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# Research article

# Asymmetries in cloud microphysical properties ascribed to sea ice leads via water vapour transport in the central Arctic

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Abstract. To investigate the influence of sea ice openings like leads on wintertime Arctic clouds, the air mass transport is exploited as a heat and humidity feeding mechanism which can modify Arctic cloud properties. Cloud microphysical properties in the central Arctic are analysed as a function of sea ice conditions during the Multidisciplinary drifting Observatory for the Study of Arctic Climate (MOSAiC) expedition in 2019–2020. The Cloudnet classification algorithm is used to characterize the clouds based on remote sensing observations and the atmospheric thermodynamic state from the observatory on board the research vessel (RV) Polarstern. To link the sea ice conditions around the observational site with the cloud observations, the water vapour transport (WVT) being conveyed towards RV Polarstern has been utilized as a mechanism to associate upwind sea ice conditions with the measured cloud properties. This novel methodology is used to classify the observed clouds as coupled or decoupled to the WVT based on the location of the maximum vertical gradient of WVT height relative to the cloud-driven mixing layer. Only a conical sub-sector of sea ice concentration (SIC) and the lead fraction (LF) centred on the RV Polarstern location and extending up to 50 km in radius and with an azimuth angle governed by the time-dependent wind direction measured at the maximum WVT is related to the observed clouds. We found significant asymmetries for cases when the clouds are coupled or decoupled to the WVT and selected by LF regimes. Liquid water path of low-level clouds is found to increase as a function of LF, while the ice water path does so only for deep precipitating systems. Clouds coupled to WVT are found to generally have a lower cloud base and larger thickness than decoupled clouds. Thermodynamically, for coupled cases the cloud-top temperature is warmer and accompanied by a temperature inversion at the cloud top, whereas the decoupled cases are found to be closely compliant with the moist adiabatic temperature lapse rate. The ice water fraction within the cloud layer has been found to present a noticeable asymmetry when comparing coupled versus decoupled cases. This novel approach of coupling sea ice to cloud properties via the WVT mechanism unfolds a new tool to study Arctic surface-atmosphere processes. With this formulation, long-term observations can be analysed to enforce the statistical significance of the asymmetries. Furthermore, our results serve as an opportunity to better understand the dynamic linkage between clouds and sea ice and to evaluate its representation in numerical climate models for the Arctic system.

## 1 Introduction

Cloud processes are among the major factors influencing the Arctic climate system. Compared to lower latitudes, Arctic clouds more commonly occur as mixed-phase clouds (MPCs). MPCs consist of ice crystals co-existing with supercooled liquid droplets and are predominantly located at low atmospheric levels (Mioche et al., 2015; Gierens et al., 2020; Korolev and Milbrandt, 2022). Because of their ubiquitous nature, MPCs have a dominant role in important processes like precipitation and the surface radiative energy balance (Korolev and Milbrandt, 2022). The latter is particularly relevant during wintertime since it has been established that MPCs have a significant impact in causing surface longwave radiative warming. This results in reductions in the surface cooling rates being linked to the rapid Arctic warming (Serreze and Barry, 2011; Wendisch et al., 2017), which results in a wintertime Arctic amplification factor. In the central Arctic this factor is about 2.5 times higher than the current Earth's global warming signal (Wendisch et al., 2023).

There are still limitations on our understanding of the Arctic's persistent low-level MPCs due to their counter-intuitive longevity despite instabilities arising from a variety of microphysical and dynamical processes. Surface-related interactions that foster turbulent and cloud-scale upward air motion are highlighted as important processes to maintain MPCs under weak synoptic-scale forcing (Morrison et al., 2012). Surface-turbulence-driven heat and moisture exchange via updraughts can lead to relative humidity increases. These updraughts, when intense enough, can lead to situations of supersaturation with respect to liquid water, hampering the ice growing at the expense of liquid but instead fostering the simultaneous growth of ice particles and supercooled liquid droplets. When dynamic forcing is absent, MPCs are generally unstable (Korolev and Milbrandt, 2022) and prone to ice growth at the expense of vapour deposition, as expected from the Wegener-Bergeron-Findeisen process (Wegener, 1911; Bergeron, 1935; Findeisen, 1938). Feedback processes between the surface and clouds can foster the resilience of mixed-phase clouds when being dynamically coupled to the surface. In addition, local feedbacks among clouds, radiation, and turbulence together with moisture intrusions can lead to the persistence of MPCs even in cases when the cloud is decoupled from the surface's energy and moisture sources (Morrison et al., 2012). Sources of surface energy and moisture in the Arctic are patches of open water in the form of polynyas or leads in the sea ice pack that is otherwise closed.

As defined by the World Meteorological Organization (WMO), leads are elongated areas of open water within the thick pack ice ranging from tens to hundreds of metres in width and tens to hundreds of kilometres in length. In the wintertime, leads are the natural sources of substantial heat and moisture flux, thus warming the atmospheric boundary layer by transferring latent and sensible heat from the ocean to the atmosphere. In wintertime this process is governed by a large temperature difference between the air and water and increases the atmospheric stability over ice floes (Lüpkes et al., 2008; Chechin et al., 2019), whereas in summertime the ocean and air temperatures are quite similar at around 0°C. Furthermore, as Lüpkes et al. (2008) concluded, when sea ice concentration is above 90 % during winter, a change of 1% in sea ice concentration causes a temperature signal of +3.5 K in the near-surface atmospheric temperature. Therefore leads provide an efficient mechanism to modify the atmospheric boundary layer and create unstably stratified conditions in contrast to the atmosphere over the surrounding ice, which is stratified stably (Andreas and Cash, 1999; Michaelis and Lüpkes, 2022). The extreme heat fluxes over leads are typically 2 orders of magnitude higher than over sea ice in winter (Andreas, 1980). As reported by Creamean et al. (2022), sampled concentrations of ice-nucleating particles (INPs), which are important for cloud processes, coincide with the occurrence of sea ice leads and melt ponds. Therefore leads and melt ponds can be thought of as sources of nucleating particles necessary for cloud formation. However, the occurrence of melt ponds happens during the Arctic summer mainly after May, when air temperature is close to 0°C or even slightly above (Creamean et al., 2022), which makes melt ponds not very efficient as sources of sensible heat, in contrast to sea ice leads. This is one reason the present paper focuses on the wintertime; thus the effects of leads can be stressed and better isolated from the cloud observations.

Recent studies based on the analysis of the lead fraction of 200 km around the North Slope of Alaska, in Utqiaġvik, have shed light on the more complex interactions between leads and low-level clouds in the Arctic. Li et al. (2020a, b) found that although open leads foster the creation of lowlevel clouds, newly re-frozen leads tend to promote the dissipation of low-level clouds due to the cut-off of moisture while heat supply is still ongoing. This counter-intuitive result emphasizes the need to study the interaction of sea ice leads with clouds at smaller scales.

The Multidisciplinary drifting Observatory for the Study of Arctic Climate (MOSAiC) expedition from October 2019 to September 2020 was an international effort to study and characterize all aspects of the Arctic atmospheric sea ice, ocean, ecology, and bio-geochemistry system in unprecedented detail, using a variety of approaches and across multiple scales (Shupe et al., 2022; Nicolaus et al., 2022). MO-SAiC is the most comprehensive measurement programme conducted over the central Arctic. The obtained data provide the optimal framework to study coupled systems such as the interaction of sea ice leads and low-level clouds. This gives us the opportunity to scrutinize the effects induced by the occurrence of leads on low-level clouds and to characterize the differences in cloud properties when leads are coupled or decoupled to the clouds.

The paper is structured as followed: in Sect. 2 the set of instrumentation used for this study on board RV *Polarstern* is presented. Section 3 gives a detailed description of the

RV Polarstern central observatory					
Instrument/VAP	Full name	Variables/products *	Resolution		Reference
			Time	Range	-
HATPRO	Humidity and Temperature Profiler microwave radiometer	LWP, IWV	1 s	_	Ebell et al. (2022)
KAZR	Ka-band ARM zenith radar	$Z_{\rm e}, V_{\rm D}, S_{\rm w}$	3 s	30 m	Johnson et al. (2020)
PollyXT	Multi-wavelength Raman lidar	$\beta$ ,	30 s	7.5 m	Engelmann et al. (2016)
CEIL10m	Ceilometer 10 m	$\beta$ , CBH	16 s	10 m	Zhang et al. (2020)
INTERPSONDE	Interpolated Sonde	$T, P, q_{\rm V}, V$	1 min	20 m	Jensen et al. (2020)
GND IRT	Ground infrared thermometer	T <sub>gnd</sub>	1 min	_	Howie and Morris (2020)
Space-borne sensors					
MODIS-AMSR2	*	SIC	1 d	1.0 km	Ludwig et al. (2020)
Sentinel-1A SAR	*	LF	1 d	700 m	von Albedyll et al. (2023)

Table 1. Specifications of instrumentation and data products used in this study.

\* See Appendix D3 for definitions of abbreviations.

methodology developed for the study, and this methodology applied to a case study is presented in Sect. 4.1. The whole MOSAiC wintertime period is statistically analysed and the statistical results for November 2019–April 2020 are presented in Sect. 4.2. Conclusions and an outlook are given in Sect. 5. Supporting material for definitions, methodology, data processing, and further statistical results is summarized in the Appendix.

#### 2 Instrumentation and data products

The suite of atmospheric remote sensing instrumentation on board RV *Polarstern* relevant for this study is mainly comprised of the Atmospheric Radiation Measurement (ARM) mobile facility AMF1 of the US Department of Energy (http://www.arm.gov, last access: 30 March 2023) and the OCEANET-Atmosphere container (hereafter referred to as OCEANET) of the Leibniz Institute for Tropospheric Research (TROPOS) (Engelmann et al., 2016). A list of instrumentation and data products utilized in this paper, along with their spatial and temporal resolutions and references, is summarized in Table 1. An extensive and detailed description of the data availability of all MOSAiC instrumentation can be found in Shupe et al. (2022), their Table B1.

# 2.1 Ground-based atmospheric remote sensing

The primary set of ship-based remote sensing instruments for the observation and characterization of clouds are the Ka-band ARM zenith radar (KAZR) and a ceilometer (from ARM) as well as a PollyXT lidar, and a microwave radiometer (MWR) of Humidity and Temperature Profiler (HATPRO) type, both from TROPOS and installed in the OCEANET container.



