ATMOSPHERIC MOISTURE IN THE ARCTIC AND ANTARCTIC

TUOMAS NAAKKA
ATMOSHERIC MOISTURE IN THE ARCTIC AND ANTARCTIC

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ACADEMIC DISSERTATION in meteorology

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Atmospheric moisture in the Arctic and Antarctic

Water vapour is an effective greenhouse gas, but clouds, which are formed when water vapour condenses into water droplets or ice crystals, may have an even greater effect on radiative energy transfer through the atmosphere. In addition, absorption or release of the latent heat of vaporization and transport of water vapour are part of the heat transport from the Tropics towards the Poles. Thus, atmospheric water vapour greatly affects the energy balance of the atmosphere and is also an important component of the water cycle.

This thesis addresses the subject of atmospheric moisture and the processes affecting it in the Arctic and Antarctic. The studies comprising the thesis are mostly based on atmospheric reanalyses. In the polar regions, meteorological observation networks are sparse, due to their remoteness and the harsh environment, and therefore traditional observations have not provided a comprehensive picture of atmospheric conditions in the polar regions. In recent years, atmospheric reanalyses have also become more accurate in remote areas, which has enabled detailed studies of atmospheric moisture in the polar regions.

In the polar regions, the mostly negative radiation budget of Earth’s atmosphere-surface system shapes the distribution of water vapour in the atmosphere, especially the vertical structure of specific humidity. The polar regions are sinks for atmospheric water vapour, due to their typically small local evaporation, and even condensation of moisture on the surface. Therefore, moisture transport from the lower latitudes balances the moisture budget in the polar regions. This type of moisture budget favours the formation of specific humidity inversions. Our results show that specific humidity inversions are common in the polar regions, and their occurrence near Earth’s surface is linked with surface conditions: radiative surface cooling, occurrence of temperature inversions in winter and cold sea surfaces or melting of sea ice in summer. Advection of warm, moist air masses over a cold surface in summer is vital for formation of specific humidity inversions.

Below the approximately 800-hPa level, interactions between the atmosphere and Earth’s surface clearly affect both the atmospheric moisture content and moisture transport. Our results show that the northward moisture transport near the surface is mostly balanced by southward transport. Moisture transport clearly shapes the spatial distribution of the atmospheric moisture content. Regional trends in atmospheric moisture content in the Arctic are also mostly the results of long-term variations in atmospheric circulation.

The negative net radiation budget, weak evaporation and extensive contribution of moisture transport to atmospheric moisture content also characterize moisture conditions in the Antarctic. The results show that, due to geographical conditions, specific humidity inversions in Antarctica are even more persistent than those in the Arctic.
This is associated with stronger isolation of air masses in inner Antarctica from advection of warm, moist air masses than in the Arctic. The results also show that when a cold, dry air mass flows from the continent towards the ocean, it undergoes adiabatic warming, which together with downward sensible heat fluxes enables evaporation on Antarctic slopes. Overall, this thesis contributes to our understanding of how the spatial distribution of atmospheric moisture content interacts with moisture transport and with physical processes such as evaporation and condensation in polar regions.
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Nimeke
Ilmakehän kosteus Arktiksassa ja Antarktiksessa

Tiivistelmä

Vesihöyry on merkittävä ilmakehän kasvihuonekaasu, mutta vielä suurempi vaikutus säteilynkulkuun ilmakehän läpi on pilvillä, jotka muodostuvat, kun vesihöyry tiivistyy joko pilvipisaroiksi tai jääkiteiksi. Lisäksi latentin lämmön sitoutuminen ja vapautuminen, sekä vesihöyrynkuljetus ovat merkittävä osa lämmönsiirtoa tropiikista kohti napa-alueita, joten vesihöyryllä on merkittävä osa ilmakehän energiatastapainossa. Lisäksi ilmakehän vesihöyry on merkittävässä roolissa veden kiertokulussa.


Noin 800hPa-painepinnan alapuolella vuorovaikutukset ilmakehän ja pinnan välillä vaikuttavat huomattavasti ilmankosteuden pystyrakenteeseen ja kosteudenkuljetukseen. Väitöskirjan tulokset osoittavat, että pinnan lähellä etelään päin suuntautunut kosteudenkuljetus suurelta osin tasapainottaa pohjoiseen päin suuntauteen kosteudenkuljetuksen. Kosteudenkuljetus vaikuttaa myöskin huomattavasti alueelliseen ilmankosteuden
jakaumaan. Väitöskirjan tulokset osoittavat, että ilmankosteuden pitkäaikaiset alueelliset muutokset liittyvät pääasiassa ilmakehän kiertoliikkeen pitkäaikaisiin muutoksiin.


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The research that has led to this thesis began in spring 2016, when prof. Timo Vihma provided me a job as a summer worker in the project he was leading. Docent Tiina Nygård had planned a study addressing atmospheric moisture in the Arctic. My research work continued after the summer, and the topic of my dissertation initiated from that study. This dissertation would not have been possible without assistance and support from several people.

First, I would like express gratitude to my supervisors Timo and Tiina. They have been my supervisor during the whole Ph.D. project. Timo has always found time for discussions in spite of his hurries. His constructive advice and positive feedback have created trust and promoted the work, when I have faced difficulties in the studies. Tiina has exhaustively taught me how to write a good scientific article, and she has a talent to ask very good and challenging questions. Answering to these questions has often required examining the problem from a different point of view. Searching answers to those questions has not only promoted the study, but also helped me to grow as a scientist. I have often had very fruitful and also fun discussions, which has provided plenty of joy for working days, with my supervisors. In addition, I would like to thank all my co-authors who have taken part in the studies of this thesis.

I am grateful to prof. Peter Langen for acting as an opponent and prof. Heikki Järvinen for acting as a custos. In addition, I would like to thank pre-examiners prof. Harald Sodemann and Dr. Felix Pithan for their time to evaluate the thesis.

Financial support for the research of this thesis has been provided by the Academy of Finland via AFEC, ASPIRE and TWASE projects, which were led by Timo Vihma, and by the Yrjö, Vilho and Kalle Väisälä foundation, which gave financial support for finalizing the thesis.

I would like to thank FMI for providing good facilities for productive work. Finally, I would like to thank all colleagues, especially polar meteorology and climatology group, for creating encouraging and inspiring working environment in spite of recent pandemic years.

Kotka, January 2022
Tuomas Naakka
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LIST OF ORIGINAL PUBLICATIONS

This thesis consists of four scientific articles identified in the text by their Roman numerals (I–IV):


The author contributed to planning of the research (I, II) and was mainly responsible for analysing the data, interpreting the results and writing the manuscript (I, II). The author further contributed to planning, analysing the data and interpreting the results (III), and was later mainly responsible for all phases of study: planning, to writing the manuscript (IV).
1. INTRODUCTION

Water vapour is a significant greenhouse gas that affects radiative energy transfer through the atmosphere. In contrast to most other greenhouse gases, especially carbon dioxide, human activities do not directly affect the amount of water vapour in the atmosphere. Instead, the amount is mostly linked with the water-vapour holding capacity of the air and the physical processes of evaporation and condensation, which determine the water cycle in the atmosphere. The water-vapour holding capacity increases exponentially with temperature. Therefore the amount of water vapour in the atmosphere tends to increase when the temperature increases, which intensifies global warming by increasing the sensitivity of the climate to forcing of other greenhouse gases. Water-vapour feedback increases the radiative forcing of well-mixed greenhouse gases by a factor of 2 to 3 (Myhre et al. 2013), because the temperature increase resulting these gases also increases the water-vapour content of the air, which strengthens the greenhouse effect.

In addition to its radiative effects, water vapour affects heat exchange between Earth’s surface and the atmosphere and redistributes heat in the atmosphere. Heat transport due to water vapour is based on the latent heat, which is absorbed/released when water vapour evaporates/condenses. On the global scale at Earth’s surface, heating by absorbed solar radiation typically exceeds cooling by emitted thermal radiation. Therefore excessive heat is transferred from the surface to the atmosphere via turbulent fluxes of sensible and latent heat. Similarly, excess heat gained in the entire surface–atmosphere system, due to net solar and thermal radiation, is greater in the Tropics than in the polar regions, where the radiative heat balance is negative. This is balanced by horizontal transport of latent heat and dry energy from the Tropics to the polar regions. The transport of water vapour together with evaporation and precipitation form the atmospheric component of the water cycle (Vihma et al. 2016). It determines the amount of precipitation and thus directly affects, e.g., the mass balance of ice sheets. Accordingly, atmospheric moisture plays an important role in energy balance and the water cycle, both globally and regionally in the polar regions.

Condensation of moisture in the atmosphere results in the formation of clouds. The influence of clouds on radiative energy transfer through the atmosphere is even larger than the influence of water vapour. In addition, uncertainty in the occurrence and properties of clouds results in a large degree of uncertainty for climate predictions. In the polar regions, clouds could have dramatic impact on surface temperature in winter, because of their capability for reducing surface cooling due to outgoing thermal radiation (Stramler et al. 2011). In contrast to water vapour, clouds may also
have a cooling effect on surface temperatures when they reduce incoming solar radiation reaching the surface. The net effect of clouds is dependent on their properties as well as the vertical and geographic distribution of their occurrence (Pavolonis and Key 2003, Shupe and Intrieri 2004, Stramler et al. 2011). Even though the properties of clouds are influenced by the cloud microphysics associated with e.g. the properties and occurrence of atmospheric aerosol particles, the occurrence of clouds is often largely controlled by moisture transport caused by large-scale circulation and distributions of atmospheric water vapour.

Atmospheric water vapour is thus an important part of climate systems in the polar regions. While the direct effect of water vapour, linked to phase changes, on atmospheric energy balance in the polar regions is small, due to the small amount of water vapour in polar atmospheres, its indirect effects are much greater, resulting from the effect of water vapour and clouds on radiation transfer through the atmosphere. However, the distribution of atmospheric water vapour and especially its vertical structure in the polar regions are not well known. Typically, the specific humidity decreases upwards in the atmosphere, but in the polar regions, layers where specific humidity increases with height, i.e. specific humidity inversions, are common (Devasthale et al. 2011, Nygård et al. 2013, 2014, Brunke et al. 2015). Specific humidity inversions are important features in polar atmospheres, because they may contribute e.g. to the occurrence of clouds (Solomon et al. 2011, 2014, Savre et al. 2015). However, the spatial and seasonal distributions of specific humidity inversions have not been comprehensively studied.

Low evaporation and even condensation on the surface (Andreas et al. 2002) together with moisture transport from the lower latitudes allow frequent occurrence of specific humidity inversions in the polar regions (Devasthale et al. 2011, Nygård et al. 2013, 2014, Brunke et al. 2015). Formation of near-surface specific humidity inversion is often associated with formation of polar air masses, due to radiative cooling at the surface. Surface-radiative cooling causes formation of a temperature inversion, and after saturated conditions are reached, it results in the formation of a specific humidity inversion due to moisture condensation (Curry 1983). In addition to moisture condensation due to radiative cooling, vertically varying moisture advection has been suggested as an important mechanism resulting in formation of specific humidity inversions (Nygård et al. 2013, 2014, Brunke et al. 2015).

Moisture transport is the most important source of atmospheric moisture in the Arctic (Serreze et al. 1995, Jakobson and Vihma 2010, Dufour et al. 2016). As such, it strongly affects moisture and cloud conditions (Nygård et al. 2019). In addition, moisture transport can affect the evolution of arctic sea ice. Moisture transport,
associated with warm-air advection to the Arctic, together with its effects on
radiation budget due to increased cloudiness, is able to reduce sea-ice growth in
winter and can affect the onset of melting in spring (Kapsch et al. 2013, Mortin et al.
2016, Woods and Caballero 2016). Moisture transport has been addressed in many
previous studies, but they have mostly focused on meridional net-moisture transport
(e.g. Serreze et al. 1995, Jakobson and Vihma 2010, Dufour et al. 2016) or strong
moisture-transport events (Woods et al. 2013, Liu and Barnes 2015). However, net-
moisture transport consists of meridional and zonal components, the last also being
important for transporting moisture from source to sink regions, even though it does
not directly affect moisture exchange between the Arctic and midlatitudes. Both the
total exchange of moisture between the Arctic and midlatitudes and zonal moisture
transport have received less awareness in recent studies.

In the Arctic, the moisture environment has undergone significant changes, due to
climate change (Serreze et al. 2012, Rinke et al. 2019). Strong climate warming in
the Arctic increases the water-vapour holding capacity of the air, which presumably
increases the atmospheric water-vapour content. However, increases in this content
have varied widely (Rinke et al. 2019). In the Arctic, the temporal distribution of
atmospheric moisture is largely determined by moisture transport (Nygård et al.
2019). Therefore, long-term changes in moisture transport due to long-term
variations in atmospheric circulation probably affect the spatial distribution of
atmospheric moisture. In addition to moisture transport, local evaporation affects
atmospheric moisture content. During recent decades, the sea-ice cover has strikingly
decreased in the Arctic Ocean, which has potentially increased local evaporation in
the Arctic. However, efficient local evaporation is possible only when a dry air mass
is advected over the open ocean (Nygård et al. 2019), and the sea-ice decline
decreases the opportunities available for formation of a dry polar air mass.

Overall, moisture environments in polar regions have not been comprehensively
studied, and knowledge about atmospheric moisture conditions in the Antarctic is
even weaker than that in the Arctic. In this thesis, I address the subject of atmospheric
moisture in the polar regions and the processes affecting it. The following research
questions were investigated:

1) How are specific humidity inversions distributed in space and time, and which
processes are responsible for formation of specific humidity inversions in the Arctic?

2) What type of vertical structure does moisture transport assume in the Arctic, and
how do northward and southward moisture transports contribute to the net poleward
moisture transport?
3) How have changes in moisture transport due to changes in atmospheric large-scale circulation affected long-term changes in atmospheric moisture content?

4) How has the sea-ice decline affected surface evaporation and increase the atmospheric moisture content in the Arctic?

5) How has the atmospheric moisture content interacted with moisture transport and physical moisture processes in the Antarctic?

These research questions were approached, utilizing atmospheric reanalyses. In both polar regions, their remoteness and harsh environment complicate making observations on the distribution of atmospheric water vapour, which hinders knowledge of its distribution and understanding of the processes controlling its content in the polar regions. In recent years, development of atmospheric reanalyses has also broadened our knowledge about climatological conditions in remote areas.

This thesis consists of four scientific articles, denoted I to IV. In the first, the focus is on the vertical structure of specific humidity, especially specific humidity inversions (I). Moisture transport, as well as evaporation and moisture condensation, shape the vertical structure of atmospheric moisture. The effects of these processes on the formation of specific humidity inversions were analysed (I), while moisture transport in the Arctic and moisture exchange between the Arctic and midlatitudes were similarly addressed (II). The impact of long-term variations in atmospheric circulation on the regional trends in atmospheric water-vapour content was determined (III), as were the effects resulting from interaction of moisture transport and atmospheric moisture content with evaporation and cloudiness during sea-ice retreat (III). The aims of the fourth article were to form a comprehensive picture of atmospheric moisture conditions in the Antarctic and to determine how moisture conditions are affected by physical processes (IV). The overall aim of this thesis is to broaden our understanding of how the spatial distribution of atmospheric moisture interacts with moisture transport as well as with physical processes, such as evaporation and moisture condensation, in the polar regions.
2. THEORETICAL BACKGROUND AND BOUNDARY CONDITIONS FOR THE POLAR ATMOSPHERES

2.1 THERMODYNAMICS OF WATER VAPOUR

Water is an important substance in Earth’s atmosphere, since it can exist in three phases: vapour, solid and liquid. Water in the vapour phase is always present in the atmosphere, but the occurrences of water droplets and ice crystals are also frequent. Formation of water droplets and ice crystals are associated with the conditions present when the partial pressure of water vapour exceeds the saturation vapour pressure and leads to condensation of water. The saturation vapour pressure is linked with air temperature, as described by the Clausius–Clapeyron equation (Equation 1),

\[ e_{\text{sat}}(T) = e_{\text{sat} \ 0} \exp \left[ -\frac{L}{R_v \left( \frac{1}{T} - \frac{1}{T_0} \right)} \right] \]  

where \( e_{\text{sat}} \) is the saturation vapour pressure, \( e_{\text{sat} \ 0} \) the saturation vapour pressure at the reference point, \( T_0 \) the temperature at the reference point, \( L \) the latent heat of vaporization (approx. 2500 kJ kg\(^{-1}\) at 0 °C) and \( R_v \) the specific gas constant of water vapour (461 J kg\(^{-1}\) K\(^{-1}\)). The triple point of water (\( T_0 = 273.16 \) K and \( e_{\text{sat} \ 0} = 612 \) Pa) is often used as a reference point. This equation indicates that the saturation vapour pressure increases exponentially with temperature in such a way that it approximately doubles when the temperature increases by 10 K (Figure 1).

The Clausius-Clapeyron equation defines the saturation vapour pressure over a flat water surface. The same equation can be applied for ice surfaces if the latent heat of sublimation is used instead of the latent heat of vaporization. The saturation vapour pressure with respect to water is slightly larger than that with respect to ice, which means that in the presence of ice, water vapour tends to condensate on the ice surface before the air reaches the saturation point with respect to liquid water. In clouds, this often causes evaporation of water droplets after ice crystals are formed. In addition, the Clausius-Clapeyron equation ignores the curvature effects on saturation vapour pressure associated with the surface energy of particles. Hence, condensation in the atmosphere in practice needs the presence of aerosol particles, condensation or ice nuclei and typically at least some slight supersaturation before condensation begins. However, the saturation vapour pressure approximately defines the upper limit of the amount of water vapour in the atmosphere and is used for defining the relative humidity of the air, which is the ratio between the actual water-vapour pressure and saturation vapour pressure (Figure 1).
Figure 1. Associations between the partial pressure of water vapour, temperature and relative humidity. The arrows describe the effects of physical processes on these variables.

Water-vapour concentrations near water or ice surfaces tend to be in equilibrium with the surface when the actual vapour pressure equals the saturation vapour pressure. Otherwise, the water molecule flux from the sea or land surface or from water droplets or ice crystals in the atmosphere attempts to balance the vapour-pressure deficit, or if the actual vapour pressure is higher than the saturation vapour pressure, the water molecules tend to condense into water droplets or ice crystals in the atmosphere or on Earth’s surface, forming dew or hoar frost. Temperature changes affect the saturation ratio of the air. Supersaturation in the atmosphere is produced by cooling of the air mass, which is usually caused by adiabatic cooling, due to upward motion of the air, or radiative cooling resulting from outgoing long-wave radiation. Mixing of air masses may also lead to supersaturation.

Adiabatic cooling is the most important process leading to moisture condensation. Most of the moisture is removed from the atmosphere, due to precipitation caused by upward motion of the air, which results in adiabatic cooling of an air mass, condensation of water vapour and formation of precipitation. Therefore, much of the removal of atmospheric water vapour occurs in the atmosphere, due to phase changes from vapour to the liquid or solid phase. In contrast, opportunities for evaporation are limited in the atmosphere. Evaporation in the atmosphere almost always occurs when precipitating particles fall through unsaturated air. Thus, most of the evaporation occurs from either land or sea surfaces. The near-surface air attempts to
be in humidity equilibrium with the surfaces beneath. However, the surfaces are often moister than the air above, if the air is not saturated, providing a moisture source for the atmosphere. The vertical asymmetry between the sinks and sources of water vapour leads to generally upward transport of water vapour, due to upward motion of moist air and downward motion of dry air.

Phase changes of water release or absorb latent heat. Evaporation requires energy for a phase change from liquid or solid phase to the gas phase, whereas condensation releases latent heat. Hence, moisture fluxes are associated with heat fluxes in the atmosphere.

2.2 PLANETARY-SCALE HEAT BUDGET AND CIRCULATION

![Figure 2. Mean net solar and thermal radiation (Wm$^{-2}$, positive values when the flux is downward), on the top of the atmosphere (TOA) and at the surface in the Northern Hemisphere in winter and Southern Hemisphere in summer (DFJ, left) and in the Northern Hemisphere in summer and Southern Hemisphere in winter (JJA, right), based on fifth-generation European Reanalysis (ERA5) reanalysis.](image)

Incoming solar radiation is not evenly distributed around the globe. In addition, the seasons affect the distribution of solar radiation. Here, the standard 3-month seasons are used, except in IV, hence winter (summer) in the Arctic (in the Antarctic) is from December to February, DJF, and summer (winter) in the Arctic (in the Antarctic) is from June to August (Northern Hemisphere summer and Southern Hemisphere
winter, JJA). Overall, the Tropics gain remarkably more solar radiation than the polar regions (Figure 2). In contrast, outgoing thermal radiation is clearly more evenly distributed. This results in the Tropics typically gaining energy due to positive net radiation, whereas the polar regions lose energy. This regional asymmetry in radiation budget is the driving force behind atmospheric circulation.

The radiation budget can be presented separately for the atmosphere and Earth’s surface. The zonally averaged seasonal mean radiation budget of the atmosphere is negative, approximately 100 Wm\(^{-2}\), at all latitudes, because the outgoing thermal radiation exceeds the sum of surface-emitted thermal radiation and solar radiation absorbed in the atmosphere. In contrast, the mean radiation budget of Earth’s surface is positive, because the solar radiation absorbed into the surface is mostly larger than the net effect of incoming and outgoing thermal radiation, except in the polar regions in winter, where the amount of incoming solar radiation is so small that the negative net thermal radiation dominates the radiation budget. The excessive heat due to the positive radiation budget is transferred from the surface to the atmosphere via sensible and latent heat fluxes (Figure 3).

![Figure 3. Mean sensible heat and latent heat surface fluxes (Wm\(^{-2}\), positive values denote upward heat flux) in the Northern Hemisphere winter and Southern Hemisphere summer (DJF, left) and in the Northern Hemisphere summer and Southern Hemisphere winter (JJA, right), based on fifth-generation European Reanalysis (ERA5) reanalysis.](image-url)
Spatial imbalance in the radiation budget is balanced by heat transport in the atmosphere and oceans from low to high latitudes (Figure 4). Uneven heating generates circulations in the atmosphere that transport heat from the equator towards the Poles. The heat transport can be divided into dry energy and latent heat, which is released when water vapour condenses. Hence, the water cycle in the atmosphere is a part of atmospheric energy transport. Availability of energy largely defines the geographical distribution of evaporation. In the Tropics and midlatitudes, the positive surface net radiation budget enables evaporation (Figure 3). On average between 60 °N and 60 °S, evaporation exceeds precipitation, and the surplus water vapour is transported to the polar regions. In the polar regions, the radiative energy budget of the atmosphere-surface system is negative, and the energy deficit is compensated for by transport of latent heat and dry energy from the lower latitudes. The contribution of dry energy transport increases towards the Poles, since cold air can contain only small amounts of water vapour. In addition, the surface net radiation budgets in the polar regions are negative, which favours downward heat fluxes from the atmosphere to the surface at least during the cold season. A major part of the downward heat flux consists of sensible heat flux, but occasionally the latent heat flux is also directed downwards.

The general circulation in the atmosphere is traditionally described with a three-cell structure in both hemispheres: the Hadley, Ferrel and Polar cell. These cells characterize the mean meridional circulation in the atmosphere, but at the mid- and high latitudes the mean meridional circulation is only responsible for part of the total heat and water-vapour transport. A large part of the heat and moisture transport is carried by transient features in the flow field, most importantly synoptic-scale cyclones and stationary planetary waves (Tietäväinen and Vihma 2008, Jakobson and Vihma 2010, Dufour et al. 2016, 2019 Fearon et al. 2021). Hence, the transports are often divided into parts of the mean meridional circulation, stationary eddies and transient eddies, using Reynolds decomposition (Palmen and Vuorela 1963). Equation 2 shows the Reynolds decomposition for meridional moisture transport.

\[
[\bar{q} \bar{v}] = [\bar{q}][\bar{v}] + [\bar{q}^* \bar{v}^*] + [\bar{q}' \bar{v}']
\]  

(2)

where the first term on the right-hand side is moisture transport due to mean meridional circulation, the second term is moisture transport caused by local departures from the zonal mean values, i.e. stationary eddies and the third term is moisture transport caused by temporal departures from local mean values, i.e.
transient eddies. In the equation, the overbar stands for a temporal mean and the square brackets for a zonal mean.

Figure 4. Mean poleward transports of dry energy and latent heat (W m\(^{-1}\)) in the Northern Hemisphere winter and Southern Hemisphere summer (DJF, left) and in the Northern Hemisphere summer and Southern Hemisphere winter (JJA, right), based on fifth-generation European Reanalysis (ERA5) reanalysis. The yellow arrows indicate the direction of meridional latent heat transport and the blue arrows indicate the direction of meridional dry energy transport.

2.3 GEOGRAPHICAL CHARACTERISTICS OF THE POLAR REGIONS AND THEIR EFFECT ON ATMOSPHERIC CIRCULATION

The large-scale radiation budget together with atmospheric meridional circulation set the boundary conditions for moisture environment in the polar regions. However, regional moisture transport and local moisture conditions are largely affected by geographical features. Due to these geographical conditions, the circulation patterns, which largely define moisture transport, differ remarkably between the Arctic and the Antarctic. In the northern polar region, the locations and orientations of mountain ranges as well as locations of the continents and oceans generate strong standing waves in the atmosphere (Wills et al. 2019). In winter, the planetary waves interacted with development of synoptic-scale cyclones generating two storm tracks over the North Atlantic and the North Pacific Oceans. In the Southern Hemisphere, the region
Figure 5. Mean sea-level pressure in the Arctic and Antarctic in the Northern Hemisphere winter and Southern Hemisphere summer (left, DJF) and in the Northern Hemisphere summer and Southern Hemisphere winter (right, JJA), based on fifth-generation European Reanalysis (ERA5) reanalysis.
near the Pole consists of Antarctica surrounded by the Southern Ocean. The geographical conditions in the southern polar region are thus rather zonally symmetric, enabling an almost zonally symmetric circulation pattern in the Antarctic.

Synoptic-scale cyclones are responsible for a large part of moisture transport to the Arctic (Jakobson and Vihma 2010, Dufour et al. 2016), thus the orientations and locations of storm tracks remarkably affect moisture conditions in the Arctic. Winds on the eastern side of a storm track are often from the south or southwest, causing poleward transport of warm, moist air masses. The climate in the Atlantic sector of the Arctic is therefore mild and the atmospheric moisture content higher than on average at the same latitudes (Jakobson and Vihma 2010, Rinke et al. 2019). The circulation often allows moisture transport to the central Arctic from the Atlantic sector (Nygård et al. 2019), and thus moisture transport to the central Arctic is largest from the Atlantic sector (Jakobson and Vihma 2010, Dufour et al. 2016). In contrast, mountain ranges in Alaska and Northwest Canada limit moisture transport from the North Pacific Ocean to the Arctic (Jakobson and Vihma 2010) in the cold seasons.

The large-scale circulation pattern in the Antarctic is rather zonally symmetric. The largest cyclonic activity occurs over the ocean around Antarctica, where the minima of mean sea-level pressure are also located (Figure 5). Synoptic-scale cyclones in the southern polar region display rather zonal tracks and relatively seldom penetrate into the continent, especially in East Antarctica, where the elevation of the ice sheet is high (Jones and Simmonds 1993, Simmonds and Keay 2000). This decreases moisture transport and causes extremely dry conditions. The occurrence of marine air masses over West Antarctica is more frequent than over East Antarctica, because the Amundsen Sea Low, seen as an area of minimum mean sea-level pressure in the Southern Hemisphere winter in Figure 5, often steers marine air masses towards the continent in the area west of the Antarctic Peninsula (Tsukernik and Lynch 2013).

2.4 SURFACE TYPES AND SEASONS

Earth’s surface conditions affect evaporation and moisture condensation on the surface. Surface conditions vary widely, regarding surface type and season, especially when associated with the ability of solar radiation to heat the surface. In the Arctic, the area near the Pole consists of ocean, which in winter is covered by ice, but much of the ice melts during summer. The Arctic Ocean is mostly surrounded by continents. In the Antarctic, the pole region is located on a continental ice sheet, which is surrounded by the ocean. Sea ice displays wide seasonal variation around Antarctica. The thermal and radiative properties of the main surface types (the open ocean, sea ice, land with seasonal snow cover and ice sheets) affect regional moisture processes.
Figure 6. Mean sensible heat flux (W m$^{-2}$, positive values denote upward heat flux) in the Arctic and Antarctic in the Northern Hemisphere winter and Southern Hemisphere summer (DJF, left) and in the Northern Hemisphere summer and Southern Hemisphere winter (JJA, right), based on fifth-generation European Reanalysis (ERA5) reanalysis.
In winter, the surface temperatures of both sea ice and land, mainly snow-covered, are constrained by local radiative budgets, due to low heat conductivity through the snow or ice. Therefore, outgoing thermal radiation is able to efficiently decrease surface temperatures under cloud-free conditions, which often causes downward sensible and latent heat fluxes (Persson et al. 2002). However, the thin sea ice and a thin snowpack on top of it allow heat conduction from the ocean through the sea ice, and increase heat fluxes from the ocean to the atmosphere, which affects turbulent heat fluxes at the surface. Clouds affect surface temperatures over snow and ice surfaces, because they remarkably increase downward thermal radiation, reducing surface-radiative cooling and resulting in a rise in surface temperature (Stramler et al. 2011) and thus weakening of the downward heat fluxes. However, moisture condensation on the surface is common over land and sea ice in winter (Figure 7), which favours the formation of specific humidity inversions (Curry 1983). In contrast, the open ocean has a relatively warm surface in winter. The surface layer of the open ocean has large heat capacity and thus the surface temperature varies only slightly, regardless of weather conditions. Since the surface of the open ocean typically is relatively warm, and availability of moisture does not limit evaporation, evaporation is often extensive over the ocean. However, specific humidity (which is often closely associated with air temperature) of the air above strongly affects evaporation efficiency. Advection of cold, dry air masses from the continent or from sea ice enable strong evaporation from the open sea surface, due to the low specific humidity of the advected air (Pithan et al. 2018, Nygård et al. 2019). Strong vertical mixing due to unstable stratification, which is a result of the upward sensible heat flux, strengthens evaporation. However, when an air mass originates from a warmer area, typically from lower latitudes, evaporation from the sea surface is weak (Nygård et al. 2019). This probably results from a small difference between the air-specific humidity and saturation-specific humidity at the sea-surface temperature, which is a result of an originally high water-vapour content of the poleward-advected air mass.

