter above and below the ice–snow interface, as measured in our laboratory prepared specimen. Our findings show via a simple one-dimensional model for energy balance that thermal contact resistance followed by decreases in the effective thermal conductivity are the primary contributors to the dramatic increases in the local temperature gradient near the ice–snow interface. Additional mechanisms for heat and mass transfer are also reviewed in our analysis, including the water vapor flux and latent heat flux.

1. Introduction

Over the last several decades, understanding the thermophyiscal properties of both seasonal and perennial snowpacks has been the focus of many field, laboratory, and modeling studies (Adams and Brown, 1983, 1990; Armstrong, 1985; Brzoska et al., 2008; Calonne et al., 2011, 2014; Colbeck and Jamieson, 2001; Colbeck, 1991; De Quervain, 1973; Flin and Brzoska, 2008; Greene, 2007; Kaempfer et al., 2005; Miller et al., 2003; Perla and Ommanney, 1985; Pinzer and Schneebeli, 2009; Pinzer et al., 2012; Riche and Schneebeli, 2013; Schneebeli and Sokratov, 2004, and others). Presumably, much of this work has been motivated by the inherent and dynamic characteristics of natural snowpacks, such that their thermophysical properties can be highly variable in space and time. Developing an adequate knowledge of these properties is dramatically important to many fields within the cryospheric sciences, including avalanche forecasting, remote sensing, hydrology, glaciology, cold regions engineering, etc. At present, some of the most recent of these studies have demonstrated success with making observations or predictions of the evolution of dry snow via in situ laboratory observations (Riche and Schneebeli, 2013), computed micro-tomography (Flin and Brzoska, 2008; Pinzer et al., 2012), scanning electron microscopy (Chen and Baker, 2010), or three-dimensional modeling (Calonne et al., 2014). However, although highly sophisticated in their approach, none of these more recent studies have begun to account for the more complex and inhomogeneous nature of natural snowpacks, such as those with inherent thermal discontinuities consisting of icy layers or crusts.

In this paper, we present the continuation of a two-part study on the thermophysical properties and processes that occur at an ice–snow interface via a simple one-dimensional model for energy balance. In Part I of this study, Hammonds et al., 2015 (hereafter referred to as Part I), we showed via in situ temperature measurements that a temperature gradient much larger than the imposed bulk temperature gradient can exist near the ice–snow interface. We also showed via in situ X-ray micro-computed tomography (μ-CT) and post-experiment scanning electron microscopy (SEM) that while new ice crystal growth was occurring from deposition upon the bottom (warmer) side of the ice lens, sublimation was occurring from the top (cooler) surface. In our μ-CT analysis, we found that the porosity was generally higher near the ice–snow interface than in the bulk and that the connectivity density increased at a much faster rate below the ice lens than above. We speculated in Part I that this difference in porosity was a likely contributor to the large temperature gradients observed near the ice–snow interface, but we did not attempt to quantify this effect. Also in Part I, we suggested that the diffusion of water vapor and subsequent latent heat flux may be acting to either enhance or diminish these large temperature gradients, but again could only speculate about the proportional contributions of the conductive vs. the latent heat flux.

Presented here in Part II, we give a detailed and quantitative analysis of the mechanisms responsible for the dramatic increase in the temperature gradient we observed near the ice–snow interface. Based on our

* Corresponding author at: Ice Research Laboratory, Thayer School of Engineering, Dartmouth College, 14 Engineering Drive, Hanover, NH 03755, USA. Tel.: +1 603 646 6511 (office); fax: +1 970 481 2745 (cell).
E-mail address: kevin.d.hammonds.th@dartmouth.edu (K. Hammonds).

http://dx.doi.org/10.1016/j.coldregions.2016.01.006

one-dimensional in situ temperature gradient measurements and a simple one-dimensional model for energy balance, we have calculated values for the conductive heat flux, latent heat flux, and water vapor flux over multiple regions of interest within our sample. We have also quantified the proportional contribution from each of these heat flux mechanisms towards the total observed temperature gradient at the ice–snow interface and have attempted to further delineate between the mechanisms responsible for the total conductive heat flux, including the effects of thermal contact resistance and the effective thermal conductivity. Contributions from convective heating within our sample have been neglected based on calculations of the Grashof and Knudsen numbers (see Part I, Section 2.2).

2. Background

2.1. Governing equations

When viewed in a Cartesian coordinate system and for a temperature gradient much greater in the z-direction than in the x- and y-directions, the three-dimensional heat equation can be reduced to a one-dimensional representation (Greene, 2007; Riche and Schneebeli, 2013), as shown in Eq. (1), where ρ (kg m^{-3}) is the density, c_p (J K^{-1} kg^{-1}) is the specific heat capacity at constant pressure, T (K) is the temperature, and t (s) is the time over which the temperature is changing.

$$\rho c_p \frac{\partial T}{\partial t} = -\frac{\partial}{\partial z} \cdot (q_{\text{conduction}} + q_{\text{latent heat}}) \tag{1}$$

For the individual contributions from each of these modes of heat transfer per unit area, Fourier's law of heat conduction (Bergman et al., 2011) and Fick's law for diffusion (Pinzer et al., 2012; Riche and Schneebeli, 2013) can be applied. These equations are given in Eqs. (2) and (3), respectively for $q_{\text{conduction}}$ (W m^{-2}) and $q_{\text{latent heat}}$ (W m^{-2}), where k (W K^{-1} m^{-1}) is the thermal conductivity of the medium, J_z (kg m^{-2} s^{-1}) is the water vapor flux in the z-direction, and L (J kg^{-1}) is the latent heat of sublimation of ice.