**Figure 1.** Synergistic observations with atmospheric remote sensing instruments on board RV *Polarstern* from 18 November 2019 during MOSAiC. (a) KAZR cloud radar reflectivity factor; (b) PollyXT lidar backscattering coefficient; (c) microwave radiometer liquid water path (LWP). Shown isotherms (horizontally aligned) and wind vectors (vertical profile lines) were obtained from radiosonde data at selected altitudes and time steps.

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Figure 1 shows a typical synergy of observations by the KAZR cloud radar, the PollyXT lidar and the MWR for the case study of 18 November 2019. These synergistic observations together with the atmospheric thermodynamic information provided by weather models are imperative for the cloud type classification and retrieval algorithms for cloud macroand microphysics as explained in Sect. 3.1.

# 2.2 Radiosondes

For the characterization of the atmospheric thermodynamic state, the main information is obtained from radiosondes launched from RV Polarstern (Maturilli et al., 2021). For this study, the high-resolution ARM value-added product (VAP) Interpolated Sonde (INTERPSONDE) is used. IN-TERPSONDE is obtained from linear interpolation of the atmospheric state variables from consecutive soundings into a fixed two-dimensional (2-D) time-height grid. The height and time resolutions are 20 m and 1 min, respectively. The grid extends from 10 m up to 40 km altitude (Jensen et al., 2020). In order to account for the thermodynamic interaction between the surface and the atmosphere, in this study the radiosonde vertical profile has been merged with the ground infrared thermometer (GND IRT; Howie and Morris, 2020) as a proxy for surface skin temperature  $T_{gnd}$  [K], which was assigned to an altitude of 0 m in the radiosonde profile.

Relevant atmospheric state variables needed for our methodology are provided or calculated from the radiosonde, e.g. pressure *P* [Pa], air temperature *T* [°C], specific humidity  $q_v$  [gg<sup>-1</sup>], relative humidity [%], wind speed [ms<sup>-1</sup>], and wind direction [°] (see Table 1 and references therein). Derived atmospheric state quantities are virtual potential temperature ( $\theta_v$ ) [K], water vapour transport (WVT) [kg s<sup>-1</sup> m<sup>-2</sup>], the bulk Richardson number (*Ri*<sub>b</sub>), planetary boundary layer height (PBLH) [m], and cloud-driven mixing layer height (CMLH) [m] above cloud top and below cloud base.

# 2.3 Satellite-based information for sea ice conditions

Space-borne sensors are the main source of information for long-term and large-scale monitoring of sea ice conditions in the Arctic. For this study, two main sea ice state variables are used: sea ice concentration (SIC) and the lead fraction (LF).

# 2.3.1 Sea ice concentration

For the observation of sea ice concentration, satellite-borne instruments like the Advanced Microwave Scanning Radiometer 2 (AMSR2) and the Moderate Resolution Imaging Spectrometer (MODIS) are the most reliable instruments in terms of spatial and temporal continuity. These types of instruments, however, have limitations intrinsic to their measurement principles. AMSR2 is a microwave radiometer that is less influenced by clouds than optical sensors and has a good spatial coverage but is limited by its low spatial resolution of about 4 km at 89 GHz or coarser at lower frequencies. On the contrary, MODIS is an optical sensor and offers a higher spatial resolution of 1 km, but its observations are restricted to cloudless conditions. In order to exploit the best features of both sensors, Ludwig et al. (2020) have developed a merged 1 km MODIS–AMSR2 product by tuning SIC from the MODIS 1 km resolution to preserve the AMSR2 average SIC.

The merged MODIS–AMSR2 sea ice product is of particular relevance for the present study since it provides the benefit of potentially detecting open water leads within sea ice due to its finer resolution. We note that leads covered with thin ice, which occur during winter conditions, are not necessarily detected. Nevertheless, for instance, the south–northaligned sea ice lead on 15 April 2020 observed by Sentinel-1 SAR (Krumpen et al., 2021, their Fig. 3) is resolved by the MODIS–AMSR2 SIC retrieval (Ludwig et al., 2020) but not by the 25 km resolution Ocean and Sea Ice Satellite Application Facility (OSI SAF) product (Lavergne et al., 2016), as shown in Fig. D2a and b, respectively.

Nonetheless, a recent study by Rückert et al. (2023) has shown that warm air intrusion events occurring during the MOSAiC drift in April 2020 fostered the formation of a large-scale surface glazing that resulted in an underestimation of SIC retrievals of about 30%. This compromises the accuracy of the ARTIST sea ice (ASI) algorithm used by the AMSR2 retrievals (Spreen et al., 2008) and thus also affects the MODIS–AMSR2 product. Therefore, in order to evaluate the accuracy of the MODIS–AMSR2 product, an alternative SIC product needs to be used. Here, the OSI SAF SIC product (Lavergne et al., 2016) was chosen, mainly because of its availability, coverage and higher accuracy during MOSAiC for April 2020 as shown by Rückert et al. (2023). The details of the MODIS–AMSR2 SIC versus OSI SAF product evaluation are described in Appendix C.

# 2.3.2 Divergence-derived sea ice lead fraction

The satellites Sentinel-1A and Sentinel-1B from the European Space Agency use active microwave synthetic aperture radar (SAR) in the C-band to capture the microwave properties of the sea ice. They are a valuable source to detect leads.

While the most common application is to classify leads from the backscatter coefficient of the SAR scenes (e.g. Murashkin et al., 2018), there is another approach that focuses on the formation process of the leads as seen in sea ice divergence (e.g. Kwok, 2002; von Albedyll, 2022).

By calculating sea ice drift and sea ice divergence from sequential SAR scenes, leads show up in the sea ice divergence whenever the ice moves apart. Such lead fractions from SARderived sea ice divergence have the advantage that they indicate the strong local change in ice velocity when a lead opens. They indicate the exact location of leads, they are independent of cloud coverage, and their magnitude is directly



**Figure 2.** (a) SIC from the MODIS–AMSR2 merged product from 18 November 2019. (b) Lead fraction from SAR Sentinel-1. The images are centred on the position of RV *Polarstern* (red star) on the given date. The RV drift is indicated by the black line; the circle indicates the 50 km radius as the region of interest. The grey cone indicates the relevant observation sector determined by the wind direction at the maximum water vapour transport (see text Sect. 3.4).

linked to the widths of the leads without requiring sensor calibration (Kwok, 2002).

Here, LF is calculated from divergence as described in von Albedyll et al. (2021), von Albedyll (2022), and von Albedyll et al. (2023). The results are interpreted as the average LF per grid cell, which is subsequently drift-corrected and rendered with a spatial resolution of 700 m. One limitation on the lead detection by the divergence-based method is that it only detects new openings. Stationary leads, i.e. leads that do not open or close further, are not detected on the days following the formation even though the leads still exist. Those stationary leads during winter will likely be covered by thin ice. Figure 2 shows a comparison of MODIS–AMSR2 and SAR Sentinel-1 sensors for the sea ice situation on 18 November 2019 around RV *Polarstern*, illustrating the contrasting capabilities of MODIS–AMSR2 SIC (Fig. 2a) and Sentinel-1 divergence-based LF (Fig. 2b).

LF from SAR divergence-based data is available for the study period, except for the time between 14 January and 15 March 2020 (vertical dashed grey lines in Fig. 3), when RV *Polarstern* was north of the latitudinal coverage extending up to 87° N of the Sentinel-1 satellite. To extract the mean LF of a certain region, e.g. 50 km around RV *Polarstern* (Fig. 3a), the average of all grid cells that are located completely or partly in the region of interest is calculated.

# 3 Methodology

The present study focuses on the MOSAiC expedition from the first to third leg, which ranged from 11 October 2019 to 16 May 2020, thus covering the main part of the transpolar drift though the central Arctic (Nicolaus et al., 2022; Shupe et al., 2022).

To obtain the relevant metrics used to identify the cloud properties that can be associated with effects induced by the presence or absence of sea ice leads, two Arctic observables need to be linked, namely the clouds and the sea ice. This section details the methods applied for this purpose.

The conceptual model proposed to identify the influence of sea ice leads on the cloud properties observed aloft RV *Polarstern*'s central observatory is depicted in Fig. 4 and described as follows:

- Leads that are spatially distributed within a 50 km radius of RV *Polarstern* are considered (Nicolaus et al., 2022).
- Leads release energy in the form of heat and moisture to the atmosphere (Andreas, 1980; Michaelis and Lüpkes, 2022).
- This release of energy can initiate a flux of water vapour along the wind direction or feed the horizontal water vapour transport (WVT) that is already present in the atmosphere.
- Given the proper wind direction, that WVT can move towards the RV *Polarstern* location.
- The WVT might favour the formation of new clouds or interact with clouds that already exist by changing their properties.
- These clouds are then observed at the RV *Polarstern* central observatory.

According to this concept, the water vapour transport is an important component which serves as a linking mechanism between sea ice leads and the clouds properties observed above the central observatory.

# 3.1 Cloud classification

The suite of remote sensing instruments, outlined in Sect. 2.1, observing the atmospheric state in the zenith-



**Figure 3.** (a) Lead fraction (LF) estimated from SAR Sentinel-1 divergence product within 50 km of RV *Polarstern* central observatory position; note the gap due to a lack of satellite overpasses at RV *Polarstern* latitude from 14 January 2019 to 13 March 2020; (b) Average fitted sea ice concentration (SIC) from MODIS–AMSR2 retrievals (blue). Shaded area corresponds to the SIC standard deviation of all pixels within 50 km.

Table 2. Cloudnet product specifications and reference for retrieval methods.

