Snow sublimation in Antarctica

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IIE Green Days, June 2022

Sublimation in conditions of drifting and blowing snow

Introduction

Snow sublimation is an important term in the surface mass balance of snow-covered regions such as Antarctica. In conditions of drifting and blowing snow, sublimation is enhanced yet difficult to quantify due to limitations in measurements and simplifications in models.
For example, large-scale models neglect or extremely simplify the contribution of the saltation layer to sublimation of drifting/blowing snow. Although this layer only comprises the lowest few centimetres of the atmosphere, it may strongly contribute to sublimation as saltation events are very frequent and the concentration of drifting snow is highest close to the surface.

Field measurements

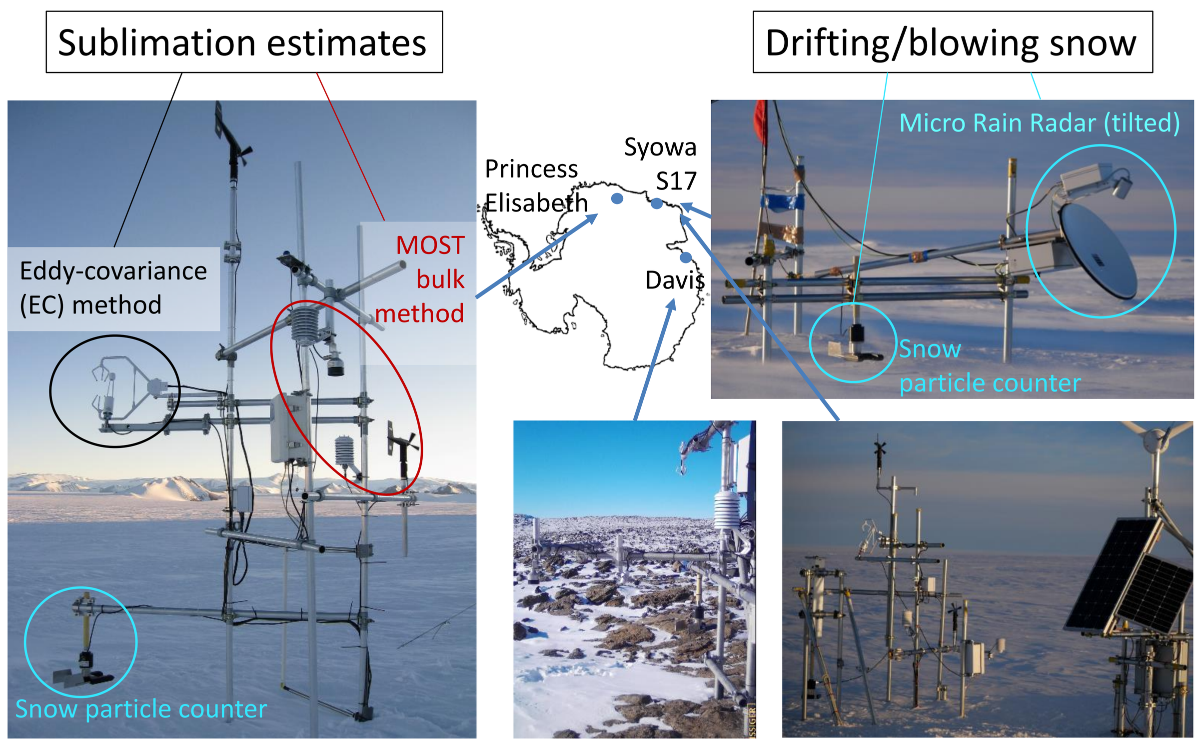


Figure 1: Measurement setups at three sites in East Antarctica.

Large differences between EC and MOST bulk methods (Fig. 2d) can arise from

- invalid assumption of height-constant flux in MOST.
- sensitivity of MOST to instrument uncertainties.
- incomplete removal of artefacts (spikes) in high-frequency EC data if blowing snow reaches sensor height.