In summer, the thermal properties of Earth’s surface are different from those in winter. Sea ice is still a sink of sensible heat (Figure 6), not primarily due to the radiation budget but rather to the latent heat required to melt the ice. Therefore, the surface temperature is bound to the melting point of ice. The radiation budget over sea ice is positive, but the energy is consumed in melting the ice, and thus the turbulent heat fluxes are generally small (Figure 6 and 7). The exception here is when a warm-air mass has been advected over the sea ice. This type of situation is able to produce remarkable downward heat fluxes and downward thermal radiation (Sotiropoulou et al. 2016, Tjernström et al. 2019, You et al. 2021). In contrast to the
winter situation, the contrast of thermal and moisture properties between the sea ice and open ocean near the sea-ice margin is small. Both surfaces are wet, and the temperature difference is small because of the large heat capacity of the surface layer of the sea. The surface albedo is smaller for the open ocean than for the sea ice, even though the albedo is smaller for wet snow than for dry snow. However, solar radiation is typically absorbed in a relatively deep layer of water, so that the effect of solar radiation on the surface temperature is small.

On land in summer, solar radiation is absorbed into the surface and it strongly affects the surface temperature, enabling large turbulent heat fluxes. In the Antarctic, land areas are mostly covered by the ice sheet. Its snow-covered surface reduces absorption of solar radiation, and the surplus radiative energy is often rather confined to melting of snow and ice than turbulent heat fluxes. Therefore, moisture condensation onto the surface occasionally also occurs over snow and ice surfaces in summer (IV). The contrast between snow and ice surfaces and bare ground is extensive in summer. A convective well-mixed boundary layer is common over snow-free and ice-free land areas in summer (Esau and Sorokina 2010). Vertical mixing, due to convection generated by sensible heat fluxes, is able to decrease relative humidity and increase surface evaporation. Thus in summer, evaporation is strongest over land areas.

Spring and autumn are transition seasons between summer and winter. In early autumn, the sea-ice cover reaches its annual minimum, which together with increasing temperature and humidity difference between open sea surface and air above during winter enables greater evaporation from the sea in autumn than in spring. In contrast, land areas react more quickly to decreasing solar radiation than does the sea, which results in decrease in surface temperature and turbulent heat fluxes over land. In spring, seasonal sea ice reaches its maximum extent, and land areas are mostly covered by seasonal snow cover. The snow cover prevents increases in surface temperature due to its high albedo, which decreases the amount of solar radiation absorbed into the surface, while most of the surplus energy is used to melt snow. Hence, the thermal contrast between land and sea in spring is not as large as in autumn, because the sea is covered by melting sea ice.
Figure 7. Mean latent heat flux (in W m$^{-2}$, positive values denote upward heat flux) in the Arctic and Antarctic in the Northern Hemisphere winter and Southern Hemisphere summer (DJF, left) and in the Northern Hemisphere summer and Southern Hemisphere winter (JJA, right), based on fifth-generation European Reanalysis (ERA5) reanalysis.
2.5 Role of Air Moisture in the Arctic Climate Change

In the Arctic the near-surface air temperature is increasing more quickly than average on Earth, which is often referred to as Arctic amplification (Graversen et al. 2008, Serreze et al. 2009, Screen and Simmonds 2010, Cohen et al. 2014). Many factors contribute to Arctic amplification, and atmospheric water vapour also plays an important role in rapid warming in the Arctic, even though the direct radiative effect of increasing amounts of water vapour in the atmosphere causes more extensive warming in the Tropics than in the Arctic (Pithan and Mauritsen 2014). Arctic amplification is closely associated with the thermal structure of the troposphere in the polar regions. In the polar regions, the troposphere is often stably stratified, and temperature inversions are common (Serreze et al. 1992, Tjernström and Graversen 2009, Devasthale et al. 2010, Pithan and Mauritsen 2014). Therefore, small increases in heat supply to the troposphere near the surface are able to cause relatively large increases in near-surface temperature (Manabe and Wetherald 1975).

Sea-ice diminishing with surface albedo feedback may be a factor behind the current rapid warming in the Arctic (Serreze and Francis 2006, Dai et al. 2019). Sea-ice diminishing as well as shortening of the seasonal snow-cover period decrease the surface albedo in the Arctic, because snow and ice surfaces reflect a large part of the incoming solar radiation, whereas the open sea or bare ground absorbs most of the incoming solar radiation, and thus increases the amount of solar radiation absorbed. However, over the Arctic Ocean, the largest increases in near-surface temperature were observed in the cold seasons (Graversen et al. 2008, Simmons and Poli 2015) when the albedo feedback is weakest, due to the very small amounts of incoming solar radiation. In summer, the increase in near-surface temperature has been modest over the Arctic Ocean (Graversen et al. 2008, Simmons and Poli 2015), due to the large heat capacity of the open sea, and because in the areas of melting ice, the near-surface temperature is strongly constrained by the melting-point temperature of ice. In contrast, the decrease in surface albedo has the greatest effect on the surface heat balance during the warm seasons, but the extra heat is confined to melting of the sea ice and is stored in seawater, which deaccelerates the growth of sea ice in winter (Serreze and Francis 2006, Stroeve et al. 2014). In winter, the near-surface air temperature varies widely between the open ocean and sea ice. Over the open ocean, sensible heat fluxes from the sea surface increases the near-surface air temperature, whereas over the sea ice, an insulating effect of sea ice allows cooling of the surface due to emitted thermal radiation, which weakens the surface heat fluxes and decreases the near-surface air temperature. Therefore, changes in the sea-ice cover substantially affect the near-surface air temperature during the cold seasons.
Several studies (Park et al. 2015, Boisvert et al. 2016, Woods and Caballero 2016) have shown that strong moisture transport to the Arctic, together with its effects on cloudiness and radiation, cause sea-ice melt or decrease sea-ice growth, resulting in diminishing of sea ice. Strong moisture transport in winter or spring is able to affect sea-ice minimum in autumn (Kapsch et al. 2013, Mortin et al. 2016). An early onset of melting in spring increases the accumulation of heat on the surface during the entire melting period (Stroeve et al. 2014, Mortin et al. 2016). Wet snow and bare ice have notably lower albedo than dry snow, which increases the amount of absorbed solar radiation on the surface and thus enhances melting. Cloudy conditions favour early start of the melting period, since clouds increase downward thermal radiation (Mortin et al. 2016). On the other hand, the effect of clouds on solar radiation mostly causes surface cooling, because clouds reflect solar radiation and therefore decrease the amount of it reaching the surface. Therefore, in summer, clouds have a cooling effect over the open seas and bare ground. Cloudy conditions over the Arctic Ocean during the cold season are often a result of warm, moist air advection from lower latitudes (Stramler et al. 2011, Nygård et al. 2019), suggesting an important contribution of atmospheric moisture and moisture transport to the rapid climate change in the Arctic.

The atmospheric moisture content is expected to increase as a result of climate change because of the temperature dependence of saturation vapour pressure. Several studies (Serreze et al. 2012, Dufour et al. 2016, Rinke et al. 2019) have confirmed that the atmospheric moisture content in the northern polar region has increased during recent decades. Decrease in sea ice potentially increases local evaporation, especially in winter when the open sea surface is relatively warm. However, efficient evaporation is possible only when the overlying air is dry, which in the Arctic practically means advection of cold air from the sea-ice zone or continents over the open ocean. These situations are typically associated with circulation patterns in which the flow is from the Arctic towards the midlatitudes, and therefore increased evaporation from the oceans probably has only a small effect on atmospheric moisture content in the central Arctic (Nygård et al. 2019).
3. MATERIAL AND METHODS

The studies comprising this thesis utilize information on the three-dimensional distribution of moisture, winds and temperature in the atmosphere; hence profile data are vital for these studies. Atmospheric profile measurements are traditionally based on radiosonde soundings. Currently, aircraft measurements are yielding increasing numbers of profile observations in densely populated areas, but not in the polar regions. Temperature and moisture profiles can also be derived from emitted or reflected radiances at the top of the atmosphere (TOA) measured by satellites. In addition, reanalyses provide three-dimensional distributions and complete time series of atmospheric variables. The studies comprising this thesis are mostly based on reanalyses, but the results of the reanalyses were compared with radiosonde soundings.

3.1 ATMOSPHERIC REANALYSES

Reanalyses are not direct observations of the state of the atmosphere, but instead are products of numerical weather prediction models. In contrast to climate model products, reanalyses attempt to simulate the actual states of the atmosphere, utilizing meteorological observations including satellite observations. In reanalyses, the observations are optimally associated with knowledge about the physics and dynamics of the atmosphere through a numerical weather prediction model to provide a comprehensive picture of the state of the atmosphere. In the studies comprising the thesis, three global reanalysis products are were utilized: the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis-Interim (ERA-Interim) (Dee et al. 2011), fifth-generation ERA (ERA5) (Hersbach et al. 2020) and the Japanese 55-year Reanalysis (JRA-55) (Kobayashi et al. 2015). The first two are produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) and the last is produced by the Japan Meteorological Agency (JMA). The specifications of these reanalyses are presented in Table 1.

Table 1. Specifications of reanalyses.

<table>
<thead>
<tr>
<th>Time/Space Specifications</th>
<th>ERA5</th>
<th>ERA-Interim</th>
<th>JRA-55</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temporal coverage</td>
<td>1950 to present</td>
<td>1979 - 2019</td>
<td>1958 to present</td>
</tr>
<tr>
<td>Temporal resolution</td>
<td>1h</td>
<td>6h</td>
<td>6h</td>
</tr>
<tr>
<td>Horizontal resolution</td>
<td>0.25° x 0.25°</td>
<td>0.75° x 0.75°</td>
<td>0.56° x 0.56°</td>
</tr>
<tr>
<td>Number of levels in the vertical</td>
<td>137</td>
<td>60</td>
<td>60</td>
</tr>
</tbody>
</table>
Reanalyses provide remarkable benefits for climatological studies in comparison with conventional observations. Firstly, they provide three-dimensional gridded presentations of the state of the atmosphere, with full spatial coverage, even over areas where observational networks are sparse. Secondly, they provide reasonably long and complete time series, mostly without spurious trends, since the reanalysis system, i.e. numerical weather prediction model and data assimilation system, is frozen throughout the entire time-series interval. However, evolution in observational networks, especially launches of new satellite instruments, may affect the accuracy of the reanalysis products and trends based on them. Thirdly, reanalyses are less affected by the weaknesses of a single instrument than many observations based only on a single instrument. For example, satellite-based observation of atmospheric profiles have problems with clouds and snow surfaces in the polar regions. However, few in-situ observations are available in the polar regions for anchoring reanalyses on the true state of the atmosphere. In addition, description of physical processes in reanalyses may possibly not be optimal for often very stably-stratified polar atmospheres. In conclusion, serious errors may occur in reanalyses, and therefore it is important to compare reanalyses with high-quality observations, which preferably are not assimilated into the reanalyses. All three reanalyses used in the studies performed reasonably well in presenting atmospheric conditions at high latitudes (Gossart et al. 2019, Graham et al. 2019a, 2019b, Jonassen et al. 2019, Wang et al. 2019).

The uncertainty in reanalyses, as well as that in numerical weather prediction models in general, is caused by the uncertainty in the estimated initial state of the model and inaccuracies in model physics and dynamics. The initial state, i.e. analysis, is based on a short forecast from previous analysis termed the background field, which is corrected by observations. The uncertainty in forecasting tends to increase during forward time integration, thus the most accurate description of atmospheric conditions can be obtained by utilizing analysis. However, analyses are affected by errors in the background field, which are caused by errors in the previous analysis and incomplete model physics and dynamics, and by the uncertainty in observations. Some variables in the reanalysis products are more vulnerable to model errors than others. For example, evaporation and precipitation in reanalyses are only products of model physics, whereas moisture transport is constrained by the specific humidity and winds observed. Therefore, the ability of reanalyses to simulate moisture conditions can be assessed by analysing the balance between the physical processes of evaporation and precipitation, and convergence of moisture transport. A well-balanced moisture budget typically indicates accurate presentations of the physical processes in reanalyses. Several studies (Dufour et al. 2016, 2019, IV) have shown
that the atmospheric water-vapour budgets in all three reanalyses utilized in the studies are reasonably well in balance in the polar regions. This is indicated by a small residual between the moisture-transport convergence and the net effect of evaporation and precipitation. However, a small residual between moisture-transport convergence and moisture tendency due to parametrized physical processes does not necessarily mean high accuracy in reanalysis. Since the residual is essentially caused by corrections made to the background field, lack of accurate observations or small weight of observations in assimilation may result in a small residual even in the case of inaccurate analysis.

3.2 Radiosonde soundings

Meteorological observational networks are sparse in the polar regions, due to the remote location and harsh environment. Permanent sounding stations are located only in land areas, since the infrastructure needed for a station in practice does not allow for establishing permanent sounding stations on unsteady sea ice. Sounding stations in Antarctica are usually located in the coastal zone, because inland stations in Antarctica are logistically challenging to reach. Therefore, most observations from polar oceans are from relative short expeditions, most of which have taken place during summer. Wintertime data are available mostly from rare drifting stations (Serreze et al. 1992, Graham et al. 2019). In conclusion, radiosonde observations done alone are not able to provide satisfying spatial and temporal coverage of atmospheric variables in the polar regions.

Radiosonde soundings are considered as accurate and reliable observations of atmospheric conditions and therefore have been used as references for other profile observations. Radiosonde observations are also an important data source for numerical weather prediction models as well as reanalyses, because they anchor the model state to the true state of the atmosphere (Naakka et al. 2019). However, especially under cold conditions, the accuracy of radiosonde measurements varies with the radiosonde type (Ingleby 2017). Under cold conditions, humidity measurements are challenging, due to the small amount of atmospheric water vapour available. In addition, contamination of ice on the humidity sensor when a sonde ascends through a cloud with supercooled liquid water causes serious errors in humidity observations (Ingleby 2017). The radiosonde observations used in this thesis were downloaded from the Integrated Global Radiosonde Archive (IGRA) and have gone through a quality check that removes distinct errors in the observations (Durre et. al. 2006, Durre and Yin 2008).
4. SUMMARY OF RESULTS

This section summarizes the results of the thesis.

Figure 8. Plot in the middle: winter, DJF, mean vertically-integrated water vapour (IWV) (kg m\(^{-2}\), presented in colour); winter mean of vertically-integrated horizontal moisture transport (kg m\(^{-1}\) s\(^{-1}\), presented by vectors). Cross sections: winter mean specific humidity (g kg\(^{-1}\), in colour, y-axis shows the pressure). The figure is based on fifth-generation European Reanalysis (ERA5) reanalysis data from the period 2000 – 2020.
Figure 9. Same as Figure 8, but for summer, JJA.
4.1 Specific Humidity Inversions in the Arctic

Figure 10. Winter (DJF) and summer (JJA) means of specific humidity inversion occurrence between the level of 800 hPa and the surface from European Reanalysis-Interim (ERA-Interim), Japanese 55-year Reanalysis (JRA-55) and radiosoundings (circles). The figure is from (I) Naakka et al. 2018, © American Meteorological Society. Used with permission.
In the Arctic, moisture profiles are strongly affected by air-mass transformation, due to radiative cooling resulting in formation of specific humidity inversions (Curry 1983, Pithan et al. 2014). Specific humidity inversions frequently occur in the Arctic, based on ERA-Interim and JRA-55 reanalysis (Figure 10). In winter, radiative cooling leads to formation of specific humidity inversions (Curry 1983, Pithan et al. 2014). Under cloud-free conditions, the strongest radiative cooling occurs at Earth’s surface, leading to formation of surface-based specific humidity inversions due to moisture condensation onto the surface, but when clouds are present, the strongest radiative cooling occurs at the top of low-level clouds, and an elevated specific humidity inversion is formed, due to moisture condensation in the clouds. However, specific humidity inversions also occur at higher altitudes in the troposphere. Specific humidity inversions were divided into two categories, based on the altitude of their occurrence (I). At high altitudes above the 800-hPa level, the spatial and seasonal occurrence of specific humidity inversions varies only slightly. These specific humidity inversions typically occur without temperature inversions and under the conditions at which relative humidity is below the saturation point. Hence, formation of these inversions is probably linked with vertically differential moisture advection, so that a moist air mass has been advected above a dry air mass.

Below the 800-hPa level, the formation of specific humidity inversions is linked with surface conditions, and therefore their spatial and seasonal occurrence and strength vary widely. In winter, specific humidity inversions are most frequent over continents and sea ice, where the frequency of occurrence exceeds 90% (I). In summer, specific humidity inversions are most frequent over the Arctic Ocean, especially near the coast. Below the 800-hPa level, specific humidity inversions often occur with temperature inversions, agreeing with previous studies (Andreas et al. 2002, Persson et al. 2002, Vihma et al. 2011, Tjernström et al. 2012, Sotiropoulou et al. 2016) and under conditions of high relative humidity. The latter suggests that moisture condensation strongly contributes to the formation of specific humidity inversions below the 800-hPa level. In winter, formation of specific humidity inversions is closely associated with formation of polar air masses as a result of radiative cooling of the surface. Radiative cooling leads to formation of a temperature inversion, and after reaching saturated conditions, radiative cooling results in moisture condensation and formation of specific humidity inversions (Curry 1983). In summer, formation of specific humidity inversions is associated with advection of warm-air masses from a continent to over the Arctic Ocean, where cooling and moisture condensation near the surface generate strong specific humidity inversions over coastal seas. In summer, near-surface air masses rapidly cool over a cold sea surface, due to downward sensible heat flux, resulting in moisture
condensation and the formation of strong specific humidity inversions, as also shown in other studies (Tjernström et al. 2015, 2019). Farther from the coast, specific humidity inversions become gradually weaker as the air mass begins to dry at the level of specific humidity maximum, due to moisture condensation in the clouds.

The effects of moisture condensation and vertically differential moisture advection on the formation of specific humidity inversions cannot be totally separated, because on one hand, summer advection of warm-air masses over a remarkably colder sea surface leads to air-mass cooling and condensation of moisture at the bottom of the advected air mass layer. On the other hand, advection of warm, moist air above a very stable boundary layer also contributes to occurrence of specific humidity inversions in winter (Devasthale et al. 2011 Nygård et al. 2013). However, the main factors behind the formation of specific humidity inversions are different between summer and winter. In summer, advection of warm, moist air is vital for formation of specific humidity inversions, because the surface temperature and often the dew-point temperature over melting sea ice are fixed at or near the melting point of snow and sea ice (0 °C) or the freezing point of ocean water (-1.8 °C), due to snow/ice melt and the large heat capacity of the ocean. Hence, specific humidity differences across the inversion are formed or maintained with moisture advection. In contrast, specific humidity inversions in winter, often occur under conditions of weak advection and extremely low surface-specific humidity. In these cases, the formation of specific humidity inversions is linked with radiative cooling and moisture condensation, which are able to form and maintain a specific humidity gradient between the surface and atmosphere, even without moisture advection. The contribution of advection to formation of near-surface specific humidity inversions is most unambiguous over sloping areas, where katabatic winds bring dry (in terms of absolute instead of relative humidity, hereafter absolutely dry), cold air from higher-elevation areas to coastal sites.

Specific humidity inversions are stronger in summer than in winter, due to the much higher water-vapour content of the air in summer (Figures 8 and 9; I). In winter, the relatively strongest specific humidity inversions occur in northern Canada and Siberia, which are the areas most remote from the moisture sources in the northern polar region and where moisture transport is weakest (Figure 8; II). Since air-mass transformation is a relatively slow process (Curry 1983, Pithan et al. 2014), it is not surprising that the relatively strongest inversions are located in areas most remote from the tracks of extratropical cyclones.

The reanalyses and radiosonde data largely agree on the spatial distribution and seasonal cycle of specific humidity inversion occurrence (Figure 10; I). However,
the specific humidity inversions in reanalyses are often weaker than in radiosonde observations, while above the 800-hPa level, the occurrence of specific humidity inversions is lower in the reanalyses than in the radiosonde observations, probably because the reanalyses are not capable of presenting all shallow inversions in the middle and upper troposphere, due to their coarse vertical resolution.

4.2 MOISTURE TRANSPORT TO THE ARCTIC

The spatial distribution of vertically-integrated moisture transport and vertical profiles of moisture transport were examined, based on ERA-Interim reanalyses (II). The results showed that moisture transport typically is vertically coherent, meaning that temporal anomalies in moisture transport at different levels in a vertical profile are linearly associated (II). Thus, an anomaly in vertically-integrated moisture transport is typically constituted by analogous anomalies throughout the troposphere. Hence, variations in vertically-integrated moisture transport characterize variations in moisture transport throughout the troposphere. However, the strength of moisture transport varies widely in the vertical. In the lower troposphere, specific humidity is often high, but winds are weaker than in the upper troposphere. Accordingly, weak winds limit moisture transport in the lower troposphere, whereas in the middle and upper troposphere the scarcity of water vapour strongly limits moisture transport.

Meridional moisture transport is a two-way process including moisture transport from the midlatitudes to the Arctic and vice versa. The net northward moisture transport represents only a minor part of the total moisture exchange between the midlatitudes and the Arctic (Figure 11; II). The contrast in moisture content between the southward- and northward-transported air masses as well as the consistence that the northward have higher specific humidity than the southward air masses affect the efficiency of net meridional moisture transport (Dufour et al. 2016). We showed, especially near the surface, that southward moisture transport is similar to northward moisture transport (II). Hence, the vertical maximum of net-moisture transport is located between the 800- and 850-hPa levels, even though the vertical maxima of southward or northward moisture-transport components are located in the layer 900–950 hPa (Figure 11). Near the surface, the southward and northward moisture transports are similar, resulting in only a small northward net-moisture transport. This is at least partly associated with air-mass transformation. The near-surface air mass reacts more rapidly to changes in surface conditions than the midtropospheric air mass. Winter advection of cold, dry air above the open ocean rapidly increases air-specific humidity in the lowermost layer near the surface, due to strong evaporation, while advection of a warm, moist air mass over snow or ice surfaces often results in cooling and moisture condensation.
Figure 11. Vertical profile of contribution of each moisture-transport strength interval on the net meridional moisture transport at 60 °N (a) and 70 °N (c) and vertical profile of meridional net-moisture transport at 60 °N (b) and 70 °N (d). The strength of the moisture transport was calculated individually at each grid point. When the strength is positive (bars on the right-hand side in (a, c)), moisture transport is northwards, and when negative (bars on the left-hand side in (a, c)) moisture transport is southwards. The figures are based on 6-hourly analyses of European Reanalysis-Interim (ERA-Interim) from years 2003–2014. The figure is from (II) Naakka et al., 2019, © Royal Meteorological Society. Used with permission.
The ratio between total and net-moisture transport varies widely (II). For example, in the Atlantic sector where net-moisture transport towards the Arctic is largest, the northward and southward moisture transports are both large, whereas on the western side of Greenland, a relatively strong meridional moisture transport is caused by a weak, but rather permanent northward moisture transport due to prevailing northward winds. In fact, the atmospheric moisture content on the western side of Greenland is low, and the northward moisture transport from there does not contribute to increasing atmospheric moisture in the central Arctic. The strength of total moisture transport is associated with the atmospheric moisture content and thus undergoes a strong seasonal cycle, peaking in late summer, whereas the seasonal cycle of net-moisture transport is remarkably smaller and undergoes its annual maximum later in autumn (II).

The moisture transport was divided into proportions of mean meridional circulation, stationary eddies and transient eddies (II). The results (II) confirmed the findings of previous studies (Serreze et al. 1995, Groves and Francis 2002, Oshima and Yamazaki 2004, 2006, Jakobson and Vihma 2010, Liu and Barnes 2015, Dufour et al. 2016) that the largest part of net-moisture transport is caused by transient eddies, followed by stationary eddy transport with a small contribution. In contrast, the mean meridional moisture transport associated with the structure of polar cells is southwards, due to the southward transport of moist air masses in the lower troposphere and northward transport of dry air masses in the mid- and upper troposphere. Even though net meridional moisture transport is mainly due to transient eddy moisture transport, the spatial distribution of net meridional moisture transport is strongly associated with stationary features of the circulation patterns (II). Stationary waves interact with transient cyclones, on one hand by steering the tracks of transient cyclones. On the other hand, however, stationary wave patterns are also affect transient cyclones. However, the locations of mountain ranges strongly influence both stationary wave structure as well as northward moisture transport. The strongest northward moisture transport occurs on the eastern sides of the Atlantic and Pacific Oceans, west of the mountain ranges in Scandinavia and in the western part of North America, agreeing with the results of previous studies (Cullather et al. 2000, Groves and Francis 2002, Bengtsson et al. 2011, Dufour et al. 2016).

Interannual variations in stationary eddy moisture transport clearly affect similar variations in net-moisture transport in winter and spring. In addition, interannual variations in moisture transport caused by strong moisture-transport events are positively correlated with stationary eddy moisture transport (II). These results suggest that similar circulation patterns that favour stationary eddy moisture
transport may also favour the occurrence of strong moisture transport events, resulting in extensive net-moisture transport into the Arctic. These results agree with studies by Woods et al. (2013) and Liu and Barnes (2015), which show that circulation patterns with strong meridional flow are a prerequisite for strong moisture-transport events. However, the seasonal cycle in moisture transport is strongly associated with the seasonal cycle in transient eddy moisture transport.

4.3 LONG TERM CHANGES IN MOISTURE TRANSPORT TO THE ARCTIC

Moisture transport not only shapes moisture conditions at short timescales, but is also the main reason for long-term variations in regional moisture conditions. Several studies have confirmed that the atmospheric moisture content in the northern polar region has increased during recent decades (Serreze et al. 2012, Dufour et al. 2016, Rinke et al. 2019). However, the increase in moisture content has not been evenly distributed (Rinke et al. 2019). The associations between long-term trends in atmospheric moisture content and moisture transport were evaluated for the 40-year period 1979 – 2018, based on ERA5 reanalysis (III).

The trends in moisture transport are closely associated with changes in atmospheric circulation in the low troposphere. These changes in moisture-transport direction are often determined by the responses in geostrophic winds to changes in mean sea-level pressure. For example, an increase in mean sea-level pressure in northern Russia in winter during 1979–2018 increased northward and eastward moisture transport in the Barents Sea and the eastern Arctic Ocean (Figure 12), while easterly moisture transport decreased on the southern side of the maximum positive mean sea-level pressure trend in western Russia. Overall, during both the entire 40-year period and the most recent 20-year period, changes in low tropospheric circulation have contributed to increases in northward moisture transport from the Nordic seas to the central Arctic.
Figure 12. Linear trends in vertically-integrated water-vapour content (kg m\(^{-2}\) per decade, presented in colour) and linear trends in vertically-integrated horizontal moisture transport (kg m\(^{-1}\) s\(^{-1}\) per decade, presented by vectors) in fifth-generation European Reanalysis (ERA5) for winter, spring, summer and autumn during 1979–2018. The red contours indicate positive (and zero) and the blue contours indicate negative linear trends in mean sea-level pressure at 0.5-hPa per decade intervals. The figure is redrawn from (III) Nygård et al. 2020, © American Meteorological Society. Used with permission.
Comparison of trends in moisture transport and atmospheric moisture content showed that the moisture content has increased in those areas that have been affected by increased moisture transport from climatologically moist regions (III). On one hand, increased atmospheric moisture content is a result of intensified moisture transport but, on the other, higher moisture content enables stronger moisture transport. However, increases in moisture content are not able to affect directional changes in moisture transport. Since directional changes in low-tropospheric winds are associated with changes in atmospheric moisture content, it allows us to interpret that moisture-transport trends, as a result of changes in low-tropospheric circulation, are responsible for many of the regional trends in atmospheric moisture content in the Arctic.