Product	Uncertainty	Input parameter (Instrument)	Reference
Liquid water content	15 %	LWP (MWR) Z <sub>e</sub> (cloud radar) <i>T</i> , <i>p</i> (radiosonde) β (lidar) (for liquid identification)	Frisch et al. (1998)
Ice water content	-30% to $50%$	Ze (cloud radar) T (radiosonde)	Hogan et al. (2006)
Liquid droplet effective radius	15 %	$Z_e$ cloud radar $\beta$ (lidar) (for liquid identification)	Frisch et al. (2002)
Ice crystal effective radius	42 % to 71 %	Z <sub>e</sub> (cloud radar) T (radiosonde)	Griesche et al. (2020), Hogan et al. (2006)



**Figure 4.** Conceptual model explaining how the presence of a sea ice lead can interact with the cloud observed above RV *Polarstern*. Water vapour transport (WVT) serves as conveyor belt for the latent and sensible heat released by the lead, which can influence the cloud over the measurement site. The thermodynamic and wind profiles are observed above RV *Polarstern* and considered constant up to the lead location. PBL stands for planetary boundary layer and IBL for intermediate boundary layer.

pointing direction comprises the main source to estimate relevant cloud properties. Several procedures and algorithms have been developed to perform atmospheric target classifications based on synergistic ground-based remote sensing observations (e.g. Shupe, 2007; Wang et al., 2020). Here we make use of the Cloudnet processing chain introduced by Illingworth et al. (2007). Cloudnet is an advanced classification algorithm specifically designed for the continuous evaluation of operational models using state-of-the-art groundbased instruments; Cloudnet provides not only the atmospheric target and cloud-phase classification but also an accurate prediction of the vertical and horizontal distribution of cloud microphysical properties like ice and liquid water content (IWC and LWC, respectively), which are analysed in Sect. 4. The main source of IWC uncertainty comes from the radar reflectivity factor  $Z_e$ ; to have an insight into the uncertainties, an estimation has been made of the relative error for the ice water path (IWP) when the reflectivity is changed



**Figure 5.** (a) Cloudnet classification for RV *Polarstern* observations on 18 November 2019. Isotherms and wind vectors are obtained from radiosonde profiles. Top colour-coded boxes indicate the type of target classification. (b) Profile of the bulk Richardson number. The selected  $Ri_c$  is visualized by a dashed black line as the PBLH. The dark-green line indicates the location of maximum  $\nabla$ WVT within the PBLH. Radiosonde launch times are indicated as vertical dash-dotted lines (RS). The wind direction profile is indicated by arrows.

by  $\pm 3\%$ ,  $\pm 10\%$ , and  $\pm 20\%$  of the original retrieved value. For the case shown in Fig. 5, it can be seen that by modifying the reflectivity by  $Z_e + 20\%$ , the IWP relative error is practically constant with a value of -12%, while for  $Z_e - 20\%$ , IWP is slightly lower than 13\% of the original retrieved value. For the cases of  $\pm 10\%$  and  $\pm 3\%$ , the retrieved IWP lies within a constant margin of  $\pm 6.5\%$  and  $\pm 1.9\%$ , respectively, regardless of the absolute value for IWP. Even though this confirms the sensitivity of IWP to the reflectivity factor, it is realistic to assume that the uncertainty in radar reflectivity is within the  $\pm 3\%$  of measured value, which gives a solid  $\pm 2\%$  of IWP uncertainty.

Basic requirements for the Cloudnet processing are having observations from a backscattering lidar, a microwave radiometer, and a Doppler cloud radar. Cloudnet is continuously being improved and developed by a large community of users under an open-source scheme (https://cloudnet.fmi. fi, last access: 30 March 2023) hosted by the Finnish Meteorological Institute and coordinated by the ACTRIS initiative (https://www.actris.eu, last access: 30 March 2023). Cloudnet has been selected as a classification algorithm for this study in its open-source version developed by the ACTRIS consortium (Tukiainen et al., 2020). Table 2 lists the input parameters and retrieval products obtained from Cloudnet, and Table 1 lists the instrumentation used with Cloudnet in the present work.

Figure 5a synthesizes an example of Cloudnet classification capabilities adapted to observations during MOSAiC, in addition to the target classification. The presented case study of 18 November 2019 shows that the stratiform lowlevel cloud present until 17:00 UTC mostly consists of a mixture of supercooled liquid droplets and ice particles, while the deep cloud system (present from 19:00 UTC onwards in Fig. 1) is mostly classified as ice-only.

A limitation of the Cloudnet classification and other synergistic retrievals happens in situations when the liquid cloud base is located below the first radar range gate, meaning the lidar signal is attenuated by low-level liquid clouds. This hampers the proper classification of liquid layers for which a lidar signal is required; hence all clouds are classified as pure ice clouds. This situation can be seen, for example, in Fig. 7b from 16:00 to 17:00 UTC and Fig. D1d (from 14:30 to 18:00 UTC). For the MOSAiC wintertime period, about 29 % of observed clouds were found to have bases below the first radar range gate; they were classified as low-level stratus according to the methodology developed by Griesche et al. (2020) based on information from the signal-to-noise ratio of the PollyXT 532 nm near-range channel.

#### 3.2 Atmospheric water vapour transport

The transport of air masses in the atmosphere is the main mechanism for the interaction of water vapour with the cloud characterized as in Sect. 3.1. One widely used concept to describe intense filament-like vapour transport in the atmosphere is the one of atmospheric rivers (ARs) (Martin-Ralph et al., 2020). Typically, either the vertically integrated water vapour (IWV) or integrated vapour transport (IVT) defined by Eq. (B1) in Appendix B is analysed to characterize ARs. Commonly ARs are identified whenever vapour flux exceeds a defined threshold relative to the zonal mean (Martin-Ralph et al., 2020; Zhu and Newell, 1998). One disadvantage of both AR characteristic variables is that IVT - being an integrated quantity – does not carry the information about the location of vapour transport in the vertical or whether one or multiple layers of vapour fluxes are present at different altitudes. The same is true for the IWV, with an additional limitation being that IWV is a wind-independent variable, and thus no transport is implicit with this metric.

For the present study, however, it is of primary interest to monitor the transport of water vapour in the lower layers of the atmosphere where the interaction with sea ice or open ocean is mostly taking place. That is why a detailed analysis of the vertical changes in vapour transport in the lower atmosphere becomes of paramount interest to locate where the most relevant flux is located. To do so, we derive the vertical gradient of WVT starting from the standard definition of IVT (Martin-Ralph et al., 2020) given by Eq. (B1), detailed in Appendix B. The vertical gradient of WVT ( $\nabla$ WVT) is calculated using radiosonde profiles of specific humidity  $q_v$  [gg<sup>-1</sup>], horizontal wind speed  $v_w$  [ms<sup>-1</sup>], and air pressure *P* [Pa], with all these at altitudes *z* [m], following Eq. (1), whose detailed derivation is described in Appendix B:

$$\nabla WVT = -\frac{10^2}{g} |q_v \cdot \boldsymbol{v}_w| \frac{\mathrm{d}P}{\mathrm{d}z},\tag{1}$$

where *g* is the constant of gravity and  $\nabla$  indicates the vertical gradient.

The advantage of using Eq. (1) is threefold: first, the altitudes at which local maximums of WVT occur can be identified separately from the flux  $|q_v \cdot v_w|$  profile; second, there is no need for thresholds to identify whether or not a layer of WVT is present (as is the case for IVT); and third, the derivative component  $\frac{dP}{dz}$  behaves as a weighting factor (inverse exponential with altitude) that naturally gives more weight to fluxes at the lowest layers and diminishes the upper ones where meso-scale ARs are more likely to be present. To constrain the relevant atmospheric layer even more, the  $\nabla$ WVT profile is analysed only below the planetary boundary layer height (PBLH; estimation is explained in the following subsection). This allows us to dismiss  $\nabla$ WVT peaks that are less likely to have interacted with sea ice in the vicinity of RV *Polarstern*.

#### 3.2.1 Estimation of the planetary boundary layer height

The Richardson number is defined as the ratio of turbulence associated with buoyancy to that associated with mechanical shear. This ratio resolved at increasing altitudes above surface level is known as the gradient Richardson number (Ri), and it is widely used to estimate the planetary boundary layer (PBL). However, when the atmospheric turbulence profile cannot be resolved with measurements at spatial and temporal resolution that is sufficiently high to resolve smallscale turbulence – as in the case with sounding the atmosphere – it is more convenient to use the bulk Richardson number ( $Ri_b$ ) as a good indicator of the stability conditions in the atmosphere. The bulk Richardson number is defined as in Eq. (2):

$$Ri_{\rm b}(z) = \frac{g}{\theta_{\rm v}} \frac{\Delta \theta_{\rm v} \Delta z}{(\Delta u)^2 + (\Delta v)^2},\tag{2}$$

where g is the constant of gravity acceleration 9.81 [m s<sup>-2</sup>];  $\theta_v$  is the virtual potential temperature profile [K],  $\Delta \theta_v = \theta_v - \theta_v(z_0)$ ; and  $\Delta u = u - u_0$  and  $\Delta v = v - v_0$  are the horizontal wind components [m s<sup>-1</sup>].  $\Delta z = z - z_0$ , with z being the altitude of the atmosphere layers [m] and the subscript 0 indicating the surface reference.

The atmospheric stability is characterized by a range of values of  $Ri_b$ , with  $Ri_b < 0$  indicating an unstable and turbulent atmosphere,  $Ri_b < Ri_c$  a neutral atmosphere, and  $Ri_b \ge$  $Ri_{c}$  a stable with almost all turbulence diminished; here  $Ri_{c}$ is the critical Richardson number. To estimate the PBLH by means of  $Ri_{\rm b}$ , the standard procedure relies on applying  $Ri_{\rm c}$ as a threshold that defines the layer above which the atmosphere is considered to be non-turbulent and laminar. The most common  $Ri_c$  value used is 0.25 although there is a wide range of values that can be found in the literature, spreading from 0.1 to 1. For this study  $\theta_v(z_0)$  and  $Ri_b$  have been calculated using the radiosonde data in combination with the ground infrared thermometer  $T_{gnd}$  (GND IRT) as surface skin temperature. For the purpose of this study, an exact estimation for the PBLH is not crucial, but rather a rough approximation is useful as an atmospheric layer top to be considered for the determination of the altitude at which VWVT reaches a local maximum. The top of the PBL is then considered to be located at the altitude when the condition  $Ri_{\rm b} > 1$  is first met, with 1 being a rather conservative critical value to cover most of the relevant mixing layers for the Arctic winter atmosphere.

In Fig. 5b the atmospheric stability based on the bulk Richardson number given by Eq. (2) and calculated from the ARM INTERPSONDE product is depicted for the 18 November 2019 case study. The atmospheric stability is colour-coded by light- to dark-blue colours ( $0 < Ri_b < Ri_c$ ) for a statically stable to neutral atmosphere. Above the critical value  $Ri_c = 1$  (light grey), the atmosphere is not considered to be significantly turbulent anymore, therefore the cloud within or above this level does not have the potential to mix with the layers below. Unstable atmospheric conditions are highlighted by the yellow to red colours corresponding to sub-zero Rib values, which, for the case of 18 November 2019, occur once the wind direction shifts to northerly and northwesterly directions and the deep cloud system is observed above RV Polarstern. For this case, the maximum of  $\nabla$ WVT within the PBLH varies between 0.1–0.4 km, which is depicted by the solid green line in Fig. 5b.