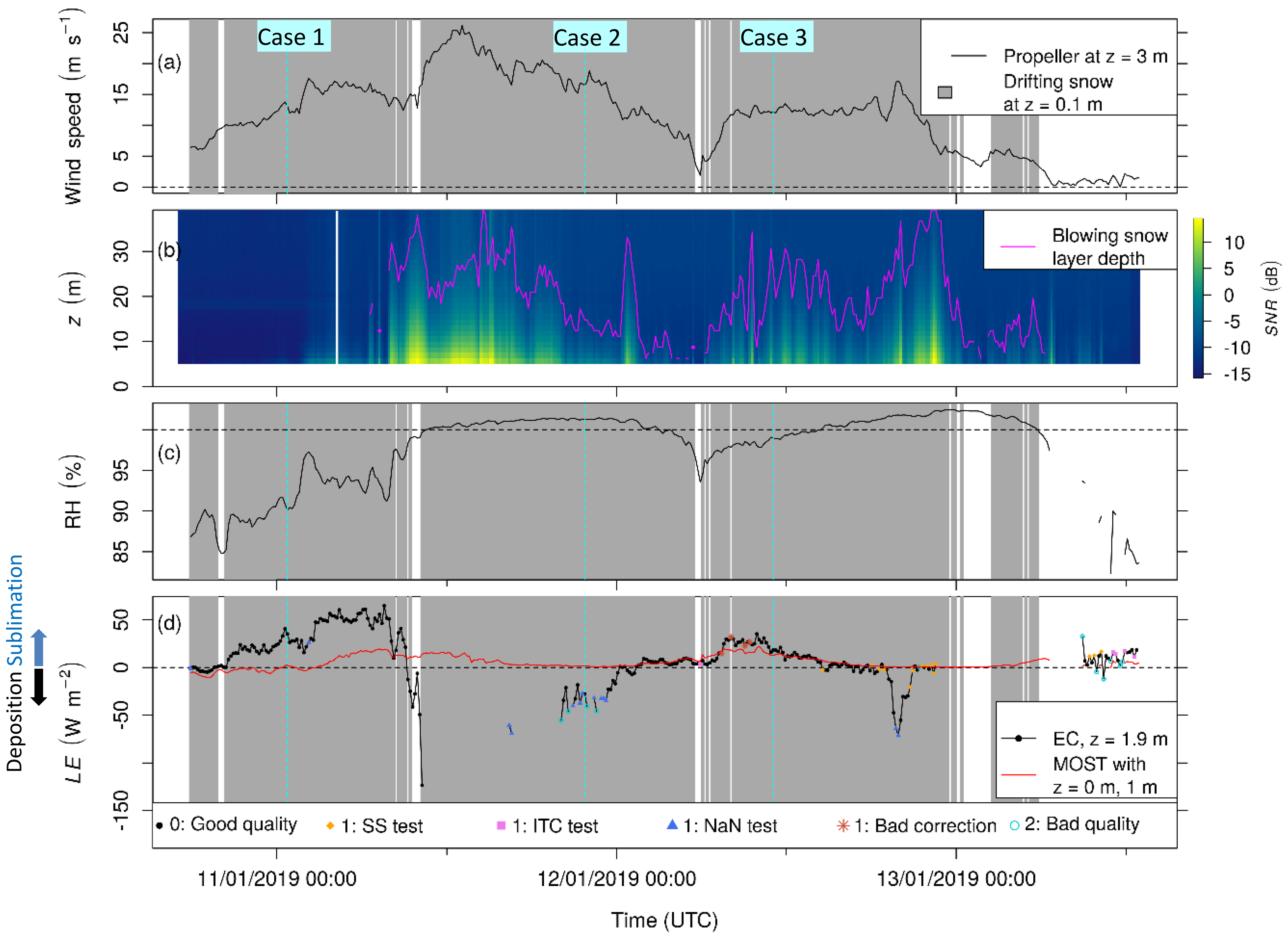


Figure 2: Storm event at Syowa S17 (10-min averages). Dashed vertical lines denote three cases reproduced in LES simulations (Fig. 3,4).

Large-eddy simulations (LES) reveal significant error in MOST bulk method

Simulation method

- LES domain: 38 x 19 x 18 m^3
- Time step: 5×10^{-5} s
- Reproducing field conditions (u, T, q) in a **10-min steady state** (Case 1)
- Drifting/blowing snow is modelled as **Lagrangian particles**, exchanging moisture, heat, and momentum with the air (Eq. 1 and 2)

Results

- Latent heat flux increases with height due to drifting snow sublimation (Fig. 3e)
- **MOST bulk formula strongly underestimates sublimation as drifting snow violates the assumption of a height-constant flux**
- LES and EC measurements agree reasonably well although LES is sensitive to upper boundary condition for humidity (Fig. 3)

$$\underbrace{\frac{\mathrm{d}m_p}{\mathrm{d}t}}_{\text{Mass change of a particle}} = \pi D d_p \underbrace{\left(\rho_{w,\infty} - \rho_{w,p}\right)}_{\text{Vapour density difference}} \mathrm{Sh} \qquad (1)$$

$$\underbrace{c_i m_p \frac{\mathrm{d}T_p}{\mathrm{d}t}}_{\Delta\text{heat storage}} = \underbrace{L_s \frac{\mathrm{d}m_p}{\mathrm{d}t}}_{\text{Latent heat}} + \underbrace{\pi k d_p \left(T_{a,\infty} - T_p\right) \mathrm{Nu}}_{\text{Sensible heat}} \qquad (2)$$

