Temporal and vertical structure of the summer Arctic boundary layer

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Abstract

This study provides an analysis of the vertical and temporal structure of the summer Arctic boundary layer, using remote-sensing data from the ice-drift of the Arctic Summer Cloud Ocean Study (ASCOS) field campaign in the summer of 2008. The ASCOS ice-drift was deployed 12 August through 1 September. The analysis focused on two time periods; one characterized by melting season conditions and was synoptically very variable due to passing weather systems and one more steady with a persistent stratocumulus, in the transition from the melt season to the fall freeze-up.

The vertical structure exhibited a well-mixed boundary layer during both periods, but indicated generally lower stability and more vigorously mixing in the later period. In this period a layering structure was also found, with a second well-mixed layer associated with the cloud layer close to the inversion.

A statistically significant diurnal cycle was found through the boundary layer for all the variables analyzed, but differences between the two periods appeared. We speculate that the vertical diurnal-variability pattern found in the later period is related to the interaction between a turbulent upper cloud layer and the lower surface layer. It is further speculated that an intermediate layer, between the top of the surface based layer and the upper cloud layer, exists. This layer is possibly sometimes mixed by turbulent eddies from clouds and sometimes not.

Three indirect different turbulence estimates, all based on remote sensing, were analyzed for their intervariability. The results show that there was no or very low correlation between the estimates and none gave any signal that could be related to cloud-induced turbulence at higher levels. A Richardson number ($Ri$) analysis did not verify a relationship between different $Ri$ regimes and the turbulence estimates. To further examine the turbulent state of the summer Arctic boundary layer, and confirm any layering of turbulence, other measurement systems must be deployed.
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1 Introduction

As an international project of the Arctic Council and the International Arctic Climate Impact Assessment, the Arctic Climate Impact Assessment (ACIA) was conducted to evaluate and synthesize scientific knowledge on climate variability and climate change in the Arctic region. According to the ACIA report (2005) anthropogenic climate change is being experienced particularly intensely in the Arctic region. Near-surface temperature are currently rising at a rate at least twice of the global average and widespread melting of glaciers and sea ice are further evidence of strong Arctic warming (also e.g. IPCC 2007, Lindsay & Zhang 2005; Serreze et al. 2007; Overland 2009). This amplification of climate change, called Arctic amplification (e.g Serreze & Francis 2006) is not news to science, but there are still many uncertainties related to this phenomenon and consensus about the primary reason for the amplification do not exist.

Using climate models to both enhance the understanding of the complex climate system and make projections of future climate is essential for climate researches. However, as showed in studies by Walsh 2002, and Chapman and Walsh, 2007, reproducing the current Arctic climate is a problem for general circulation models (GCMs) and the discrepancies between different models are large. Holland and Bitz (2003) further showed in their model study that the projected polar amplification of temperature increase varies between two to four times of the global average increase at 2xCO₂ conditions in different models. This implies that the inter-model spread in future climate projections is the largest in the Arctic region.

It is known that climate change signal in the Arctic region is amplified by several strong feedback processes that are characteristics for the Arctic climate (e.g. Sorteberg et al. 2005) and difficulties simulating the Arctic climate related partly to insufficient understanding of these processes. The feedbacks are especially related to the radiation surface balance that is modified by presence of clouds and the special surface settings due to the semi-permanent ice sheet. As an example, snow/ice albedo feedback plays an important role in the radiation balance of the Arctic environment; warming of the climate reduces ice and/or snow cover, which reduces albedo and further enhances the warming.

This current study deals with the structure of the Arctic boundary layer during the transition from summer melt-conditions to fall freeze-up. During the Arctic summer, when the sun never sets, the additional energy input from the sun will melt the ice and snow instead of heating the surface. Energy loss will, in open-water conditions, freeze the water instead of cooling the surface. Thus, melting and freezing efficiently regulates near surface temperatures, which will remain close to freezing, and near surface moisture. This, together with the small variation solar radiation, implies that the diurnal cycle, which is fundamental for mid-latitude boundary layer structure, is often assumed to be weak or even absent.

