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# Weighting alternatives for water stable isotopes: Statistical comparison between station- and firn/ice-records

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**Abstract**: It is generally accepted that ice cores archive amount-weighted water stable isotope signals. In order to achieve an improved understanding of the nature of water stable isotope signals stored in ice cores annual $\delta^{18}O$ and $\delta^{2}H$ averages (*i.e.* amount-weighted) were calculated for two Antarctic meteorological stations, Vernadsky and Halley Bay, using monthly precipitation amount and monthly net accumulation as weights, respectively. These were then compared with the annual mean $\delta^{18}O$ and $\delta^{2}H$ records of the nearest available ice cores. In addition, at the stations, both arithmetic means (*i.e.* time-weighted) and amount-weighted (precipitation amount and net accumulation used as weights) annual air temperature averages were calculated and then compared to amount weighted annual mean $\delta^{18}O$ and $\delta^{2}H$ using correlation- and regression analyses. The main hypothesis was that amount weighted annual mean water isotope and temperature records from the stations would be able to replicate the annual water isotope signal stored in ice cores to a higher degree. Results showed that (i) amount weighting is incapable of ameliorating the signal replication between the stations and the ice cores, while arithmetic means gave the stronger linear relationships; (ii) post depositional processes may have a more determining effect on the isotopic composition of the firn than expected; and (iii) mean annual air temperature provided the closest match to ice core derived annual water isotope records. This latter conveys a similar message to that of recent findings, in as much as ambient temperature, via equilibrium isotope fractionation, is imprinted into the uppermost snow layer by vapor exchange even between precipitation events. Together, these observations imply that ice core stable water isotope records can be a more continuous archive of near-surface temperature changes than hitherto believed.

Key words: Antarctica, amount-weighted annual average, $\delta^{18}O$ and $\delta^{2}H$ records, polar precipitation, snow accumulation, temperature proxy, vapor exchange.

## Introduction

Water stable isotopes preserved in firn/ice deposits have for decades been considered a valuable climatic proxy (Johnsen *et al*. 1972; Robin 1977), providing complex information on the origin of water vapor and the conditions prevailing during precipitation in glaciated areas, ranging from the tropics (Thompson 2000; Thompson and Davis 2007), through mid-latitudes (Zdanowicz *et al*. 2014), to the Polar Regions (NGRIP 2004; Jouzel *et al*. 2007). In this connection, Antarctica is of exceptional importance, as ice cores from the continent document significant climatic changes over the last several hundred thousand (EPICA 2006) to perhaps million years (Fischer *et al*. 2013). Moreover, due to the strong physical link between ambient temperature and the stable isotope composition of the precipitation (Jouzel and Merlivat 1984; Różański *et al*. 1992), these serve as archives of temperature signals (Steig 2003). Unfortunately, direct long-term temperature measurements at the drill sites are rare, thus, in the exploration of related phenomena, temperature records available from the meteorological stations closest to the drill sites have to be used. According to the classical interpretation, precipitation weighted isotope signals are archived in ice cores through time (Steig *et al*. 1994), serving as discrete, but possibly biased records of ambient temperatures (Jouzel *et al*. 1997).

Nonetheless, approximating the signal stored in ice cores by amount weighting using precipitation fallen is hardly as straightforward as it may at first seem. Having received the initial impetus from a debate in the mid-1990s concerning seasonality acting as a sort of natural weighting (Steig *et al*. 1994; Charles *et al*. 1995), exciting advances have been made in the recent past in the quest to discover whether or not a stronger link exists between amount-weighted stable isotope records (Persson *et al*. 2011) in precipitation and ice cores.

There have been attempts to assess the dependence of the mean composition of isotopic species in annual layers (Isaksson *et al*. 1996; Fernandoy *et al*. 2010; Mulvaney *et al*. 2012), or the dependence of stable oxygen isotope composition and air temperatures recorded on site, in association with accumulation events in a firn core (McMorrow *et al*. 2001, 2004).

In addition, the relationship between annual mean- and amount-weighted temperatures and water stable isotope composition from three ice cores from various points on the Antarctic Peninsula were compared with model derived climate data such as temperature, precipitation, accumulation defined as precipitation minus evaporation (Sime *et al*. 2009). This approach may have its drawbacks, due to the varying spatiotemporal accuracy of the reanalysis products (Miles *et al*. 2008). Nevertheless, the results undoubtedly underline the discrepancies between mean- and amount weighted air temperature signals (Sime *et al*. 2009).

In addition, to complicate the picture further, recent field measurements have shed light on the phenomenon of changes in the isotopic composition of surface snow (upper 5 mm) between precipitation events following the changes in surface isotopic vapor composition (Steen-Larsen *et al*. 2014).

In the present paper, phenomena related to amount-weighted annual averaging is examined in an Antarctic sub-region where the crucial data requirements were met. High-resolution ice core records were available to test the representativeness of the different weighting schemes, which in addition, were reasonably close to a station where not only long-term, and continuous meteorological measurements, but also precipitation stable isotope measurements are conducted. Two research questions were framed: (i) to what extent does amount weighting improve the relationship between the stable isotope composition of precipitation and air temperature at two Antarctic stations and (ii) which weighting approach approximates the actual signal stored in the nearby ice cores better.