Changes in local evaporation also contribute to trends in atmospheric moisture content. Our results showed that trends in evaporation are mostly positive over the entire polar region north of latitude 60 °N and most are even statistically significant during the warm season (III). During cold seasons, the strongest trends in evaporation occur in the vicinity of the sea-ice margin and are probably associated with diminishing of sea ice. Decrease in sea-ice concentration intensifies evaporation, especially when cold air masses from sea ice are advected over the open ocean. Further examination of the effects of sea-ice retreat on evaporation showed that evaporation has increased in the area where sea ice has recently disappeared, but decreased farther south from the sea-ice margin, because a prerequisite for efficient evaporation is the occurrence of absolutely dry air over the surface. Sea-ice retreat caused a stepwise increase in evaporation when the sea ice disappeared, but farther south of the sea-ice margin evaporation became gradually weaker (III). Analyses of the interannual variation in mean intensity of moisture transport and atmospheric moisture content showed that moisture transport and atmospheric moisture content are strongly correlated throughout the circumpolar Arctic, whereas evaporation is typically weak when the atmospheric moisture content is high, except near the sea-ice margin (III). Since a high atmospheric moisture content efficiently limits local evaporation, strong evaporation occurs over the open sea near the sea-ice margin, when air flows from the sea ice to the open ocean during cold seasons. This typically occurs when air flows from the Pole towards lower latitudes in the Arctic (Nygård et al. 2019). This evaporated moisture usually does not contribute to atmospheric moisture content in the central Arctic, because much of it is transported away from the Arctic by southward advection (Nygård et al. 2019). In conclusion, long-term changes in moisture transport due to variations in circulation patterns are responsible for most of the regional changes in atmospheric moisture content, whereas local evaporation probably plays only a minor role in regional patterns of moistening in
the Arctic. This is natural, because moisture transport is an important source for atmospheric moisture in the polar regions, and efficient evaporation is often possible only after formation of cold, dry polar air masses.

4.4 MOISTURE CONDITIONS IN THE ANTARCTIC

![Figure 13](image-url)

**Figure 13.** Plot in the middle: winter, April to October, mean vertically-integrated water vapour (IWV) (kg m\(^{-2}\), presented in colour) and winter mean of vertically-integrated horizontal moisture transport (kg m\(^{-1}\) s\(^{-1}\), presented as vectors). Cross sections: winter mean specific humidity (g kg\(^{-1}\), in colour) and winter mean meridional (kg m\(^{-1}\) hPa\(^{-1}\) s\(^{-1}\)) and vertical (kg m\(^{2}\) s\(^{-1}\)) moisture flux (in vectors; y-axis shows the pressure). The thin magenta lines indicate the locations of turning points of mean vertical moisture transport. The figure is based on fifth-generation European Reanalysis (ERA5) reanalyses and is taken from (IV) Naakka et al. 2021, © American Meteorological Society. Used with permission.
Figure 14. Same as Figure 13 but for summer, November to March. The figure is from (IV) Naakka et al., 2021, © American Meteorological Society. Used with permission.

Previous studies have shown that the atmospheric moisture content in large parts of Antarctica is extremely low (Tietäväinen and Vihma 2008) and that weather conditions over the East Antarctic Plateau are characterized by strong temperature inversions, high relative humidity and weak clear-sky precipitation (King and Turner 1997). On slopes, low relative humidity resulting from katabatic winds allows evaporation of moisture from precipitation particles and causes frequent occurrence of virga events, i.e. precipitation that does not reach the surface (Vignon et al. 2019). However, atmospheric moisture conditions in the Antarctic have not been as comprehensively studied as those in the Arctic. Observation networks in the Antarctic are even sparser than in the Arctic, resulting in challenges to the study of moisture climates in the Antarctic. Recent development of reanalyses, especially
assimilation of satellite observations, has improved skills of reanalyses at replicating atmospheric conditions in remote areas. In the fourth article Atmospheric moisture conditions and moisture processes in the Antarctic were examined, based on ERA5 reanalysis.

Figure 15. Schematic illustration of processes affecting moisture conditions in the Antarctic. The thin blue vectors represent moisture transport along the air flow, and the thick vectors represent surface moisture fluxes. The yellow area represents low relative humidity, whereas the cyan area represents high relative humidity and saturated conditions. The thin black lines represent isentropes. The figure is from Naakka et al. 2021, © American Meteorological Society. Used with permission.

Regional differences in moisture conditions in the Antarctic are vast, and various physical processes dominate moisture conditions in different areas (IV). Radiative cooling characterizes conditions over the continent, leading to formation of strong surface-based temperature inversions and to simultaneous formation of almost permanent specific humidity inversions due to moisture condensation (IV). Even though vertical motion of air in the inner continent is mostly downwards, the results indicated that radiative cooling keeps relative humidity close to the saturation point (IV), which is in line with observations of clear-sky precipitation (King and Turner 1997). Since near-surface temperatures are extremely low, condensation of moisture causes formation of absolutely extremely dry near-surface air masses. Absolutely dry air masses are transported downwards from the plateau by katabatic winds. On the slope, downward motion of air leads to adiabatic warming, which decreases
relative humidity and strengthens downward sensible heat flux (IV). Heating of the surface together with the very dry air mass above enables surface evaporation, which decreases the occurrence of surface-based specific humidity inversions on the slopes. However, elevated specific humidity inversions are common, and formation of these inversions is probably associated with moisture advection.

Analyses indicated that the amount of evaporation in Antarctica is generally low. Even though evaporation on the slope is common, the amount of evaporated water vapour is small, due to low temperatures. Over high-elevation areas, the turbulent surface moisture flux is downwards, even in summer, and the mean annual evaporation is practically zero throughout the East Antarctic Plateau. Due to low evaporation, precipitation is several times higher than evaporation in Antarctica. Most of the precipitation occurs on the coastal slopes, and the amount of precipitation steeply decreases towards the inland.

Sea ice in the Southern Ocean undergoes extensive seasonal evolution. In winter, sea ice extends as far north as latitude 60 °S, but in summer sea ice does not even form a continuous belt around Antarctica. The sea-ice region is a transition zone between continental and marine moisture conditions. In winter, near-surface conditions are affected by cold, dry continental air masses. In addition, sea ice acts as an insulating layer between the ocean and atmosphere. Hence, specific humidity is often rather low over sea ice, which allows strong evaporation from leads and polynyas. This probably decreases the occurrence of surface-based specific humidity inversions. However, the results showed that elevated specific humidity inversions are common, which is probably associated with advection of dry continental air near the surface and moister marine air above it (IV). In summer, solar radiation heats the surface, increasing evaporation, which erodes specific humidity inversions.

The influence of transient cyclones on moisture conditions increases from the continent towards the ocean. The cyclones mostly determine the direction of meridional moisture transport over the ocean, since the mean flow is mostly zonal (IV). We showed that the direction of moisture transport notably affects moisture conditions over the ocean and on slopes of the Antarctic ice sheet. Southward moisture transport increases the vertically-integrated water-vapour content, which weakens evaporation. In addition, the cloud water content is often larger when the air mass originates from lower latitudes than from the continent or sea ice, except near the sea-ice margin in winter. Northward moisture transport brings cold, dry air from the sea-ice zone over the ocean, especially in winter when the contrasts between the moisture conditions on sea ice and over the open ocean are largest. Advection of cold, dry air over the open ocean strengthens vertical mixing and evaporation. This
results in convection over the open ocean, causing condensation of moisture and formation of clouds in the upper part of the well-mixed boundary layer. This causes positive cloud water anomalies in the upper part of the boundary layer downwind of the sea-ice margin during southerly winds. However, although the local evaporation over the open ocean is clearly larger than over the continent and sea ice, it is only about half as large as the amount of precipitation. Hence, moisture transport from the lower latitudes is also an important moisture source for the atmosphere over the open ocean.
5. CONCLUSIONS

The topic of this thesis is atmospheric moisture in the polar regions. In the Arctic and Antarctic, the surface-atmosphere system loses heat, due to negative net radiation, which strongly shapes the moisture environment. The negative net radiation budget of Earth’s surface during cold seasons causes cooling of the near-surface air mass, which results in condensation of moisture either on the surface or in clouds and fog near the surface, affecting the vertical profile of atmospheric water vapour. The results confirm that moisture condensation due to radiative cooling is the main reason for formation of near-surface specific humidity inversions (I). This is indicated by their frequent co-occurrence with temperature inversions and occurrence in conditions of high relative humidity (I).

Evaporation in the polar regions is mostly small or even negative, except on snow-free continents in summer and over the open ocean in winter. Hence, atmospheric transport is an important source of air moisture. Moisture transport affects the vertical structure of specific humidity and maintains atmospheric moisture without a notable contribution of local surface evaporation. In the Arctic, the net northward moisture transport is maximal between the 800- and 850-hPa levels, which is above the layer where specific humidity inversions are most common (I, II). However, both the northward and southward components of moisture transport are on average strongest below the 900-hPa level (II). Near the surface, the northward moisture transport is mostly balanced by southward transport, suggesting that the humidity differences between the northward and southward moisture transports are small, because the frequencies of occurrence of the northward and southward transports are similar (II).

We conclude that interactions between the atmosphere below approximately the 800-hPa level and Earth's surface clearly affect the profiles of atmospheric moisture and moisture transport, whereas above the 800-hPa level, the interactions of atmospheric moisture conditions with the surface become weaker (I, II). This suggests that in the Arctic below the 800-hPa level, physical processes associated with atmospheric moisture, such as cloud formation, are strongly associated with surface properties, while changes in surface properties, e.g. decline in sea ice, probably modify these processes substantially. However, above the 800-hPa level the link between surface properties and atmospheric moisture becomes weaker, and at those altitudes large-scale circulation is probably the factor mainly responsible for variations in atmospheric moisture.
Moisture transport shapes the spatial distribution of atmospheric moisture, such that atmospheric moisture content is greatest in those areas where poleward moisture transport is strongest. Moisture transport affects not only the short-term (from a couple of days to a week) variations in atmospheric moisture and moisture processes, but also long-term (decades) regional changes (III). In addition, moisture exchange between the surface and atmosphere and cloud formation also interact with atmospheric moisture content. Therefore, changes in evaporation may also affect long-term trends in atmospheric moisture content. However, surface evaporation is strongly constrained by the water-vapour content of the air mass above (Nygård et al. 2019, III). A prerequisite for efficient evaporation is advection of absolutely dry air over a warm, moist surface. In the Arctic, this typically occurs when a cold air mass is advected over the open ocean. We showed that large changes in surface evaporation have only occurred in those areas where sea ice has retreated. In these areas, evaporation has increased, which is mostly associated with advection of cold, dry air masses over a warm sea surface, but at the same time evaporation has decreased in those areas farther south from the sea-ice margin (III). Accordingly, the area of strong evaporation has been shifted northwards with the retreating sea ice (III). In conclusion, moisture transport plays a dominant role in regional long-term trends in atmospheric moisture content (III). In contrast, the role of local evaporation in regional trends of Arctic atmospheric moisture content was not as clear, and the changes in evaporation were strongly associated with humidity differences between the surface and the air mass above. Sea-ice decline potentially increases the area where extensive evaporation is possible, but at the same time decreases the area available for formation of cold, dry air masses. In the light of these results, these two factors have contrasting effects on evaporation, even though both effects contribute to increase in atmospheric moisture in the Arctic, as the area of efficient evaporation moves northward.

We also focused on moisture conditions in the Antarctic (IV). A comprehensive climatology of atmospheric moisture, moisture transport and physical processes associated with atmospheric moisture was formed. Even though the boundary conditions resulting from large-scale circulation are rather similar to those in the Arctic, a highly different geographic environment results in vast differences in moisture conditions between the Antarctic and the Arctic. This study provided a basis for more detailed studies of Antarctic atmospheric moisture and clouds.

In this thesis, we examined the distribution of atmospheric moisture content and the processes affecting it. Moisture transport is part of the energy transport from the Tropics to the Poles, but at high latitudes a more important aspect is the effect of moisture on radiative heat fluxes. For example, radiative fluxes play an important
role in the current rapid climate warming in the Arctic (Cohen et al. 2014). Clouds have an especially large effect on radiative fluxes. In this thesis, however, clouds received limited regard, because we mostly focused on atmospheric water-vapour content. However, the vertical structure of atmospheric water vapour and moisture transport do affect the formation of clouds (Solomon et al. 2011, 2014, Savre et al. 2015). In warmer climates, the sea-ice coverage over the Arctic Ocean declines, which decreases the area available for the formation of cold polar air masses and increases the area available for warm, wet surfaces in the Arctic. This probably increases the atmospheric moisture content and also affects cloudiness. Understanding of moisture conditions and the physical processes affecting them may also shed the light required for a better understanding of clouds. However, further research is needed to more fully understand the interaction of clouds and atmospheric moisture and how the resulting changes in atmospheric moisture affect cloudiness in the current warming global climate.
REFERENCES


Devasthale, A., U. Willén, K. G. Karlsson, and C. G. Jones, 2010, Quantifying the clear-sky temperature inversion frequency and strength over the Arctic Ocean during summer and winter seasons from AIRS profiles. Atmospheric Chemistry and Physics, 10(12), 5565-5572, https://doi.org/10.5194/acp-10-5565-2010

Devasthale, A., J. Sedlar, and M. Tjernström, 2011, Characteristics of water-vapour inversions observed over the Arctic by Atmospheric Infrared Sounder (AIRS) and radiosondes. Atmospheric Chemistry and Physics, 11(18), 9813-9823, https://doi.org/10.5194/acp-11-9813-2011


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Oshima, K. and K. Yamazaki, 2004, Seasonal variation of moisture transport in polar regions and the relation with annular modes. Polar Meteorology and Glaciology, 18, 30-53


Vignon, É., O. Traullé, and A. Berne, 2019, On the fine vertical structure of the low troposphere over the coastal margins of East Antarctica. Atmospheric Chemistry and Physics, 19(7), 4659 – 4683, https://doi.org/10.5194/acp-19-4659-2019


Arctic Humidity Inversions: Climatology and Processes

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ABSTRACT

The occurrence and characteristics of Arctic specific humidity inversions (SHIs) were examined on the basis of two reanalyses (ERA-Interim and JRA-55) and radiosonde sounding data from 2003 to 2014. Based on physical properties, the SHIs were divided into two main categories: SHIs below and above the 800-hPa level. Above the 800-hPa level, SHIs occurred simultaneously with relative humidity inversions and without the presence of a temperature inversion; these SHIs were probably formed when a moist air mass was advected over a dry air mass. SHIs below the 800-hPa level occurred simultaneously with temperature inversions in conditions of high relative humidity, which suggests that condensation had an important role in SHI formation. Below the 800-hPa level, SHI occurrence had a large seasonal and spatial variation, which depended on the surface heat budget. In winter, most SHIs were formed because of surface radiative cooling, and the occurrence of SHIs was high (even exceeding 90% of the time) on continents and over the ice-covered Arctic Ocean. In summer, the occurrence of SHIs was highest (70%–90%) over the coastal Arctic Ocean, where SHIs were generated by warm and moist air advection over a cold sea surface. In the reanalyses, the strongest SHIs occurred in summer over the Arctic Ocean. The comparisons between radiosonde soundings and the reanalyses showed that the main features of the seasonal and spatial variation of SHI occurrence and SHI strength were well represented in the reanalyses, but SHI strength was underestimated.

1. Introduction

Water vapor is an important component of the Arctic climate system (Serreze et al. 2006; Francis et al. 2009; Vihma et al. 2016). Water vapor is present in the Arctic atmosphere because of local evaporation and evapotranspiration (Bring et al. 2016; Boisvert and Stroeve 2015) and transport from lower latitudes, most of which is carried by synoptic-scale cyclones (Jakobson and Vihma 2010; Dufour et al. 2016). Under clear skies, water vapor enhances the atmospheric emissivity for longwave radiation, seen as increased downward longwave radiation at Earth’s surface (Prata 1996; Zhang et al. 2001; Devasthale et al. 2011). The increase in the atmospheric emissivity is even larger when water vapor condenses into cloud or fog droplets (Shupe and Intrieri 2004). Arctic clouds have a warming effect on the surface during most of the year because their effect of increasing the downward longwave radiation dominates their effect of reducing the net solar radiation over high-albedo snow and ice surfaces. In summer, however, clouds typically have a cooling effect on surface types with a lower albedo, such as the open sea, melting sea ice, and ground (Intrieri et al. 2002a; Shupe and Intrieri 2004).

The altitude where water vapor condenses to clouds or fog depends on the vertical profiles of air temperature and specific humidity. On the global scale, a typical situation is that both air temperature and specific humidity decrease with height. In the Arctic, however, the lower troposphere often includes inversion layers where the temperature or specific humidity or both increase with height (Devasthale et al. 2011; Sedlar et al. 2012; Shupe et al. 2013; Nygård et al. 2014; Brunke et al. 2015). A specific humidity inversion (SHI) can contribute to cloud formation and maintenance. An SHI above a cloud layer can provide an adequate moisture source for Arctic stratocumulus clouds, and this feature probably allows for the maintenance of extensive cloudiness in summer (Solomon et al. 2011, 2014; Sedlar et al. 2012; Sedlar 2014). If a cloud is decoupled from Earth’s surface, an SHI may be the only moisture source for the cloud (Sedlar et al. 2012; Savré et al. 2015).

SHIs often occur simultaneously with temperature inversions (TIs) (Andreas et al. 2002; Persson et al. 2002; Vihma et al. 2011; Tjernström et al. 2012;
Sotiropoulou et al. 2016). In winter, surface cooling related to the negative net radiation leads to the formation of TIs and increases the relative humidity, and further leads to the formation of SHIs via moisture condensation (Curry 1983). Therefore, the properties of SHIs and TIs are partly connected. Devasthale et al. (2011) and Nygård et al. (2014) found that the strength of SHIs and TIs are often linked. Nygård et al. (2014) also found that the strength, base height, and depth of the strongest SHI and TI in each sounding profile were correlated. Another mechanism affecting the formation of SHIs is humidity advection. Nygård et al. (2014) showed that approximately half of all Arctic SHIs occurred without the presence of TIs, and Nygård et al. (2013) showed that vertical changes in humidity advection, and especially the near-surface advection of dry air by katabatic winds, was an important factor in generating SHIs in the Antarctic. Geographical conditions on the coast of Antarctica are unique and, therefore, it is probable that katabatic processes do not play a major role for Arctic SHIs, except in regions with large slopes, as in Greenland. Brunke et al. (2015) showed that in the Arctic, humidity advection has the largest effect on specific humidity tendencies in SHI layers, whereas moist physics (condensation and evaporation) has a much smaller contribution to the tendencies. However, Brunke et al. (2015) showed that humidity advection has both weakening and strengthening effects on SHI, and they suggested that humidity advection may not be as important a process for the formation of SHIs as the examination of specific humidity tendencies in SHI layers implied.

SHIs typically occur simultaneously on multiple levels (Devasthale et al. 2011; Nygård et al. 2013, 2014). Based on satellite retrievals of the Atmospheric Infrared Sounder (AIRS), Devasthale et al. (2011) showed that in winter under clear skies the occurrence of SHIs exceeds 50% in most of the continental Arctic and the Arctic Ocean, and an even higher SHI occurrence was found in radiosonde sounding data (Devasthale et al. 2011; Nygård et al. 2014). SHI occurrence mostly exceeded 90% at the radiosonde sounding stations north of 65°N (Nygård et al. 2014). Brunke et al. (2015) showed that an SHI is found in wintertime monthly mean specific humidity profiles over the majority of the Arctic based on reanalyses. In summer, the reported occurrence of SHIs has varied considerably between studies. Devasthale et al. (2011) showed a very low, approximately 10%, SHI occurrence in the Arctic in summer based on AIRS satellite retrievals, but notably higher SHI occurrence in the radiosonde sounding observations. Nygård et al. (2014) reported that SHI occurrence in summer was high, on average only 10% lower than in winter. It is noteworthy that the results of AIRS satellite retrievals (Devasthale et al. 2011) were for clear-sky conditions only, and therefore, they were not very representative for the generally cloudy conditions over the Arctic Ocean in summer. In addition, Gettelman et al. (2006) showed that AIRS retrievals are not able to capture very fine details of specific humidity profiles, and the prior information used for AIRS retrievals (Susskind et al. 2014) could affect the low occurrence of SHIs. The strength of SHIs (i.e., the specific humidity difference across the inversion layer) has been found to be larger in summer, even though SHIs are more frequent in winter (Devasthale et al. 2011; Nygård et al. 2014; Brunke et al. 2015).

The regional horizontal distribution of SHI occurrence has been well reported in recent climatological studies (Devasthale et al. 2011; Nygård et al. 2014; Brunke et al. 2015), but the vertical distribution of SHIs and its regional and seasonal variations have remained unstudied. In addition, the strong dependence between the properties of SHIs and TIs and the processes leading to the formation of SHIs have not been well explained so far. In this study, we examine the vertical and regional distributions of SHI occurrence and the regional distributions of SHI strength based on two atmospheric reanalyses and radiosonde soundings. In addition, physical processes behind the formation of SHIs are suggested. The paper is structured as follows: In section 2, the data and calculation methods are presented, and the potential sensitivity of the results to differences between the datasets is discussed. In section 3a, the reanalyses are compared to the radiosonde sounding data to evaluate the accuracy of the reanalyses in representing the atmospheric moisture distribution, and in section 3b, an overview of specific humidity distributions in the Arctic is presented. Then, SHIs above the 800-hPa level (section 3c) and below it (section 3d) are examined separately, because the processes responsible for the formation of SHIs are different between the low troposphere and the atmosphere aloft. In section 4, the key results of this study are compared with previous studies, and the processes behind SHI formation and data accuracy are discussed. The main conclusions of this study are presented in section 5.

2. Study region, material, and methods

We examined the occurrence and strength of SHIs on seasonal time scales and focused on winter (DJF) and summer (JJA), which represent extremes in atmospheric moisture and have clear differences in SHI occurrence and strength, whereas spring and autumn are transition periods between winter- and summer-type SHIs. The study area is the area north of 60°N. The area
is divided into six regions based on the spatial patterns of specific humidity conditions in the lower troposphere; the variation of conditions inside each region is much smaller than the variation between regions. These regions are 1) the Arctic Ocean, 2) North Atlantic, 3) northern Europe, 4) Siberia, 5) Alaska and Canada, and 6) Greenland. The regions are outlined in Fig. 1. The study period is 12 years, from January 2003 to December 2014, which is rather short for climatology, but close to the periods addressed by previous studies on Arctic SHIs (Nygård et al. 2014 and Devasthale et al. 2011). The period is short for the evaluation of the trends of SHIs but probably long enough to determine the main features of SHIs and the processes behind their formation.

a. Datasets

Two modern reanalyses, the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim (Dee et al. 2011) and the Japan Meteorological Agency (JMA) JRA-55 (Kobayashi et al. 2015), are used for the examination of SHIs in the Arctic. Variables used in this study have been stored in a regular latitude–longitude grid with a horizontal resolution of 0.75° in ERA-Interim and in a Gaussian grid with a horizontal resolution of 0.56° in JRA-55. Both reanalyses have 60 hybrid levels, with very similar vertical resolutions in the troposphere. The model-level fields were used because they have a better vertical resolution in the lower troposphere than the pressure-level fields. Both reanalyses use a four-dimensional variational data assimilation (4D-Var) system, and the analysis fields have been produced with 6-h intervals. As no surface flux data were assimilated into the reanalyses, the products for these variables are only based on short-term forecasts.

The reanalysis products are compared with radiosonde sounding data taken from the Integrated Global Radiosonde Archive (IGRA) (Durre et al. 2006; Durre and Yin 2008). Soundings from 36 stations north of 65°N are used in this study. Quality-assured IGRA sounding data have undergone several quality assurance checks, which have removed outliers from observations, providing a consistent dataset. The quality assurance checks do not remove biases caused, for example, by radiosonde types. IGRA data consist of observed variables and variables derived from the observed variables. Radiosonde soundings are taken mainly twice a day at 0000 and 1200 UTC. Specific humidity has a weak diurnal cycle. Hence, the differences in time resolution have probably only a minimal effect on the differences between the sounding data and the reanalyses. However, in summer over continents, the diurnal cycle of specific humidity may cause a difference between the soundings and reanalyses in SHI occurrence, if no soundings are carried out between 0000 and 0600 local solar time (see the discussion in section 4). The sounding data from the stations located in the six study regions are averaged for each region. As conditions at the sounding stations are possibly not representative for the whole region, the comparisons between the reanalyses and the sounding data are made on the basis of the values at the grid points closest to each sounding station (one grid point per station). The representativeness of the soundings is probably weakest over the oceans because all sounding stations are located on land. Three coastal stations (Eureka, Barrow, and Tiksi Bay), which are located near sea level, were utilized to represent the Arctic Ocean, and stations on small islands, Jan Mayen and Bear Island, and Ny Ålesund were used to represent the Atlantic Ocean.
b. Methods

Our focus was on the occurrence and strength of SHIs, and their spatial and seasonal variation. SHI occurrence is computed by comparing specific humidity values between consecutive model levels up to the 400-hPa level. For analyses, SHI occurrences between individual model levels are combined into thicker layers consisting of several model levels. This method does not yield the probability of SHI occurrence in a single vertical level but the probability of occurrence in the whole layer. SHI occurrence is calculated from the sounding data in the same way as from the reanalyses, except that, instead of model levels, all vertical levels in soundings, that is, mandatory and significant levels are utilized. A vertical resolution of 100 hPa is applied for the analyses of the vertical structure of SHI occurrence. Layers that are 100-hPa thick are used instead of 50-hPa-thick layers because SHI occurrence in the sounding data was systematically higher in the 50-hPa-thick layer above the mandatory pressure levels 850 hPa and 700 hPa than in the 50-hPa-thick layer below these levels. This feature was noticeable at most sounding stations and in addition, at the stations where the upward-decreasing trend of SHI occurrence was strong, the values above a mandatory level were not higher than below the level but deviated similarly from the trend. Those peaks in SHI occurrence are probably caused by the reporting method of sounding variables instead of real physical conditions. The sounding data capture finer details than the reanalyses, especially above the 800-hPa level, which may increase the probability of SHI occurrence in the sounding data. This is because the vertical resolution of reanalyses decreases from nine levels between 900 and 1000 hPa (when the surface pressure is 1000 hPa) to two or three levels per each 100-hPa interval above the 800-hPa level.

SHI strength is computed by subtracting the value of specific humidity at the lowest model level from the vertical maximum value of specific humidity. Therefore, it does not take into account local vertical gradients below the specific humidity maximum, nor the SHIs located above the specific humidity maximum. SHI strength is sensitive to the specific humidity at the lowest model level, which is not at Earth’s surface but at an altitude of approximately 10 m in both reanalyses. Hence, SHI strength does not exactly represent the specific humidity difference between the atmospheric maximum and the surface. When calculating the seasonal-mean SHI strength, zero values from the analysis times when the maximum is located at the lowest model level are included.

The vertical gradient of specific humidity $q$ can be written as follows:

$$
\frac{\partial q}{\partial z} = \frac{\partial S \rho_{v,\text{sat}} / \rho}{\partial z} = q \left( \frac{1}{S} \frac{\partial S}{\partial z} + \frac{1}{\rho_{v,\text{sat}}} \frac{\partial \rho_{v,\text{sat}}}{\partial T} \frac{\partial T}{\partial z} - \frac{1}{\rho} \frac{\partial \rho}{\partial z} \right),
$$

where $S$ is the saturation ratio, $\rho_{v,\text{sat}}$ is the saturation density of water vapor, $\rho$ is air density, $T$ is air temperature, and $z$ is the vertical coordinate. The first term on the right-hand side is the effect of the relative humidity vertical gradient on the vertical gradient of specific humidity. The second term on the right-hand side is the effect of saturation vapor density on specific humidity.

As the saturation vapor density only depends on temperature, the second term depends on the vertical gradient of temperature. The third term is the effect of the upward decrease in air density on specific humidity. Accordingly, the occurrence of an SHI is related to the occurrence of a relative humidity inversion (RHI), a TI, an upward-decreasing air density, or the occurrence of two or three of these factors simultaneously.

The formation of an SHI can result from humidity advection, condensation, or evaporation. Evaporation from rain or snowfall can generate an SHI only when the dry layer is located below a cloud. Typically the formation of an SHI is connected to moisture condensation or vertically differential humidity advection. In Fig. 2a, the lowest SHI layer (from the surface to 950 hPa), where the dewpoint temperature curve is tilted clockwise from specific humidity isolines, was associated with a cold air mass, and the formation of the SHI was related to moisture condensation due to surface radiative cooling. Surface cooling led to the formation of a TI, and after saturation was reached, further cooling led to moisture condensation and the formation of an SHI (process 3 in Fig. 3). In contrast, the formation of the SHI layer between the 975–950-hPa levels in Fig. 2b was largely contributed by humidity advection, even though the SHI occurred with a TI in saturated conditions (process 2 in Fig. 3). The strong SHI and TI were formed because of temperature and humidity advection due to southwesterly winds. However, the formation of the cold and dry layer near the surface was probably related to radiative surface cooling, and also saturated conditions in the SHI layer suggested the moisture condensation. Accordingly, both processes, humidity advection and condensation, had contributed to the evolution of the SHI. In both Figs. 2a and 2b, SHIs at upper tropospheric levels occurred with RHIs and unsaturated conditions, hence the formation of these SHIs was caused by humidity advection (process 1 in Fig. 3). In unsaturated conditions, when SHIs occurred...
simultaneously with RHIs, SHIs were formed because of differential humidity advection, either upward-increasing moist air advection or downward-increasing dry air advection. Moisture condensation or evaporation cannot form or strengthen an SHI when it occurs with an RHI, except in the case of evaporation from rain or condensation in supersaturated conditions when the availability of condensation nuclei determines the occurrence of condensation, for example, when supersaturation respect to ice occurs near the surface, and water vapor condenses onto the surface because of lack of ice nuclei.