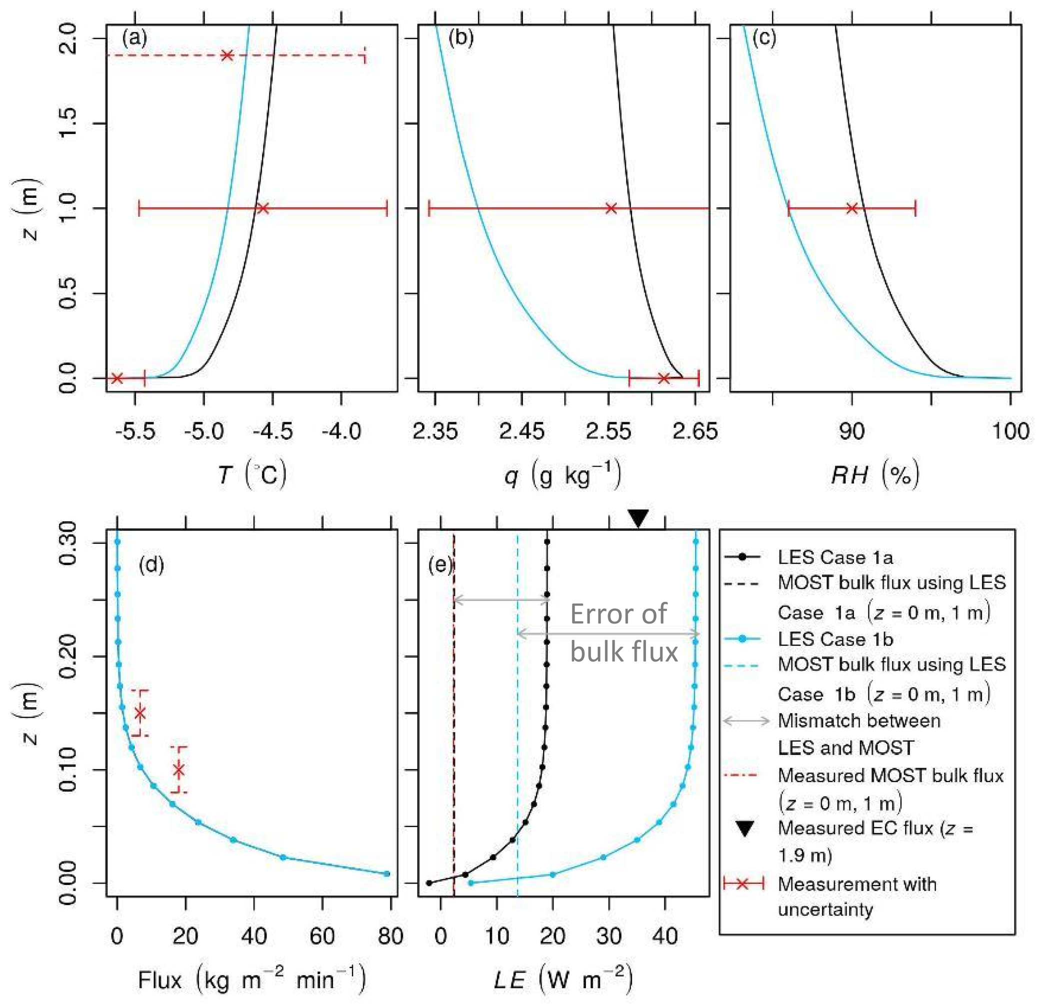


Figure 3: Simulated steady-state average vertical profiles for two different upper boundary conditions for humidity (Cases 1a and 1b).

Parametrizing sublimation in the saltation layer

Approach of Thorpe and Mason (TM, 1966)

- Solving Eq. 1 and 2 by **assuming the change in particle temperature to be zero**
- Sublimation of a drifting snow particle depends on RH, T_a, d_p
- Suitable for suspended particles, staying aloft for a long time
- Can lead to **significant errors in the saltation layer** (Fig. 4) where particle temperatures change along short, ballistic trajectories.