## Material and Methods

The two main focal points of the study are (i) Station Vernadsky (VK) at the Northern tip of the Antarctic Peninsula, along with an ice core from nearby James Ross Island (JR), and (ii) Station Halley Bay (HB), located on the Brunt Ice Shelf on the Weddell Sea, and the ice cores in its vicinity (Fig. 1C). These were selected because the longest and most complete stable water isotope and precipitation/accumulation records simultaneously available at these stations, the like of which was unavailable for any other place in Antarctica. As an additional advantage, the two assessed stations represent the characteristic climate zones of the polar region. VK has the characteristics of a windy, wet polar maritime environment (Turner 2015), with high precipitation rates throughout the year, and monthly average temperatures (T) above 0°C in January and February. Average annual precipitation at VK was 395 mm, average annual T was -3.5°C (reference period: 1964–2007). The warmest year in this period was 1989 (-1.2°C), and the coldest 1980 (-6.1°C) (Fig. 1A). By way of contrast, HB has a polar continental climate (Turner 2015), with relatively low accumulation and mean monthly T falling below -25°C, and showing a larger seasonal amplitude than at VK. Average annual net accumulation was 93 mm, and an average annual T was -18.8°C (reference period: 1972–2008). The coldest year in this period was 1997 (-21.2°C), and the warmest 1975 (-16.6°C) (Fig. 1B). It should be noted that the local climatic characteristics are well reflected in the isotopic composition of precipitation at the stations (Różański *et al*. 1993) (Figs. 1A, B). The stable isotope composition of water samples is conventionally expressed in δ notation – multiplied by 1000 – representing a relative deviation from

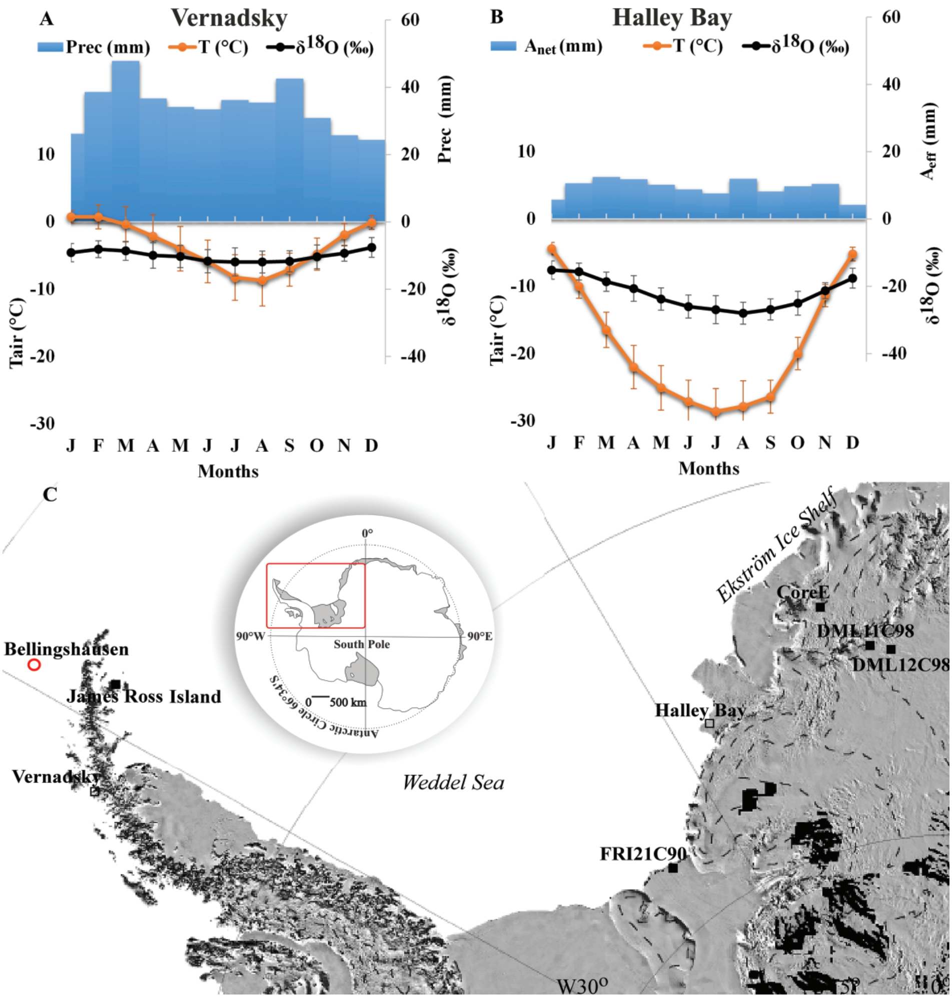

Fig. 1. Location and climate of the study area. Seasonal course of precipitation/accumulation (Prec/$A_{net}$), temperature (T) and precipitation $\delta^{18}O$ at Stations Vernadsky (**A**) and Halley Bay (**B**). The error bars mark the monthly standard deviation of the variables. Location of the stations (with both precipitation stable isotope composition and T measurements – empty squares; with only T measurements – empty circle) and the ice cores (filled black squares) used in the study. The rectangle in the inset map marks the study area, and the black shadings mark the major wind-scour areas redrawn after Das *et al.* (2013).

a standard, thus: $\delta_{sample} = (R_{sample} - R_{standard})/R_{standard}$, where R is $^{18}O/^{16}O$ for oxygen and $^{2}H/^{1}H$ for hydrogen.