Figure 2 illustrates the processes behind the formation of SHIs. Our hypothesis was that the simultaneous occurrence of SHIs and RHIs suggests that SHIs are formed as a result of humidity advection, whereas the simultaneous occurrence of SHIs and TIs in saturated conditions suggests that moisture condensation has impacted the formation of SHIs (Fig. 3). The criterion defining simultaneous occurrence was that inversions, either SHIs and TIs or SHIs and RHIs, should occur simultaneously, overlapping, in the layer examined. The simultaneous occurrence was examined for 50-hPa-thick layers. Hence, the probability of the simultaneous occurrence of SHIs with TIs or RHIs is not directly comparable to SHI occurrence. RHIs were computed from the relative humidity with respect to water when the air temperature was above 0°C and with respect to ice when the air temperature was below 0°C. A relative humidity of 99%, instead of 100%, was used for the limit of saturated conditions, because of possible numerical inaccuracy. Condensation often occurs even before the gridcell
mean relative humidity reaches 100%, but saturation in the entire gridcell volume is probably needed before the condensation substantially decreases the gridcell mean specific humidity. However, supersaturation with respect to ice is typical in the absence of ice nuclei. Therefore, it is not possible to set a universally valid relative humidity threshold for condensation.

The contribution of horizontal humidity advection to the strengthening of SHIs was examined by comparing the observed 6-h specific humidity tendencies against the tendencies calculated on the basis of horizontal specific humidity advection. Horizontal specific humidity advection was calculated by multiplying the specific humidity gradient along model-layer surfaces with the local horizontal wind speed. Linearity in the changes of humidity advection between analysis times was assumed. Only analysis times when the strength of SHIs increased between two consecutive model levels were used for the statistics. Based on these specific humidity tendencies, we calculated the contribution of differential humidity advection to the change of the specific humidity vertical gradient. For a further analysis, the contributions of humidity advection on SHI strength were averaged from layers between consecutive model levels to thicker layers. The computation of horizontal humidity advection from analysis fields with a low temporal resolution and over areas with substantial topographic variations is vulnerable to discretization errors (Seager and Henderson 2013). In addition, the assimilation of observations affects the tendencies in the reanalyses. Therefore, the advection calculations were not able to give quantitatively accurate results of the contribution of humidity advection to the strengthening of SHIs.

3. Results

a. Comparison of reanalyses and soundings

The accuracy of the reanalyses was evaluated by comparing them with the sounding data. Away from radiosonde sites, the reanalyses might be less accurate because of the lack of assimilation of radiosonde data and, particularly over sea ice and snow/ice-covered land, limited assimilation of infrared and microwave sounding data from satellites. However, radiosonde soundings may include errors. Typical errors are due to humidity sensor time lag, sensor icing, and sensor drying due to the heating of solar radiation (Anderson 1995; Miloshevich et al. 2006; Ingleby 2017). The time lag does not cause a bias in mean values but solar heating may cause a small dry bias, and sensor icing during an ascent through a cloud leads to a moist bias above the cloud (Miloshevich et al. 2006). Sensitivity to errors varies between radiosonde types (Miloshevich et al. 2006; Ingleby 2017). Hence, the differences between reanalyses and soundings do not always indicate the inaccuracy of reanalyses. The comparison at individual sounding sites was performed by choosing simultaneous data from soundings and from the closest grid point of the reanalyses. After that, the results were averaged for each region.

In both reanalyses, the mean profiles of specific humidity match the sounding data quite well (Figs. 4 and 5, left). However, the specific humidity in both reanalyses was typically lower than in the sounding data in winter (Fig. 4, center). JRA-55 was mostly drier than the soundings in all seasons with the largest differences occurring in summer over Siberia, where the difference exceeded 0.2 g kg\(^{-1}\) in the 500–700-hPa layer (Fig. 5k). In ERA-Interim, the biases were smaller than in JRA-55, except in summer in the layer below 800 hPa (Fig. 5, center), where ERA-Interim was too moist. The summertime near-surface moist bias in ERA-Interim over the central Arctic Ocean has been reported in previous studies (Lüpkes et al. 2010; Jakobson et al. 2012; Wesslén et al. 2014). Our results indicated that the near-surface moist bias also occurs over continents. Another difference between ERA-Interim and the sounding data was in the midtroposphere over northern Europe and Siberia, where ERA-Interim was too dry. Over northern Europe and Siberia, both reanalyses had a dry bias in the 500–800-hPa layer. At least part of the dry bias in the reanalyses in comparison with the soundings can be explained by sensor icing problems in the radiosondes, notably those used in Russia (Ingleby 2017). On average, ERA-Interim had a smaller bias and a root-mean-squared error than JRA-55 in the comparison with radiosonde soundings.

SHIs were on average stronger and more frequent in the sounding data than in the reanalyses (Fig. 6 and Table 1). Above the 900-hPa level, SHIs were more frequent in JRA-55 than ERA-Interim and vice versa below the 900-hPa level, but in the sounding data SHI occurrence was higher than in either of the reanalyses (Fig. 7). The difference in SHI occurrence between the reanalyses and the sounding data could be affected by their different vertical resolution. In the reanalyses, the vertical resolution decreases upward. Accordingly, at higher altitudes, reanalyses cannot resolve as many thin inversion layers as they can at lower altitudes, which may contribute to the upward-decreasing SHI occurrence in the reanalyses. The upward decrease in SHI occurrence was indeed larger in the reanalyses, especially in ERA-Interim, than in the soundings.
FIG. 4. (left) DJF mean specific humidity from ERA-Interim, JRA-55, and soundings. (center) Bias and (right) root-mean-square difference of ERA-Interim (dashed line) and JRA-55 (solid line) in the comparison with soundings in winter.
Fig. 5. As in Fig. 4, but for JJA.
atmospheric conditions were different between the layers which occurred simultaneously with SHIs indicated that atmospheric humidity is found (Fig. 8). Even though water vapor in the lower troposphere, where most of humidity was usually high in the lower troposphere in Greenland, and the Canadian archipelago. Relative lowest values occurred over three cold areas: Siberia, except for the low values in Greenland, but in winter the por content was almost symmetric zonally in summer, Greenland. The field of vertically integrated water va-

values exceeded 10 kg m$^{-2}$ almost everywhere north of 60$^\circ$N, but in summer the values exceeded 10 kg m$^{-2}$ everywhere, except over Greenland. The field of vertically integrated water vapor content was almost symmetric zonally in summer, except for the low values in Greenland, but in winter the lowest values occurred over three cold areas: Siberia, Greenland, and the Canadian archipelago. Relative humidity was usually high in the lower troposphere in winter, therefore air temperature limited the amount of water vapor in the lower troposphere, where most of atmospheric humidity is found (Fig. 8). Even though specific humidity typically increased downward, the layers where specific humidity decreased, that is, SHI layers, occurred frequently in the Arctic troposphere (Fig. 7). The occurrence of RHIs and TIs (Figs. 9 and 10), which occurred simultaneously with SHIs indicated that atmospheric conditions were different between the layers above and below approximately the 800-hPa level, so we present the properties of SHIs separately for the layers above and below the 800-hPa level.

c. Specific humidity inversions above the 800-hPa level

SHIs were present less than 40% of the time in each 100-hPa-thick layer between the 400- and 800-hPa levels in both reanalyses, except in Greenland and in winter in Siberia and in the Alaska/Canada region, where SHIs were more frequent (Fig. 7). In the sounding data, the occurrence was higher, mostly between 40% and 60%. The difference between the reanalyses and the sounding data could not be explained by the locations of the sounding stations, because the occurrence profiles that were computed only from the grid points nearest to the sounding stations (dashed lines in Fig. 7) were representative for the whole region, but differed from those based on the sounding data. The seasonal and regional variations in SHI occurrence were smaller than the differences between the datasets.

Above the 800-hPa level, the formation of SHIs was mostly due to specific humidity advection, because almost all SHIs occurred simultaneously with RHIs but almost none with TIs in the 400–700-hPa layer in the reanalyses (Figs. 9 and 10). In the reanalyses, TIs were rare in this layer (occurrence mostly below 1% in each 100-hPa-thick layer). Mean relative humidity was mostly below 50% in the 400–800-hPa layer when SHIs occurred in both reanalyses. In the sounding data, the proportion of the SHIs occurring simultaneously with RHIs was smaller (Fig. 10), and the proportion of the SHIs occurring simultaneously with TIs was larger than in the reanalyses (Fig. 9).

d. Specific humidity inversions below the 800-hPa level

SHI occurrence below the 800-hPa level was dependent on the surface heat budget. In winter SHI occurrence was high over the continents and ice-covered seas (Figs. 11a,b), where the negative net longwave radiation effectively cooled the surface. In summer SHI occurrence in the reanalyses was highest over the Arctic Ocean (Figs. 11c,d), where sea ice and snowmelt and the large heat capacity of the open sea limited the seasonal increase of the near-surface air temperature. Spring and autumn were transition periods between winter and summer types of SHIs (Fig. 12). In addition, over sloping surfaces, especially over slopes near the coasts of Greenland, SHIs associated with RHIs and formed via differential humidity advection (Fig. 13) were identified. The formation mechanism of these SHIs was probably a consequence of dry and cold air advection from an ice sheet to a near-surface coastal layer as suggested for Antarctica by Nygård et al. (2013).
TABLE 1. Mean SHI occurrence and strength for regions of datasets. The last column, "Arctic," covers the whole area north of 60° N. In each column, the first value is the mean value over whole region, and the second value is the mean value of the locations of the sounding stations only. SHI occurrence includes only the analyses and the grid points for which the surface pressure was higher than 800 hPa (for most of Greenland, the surface pressure is always smaller).

<table>
<thead>
<tr>
<th></th>
<th>Arctic Ocean</th>
<th>North Atlantic</th>
<th>Northern Europe</th>
<th>Siberia</th>
<th>Alaska/Canada</th>
<th>Greenland</th>
<th>Arctic</th>
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<tr>
<td><strong>SHI occurrence below 800 hPa</strong></td>
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<td>0.43, 0.50</td>
<td>0.81, 0.83</td>
<td>0.98, 0.99</td>
<td>0.98, 0.98</td>
<td>0.82, 0.85</td>
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<td>0.77, 0.75</td>
<td>0.97, 0.99</td>
<td>0.94, 0.99</td>
<td>0.74, 0.89</td>
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<td>Soundings</td>
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<td>—, 0.73</td>
<td>—, 0.86</td>
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<td>—, 0.99</td>
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<td>0.26, 0.26</td>
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<td>—, 0.53</td>
<td>—, 0.47</td>
<td>—, 0.21</td>
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<td>—, 0.23</td>
<td>—, 0.37</td>
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<tr>
<td></td>
<td>No. of sounding stations, No. of soundings</td>
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<td>12, 19 376</td>
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<td>3, 5386</td>
<td>9, 15 789</td>
<td>12, 19 452</td>
<td>8, 16 946</td>
<td>4, 7504</td>
<td>36, 65 077</td>
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1) Specific Humidity Inversions in Winter

The formation of winter SHIs began in October over the northernmost land areas and ice-covered seas (Fig. 12). The strength and the areal extent of SHIs increased during November and December with decreasing solar radiation and decreased in spring with increasing solar radiation (Fig. 12.). In winter, SHI occurrence below the 800-hPa level was high, exceeding 90%, over cold surfaces over continents and the Arctic Ocean, and low, less than 50%, over the warm sea surface of the North Atlantic in both reanalyses (Fig. 11 and Table 1). Vertically, SHIs were most common in the layer below 900 hPa (Fig. 7). The differences between the reanalyses and between the reanalyses and the sounding data were small in the areas of high SHI occurrence. In JRA-55, SHI occurrence was slightly higher over sea ice than in ERA-Interim (Fig. 11). In Greenland, SHI occurrence could not be properly presented at constant pressure levels because of the orography. In Figs. 7, 9, and 10 the curves below the 800-hPa level only represent the coastal areas of Greenland, because the surface pressure is always lower than 800 hPa in most of Greenland (masked area in Fig. 11). Nevertheless, SHIs between the two lowest model levels were most frequent in Greenland.

On continents and the Arctic Ocean, most wintertime SHIs below the 800-hPa level occurred simultaneously with TIs (Fig. 9), and the occurrence with RHIs was smaller than at higher altitudes (Fig. 10). At low levels, SHIs normally occurred in conditions of high relative humidity, and saturated conditions (relative humidity over 99%) were present on average in 10%–30% of the cases when an SHI occurred in the reanalyses. Accordingly, SHIs occurred in conditions when condensation was able to form or strengthen SHIs. The importance of the contribution of condensation in the formation of SHIs was supported by the result that most continental areas and the Atlantic Ocean, the contribution of vertically differential horizontal humidity advection was less than 50% (Fig. 13).

The mean SHI strength was lower in the reanalyses, especially in JRA-55, than in the sounding data (Fig. 6). The highest mean SHI strength occurred inland in northern Canada and Alaska (Fig. 14) in all datasets, but the differences in the spatial distribution of the mean SHI strength between the reanalyses were much larger than in SHI occurrence. In ERA-Interim, the highest values of the mean SHI strength occurred over continents and near the eastern coast of Greenland. In addition to the areas of strong SHIs in northern Canada and Alaska and the eastern coast of Greenland, high values of the mean SHI strength were also found over
FIG. 8. (a) DJF and (b) JJA means of vertically integrated water vapor from ERA-Interim and soundings (circles). DJF and JJA means of specific humidity cross sections along latitudes (c),(d) 65°N and (e),(f) 75°N from ERA-Interim.
sea ice near open seas near the Bering Strait and north of Svalbard in JRA-55. On continents, the mean SHI strength was higher in ERA-Interim than in JRA-55 because of a larger vertical gradient of specific humidity in SHIs. On the contrary, over sea ice, the mean SHI strength in JRA-55 was higher than in ERA-Interim. In ERA-Interim, the mean SHI strength was smaller over sea ice than over land, but in JRA-55, there was no difference in the mean SHI strength between these surface types. Over the Arctic sea ice, the latent heat flux was upward in ERA-Interim, whereas in JRA-55 it was downward, which probably at least partly explains the moister conditions at the lowest model level and weaker SHIs over sea ice in ERA-Interim than in JRA-55. However, the magnitudes of the latent heat fluxes were small in both reanalyses.

2) SPECIFIC HUMIDITY INVERSIONS IN SUMMER

The strongest SHIs in summer formed over seas when warm, moist air was advected from land to over a relatively cold sea surface. Even though SHIs were also present over continents, they were fewer (Fig. 11) and weaker (Fig. 14) than SHIs over cold seas. The season of summer-type SHIs over the Arctic Ocean began in June, when the air specific humidity over continents exceeded the saturation specific humidity of the sea surface, and ended in September (Fig. 12).

An example of the formation of summer-type SHIs is presented in Fig. 15. The formation of SHIs over the East Siberian Sea and the Arctic Ocean on 6–8 August 2014 was caused by the offshore flow of moist air from Siberia. The specific humidity maximum occurred below the 900-hPa level near the Siberian coast, but its height increased with the fetch over the sea, as did the height of the air temperature maximum; these increases were due to upward-increasing moisture and heat transport and the cooling and drying of the lowest layers of the advected air mass. The wind direction was almost the same in the entire layer of the major humidity advection from the surface up to the 700-hPa level, and wind speed increased only slightly upward from the 900- to 750-hPa level. The maximum moisture transport occurred between the 850- and 900-hPa levels, almost collocating with the layer of the maximum specific humidity. Relative humidity was at the saturation point below the humidity maximum, and both sensible and latent heat fluxes were downward below the SHI, resulting in cooling and drying, which was also contributed to by the condensation of moisture of the advected air mass. These processes also decreased specific humidity at the level of the humidity maximum, causing a weakening of SHI with increasing distance from the coast.

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**Fig. 9.** DJF and JJA means of the fraction of SHIs that occurred simultaneously with TIs in 50-hPa-thick layers for the study regions from ERA-Interim and JRA-55 (solid lines) and for the locations of the sounding stations from ERA-Interim, JRA-55, and soundings (dashed lines).
At the same time as the example case, the Swedish icebreaker *Oden* was in the East Siberian Sea slightly eastward of the location of the example case, where almost similar offshore humidity advection occurred a couple of days later. The ERA-Interim values for the maximum specific humidity, over 10 g kg⁻¹, and the altitude of the humidity maximum near the coast in the example case (Fig. 15) were close to the values observed at *Oden* by Tjernström et al. (2015), but the surface fluxes of sensible heat and latent heat had a much larger magnitude than the measured fluxes reported by Tjernström et al. (2015).

In summer, SHIs were most frequent over the Arctic Ocean (Table 1). The maximum occurrence was in both reanalyses near the coast of the Arctic Ocean, and the occurrence was higher in ERA-Interim than JRA-55 (Fig. 11). Over the Arctic Ocean, the difference between the reanalyses increased with distance from the coast, so in ERA-Interim SHIs frequently penetrated farther over the sea. In summer, near-surface SHIs (SHIs between the two lowest model levels) were frequent in central Greenland but they occurred above the 800-hPa level.

The mean SHI strength was largest near the coast of the Arctic Ocean and decreased toward the North Pole and the northern coast of Greenland (Fig. 14). Summer SHIs near the coast of the Arctic Ocean were even stronger than winter SHIs over continents (Fig. 14). Stronger SHIs in summer than in winter have been also reported in previous studies (Devasthale et al. 2011; Nygård et al. 2014; Brunke et al. 2015). In JRA-55, the mean SHI strength over the Arctic Ocean was only 40% of that in ERA-Interim (Table 1). In the sounding data, SHIs were weak in summer, and the differences between the sounding data and the reanalyses were smaller in summer than in winter (Fig. 6). However, the strongest SHIs were not well represented in the sounding data, because the sounding stations are located on land. As strong SHIs over the sea occurred during offshore winds, the sounding stations were in these cases under an influence of continental conditions. However, when strong SHIs occurred over land near the coast of the Arctic Ocean, especially in Siberia, the near-surface winds were directed from sea to inland areas.

3) SPECIFIC HUMIDITY INVERSIONS GENERATED BY DOWNSLOPE WINDS

Over a sloping surface, especially in the coastal zone of Greenland, SHIs were often generated by the advection of dry air due to katabatic or other downslope winds. For these areas, the effect of advection dominated the formation of SHIs (Fig. 13). In these
conditions, the relative humidity at the lowest model level was even below 70%, and the relative humidity was lower at the lowest model than at the level of the specific humidity maximum, but the temperature difference between those levels was small. Over the slopes of the Greenland ice sheet, high simultaneous occurrences of SHIs and RHIs were found from the approximately 30-hPa level above the surface when SHI occurrence was examined on model levels. Accordingly, an RHI rather than a TI explained the formation of SHIs over

Fig. 11. (a),(b) DJF and (c),(d) JJA means of SHI occurrence between the level of 800 hPa and the surface from ERA-Interim, JRA-55, and soundings (circles).
FIG. 12. Monthly means of SHI strength from ERA-Interim.
the slopes of Greenland. The formation of a dry-air layer requires the formation of a cold, dry air mass on a highly elevated area and the downslope advection of this air mass. When adiabatic warming decreased the relative humidity faster than evaporation increased it, a layer of low relative humidity formed near the surface, which could also be seen in the reanalyses’ surface fluxes: a strong downward sensible heat flux was present in these areas, but the latent heat flux was upward.

4. Discussion

We examined the spatial distribution of SHI occurrence based on reanalyses and sounding data. In general, SHI occurrence was as high as in previous studies (Brunke et al. 2015; Nygård et al. 2014; Devasthale et al. 2011). We found that the seasonal variation of SHI occurrence was large over the continents but smaller over the oceans, and seasonal and spatial variations occurred mostly below the 800-hPa level. The amount of spatial variation in SHI occurrence depended on the thickness of the examined layer, in such a way that, for large areas in winter, the occurrence was close to 100% in the layer below the 800-hPa level (Fig. 11). The spatial variation of SHI occurrence became smaller when a thicker layer was examined. The above-mentioned effect is a consequence of the simultaneous occurrence of SHIs in several layers. Nygård et al. (2014), addressing the layer up to the 500-hPa level, reported seasonal and spatial variations smaller than found here in the layer up to the 800-hPa level. They reported that in the layer up to the 500-hPa level in every season, SHI occurrence was over 70% at all Arctic sounding stations, except Turuhansk in Siberia. Our findings based on the sounding data indicated that SHI occurrence below the 800-hPa level was below 70% at almost half of the sounding stations in summer and at three stations (near the Atlantic Ocean) in winter. In winter, the spatial distribution of SHI occurrence below the 800-hPa level was rather similar to that presented by Devasthale et al. (2011) based on AIRS data for the layer between the surface and the 400-hPa level. In summer over the Arctic Ocean, our results indicated a much higher SHI occurrence than the results of Devasthale et al. (2011). At least a part of the difference could be explained by the fact that their data did not well represent summer conditions over the Arctic Ocean, because they could only use clear-sky observations, but cloudy conditions prevail over the Arctic Ocean (Intrieri et al. 2002b).

Nygård et al. (2014) and Brunke et al. (2015) have discussed the role of advection and moisture condensation for the formation of SHIs. We found that atmospheric conditions related to SHIs were notably different below and above the 800-hPa level. Above the 800-hPa level during the occurrence of SHIs, the stratification was weaker and relative humidity was lower than in the layer below 800 hPa, which allowed upward-increasing heat and humidity advection and the formation of SHIs without the formation of TIs. Above the 800-hPa level, SHIs occurred at the boundaries between dry and moist air masses, which were probably formed when a moist air mass was advected above a dry air mass. The computation of the effect of differential humidity advection on strengthening SHIs showed that
differential humidity advection explained only approximately 50% of the increase in SHI strength among all strengthening SHIs above the 800-hPa level. However, the method to evaluate the evolution of SHI strength from advection profiles did not take into account the effect of data assimilation, which could affect this result.

In the boundary layer, the cooling of an air mass over a cold surface leads to high relative humidity and very stable stratification and then to the occurrence of TIs and SHIs. Our results clearly indicated that the frequent occurrence of near-surface SHIs in the Arctic is closely connected to 1) the surface cooling caused by the negative net radiation in winter and 2) the cold surfaces of both ice-covered and open parts of the Arctic Ocean in summer (Fig. 11). This suggests that moisture condensation related to near-surface air mass cooling is an

FIG. 14. (a),(b) DJF and (c),(d) JJA means of SHI strength from ERA-Interim, JRA-55, and soundings (circles).
FIG. 15. (a),(c),(e) SHI strength and moisture flux density at the level of 850 hPa (black arrows) from ERA-Interim. (b),(d),(f) Specific humidity cross section along longitude 150°E [marked with a red line in (a), (c), and (e)] and meridional moisture flux density (black arrows) from ERA-Interim.
important factor for generating SHIs below 800 hPa, which is supported by the fact that most of the SHIs below the 800-hPa level occurred simultaneously with TIs (Fig. 9) and high relative humidity. In winter, the formation of TIs and SHIs is conventionally thought to be due to radiative surface cooling in the conditions of clear skies or optically thin clouds, which is often the case in the Arctic (Stramler et al. 2011), but SHIs are also present in cloudy conditions (Stramler et al. 2011; Sedlar et al. 2012; Nygård et al. 2014; Sedlar 2014). Clouds modify radiative energy transfer, increasing the downward longwave radiation at the surface (Shupe and Intrieri 2004). In winter, cloudy conditions (Stramler et al. 2011) and increased humidity advection (Woods et al. 2013) are often associated with synoptic-scale cyclones, and both of them probably affect the specific humidity profile. Above the 800-hPa level, a small part of SHIs occurred without the presence of a TI (Fig. 9), and the majority of SHIs (Fig. 10) occurred simultaneously with an RHI, suggesting the effects of humidity advection on SHI occurrence (process 1 in Fig. 3). Nygård et al. (2013, 2014) proposed that an upward-increasing horizontal moisture transport–related large-scale moisture convergence may be important for the formation of SHIs, and Brunke et al. (2015) showed that the effect of differential humidity advection can strengthen or weaken SHIs depending on the location. Our results also indicate that the differential humidity advection contributed to the strengthening of SHIs but did not entirely explain the formation of SHIs (Fig. 13). Devasthale et al. (2011) suggested that a TI prevents effective moisture transport near the surface and the downward turbulent mixing of moisture from the layer of the maximum horizontal moisture transport, and in that way, affects the formation of SHIs (process 5 in Fig. 3). Accordingly, humidity advection may contribute to the occurrence of an SHI or also increase its strength in cases when the SHI was originally generated via surface cooling and condensation (process 4 in Fig. 3).

The main formation mechanism of summer SHIs is the advection of warm, moist air over a cold surface, causing condensation and the removal of moisture from the lowest parts of the boundary layer. An important difference between summer and winter is that in summer the Arctic Ocean surface temperature typically remains between −2°C and 0°C (Lümpkes et al. 2010; Tjernström et al. 2012; Sotiropoulou et al. 2016) because of the large heat capacity of the ocean and the presence of (melting) snow or ice. Therefore, in summer SHI strength is related to the specific humidity of the advected air mass, which agrees with the suggestion of Brunke et al. (2015) that regional moisture convergence in the Arctic is related to the mean strength of SHIs. Instead, in winter, the surface cooling due to a negative net radiation may result in the deepening of TIs and SHIs without heat or humidity advection. In summer over the Arctic Ocean, air mass cooling and moisture condensation at the level of the specific humidity maximum result in weakening SHIs with increasing fetch over the cold sea, which reduces the area of the occurrence of strong SHIs over a cold sea surface. Previous studies (Solomon et al. 2011, 2014; Sedlar et al. 2012; Shupe et al. 2013) have shown that an SHI layer above a cloud layer can be a moisture source for the cloud layer. Thus, turbulent downward moisture transport from the SHI layer to the cloud layer together with moisture condensation in clouds or in fog are probably important for the drying of the advected air mass. This is supported by our result that the layer of near-surface cooling and drying grew higher with increasing distance from the coast (Fig. 15), as well as by previous observations of the frequent occurrence of 100% relative humidity (Tjernström et al. 2012; Sotiropoulou et al. 2016), fog, and extensive low cloudiness (Intrieri et al. 2002b; Tjernström 2005; Sotiropoulou et al. 2016), which indicate that moisture condensation is common over the Arctic Ocean. Upward motions with adiabatic cooling and moisture condensation can also contribute to air mass drying.

Based on the comparison with radiosonde soundings, the reanalyses are able to represent the main features of the spatial variation (between sounding sites) of the occurrence and strength of SHIs. However, the mean SHI strength was underestimated in both reanalyses. In winter, the error was notably larger for JRA-55 than ERA-Interim, and this was due to a smaller vertical specific humidity gradient in JRA-55. This suggests that vertical mixing in JRA-55 is probably too strong in very stable conditions, which prevents the formation of strong inversions. The stronger vertical mixing in JRA-55 can also explain the moister conditions at the lowest model level over continents and the differences in SHI properties over the Arctic Ocean in summer: weaker vertical mixing in ERA-Interim allowed moist air advection to penetrate farther to the Arctic Ocean without strong cooling and drying by turbulent heat fluxes. Soundings are mostly performed twice a day, whereas the time resolution of the reanalysis products consisted of four analyses a day. This probably only caused a minimal effect on the comparison of the reanalyses and soundings, because the diurnal cycle of specific humidity was weak. In the reanalyses, the largest amplitude of the diurnal cycle was observed over continents in summer, where it possibly affected SHI occurrence near the surface. ERA-Interim showed that the near-surface specific humidity had two maximums and minimums a
day. The maximums occurred between 0600–1200 and 1800–2400 local solar time, and the minimums occurred between 0000–0600 and 1200–1800 local solar time. The nocturnal minimum was related to the SHI caused by surface radiative cooling. If soundings had not been performed in nighttime (0000–0600 local solar time), this SHI could not be detected.

SHI occurrence, and especially SHI strength, was sensitive to surface properties and especially surface heat fluxes. Upward latent heat fluxes from ice-covered seas to the atmosphere in ERA-Interim compared with downward latent heat fluxes in JRA-55 were probably at least partly responsible for the weaker SHIs over sea ice in ERA-Interim than in JRA-55. For the summer cruise examined by Wesslén et al. (2014), the upward latent heat flux was overestimated in ERA-Interim over the central Arctic Ocean, and the downward sensible and latent heat fluxes in our example case were larger than the fluxes measured by Tjernström et al. (2015) during a similar episode as the example case. Further, Tastula et al. (2013) showed that reanalysis products for the similar episode as the example case. The comparison to radiosonde soundings showed that reanalyses were able to describe the mean vertical profiles of specific humidity and the seasonal and spatial variations of SHI occurrence and strength. However, SHI occurrence above the 800-hPa level and SHI strength were underestimated in the reanalyses, the latter particularly in JRA-55. The vertical resolution decreases upward in the reanalyses, so that very thin SHI layers cannot be resolved, which generated differences between the reanalyses and the sounding data. Uncertainty in turbulent surface fluxes and vertical mixing probably caused differences in SHI strength between the reanalyses and the sounding data.

These results provide a more detailed view of the processes involved in SHI formation. To better understand the Arctic climate system, further studies on the interactions between SHIs and the vertical distributions of clouds and moisture transport to the Arctic, as well as the associated radiative and turbulent processes, are needed.

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REFERENCES
Boisvert, L. N., and J. C. Stroeve, 2015: The Arctic is becoming warmer and wetter as revealed by the Atmospheric Infrared


——, and Coauthors, 2012: Meteorological conditions in the central Arctic summer during the Arctic Summer Cloud Ocean Study (ASCOS). *Atmos. Chem. Phys.*, 12, 6863–6889, https://doi.org/10.5194/acp-12-6863-2012.