In the following section, it is explained how  $\nabla WVT$  is exploited as the mechanism responsible for the interaction between sea ice or open ocean and the cloud observations above RV *Polarstern*.

#### 3.3 Cloud coupling

#### 3.3.1 Cloud mixing layer

The cloud-driven mixing layer below the cloud base is determined by calculations based on the degree of variability in the virtual potential temperature  $\theta_v$  profile (Eq. A1). A quasiconstant  $\theta_v$  profile below cloud-base height (CBH) implies a well-mixed layer. A departure from quasi-constant  $\theta_v$  indicates a thermodynamic inversion and thus decoupling from the layer beneath.



**Figure 6.** Examples for  $\nabla$ WVT profile (black axes) coupling (**a**, **b**) and decoupling (**c**) to the cloud for 18 November 2019. The liquid cloud layer is shown as a grey-shaded area. The coupling status is determined by the position of the maximum  $\nabla$ WVT (dashed green) relative to the decoupling height CMLH (dashed red): (**a**) the max  $\nabla$ WVT is above CMLH and below the cloud base but is still considered coupled to the cloud and (**b**) max  $\nabla$ WVT is inside the cloud. Panel (**c**) shows a case where the cloud is decoupled. For reference  $\theta_v$  is indicated by the solid line, and the *x* axis is in blue.

To estimate the cloud mixing layer height (CMLH) the relative variability in  $\theta_v$  starting at the cloud base ad going downwards is analysed. This concept has been extensively utilized by Sotiropoulou et al. (2014) and Gierens et al. (2020) for classification of surface-coupled clouds. The criterion used in this study consists of calculating the cumulative variance of  $\theta_v(i)$  ( $\sigma_{\Sigma}^2(z)$  defined by Eq. A2) starting from the cloud base and going towards the surface level or i = 0. Thus CMLH is equal to the *z* value at which the criterion  $\sigma_{\Sigma}^2(z) \ge 0.01 \text{ K}^2$  is first met. The same procedure is applied to estimate the cloud-driven mixing layer above the cloud top. The CMLHs below and above the cloud base and -top are quantities used to estimate the coupling or decoupling state of the cloud with the water vapour transport, as described in the following section.

# 3.3.2 Cloud coupling classification

The observations are sorted into two classes depending on the likelihood of interaction between the sea ice situation upwind and the cloud observed aloft, and they are linked by the water vapour transport as a conveying mechanism for moisture and sensible heat to the clouds above the central observatory, i.e. whether or not the WVT is coupled or decoupled to the cloud. A cloud observation is considered to be coupled to WVT when the location of maximum  $\nabla$ WVT is found to meet one of the following criteria: be in the cloud, or between the cloud's CMLH below and above the cloud base and top, respectively. Conversely it is considered decoupled when the maximum of  $\nabla WVT$  happens to be either above the cloud top's CMLH or below the cloud base's CMLH. Those cases are illustrated in Fig. 6 for coupled (Fig. 6a and b) and decoupled (Fig. 6c) situations. It is important to note that in the present study we are departing from the canonical concept of surface-cloud coupling generally found in the literature. This is due to the fact that the location of sea ice lead occurrence is not strictly co-located with the position of RV Polarstern (see for instance Figs. 2 and D2); thus the sea ice–cloud interaction is not expected to take place vertically within a static column but rather to be dynamically ascribed by the air mass movement from afar. Moreover, the persistent presence of a surface temperature inversion in the Arctic makes the case of a vertically static column cloud–surface coupling a rare event.

Hence our definition of the sea ice–cloud coupling status is governed by the following criteria:

- I. *Coupled.* This is when the maximum of  $\nabla WVT$  is localized within the cloud-driven mixed height above and below the cloud top and base, respectively, in Fig. 6a and b.
- II. *Decoupled.* This is when the maximum of  $\nabla WVT$  is found to be outside the cloud layer limited by the top or bottom CMLH. as shown in Fig. 6c.

According to the above, the classical definition of cloud– surface coupling is only a special case of this more general approach when the CMLH below the cloud base reaches the surface level; i.e. z = 0 in Eq. (A2). We found, however, that during the MOSAiC wintertime, this situation only comprises 4.7 % of all cloudy observations and 7.3 % of all cases that fulfil criterion (I), "coupled".

# 3.4 Sea ice concentration in the direction of WVT

Information about the state of the sea ice is considered within a circular area of 50 km radius centred on the RV *Polarstern*. This particular radius has been chosen as a compromise to cover the sea ice conditions representative of the observations at the central observatory based on SIC comparisons of circles with 6, 50, and 100 km radius (see Krumpen et al., 2021).

Within the 50 km circular area, sea ice conditions relevant for the interaction with the cloud observations are extracted



**Figure 7.** From top to bottom: (a) median lead fraction time series (black line, left axis) with the interquartile region (shaded region) and sea ice concentration (light blue, right axis) after considering only grid cells within a conical sector centred on RV *Polarstern*. The direction of the conical sector as a function of the wind direction is shown by the blue arrows and values on top. The horizontal dashed red line indicates an average LF value for totally covered sea ice. (b) The same as Fig. 5 but magnified to the lower 2 km where the post-processing cloud edge detection is highlighted with black lines and the lidar cloud base is shown in grey dots. The height of maximum  $\nabla WVT$  is shown in dark green, and the PBLH is in light green (dashed). (c) IWP and LWP within the detected cloud layer (b). The coupled–decoupled status flag is highlighted in yellow and magenta, respectively.

from a conical sector centred on RV Polarstern and extended up to a 50 km radius and angular span of 5°. The azimuth angle of this conical sector is adjusted every minute based on the wind direction measured at the altitude of maximum  $\nabla$ WVT (green lines in Figs. 6 and 5b). For instance, for the sea ice situation on 18 November 2019 only the LF and SIC highlighted within the grey lines in Fig. 2 are associated with the zenith-pointing cloud observations. To ensure that the considered wind direction is still representative within the 50 km range, back-trajectory analysis was performed using the Lagrangian back-trajectory tool LAGRANTO (Sprenger and Wernli, 2015). The trajectory of WVT was tracked backwards from the altitude where maximum  $\nabla WVT$  occurs; it was found that the back trajectories show a considerable agreement with the assumed wind direction within the 50 km radius (see Supplement).

Although RV *Polarstern*'s ice floe drifted (i.e. the geographical position for the centre of the 50 km circular area had an average drifting speed of  $8.52 \text{ km d}^{-1}$ ; Krumpen et al., 2021; Nicolaus et al., 2022), we update the azimuth angle of the sea ice conical sectors every minute in synchrony to the available vertical wind profiles given by ARM's INTERP-SONDE product. Since LF and SIC information is only available on a daily basis, the centre of the 50 km circular area also needs to be updated accordingly to avoid abrupt changes in the relative position of the leads with respect to RV *Polarstern*.

#### 4 Results

# 4.1 Case study of 18 November 2019

Besides the Cloudnet retrieval products summarized in Table 2, further macro-physical and thermodynamical properties of the cloud are estimated from the Cloudnet target classification and from radiosonde. These properties include cloud-base height and cloud-top height (CBH and CTH, respectively) as well as the temperature at the cloud base and top. Figure 7 summarizes all those properties as well as the coupled–decoupled status of every observation according to Sect. 3.3 for the case study of 18 November 2019. Results presented in Fig. 7c are based on vertical integrals from single-layer CBH to CTH of Cloudnet-determined LWC and IWC (see Table 2). LWP is determined for liquid-only clouds and MPCs, whereas IWP is determined for MPCs and pure ice clouds and includes falling solid precipitation (snowfall).

The case study in Fig. 7a shows the corresponding 1 min resolution time series for the sea ice statistics of all grid cells located within the conical sector aligned with the wind direction (see grey cone in Fig. 2) as described in Sect. 3.4. The conical sector is adjusted as a function of the wind direction (blue arrows in the top panel of Fig. 7) given at the height of maximum  $\nabla$ WVT occurrence (Fig. 7b, solid green line).

From 00:00 to 16:00 UTC on 18 November 2019, latent heat and sensible heat were advected towards RV *Polarstern* from north-northeasterly directions where a sea ice lead had formed (Fig. 2, right panel). LF within the conical sector has median values ranging from 0.1–0.2 and with an interquartile region (IQR) of up to 0.4 shown in Fig. 7 (top panel). During this time period of high LF, a stratiform low-level MPC was observed (Figs. 1, 5a, and 7b).

At approximately 16:00 UTC, the wind direction changed towards the northwest to west, where no leads were located. This cut-off of the heat and moisture supply led to dissipation of the low-level MPCs. After about 17:00 UTC, a deep cloud system related to a storm was observed above RV *Polarstern*. This storm was associated with sublimation just above the maximum  $\nabla$ WVT and also led to an increase in turbulence in the lowest atmospheric layers (see Fig. 5b, red colours). SIC shows a slight decrease from values near 100 % to about 98 % towards the end of the case study period. The coupling status is highlighted by the yellow (coupled) and magenta (decoupled) flag at the bottom of Fig. 7c.



**Figure 8.** Results for 18 November 2019: single-cloud-layer relationships between LF and micro- and macro-physical cloud properties. Observations are represented as averages (circles and triangles) with standard deviation (bars) within intervals of 15 min; coupled and decoupled cases are marked as blue circles and orange triangles, respectively. (a) Mean single-cloud-layer LWP vs. LF (black line in Fig. 7, top panel). (b) The same but for IWP of the same cloud layer. (c) Cloud layer average effective radius for liquid droplets. (d)  $\Gamma_{cloud}$  as defined in Eq. (4) vs. LF; the horizontal dashed line represents a moist adiabatic  $\Gamma_m$ . In all panels the best fit is shown as a dark-blue line and the corresponding coefficient of determination  $r^2$  is based only on coupled data points.

From Fig. 8a, a relationship between LWP and LF can be seen; i.e. the larger the LF, the higher the LWP. For IWP (Fig. 8b), a more scattered relationship is found, with a wide range of IWP values occurring independently of the magnitude of the observed LF. The only clear feature is the clustering of larger IWP values at low LF which correspond to the decoupled profiles of the deep cloud present after 17:00 UTC. Note that between 16:00 and 17:30 UTC (in Fig. 7b) the lidar detects a liquid layer below the lowest available Cloudnet classification height, meaning Cloudnet could not relate this period with the occurrence of liquid droplets but instead misclassifies it as ice cloud only. This is reflected in the total water path calculated within the lowest and top cloud limits as indicated in Fig. 7c.