In order to calculate the weighted annual averages, monthly records of primary isotopic ($\delta^{18}O$, $\delta^{2}H$) values were acquired from GNIP ISOHIS (IAEA 2013), while the corresponding meteorological parameters were provided by the National Antarctic Scientific Center of the Ukraine, and the British Antarctic

Survey (READER 2013) for VK and HB stations, respectively. Precipitation (Prec) was only available at VK, and monthly net balance available only at HB (Table 1). At HB, monthly net balance was derived from snow gauge field measurements, i.e. 9 snow stakes arranged in a square grid at a spacing of 20 m about 2 km E of the station. The snow levels at the stakes were all measured together once a week and averaged (S. Colwell personal communication). Annual mean water stable isotope records ($\delta^{18}O$, $\delta^{2}H$) were acquired from ice cores located reasonably close to the stations (Fig. 1C and Table 1) to compare with the weighting results obtained at the stations. The chosen scale was annual, since from previous studies it has become obvious that the sub-annual relationship between δ from firn cores (indexed as: $\delta_{ic}$) and T is complicated (Fernandoy *et al*. 2012). Moreover, this focus on annual mean isotopic data is in agreement with the approach of a recent ice core study from Greenland (Masson-Delmotte *et al*. 2015).

The availability of a multi-decadal time span and the monthly resolution (thus free of chronological error) offered an invaluable opportunity, which was exploited by calculating amount weighted annual averages of the GNIP precipitation stable isotope data, described above. These can be called synthetic ice cores. Their mean annual values serve as benchmarks for an on-site sedimentary ice record free from post deposition alterations. These synthetic ice cores can be compared to the annual stable isotope composition of drilled ice cores. A similar approach and an identical term (“synthetic ice core”) was used in a Greenland study (Persson *et al*. 2011); there however, to obtain the signal of their synthetic ice cores, model simulations were used. As was remarked in that study, additional observational constraints should reaffirm their findings on the impact on precipitation weighting. The present study can be considered as an attempt to meet this challenge.

It is a well-known fact that interstitial diffusion dampens the original variability of the isotope signal (Johnsen 1977). Thus, the average surface diffusion lengths for the drill sites were calculated following Johnsen *et al*. (2000), and using the updated ice-vapor equilibrium fractionation factors of Ellehoj *et al*. (2013). It was found that the diffusion length at the different sites never exceeds 0.18% of the average annual accumulation (Table 1). Therefore, as a conservative constraint the 0.18% was multiplied by the number of years, since in real life diffusion diminishes down-core. Thus, the diffusion length did not exceed ~8% of the annual mean accumulation in either studied core. Furthermore, according to recent experimental findings, diffusivity in the firn may be expected to be substantially lower than early estimates (van der Wel *et al*. 2015). Thus, the suspected bias of diffusion and/or its potentially imperfect de-convolution plays an insignificant role in the assessed inter-annual timescale, not interfering with the actual processes assessed.

The moisture source of the stations and the ice cores in the two regions assessed are uniformly under ~1.5 km regarding the annual average of air-parcel

Table 1

Summary table of the dataset ice core record sources and reported uncertainties: [a] ±1yr, Mulvaney *et al*. (2012); [b] no reported dating uncertainty, Graf *et al*. (1994); [c] ~2% of the time interval to the nearest reference horizon corresponding to ~0.5 yr for the section studied, Graf *et al*. (2002); [d] ±3yrs, Isaksson *et al*. (1996).

| | Stations | | Ice cores | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | Halley Bay | Vernadsky | | | JR[a] | FRI21C90[b] | DML11C98[c] | DML12C98[c] | Core E[d] |
| Lat (°, S), Lon (°, W) | 75.58; 26.56 | 65.07; 63.97 | Lat (°, S), Lon (°, W) | | 64.2; 57.68 | 78.31; 39.43 | 74.85; 8.49 | 75.0; 6.49 | 73.6; 12.43 |
| Elev (m asl) | 30.8 | 1 | Elev (m asl) | | 1392 | 76 | 2578 | 2648 | 773 |
| Precipitation (mm) | – | 1964–2008 | Dist. from nearest station (km) | HB | – | 450 | 530 | 580 | 480 |
| $A_{net}$ (mm) | 1972–2012 | – | | VK | 330 | – | – | – | – |
| T (°C) | | 1964–2007 | Annual A (m w.e.) | | 0.63 | 0.22 | 0.19 | 0.06 | 0.32 |
| | | | Mean $diff_{length}$ (m) | | $3.04\times10^{-3}$ | $1.19\times10^{-3}$ | $2.97\times10^{-4}$ | $2.08\times10^{-4}$ | $1.52\times10^{-3}$ |
| $\delta^{18}O$ (‰) | | 1964–2008 | $\delta^{18}O$ (‰) | | – | – | 1960–1997 | 1801–1997 | 1932–1991 |
| $\delta^{2}H$ (‰) | | | $\delta^{2}H$ (‰) | | 1807–2007 | 1966–1989 | – | – | – |

altitudes 5-days-prior to the precipitation events (Suzuki *et al*. 2013). Moreover, the moisture source is the same with regard to the mean source region latitude and longitude in the case of VK and the JR ice cores. Also, it is within a 2° tolerance in relation to the mean source region latitude and the same for longitude in the case of HB and the corresponding ice cores (Sodemann and Stohl 2009).

In Antarctica, the snow deposited on the surface may travel large distances before it finally settles and forms firn and ice deposits. In certain windy areas of Antarctica for example, winds can mobilize more than half of the snowfall (Barral *et al*. 2014). In the two focus areas, the mean (1989–2009) ratio of drifting snow to snowfall was below 10% (Lenaerts *et al*. 2012).