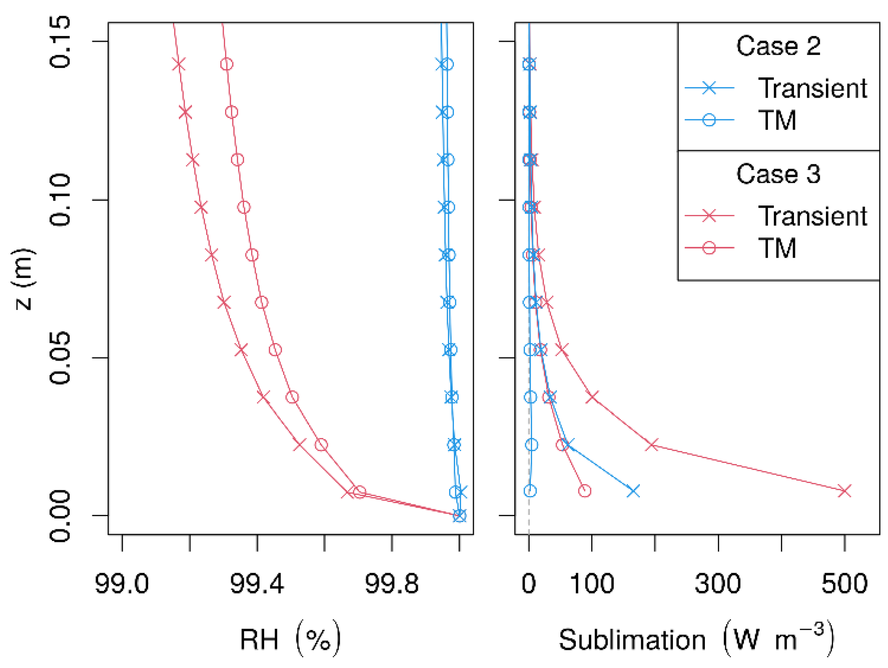


Figure 4: Comparison of LES simulations with transient particle temperature or the Thorpe-Mason (TM) formula for Cases 2 and 3.

Reference:
Thorpe, A.D. and Mason, B.J. (1966). The evaporation of ice spheres and ice crystals, British Journal of Applied Physics 17, 541–548.

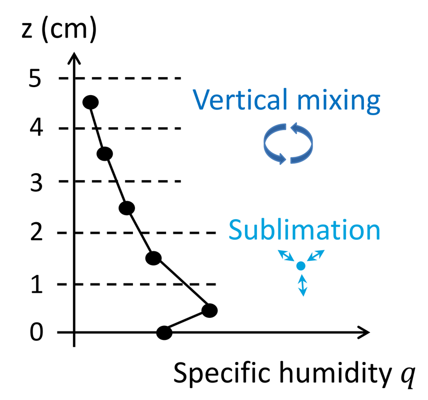
New concept

- Modify the TM formula by including an **empirical term for the change in particle temperature**, to be derived from LES output:

$$\frac{\mathrm{d}T_p}{\mathrm{d}t} = f(T_a - T_0, d_p, z, u_*)$$

- Consider several vertical levels in the saltation layer
- **Prognostically compute humidity and air temperature**:

$$\frac{\mathrm{d}q}{\mathrm{d}t} = K\frac{\mathrm{d}^2 q}{\mathrm{d}z^2} + S_q$$



- Estimate drifting snow concentration with an exponential vertical profile in the saltation layer
- Assume d_p to follow a gamma distribution with $\overline{d_p} = 200$ µm and $\sigma_{d_p} = 115$ µm

Effects of surface sublimation and air mass origin on water vapour δ^{18}O

Modelling approach

1) δ^{18}O of snow and sublimate across Antarctica based on ERA5 hourly weather data

Snowfall δ^{18}O
Linear function of $\overline{T}_a$ (Stenni et al., 2016)