Atmospheric moisture transport between mid-latitudes and the Arctic: Regional, seasonal and vertical distributions

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Horizontal moisture transport has a manifold role in the Arctic climate system as it distributes atmospheric water vapour and thereby shapes the radiative and hydrological conditions. Moisture transport between the Arctic and the mid-latitudes was examined based on ERA-Interim reanalysis. The meridional net transport is only a small part of the water vapour exchange between the Arctic and mid-latitudes and does not give a complete view of temporal and spatial variations in the transport. Especially near the surface, most of the northwards moisture transport is balanced by the southwards transport, and therefore the meridional net moisture transport at 60°–70°N peaks approximately at 100 hPa higher altitude than the northwards and southwards moisture transports. The total moisture transport (sum of absolute northwards and southwards moisture transports) has a much larger seasonal variation than the net transport (mean meridional transport), and the strength of the total transport is related to atmospheric humidity rather than the wind field. Strong individual moisture transport events contribute to a large part of the northwards moisture transport. This is consistent with the result that the net moisture transport is essentially generated by temporal variations of moisture fluxes. The moisture transport due to stationary zonal variation in the mass flux mostly defines the spatial distribution of the meridional moisture transport. The seasonal cycle of the net moisture transport is related to the seasonal cycle of transient eddy moisture transport but inter-annual variations of the net moisture transport are largely influenced by the stationary eddy moisture transport.

KEYWORDS
climate, humidity, moisture transport, polar, reanalysis, troposphere

1 INTRODUCTION

Atmospheric water vapour is an effective greenhouse gas, and thereby it directly affects radiative transfer. It is also the source for formation of clouds and fog which likewise largely influence the radiation reaching Earth’s surface. Clouds increase the downwelling longwave radiation and for most of the year as well as the net radiation on snow and ice-covered surfaces in the Arctic, as the effect of clouds on the amount of solar radiation absorbed in the snow and ice surface is small for most of the year (Shupe and Intrieri, 2004). Positive anomalies of atmospheric water vapour, clouds and downwelling longwave radiation have been shown to initiate spring onset of snow melt over the Arctic sea ice (Maksimovich and Vihma, 2012; Mortin et al., 2016). Years with an anomalously low end-of-summer sea ice extent have been connected to a significantly enhanced transport of moisture during spring into the region where the ice retreat is encountered (Kapsch et al., 2013). Radiative impacts of water vapour can also be seen on shorter timescales as a passage of an intense water vapour intrusion event can cause significant changes to the surface energy budget (Doyle et al., 2011) and a retreat of the ice in the marginal sea ice zone lasting for several days (Woods and...
Caballero, 2016). From a hydrological perspective, atmospheric moisture flux convergence acts as a major freshwater input to the Arctic Ocean, either directly through precipitation over the sea or as river discharge initiating from precipitation over the river basins on adjacent land (Serreze et al., 1995; Groves and Francis, 2002; Oshima and Yamazaki, 2004; Oshima and Yamazaki, 2006; Liu and Barnes, 2015). The mean meridional circulation is related to the Polar cell causing southwards flow near the surface and northwards flow aloft (Jakobson and Vihma, 2010; Dufour et al., 2016). SEs represent moisture transport due to deviations of specific humidity and wind speed from the zonal mean, whereas TEs represent deviations from the temporal mean. TE moisture transport is responsible for the majority of northwards net moisture transport, whereas the contribution of SE is rather small (5–9%) (Jakobson and Vihma, 2010). As an alternative approach to the division of moisture transport between the mean meridional circulation and TEs, Newman et al. (2012) separated time-varying departures from the seasonally varying basic state into contributions of synoptic (<10 days) and low-frequency anomalies (>10 days) and concluded that low-frequency anomalies drive two thirds of the transport to the Arctic throughout the year. Instead of dividing moisture transport based on the timescale of anomalies, Graversen and Burtu (2016) divided moisture transport into parts of planetary and synoptic waves based on wavelengths. Their conclusion was that the planetary waves caused approximately 60% of meridional net moisture transport at 60°N.

The moisture transport to the Arctic is largely driven by extreme events, that is, moisture intrusions, in which the moisture flux is larger than the 90% percentile (Woods et al., 2013; Liu and Barnes, 2015). These extreme transport events are typically related to blocking high-pressure systems that steer the cyclones towards the pole (Woods et al., 2013) and Rossby wave breaking (Liu and Barnes, 2015). In addition, Liu and Barnes (2015) showed that the transport paths of moisture are typically situated within warm tongues, which limits the loss of water vapour due to condensation. This agrees with the concept of atmospheric rivers, narrow plumes with high water vapour content and strong low-level winds (Zhu and Newell, 1998; Gimeno et al., 2014; Baggett et al., 2016; Komatsu et al., 2018). Zhu and Newell (1998) proposed that 90% of the moisture transport at any given time at any given latitude takes place within only 10% of the total longitudinal length.

The vertical structure of moisture transport has received surprisingly little attention although it potentially has remarkable importance for the cloud climatology and radiative fluxes. Based on radiosounding data from the circumpolar Arctic, the zonally averaged meridional net moisture transport has been reported to peak at the 850-hPa level (Serreze et al., 1995; Dufour et al., 2016), whereas estimations for the peak altitude based on reanalyses have varied between 750 and 990 hPa (Jakobson and Vihma, 2010; Bengtsson et al., 2011; Dufour et al., 2016). The strength of the moisture transport depends on wind speed and availability of moisture. Wind speed typically increases with height,
whereas specific humidity decreases with height, except in low-tropospheric humidity inversion layers (Devasthale et al., 2011; Sedlar et al., 2012; Nygård et al., 2014; Naakka et al., 2018). Scarcity of humidity limits moisture transport in the middle and upper troposphere. However, the meridional net moisture transport is the difference between northwards and southwards moisture transports, and thus the net moisture transport is largely affected by the meridional moisture gradient and the correlation between specific humidity and meridional wind speed. On decadal timescales, moisture transport to the Arctic has been projected to increase in the 21st century due to an increase in atmospheric moisture content and a larger northwards moisture gradient (Skific et al., 2009; Vihma et al., 2016). The latter is due to the Clausius–Clapeyron relation stating an exponential dependency of water-vapour pressure on temperature implying that the moisture content will increase more at low than at high latitudes during global warming. However, Dufour et al. (2016) found that although the atmospheric moisture content has already increased in the Arctic, there has been no significant change in the northwards moisture transport because the correlation between meridional wind and specific humidity anomalies has decreased.

Previous studies have provided a comprehensive understanding on the vertically integrated net moisture transport to the Arctic, including decomposing the roles of mean meridional circulation, SEs and TEs. However, regional, seasonal and especially vertical distributions of moisture transport have not been studied in sufficient detail to understand the manifold role of moisture transport in the Arctic climate system. For example, the net transport, on which the major research focus has been, does not represent the actual moisture transport in all regions, as northwards and southwards transports partly compensate for each other. In this study, we investigate the northwards, southwards and net moisture transports between the mid-latitudes and the Arctic with a specific focus on regional, seasonal and vertical aspects. In section 2 we present the data sets used and in section 3.1 the results of vertically integrated moisture transport. Section 3.2 focuses on the moisture transport profiles.

2 | DATA AND METHODS

2.1 | Data sets

In this study, moisture transport is examined based on ERA-Interim reanalysis from the European Centre for Medium-Range Weather Forecasts (Dee et al., 2011). In Text S1, the results of ERA-Interim are compared with the 55-year reanalysis of the Japan Meteorological Agency (JRA-55; Kobayashi et al., 2015) and with estimates of moisture transport based on specific humidity profiles retrieved from the Atmospheric Infrared Sounder (AIRS; Chahine et al., 2006; Susskind et al., 2014) and combined with ERA-Interim mass fluxes. In addition, specific humidity profiles from AIRS and the reanalyses are compared with radiosoundings from the Integrated Global Radiosonde Archive (IGRA; Durre et al., 2006; Durre and Yin, 2008) to investigate uncertainty related to humidity profiles (Text S1).

The study period is from 2003 to 2014. The 12-year period is likely incapable to capture all variations that occur in the current Arctic climate, but the period is very likely long enough to estimate the main properties of moisture transport between mid-latitudes and the Arctic. Climate change across the Arctic also affects moisture transport (Gimeno-Sotelo et al., 2018), and the choice of a relatively short period limits the impacts of climate trends in a decadal scale, and puts more focus on the moisture transport of the current Arctic climate system.

2.1.1 | Reanalyses

Reanalyses provide a gridded and coherent representation of the atmospheric state by assimilating various systems of observations into a numerical weather prediction model. ERA-Interim represents the state-of-the-art of atmospheric reanalyses applying a four-dimensional variational data assimilation method to assimilate a wide range of observations. The spectral model resolution of ERA-Interim is T255, and the data set used in this study has a horizontal resolution of 0.75 × 0.75°. We mainly utilized 6-hourly output from the pressure levels. Bengtsson et al. (2011) warned about integrating moisture fluxes on pressure levels, especially in the regions with complex topography, but Dufour et al. (2016) showed that the moisture convergence results for north of 60°N based on pressure levels and model levels agree within 2%. The use of pressure coordinates also simplifies the comparison and combination of different data sets, and therefore the pressure coordinate was considered as the most appropriate vertical coordinate for this study.

It is recognized that reanalyses, especially those of the older generation, suffer from incorrect mass fluxes associated with spurious winds, which, in turn, cause errors in advection of energy and momentum, including advection of latent energy (Trenberth, 1991; Graversen et al., 2007). Trenberth (1991) proposed a method for a mass correction, which has later been applied in several studies. Calculations of Dufour et al. (2016) for the residual term of moisture convergence in the Arctic (mean precipitable water times the mean meridional mass flux) gave indications that the non-closure of the mass budget might introduce an error of an order of 1%. In Text S2 we show that, for the latent energy transport, the impact of mass-correction is, indeed, very small in ERA-Interim. Based on these calculations and estimations, the mass-correction was neglected in this study.

Accurate wind and moisture fields are a prerequisite for a robust moisture transport study. Evaluation of moisture transport based on reanalyses against transport calculated from Arctic radiosonde data has been made by Dufour et al. (2016).
Their evaluation indicated that meridional net moisture transport was about 10% higher based on reanalyses than radiosonde soundings. The difference was mainly due to a higher moisture transport in reanalyses below 700-hPa level. Their results also indicates that the moisture transport profiles of ERA-Interim and NCEP CFSR (National Centers for Environmental Prediction, Climate Forecast System Reanalysis) correlated best with the transport profiles of radiosonde soundings averaged over the sounding sites in the comparison of seven reanalyses. Based on these previous evaluation results, we conclude that despite the slight overestimated net transport, reanalyses can well capture the main spatial and temporal patterns of moisture transport in mid- and high latitudes, providing the motivation to focus our analysis here on ERA-Interim data. We note that although radiosoundings yield the most accurate vertical profiles of atmospheric moisture, wind vector and pressure, the sounding network is sparse, and reanalyses are needed for circumpolar moisture transport calculations.

2.2 Methods

Vertically integrated moisture transport from the reanalyses was calculated from the Earth’s surface to the 300-hPa level by integrating moisture fluxes from the pressure level data. Data have been stored with 25-hPa vertical resolution from 1,000 to 750 hPa and with 50-hPa vertical resolution from 750 to 300 hPa. In addition, we used surface pressure data for defining lower boundary for vertical integral. Moisture flux \( f_{qi} \) between consecutive pressure levels \( p_i \) and \( p_{i+1} \) is

\[
f_{qi} = \frac{1}{2g} (q_i v_i + q_{i+1} v_{i+1}) (p_i - p_{i+1}),
\]

where \( q \) is specific humidity, \( v \) is meridional wind component, \( p \) is air pressure and \( g \) is the gravitational acceleration. Subscript \( i \) refers to the index of each pressure level and subscript \( j \) refers to the index of layer between consecutive pressure levels \( p_i \) and \( p_{i+1} \). The vertically integrated, meridional moisture transport between the surface and the 300-hPa level \( (F_q) \) is

\[
F_q = \sum_{j=1}^{n-1} f_{qj} = \sum_{i=1}^{n-1} \frac{1}{2g} (q_i v_i + q_{i+1} v_{i+1}) (p_i - p_{i+1}).
\]

The upper boundary \( (p_1) \) is the 300 hPa level and the lower boundary \( (p_n) \) was the actual surface pressure on each time step \( (P_{surface}) \). The pressure levels where pressure was higher than the surface pressure have been ignored. The flux in the lowest layer above the surface was calculated in such a way that the flux density of the lowest pressure level was used for the entire layer. For zonal moisture transport, \( v \) in Equations 1 and 2 was replaced by the zonal wind component, \( u \).

We decomposed the flux into parts of the mean circulation, SEs and TEs applying Reynolds decomposition. However, the first method of calculating moisture flux (Equation 1) could not be directly applied to accomplish a complete decomposition, because by applying Reynolds decomposition directly to specific humidity and wind, and taking vertical integrals after averaging over time and latitude, the effect of temporal and zonally varying surface pressure on the moisture flux was not correctly taken into account. This method would lead to incorrect estimates of moisture flux in the layers, where changing surface pressure affects the thickness of the layer. In the mass flux method, the contributions of varying wind and surface pressure are combined into a mass flux term and the Reynolds averaging is now applied on humidity and mass flux. Hereby the moisture transports by the mean circulation, SEs and TEs constitute a complete decomposition of the net moisture transport. Thus, we divided the moisture flux into products of specific humidity and mass flux and applied Reynolds decomposition to both components yielding the following equation:

\[
q f_{mj} = \left( [q_j] + q^* j \right) \left( [f_{mj}] + f^* m + f_{mj} \right),
\]

| TABLE 1 | Zonally averaged seasonal and annual means (mean) of vertically integrated (surface–300 hPa) meridional net moisture transport (kg m\(^{-1}\) s\(^{-1}\)), and its standard deviation (SD) between years based on ERA-Interim from time period 2003–2014 |
|---|---|---|---|---|---|---|
| 60° N | MMC Mean | SE Mean | TE Mean | Net transport Mean | | |
| Winter | 0.0 | 5.3 | 8.5 | 13.8 | | |
| Spring | −0.3 | 3.0 | 9.3 | 12.0 | 0.9 | |
| Summer | 0.9 | 1.3 | 11.9 | 14.1 | 1.1 | |
| Autumn | 0.7 | 3.8 | 12.9 | 17.4 | 1.2 | |
| Year | 0.3 | 3.2 | 10.5 | 14.0 | 0.6 | |
| 70° N | | | | | | |
| Winter | −0.4 | 1.7 | 4.2 | 5.4 | 1.2 | |
| Spring | −0.7 | 1.2 | 5.0 | 5.5 | 0.9 | |
| Summer | −1.7 | 1.8 | 9.8 | 9.9 | 1.2 | |
| Autumn | −0.8 | 1.3 | 6.8 | 7.3 | 0.9 | |
| Year | −0.9 | 1.5 | 6.4 | 6.9 | 0.5 | |
where $f_{m,j}$ is mass flux in the layer $j$ and $q_j$ is mean specific humidity in the layer $j$. The square brackets denote the zonal average and the overbar the time average over a month. $q'$ and $f_{m}'$ are deviations from the monthly mean, and $q^*$ and $f_{m}^*$ are deviations from the zonal average. The time averages are computed over a month. Taking the zonal and time mean of Equation 3 yields:

$$\left[q f_{m,j}\right] = \left[q_j\right] \left[f_{m,j}\right] + \left[q_j f_{m}^*\right] + \left[q^* f_{m,j}\right].$$

(4)

where the first term on the right-hand side is the moisture flux due to mean circulation, the second term is the moisture flux due to SEs, and the third term is the moisture flux due to TEs. The mass fluxes ($f_{m,j}$) and mean specific humidity ($q_j$) between consecutive pressure levels $p_i$ and $p_{i+1}$ were calculated as follows:

$$f_{m,j} = \frac{1}{2g} (v_i + v_{i+1})(p_i - p_{i+1}),$$

(5)
yielding the moisture flux calculated as follows:

$$q_i = \frac{1}{2}(q_i + q_{i+1}),$$  \hspace{1cm} (6)

yielding the moisture flux calculated as follows:

$$q_j f_m = \frac{1}{4g} (q_i + q_{i+1})(v_i + v_{i+1})(p_i - p_{i+1}).$$  \hspace{1cm} (7)

The zonal mass flux was calculated analogously to the meridional mass flux, using the zonal wind component instead.

We used the mass flux method to avoid the problem associated with varying surface pressure, and thereby to improve the agreement between the sum of all terms in the decomposition and the moisture transport calculated based on Equation 1. Nevertheless, the mass flux method (Equation 7), that is, the sum of moisture transports due to the mean circulation, SEs and TEs, does not yield exactly the same equation for moisture flux than the first method (Equation 1). In Equation 1, moisture flux densities are first calculated on pressure levels then multiplied by thickness of the layer, whereas in Equation 7, wind speed is first multiplied by thickness of the layer yielding mass flux in the layer, and then the product is multiplied by mean specific humidity inside the layer. Therefore, the sum of moisture transports due to the mean circulation, SEs and TEs did not exactly match the net moisture transport, which was calculated using the first method (Equation 1), and neither the sum of moisture transport due to the mean circulation, SEs and TEs was equal to net transport in Table 1, but the difference was always smaller than 5%. In the mass flux method, the mass flux and specific humidity are both individually averaged inside layers between the pressure levels. This method provides accurate results if the changes in the mass flux and specific humidity are linear between the pressure levels. On the other hand, the first method (Equation 1) only requires vertical averaging of the moisture flux and provides accurate results if the change in the moisture flux is linear between the pressure levels. Anyhow, the difference between these methods in zonally averaged meridional moisture transport was small but it shows the uncertainty related to calculation methods of moisture transport applying data from pressure levels.

In addition to moisture transport due to mean circulation, SEs and TEs, we examined the moisture flux caused by the product of deviations of mass flux from the zonal mean and the entire specific humidity field, including both parts, zonal mean and deviation from zonal mean of the specific humidity field. Accordingly, this transport consists of two parts: (a) the product of the deviations of specific humidity and mass flux from zonal means \((q_i f_m^*)\) and (b) the product of zonal mean specific humidity and the deviation of mass flux from the zonal mean \(\langle q_i f_m^* \rangle\). This approach allows the examination of spatial variation in moisture transport due to the stationary longitudinal variation of mass flux.

To represent the magnitude of meridional transport, we calculated the total moisture transport, which is a mean of the absolute values of the instantaneous meridional moisture transports, whereas the net transport is the mean of the instantaneous meridional transport. For a more precise view of temporal variations of moisture transports, we computed frequency distributions of the strength of moisture transport. The frequency distribution represents the relative frequencies of occurrence of moisture flux within magnitude intervals. In the case of vertically integrated moisture transport, the frequency distribution of moisture transport was calculated separately for every grid point, and separately for every grid cell in the case of the vertical profile of moisture transport. In the same way, the contribution of the moisture transport portions, within the individual magnitude intervals, to the whole moisture transport was also examined. The sum of absolute values of these portions is equal to the total...
moisture transport, and the difference between the positive (northwards) and negative (southwards) portions is equal to the net moisture transport.

3 | RESULTS

3.1 | Vertically integrated moisture flux

Horizontal moisture fluxes were typically vertically coherent in the sense that the vertically integrated moisture flux well represented temporal and spatial variations in the horizontal moisture flux at all vertical levels. Linear correlations of the time series between the grid point values of the vertically integrated moisture transport and the grid point values of the moisture transport on each level were high; and so were the correlations between the fields of the vertically integrated moisture transport and the moisture transport on each level at each analysis time. On average, the correlations mostly exceeded 0.9 between 700 and 800 hPa, and 0.8 between 500 and 950 hPa. The high correlations indicate that relationships of temporal and spatial anomalies between

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**FIGURE 3** Frequency distribution of the vertically integrated meridional moisture transport strength at individual grid points at 60°N (a) and 70°N (b) and contribution of each moisture transport strength interval to meridional net moisture transport at 60°N (c) and 70°N (d). The figures are based on 6-hourly analyses of ERA-Interim from time period 2003–2014.
Vertically integrated moisture transport and moisture transport on each level as well as relationships between anomalies of moisture transport on different levels were close to linear in low troposphere. Accordingly, the anomaly in vertically integrated moisture transport was typically contributed by almost analogous anomalies on all pressure level in low troposphere.

The general pattern of net moisture transport persevered throughout the year (Figure 1), even though the magnitude of moisture transport was larger in summer than other seasons (Figure 2). South of 75°N, moisture transport due to westerly winds characterized the transport (Figure 1). In particular, westerly winds transported moisture from the Atlantic via northern Europe to Siberia. Our result agree well with earlier studies (Cullather et al., 2000; Groves and Francis, 2002; Bengtsson et al., 2011; Dufour et al., 2016), depicting that the largest northwards meridional net moisture transports occurred over the eastern Atlantic and Pacific Oceans and over western Greenland. In winter, the transport was largest over the Atlantic and Pacific Oceans but in summer, the northwards moisture transport increased across western Alaska and near the Bering Strait as well as the west coast of Greenland. Seasonal mean values showed that southwards moisture transport occurred primarily over central Canada and on the eastern side of Greenland through the Fram Strait.

Total moisture transport (sum of the magnitudes of northwards and southwards directed moisture transports) was several times larger than the net transport. Accordingly, most of northwards moisture transport was balanced by southwards transport (Figure 2). The magnitudes of total and net transports were the largest in summer or early autumn. The maximum of monthly means of total transport was found in July at all latitudes examined, except at 60°N, where the maximum occurred in August. At 60°N and 65°N, the annual maximum of net moisture transport occurred in September, which agreed with the net moisture transport results of Dufour et al. (2016) for the locations of sounding stations north of 60°N based on radiosonde soundings and reanalysis products. However, further north at 70°N and 75°N, the annual maximum of net moisture transport occurred in July at the same month as the
maximum of total moisture transport. The seasonal cycle of total transport was larger than the seasonal cycle of net transport. The ratio between the annual maximum and the annual minimum was approximately 1.5 times larger for the total moisture transport than net moisture transport. Atmospheric moisture content has a strong seasonal cycle in high latitudes (Naakka et al., 2018), and the amount of total moisture transport was closely related to this cycle, because the total moisture transport is simply a product of the atmospheric moisture content and scalar meridional wind speed. However, the net transport was also affected by the dynamical setting (cyclonic activity and large-scale circulation pattern) and the humidity gradient between the pole and lower latitudes. These properties affected the moisture differences between cases of southerly and northerly moisture transport and, hence, the strengths of the transports themselves. This further shaped the seasonal cycle of net transport.
Moisture transport was mainly within ±100 kg m\(^{-1}\) s\(^{-1}\) in winter (91% of time at 60°N and 97% of time at 70°N) and within ±200 kg m\(^{-1}\) s\(^{-1}\) in summer (91% of time at 60°N and 96% of time at 70°N) but a relatively large proportion of moisture transport was associated to events of strong northwards and southwards transports (Figure 3c,d). At 60°N, northwards moisture transports with a magnitude exceeding 200 kg m\(^{-1}\) s\(^{-1}\) were responsible for 29% (42%) of northwards moisture transport in winter (summer), even though their relative frequency of occurrence was only 3% (7%). In winter, very intensive northwards moisture transports (exceeding 200 kg m\(^{-1}\) s\(^{-1}\)) occurred most frequently over the Atlantic and Pacific Oceans (Figure 4e). Southwards moisture transports with magnitude over 200 kg m\(^{-1}\) s\(^{-1}\) were almost absent in winter (Figure 4a). In summer, northwards and southwards moisture transports with a magnitude over 200 kg m\(^{-1}\) s\(^{-1}\) (Figure 4b,f) occurred across a much wider area than in winter. Accordingly, because of a higher specific humidity, intensive moisture transports were more common in summer than in winter.

In general, the northwards moisture transport was stronger than southwards transport. An analysis of moisture transport strength on individual grid points showed that, on average over the whole latitude band, the frequency of weak southwards transport was higher than the frequency of weak northwards moisture transport, but strong moisture transport occurred more often northwards than southwards (Figure 3a,b). However, the frequency of northwards and southwards moisture transport was practically equal. The relative frequency of northwards or southwards moisture transports had large spatial variation (Figure 4c,d) which, on one hand, was only weakly linked to the strength of the net moisture transport. The most permanent northwards moisture flux was located on the western side of Greenland, where the maximum frequency of northwards moisture transport was approximately 90%, but over the North Atlantic, where the strongest northwards moisture transport was located, the northwards moisture transport was only slightly more frequent than southwards moisture transport. Over the Atlantic Ocean northwards moisture transport was related to strong moisture transport both in southwards and northwards directions and a relatively small difference in their magnitude, whereas on the western side of Greenland strong northwards moisture transport was caused by relatively weak but very persistent northwards moisture transport (Figure 5). On the other hand, all areas where the net transport was directed to the south (Figure 1) collocated with areas where relative frequency of southwards transport exceeded 50% (Figure 4c, d), that is, the median transport was directed southwards. The most permanent southwards moisture flux was located in Northern Canada, where also the strongest net southwards moisture transport occurred.

Meridional moisture transport was divided into moisture transports due to the mean circulation, SEs and TEs. Most of the net moisture transport was caused by TEs, which were responsible for approximately 70% (90%) of the net meridional moisture transport at 60°N (70°N) (Table 1). The proportion of SE moisture transport on the net meridional moisture transport was 10–30%. The mean circulation had a small negative contribution on the net meridional moisture transport north of 60°N. A large part of the annual cycle of the net moisture transport was caused by TEs, especially at 70°N. Instead, moisture transport due to SEs and TEs had an approximately equal contribution to the inter-annual variation of the net moisture transport. The SE moisture transport

**FIGURE 6** Correlation of inter-annual time series of seasonal means (DJF, MAM, JJA, SON) from top to bottom in each column of numbers) between meridional net moisture transport (NET), moisture transport due to strong (>200 kg m\(^{-1}\) s\(^{-1}\)) northwards moisture transport events (STRONG), stationary eddy moisture transport (SE) and transient eddy moisture transport (TE). The figures are based on 6-hourly analyses of ERA-Interim from time period 2003–2014.

**Figure 3a,b** presents frequency distributions of moisture transport strength computed from grid point values. Vertically integrated meridional moisture transport was weak for most of the time, seen as a small frequency of occurrence of cases of strong northwards and southwards transport. Moisture transport was mainly within ±100 kg m\(^{-1}\) s\(^{-1}\) in winter (91% of time at 60°N and 97% of time at 70°N) and within ±200 kg m\(^{-1}\) s\(^{-1}\) in summer (91% of time at 60°N and 96% of time at 70°N) but a relatively large proportion of moisture transport was associated to events of strong northwards and southwards transports (Figure 3c,d). At 60°N, northwards moisture transports with a magnitude exceeding 200 kg m\(^{-1}\) s\(^{-1}\) were responsible for 29% (42%) of northwards moisture transport in winter (summer), even though their relative frequency of occurrence was only 3% (7%). In winter, very intensive northwards moisture transports (exceeding 200 kg m\(^{-1}\) s\(^{-1}\)) occurred most frequently over the Atlantic and Pacific Oceans (Figure 4e). Southwards moisture transports with magnitude over 200 kg m\(^{-1}\) s\(^{-1}\) were almost absent in winter (Figure 4a). In summer, northwards and southwards moisture transports with a magnitude over 200 kg m\(^{-1}\) s\(^{-1}\) (Figure 4b,f) occurred across a much wider area than in winter. Accordingly, because of a higher specific humidity, intensive moisture transports were more common in summer than in winter.

In general, the northwards moisture transport was stronger than southwards transport. An analysis of moisture transport strength on individual grid points showed that, on average over the whole latitude band, the frequency of weak southwards transport was higher than the frequency of weak northwards moisture transport, but strong moisture transport occurred more often northwards than southwards (Figure 3a,b). However, the frequency of northwards and southwards moisture transport was practically equal. The relative frequency of northwards or southwards moisture transports had large spatial variation (Figure 4c,d) which, on one hand, was only weakly linked to the strength of the net moisture transport. The most permanent northwards moisture flux was located on the western side of Greenland, where the maximum frequency of northwards moisture transport was approximately 90%, but over the North Atlantic, where the strongest northwards moisture transport was located, the northwards moisture transport was only slightly more frequent than southwards moisture transport. Over the Atlantic Ocean northwards moisture transport was related to strong moisture transport both in southwards and northwards directions and a relatively small difference in their magnitude, whereas on the western side of Greenland strong northwards moisture transport was caused by relatively weak but very persistent northwards moisture transport (Figure 5). On the other hand, all areas where the net transport was directed to the south (Figure 1) collocated with areas where relative frequency of southwards transport exceeded 50% (Figure 4c, d), that is, the median transport was directed southwards. The most permanent southwards moisture flux was located in Northern Canada, where also the strongest net southwards moisture transport occurred.