At HB, net accumulation ($A_{net}$) was used as a weight derived from the original monthly net balance time series obtained from the British Antarctic Survey. When the monthly net balance was negative (*e.g*., December 1993; Fig. 2), in that particular month, the monthly $A_{net}$ value is considered to be zero. The corresponding amount is then subtracted from the monthly net balance of the previous month(s) – in this particular case from November 1993; Fig. 2 – and this back-corrected monthly net balance then gives the $A_{net}$ for that particular month (Fig. 2). This procedure is analogous to that employed by McMorrow *et al*. (2001, 2004), and the derived values practically represent net accumulation over monthly periods (Cogley *et al*. 2011). It might therefore be expected that weighting chemical or isotopic tracers with $A_{net}$ makes it possible to simulate more realistically the signal stored in ice and firn (McMorrow *et al*. 2004).

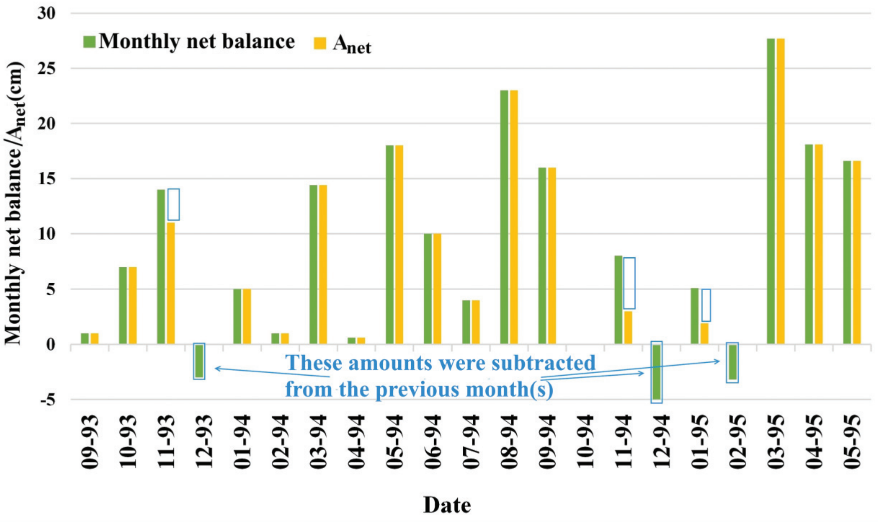

Fig. 2. An example of monthly net balance, and derived monthly net accumulation ($A_{net}$) at Halley Bay (September 1993 – June 1995). Monthly net balance was measured by the British Antarctic Survey.

Therefore, besides arithmetic means (time-weighting), net accumulation (at HB) and precipitation amount (at VK) were the weighting alternatives used in the calculation of the amount weighted annual averages. The procedure may be described as follows:

(i) the calculation of the alternatively weighted time series of $\delta_{ic}$ and T and determination of the correlation coefficient ($r$) and the slope of their linear regression models (Box *et al.* 1994) at each station separately;

(ii) the calculation of the alternatively weighted time series of $\delta^{18}O/\delta^{2}H$ and T at each station and the determination of the correlation coefficient and the slope of their linear regression models with the stable isotope records of the nearby ice cores (Fig. 1C). The hypothesis is that amount weighting should provide the strongest linear relationship, since it takes wind erosion into account.

The amount-weighted annual averages were calculated as

$$A_w = \frac{\sum_{i=1}^{N} A_i \times w_i}{\sum_{i=1}^{N} w_i \times g_i}$$

where N is the number of samples (months) used from each year. $A_i$ denotes the variable ($\delta^{18}O$, $\delta^{2}H$ or T) and $w_i$ denotes the weights ($A_{net}$, Prec or 1 in case of arithmetic mean) in the i^th^ month. The denominator is the summarized monthly weights, where only those events are taken into account which had corresponding $A_i$ (*i.e.* measured δ or T) available. Thus, $g_i = 0$ if there is no $A_i$ and $g_i = 1$ if there was corresponding $A_i$. To avoid major bias, if the number of missing values for a year exceeded 50% for either A or w, than that particular year was considered non-representative.

## Results

**Relationship of weighted δ – T at the stations**. — The amount weighting approach indicated a significantly strengthened linear δ – T relationship at HB (Table 2). Conversely, at VK there was a non-significant improvement in $r$ observed between the various weighting schemes (Table 2A). As for the slopes at both stations for both $\delta^{18}O$ and $\delta^{2}H$, these turned out to be less steep when the temperature time series were weighted than in the case of arithmetic mean ($T_1$) (Table 2B). The patterns observed for the longest individual periods were also similar to the longest common time interval.