$$q_{\text{conduction}} = -k\frac{dT}{dz} \tag{2}$$

$$q_{\text{latent heat}} = J_z L \tag{3}$$

In calculating J_z, we have followed the approach given in Riche and Schneebeli, 2013, shown in Eq. (4), where D_{H_2O} (m^2 s^{-1}) is the water vapor diffusivity calculated as a function of the mean temperature T_M (K) of the region of interest (Massman, 1998), ΔC_z (Pa m^{-1}) is the gradient of vapor pressure, and M_{H_2O} (kg) is the mass of a water molecule. In Eq. (5), ΔC_z is calculated as a function of the equilibrium vapor pressure at the top e_t (Pa) and bottom e_b (Pa) of the region of interest (Murphy and Koop, 2005), where ΔZ (m) is the thickness of the region of interest in the sample, and κ_B (J K^{-1}) is the Boltzmann constant. Equations for D_{H_2O} and e are given in Appendix A.3, while all constants are listed in Appendix A.2.

$$J_z = D_{H_2O}\Delta C_z M_{H_2O} \tag{4}$$

$$\Delta C_z = \frac{|e_b - e_t|}{\Delta Z \kappa_B T_M} \tag{5}$$

2.2. Porous Media

By definition, dry snow is a saturated porous medium consisting of an ice matrix and a fluid mixture of air and water vapor (Bergman et al., 2011; Pinzer et al., 2012). If the fluid saturating the pore space is stationary, meaning that any contributions from convection can be neglected, then the thermal conductivity of the porous medium can be represented by the relative contributions to conduction from the fluid and the solid phase (Bergman et al., 2011). Under steady-state heat flow conditions, the thermal conductivity of the porous medium is best represented by an effective thermal conductivity k_{eff} (W K^{-1} m^{-1}), which when substituted into Eq. (2), gives Eq. (6) for the conductive heat flow in the z-direction, q_z (W m^{-2}).

$$q_z = -k_{\text{eff}}\frac{\Delta T}{\Delta Z} \tag{6}$$

In this treatment, k_{eff} accounts only for the contribution to conductive heat flow from the thermal conductivity of the fluid k_f (W K^{-1} m^{-1}) and solid k_s (W K^{-1} m^{-1}) of the porous medium and does not include any additional contribution from $q_{\text{latent heat}}$. By not allowing contributions from $q_{\text{latent heat}}$, k_{eff} becomes a function of the porosity ε (m^3 m^{-3}) or void fraction of the medium. Eq. (6) can then be written as a function of ε, k_f, and k_s, as shown in Eq. (7) (Bergman et al., 2011, see their Eq. 3.22, p. 120). Lower and upper bounds (Voigt–Reuss bounds, Caruta et al., 2007, p. 99) for k_{eff} can also be derived for a stratified medium if considering heat conduction perpendicular to the layers (Reuss bounds) or parallel to the to the layering (Voigt bounds), as given, respectively, in Eqs. (8) and (9) for $k_{\text{eff,min}}$ (W K^{-1} m^{-1}) and $k_{\text{eff,max}}$ (W K^{-1} m^{-1}) (Bergman et al., 2011, see their Eqs. 3.23 and 3.24, p. 120, respectively).

$$q_z = \frac{\Delta T}{\frac{(1-\varepsilon)\Delta Z}{k_s} + \frac{\varepsilon \Delta Z}{k_f}} \tag{7}$$

$$k_{\text{eff, min}} = \left[\frac{1-\varepsilon}{k_s} + \frac{\varepsilon}{k_f}\right]^{-1} \tag{8}$$

$$k_{\text{eff, max}} = \varepsilon k_f + (1-\varepsilon)k_s \tag{9}$$

Through the calculation of $k_{\text{eff,min}}$ and $k_{\text{eff,max}}$, we can provide a theoretical constraint on the degree to which k_{eff} should be allowed to vary when attempting to satisfy the conditions of energy balance within our system.

3. Methods

As first described in Part I, Section 3.5, our in situ temperature gradient measurements were made using a micro-thermocouple array that had been integrated into the polycarbonate tube that acted as the housing for our ice–snow specimen. We recorded a total of 12 in situ temperature measurements with 4 micro-thermocouples placed within 1 mm above and below the ice lens and 2 additional micro-thermocouples at distances of 8 and 16 mm away from the ice lens. To facilitate our thermodynamic investigations, we have segregated our temperature gradient measurements into five distinct regions, R_1, R_2, R_3, R_4, and R_{ice}. These segments correlate to the five unique temperature gradients we observed over the height of our specimen under conditions of steady-state heat flow and a 24 h recording period. The imposed temperature gradient for this 24 h recording period was held at -100 K m^{-1}. The location of each of these regions is illustrated schematically in Fig. 1, which has been adapted from Part I, Fig. 13. As described in Part I, Section 3.5, all calculations for R_2 and R_3 in Part I were for our target ΔZ of 0.8 mm. In the analysis presented here in Part II, we have allowed ΔZ to vary from 0.6 mm to 1 mm, as the precise position of our micro-thermocouples to within 0.1 mm was unknown. We have added this variability to all of the following analysis in the form of a box-and-whiskers plot that has been overlaid on top of bar plots given to display our quantitative findings.

In order to satisfy the requirements for an energy balance (Kakac and Yener, 1993), we assume that the total heat flux Q_{tot} (W m^{-2}), as