Meridional moisture transport was divided into moisture transports due to the mean circulation, SEs and TEs. Most of the net moisture transport was caused by TEs, which were responsible for approximately 70% (90%) of the net meridional moisture transport at 60°N (70°N) (Table 1). The proportion of SE moisture transport on the net meridional moisture transport was 10–30%. The mean circulation had a small negative contribution on the net meridional moisture transport north of 60°N. A large part of the annual cycle of the net moisture transport was caused by TEs, especially at 70°N. Instead, moisture transport due to SEs and TEs had an approximately equal contribution to the inter-annual variation of the net moisture transport. The SE moisture transport
had the largest contribution to inter-annual variation in winter and spring, whereas the TE moisture transport had the largest contribution in summer. Inter-annual variations of the seasonal mean SE and TE moisture transport partly balanced each other, seen as negative correlations between their strengths in all seasons (Figure 6). The correlation of inter-annual variation between time series of net moisture transport and SE moisture transport was strongest at 60°N in spring and at 70°N in winter, and smallest in summer at both latitudes. The correlation between meridional net moisture transport and TE moisture transport had an opposite seasonal cycle, being strongest in summer and weakest in either winter or spring.

Moisture transport caused by stationary zonal variation in mass flux was mostly responsible for the regional pattern of the mean meridional moisture transport (Figure 7a). This meridional moisture transport consisted of two components: (a) the product of the deviations of specific humidity and mass flux from zonal means \( (\bar{q}_f \bar{f}_m) \) and (b) the product of zonal mean specific humidity and the deviation of mass flux from the zonal mean \( (\bar{q}_f \bar{f}_m) \), but only the first component (Figure 7b) affected the zonally integrated net moisture transport. The second component had a regionally larger magnitude (Figure 7a) than the first component but its zonal mean was zero. The largest moisture fluxes related to the zonal variation of meridional mass flux occurred over eastern parts of Atlantic and Pacific oceans and adjacent coasts, where mountains or highlands steer westerly winds towards the north.

Moisture transport caused by TEs was mostly meridional (Figure 7c) related to the mean gradient of vertically integrated atmospheric moisture content. TE moisture transport was northwards almost everywhere north of 60°N, except near the north coast of Greenland in every season and in Siberia and Siberian sector of the Arctic Ocean in winter (not seen in the annual mean) where the atmospheric moisture content decreased southwards. A large northwards moisture transport due to TEs was located over oceans throughout the year, but in summer, the TE moisture transport was also large over continents.

### 3.2 Vertical profile of moisture transport

Vertically, the largest annual mean net moisture transport occurred in the 800–850-hPa layer at 60°N and 70°N (Figure 7b,d) but the net moisture transport had only a small vertical variation between 700 and 950 hPa. However, the zonally averaged meridional net moisture transport profile varied seasonally (Figure 9). The vertical maximum of zonally averaged net moisture transport was located the lowest (below the 900-hPa level) in autumn and winter at 55°–60°N and near the North Pole, and the highest (approximately the 825-hPa level) in spring and summer at 60°–70°N (Figure 9). The increase of altitude of the maximum transport in summer was a consequence of increased southwards moisture transport near the surface, that is, strengthening of the moisture transport of the lower branch of the mean meridional circulation. This occurred especially over continents (Figure 10), which suggests that surface...
evaporation caused strengthening of the southwards moisture transport. Moisture transport caused by the mean meridional circulation in summer was indeed twice as strong as its annual average (Table 1). In addition, the level of maximum net moisture transport had notable longitudinal variation (Figure 10). For example, at 70°N, the maximum moisture transport occurred lower than along the western side of Greenland.

The maxima of both mean northwards and mean southwards moisture transports were located lower (Figure 8a,c) than the maximum of net meridional moisture transport (Figure 8b,d). This was due to the cancellation of net

![Graphs showing vertical profile of contribution of each moisture transport strength interval on net meridional moisture transport at 60°N (a) and 70°N (c) and vertical profile of meridional net moisture transport at 60°N (b) and 70°N (d). The strength of moisture transport was calculated individually at each grid point. When the strength is positive (bars on right-hand side in (a, c)) moisture transport is northwards and when the strength is negative (bars on left-hand side in (a, c)) moisture transport is southwards. The figures are based on 6-hourly analyses of ERA-interim from time period 2003–2014 [Colour figure can be viewed at wileyonlinelibrary.com]](image-url)
moisture transport near the surface because of nearly equal southwards and northwards moisture transport components. The largest southwards and northwards moisture transport occurred in the 900–950-hPa layer. In the 950–1,000-hPa layer moisture transport was smaller than above it, which was partly because surface topography substantially decreased the moisture flux in the lowermost layer. Strong moisture transport events were also most frequent below the 900-hPa level (Figure 11). This suggested that a large part of the total moisture transport occurred in a rather shallow layer above the surface, even though the net moisture transport peaked higher. The mean southwards transport decreased upwards slightly faster than the mean northwards transport (Figure 8a,c). This was also the case for the frequency of occurrence of intensive moisture transports events, indicating that strong northwards moisture transport events had slightly larger vertical extent than southwards moisture transport (Figure 11). The relative difference between the southwards and northwards moisture transports was the largest in the 600–700-hPa layer at 60°N and 70°N. In this layer the net transport was 27% (23%) of the total transport at 60°N (70°N).

The Polar cell structure was recognizable in the zonally averaged mean meridional circulation (Figure 12a). Between 60°N and 80°N the mean meridional circulation resulted in southwards net moisture transport below approximately the 700-hPa level, and above it the net moisture transport was northwards. The zonally averaged SEs and TEs moisture transports were directed almost entirely northwards (Figure 12b,c). The vertical maximum of SE moisture transport was approximately at the 900-hPa level. This was due to the large spatial variation in the near-surface specific humidity, which was strongly affected by the Earth’s surface properties. However, the spatial variation of meridional wind speed increased upwards up to the 300-hPa level. TE moisture transport peaked approximately at the 800-hPa level at 60°N but the level of the maximum descended towards the north, and north of 80°N the transport peaked below the 900-hPa level. These results agreed with Dufour et al. (2016) who showed that the level of maximum net moisture transport due to SEs occurred lower than moisture transport due to TEs. The largest temporal variation in the specific humidity field occurred at higher altitudes than the largest zonal variation. This was consistent with the fact that the temporal variations are mostly caused by internal atmospheric dynamics, whereas the spatial variations are strongly affected by the varying specific humidity over the different surfaces.

4 | DISCUSSION AND CONCLUSIONS

Moisture transport between the Arctic and mid-latitudes was studied based on ERA-Interim reanalysis. Moisture transport was vertically coherent, seen as a high temporal and spatial correlation between the vertically integrated moisture transport and moisture transport on individual levels in most of the lower and middle troposphere. This allowed the examination of the spatial and temporal variation of moisture transport based on the vertically integrated moisture transports. As an annual mean, the total moisture transport was several times larger than the net moisture transport. Accordingly, most of the northwards moisture transport was balanced by southwards moisture transport. This was the case especially near the surface, and therefore the meridional net moisture transport peaks at a higher altitude than both northwards and southwards moisture transports. The altitude of the maximum annual mean net moisture transport was located below the 800-hPa level, which agrees with the results of Dufour et al. (2016). On the circumpolar scale, the level of maximum meridional net moisture transport indicated the level of maximum moisture flux convergence. However, the local moisture flux convergence, which is important for the local humidity profile, and further for cloud formation, is influenced by the local temporally varying moisture transport, which largely deviates from the zonal mean. Regionally, zonal moisture transport is important for humidity conditions, in particular in areas such as northern Europe and western Russia, where the zonal moisture
transport has a much larger magnitude than the meridional component.

Most of time the moisture transport was weak in both southwards and northwards directions, but strong transport events were more frequent northwards than southwards. Our results, as well as previous studies (Woods et al., 2013; Liu and Barnes, 2015), show that strong moisture transport events caused a relatively large contribution of moisture transport, which emphasizes their importance. This is also in line with a concept of atmospheric rivers, narrow areas of strong moisture transport, which are responsible for a large part of the northwards moisture transport (Zhu and Newell, 1998). In winter, strong moisture transports events occurred over the Atlantic and Pacific oceans, collocating with the

FIGURE 10  Cross section of seasonal mean (DJF, MAM, JJA, SON) meridional net moisture transports at 60°N and 70°N. The figures are based on ERA-Interim from time period 2003–2014.
areas of extreme northwards moisture transport identified by Woods et al. (2013) and Liu and Barnes (2015).

A large part of the meridional net moisture transport was caused by TE moisture transport and the proportion of TEs was larger in summer (84%/99% at 60°N/70°N) than in winter (62%/78% at 60°N/70°N) (Table 1). These results are in line with previous studies, which have suggested that approximately 90% of the annual net moisture transport (Dufour et al., 2016), or approximately 80% of net moisture transport in winter and 90% in summer (Jakobson and Vihma, 2010) at 70°N is caused by TEs. However, the regional pattern of meridional moisture transport is mostly caused by stationary zonal variation in meridional mass flux. The seasonal cycle of the net moisture transport was largely explained by variations in TE moisture transport. At 70°N, the net moisture transport and moisture transport due to TEs peaked in summer, whereas at 60°N they both peaked in autumn. Oshima and Yamazaki (2006) suggested that the meridional moisture gradient is mostly responsible for the seasonal cycle of the TE moisture transport to the Arctic. They showed that the TE moisture transport is dependent on the meridional moisture gradient, which is largest in summer, and cyclonic activity, which is largest in winter. The difference between latitudes 60°N and 70°N was probably due to the meridional moisture gradients nearby those latitudes. The meridional moisture gradient has a much larger seasonal cycle nearby 70°N than 60°N because of geographical conditions: 70°N is near the coast of the Arctic Ocean and in summer the specific humidity is much higher over warm continents than over the Arctic Ocean.

The largest part of inter-annual variation of net moisture transport in winter and spring was due to SE moisture transport, whereas in summer the contribution of TE moisture was the largest. This is in agreement with the results of Jakobson and Vihma (2010), who found that the inter-annual variations of SE moisture transport was often larger than the inter-annual variations of TE moisture transport, especially in winter. In our analysis, SE transport includes all spatial variation in meridional mass flux and specific humidity, whose timescales were longer than a month. Large inter-annual variation of SE moisture transport indicated that SE transport is not produced only because of geographical features, such as land-sea variation or high topography, but it is largely affected by large-scale atmospheric dynamics. Our results indicated a positive correlation between inter-annual time series of meridional net moisture transport and northwards moisture transport of strong moisture transport events (strength more than 200 kg m$^{-1}$ s$^{-1}$) and a negative correlation between SE and TE moisture transports in all seasons. In winter, the correlations between the inter-annual time series of SE moisture transport and both net moisture transport and strong transport events were positive (Figure 6). Accordingly, in winter large-scale circulation patterns permitting strong moisture transport events and large meridional moisture transport are likely a permanent enough feature so that they also favour the moisture transport by SEs instead of TEs. The permanent features in circulation patterns may be related to blocking situations or Rossby wave breaking, which was consistent with results of Woods et al. (2013)
and Liu and Barnes (2015) who showed that the occurrence of strong moisture transport events was related to blocking situations or Rossby wave breaking. In summer, the net meridional moisture transport and the northwards moisture transport due strong moisture transport events are both correlated with the TE moisture transport.

Comparison of moisture transport between ERA-Interim and JRA-55 showed relatively small differences between the reanalyses (Text S1). The twice-daily temporal resolution of AIRS Level3 specific humidity profiles were averaged to produce daily mean profiles; therefore, the combined product of AIRS specific humidity profiles and ERA-Interim mass fluxes was not directly comparable with reanalysis. Using the daily averaged time resolution of AIRS instead of reanalyses (4 analyses per a day) resulted in an underestimation of TE moisture transport, because a significant part of this moisture transport is produced by variations that act on timescales shorter than 1 day. Accordingly, the temporal resolution of AIRS was not sufficient for moisture transport calculation. However, the larger difference between ERA-Interim and CAE (using time resolution 1 analysis per day in both products) than ERA-Interim and JRA-55 (using time resolution 4 analyses per day in both products) suggested that the differences in specific humidity fields are able to cause relatively large uncertainty in moisture transport and therefore, the difference between the reanalyses may not represent the entire uncertainty of reanalyses (Text S1).

We showed that the meridional net moisture transport was a relatively small part of the total meridional moisture transport, and the profiles varied spatially, seasonally, and even slightly offset vertically. This suggests that noticing spatial and seasonal variations of moisture transport as well as identifying the differences between the total and the net moisture transport is important for understanding the interaction of moisture transport and vertical humidity profiles as well as physical processes associated with atmospheric moisture. Inter-annual variations of moisture transport showed linkages between the net, SE and TE moisture transports as well as the moisture transport during strong events. For better understanding of the inter-annual variations of moisture transport, it is important to examine the circulation patterns responsible for the moisture transport due to SEs and TEs.

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REFERENCES


**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section at the end of the article.

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Horizontal Moisture Transport Dominates the Regional Moistening Patterns in the Arctic

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ABSTRACT

Along with the amplified warming and dramatic sea ice decline, the Arctic has experienced regionally and seasonally variable moistening of the atmosphere. Based on reanalysis data, this study demonstrates that the regional moistening patterns during the last four decades, 1979–2018, were predominantly shaped by the strong trends in horizontal moisture transport. Our results suggest that the trends in moisture transport were largely driven by changes in atmospheric circulation. Trends in evaporation in the Arctic had a smaller role in shaping the moistening patterns. Both horizontal moisture transport and local evaporation have been affected by the diminishing sea ice cover during the cold seasons from autumn to spring. Increases in evaporation have been restricted to the vicinity of the sea ice margin over a limited period during the local sea ice decline. For the first time we demonstrate that, after the sea ice has disappeared from a region, evaporation over the open sea has had negative trends due to the effect of horizontal moisture transport to suppress evaporation. Near the sea ice margin, the trends in moisture transport and evaporation and the cloud response to those have been circulation dependent. The future moisture and cloud distributions in the Arctic are expected to respond to changes in atmospheric pressure patterns; circulation and moisture transport will also control where and when efficient surface evaporation can occur.

1. Introduction

During the recent decades, the Arctic has experienced drastic changes, the most recognized being the amplified warming and dramatic decline in sea ice concentration and thickness (IPCC 2019). The horizontal moisture transport and especially the associated increase in downward longwave radiation have been recognized among factors contributing to these major changes (Kapsch et al. 2013; Park et al. 2015; Graversen and Burtu 2016; Kapsch et al. 2016; Gong et al. 2017; Lee et al. 2017; Yang and Magnusdottir 2017; Dai et al. 2019; Hao et al. 2019). An example of impacts of moisture transport is the spring onset of surface melt on Arctic sea ice, which is predominantly driven by the increased downward longwave radiation associated with increased moisture transport to a region (Maksimovich and Vihma 2012; Persson 2012; Mortin et al. 2016; Cao et al. 2017).

Furthermore, years with anomalously large poleward moisture transport in the spring have been followed by an anomalously low autumn sea ice concentration (Kapsch et al. 2013; Mortin et al. 2016). On daily time scales, moisture transport largely determines the spatial distributions of water vapor, cloud water, and surface downward longwave radiation in the Arctic (Devasthale et al. 2012; Nygård et al. 2019).

Climate model simulations have demonstrated that horizontal moisture transport responses to an increase in CO₂ concentration (Hwang et al. 2011), but the response is strongly seasonal (Singh et al. 2017). In previous studies, which have addressed the Arctic as a whole, the estimates of the net moisture transport trend have ranged from a decreasing, although statistically insignificant, trend across 70°N in 1979–2013 (Dufour et al. 2016) to a statistically significant, positive trend across 60°N since 1959 (Villamil-Otero et al. 2018). In winter, there has been a significant increase in the number of intensive moisture intrusions (Woods and Caballero 2016), which contribute substantially to the moisture transport to the Arctic (Woods et al. 2013; Liu and Barnes 2015; Baggett et al. 2016).

In general, temporal changes in moisture transport can be attributed to some combination of changes in
atmospheric circulation (Woods et al. 2013; Zhang et al. 2013; Vázquez et al. 2016; Gong and Luo 2017; Yang and Magnusdóttir 2017; Kapsch et al. 2018; Zhong et al. 2018; Nygård et al. 2019), evaporation especially in the seasonally varying key source regions (Singh et al. 2017; Gimeno-Sotelo et al. 2018), and removal of water vapor by condensation. Atmospheric circulation patterns are responsible for determining whether the moist air masses from the source areas find their way to the Arctic, and for determining the regional variability of this transport. Midlatitude cyclone activity has a critical role in this (Villamil-Otero et al. 2018). Interactions between atmospheric circulation and moisture transport are in fact acting in two directions. For example, in the Barents and Kara Seas, recent studies have identified a positive feedback loop between Ural blocking and Arctic moisture transport; an increased frequency of the Ural blocking enhances poleward moisture transport and sea ice decline, while also regionally reducing the background meridional temperature gradient (Luo et al. 2016; Gong and Luo 2017; Zhong et al. 2018). This regional reduction in the meridional temperature gradient further enhances background conditions favorable for an increased frequency of the Ural blocking.

Here we address the regional and seasonal trends in the horizontal moisture transport, redistributing the atmospheric moisture in the Arctic during the last four decades, in 1979–2018. We demonstrate the dominant role of moisture transport for determining the regional moistening patterns and causing regionally uneven greenhouse effect. We also show that changes in local evaporation in the Arctic, although they previously received much attention (Screen and Simmonds 2010; Bengtsson et al. 2011; Bintanja and Selten 2014; Boisvert and Stroeve 2015; Boisvert et al. 2015; Morrison et al. 2018), have only had a minor role in shaping the moistening patterns. Compared to previous studies (Bintanja and Selten 2014; Dufour et al. 2016; Villamil-Otero et al. 2018), the advantage of our approach is that 1) the role of horizontal moisture transport is not only limited to transport to and from midlatitudes across a certain latitudinal belt (e.g., 70°N) and 2) evaporation in the Arctic is not only counted as a circumpolar mean. Accordingly, we specifically address the regionally varying moisture transport and evaporation within the Arctic.

2. Data and methods

a. Reanalysis data

The study is based on the most modern global reanalysis ERA5 (Copernicus Climate Change Service 2017), produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). In ERA5, a variety of atmospheric observations (e.g., from radiosoundings and satellites) have been assimilated into a numerical weather prediction model, applying a four-dimensional variational data assimilation method. The spectral model resolution of ERA5 is T639, and the horizontal grid of the data used in this study is 0.25° × 0.25°. The amount of vertical model levels is 137. Our analysis is based on the surface level and vertically integrated products of the reanalysis at 6-h intervals. Northward and eastward components of vertically integrated moisture transport are direct output variables of ERA5, calculated for the air column from the surface to the top of the atmosphere. ERA5 is the fifth generation of ECMWF atmospheric reanalyses, published in 2018. According to the producers of ERA5, strengths of ERA5 compared to its predecessor ERA-Interim (Dee et al. 2011) include its much higher spatial resolution, better global balance of precipitation and evaporation, and more consistent sea surface temperature and sea ice. ERA5 is generally warmer than ERA-Interim in summer and autumn over Arctic sea ice, although both of these reanalyses have a warm bias (Wang et al. 2019). ERA5 provides the best estimates of downward radiative fluxes in spring and summer, suggesting more realistic representation of Arctic cloud cover in it (Graham et al. 2019). Based on our own analyses, we found that the imbalance between moisture and cloud water convergence compared to the net precipitation (precipitation minus evaporation) in ERA5 is rather small in the region north of 50°N; the convergence is 6% larger than the net precipitation.

It has been previously reported that different reanalyses mostly agree on the sign of trend in precipitable water, but not in its magnitude (Rinke et al. 2019). In the online supplemental material, the trends in mean sea level pressure (MSLP) and moisture transport in ERA5 are compared to those in ERA-Interim (Dee et al. 2011), the Japanese Meteorological Agency (JMA) 55-Year Reanalysis (JRA-55) (Kobayashi et al. 2015), and radiosonde observations. The comparison indicates that the MSLP and moisture transport trends are consistent in these different datasets, which suggests that our conclusions on these trends are robust. However, uncertainty related to trends in evaporation and total cloud water in ERA5 is expected to be higher. The total cloud water is especially vulnerable to errors, due to possible misrepresentation of cloud-related processes in the reanalysis (Vihma et al. 2014). Nevertheless, the uncertainties of evaporation and total cloud water in ERA5 are assumed to be small enough to allow for an assessment of the directions of their long-term changes.
b. Trend analyses

Linear trends in seasonal variables were calculated from the annual seasonal mean values of the variables at each grid point of the reanalyses. The traditional seasonal division into the 3-month periods (winter: December–February (DJF), spring: March–May (MAM), summer: June–August (JJA), and autumn: September–November (SON)) provided optimal groups for characterizing trends in MSLP and moisture transport. The trends were analyzed for three time periods: 1979–2018 (40 years), 1979–98 (20 years), and 1999–2018 (20 years). The 40-yr period indicates the direction of the long-term change, whereas the 20-yr periods give indications whether the trends have changed over the time. The annual average of Arctic sea ice extent was approximately 10% lower during the latter 20-yr period compared to the first 20-yr period. Statistical significance of linear trends was tested with the two-tailed Student’s t test, applying the 90% confidence level, which is commonly applied in studies related to atmospheric circulation and cyclones (Rudeva and Simmonds 2015; Zahn et al. 2018). As the horizontal moisture transport is a vector, with a magnitude and direction, the statistical significance was tested for the meridional and zonal components separately, and the trend was considered as statistically significant if at least one of the components had a statistically significant trend. The trend in interannual variations of net horizontal moisture transport (see Fig. S4 in the online supplemental material) was calculated by first determining how much each value of the detrended time series of annual seasonal means deviated from the detrended 40-yr seasonal mean, and then calculating a linear trend based on the absolute values of the annual deviations. Linear correlations between moisture and cloud variables were calculated from their time series of annual seasonal means, without detrending. The correlations were considered statistically significant when the 90% confidence level was reached.

3. Results

a. Changes in atmospheric circulation and moisture transport

Regional trends of moisture transport, and in particular trends in its direction, are strongly associated with the trends in mean sea level pressure (MSLP). This is because wind direction and speed in the lower troposphere, where most of the atmospheric moisture is located and being transported (Naakka et al. 2019), are largely regulated by MSLP patterns (Fig. 1). The distribution of MSLP in the Arctic has changed during the last four decades (Fig. 2). In Fig. 2, the trends in MSLP are shown with a color shading, and the trends in moisture transport are visualized with trend vectors, which contain information about the change in magnitude and direction of the transport. These trends (Fig. 2) need to be interpreted together with the mean moisture transport (Fig. 1) to perceive whether these trends have actually increased or weakened the moisture transport to a certain direction. Linear trends in MSLP for the 40-yr period (1979–2018) are on the order of ±1 hPa decade$^{-1}$, and for the latter 20-yr period (1998–2018) on the order of ±3 hPa decade$^{-1}$; seasonal and regional variability in MSLP trends is large. Regional trends in the magnitude of moisture transport have been drastic and coincident with the changes in the circulation, being at largest more than 10 kg m$^{-1}$ s$^{-1}$ decade$^{-1}$.

In winter (DJF), the 40-yr trend of MSLP is characterized by increasing MSLP in Eurasia and decreasing MSLP in the North Atlantic (Fig. 2a). During the latter 20-yr period, these trends have been stronger and more confined to the western part of Russia and the Barents Sea, and in contrast to the 40-yr period, MSLP in the Pacific sector has increased by 3 hPa decade$^{-1}$ (Fig. 2e). In the Atlantic sector, these trends have increased the zonal pressure gradient between the North Atlantic and Eurasia, and as a consequence, more moisture is transported over the North Atlantic and Scandinavia toward the pole. The regional trends in the magnitude of net moisture transport in the Arctic are substantial as the meridional and zonal components separately, and the trend was considered as statistically significant if at least one of the components had a statistically significant trend. The trend in interannual variations of net horizontal moisture transport (see Fig. S4 in the online supplemental material) was calculated by first detrending how much each value of the detrended time series of annual seasonal means deviated from the detrended 40-yr seasonal mean, and then calculating a linear trend based on the absolute values of the annual deviations. Linear correlations between moisture and cloud variables were calculated from their time series of annual seasonal means, without detrending. The correlations were considered statistically significant when the 90% confidence level was reached.

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moisture transport on its eastern side, and the decrease in MSLP in the Barents and Kara Seas, enhancing the zonal moisture transport along the Russian coast. The main difference is the increase in MSLP during the latter 20-yr period over central and eastern Russia, which has further strengthened the moisture transport along the coast and made the circulation over the Arctic Ocean more cyclonic. The 40-yr summer trends in moisture transport correspond to $-4\%$ to $+8\%$ change per decade compared the mean magnitude, and the recent 20-yr trends to $-16\%$ to $+15\%$.

In autumn (SON), MSLP over Greenland has had a negative trend, which has become stronger during the latter 20-yr period (Figs. 2d,h). The positive trend in MSLP over Scandinavia is seen during both of these periods, whereas increase in MSLP over North America and central Siberia has only occurred during the latter 20-yr period. These changes have significantly strengthened the meridional moisture transport from the North Atlantic all the way to the central Arctic Ocean. At the same time, interannual variations of moisture transport magnitude have become larger in the North Atlantic (Fig. S4). Over Russia, the magnitude of dominantly zonal moisture transport has decreased, but mostly statistically insignificantly.

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**Fig. 1.** Seasonal mean sea level pressure in (a) December–February, (b) March–May, (c) June–August, and (d) September–November in 1979–2018 in ERA5. Vectors represent the mean vertically integrated net horizontal moisture transport.
The trends in moisture transport have thus distinct characteristics in all four seasons, but all these trends are largely driven by changes in atmospheric circulation. The seasonal trends can be briefly summarized as follows: Winter has been characterized by enhanced northward moisture transport in the Atlantic and Pacific sectors, and spring by the increased moisture transport across the Arctic Ocean. In summer, the zonal moisture transport has been intensified along the continental coasts, whereas in autumn the role of the moisture transport from the North Atlantic has increased.

b. Interactions with trends in total column water vapor and local evaporation

The role of atmospheric circulation in distributing the moisture in the Arctic is evident: Positive trends in total column water vapor, which is defined as the total amount of water vapor in an atmospheric column, accompany the wind direction trends along the main moisture transport pathways, especially during the cold seasons: winter, spring, and autumn (Fig. 3). For example, in autumn, the trend toward more northward wind and moisture transport from the North Atlantic, together with increased northward transport from the Pacific sector have been associated with an increase in total column water vapor over most of the Arctic sea areas (Figs. 2d,h and 3d,h). During all the seasons, except in summer, annual means of total column water vapor are strongly correlated (correlation coefficient $r > 0.8$) with the magnitude of the net moisture transport. Even if the close interaction between these variables is apparent, it is difficult to determine the relative importance of circulation changes that enhance moisture transport versus an increase in moisture available for transport. Such a distinction is outside of the scope of this paper as the focus here is on comparison of the roles of transport driven or enhanced by the large-scale circulation, and local evaporation.

Changes in evaporation have also contributed to the trends in total column water vapor. Comparison of the earlier 20-yr period with the latter one clearly indicates that the area with a strongly positive evaporation trend (Fig. 4) is mostly limited to the marginal ice zone (Fig. 5) during the cold seasons (winter, spring, and autumn). In winter, the positive trend in evaporation has moved from the central Barents Sea (Fig. 4e) to the northern Barents Sea and Kara Sea (Fig. 4i), following the retreat of sea ice (Figs. 5a,e). In autumn, the positive trends in evaporation have occurred on the coastal seas off Russia and North America (Figs. 4d,h,l).
Over the sea areas, the marginal ice zone is practically the only region where the annual means of evaporation and total column water vapor in winter, spring, and autumn correlate positively ($r > 0.6$) (Fig. 6b; see also Figs. S5b,g); however, the mean evaporation there is small. This positive correlation may be causally related to evaporation, or it may also be a consequence of interactions between horizontal moisture transport and sea ice retreat. Analyses support the latter one, as the correlation between evaporation and moisture transport magnitude is positive in the marginal ice zone, but mostly negative or small in the other sea areas (Fig. 6c and Figs. S5c,h). Furthermore, correlation between moisture transport magnitude and sea ice fraction in the marginal ice zone is negative (not shown). This correlation clearly indicates that in the marginal ice zone, winters, springs, and autumns with strong moisture transport have been associated with a larger sea ice retreat and thus with a warmer surface. However, it remains as an open question whether 1) the sea ice retreat has actually been caused by the increased moisture transport [by decreasing the sea ice growth (Persson et al. 2017) and by the associated southerly winds pushing the ice edge northward (Stroeve and Notz 2018)] or 2) the increased horizontal moisture transport is a consequence of the warmer surface allowing for larger moisture transport due to reduced cooling and condensation. In any case, the sea ice retreat has also allowed for an increase in surface evaporation, which has provided additional moisture available for transport. Hence, both increased transport and evaporation have probably increased total column water vapor in the marginal ice zone.

South of the marginal ice zone, in the open water areas where the mean evaporation is relatively high in the cold seasons, the trend in evaporation is mostly negative, although not everywhere statistically significant (Fig. 4). Here, evaporation correlates negatively with total column water vapor (Fig. 6b and Figs. S5b,g), suggesting that winters, springs, and autumns with enhanced evaporation have occurred when the air has been relatively dry allowing for efficient evaporation. A negative correlation between evaporation and total column water vapor is also seen in summer (Fig. 6g), although the mean evaporation over the sea is then small. Increases in total column water vapor in the open water areas are thus likely associated with circulation and moisture transport changes rather than local evaporation. A clear
example of this relationship is seen in the area south of Greenland in winter, where evaporation has a trend of more than +0.3 mm day$^{-1}$ decade$^{-1}$ (Fig. 4i), negative correlation between evaporation and total column water vapor is particularly strong ($r = -0.8$) (Fig. 6b), and also evaporation and moisture transport are anti-correlated (Fig. 6c).