Figure 8c depicts the liquid effective radius retrieved by Cloudnet and averaged over the cloud layer defined by

$$\overline{r}_{\rm eff} = \frac{\int_{\rm clb}^{\rm cth} N(z) r_{\rm e}(z) dz}{\int_{\rm clb}^{\rm cth} N(z) dz},\tag{3}$$

where N(z) is the droplet number concentration and  $r_e(z)$  is the effective radius corresponding to the altitude z within CBH and CTH. The best-fit curve indicates a slightly positive correlation as LF increases.

Shown in Fig. 8d is the in-cloud temperature lapse rate defined as follows:

$$\Gamma_{\text{cloud}} = -\frac{\mathrm{d}T}{\mathrm{d}h} = -\left(\frac{T_{\text{top}} - T_{\text{base}}}{\mathrm{CTH} - \mathrm{CBH}}\right). \tag{4}$$

Figure 8d indicates that  $\Gamma_{cloud}$  is often close to the moist adiabatic lapse rate of 6.0 °C km<sup>-1</sup> (dashed horizontal line in Fig. 8d), especially the decoupled cases (orange triangles) independent of LF. The negative  $\Gamma_{cloud}$  values represent cases with a temperature increase within the cloud layer or inversion at the cloud top, and they are mostly related to coupled clouds (blue circles).

The 18 November 2019 case study encompasses a situation where the observed clouds have a well-defined correlation with LF situation upwind, mainly due to the occurrence of a single cloud layer. This is not always the case, as can be seen in Fig. D1, where the cloud properties correlated to LF are more subtle. To assess the robustness of the case study results over a wide range of cases, a statistics analysis is performed based on the same methodology applied to the whole wintertime MOSAiC expedition.

Description Number of avail		er of available obse	able observations	
	Total	Coupled [%]	Decoupled [%]	
During wintertime	259 200			
With cloudy sky	199 926	121 970 [61 %]	77 955 [39 %]	
Cloudy and LF available	124 787			
$LF \le 0.02$	104 025	66 432 [64 %]	37 593 [36 %]	
LF > 0.02	20762	13 081 [63 %]	7681 [37 %]	
Cloudy and SIC available	199 333			
SIC > 98 %	108 979	69 332 [64 %]	39 647 [36 %]	
$SIC \le 98 \%$	15 808	10 181 [64 %]	5627 [36%]	
Liquid clouds detected below first radar range gate	68 998 [29 %]			

**Table 3.** Number of available observations at 1 min resolution during the MOSAiC wintertime for the statistical analysis. The total of observations includes cloudless and cloudy cases. For cloudy situations, a distinction is made between WVT coupled and decoupled. The frequency of occurrences of different intervals of cloud water path sorted by the coupling status is presented in Appendix D2, Fig. D3a.

# 4.2 Statistical analysis

The methodology introduced for the case study in Sect. 4.1 was applied to the whole wintertime MOSAiC period (November 2019-April 2020). Table 3 summarizes the obtained dataset that was statistically analysed after splitting between cases with LF less than or equal to 0.02 (LF  $\leq 0.02$ ) and LF greater than 0.02 (LF > 0.02). In that way we try to isolate cases in which sea ice leads have most likely been interacting with the observed clouds. The motivation to separate the coupling state when  $LF \le 0.02$  is to have an insight into situations where WVT is present and leads can be located at ranges further than 50 km. In the following analysis when the probability distribution function (PDF) of a certain cloud property shows a similar shape for coupled cases with LF < 0.02 and LF > 0.02, it might imply that leads located further away produce similar PDFs, e.g. having the same PDF maximum location but being less frequent. On the contrary, when the PDFs are different for cases of coupled with LF > 0.02 and decoupled (e.g. multiple peaks versus monomodal distributions), it is an indication that the leads-WVTcloud coupling system is separating the observations into two distinguishable distributions. This section presents the relevant statistical differences in relation to the cloud microand macro-physics and thermodynamic properties between clouds classified as coupled or decoupled to the WVT.

Figure 9 depicts the PDFs of different macro- and microphysical cloud properties. The data are separated into four groups: WVT coupled cases (blue), decoupled cases (orange), cases with LF  $\leq$  0.02 (dashed lines), and cases with LF > 0.02 (solid lines). In Fig. 9a the PDFs of the liquid layer base height presented for coupled cases are generally comprised of low-level clouds with a probability of occurrence significantly enhanced for the subset corresponding to LF > 0.02, with a main peak at 250 m. In contrast, decoupled clouds do not have a pronounced peak in liquid layer base height occurrence but are more homogeneously distributed over a range of a few hundred metres to 1 km. Furthermore, Fig. 9b exposes the fact that coupled clouds also tend to be thicker, with a peak in the PDF at around 400 m, whereas decoupled clouds are equally likely to have thicknesses ranging between 40 and 500 m. Statistics of cloud-top temperature are shown in Fig. 9c, where two distinct features appear: clouds related to  $LF \le 0.02$  have a maximum probability of cloud-top temperature at around -22 °C regardless of their coupling state. Conversely, for LF > 0.02, coupled clouds are generally warmer with their maximum PDF of cloud-top temperature at about -12 °C and a second minor PDF peak at -29 °C. For decoupled clouds and LF > 0.02, the cloud-top temperature PDF spreads out to colder temperatures with a primary peak at -22 °C and a second peak at -36 °C. Moreover cloud-top temperatures below -40 °C are considerably more frequently observed for decoupled than coupled clouds.

In Fig. 9d the PDF of the cloud layer mean liquid droplet effective radius  $\bar{r}_{eff}$  of coupled cases peaks at 12 µm for both LF classes. The PDF of  $\bar{r}_{eff}$  for the WVT decoupled cases exhibits a bimodal distribution peaking at 8 and 17 µm irrespective of LF. Likewise, the average effective radius for ice particles (Fig. 9e) exposes a bimodal PDF of ice  $\bar{r}_{eff}$  with peaks at 42 and 49 µm almost equally likely to occur for coupled cases, while the PDF for decoupled cases has one minor peak at 32 µm and a major peak at 48 µm. The PDFs for LF  $\leq 0.02$  (dashed lines) have single maximums at 48 and 32 µm for coupled and decoupled cases, respectively.

To complement the thermodynamic features of the cloud, Fig. 9f shows the PDF of the cloud layer temperature lapse rate. The main feature found is that for LF  $\leq$  0.02, the decoupled PDF indicates a maximum at the nominal value for the moist adiabatic lapse rate  $\Gamma_m$ , while decoupled clouds at LF > 0.02 show a slightly lower lapse rate.



**Figure 9.** Probability distribution functions for all observations from November 2019 to April 2020. Data are sorted by  $LF \le 0.02$  (dashed lines), LF > 0.02 (solid lines), coupled ("co.", blue), and decoupled ("de.", orange) cases of cloud properties. (a) For liquid layer base height, separated for coupled (blue lines) and decoupled (orange lines) cases. Dashed lines correspond to  $LF \le 0.02$  and solid lines to LF > 0.02. (b) The same but for the cloud depth. (c) PDF for cloud-top temperature. (d) PDF for cloud layer mean effective radius of liquid droplets. (e) Cloud layer mean effective radius of ice crystals. (f) In-cloud temperature lapse rate, with the dashed black line indicating a nominal moist adiabatic lapse rate.

The coupled  $\Gamma_{cloud}$  PDFs are biased and skewed towards negative lapse rates with the most probable values found to be 0 and  $-2 \,^{\circ}C \, km^{-1}$  for LF  $\leq 0.02$  and LF > 0.02, respectively. The latter also has a minor peak in the PDF at around  $-5 \,^{\circ}C \, km^{-1}$ . The dominant feature found in Fig. 9f is that the  $\Gamma_{cloud}$  PDFs of clouds coupled to the WVT mechanism are displaced towards lower or even negative  $\Gamma_{cloud}$  values (i.e. temperature inversion above the cloud base), and this characteristic is enhanced when LF > 0.02.

# 4.2.1 Fraction of ice water content in cloud

Hereafter we define the fraction of ice water content in the clouds relative to the total condensed water in the cloud (ice and liquid) as follows:

$$\chi_{\rm ice} = \frac{\rm IWP}{\rm IWP + \rm LWP}.$$
(5)

The definition given in Eq. (5) is in line with the Korolev and Milbrandt (2022) phase composition of clouds but differs from most of the studies based on space-borne Arctic observation (Coopman et al., 2018) where  $\chi_{ice}$  is defined as the number of grid cells considered to be ice divided by the number of grid cells considered to be either liquid or ice.

Similarly, the definition in Eq. (5) differs from other groundbased and ship-based observations over mid-latitudes and the Arctic, where the fraction of ice-containing clouds with respect to all observed clouds is considered (Kanitz et al., 2011; Westbrook and Illingworth, 2011; Griesche et al., 2021) regardless of the water content in those clouds. Therefore the results based on ice water fraction analysis presented here cannot be directly related to the previously mentioned work. We are mainly interested in the features controlling the cloud microphysical properties such as LWP and IWP, since those are the dominant drivers of the cloud-surface interaction. Furthermore, note that in the following analysis, Eq. (5) is only applied to the single cloud layer relevant for the classification of coupled or decoupled to the WVT; therefore the results are not representative of the whole atmosphere in the case of multi-layer cloud situations.

Figure 10 depicts a clear difference in the ice water fraction when separated by the cloud coupling to the WVT status (blue for coupled and orange for decoupled) between -15 and -25 °C for cases with LF  $\leq 0.02$  (a) and between -12 and -30 °C for cases with LF > 0.02 (b). For the situation in (a) the ice water fraction, for coupled and decoupled cases, increases until a local maximum corresponding



**Figure 10.** Top panels: histograms of the number of occurrences for liquid (bars) and ice (lines) water paths for coupled ("co", blue) and decoupled ("de", orange) cases. For visualization purposes, the decoupled number of occurrences has been scaled by 2 (2×). Bottom panels: ice water fraction, as defined by Eq. (5), versus cloud-top temperature for (a) all cases and (b) cases where LF > 0.02. Coupled (decoupled) cases are depicted as blue circles (orange triangles); shaded areas represent 1 standard deviation of the ice water fraction sorted within 1 °C temperature bins. The best fit to the coupled data is shown as a dashed blue line.

to a cloud-top temperature of -15°C is reached, as indicated by the vertical dashed light-blue line. This can be explained by the fact that at approximately -15 °C the maximum ice growth takes place due to the largest difference between saturation water vapour pressure over ice and water (Rogers and Yau, 1991). Below -15 °C cloud-top temperature, the coupled and decoupled cases depart significantly until approximately -25 °C, with the coupled cases showing a steady increase in ice water fraction, presumably due to the intake of humidity provided by the coupling with the WVT and thus fostering the formation of ice particles, whereas the decoupled case indicates a drop of ice water content up to  $\chi_{ice} = 25 \%$  at a cloud-top temperature of  $-20 \degree C$ , whereafter the heterogeneous freezing process continues. Both  $\chi_{ice}$ curves reach 50 % at about the same temperature of -22and -24 °C for the decoupled and coupled cases, respectively. For  $\chi_{ice} = 75$  % and higher, the coupled and decoupled  $\chi_{ice}$  curves behave similarly towards homogeneous freezing, which has been found to occur within the range -37 to -40 °C (Rogers and Yau, 1991; Pruppacher and Klett, 1997).