Sublimation
Equilibrium fractionation

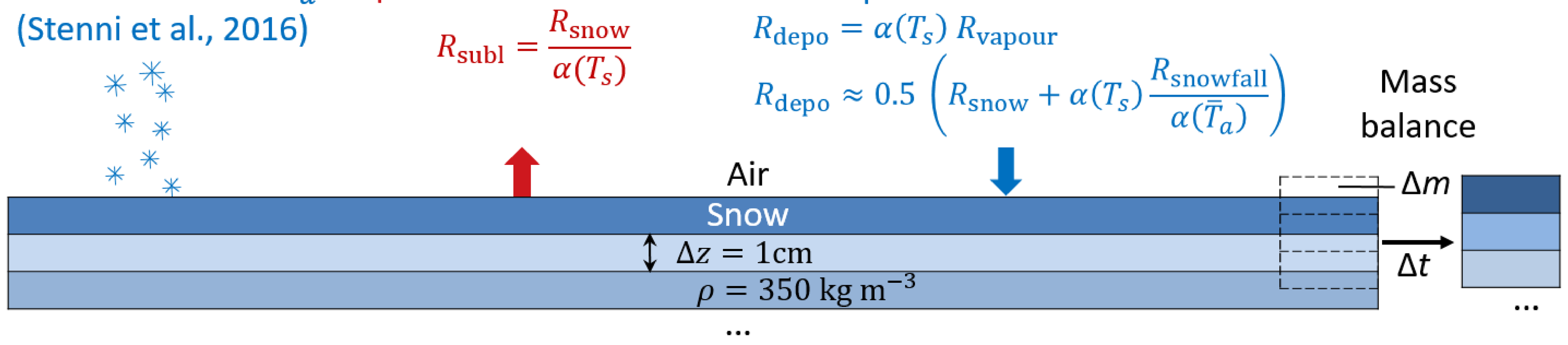
$$R_{\mathrm{subl}} = \frac{R_{\mathrm{snow}}}{\alpha(T_s)}$$

Vapour deposition
Equilibrium fractionation

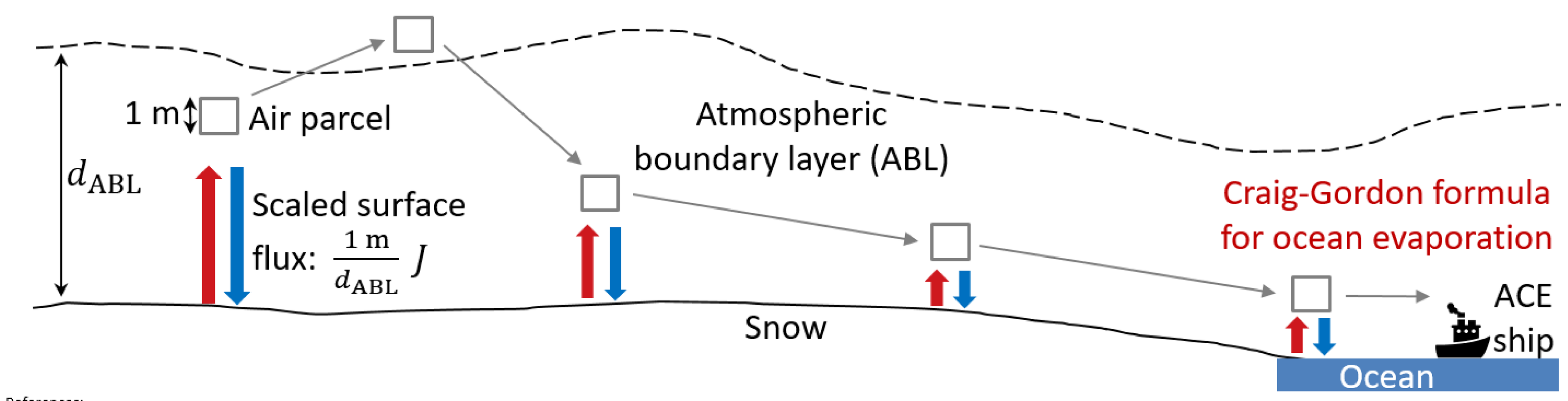
$$R_{\mathrm{depo}} = \alpha(T_s)\, R_{\mathrm{vapour}}$$

$$R_{\mathrm{depo}} \approx 0.5\left(R_{\mathrm{snow}} + \alpha(T_s)\frac{R_{\mathrm{snowfall}}}{\alpha(\overline{T}_a)}\right)$$