**Relationship between the weighted isotope and temperature signals at the stations, and the ice core derived water stable isotope records**. — The comparison of the weighted δ time series of the synthetic ice cores (derived

Table 2

Correlation coefficients (**A**) and slopes (**B**) of the regression models at HB and VK stations between $\delta^{18}O$ and T and $\delta^2H$ and T, respectively, for the longest individual periods (left) and the longest common period of the stations (1972–2007 right). The corresponding slopes of their regression models are given in ‰ (°C)$^{-1}$. $T_1$ stands for arithmetically weighted temperatures, * marks $\alpha = 0.1$, ** marks $\alpha = 0.05$, while *** marks $\alpha = 0.01$ significance level; a dash (–) indicates that a comparison was not made due to the lack of data. The corresponding time series are displayed in Fig. 3.

| | Halley Bay | | Vernadsky | | Halley Bay | | Vernadsky | |
|---|---|---|---|---|---|---|---|---|
| | 1972–2012 | | 1964–2007 | | 1972–2007 | | | |
| | $T_1$ | $T_{Anet}$ | $T_1$ | $T_{Prec}$ | $T_1$ | $T_{Anet}$ | $T_1$ | $T_{Prec}$ |
| **A** | Correlation coefficients | | | | | | | |
| $\delta^{18}O_{Anet}$ | 0.27 | 0.61** | – | – | 0.26 | 0.62*** | – | – |
| $\delta^{18}O_{Prec}$ | – | – | 0.63*** | 0.69*** | – | – | 0.7*** | 0.71*** |
| $\delta^2H_{Anet}$ | 0.34* | 0.63** | | | 0.36* | 0.65*** | | |
| $\delta^2H_{Prec}$ | – | – | 0.59*** | 0.64*** | – | – | 0.64*** | 0.70*** |
| **B** | Slopes | | | | | | | |
| $\delta^{18}O_{Anet}$ | 0.57 | 0.58 | – | – | 0.60 | 0.59 | – | – |
| $\delta^{18}O_{Prec}$ | – | – | 0.67 | 0.67 | – | – | 0.75 | 0.69 |
| $\delta^2H_{Anet}$ | 6.16 | 5.01 | | | 6.83 | 5.11 | | |
| $\delta^2H_{Prec}$ | – | – | 4.19 | 3.92 | – | – | 4.54 | 4.43 |

from the station δ records) and annual mean stable water isotope records of the nearby ice cores provided an insight to answering the question of whether amount weighting improves the correlations or not. The analysis indicated that at HB and the nearby ice cores, weighting with $A_{net}$ did not result in an ameliorated relationship between the station and the cores (Table 3). The slopes of the regression models varied around zero (Table 3). In the case of the $\delta_{ic}$ – T relationships near HB, a similar pattern was observed for all the cores. Only $T_1$ at HB had any significant correlation with the $\delta^{18}O_{ic}$ (Table 3). It should be noted that most of these values do not attain any level of significance; however, their coherent pattern is definitely informative, providing an overall picture (Table 3). As for the slopes, the weighting decreased their values (Table 3).

In the case of VK, all the variables measured at the station followed the $\delta^2H$ records from JR (Fig. 4). The correlation between the precipitation weighted $\delta^2H$ from the station and the ice core was much stronger than that seen in the δ – δ relationships at HB (Table 3). In addition, the slope for $\delta^2H$ between VK and JR was 0.3‰ (°C)$^{-1}$. In the meanwhile, $T_1$ showed a stronger, yet still insignificant, linear relationship, with a slope of 1.83‰ (°C)$^{-1}$, with the $\delta^2H$

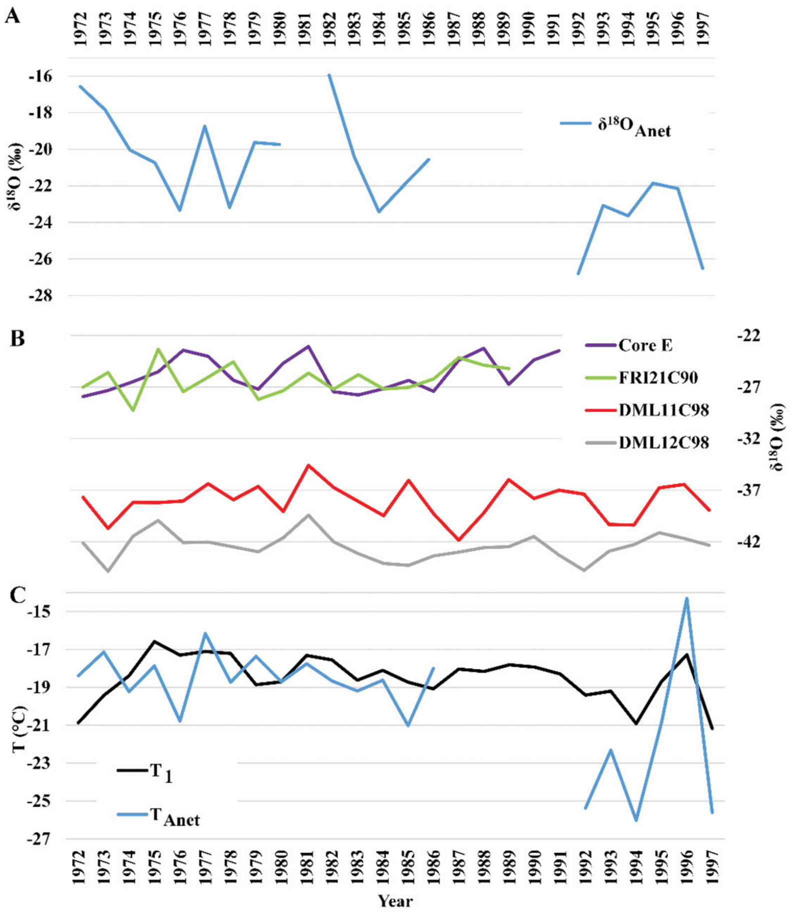

Fig. 3. Comparison of annual means of differently weighted $\delta^{18}O$ (**A**) and temperature (**C**) from Halley Bay, and $\delta^{18}O$ time series of four nearby ice cores (**B**) available between 1972 and 1997.

from core JR than it did with the precipitation weighted ones (Fig. 4; slope: 0.93‰ (°C)$^{-1}$). The annual mean $\delta^2H_{ic}$ at JR was compared to annual mean T at two closely located stations (Abram *et al*. 2011), which provided stronger correlations compared to the annual mean T shown above.

