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Spatial distribution of ^{17}O-excess in surface snow along a traverse from Zhongshan station to Dome A, East Antarctica

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ABSTRACT

The influence of temperature on the triple isotopic composition of oxygen in water is still an open question and limits the interpretation of water isotopic profiles in Antarctic ice cores. The main limitation arises from the lack of ^{17}O-excess measurements in surface snow and especially for remote regions characterized by low temperature and accumulation rate. In this study, we present new ^{17}O-excess measurements of surface snow along an East Antarctic traverse, from the coastal Zhongshan station to the highest point of the Antarctic ice sheet at Dome A. The ^{17}O-excess data significantly decrease inland, with a latitudinal gradient of -1.33 ± 0.41 per meg/degree, an altitudinal gradient of -0.48 ± 0.17 per meg/100 m, and a temperature gradient of 0.35 ± 0.11 per meg/°C. Theoretical calculations performed using a Rayleigh model attribute this inland decrease to kinetic isotopic fractionation occurring during condensation from vapor to ice under supersaturation conditions at low temperatures. However, large heterogeneity of ^{17}O-excess in Antarctic precipitation cannot only be explained by temperature at condensation and/or influences of relative humidity in the moisture source region.

1. Introduction

The relative abundance of water isotopologues $H_2{}^{18}O$ and HDO (expressed as $\delta^{18}O$ and δD) in polar ice cores have long been used as a proxy for past local condensation temperature (Dansgaard et al., 1969; Johnsen et al., 1972; Jouzel et al., 1987; EPICA community members, 2004). The second-order parameter d-excess, defined as d-excess $= \delta D - 8\delta^{18}O$ (Dansgaard, 1964), mostly depends on the sea surface temperature (SST) and relative humidity (RH) at the moisture source region (Merlivat and Jouzel, 1979; Stenni et al., 2001; Uemura et al., 2008; Jouzel et al., 2013; Steen-Larsen et al., 2014b). The d-excess is also modified significantly by the distillation processes along the air mass trajectory (Hendricks et al., 2000; Xiao et al., 2013) and therefore affected by changes in condensation temperature (Uemura et al., 2004; Masson-Delmotte et al., 2008). Thus, the d-excess records in polar ice cores have been used to estimate past climate conditions (SST and/or RH) in the moisture source region (Jouzel et al., 1982; Vimeux et al., 1999; Stenni et al., 2003; Uemura et al., 2012). However, these reconstructions are limited because there are three unknown parameters (condensation temperature, source SST and RH) and only two observational constraints ($\delta^{18}O$ and δD), leading to an under-constrained inverse problem.

With the improvement of water fluorination technique (Barkan and Luz, 2005), the ability to measure $H_2{}^{17}O$ ($\delta^{17}O$) in water with high precision led to the definition of another second-order parameter: ^{17}O-excess $= 10^6 \times (\ln(\delta^{17}O/1000 + 1) - 0.528 \times \ln(\delta^{18}O/1000 + 1))$ (Landais et al., 2006; Barkan and Luz, 2007). Unlike the d-excess, the ^{17}O-excess is mainly controlled by the relative humidity at the moisture source region and is by construction insensitive to source temperature (Landais et al., 2008; Risi et al., 2010; Uemura et al., 2010). First studies with this parameter suggest that precipitation ^{17}O-excess in polar regions is expected to give information on relative humidity at the evaporation regions

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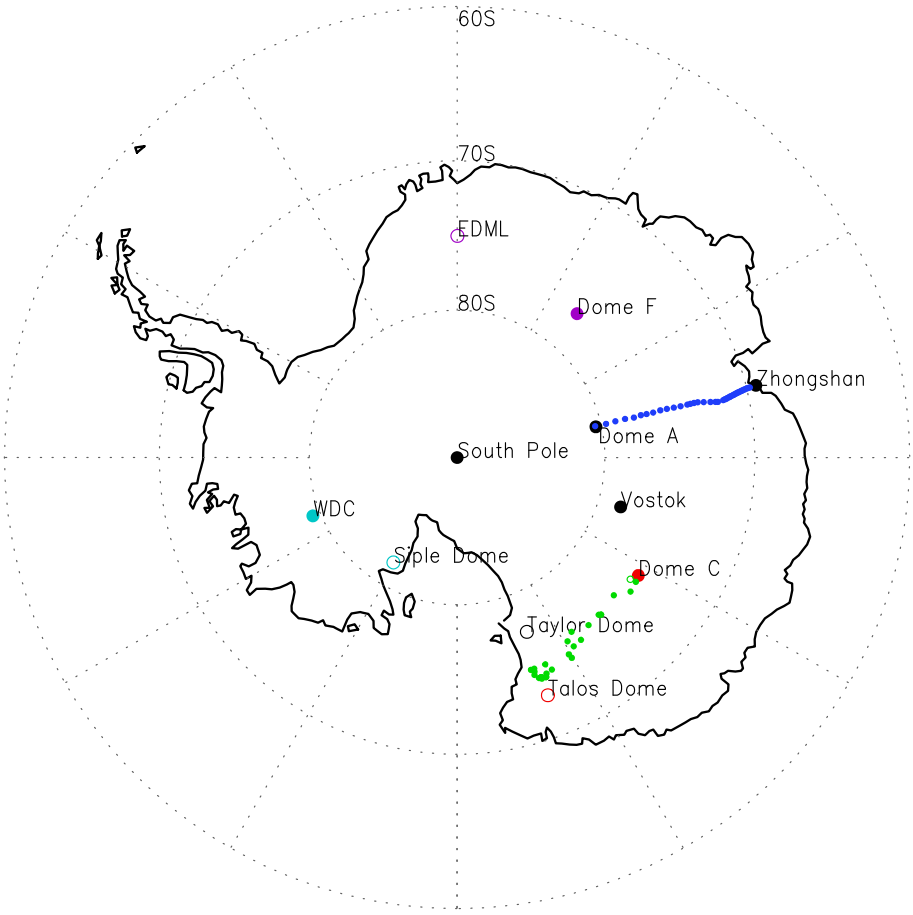