Mass balance

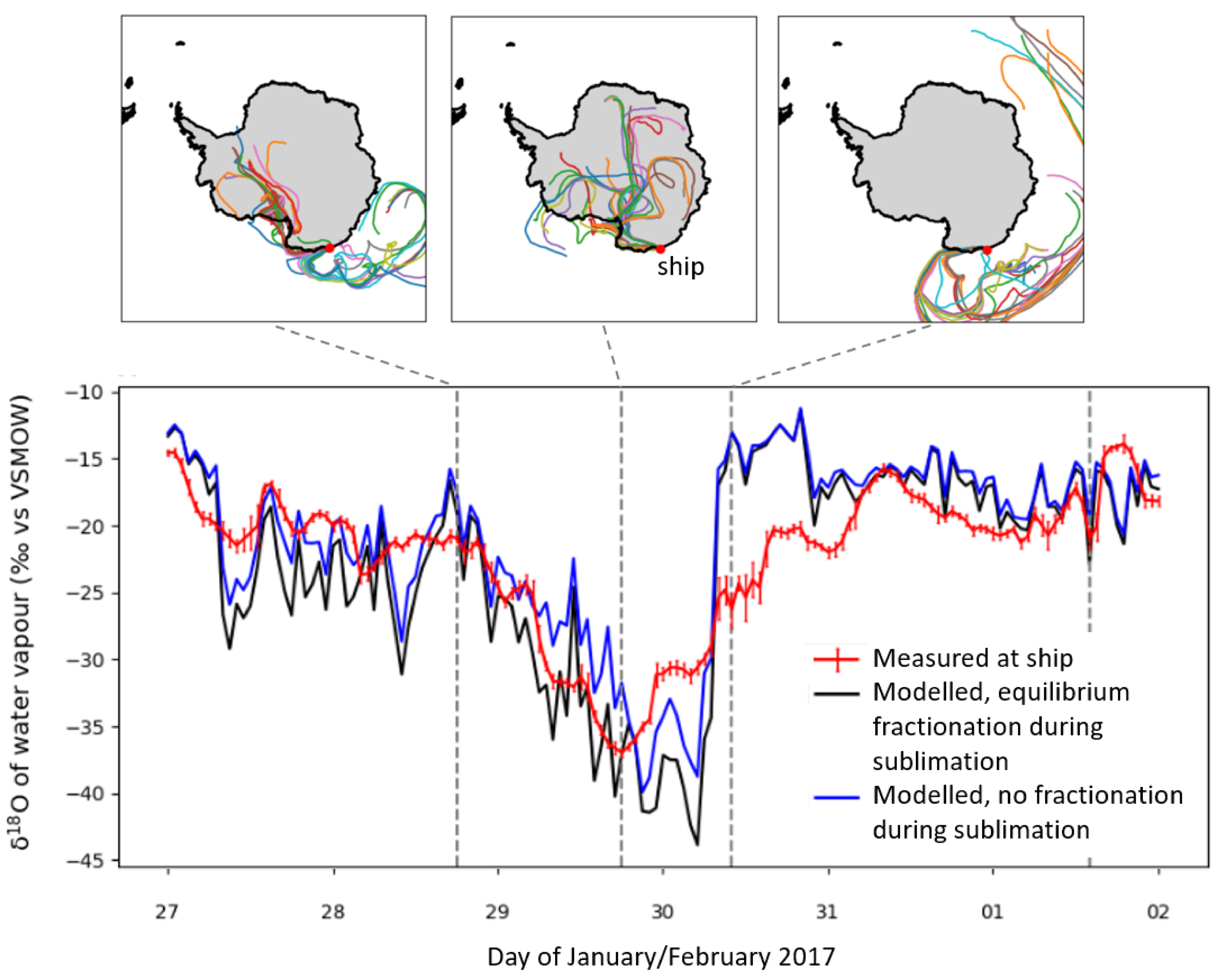


2) Vapour δ^{18}O along backward trajectories (Thurnherr et al., 2020) arriving at coast (Mertz glacier)



References:
Stenni, B., Scarchilli, C., Masson-Delmotte, V., Schlosser, E., Ciardini, V., Dreossi, G., . . . others (2016). Three-year monitoring of stable isotopes of precipitation at Concordia station, East Antarctica. The Cryosphere, 10 (5), 2415–2428.
Thurnherr, I., Wernli, H. and Aemisegger, F. (2020). 10-day backward trajectories from ECMWF analysis data along the ship track of the Antarctic Circumnavigation Expedition in austral summer 2016/2017. (Version 1.0) [Data set]. Zenodo. DOI: 10.5281/zenodo.4031705

Results and discussion



Top panels: 10-day backward trajectories in random colours for three example situations.
Bottom panel: Model-measurement comparison of water vapour $\delta^{18}O$ at the ACE ship close to the Mertz glacier in a 6-day period.

- Simple model **reproduces the striking minimum** in δ^{18}O (−37 ‰)
- Shifts between air masses with **marine and continental origin** are the main driver for vapour δ^{18}O
- Fractionation during snow sublimation has a minor effect on the vapour δ^{18}O, which does not improve the agreement with the measurements