In contrast to the cold seasons, most of summertime evaporation is taking place on land (Vihma et al. 2016). Over land, summers with high evaporation have typically been associated with a high amount of total column water vapor (Fig. 6g) but a low amount of total cloud water (Fig. 6j), suggesting that evaporation is most efficient when the reduction of clouds allows more surface radiative heating. However, regional trends of evaporation in summer (Figs. 4c,k) have not directly been translated into increased total column water vapor (Figs. 3c,g).

We summarize the interactions between moisture transport, surface evaporation, and total column water vapor as follows: Probably both increased moisture transport and local evaporation have increased total column water vapor at the marginal ice zone, whereas evaporation is of limited importance elsewhere. Over the open sea, years with a high amount of total column water vapor have been linked to large moisture transport but relatively low evaporation. In general, large evaporation strengthens the horizontal transport of moisture, whereas large horizontal moisture transport tends to suppress local evaporation by decreasing the
humidity difference between the surface and the air above.

c. Trends in relative humidity and cloud water

Long-term changes in atmospheric moisture in the Arctic have also included changes in relative humidity (RH) in the whole air column from the surface to 300 hPa. During the latter 20-yr period, RH at 850 hPa has had positive trends in the regions, where the mean sea level pressure has had a negative trend (Fig. 7), suggesting that increased cyclonic activity and/or weaker high pressure patterns with reduced subsidence have been associated with the increased RH. Averaged over the whole Arctic (from 60°N northward), the 40-yr trend in RH has been slightly negative at all pressure levels, whereas the 20-yr trend has been negative below 950 hPa and above 500 hPa and positive at 950–500 hPa. The magnitude of 40- and 20-yr trends for the whole Arctic has been relatively small (on the order of 0.5% decade⁻¹), but the largest regional and seasonal trends of RH are on the order of 5% decade⁻¹.

The largest regional trends in total cloud water (Fig. 8) are collocated with the largest trends in total column water vapor (Fig. 3). In the trends of total cloud water (Fig. 8), there is a sharp division between the marginal ice zone and open water area south of it (Fig. 5). In the marginal ice zone, winter, spring, and autumn means of total cloud water correlate positively with moisture.

**Fig. 5.** Seasonal mean sea ice concentration in ERA5 for the periods (top) 1979–98 and (middle) 1999–2018, and (bottom) the difference in concentration between those two periods (1979–98 minus 1999–2018). The results are divided into (a),(e),(i) winter, (b),(f),(j) spring, (c),(g),(k) summer, and (d),(h),(l) autumn.
FIG. 6. Linear correlations between annual means of (a),(f) moisture transport magnitude and total column water vapor, (b),(g) evaporation and total column water vapor, (c),(h) moisture transport magnitude and evaporation, (d),(i) moisture transport magnitude and total cloud water, and (e),(j) evaporation and total cloud water in ERA5 during the period 1979–2018, divided into winter in (a)–(e) and summer in (f)–(j). The arrows indicate the variables included in the correlations. The dotted areas denote the correlations that are statistically significant ($p < 0.10$). The gray areas mask the regions where the seasonal mean magnitude of evaporation is smaller than 0.5 mm day$^{-1}$. 
transport magnitude and evaporation (Figs. 6d,e and Figs. S5d,e,i,j), suggesting that the positive trend in cloud water (Fig. 8) is in most regions linked to the displacement of the sea ice zone. This is the zone where northward-transported moisture meets the cold sea ice surface, leading to lifting due to the upward-tilting isentropes (Komatsu et al. 2018), cooling (Vihma et al. 2003), and condensation; the lifting does not strictly occur along the isentropes, due to diabatic processes. In the same zone, southward-moving cold and dry air masses from the sea ice trigger high evaporation and convective cloud forming when meeting the open water. On the open water side in the Barents Sea, the total cloud water has had a negative trend during the cold
seasons. This is the only area, except regions with orographically induced cloud decay, where annual winter means of net moisture transport magnitude only weakly correlate with total cloud water (Fig. 6d and Figs. S5d,i). The weak correlation is presumably due to flow dependency, because the on-ice and off-ice flows may have compensating effects on cloud water (see section 4).

It is thus evident that long-term changes in atmospheric circulation and moisture transport have also been reflected to the trends in RH and total cloud water, and further to trends in downward longwave radiation at the surface, which are discussed in the online supplemental material.

4. Discussion

Our results show that the regional moistening patterns in the Arctic (Rinke et al. 2019) have predominantly been shaped by trends in moisture transport rather than trends in local evaporation. Trends in moisture transport are associated with changes in the atmospheric circulation and have most likely arisen from a combination of inherent variability of atmospheric dynamics (Ding et al. 2017; Gong et al. 2017) and response to the increased amount of anthropogenic greenhouse gases (Hwang et al. 2011; Singh et al. 2017). We have demonstrated that not only local evaporation but also the magnitude of horizontal moisture transport is affected by the diminishing sea ice cover during the cold seasons, from autumn to spring. Retreat of the sea ice from a region causes an evident stepwise increase in evaporation, due to removal of the insulating layer. This increase has been visible in previous studies in which evaporation trend from a single period (Boisvert et al. 2015) has been addressed, or mean evaporation amounts of two periods (Bintanja and Selten 2014; Singh et al. 2017) have been compared. This study is the first to show a negative trend in open water evaporation after the sea ice has disappeared from a region. Hence, the positive trends in evaporation during a local sea ice decay are restricted to the vicinity of sea ice margin over a limited period. Negative trends in wintertime evaporation, due to reduced surface-air specific humidity difference, have previously been reported by Boisvert et al. (2013) in the Kara/Barents Seas, East Greenland Sea, and Baffin Bay regions where there is
open water year round. We suggest that the negative trend in evaporation in open water areas, including the newly opened areas, is tightly linked to the influence of moisture transport to suppress the local evaporation (Park et al. 2015; Nygård et al. 2019).

In the vicinity of the marginal ice zone, the trends and impacts of moisture transport and evaporation are dependent on the flow direction. In a schematic figure (Fig. 9), we summarize the current understanding and the new results of flow dependency of the trends in moisture transport and evaporation, and the cloud response to those. When the flow is from the open ocean, the retreat of sea ice (Figs. 9a,c) is not associated with major changes in the open water zone (zone 1 in Fig. 9c), whereas the zone from where the sea ice has been removed (zone 2 in Fig. 9c) experiences higher evaporation due to the warmer surface. Also the moisture transport is larger, as the evaporation and lack of air mass cooling and condensation allow for more water vapor in the air. Low clouds and fog, forming due to cooling downstream of the ice edge, are displaced farther north to the retreated sea ice zone (zone 3 in Fig. 9c). The moisture transport over this retreated sea ice zone (zone 3) (Fig. 9c) is larger than previously in this zone (Fig. 5a), because the air mass has had shorter time and fetch to gradually cool and dry.

Conversely, when the flow is from the sea ice zone (Figs. 9b,d), the retreat of sea ice induces much higher evaporation in the newly opened sea zone, leading to convective cloud formation (Kay and Gettelman 2009) and increased southward moisture transport (Fig. 9d). Originally cold and dry air mass rapidly moistens over the warmer sea surface. Consequently, farther on the open sea, evaporation is not as efficient as near the sea ice edge, but moisture transport is larger (zone 1 in Fig. 9d).

The increased amount of clouds in the vicinity of the marginal ice zone has been in many previous studies attributed solely to increased local evaporation and strong air–sea coupling (Kay and Gettelman 2009; Boisvert et al. 2015; Morrison et al. 2018; Taylor et al. 2018; Morrison et al. 2019), but we emphasize the fundamental role of moisture transport and the flow dependency of the cloud response. An example of the flow dependency is that the moisture transport magnitude presumably increases in the newly opened sea (zone 2 in Figs. 9c,d) both in southward and northward flows, but these flows are most likely linked to opposing effects on cloud water in this zone. Indeed, the weak correlation between cloud water and moisture transport magnitude (Fig. 6d and Fig. S5) suggests that these processes related to cloud formation partly compensate each other.

Generally, an increase of total column water vapor with the increasing temperature in the Arctic has been expected, as a consequence of the Clausius–Clapeyron relation. However, the increase in the Arctic moisture is not only a direct consequence of the temperature increase. On one hand, locally originating increase of surface temperature (e.g., due to decreased sea ice concentration) allows for increases in local evaporation, sensible heat flux, and air temperature, enabling an increase in moisture transport at the location. On the other hand, temperatures in many Arctic subregions are driven by midlatitude circulation (McGraw and Barnes 2020). Possibilities for poleward transport of an air mass that is both warm and dry are very limited in the cold seasons and effective moisture transport from the mid-latitudes has to be associated with transport of warm air, as cold air cannot hold and carry moisture. As a consequence, temperature and moisture in the Arctic are highly correlated. However, we want to emphasize that moisture content is not a passive component responding to the temperature changes in the Arctic; a large body of evidence has demonstrated the two-way feedback between atmospheric moisture and temperature (Kapsch et al. 2013; Park et al. 2015; Graversen and Burtu 2016; Kapsch et al. 2016; Gong et al. 2017; Lee et al. 2017; Yang and Magnusdottir 2017; Dai et al. 2019; Hao et al. 2019). If transport of moist and inevitably warm air increases to a region, it is seen as an increase in temperature not only due to the direct effect of transport of dry static energy but also because of the release of the transported latent heat. In addition, temperatures along the moisture transport path are typically further affected by radiative effects of water vapor and clouds (Nygård et al. 2019). Hence, individual impacts of temperature and moisture are difficult to isolate, and the increase in the Arctic moisture is not only a direct consequence of the temperature increase. Furthermore, we have shown, based on our analyses (section 3c), that the often-made assumption of constant relative humidity in the changing climate is not valid at regional and local scales, especially when MSLP and occurrence of certain circulation patterns in a region have changed.

5. Conclusions

In this paper, we have addressed the regional and seasonal trends in the horizontal moisture transport in the Arctic during the last four decades (1979–2018). We have also compared the roles of moisture transport and local evaporation for the moistening of the Arctic atmosphere. Our main conclusions are the following:

1) Regional and seasonal trends of moisture transport are strongly associated with the changes in atmospheric circulation.
2) Long-term changes have most likely arisen from a combination of inherent variability of atmospheric dynamics and response to the increased amount of anthropogenic greenhouse gases. In winter, the changes are characterized by enhanced northward moisture transport in the Atlantic and Pacific sectors, in spring by the increased moisture transport over the Laptev Sea and farther toward the Canadian Archipelago, in summer by the intensified zonal moisture transport along the Eurasian coast, and in autumn by the increased moisture transport from the North Atlantic.

3) Moisture transport has had a dominant role for determining the regional changes in total column water vapor, whereas the role of local evaporation has been minor.

4) Increases in evaporation have been restricted to the vicinity of the sea ice margin over a limited period during the local sea ice decline. After the sea ice has disappeared from a region, evaporation over the open sea has had negative trends due to the effect of horizontal moisture transport to suppress evaporation.

5) At the marginal ice zone, both increased moisture transport and local evaporation have probably increased total column water vapor, whereas evaporation is of limited importance elsewhere. Near the sea ice margin, the trends in moisture transport and evaporation and the cloud response to those have been circulation dependent.

6) Over the open sea, years with a high amount of total column water vapor have been linked to large moisture transport but relatively low evaporation.

7) The trends in total cloud water are in many regions collocated with the trends in total column water vapor and moisture transport. Relative humidity, in turn, has increased in the regions, where the mean sea level pressure has had a negative trend.

In the future, there will be more open water area in the Arctic (IPCC 2019) to potentially increase evaporation. However, the sea area where cold, dry air masses can be formed is simultaneously diminishing (Pithan et al. 2015). Hence, the sea areas where cold, dry air masses can be injected to increase (Bengtsson et al. 2013; Skific and Francis 2013), it is likely that also in the future the atmospheric circulation and horizontal moisture transport patterns continue to distribute the moisture and control where and when efficient evaporation can take place. The regional moistening and cloud trends in the Arctic will probably be very sensitive to future changes in circulation patterns.

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REFERENCES


ABSTRACT: Atmospheric moisture is a key component in the water cycle and radiative transfer. In this study, a comprehensive picture of air moisture climatology and related physical processes is presented for the first time for the circumpolar area south of 50°S. The results are based on the most modern global reanalysis, ERA5, which manages reasonably well to close the Antarctic water budget. We show that over the ocean transient cyclones have the dominant role in determining moisture conditions, whereas over the continent the moisture conditions are largely affected by the mean circulation. Over the open sea, moisture transport from lower latitudes is an equally important source of moisture compared to the local evaporation, but practically all precipitating moisture over the plateau is provided by the horizontal transport. Over the ocean and continental slopes, southward moisture transport brings warm and moist air masses from lower latitudes, notably increasing atmospheric water vapor and cloud water, and simultaneously decreasing local evaporation over the open sea. On the Antarctic plateau, radiative cooling leads to high relative humidity and causes condensation of moisture especially near the surface, causing a nearly permanent specific humidity inversion layer. As a consequence, dry air masses with extremely low specific humidity are formed. This dry air masses are transported downward from the plateau by katabatic winds, experiencing adiabatic warming. This leads to a decrease in relative humidity and to a downward-directed sensible heat flux, which enable efficient surface evaporation on the coastal slopes and farther over coastal polynyas and leads.

KEYWORDS: Antarctica; Clouds; Evaporation; Moisture/moisture budget; Water vapor; Reanalysis data

1. Introduction

Atmospheric moisture has an important role in the Antarctic climate system, especially via its effects on the radiative energy transfer through the atmosphere, on precipitation, and further on the mass balance of the Antarctic ice sheet. Water vapor itself is an important greenhouse gas, but clouds, formed via condensation to water droplets or ice crystals, have an even larger influence on the radiative energy transfer. Clouds may have either a cooling effect on the surface by reflecting incoming solar radiation or a warming effect by reducing atmospheric transparency for thermal radiation. In polar regions, the latter effect typically dominates because of the small amount of incoming solar radiation, especially in wintertime, and the large surface albedo of snow (Pavolonis and Key 2003; van den Broeke et al. 2006). However, clouds often have a cooling effect on the surface over the open sea in the Antarctic (Pavolonis and Key 2003).

Much of the moisture in the polar regions is transported from the lower latitudes. On one hand, occurrence of clouds is sensitive to large- and synoptic-scale circulation patterns and moisture transport (Silber et al. 2019; Nygård et al. 2019). On the other hand, radiative cooling in the polar regions drives the atmospheric circulation, and clouds are thus able to affect the circulation by reducing the near-surface radiative cooling (Lubin et al. 1998). In this study, we address atmospheric moisture conditions in the high southern latitudes, focusing on moisture transport, surface moisture fluxes and clouds, and their interactions, based on the state-of-the-art reanalysis data.

Even though clouds have a remarkable influence on the radiation budget, their occurrence and properties are poorly known in the high southern latitudes, especially over the Antarctic ice sheet (Bromwich et al. 2012). In general, cloudiness is largest over the open ocean and decreases toward the interior of the Antarctic continent due to the decreasing amount of atmospheric moisture and the general lack of synoptic-scale cyclones over the continent (Bromwich et al. 2012; Simmonds and Keay 2000; Grieser et al. 2018). However, near-surface relative humidity is high (Genthon et al. 2017), but probably the rarity of condensation or ice nuclei prevents cloud formation (Bromwich et al. 2012). Occasionally the transport related to synoptic-scale cyclones or a flow around a blocking anticyclone is able to provide notable amounts of moisture and increase the cloudiness on the continent (Naithani et al. 2002).

Moisture transport caused by transient cyclones is the most important source of mass of the Antarctic ice sheet (Connolley and King 1993; Tietäväinen and Vihma 2008; Tsukernik and Lynch 2013; Papritz et al. 2014; Dufour et al. 2019). Transient cyclones occur on both the synoptic scale and the mesoscale (Irving et al. 2010; Uotila et al. 2011). Mesoscale cyclones have also an important contribution to moisture transport at least locally (Carrasco et al. 2003). Synoptic-scale cyclones are sometimes associated with atmospheric rivers, which typically generate the strongest moisture transport events. Their contribution to accumulation of snow on the ice sheet through precipitation is remarkable (Schlosser et al. 2010; Gorodetskaya et al. 2014; Turner et al. 2019). However, the strong transport events can also cause extensive surface melt, mainly at areas of a...
low surface elevations, due to transported sensible and latent heat as well as associated abundant cloudiness (Bozkurt et al. 2018; Wille et al. 2019). Atmospheric rivers are linked to meridional circulation patterns and blocking situations, which enable meridional moisture transport far from the ocean to the Antarctic continent (Naithani et al. 2002; Massom et al. 2004; Nicolas and Bromwich 2011; Gorodetskaya et al. 2014; Wille et al. 2019).

The moisture conditions on the Antarctic slope areas are largely affected by the prevailing flow features such as katabatic winds. On the coast of East Antarctica, moisture transport has been found to be divided into two vertical layers with different characteristics (Dufour et al. 2019); below the 900-hPa level, moisture transport is directed away from the continent due to katabatic winds, but above, moisture is transported inwards to the continent driven by the large-scale atmospheric circulation. Katabatic winds also have a strong influence on the vertical structure of specific and relative humidity (Nygård et al. 2013; Vignon et al. 2019). Nygård et al. (2013) suggested that advection of cold and dry air from the continent due to katabatic winds and advection of moist air mass from the ocean contribute notably to the formation of specific humidity inversions, which were frequently observed at the coastal Antarctic sounding stations. Adiabatic warming of air during downward flow decreases relative humidity (Vignon et al. 2019) and enables evaporation from the surface and from precipitation particles, notably affecting the amount of precipitation reaching the surface (Grazioli et al. 2017; Jullien et al. 2020).

Although the vertical structures of specific and relative humidity have important direct and indirect effects on radiative energy transfer and the mass balance of the Antarctic ice sheet, the vertical structures of humidity at high southern latitudes have not been comprehensively studied so far. The lack of observations limits the data available for analyses of air moisture and physical processes affecting the atmospheric moisture in the high southern latitudes. Furthermore, older-generation reanalyses have not provided sufficient accuracy for detailed studies of atmospheric moisture, because they have a very high uncertainty in the water cycle (Tietäväinen and Vihma 2008; Jakobson and Vihma 2010). However, a better availability and utilization of satellite data and development of data assimilation methods, as well as increased spatial resolution have improved the skill and brought the most recent global reanalysis, ERA5 (Copernicus Climate Change Service 2017; Hersbach et al. 2020), to the level of accuracy required for detailed studies of moisture conditions and related processes (Gossart et al. 2019; Graham et al. 2019; Vignon et al. 2019).

In this study, we utilized data of ERA5 for building, for the first time, a comprehensive picture of air moisture climatology and related physical processes (i.e., moisture transport, evaporation, and condensation) on the high southern latitudes. In particular, this approach enables analysis of interactions of vertical atmospheric moisture profiles with surface moisture fluxes and cloud formation, as well as with circulation-driven moisture transport over the whole Antarctic region. Moisture profiles and physical processes are separately presented for five regions: the open ocean, sea ice zone, coastal slopes of East Antarctica, the high plateau of East Antarctica, and West Antarctica (Fig. 1). Each of these regions has its own principal physical processes determining the moisture conditions inside the region.

2. Material and methods

This study is based on ERA5 reanalysis (Copernicus Climate Change Service 2017; Hersbach et al. 2020), which is the latest
version of reanalysis products produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). ERA5 provides a better horizontal and vertical resolution than its precursor ERA-Interim, and therefore ERA5 is potentially more capable to capture small-scale features. The high vertical resolution is needed especially over Antarctica, where the static stability near the surface efficiently limits the vertical extent of near-surface features and causes large vertical gradients in the flow field, above all in the case of katabatic winds. Recent studies (Gossart et al. 2019; Graham et al. 2019; Vignon et al. 2019) have demonstrated that ERA5 is capable of accurately simulating climatological conditions in the polar regions. Gossart et al. (2019) showed that ERA5 performs best among four modern reanalyses (ERA5, ERA-Interim, NCEP CFSR, MERRA-2) in simulating near-surface temperature over the Antarctic ice sheet in a comparison with nonassimilated surface observations. However, ERA5 has a warm bias in the interior of Antarctica, especially in winter and it underestimates relative humidity in winter (Gossart et al. 2019). Gossart et al. (2019) also showed that net precipitation (i.e., precipitation minus evaporation) in ERA5 was in the closest agreement with the observed surface accumulation on the Antarctic ice sheet. Graham et al. (2019) concluded that ERA5 simulates temperature, humidity, and wind profiles most accurately among five reanalyses (ERA5, ERA-Interim, JRA-55, NCEP CFSR, MERRA-2) in a comparison with nonassimilated radiosonde soundings in Fram Strait in the Arctic. Vignon et al. (2019) showed that ERA5 was generally better than ERA-Interim in simulating low tropospheric temperature, humidity, and wind speed profiles in coastal East Antarctica, when compared with assimilated radiosounding data. However, relative humidity in ERA5 was noticeably underestimated in the lowest 1500 m above surface. Nygård et al. (2020) stated that ERA5 can rather well produce the atmospheric water cycle in the Arctic, and our own analyses indicated that ERA5 reasonably well closes the water budget also in the Antarctic (Table 1). In the Antarctic, regional residuals between

| TABLE 1. Regional mean values of moisture variables for the five study regions. Mean values are shown for austral winter (April–October) and summer (November–March), and mean cumulative values for evaporation, precipitation, and net precipitation as well as the convergences of water vapor and cloud water for the entire year. Net precipitation is precipitation minus evaporation. Cloud water contains both liquid water and ice. |
|----------|--------|-------------|-----------------|-----------------|----------------|
|          | Open ocean | Sea ice | East Antarctic slopes | East Antarctic Plateau | West Antarctica |
| IWV (kg m⁻²) |        |          |                  |                  |                |
| Winter    | 8.27   | 4.00    | 1.36             | 0.37             | 1.83           |
| Summer    | 10.19  | 6.12    | 2.65             | 0.78             | 3.07           |
| Evaporation (mm day⁻¹) |        |          |                  |                  |                |
| Winter    | 1.61   | 0.40    | 0.10             | -0.01            | 0.00           |
| Summer    | 1.00   | 0.68    | 0.24             | 0.01             | 0.08           |
| Year      | 495    | 189     | 59               | -1               | 13             |
| Precipitation (mm day⁻¹) |        |          |                  |                  |                |
| Winter    | 2.95   | 1.78    | 0.96             | 0.19             | 1.01           |
| Summer    | 2.50   | 1.44    | 0.86             | 0.17             | 0.75           |
| Year      | 1009   | 598     | 334              | 65               | 328            |
| Net precipitation (mm day⁻¹) |        |          |                  |                  |                |
| Winter    | 1.35   | 1.38    | 0.86             | 0.20             | 1.00           |
| Summer    | 1.50   | 0.76    | 0.62             | 0.16             | 0.67           |
| Year      | 514    | 410     | 276              | 66               | 316            |
| Water vapor convergence (mm day⁻¹) |        |          |                  |                  |                |
| Winter    | 1.30   | 1.35    | 0.71             | 0.16             | 0.80           |
| Summer    | 1.48   | 0.84    | 0.53             | 0.13             | 0.54           |
| Year      | 502    | 415     | 231              | 54               | 252            |
| Cloud water convergence (mm day⁻¹) |        |          |                  |                  |                |
| Winter    | 0.02   | 0.00    | 0.06             | 0.03             | 0.08           |
| Summer    | 0.02   | -0.01   | 0.04             | 0.02             | 0.05           |
| Year      | 7      | -1      | 20               | 8                | 25             |
| Latent heat flux (W m⁻²) |        |          |                  |                  |                |
| Winter    | 46     | 12      | 3                | -0               | 0              |
| Summer    | 29     | 22      | 8                | 0                | 3              |
| Sensible heat flux (W m⁻²) |        |          |                  |                  |                |
| Winter    | 16     | 7       | -36              | -31              | -24            |
| Summer    | 2      | 7       | -18              | -14              | -10            |
| Cloud water (g m⁻²) |        |          |                  |                  |                |
| Winter    | 96     | 53      | 31               | 10               | 31             |
| Summer    | 101    | 57      | 27               | 8                | 25             |
| Specific humidity inversion occurrence (%) |        |          |                  |                  |                |
| Winter    | 24     | 64      | 85               | 99               | 95             |
| Summer    | 19     | 19      | 57               | 88               | 67             |
monthly mean net precipitation and monthly mean water (vapor and cloud condensate) flux divergence did not exceed 16% of water flux divergence in any subregion. ERA5 has a slightly better water budget closure over the ocean than its precursor ERA-Interim but a slightly poorer one over the continent. This, indeed, suggests that ERA5 is accurate enough for analyzing moisture conditions and related processes in detail.

ERA5 utilizes 4D variational data assimilation system of the IFS model of ECMWF. The spectral resolution of ERA5 is T639 leading to a 0.28° horizontal resolution. Reanalysis fields are available for every hour. However, in this study a lower (0.5° 5° horizontal resolution and a 6-h temporal resolution as well as a 15-yr time period were used to keep the computational resources required on a manageable level. ERA5 has 137 hybrid levels, of which the lowest 58 levels were used in this study to cover the troposphere up to the 300-hPa pressure level; the amount of water above 300 hPa is very small. Vertical spacing of levels is denser near the surface. There are approximately seven levels in the lowest 200 m and 20 levels in the lowest 1000 m, but the spacing gradually becomes coarser upward.

An accurate presentation of physical processes is also important for the accuracy of reanalysis products. In the perspective of atmospheric moisture, the most essential physical processes are surface evaporation/condensation and cloud formation. In ERA5, evaporation is calculated between the surface and the lowest model level by applying a bulk aerodynamic equation similar to Eq. (1). Wind speed and specific humidity of air are taken from the lowest model level, and the specific humidity of water, ice and snow surfaces is the saturation specific humidity corresponding to the surface temperature. The turbulent exchange coefficient includes the stability parameter, which is itself a function of the surface fluxes. The stability dependence is calculated based on the Monin–Obukhov formulation, applying empirical forms of the dimensionless gradient functions.

The cloud scheme of ERA5 has five prognostic variables, cloud liquid water, cloud ice, cloud fraction, rainwater, and snow, which are advected by the airflow. Evolutions of cloud liquid water and ice are prognosed separately, which allows the occurrence, but are advected by the airflow. Evolutions of cloud liquid water and ice, cloud fraction, rainwater, and snow, which are advected by the airflow. Evolutions of cloud liquid water and ice, cloud fraction, rainwater, and snow, which are advected by the airflow. An empirical forms of the dimensionless gradient functions. In our analyses, the circumpolar study area south of 50°S was divided into five subregions: the open ocean, sea ice zone, coastal slopes of East Antarctica, the high plateau of East Antarctica, and West Antarctica (Fig. 1). On the basis of the 6-hourly data, the mean sea ice concentration over 2004–18 in each grid box was calculated separately for each month. A 50% threshold value was applied to define the mean (over 2004–18) monthly sea ice zone. Instead of the 15% threshold used for sea ice extent (e.g., Stroeve et al. 2012), a 50% threshold was used because it better represents the effects of sea ice on evaporation. The sea ice extent has large seasonal variation, having maximum in September, when the sea ice reaches almost all the way to 60°S and even beyond, and minimum in February, when there is no continuous belt of sea ice around the continent (Fig. 1).

For total column vertically integrated variables (vertically integrated moisture transport and moisture transport convergence, vertically integrated water vapor, vertically integrated cloud liquid water and ice content) we directly utilized vertically integrated products from ERA5 reanalysis data. For vertical structures and inversion statistics, we utilized the native model level data for taking the full advantage of the high vertical resolution. This was particularly important for calculations of moisture transport and inversion layer statistics. All calculations (i.e., temporal averaging, inversion occurrence calculations, and correlation calculations) except spatial averaging were made using the model level data, and afterward results were converted to a pressure coordinate system with 25-hPa vertical resolution, using multiyear monthly mean pressures of each model level. Variables were linearly interpolated from the closest model levels to the target pressure level, except horizontal moisture transport, which was first vertically integrated from the model level data, and then converted to pressure levels. Detailed description of the moisture transport calculation method is presented in the online supplemental material.

When interpreting the factors controlling evaporation \( E \), we refer to the well-known bulk aerodynamic equation (e.g., Deardorff 1968; Launiainen and Vihma 1990; Boisvert et al. 2020)

\[
E = \rho C_D (q_s - q_a) V, \tag{1}
\]

where \( \rho \) is the air density, \( C_D \) is the turbulent transfer coefficient for moisture, \( q_s \) is the surface specific humidity (for water, snow and ice surfaces only depending on the surface temperature), \( q_a \) is the air specific humidity, and \( V \) is the wind speed.

In this study, relative humidity is defined with respect to water at temperatures above 0°C, whereas at temperatures below 0°C relative humidity is defined with respect to ice.