For the subset of data with LF > 0.02 (Fig. 10b), the  $\chi_{ice}$  coupled and decoupled cases reach a first local maximum at a colder temperature of -18 °C. Note, however, that the number of cases for LF > 0.02 are much smaller by a factor of ~6 (Table 3). Furthermore this local maximum at -18 °C coincides with  $\chi_{ice} = 50$  %, which means it reaches the 50 % ice water fraction at a warmer temperature as compared to Fig. 10a. Moreover, it can also be seen that  $\chi_{ice}$  for LF > 0.02 follows more closely the empirical model given as  $\chi_{ice}(T) = 0.5[1 + \tanh(-\beta_0(T - \beta_1))]$  by Coopman et al. (2018) based on satellite-based observations for the Arctic between 2005 and 2010 during March to September. A slight

modification to the empirical model for  $\chi_{ice}$  was made in this study so that  $\beta_1$  fits the temperature corresponding to  $\chi_{ice} = 50 \%$ . The decoupled curve shows a steep drop in the ice fraction to a minimum of about 15 % at the temperature of -20 °C (a similar situation to that observed in Fig. 10a), followed by an abrupt increase in  $\chi_{ice}$  to 95 %. The best-fit curve for  $\chi_{ice}$  values of coupled cases has a more monotonic increase but with a less steep slope than the empirical model (dashed blue curve).

# 4.2.2 Liquid and ice water content as a function of sea ice

With the aim to confirm the relationship between cloud properties with the sea ice lead fraction found in Sect. 4.1, the whole dataset is analysed and summarized in Fig. 11. The number of occurrences for the whole dataset of observations containing either ice or MPCs is shown colour-coded in the background (corresponding to the total cloudy sky in Table 3). Overlaid on every panel are the data for LF > 0.02 (see Table 3 for references to the number of observations for LF > 0.02) depicted as binned averages for water vapour transport coupled (blue circles) and decoupled (orange triangles), with the corresponding standard deviation indicated by the bars. The best-fit curve to the coupled data (blue circles) is indicated by the black curves, and the coefficient of determination  $r^2$  is given.

The positive correlation between LWP and LF is evident from Fig. 11a, with the latter being responsible for 63 % of the variability observed in LWP as indicated by  $r^2$ . Although there is an apparent reduction in LWP at LF between 0.02 and 0.08, this can be due to the circumstantial lack of LWP observations around an LF of 0.1. However, the fit is robust for



**Figure 11.** Data for the period November 2019–April 2020: top row (**a**, **b**), distribution of LWP as a function of observed LF (**a**) and SIC (**b**); bottom row (**c**, **d**), distribution of IWP as a function of LF (**c**) and SIC (**d**). The symbols are the average of the observations within a fixed LF bin width, while the bars indicate 63 % variability within the LF bin. The colour scale indicates the number of observations within the bins for the whole dataset, whereas the symbols represent only the data corresponding to LF > 0.02. The coupling status is indicated by orange triangles (decoupled) and blue circles (coupled). The black curves represent the best fit for only the coupled blue circles, and the coefficient of determination is given by  $r^2$ .

data with an LF of 0.1 and higher. When comparing LWP to sea ice concentration, the positive relation is certainly weak or arguable nonexistent since the analysis shows that SIC can only explain 3 % of variability in LWP values (Fig. 11b). This result is strongly influenced by the LWP values that are less than  $40 \text{ g m}^{-2}$  paired with SIC between 90 % and 80 %. When excluding these data points, the correlation of LWP and SIC is enhanced. It is important to remark that due to the different sensors and retrieval methods, SIC and LF are not equivalent or interchangeable, meaning that higher LF observations do not necessarily correspond to low SIC, as demonstrated in Fig. 3. Therefore even after SIC is adjusted to reduce retrieval underestimations during April (see Appendix C), SIC can still contain ill-posed SIC retrievals that can attribute LWP values to uncertain SIC, e.g. a range of LWP values mapped to low SIC.

Regarding the IWP sensitivity to changes in LF and SIC, Fig. 11, bottom row, indicates a moderate positive correlation with LF ( $r^2$  of 0.25 for IWP vs. LF). On the contrary, when IWP is related to SIC, the relation is opposite, as with the case of IWP versus LF, mainly for the region of SIC between 80% and 96%, with only a moderate increase in IWP when SIC change from 100% to 97%. It is important to note that most of the observed high values of IWP are related to deep precipitating cloud systems.

To highlight the effect introduced by deep convective clouds in the statistical analysis, Fig. 11 is reproduced (see Saavedra Garfias, 2023) by only considering cloud-top heights below 2.5 km. This reveals that the robust positive relationship between LWP and LF originates mainly from the low-level clouds (with  $r^2 = 0.58$  for the case of 2.5 km versus  $r^2 = 0.63$  for all cloud heights in Fig. 11). Regarding IWP, the coefficient of determination with respect to LF is reduced to  $r^2 = 0.03$  (for the cases with 2.5 km cloud-top height) as compared to  $r^2 = 0.25$  in Fig. 11c. Moreover, it can be seen that when cloud-top height is constrained to 2.5 km, the clouds present values for IWP below  $100 \,\mathrm{gm^{-2}}$ , with the coupled cases having systematically larger IWC than the decoupled cases. Therefore, it is concluded that the main sources of larger ice water content are deep precipitating systems rather than stratiform low-level clouds.

# 5 Discussion and conclusions

Based on the methodology developed in Sect. 3 applied to the case study of 18 November 2019 and to the complete MO-SAiC wintertime observations (November 2019–April 2020) for statistical analysis, the following can be concluded:

The WVT coupled cloud observations outnumber the decoupled cases by at least a factor of 1.6 (61% vs. 39% from all cloudy cases in Table 3). However, this

factor can range between 10 and 100 for a given total water path (Fig. D3). With LWP, results are found to be mostly below  $150 \text{ gm}^{-2}$ , whereas for IWP most of the data points are above  $100 \text{ gm}^{-2}$ .

- When LF > 0.02, coupled clouds are statistically lower and thicker in height and depth, respectively.
- Coupled clouds have significantly warmer cloud-top temperatures, with a vast majority of clouds having a temperature inversion at the cloud top, thus implying a stratified stable cloud layer.
- The cloud microphysical properties such as droplet effective radii distribution do not show any clear indication of relevant differences between LF  $\leq 0.02$  and LF > 0.02 for WVT coupled clouds, with only a slight increase in the probability of occurrences of  $\bar{r}_{\text{eff}}$  between 10 and 25 µm for decoupled cases. The effective radius for ice does not show a clear dependence on LF for coupled cases. For decoupled cases the  $\bar{r}_{\text{eff}}$  values are larger for LF > 0.02 than for LF < 0.02. This suggests that the ice  $\bar{r}_{\text{eff}}$  of decoupled clouds seems to be more sensitive to LF, which is counter-intuitive.
- The distribution of LWP and IWP as a function of the sea ice lead fraction and sea ice concentration reveals that only LWP and LF are strongly related ( $r^2 = 0.63$ ). For the case of LWP versus SIC, the correlation is weakly negative ( $r \sim -0.17$ ), with SIC only explaining 3% of the LWP variability. It is important to note that SIC < 90 % was mainly observed during April 2020 (Fig. C1); this MOSAiC period is being extensively studied due to the occurrence of warm air intrusions into the central Arctic that are conducive to inaccuracies in AMSR2-based retrievals (Krumpen et al., 2021; Rückert et al., 2023). Given that in April there was not 8 % open water reported around RV Polarstern, the bias correction to the AMSR2-MODIS merged product (presented in Appendix C) still has inaccuracies, leading to the attribution of LWP at lower SIC values and hence affecting the correlation between LWP and SIC.
- The occurrence of LF > 0.02 is correlated with increases in IWP. This is however contradictory to the result found for IWP versus SIC, where decreasing SIC is strongly correlated with a reduction in IWP. Important to note, however, is that when cases with cloud depth larger than 3 km are excluded (not shown), the relation between LWP and LF/SIC does not differ from the results in Fig. 11. The main difference is that for these situations no significant relation between IWP and LF ( $r^2 = 0.0$ ) is found. For IWP versus SIC, when only cloud depths below 3 km are considered, the pattern remains similar ( $r^2 = 0.33$ ) when all cloud depths are considered. This is an indicator that sea ice leads have

no trivial effect on IWP but are only strongly correlated to LWP.

- The ice fraction of total water content  $\chi_{ice}$  depicts major differences when the entire dataset is compared to cases with LF > 0.02. Mainly in the region of heterogeneous ice formation between -10 and -32 °C, the observations with  $LF \le 0.02$  have a pronounced peak at about -15 °C (temperature where maximum growth of ice crystals takes place; Rogers and Yau, 1991). For the subset of data with LF > 0.02, the maximum  $\chi_{ice}$ peaks are displaced to a slightly colder temperature  $(-17 \,^{\circ}\text{C})$ .  $\chi_{\text{ice}} = 50 \,\%$  is reached at warmer temperatures  $(-18 \degree C)$  for LF > 0.02 as compared to the entire dataset (-22 °C). Both cases starkly contrast results reported by Westbrook and Illingworth (2011) for midlatitude observation where  $\chi_{ice} = 50 \%$  was reported to happen at -27 °C, based, however, on cloud observations only for temperatures below  $-10 \,^{\circ}$ C.

The findings presented here can be used as valuable constraints to evaluate cloud microphysical parameterizations for the Arctic system. Since sea ice leads are not explicitly resolved in such models, lead-averaged surface heat flux, as well as its influence on clouds, is of considerable interest for the parameterization of energy exchange (Gryschka et al., 2023). The different features of ice water fraction  $\chi_{ice}$ , as a function of cloud-top temperature, found for coupled and decoupled cloud cases are a result that deserves to be deeply investigated by validating it not only with long-term observations but also by a better understanding of the modelling of cloud microphysics that can lead to explaining the finding.