## Discussion

**Relationship of weighted δ – T at the stations**. — As expected, since the T conditions during precipitation events are one of the most determining factors of the δ values (Schlosser 1999), the amount-weighted δ – T gave

Table 3

Correlation coefficients for the δ ice core – station ($\delta_{ic} - \delta$) and the δ ice core – station temperature ($\delta_{ic}$ – T) relationships between the weighted $\delta^{18}$O time series at HB and the nearby ice cores (1972–1997). Corresponding slopes of their regression models are given in ‰ (‰)$^{-1}$ and ‰ (°C)$^{-1}$. Significance levels: α = 0.1 (*); α = 0.05 (**).

| | | Variable | Core E | FRI21C90 | DML11C98 | DML12C98 |
|---|---|---|---|---|---|---|
| Corr. coeff. | $\delta_{ic} - \delta$ | $\delta^{18}O_{Anet}$ | -0.35 | -0.12 | 0.17 | 0.2 |
| | $\delta_{ic}$ – T | $T_1$ | 0.51** | 0.38 | 0.38* | 0.37* |
| | | $T_{Anet}$ | -0.01 | 0.31 | 0.34 | 0.24 |
| Slope | $\delta_{ic} - \delta$ ‰ (‰)$^{-1}$ | $\delta^{18}O_{Anet}$ | -0.21 | -0.08 | 0.08 | 0.09 |
| | $\delta_{ic}$ – T ‰ (°C)$^{-1}$ | $T_1$ | 0.9 | 0.57 | 0.53 | 0.4 |
| | | $T_{Anet}$ | -0.01 | 0.36 | 0.17 | 0.11 |

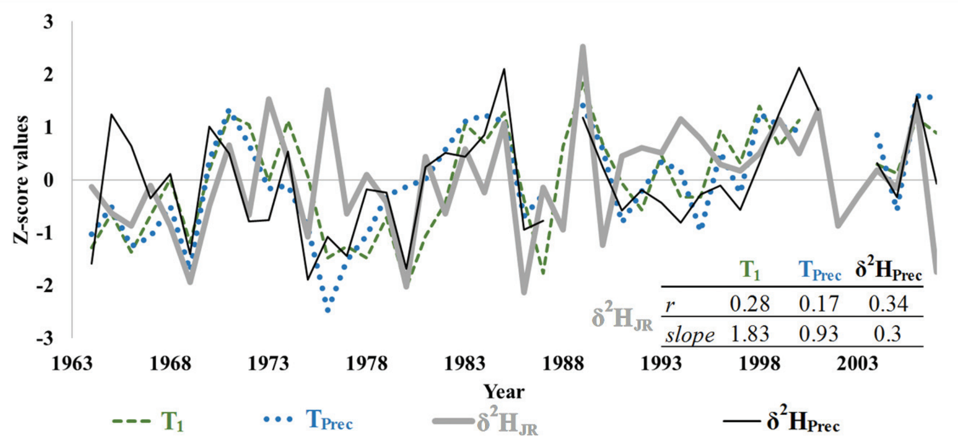

Fig. 4. Comparison of differently weighted temperatures from Vernadsky and $\delta^2$H records from the nearby James Ross Island ice core (1964–2007) abbreviated in the figure as $\delta^2H_{JR}$. The inset table presents the correlation coefficients and slopes. The slopes of the regression models are given in ‰ (°C)$^{-1}$.

an improved linear relationship at the stations, most explicitly at HB. Here, the correlations of the weighted $\delta^{18}$O – T were of the same magnitude as those in the studies of Schlosser *et al*. (2004) from Neumayer station (1981–2000, mostly for precipitation events), and in that of Fernandoy *et al*. (2012) conducted on monthly δ – T records from two maritime stations on the Antarctic Peninsula. Moreover, the slopes of the obtained $\delta^{18}$O – T regressions were also in accordance with those found in literature (Schlosser *et al*. 2004; Fernandoy *et al*. 2012). At HB the improvement in *r* due to the weighting was significant, while at VK no such clear distinction arose. Regarding the slopes, those mostly decreased with weighting.

The phenomenon of the changes being more remarkable at HB, while those are more modest at VK may be related to the relatively lower seasonal variability at VK (Fig. 1A). The coefficient of variation of $A_{net}$ at HB was 89%, while at VK this was only 70% for precipitation. This is in line with the observation and related interpretation that precipitation-weighted temperature records calculated from the ERA-Interim data did not increase the correlation with isotopes in the James Ross Island ice core, indicating that the interannual temperature information preserved in the stable water isotopes at this site is not significantly biased by precipitation variability (Abram *et al*. 2011). Another meaningful difference between the $\delta^{18}O$ signals from a maritime (VK) and a continental (HB) Antarctic station was found in their intra-seasonal variance. The standard deviation (SD) of $\delta^{18}O$ monthly values was calculated for the summer- (defined here as November–April) and winter half years (May–October) for the two stations. While the SD of $\delta^{18}O$ for both summer and winter was roughly 1.7‰ at VK, a pronounced difference (summer $\delta^{18}O$ SD: 1.8‰; winter $\delta^{18}O$ SD 2.1‰) was found at HB. In the Northern Hemisphere, it has been observed that the winter variability of T or δ plays a dominant role governing interannual variability (Persson *et al*. 2011). Consequently, the observation that winter variability is higher at HB and the fact that winter variability was found to be more important regarding the interanual variability (Persson *et al*. 2011) concurs with our observation: a strengthened linear relationship between amount-weighted δ – T, especially at HB.

