

# A model for co-isotopic signatures of evolving ground ice in the cold dry environments of Earth and Mars



David Andrew Fisher<sup>a,\*</sup>, Denis Lacelle<sup>b</sup>

<sup>a</sup> Department of Earth Sciences, University of Ottawa, Ottawa, Ontario, Canada

<sup>b</sup> Department of Geography, University of Ottawa, Ottawa, Ontario, Canada

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## ABSTRACT

The basics of the model (REGO) are summarized including diffusive water vapor transport, ice content, adsorption and the isotopic ratios of the various water species. Inclusion of a temperature and stress-strain solution is also described as are the effects of cracking when the tensile breaking strengths are exceeded. Cracking is viewed as a multi-scale phenomena, whereby the larger cracks are produced by chaining together smaller ones. After some validation, the model is applied to the upper University Valley (Antarctic Dry Valleys) data and to the Phoenix landing site at lat. 69N on Mars. For the University Valley, the model when forced by measured diurnal temperature cycles and known relative humidity reproduces some of the complex the ice content and isotope structure found in the upper University Valley ground ice cores. The model takes a few hundred years to fill the pores just under the ice table. The excess ice observed in the University Valley cores can be produced by thermal cracking combined with the diurnal temperature cycle. Time taken to achieve the observed excess ice content is of order  $10^4$  years. There is limited subsurface data for the Phoenix site, but there are measured ice table depths, surface thermal properties and Met variables. There is enough to do diurnal and seasonal runs of the REGO model. As with the University Valley, the upper few tens of centimeters icing are controlled by the diurnal temperatures, largely in the warmer part of the year. The observed ubiquitous pore full ice right at the ice table could be produced from a dry soil in about  $10^3$  a and the nearly pure ice patches could be produced by the cracking-diffusion mechanism in about million years, if shear stresses are included. Emplacement of ice down to  $\sim 10$  m is possible using the seasonal temperature cycle and cracking. Excess ice in the upper meter could produced by the model in about  $6 \times 10^6$  a if the temperature at the Phoenix site was 10–20 deg warmer.

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## 1. Introduction

On Earth, stable water isotope ratios are close to those of Earth's oceans (Standard Mean Ocean Water), SMOW, being  $2005.2 \times 10^{-6}$  and  $155.8 \times 10^{-6}$  for  $^{18}\text{O}/^{16}\text{O}$  and D/H, respectively; isotopic ratios, R, hereafter are normally expressed relative to these SMOW values, e.g.  $R_{\text{sample}}(\text{D}) = (\text{D}/\text{H})_{\text{sample}}/(\text{D}/\text{H})_{\text{SMOW}}$ , or equivalently in delta form,  $\delta_{\text{sample}}(\text{D}) = (R_{\text{sample}}(\text{D}) - 1) \times 1000$ ). The Earth's history of the water cycle over the last 1.5 Ma is archived in the stable isotopic ratios ( $^{18}\text{O}/^{16}\text{O}$  and D/H) of the snow and ice layers of glaciers (Reeh et al., 1991; Fisher, 1993; Jouzel et al., 2007). This is due to the fact that terrestrial water isotopes along the water cycle are largely controlled by the distance to source water and temperature

changes from source to final precipitation (Dansgaard, 1961; Fisher, 1990; Johnsen et al., 1997). However, the Earth's large oceans do not allow the ratios of any water type along the water cycle to deviate very far from SMOW; even during ice-ages, the oceans isotopic R varied little from 1 (Dansgaard and Tauber, 1969; Emiliani, 1966).

On Mars, the two large interacting water reservoirs, the ice caps and the ice-bearing permafrost (Squyres et al., 1992; Clifford, 1993) have as yet unknown isotopic ratios (Yung et al., 1988; Fisher, 2007). There are only a few atmospheric values for  $^{18}\text{O}/^{16}\text{O}$  and D/H (Novak et al., 2005; Fisher et al., 2008; Niles et al., 2010) and meteorite values (Jakosky, 1991). These studies suggested that the oxygen isotopes in  $\text{H}_2\text{O}$  and  $\text{CO}_2$  of the martian atmosphere are well mixed in about  $10^4$  years (McElroy and Yung, 1976) so that the  $^{18}\text{O}/^{16}\text{O}$  ratio in water is not a simple function of the water cycle; however, the D/H ratio of water probably is (Yung et al., 1988; Fisher, 2007). If the atmospheric value (average

\* Corresponding author.