3. Results

In this section, the vertical structure of atmospheric moisture and the physical processes affecting the moisture are studied regionally. The study regions—the open ocean, sea ice zone, East

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Antarctic slopes, East Antarctic plateau, and West Antarctica—
are addressed by starting from the moistest region (i.e., the open
ocean) and then proceeding to the drier regions. An exception to
this order is West Antarctica, which is addressed last, because it
includes features typical for both the East Antarctic slopes and
the East Antarctic plateau. In this section, we also discuss effects
of meridional moisture transport direction on the moisture con-
ditions, but the related figures and tables are presented in the
online supplemental material.

a. Atmospheric moisture over the open ocean

Over the open ocean, the spatial distribution of vertically
integrated water vapor (IWV) is zonally symmetric but the
amount of IWV notably decreases toward the south (Figs. 2
and 3). At latitude 50°S, for both winter and summer, the
seasonal mean IWV is approximately twice as large as near the
sea ice margin. In addition, IWV has a clear but fairly weak
seasonal cycle; in summer IWV is 23% higher than in winter
(Table 1). The sources of IWV are evaporation and moisture
transport. The seasonal cycle of evaporation is opposite to that
of IWV. Evaporation is 61% larger in winter than in summer
(Table 1). Evaporation increases from the sea ice edge toward
lower latitudes (Figs. 4 and 5). Even though evaporation over
the open ocean is relatively large compared with the other
parts of the Antarctic, it only covers approximately a half of the
mean precipitation in this subregion (Table 1). This means
that a major part of the water precipitating in the Antarctic region, including the open ocean south of 50°S, is transported from latitudes lower than 50°S.

Over the open ocean north of latitude 65°S, the mean moisture transport consists of an eastward directed and zonally symmetric zonal component, due to strong westerly winds, and a mostly southward meridional component (Figs. 2 and 3). In most parts of the subregion the zonal moisture transport component is eastward over 80% of the time but direction of meridional moisture transport is almost as often poleward as equatorward (Supplement Figs. 2 and 3). The eastward moisture transport becomes weaker toward the sea ice margin due to weakening of the westerlies and decreasing atmospheric moisture content. South of 65°S the mean moisture transport becomes more meridional and turns westward in the vicinity of the continent.

Moisture transport has remarkable effects on the spatial and temporal variations in atmospheric moisture content. IWV is correlated with vertically integrated meridional moisture transport (Supplement Table 1) being on average one-third higher when the air masses are originating from the north compared with air masses originating from the south (Supplement Figs. 4 and 5). Moisture transport also affects moisture exchange between the atmosphere and sea surface (Supplement Table 2). Efficient evaporation is associated with northward flow, as evaporation is on average 87% higher during northward moisture transport compared with southward transport (Supplement Fig. 6).

On average, specific humidity decreases with height over the open ocean (Figs. 2 and 3). Specific humidity inversions (i.e., conditions when the vertical specific humidity maximum is not located at the surface) occurred only 16% (19%) of the time in

Fig. 3. As in Fig. 2, but for austral summer (November–March).
winter (in summer) (Fig. 6). Of those specific humidity inversions, 82% occurred when moisture transport was directed southward. The decrease in specific humidity from the surface to approximately 300 m above the surface is relatively small, which is probably related to frequent occurrence of a shallow well-mixed boundary layer. This well-mixed boundary layer enhances surface evaporation (Fig. 7), which further decreases occurrence of specific humidity inversions.

Of the five Antarctic subregions, the highest vertically integrated cloud water content occurs over the open ocean (Figs. 8 and 9 and Table 1). In this study, cloud water includes both ice and liquid phases. Over the open ocean, vertical maximum of cloud water content occurs between the 900- and 850-hPa levels. Formation of clouds above the well-mixed boundary layer is probably contributed by adiabatic cooling of the air mass holding moisture evaporated from the surface. This is supported by the mean relative humidity increasing from the surface to the 900-hPa level (Figs. 4 and 5) and by the upward-directed mean vertical water vapor transport in the lower troposphere (Figs. 2 and 3). However, a large part of precipitated moisture is transported from lower latitudes, which probably has a large contribution of clouds associated with transient cyclones. This is supported by the fact that vertically integrated cloud water is on average twice as high when the moisture transport was from the north compared with transport from the south (Supplement Figs. 7 and 8). As an
exception, the cloud water content in the low troposphere near the sea ice margin in winter is higher, when moisture transport is from the south. This probably due to advection of cold air from sea ice to open ocean, which increases evaporation and likely leads to formation of convective clouds on the top of the boundary layer.

In summary, typical conditions over the open ocean consist of a shallow unsaturated and often well-mixed layer near the surface, which enables relatively large surface evaporation. Above the layer, stratification is more stable and upward-transported moisture reaches saturation leading to formation of clouds and decrease of specific humidity. However, local evaporation provides only a part of moisture for precipitation, which is approximately twice as large as evaporation in the region. Therefore, moisture transport from lower latitudes is approximately an equally important source of moisture as local evaporation. The southward moisture transport increases IWV and decreases surface evaporation. Over the open ocean, evaporation is larger in winter than in summer, but overall, seasonal variations in moisture transport, vertical structure of specific and relative humidity, and cloud water content are relatively small.

b. Atmospheric moisture over the seasonal sea ice region

Over the sea ice, IWV has relatively large seasonal and spatial variation. IWV is approximately 50% higher in summer than in winter and several times higher near the sea ice margin than near the coast in winter (Figs. 2 and 3, Table 1). Over the sea ice, the contribution of local evaporation to the atmospheric moisture content is considerably smaller than over the open ocean, whereas contribution of moisture transport is
larger, because the annual precipitation is approximately 3 times larger than annual evaporation (Table 1). The seasonal cycle of evaporation is opposite to that on the open sea. In summer, evaporation is 70% larger than in winter (Table 1), whereas precipitation is approximately a quarter larger in winter than in summer (Fig. 10).

Over the sea ice region, vertically integrated moisture transport changes from the westerly transport in the northern part of the region to easterly transport in the southern part of region (Figs. 2 and 3). This is caused by a change in the direction of moisture transport at low altitudes, whereas at higher altitudes transport remains westerly. The low-level change is related to the geostrophic wind and mean sea level pressure; the latter of which is lowest at these latitudes. The meridional component of mean moisture transport is mostly southward over the sea ice, except near the continent on the western side of Ross Sea and Amery Ice Shelf, where the outflow from the continent is strongest, and in the area east of the Antarctic Peninsula and the western Weddell Sea. In the former case, the transport pattern is related to the prevailing Amundsen Sea low, so that the northward transported moisture mostly originates from over the Amundsen and Bellingshausen Seas but makes a loop over West Antarctica.

Correlations between meridional moisture transport and IWV are slightly weaker than over the open sea (Supplement Table 1). The mean IWV was 68% (33%) larger when the moisture transport was from the north compared with situations when the moisture transport was from the south in winter (in summer) (Supplement Figs. 4 and 5). However, the effects of moisture transport on evaporation are weak and spatially varying over the sea ice (Supplement Table 2 and Supplement Fig. 6).

Vertical structure of atmospheric moisture has large seasonal variations over the sea ice. In winter, katabatic winds bring cold, dry (both absolutely and relatively) air masses from the continent over the sea ice, where stable stratification, often with temperature inversions, prevails in the layers below the 900-hPa level, preventing mixing with the layers above. Specific humidity inversions are also common in winter (Table 1), being most frequent between the 900- and 800-hPa levels. Formation of specific humidity inversions is probably related to vertically varying moisture advection, with advection of a dry and cold continental air mass near the surface and advection of an oceanic air mass at higher altitudes. In winter, the insulating effect of sea ice is seen as a low surface specific humidity (Fig. 2). The generally low specific humidity allows effective evaporation from leads and polynyas. The largest evaporation occurs in the locations where the mean sea ice concentration is lowest (Fig. 7), demonstrating the large contribution of leads and polynyas to surface evaporation. Evaporation decreases the occurrence of surface-based specific humidity inversions, but on the Weddell Sea, where surface evaporation is small or even negative, surface-based humidity inversion are most common, and their occurrence is between 30% and 50% during midwinter. In summer, solar radiation increases the gain of radiative energy at the surface and enables upward directed sensible heat flux (not shown) and evaporation (Fig. 5). Upward sensible heat flux weakens the stratification, and there is no near-surface cold air layer capped with temperature inversion, therefore specific humidity and temperature inversions are rare in summer (Table 1).
Over the sea ice, vertically integrated cloud water decreases toward the continent but the vertical structure of cloud water content is mostly similar to that over the open ocean (Figs. 8 and 9). Above the 900-hPa level, cloud water content is largely affected by the direction of meridional moisture transport (Supplement Figs. 7 and 8 and Supplement Table 3). On average, vertically integrated cloud water is almost 4 times (more than 2 times) larger in winter (in summer) when the moisture transport is southward compared with northward moisture transport.

In summary, the sea ice region is a transition zone between continental and oceanic moisture conditions, where the near-surface moisture environment undergoes large seasonal variations. In winter, katabatic winds bring dry, cold air from the continent into a layer near the surface, enabling efficient evaporation from leads and polynyas. In summer, solar radiation strengthens the heat and moisture fluxes from the sea ice surface, weakens the stratification and erodes humidity and temperature inversions. Synoptic-scale variations of meridional moisture transport have a large effect on the vertically integrated water vapor and cloud water, but only a weak influence on evaporation.

c. Atmospheric moisture over the slopes of East Antarctica

On the coastal slopes of East Antarctica, IWV decreases rapidly poleward, being approximately twice as large near the coast as at the altitude of 2000 m on the slopes (Figs. 2 and 3). The decrease in IVW is not only caused by the higher surface elevation, but it is also affected by the horizontal gradient in specific humidity. Even though most of atmospheric moisture is located in low elevations due to higher temperatures, the decrease in IWV is not only due to the temperature–altitude dependence but also due to the horizontal gradient in specific humidity, which is related to meridional temperature changes. On the slopes, moisture transport is the main source of atmospheric water vapor, as precipitation is more than
5 times larger than evaporation (Table 1). Convergence of moisture transport thus balances the strongly positive net precipitation. Precipitation on the slopes is largest near the coastline, but it sharply decreases toward the inner continent (Fig. 10).

Vertically integrated mean moisture transport is directed almost parallel with the surface elevation isolines but has a mostly small component toward the inner continent (Figs. 2 and 3). However, near the surface, moisture transport is more than 80% of the time directed downward along the slope (i.e., typically approximately northward), indicating that the katabatic winds cause rather continuous moisture transport from the continent toward the ocean, despite the rather low humidity of the transported air mass. Mean moisture transport turns toward the inner continent with altitude.

Northward moisture transport is usually associated with dry air advection, which contributes to a decrease of specific humidity at a fixed location. This is the typical case in all the five study regions, but the effect is, on average, strongest over the coastal regions, where the climatological moisture gradient is largest. IWV correlates positively with vertically integrated southward moisture transport (Supplement Table 1). In winter (in summer), mean IWV is 53% (34%) larger, when the vertically integrated moisture transport is from the north than from the south. However, the direction of meridional moisture transport only weakly affects evaporation (Supplement Table 2).

As the near-surface air masses over the slopes of East Antarctica commonly originate from the inner continent, air specific humidity in the low troposphere is often low. Adiabatic
warming increases the temperature of the downward-flowing air mass and, due to the turbulent heat exchange between the air and snow surface, the surface temperature is also increased. Hence, the (saturation) specific humidity of the surface is increased but the air specific humidity is not directly affected by the downslope flow (in a Lagrangian framework). This, together with the increasing near-surface wind speeds along the slope, strengthens surface evaporation [see Eq. (1)]. Evaporation increases air specific humidity most effectively near the surface and thus decreases the occurrence of surface-based specific humidity inversions, but it is not able to increase relative humidity close to the saturation point. Besides, the influence of surface evaporation on the vertical profile of specific humidity is only limited to a shallow layer due to stable stratification. However, elevated specific humidity inversions are common also near the coastline (Fig. 6).

Relative humidity is typically below 60% in the layer 200–500 m above the slopes of East Antarctica, being the lowest mean near-surface relative humidity in the entire Antarctic (Figs. 4 and 5). On the slopes, there is less vertically integrated cloud water than over the ocean (Figs. 8 and 9; Table 1), mostly due to the higher surface altitude. However, in the middle and upper troposphere, the largest decrease in cloud water toward the inner continent does not occur until the upper part of the slopes. The direction of meridional moisture transport strongly affects vertically integrated cloud water (Supplement Table 3), which in winter (in summer) is almost 7 (5) times larger in cases when vertically integrated moisture transport is from the north than when it is from the south. The large difference in cloud water suggests that the cases of flow from ocean to continent are associated with thick clouds and the cases of flow from the continent toward the ocean with notably less clouds.

Fig. 9. As in Fig. 8, but for summer (November–March).
Over the slopes, near-surface moisture conditions are dominated by the effects of dry air advection related to katabatic winds. In this layer, downward winds bring dry air from the continent, and adiabatic warming of the air contributes to higher surface temperatures, enabling more evaporation from the surface [see Eq. (1)]. This decreases the occurrence of surface-based specific humidity inversion. However, specific humidity inversions are common in the upper part of the katabatic layer. These inversions probably separate near-surface air mass of continental origin from the overlying air mass, which presumably more often has an oceanic origin. The direction of vertically integrated moisture transport noticeably affects IWV. In addition, the large relative difference in vertically integrated cloud water between the cases of southward and northward moisture transport suggests that cloud formation, to which orographic lifting contributes, is associated with southward moisture transport.

**d. Atmospheric moisture over the East Antarctic Plateau**

Of the whole Antarctic region, the lowest IWV occurs on the highest parts of the East Antarctic Plateau (Figs. 2 and 3; Table 1). On the plateau, the mean surface flux of moisture in summer is slightly positive meaning evaporation, but in winter the flux is slightly negative meaning condensation of moisture to the surface (Table 1); the small magnitudes are related to the low temperature and weak winds [see Eq. (1)]. The annual mean evaporation is slightly negative, meaning that condensation exceeds evaporation (Table 1) and practically all the moisture precipitated on the plateau is transported from outside of the area.

Even the strongest moisture transport events on the plateau are weak, because synoptic-scale cyclones are typically not able to efficiently transport moisture to the high plateau. Mean horizontal moisture transport is mostly directed along the surface elevation isolines as on the coastal slopes (Figs. 2 and 3). Over the highest areas, vertical moisture transport is downward in the entire layer from the ice sheet surface to the 300-hPa level. The downward moisture transport is related to the mean meridional circulation with downward motion of the air near the pole.

Moisture transport affects IWV also over the plateau, even though its effect is smaller than on the slopes (Supplement Figs. 4 and 5). On the plateau, direction of moisture transport does not as clearly define the moisture properties of the air mass as in the other areas. One reason for this is that air flowing across the pole toward the center of plateau typically brings moisture to the plateau, even though moisture transport is toward lower latitudes. The strongest correlation between meridional moisture transport and IWV is found in the area toward the coastal slopes, and the correlation decreases toward the highest areas. Moisture transport also affects surface flux of moisture in winter (Supplement Fig. 6). Transport of moisture from the areas of a lower surface elevation increases the surface condensation. In summer, there is almost no correlation between the surface moisture flux (evaporation or condensation) and the horizontal moisture transport (Supplement Table 2).

Condensation of moisture, especially near the surface, has a strong influence on the vertical structure of atmospheric moisture on the plateau. In winter, specific humidity inversions are practically always (i.e., 99% of the time) present on the plateau (Fig. 6; Table 1). In summer, specific humidity inversions are slightly less common, occurring 88% of the time. The vertical maximum of specific humidity occurs, on average,
700 m (600 m) above the surface in winter (in summer), and a strong surface-based humidity inversion occurs below this level (Figs. 2 and 3). In winter, specific humidity is approximately 3 times larger on the level of specific humidity maximum than at the surface. In summer, the absolute strength of specific humidity inversion is almost the same as in winter but, the mean surface specific humidity is 4 times larger in summer.

Over the plateau, the amount of cloud water is small, even though relative humidity is high (Figs. 4, 5, 8, and 9). In contrary to the other areas, the largest cloud water content is found near the surface, as is the highest relative humidity. Near the surface, the mean relative humidity exceeds 95% (90%) in winter (in summer) and supersaturation with respect to ice is common. Saturated conditions with respect to ice are frequent, especially near the surface; they occur 40%-60% (30%-40%) of the time in winter (in summer). The high relative humidity and frequently occurring saturated conditions, despite the downward vertical motion and related adiabatic heating, suggest that the radiative cooling has a dominating influence on the moisture conditions.

In summary, the East Antarctic Plateau acts as a sink for atmospheric moisture and as a formation area of continental polar air. All precipitating moisture is transported to the area, as there is no local source of water vapor. Instead, there is condensation onto the ice sheet surface. Despite prevailing subsidence, the relative humidity is high, especially in winter, which is due to the radiative cooling of air masses. The cooling is strongest close to the ice sheet surface, and the cold, dry (in terms of specific humidity), near-surface air masses drain out of the high plateau and affect moisture conditions on the slopes and as far as on the sea ice. Over the plateau, specific humidity inversions are strong and persistent.

e. Atmospheric moisture in West Antarctica

In West Antarctica, IWV has large spatial variation inside the region; IWV is the smallest in the region near the South Pole and the largest in the Antarctic Peninsula (Figs. 2 and 3). Near the South Pole, the annual mean IWV is approximately only one-tenth of the IWV in the coastal areas. Mean IWV over the whole region is approximately 70% higher in sumer than in winter (Table 1). In West Antarctica, mean precipitation is relatively large, mainly due to large contributions of the coastal area and the Antarctic Peninsula (Fig. 10). Mean precipitation rate is 35% larger in winter than in summer. Local evaporation covers only 4% of the precipitation, and practically nearly all evaporation happens in summer (Table 1).

Precipitation in West Antarctica is largely contributed by moisture convergence over the Antarctic Peninsula and the coastal area between the Peninsula and the Ross Ice Shelf. On the coast between the Ross Ice Shelf and Antarctic Peninsula, 150°–65°W, largely due to the Amundsen Sea low, the mean moisture transport is southward from the Amundsen Sea toward the continent, and over the Antarctica Peninsula the mean transport is eastward (Figs. 2 and 3). In these areas, moisture transport is from the ocean toward the high-elevation areas, causing moisture convergence. Near the coast, between the Ross Ice Shelf and the Antarctic Peninsula, southward moisture transport is only 0%–20% more frequent than northward moisture transport, suggesting that transient cyclones are responsible for the major part of the moisture transport in the area. In inner parts of West Antarctica, the mean moisture transport is affected by both the Amundsen Sea low and local surface topography. Moisture transport is mainly southward in Marie Byrd Land and northward over the Ross Ice Shelf, Ellsworth Land, and the east side of the Antarctic Peninsula. A rather persistent direction of moisture transport (Supplement Fig. 3) suggests a weaker role of transient cyclones in moisture transport in these areas.

Temporal variation of southward moisture transport is mostly positively correlated with temporal variations of IWV (Supplement Table 1), except on the west side of the Ross Ice Shelf, where IWV typically decreases when moisture transport is southward (Supplement Figs. 4 and 5). The maximal decrease occurs close to the Transantarctic Mountains, suggesting that the contribution of dry-air advection from East Antarctic Plateau to atmospheric moisture is stronger during southward flow. On average over the region, IWV is 50% (30%) larger when moisture transport is from the north than from the south in winter (in summer). However, impacts of moisture transport on evaporation are spatially varying (Supplement Fig. 6).

Condensation at the surface is common in winter (Fig. 4), and it is associated with frequent occurrence of specific humidity inversions (Fig. 6). In winter, specific humidity inversions are most common in a 500-m-thick layer near the surface, and their occurrence decreases above the layer. In summer, surface evaporation is higher than in winter (Fig. 5), which decreases the occurrence of specific humidity inversion. The difference between winter and summer in specific humidity inversion occurrence and surface moisture flux is largest in the low elevation and slope areas, but on the dome the mean surface moisture flux is downward even in summer and specific humidity inversions are common. Katabatic winds have locally a large impact on the vertical structure of humidity also in West Antarctica. The lowest occurrence of specific humidity inversions as well as the largest evaporation is located southeast from the Ross Ice Shelf and on the west side of Ross Ice Shelf, where northward moisture transport is the most frequent (Supplement Fig. 3). Over this area, the almost persistent northward moisture transport is associated with dry air advection from the higher elevation areas, which enables surface evaporation and decreases the occurrence of specific humidity inversions.

Cloud water content has large spatial variation in West Antarctica (Figs. 8 and 9). The mean vertically integrated cloud water is approximately 10 times larger in the west side of the Antarctic Peninsula than near the pole. Vertically integrated cloud water is also high in the coastal areas between the Ross Ice Shelf and Antarctic Peninsula, especially in winter. In these areas, orographic lifting and moisture convergence result in formation of vertically thick clouds. In the high elevation areas, near-surface relative humidity is as high as in the East Antarctic Plateau, but the layer with high relative humidity is shallower than in East Antarctica, suggesting that condensation due to radiative cooling is limited to a shallow layer near the surface. This may be related to more frequent advection of warm, moist air masses to West Antarctica than to East Antarctica.
We summarize that moisture conditions have large spatial variation in West Antarctica. On the coast between the Ross Sea and Antarctic Peninsula, moisture transport from the ocean to the continent causes moisture convergence and leads to large precipitation on the coastal areas. Moisture conditions on the dome areas of West Antarctica resemble conditions typical for the East Antarctic Plateau, whereas on the slope areas between the Dome and the Ronne Ice Shelf moisture conditions resemble those over the East Antarctic slopes. Downslope winds have also a large influence on the moisture conditions on the west side of Ross Ice Shelf, where almost a permanent flow of dry air from the higher elevation areas enhances surface evaporation decreasing specific humidity inversion occurrence.

4. Discussion

Moisture conditions in the Antarctic are largely controlled by radiative cooling on the Antarctic plateau, katabatic winds and associated adiabatic warming on the sloping ice sheet, and transient cyclones over the ocean and coastal slopes. Over the high plateau, the strong radiative cooling causing moisture condensation results in the formation of cold air masses with a very low specific humidity. Near-surface relative humidity is high and supersaturation with respect to ice is common on the dome areas as observed by Genthon et al. (2017). On the sloping ice sheet, the downward flowing air warms adiabatically, which is also reflected in a higher snow surface temperature, due to the downward sensible heat flux, and accordingly to a higher surface (saturation) specific humidity. However, air specific humidity is not directly affected by adiabatic warming, as indicated by low relative humidity. Hence, the specific humidity on the snow surface is higher than in the air above, enabling surface evaporation, which is further strengthened by the strong winds on the slopes [Eq. (1)].

Our results confirm that, on the coast of East Antarctica, the mean moisture transport is from the continent toward the ocean in the near-surface layer, as earlier shown by Dufour et al. (2019). Based on radiosonde observations, Vignon et al. (2019) showed that in the same layer relative humidity is typically low, and wind is often directed out of the continent at coastal sounding stations. Our results confirm the presence of a layer of low relative humidity in the low troposphere on the slopes of East Antarctica. The low relative humidity allows efficient sublimation from precipitation particles. According to Grazzioli et al. (2017) and Jullien et al. (2020) a notable part of the precipitating particles sublimates before reaching the ice sheet surface. However, the relative humidity and evaporation in our results may be underestimated because drifting and blowing snow are not simulated in the ERA5 reanalysis (Gossart et al. 2019). Katabatic flows also often cause strong surface winds, which are able to lift snow from the surface and form drifting and blowing snow (Lenaerts and van den Broeke 2012; Barral et al. 2014; Gossart et al. 2017). Blowing snow can significantly increase sublimation, and thus affect the mass balance of the ice sheet (Lenaerts and van den Broeke 2012), as well as increase relative humidity of air (Barral et al. 2014).

The influence of katabatic winds and drainage of the cold, dry air mass is not only limited to the slopes of the continent, but the effects also reach to the sea. Katabatic winds contribute to formation of coastal polynyas (Adolphs and Wendler 1995; Zhang et al. 2015), which considerably increase evaporation in the sea ice zone (Boisvert et al. 2020). Evaporation (latent heat flux) and sensible heat flux cause strong cooling and ice production in leads and polynyas, which contributes to formation Antarctic Bottom Water (Ohshima et al. 2016; Thompson et al. 2020). Evaporation from polynyas is high due to strong winds of the katabatic flow, the dry air transported from the continent over the open sea surface, and the relatively high surface temperature of the open sea. Our results show larger surface evaporation over coastal polynyas than surrounding areas, especially in winter. Equally strong evaporation occurs downwind of the sea ice margin during off-ice flows.

Although the mean circulation, with persistent katabatic winds, largely determines the moisture conditions over the Antarctic plateau and the slopes, transient cyclones and frontal structures associated with them have a vital role for moisture and cloud conditions in the high southern latitudes. Previous studies (Tietäväinen and Vilhama 2008; Tsukernik and Lynch 2013; Papritz et al. 2014; Dufour et al. 2019; Sinclair and Dacre 2019) have shown that transient cyclones are responsible for the large part of the moisture transport and precipitation to the Antarctic continent. The synoptic-scale and mesoscale cyclone activity is largest at latitudes between 60° and 65°S (Simmonds and Keay 2000; Uotila et al. 2009; Pezza et al. 2016). The structure of most synoptic-scale cyclones located in the baroclinic zone between polar and midlatitude air masses can be probably explained approximately by the Norwegian cyclone model consisting of southward advection of the warm air mass in front (east side) of the system, related to a southward displacement of the frontal zone, and northward advection of cold air mass in behind (west side) of the system, related to a northward displacement of the frontal zone. The track and evolution of the cyclone define how far southward the warm air mass is able to penetrate (Sinclair and Dacre 2019). In contrast to Northern Hemispheric extratropical cyclones, the Southern Hemispheric ones have rather zonal tracks, and they are typically not able to penetrate to the inner continent (Jones and Simmonds 1993; Simmonds and Keay 2000). Thus, they cannot efficiently provide moisture to the inner parts of the continent (Sinclair and Dacre 2019), which may explain the very small moisture transport and low atmospheric moisture content on the East Antarctic plateau. Moisture transport from the lower latitudes is linked to a warm conveyor belt in the warm sector of transient cyclones. Our qualitative analysis of a small set of synoptic situations suggests that synoptic-scale cyclones have often started to occlude before the warm moist air mass reaches the continent. This would prevent advection of the warmest and moistest air masses to the near-surface layer over the continent and would explain the rather small influence of moisture transport variation on the surface moisture fluxes over sea ice and slopes. However, transient cyclones affect the surface moisture flux over the open ocean by decreasing evaporation when warm moist air is advected over the surface. Instead, northward flow increases evaporation due to a lower
specific humidity and increased vertical mixing, the latter because of a less stable near-surface stratification. These are in line with the results of Truong et al. (2020).

5. Conclusions

Our most important findings on features and processes related to moisture conditions in the Antarctic are summarized below and in Fig. 11.

1) Dry air masses with extremely low specific humidity are formed near the surface on the Antarctic plateau, where the radiative cooling increases relative humidity and causes condensation of moisture, especially near the surface. Here, relative humidity is always close to the saturation point despite of the downward motion of air, and there is a practically a permanent specific humidity inversion layer.

2) Katabatic winds transport air masses with very low specific humidity from the inner continent toward the coast. On the coastal slopes, descent of air causes adiabatic warming and increase of the specific humidity difference between the snow surface and air, which, together with strong winds, enables efficient surface evaporation on the coastal slopes. Katabatic winds also contribute to formation of coastal polynyas, which further increases evaporation in the sea ice region in winter.

3) Transient cyclones, and meridional moisture transport associated with them, strongly affect atmospheric water vapor and cloud water content over the ocean and continental slopes, but have only a small influence on the surface moisture fluxes on the sea ice and coastal slopes. Southward moisture transport typically increases IWV and cloud water content. In winter, off-ice airflow enhances evaporation on the open sea, in the vicinity of the sea ice margin, and leads to a large cloud water content in the upper part of boundary layer.

4) Precipitation exceeds evaporation in the high southern latitudes. Even on the open ocean, precipitation is approximately twice as large as evaporation, and evaporation decreases southward at a higher rate than precipitation. The contribution of moisture transport to precipitation thus increases toward the inner parts of the continent, where practically all precipitating moisture is transported from outside the region, as indicated by the negative mean evaporation.

5) In general, the mean moisture conditions on the continent are largely associated with the mean meridional circulation and topographical adjustments to it, but over the ocean and coasts, transient cyclones have an important role. They are typically not able to efficiently transport moisture to the inner continent (Naithani et al. 2002), but rare, strong moisture transport events, related to the warm conveyor belt and frontal structures of synoptic-scale cyclones (Papritz et al. 2014; Catto et al. 2015), have a major contribution to the water cycle, especially to accumulation of snow on the high-altitude ice sheet (Gorodetskaya et al. 2014).

We conclude that the recent development, especially increases in satellite observations and improvements in utilization of these observations in data assimilation, has brought the global reanalysis, ERA5, to the level of accuracy sufficient for detailed analyses on moisture conditions and related processes also in the high southern latitudes. In this study, we utilized the unprecedented level of accuracy to provide, for the first time, a comprehensive description of atmospheric moisture conditions in the Antarctic. Further research is needed, among others, to better understand and quantify 1) evaporation from drifting and blowing snow and 2) the combined effects of the often simultaneous transports of moisture and dry static energy on cloud formation, precipitation, and the surface mass balance of the Antarctic ice sheet.

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Data availability statement. All ERA5 data are openly available from Climate Data Store (CDS) of Copernicus Climate Change Service. Single level data are available at https://doi.org/10.24381/cds.adbb2d47 and model level data can be downloaded via Climate Data Store (CDS) of Copernicus Climate Change Service at https://cds.climate.copernicus.eu/cdsapp#!/home

REFERENCES