**Relationship between the weighted isotopic signals at the stations and the ice core derived water stable isotope records**. — The correlation of weighted isotopic signals (δ records) between the stations and the nearby ice cores was inconsistent with the assumption that amount weighting would give an improved weighted signal. The correlation coefficients drew attention to the observation that weighting seems to be incapable of approximating the actual annual signal stored in the nearby ice cores to a higher degree of accuracy.

Except for the lack of interpretability of the negative correlations (Table 3), the strength of the linear relationships between the $\delta^{18}O_{ic}$ and those of precipitation at the stations was between 0.17 and 0.20, remaining under $\alpha=0.1$. This insignificance is not unusual, since monthly means from firn cores and station precipitation δ records at another site located in the maritime Antarctic Peninsula (Fernandoy *et al*. 2012) were also found to be insignificant. This was a case in which the weak correlations and the incoherent pattern were vaguely explained by the high spatial variability of meteorological conditions. Studies dealing with $\delta_{ic}$ – δ relationships, specifically regression models and/or numerical discussions of ice cores and precipitation collected at Antarctic meteorological stations are rather scarce (Fernandoy *et al*. 2012). Therefore, the fact that the slopes vary around 0‰ (‰)$^{-1}$ is quite thought-provoking, and calls for further studies to deepen our understanding of them. In an ideal situation, the slope should be equal to that of the 1:1 line, formally $f(x) = x$.

In our case, however, the phenomena behind the less pronounced correlations are presumed to be post depositional processes, diffusion in the firn (Johnsen 1977; van der Wel *et al*. 2015), the distance between and the geographical setting of the stations and drill sites, the rate of accumulation at the sites (Vinther *et al*. 2010), and some sort of dating error biasing the ice core chronology (Cuffey and Paterson 2010). Although, melt or snowdrift could have also been an additional factor, at the studied drill sites both were insignificant (Picard and Fily 2006; Lenaerts *et al*. 2012).

In addition, it should be noted that the relatively long distance between the stations, those from which the synthetic ice core records were estimated, and the drilled ice cores, stands in need of additional controlled field experiments being conducted (McMorrow *et al*. 2004; Steen-Larsen *et al*. 2014). Thus, such a complex isotope hydrometeorological monitoring over several years and the drilling of shallow ice cores and/or the digging of snow pits at the same site can provide an ideal comparison.

**Relationship between the weighted temperature signals at the stations and the ice core derived water stable isotope records**. — The strong linear relationship between $\delta^{18}O$ and $\delta^{2}H$ in polar precipitation and ambient temperature has been well known for many decades (Dansgaard 1964; Lorius and Merlivat 1977; Masson-Delmotte *et al*. 2008; Steig 2003). The results obtained regarding the relationship between the weighted temperature signals at the stations and the ice core derived water stable isotope records (Table 3) were in harmony with previous findings, in which:

(i) The relationship between $\delta^{18}O$ values from Core E and temperature at HB have been explored for the period 1957–1989 ($r=0.36$) (Isaksson *et al*. 1996; Isaksson and Melvold 2002). In the present study however, the assessed time interval for Core E was two years longer, 1972–1991. The reason why it starts only in 1972 is explained by the need for the ice core water stable isotope record to be comparable when the different weighting methods are tested (see Materials and Methods section). Within the chosen time interval (1972–1991), the linear relationship of $\delta^{18}O_{ic} – T_1$ was $r=0.51$, being stronger than for 1957–1989. If, however, the full time span of the available data (1957–1991) is used for comparison, a fine agreement is obtained with the previous studies.

(ii) Regarding the other three ice cores with $\delta^{18}O$ records, it may be said in general that the linear relationships observed between the $\delta_{ic}$ and the T from the stations are of the same magnitude as those found in ice cores and T records from the Antarctic Peninsula ($0.33<r<0.51$: Peel *et al*. (1988)) and from the vicinity of JR ($\delta^{18}O_{ic}$ – T r=0.42 and 0.46: Abram *et al*. (2011)). In an additional study exploring the influence of precipitation weighting on interannual variability of $\delta^{18}O$ in Greenland (Persson *et al*. 2011), the correlations between the simulated regional T signals and an array of $\delta^{18}O_{ic}$ ranged from ~0.05 to ~0.55. These were of the same magnitude as the results presented here. They were, however, lower than those obtained from the $\delta^{18}O_{ic}$ from the Ekström Ice Shelf, Dronning Maud Land and T recorded at Neumayer Station (0.54<r<0.71: Fernandoy *et al*. (2010)). In contrast, the obtained slopes in the case of the arithmetically weighted models in the present study were all higher than in the one conducted on the Ekström Ice Shelf region (Fernandoy *et al*. 2010). Nevertheless, as a consequence of the amount weighting, the model slopes decreased to the level of those found in literature (Fernandoy *et al*. 2010).