E-mail addresses: [dafisher2@sympatico.ca](mailto:dafisher2@sympatico.ca) (D.A. Fisher), [denis.lacelle@gmail.com](mailto:denis.lacelle@gmail.com) (D. Lacelle).

$R \sim 5.0\text{--}10.0$ ) reflected the value of Mar’s large water reservoirs, it would mean that there had been substantially more water in the past than at present. The absence of large open water bodies on recent Mars allows for the D/H ratios to deviate significantly from 1 and, can at certain seasons, reach +10 relative to SMOW, due to the interaction between space losses of water and Milankovitch-driven fluxes from the accessible water reservoirs (Yung et al., 1988; Fisher, 2007; Fisher et al., 2008; Schorghofer, 2008).

An estimate of the initial water mass on Mars could be derived from a cross-reservoir water isotope average. But because studies have used only the available atmospheric D/H values to make an estimate of the initial water mass on Mars, they have reached a wide range of answers with large error bars (Yung and Kass, 1998). Obtaining an average D/H value from the two largest water reservoirs on Mars would help to correct the interpretations of the atmospheric measurements and allow making a reasonable estimate of the initial amount of water on Mars. To derive a more realistic quantitative assessment of water on Mars, one needs to develop a water cycle model for Mars that connects the ice caps, ground ice and atmospheric D/H values, because it is only the atmospheric values at various latitudes and seasons that are currently easily accessible and measurable (Novak et al., 2005).

To help better understand water exchanges between the atmosphere and regolith, the objective of this study is to expand the Fisher (2005) diffusion/thermal contraction–expansion model by incorporating fractionation by phase changes (vapor ↔ ice), Graham fractionation (molecular sieving) and adsorption for all the various species of water. The model calculates the ground ice concentration, stable isotopes ratios ( $^{18}\text{O}/^{16}\text{O}$  and D/H) of the ice and vapor phases and their evolution in the uppermost meters of permafrost. The model traces the vapor flux and icing caused by diurnal and seasonal temperature variations. Diurnal and seasonal temperature waves die out in dry soils by depths  $\sim 0.4$  and  $\sim 10$  m respectively and the runs are over these depth ranges. The stress/strain component in the model allows for arbitrary surface temperatures to drive daily/annual temperature waves into the frozen soils, which can cause cracking and build-up ice density above that allowed by the dry soil porosity.

The structure of the paper is as follows. First, the model (henceforth referred to as REGO), is summarized including diffusive water vapor transport, ice change, adsorption and the development isotopic ratios of the various water species. Inclusion of a temperature and stress–strain solution is also described, as are the effects of cracking when the tensile breaking strengths are exceeded. The details of REGO are given in the Appendices, where their dimensionless forms and difference equations are presented. Following the description of the model, REGO’s ice up rates are compared to the experimental results of Hudson et al. (2009) obtained using a Mars chamber. To verify REGO’s handling of water isotopes, the REGO results are compared to a very simple Rayleigh model for isotope fractionation from a fixed mass of water vapor with specified start co-isotopes that is progressively cooled. Even though the Hudson et al. (2009) work did not include isotopic measurements, the isotopic ratios of forming ground ice from their experiments are predicted and discussed, including the effects of adsorption.

The REGO model is then applied to a terrestrial analogue site in the upper Dry Valleys of Antarctica using measured daily/annual temperature and humidity values. REGO-predicted ground ice content and  $\delta(\text{D})\text{--}\delta(^{18}\text{O})$  composition is compared to measured values. Once the functionalities of the model are demonstrated, REGO is used to estimate the ice-up rates and possible water isotope structure at the Phoenix site on Mars using the local environmental conditions (Mellon et al., 2009; Taylor et al., 2010). These results are discussed in the context of daily and seasonal temperature

fluctuations. Finally, the study touches on the further development of such models.