The results found in this study are presented using the lead fraction as a constraint to distinguish effects on cloud properties. Based on SIC data from the MODIS-AMSR2 merged products, a spatial resolution of 1 km is sufficient to detect large leads. However, this resolution and merged retrieval product are not sufficient to resolve most small leads. Thus the novel product of the lead fraction (LF) with a spatial resolution of 700 m based on Sentinel-1 SAR satellite divergence data is of utter importance to determine the influence of sea ice leads over cloud properties and because of its ability to only detect leads when sea ice opens, avoiding therefore the consideration of newly frozen leads, which have been argued to serve as a dissipation mechanism for low-level clouds (Li et al., 2020b). Regarding aerosols as key components in cloud processes, although no direct or remote sensing measurements of advected aerosols along the WVT path were available during MOSAiC, aerosols and ice-nucleating particles (INPs) were sampled at the RV Polarstern location (Creamean et al., 2022). INP concentrations are found to be persistent among the months from October to April mainly at temperatures ranging from -25 to -15 °C, with large INP sampling during periods with high lead occurrence and wind speeds above  $5 \text{ m s}^{-1}$ . Therefore, as highlighted by Creamean et al. (2022), the high fractional occurrence of ice

in clouds below 3 km in winter implies that observed small INPs could serve an important role in cloud ice formation. Since the surface is predominantly frozen, the local source of INPs is locally limited; thus it is plausible to support the hypothesis that leads play an important role as sources of sea spray in windy conditions during the wintertime. It is feasible that the sea ice leads as sources of INPs, like sea spray, can be advected along the WVT and therefore be included in our analysis as part of the coupled–decoupled classification. However, since no continuous INP sampling has been performed, our dataset cannot be separated based on INP concentration, but such a type of analysis is an important source of information to narrow down the leads' effects on cloud properties.

The MOSAiC observations of LWP and IWP for cloud layers coupled to WVT have been analysed as a function of LF and depict a clear positive relation between LWP and IWP with LF for coupled cases. When compared to SIC, LWP has a less pronounced positive relation and IWP even exhibits a negative correlation. When cases with cloud-top heights below 2.5 km are considered, the strong relationship between LWP and LF is preserved, which points to the conclusion that sea ice leads have the dominant signal in LWP. On the contrary, the positive relationship between IWP and LF practically disappears when cases with cloud-top height below 2.5 km are considered, with IWP values reduced to below  $100 \,\mathrm{g}\,\mathrm{m}^{-1}$ . Based on these findings, the interpretation is that the effect of sea ice leads on low-level MPCs is that the former favour an increase in liquid water while tending to keep ice water steady. The dataset constrained to LF > 0.02 comprises only about 10% of the total data containing clouds; nevertheless it exhibits significant differences compared to the whole dataset regarding cloud liquid base height, cloud thickness, cloud-top temperature, and the lapse rate.

Previous studies have already shown differences in various cloud properties when classified by surface coupling or observations over ocean or sea ice. For instance, Gierens et al. (2020) found, using observation from Ny-Ålesund in Svalbard, that surface-coupled persistent MPCs contain about twice as much liquid as the decoupled clouds. The total amount of condensed water was higher for coupled persistent MPCs, which led the authors to suggest that a humidity source existed which is not available for the decoupled MPCs. This suggestion can be confirmed by the present study, since the WVT serves as a humidity source from sea ice leads. Papakonstantinou-Presvelou et al. (2022), using satellite products for large-scale clouds below 2 km in the Arctic, found a strong ocean-sea ice contrast in terms of ice crystal number concentration, with this difference between sea ice and ocean being enhanced for temperatures between 0 and -10 °C and clouds located south of 70° N latitude. Although our study encompasses a different scale, we found that the highest values of IWP are concentrated at low LF or high SIC.

Griesche et al. (2021) reported contrasting ice formation in summer Arctic clouds when separated by surface coupling from observations on board RV *Polarstern* in 2017, with a larger number of ice-containing clouds corresponding to surface-coupled clouds between -10 and -5 °C cloud minimum temperature. Although we use the ice water fraction instead, our study found contrasting differences for WVT coupled versus decoupled cases, but the differences were mainly located in the range between -15 and -25 °C cloud-top temperature.

Danker et al. (2022), using CloudSat-CALIPSO DAR-DAR product for clouds below 2.5 km, have also reported the increase in the occurrence of mixed-phase clouds and decrease in supercooled liquid clouds at temperatures around -15 °C, although they only consider cloud-top temperatures up to -20 °C. For similar dips in  $\chi_{ice}$  at lower temperatures, no other references have been found, which exposes the significant importance of understanding the microphysical processes related to the interaction between water vapour, liquid, and ice growth revealed by the ice water fraction. This exposed a clear asymmetry when data were separated by the coupling to WVT. This result requires further investigation, since such an impact of WVT on cloud properties has not been reported previously.

The presented study puts into consideration a methodology to study the influence of sea ice on cloud properties based on the observations from the MOSAiC expedition. Although MOSAiC is unprecedented in terms of providing a detailed dataset, it only comprises one winter. Therefore, a similar study is being extended to a period from 2012 to 2022 in the western Arctic using data from the ARM North Slope of Alaska (NSA) site in Utqiaġvik, which has an instrumental suite comparable to that of RV *Polarstern* during MOSAiC. Moreover, recent improvements in cloud-phase classification are being implemented. This refers to Schimmel et al. (2022), who use a radar Doppler spectrum for the detection of liquid layers above lidar attenuation and thus provide the potential to significantly improve cloud-phase target classification, which can then be used to support the findings of this study.

#### Appendix A: Cloud decoupling criteria

The virtual potential temperature is defined as

$$\theta_{\rm v} = \theta \left( \frac{1 + \frac{q_{\rm r}}{\epsilon}}{1 + q_{\rm r}} \right),$$
(A1)

where  $q_r$  is the water vapour mixing ratio  $[gg^{-1}]$ ;  $\epsilon \approx 0.622$  is the ratio of dry-air to wet-air gas constant; and  $\theta$  is the potential temperature  $\theta = T(\frac{P_0}{P})^{\kappa}$  [K], with  $\kappa \approx 0.286$  being the ratio of the dry-air gas constant and specific heat capacity for constant pressure.

In order to estimate the mixing layer below the cloud, the radiosonde profiles are used first to compute the virtual potential temperature according to Eq. (A1), and then the cumulative variance of the potential temperature is calculated as follows:

$$\sigma_{\Sigma}^{2}(z) = \sum_{i}^{z} \theta_{v}^{2}(i) - \left[\sum_{i}^{z} \theta_{v}(i)\right]^{2} \forall \quad i = \{\text{CBH}, \dots, 0\}.$$
(A2)

Equation (A2) is evaluated starting at cloud-base height (CBH) downwards until surface level; i.e. i = 0. The mixing layer below the cloud is then assigned to the altitude where Eq. (A2) first surpasses a threshold value of 0.01 K<sup>2</sup>, as explained in Sect. 3.3 of the main text.

Similarly to the case of the cloud-driven mixing layer above the cloud top, Eq. (A2) is evaluated from cloud-top height (CTH) upwards until a threshold of  $0.01 \text{ K}^2$  is fulfilled. A stricter criterion for the upper threshold is due to the fact that there are cases where the temperature inversion happens inside the cloud and not necessarily above CTH. For these cases,  $\theta_v$  can already be in a regime of adiabatic cooling; thus its variability might be small. This was also reported by Sedlar et al. (2012) for the central Arctic Ocean, where they found cloud top was frequently located at 100–200 m above the temperature inversion base.

# Appendix B: Derivation for the gradient of water vapour transport

The integrated vapour transport (IVT) is defined as the vertical integral of horizontal vapour fluxes (Zhu and Newell, 1998) and is normally used to identify ARs whenever either the IVT threshold of  $250 \text{ kg s}^{-1} \text{ m}^{-1}$  is exceeded or the IVT exceeds the 85th percentile of a climatologically varying value. IVT can be calculated by integrating the module of the wind vector  $v_w$  times the specific humidity  $q_v$ :

$$IVT = -\frac{10^2}{g} \int_{P_0}^{P_1} |q_v \cdot \boldsymbol{v}_w| dP, \qquad (B1)$$

where  $v_w$  is the horizontal wind speed [ms<sup>-1</sup>];  $q_v$  the specific humidity [gg<sup>-1</sup>]; *P* atmospheric pressure [hPa]; and *g* the constant of gravity, 9.81 ms<sup>-2</sup>. The factor 10<sup>2</sup> in Eq. (B1) expresses IVT in units of kgm<sup>-1</sup>s<sup>-1</sup> after the integration normally performed from surface reference pressure  $P_0$  to a nominal  $P_1 = 300$  hPa.

The vertical gradient of WVT is obtained starting from the definition of IVT given by Eq. (B1) and applying the derivative with respect to the vertical component z:

$$\frac{\mathrm{d}}{\mathrm{d}z}\mathrm{IVT}(P) = -\frac{10^2}{g}\frac{\mathrm{d}}{\mathrm{d}z}\left[\int_{P_0}^{P} |q_{\mathrm{v}} \cdot \boldsymbol{v}_{\mathrm{w}}|\mathrm{d}P\right].$$
(B2)

Using the chain rule, the integration variable can be changed from P to z as follows:

$$\frac{\mathrm{d}}{\mathrm{d}z}\mathrm{IVT}(z) = -\frac{10^2}{g}\frac{\mathrm{d}}{\mathrm{d}z}\left[\int_{z_0}^{z} \left(|q_{\mathrm{v}}\cdot\boldsymbol{v}_{\mathrm{w}}|\frac{\mathrm{d}P}{\mathrm{d}z}\right)\mathrm{d}z\right].$$
(B3)

Thus the integral can be cancelled out by the outer derivative, resulting in

$$\frac{\mathrm{d}}{\mathrm{d}z}\mathrm{IVT}(z) = -\frac{10^2}{g} \Big( |q_{\mathrm{v}} \cdot \boldsymbol{v}_{\mathrm{w}}| \frac{\mathrm{d}P}{\mathrm{d}z} \Big). \tag{B4}$$

Hereafter we rename the derivative of IVT with respect to the z variable in Eq. (B4) as the gradient along the vertical for the water vapour transport as  $\nabla_z WVT(z)$ , which has units of kg s<sup>-1</sup> m<sup>-2</sup> and results in

$$\nabla_z WVT = -\frac{10^2}{g} |q_v \cdot \boldsymbol{v}_w| \frac{\mathrm{d}P}{\mathrm{d}z}.$$
 (B5)

Note that we started from the definition of IVT given by Eq. (B1), where the horizontal wind speed is used. Some authors prefer to define IVT by means of the wind zonal and meridional components U and V, respectively. It is important to mention that both definitions are not mathematically identical and thus produce slightly different results.