By way of further verification, the $\delta^{2}H_{ic}$ values from JR were compared with both the precipitation weighted and arithmetical mean T for 1969 to 2005 from a closer meteorological station (Bellingshausen, located at 250 km from JR on King George Island; Fig. 1C; AARI (2015)). In this case, again the arithmetic annual mean temperatures showed a stronger correlation (r=0.45) and a higher slope (4.44‰ $(°C)^{-1}$) with the $\delta^{2}H_{ic}$ record than the precipitation weighted ones (r=0.31; a=2.26‰ $(°C)^{-1}$), corroborating the findings from VK (inset table in Fig. 4). By choosing ice cores closer to the stations, the correlation as well as the slope increased. This observation is further reinforced by the improved correlation reported for the JR $\delta^{2}H_{ic}$ and annual mean T from short-distance stations (Abram *et al*. 2011). This, in turn, indicates a stronger and indeed closer relationship, reaching values only slightly below the $\delta^{2}H_{Prec}$ – T correlation observed on site at VK (Table 2).

According to the classical interpretation, comparing the $\delta^{2}H$ and $\delta^{18}O$ signature of solid precipitation (snow) with air temperature recorded at the ground can be misleading in polar regions. In theory, the temperatures prevailing at the cloud formation level is preserved in the water isotope composition of solid precipitation (Dansgaard 1964). This temperature can stand in a varying relation to ground-level temperature and may blur the apparent relationship between the isotopic composition of precipitation and the surface temperature (Jouzel and Merlivat 1984). Nevertheless, it has been documented in coastal Antarctica that a strong relationship does exist between contemporaneous $\delta^{18}O_{ic}$ and local air T on an event-by-event basis (McMorrow *et al*. 2004).

The stronger correlation between $T_1$ and the ice core water stable isotopes compared to amount weighting alternatives challenges their classical interpretation. Nevertheless, it is in harmony with, and even reinforces, recent field evidence from Greenland. Here, it was found that not only the temperatures prevailing during precipitation events are archived in polar snow (Steen-Larsen *et al*. 2014), but also the stable isotope composition of atmospheric water vapor has a close and dynamic relationship with the corresponding weather conditions (Steen-Larsen *et al*. 2013). The isotopic composition of water vapor is taken up by the firn with the exchange between interstitial and atmospheric water vapor (Steen-Larsen *et al*. 2011, 2014), transferring the current temperature signal continuously into the firn even between precipitation events. The previously discussed process was further supported by geostatistical evidence of related data (Hatvani *et al*. 2017).

**Spatial extension of the temporal relationships**. — The individual characteristics of temporal $\delta - T$ relationships obtained at the stations can be compared/extended using a spatial approach (Masson-Delmotte *et al*. 2008). Both the correlation coefficients and the temporal slopes obtained from the $\delta - T$ regressions from the stations are of the same magnitude, but less than those found in a continental-scale study of Antarctica (Masson-Delmotte *et al*. 2008). If we zoom to the regions of the present study, this difference is even clearer. The temporal $\delta^2H - T$ correlations and spatial slopes near VK from the present study are again smaller than those from the Antarctic Peninsula. Conversely, although the temporal correlations and slopes obtained from the $\delta^{18}O - T_1$ models from the surroundings of HB are lower than those in the literature (Masson-Delmotte *et al*. 2008), the weighted $\delta^{18}O - T$ temporal slopes approximated the spatial ones. The slopes of the temporal $\delta_{ic} - T$ regression models of the different amount-weighted schemes at HB (Table 3) and at VK (inset table in Fig. 4) are considerably below those found in the literature (Masson-Delmotte *et al*. 2008), while the slopes of the $T_1$ models at both sites are better able to approximate those found in the literature (Masson-Delmotte *et al*. 2008), as previously seen at Bellingshausen Station. It should be noted that the relative differences between the temporal and spatial slopes discussed in the present study are in accordance with the well-known discrepancy that in general the temporal $\delta - T$ slopes are slightly lower than the spatial ones (Johnsen *et al*. 1995; Jouzel *et al*. 1997). This result fits in with studies in the literature regarding both the correlations and the slopes, in both cases highlighting areas which require further exploration, and points where there is concurrence with recent evidence.

## Conclusions

The research was conducted on stable water isotopes of precipitation, ambient temperature, monthly precipitation amount and monthly net accumulation measured at the Antarctic stations Vernadsky and Halley Bay, respectively, and stable water isotope records from their nearby ice cores. It confirmed that amount weighting, especially at Halley Bay, improves the correlation between precipitation stable isotopes and temperature at the stations against arithmetic annual means, probably due to its climatic characteristics.

As for the $\delta_{ic}-\delta$ relationships between the synthetic- and the real ice cores, the hypothesis that amount weighting would give a fundamentally better replication of the signal had to be rejected. The results highlight that in this case and this setting, amount weighting did not further improve the expected linear relationships. Neither did the amount weighting further improve the $\delta_{ic}-T$ relationships between the ice cores and the nearby stations' records, where arithmetic means ($T_1$) gave the best approximation. This became more pronounced at an additional and closer station (Bellingshausen), (i) reinforcing the importance of taking distance into consideration in amount weighting, and (ii) corroborating the idea that ambient temperatures not only during precipitation events are imprinted in the firn, but also in-between them.

The result of the comparison of the temporal- and spatial slopes was in accordance with that found in the literature. Moreover, the study extended this knowledge by showing that weighting decreased the slopes in most of the cases.

The presented results complement previous studies both over a lengthened time-scale (closer to the present) and in space, as well using additional ice cores and considering all the stations in the region where amount indicators of precipitation and stable isotopes are both measured. Although the findings presented call for further studies, they have meaningful implications for ice core data-assessment and interpretation as well as for the comparison of models.

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