## 2. Theory

### 2.1. Basic mass conservation equation for water including isotopes

The model concerning just  $\text{H}_2^{16}\text{O}$  vapor and ice is described in Fisher (2005). The vertical coordinate  $z$  is positive downwards and 0 at some surface. Time is “ $t$ ”. Mass conservation in a diffusive regolith is:

$$f \frac{\partial q_v}{\partial t} = \partial \left[ \frac{f D_e \partial q_v / \partial z}{\partial z} \right] - S_i \frac{\partial q_i}{\partial t} - S_a \frac{\partial q_a}{\partial t} q_{rock} \quad (1)$$

where  $f$  is the (still open) porosity of the regolith,  $q_v$  is the vapor density of  $\text{H}_2\text{O}$  vapor in the pores,  $q_i$  and  $(q_a q_{rock})$  are the ice and adsorption bulk water/ice densities within the regolith as a whole (g/cc), and  $D_e$  is the diffusivity of  $\text{H}_2\text{O}$  vapor in the regolith. Each term in Eq. (1) has the units of  $\text{g} (\text{cm}^3 \text{a})^{-1}$  where “ $\text{cm}^3$ ” refers to a cubic centimeter of regolith, including particulates, ice and open pores (Zent et al., 1986). The adsorption model is, that presented in Zent et al. (2001) and its use here, is developed in Appendix 1. The numerical switches,  $S_i$  and  $S_a$  are either 1 or 0 for ice-change and adsorption-change inclusion or exclusion respectively. The dimensionless version of Eq. (1) and the difference equation is given in Appendix 2.

Fig. 1 presents the main variables. The runs presented here are from the ice table (IT) horizon downwards and the amplitude of the temperature variation at the ice table is known or calculated from known surface temperatures and thermal diffusivity. Calculation of the ice table depth will be the subject of a later paper.

Equivalent versions of Eq. (1) are also solved for  $q_{v18}$ : for  $\text{H}_2^{18}\text{O}$ , and  $q_{vD}$  for  $\text{HD}^{16}\text{O}$ . The  $\text{H}_2^{16}\text{O}$  equation is the lead one and it is assumed that  $q_v = q_{v16}$ , i.e. the overall water vapor density is closely given by that of the major species, namely  $\text{H}_2^{16}\text{O}$ . The fractionation coefficients connecting vapor to ice phases are the standard ones that are not well known below about  $-40^\circ\text{C}$ . Details for the  $\text{H}_2^{18}\text{O}$ , and  $\text{HD}^{16}\text{O}$  equations are given in Appendices 2 and 3.

Fractionation coefficients for vapor-adsorption changes are unfortunately rough estimates for temperatures above  $0^\circ\text{C}$  (Caballero and de Cisneros, 2011; Richard et al., 2007) and some discussion appears in Appendix 4. Using them can only indicate rough tendencies in the stable isotopes of adsorption water.

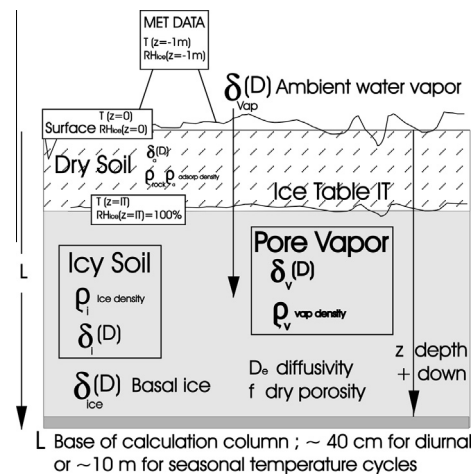


Fig. 1. Basics of the REGO model for moving water vapor into an initially dry soil where it is held as ice or adsorption water. The variables are those used in the model.

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