Fig. 1. Map of Antarctica showing sampling locations of surface snow from Zhongshan station to Dome A (the filled blue circles). The sampling locations from Terra Nova Bay to Dome C (Landais et al., 2008) (the filled green circles) together with other Antarctic ice cores sites (the filled cyan circle: WDC, the WAIS Divide ice core; the unfilled cyan circle: Siple Dome; the unfilled black circle: Taylor Dome; the unfilled red circle: Talos Dome; the filled red circle: Dome C; the filled black circle: Vostok; the filled purple circle: Dome F; the unfilled purple circle: EDML) are also indicated.

(Landais et al., 2008, 2012b; Barkan and Luz, 2007; Winkler et al., 2012). As a consequence, combining the d-excess and ^{17}O-excess parameters opens the possibility to constrain past changes in evaporation conditions. However, recent studies have suggested that other processes could also impact precipitation ^{17}O-excess: precipitation re-evaporation (Landais et al., 2010), mixing between vapors from different origins (Risi et al., 2010), kinetic isotopic fractionation associated with condensation of vapor over ice crystals at low temperatures under supersaturation conditions (Angert et al., 2004; Landais et al., 2012a; Risi et al., 2013; Schoeneman et al., 2014), and possible stratospheric vapor intrusions at sites in interior Antarctica where the temperature and snow accumulation rate are very low (Miller, 2008; Winkler et al., 2013). As for d-excess, processes controlling ^{17}O-excess in polar regions are complex, not fully understood, with only sparsely available data. Up to now, there is only one Antarctic transect from Terra Nova Bay to Dome C (hereafter we call it Dome C transect) where the ^{17}O-excess in surface snow samples was measured by Landais et al. (2008) and there are only several locations in Antarctica where the ^{17}O-excess measurements in core cores were performed (Fig. 1 and Table 1). Documenting the spatial variability of ^{17}O-excess in Antarctica is an important gap to fill to progress toward a quantitative understanding of this new proxy.

In this study, we present new ^{17}O-excess measurements in surface snow along an East Antarctica traverse from Zhongshan station to Dome A (Fig. 1), in an undocumented area which encompasses the highest point of Antarctic ice sheet and the most intense present-day isotopic depletion. Section 2 describes the sampling and analytical protocols. The results are described in Section 3, and compared with theoretical distillation calculations performed with a Rayleigh-type model introduced in Section 4. The model-data comparison and discussion (Section 5) is followed by our conclusions.

2. Sampling and isotopic measurements

Surface snow samples were collected along a route from Zhongshan Station to Dome A (80°22′51″, 77°27′23″, 4093 m a.s.l., the highest point of Antarctic ice sheet) during the 26th Chinese National Antarctic Research Expedition (CHINARE) from December 2009 to January 2010. The CHINARE route is located approximately along the 77°E longitude line and is about 1250 km long (Fig. 1). During the journey to Dome A, 42 surface snow samples (named B1, B2, ..., B42) were collected along the route with an average step of approximately 30 km. For each location, the surface snow layer (10 cm in depth) was sampled, and collected using low density polythene plastic bags. All the samples were kept frozen during transportation and storage. According to field observations by stakes on surface mass balance at 2 km intervals along the Zhongshan station to Dome A conducted in 2005 and 2008 (Ding et al., 2011), the percent of surface 10 cm snow layer to its annual net accumulation ranges from 21% to 38% at the coastal region (elevation below 2000 m a.s.l.), and the average percent 102% over the Dome A region (elevation above 4000 m a.s.l.). Water isotopologues (δ^{17}O, δ^{18}O and δD) measurements were performed at the Laboratoire des Sciences du Climat et de l'Environnement