#### Appendix C: Sea ice concentration offset correction

As reported first by Krumpen et al. (2021) and extensively studied by Rückert et al. (2023), there were several events of warm air mass intrusion (WAI) during the MOSAiC expedition, mainly in spring 2020. Those WAI events have fostered inaccuracies in some sea ice concentration retrievals (Figs. 1 and 9 in Rückert et al., 2023, and Krumpen et al., 2021, respectively). This is particularly the case for products from algorithms that use MWR polarization information at 36 GHz, i.e. the ASI algorithm. In the context of the present study those inaccuracies have fostered a misclassification of cloud properties when sorted as a function of SIC due to the SIC offset. Therefore it is paramount to correct the SIC product when the offset due to WAI is present. This section describes the details of this correction.

The underestimation by MODIS–AMSR2 SIC retrievals can be observed from Fig. C1b. For the period of mid-February to the end of May 2020, considerable disagreement has been found in the SIC products obtained by the MODIS–AMSR2 and the OSI SAF retrievals. Unfortunately, the resolution of the OSI SAF product is 25 km, which is not enough to resolve small leads relevant to this study. Therefore, OSI SAF is only being used as a reference product, and we do not imply OSI SAF provides error-free retrieval or absolute values of SIC.

In Fig. D2 can be seen the advantage of detecting leads by the MODIS–AMSR2 product (Fig. D2a) as compared to the



**Figure C1.** From top to bottom: (**a**) the ratio of OSI SAF by MODIS–AMSR2 SIC products. A ratio close to 1 indicates that both products are on average similar within the sector of study. (**b**) SIC from MODIS–AMSR2 (blue) and from the OSI SAF product (green) averaged within 50 km of RV *Polarstern*. The initial period and end of reported warm air intrusions are marked by the dashed vertical red and grey lines. (**c**) The same as (**b**) but with the MODIS–AMSR2 SIC corrected (blue). In all panels the shaded areas correspond to 1 standard deviation.

OSI SAF (Fig. D2b). Conversely OSI SAF sea ice concentration has the advantage of not being affected by the WAI events due to its different retrieval algorithm, which has been corroborated by Krumpen et al. (2021) and Rückert et al. (2023). Therefore those two products are being used in order to fix the SIC bias with OSI SAF while still keeping the high-resolution variability in the MODIS–AMSR2 product.

The time series of the OSI SAF and merged MODIS– AMSR2 products averaged over the 50 km radius around RV *Polarstern* is shown in Fig. C1. It clearly confirms the WAI compromised sea ice concentration mainly from February to May 2020 as reported by Rückert et al. (2023).

In order to ensure both products are statistically comparable, the MODIS-AMSR2 SIC is averaged by assuming a truncated normal distribution with 0 and 100 as lower and upper distribution limits. The averaging was performed within a 10 km grid centred on every OSI SAF coordinate grid; thus a MODIS-AMSR2 and OSI SAF dataset is achieved with the same spatial resolution as the OSI SAF grid. Then the ratio of those two products is calculated as an indicator of overor underestimation of ASI relative to OSI SAF. The ratio is shown in Fig. C1a for the entire period of interest. It can be seen that the greatest discrepancy occurs from mid-February to May (in agreement with findings by Rückert et al., 2023) with the OSI SAF SIC retrieval being up to 15 % higher than that of MODIS-AMSR2. On the other hand the ratio rarely reaches  $\pm 2\%$  for lower or higher MODIS-AMSR2 SIC outside the WAI period (mainly winter 2019), corroborating the fact that when no WAI events are experienced both products are comparable.

Since both SIC products are provided at a 24 h temporal resolution, the SIC ratio has been estimated on a daily basis. Every MODIS–AMSR2 SIC section of interest around RV *Polarstern* is then corrected by the SIC ratio to match on average the OSI SAF but keep the high spatial resolution and the variability within this sector of study. A comparison of this procedure is depicted in Fig. C1c, where the offset-free SIC is overlapped with the OSI SAF as a visual assessment for the feasibility of the method.

# **Appendix D: Supporting material**

#### D1 Case study for 15 April 2020

Sea ice observations for 15 April 2020. In this example, both MODIS–AMSR2 (1 km resolution) and Sentinel-1 (700 m) detected the lead, whereas OSI SAF (25 km resolution) is not able to detect the lead.



Figure D1. Synergy of remote sensing instruments on board RV *Polarstern* for 15 April 2020. (a) KAZR cloud radar reflectivity factor; (b) PollyXT lidar backscattering coefficient; (c) liquid water path from the microwave radiometer. The isotherms and wind vectors are obtained from the weather model at selected altitudes and time steps. (d) Cloudnet classification.



**Figure D2.** Satellite-based sea ice observations for 15 April 2020. (a) SIC from the MODIS–AMSR2 merged product after being offsetcorrected (see Appendix C). (b) SIC from OSI SAF. (c) LF from SAR Sentinel-1. The images are centred on the position of RV *Polarstern* (red star) on the given date. The RV drift is indicated by the red line; the dashed circle indicates the 50 km radius region of interest.

# D2 Statistical analysis additional results

Figure D3a shows the frequency of occurrence of coupledversus-decoupled cases when binned within intervals of total water path. For a total water path larger than  $50 \text{ gm}^{-2}$ (light- to dark-green colours), the coupled cases can be approximately 10 times more frequent than the decoupled cases (middle dashed red line), whereas cases with the total water path lower than  $50 \text{ gm}^{-2}$  are the most frequent observations (yellow colours) with a ratio of  $\sim 3/2$ . It can be seen that liquid clouds are mostly found below  $200 \text{ gm}^{-2}$  and are more frequent than the ice clouds.

Finally the PDFs for surface skin temperature are presented in Fig. D3b, showing an asymmetric coupled– decoupled bimodal distribution. Cases with LF > 0.02 are associated with a considerable warm surface for coupled cases, with one peak located at -6 °C, whereas the decoupled PDF shows one peak at -17 °C.



Figure D3. (a) The distribution of cloud observations as a function of the integrated water path for coupled versus decoupled cases. The colour scale indicates the cloud water path for ice (stars) and liquid (circles). (b) PDF for the surface skin temperature from the GND IRT sensor.

#### D3 List of abbreviations and variables

Abbreviation	Description
KAZR	Ka-band ARM zenith radar
MWR	Microwave radiometer
HATPRO	Humidity and Temperature Profiler
CEIL10m	Ceilometer 10 m resolution
INTERPSONDE	Interpolated Sonde
RV	Research vessel
AMSR2	Advanced Microwave Scanning Radiometer 2
MODIS	Moderate Resolution Imaging Spectroradiometer
MPCs	Mixed-phase clouds
ASI	ARTIST sea ice algorithm
OSI SAF	Ocean and Sea Ice Satellite Application Facility
SAR	Synthetic aperture radar
VAP	Value-added product

Symbol	Description	Unit
LWP	liquid water path	$gm^{-2}$
IWV	integrated water vapour	$kgm^{-2}$
LWC	liquid water content	$gm^{-3}$
IWC	ice water content	$gm^{-3}$
IWP	ice water path	$\mathrm{gm^{-2}}$
IVT	integrated water vapour	$kg s^{-1} m^{-1}$
	transport	
$\nabla_z WVT$	vertical gradient of water	$g s^{-1} m^{-2}$
	vapour transport	
CBH	cloud-base height	m
CTH	cloud-top height	m
CMLH	cloud mixing layer height	m
LF	sea ice lead fraction	_
SIC	sea ice concentration	%
Ze	radar equivalent reflectivity	dBZ
	factor	
$V_{\rm D}$	mean Doppler velocity	$m s^{-1}$
SWD	Doppler spectral width	$\mathrm{ms^{-1}}$
eta	lidar-attenuated	$m^{-1} sr^{-1}$
	backscattering coefficient	
δ	lidar depolarization ratio	-
Р	atmospheric pressure	Pa
$q_{ m v}$	specific humidity	$gg^{-1}$
$W_{\rm s}$	horizontal wind speed	$\mathrm{ms^{-1}}$
Wd	wind direction from the north	0
$T_{\rm V}$	virtual temperature	K
$\theta_{\rm v}$	virtual potential temperature	K
$\Sigma \sigma_i^2(\cdot)$	cumulative variance function	$K^2$
Rib	bulk Richardson number	_
Tgnd	skin surface temperature	K
$\Gamma_{cloud}$	cloud layer temperature lapse	$^{\circ}$ C km <sup>-1</sup>
	rate	
Xice	ice water fraction in cloud	_
_	layer	
$\bar{r}_{\rm eff}$	cloud layer average effective radius	μm

 Table D1. Symbols, description, and corresponding units of physical quantities relevant in this study.

**Code and data availability.** Data were obtained from the Atmospheric Radiation Measurement (ARM) user facility, a US Department of Energy (DOE) repository: https://arm.gov/data (ARM, 2023). The merged MODIS–AMSR2 1 km SIC product is publicly available at the University of Bremen data repository https: //seaice.uni-bremen.de (Sea Ice Remote Sensing, 2023). HATPRO MWR data are public (https://doi.org/10.1594/PANGAEA.941389, Ebell et al., 2022). PollyXT and Cloudnet classification is available at the Cloudnet data portal under https://hdl.handle.net/21.12132/1.8865776cccd74da3 (Engelmann et al., 2023). Lead fraction data based on the divergence product from Sentinel-1 are provided following von Albedyll et al. (2023). The Cloudnet retrievals were processed using a Julia package – https://doi.org/10.5281/zenodo. 7245669 (Saavedra Garfias, 2022) – and the dataset for the sta-

tistical analysis is public (https://doi.org/10.7910/DVN/DZSUV7, Saavedra Garfias and Kalesse-Los, 2023).

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Author contributions. PSG performed the data analysis and led the manuscript writing. HKL as project leader contributed to the research design and result analysis. LvA produced the sea ice divergence and lead fraction from SAR Sentinel-1 data. HG provided Cloudnet and the low-level cloud flag product for MOSAiC. GS provided the sea ice concentration products. All co-authors contributed with discussion and revision of the manuscript.

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