Low level clouds are present at almost all times during the Arctic summer (Curry & Evert 1992, Tjernström 2005). Because of the unusual surface settings, clouds are the most influential factor in determining the surface energy balance in the Arctic boundary layer (e.g. Intrieri et al 2002, Sedlar et al. 2011) and feedbacks that are related to the clouds are potentially very important in the Arctic climate system. Intrieri et al. (2002) have further found that the Arctic low level clouds act to warm the Arctic surface for most of the annual cycle, in contrast to similar clouds at lower latitudes. Moreover, since both short- and longwave radiation are modified by cloud-radiative interactions (e.g Curry et al. 1996), the presence of clouds affect energy budgets at both the surface and the top of the atmosphere. This means that changes in cloud cover or in the optical properties of the clouds, which may be caused by a changing climate, can lead to large impacts of the energy balance in the Arctic.

All physical processes involving clouds (both micro- and macrophysical properties), radiation, sea ice and snow cover that are fundamental for the Arctic climate system, and are therefore essential to be represented correctly in climate and weather forecast models. However, because of their small
temporal and spatial scale these processes need to be parameterized in the numerical models, and how this is done may determine the quality and sensitivity of the climate models. Since our detailed understanding of these feedbacks and processes that are specific for the Arctic climate is still limited they are most uncertain, and to parameterize them for numerical modeling is therefore not an easy task to encounter.

One reason for our lack of understanding of the Arctic climate is due to the harsh environmental conditions and low population density of the Arctic region that limit surface observations. In-situ process-level observations in the central Arctic are much sparser than elsewhere. The parameterizations of the processes for numerical modeling, which also must be evaluated against observational data, tend to rely more on lower latitudes observation for development and validation (Tjernström 2005). Since these data may not be valid for the Arctic climate system, there is a great need for observations that can lead to a process-level understanding for the Arctic.

To provide in-situ process level data necessary to constrain or improve models, several of targeted experiments have been conducted in the central Arctic Ocean over the past years. Because of more accessible conditions most of these experiments have focused on the summer season. Between August to early September, 2008, the Arctic Summer Cloud Ocean Study (ASCOS) expedition took place in the central Arctic Ocean. One of the main aims of the meteorological programs of the expedition was to provide scientific data on clouds and boundary layer structure of the summer Arctic atmosphere, and during the expedition the vertical structure of the lower troposphere was continuously monitored at a high temporal resolution.

In this study we have focused the analysis on the vertical structure of the summer Arctic boundary layer during a three week long ice drift that was deployed during the ASCOS expedition. This ice-drift took place during the transition from summer melt period to fall freeze-up. Layering, in situation with boundary-layer clouds, has previously been observed (Herman and Goody 1976, Curry et al. 1996) and has been attributed to the interplay between a weak surface forcing, turbulence generation from cloud-top cooling and radiation. Since transport processes in the boundary layer, related to turbulence, leave a mark in the vertical structure, analyzing this structure for different meteorological parameters may lead to a further conceptual understanding of these processes that governs the interactions.

We have also studied the temporal variability of the boundary layer structure. Since different processes interact on different time scales, studying the diurnal cycle within the boundary layer can also indicate how the surface layer interact with the cloud layer and above. The magnitude and the height of any diurnal cycle may reach also indicate the strength of vertical mixing. Further, a clear diurnal cycle indicates an impact by solar radiation, while higher frequency variability is indicative of turbulent processes. Since turbulence is a key process, we have also analyzed different turbulence estimates within the boundary layer, and studied their intervariability.

We know that processes that occur in the Arctic are linked to both global ocean and atmospheric circulations, and changes in the Arctic climate are expected to have impact at lower latitudes (e.g. Overland et al. 2011). It is of great importance to increase our general understanding of the Arctic climate, and to be able to accurately represent Arctic meteorological processes in global models, both for weather forecasts and for climate change projections, globally and in the Arctic. This study is one of several that deal with the structure of the Arctic summer boundary layer. With only a limited amount of process-level data from the Arctic region each study result is contributing to the accumulated knowledge and enhanced understanding of the Arctic climate. Together with an enhanced conceptual understanding of the Arctic summer boundary layer, we believe that this type of study ultimately may lead to better descriptions of the processes that governs Arctic boundary layer interactions in climate models.
In this report we first provide a description of the ASCOS expedition and of the instrumentation and methods used for the analysis, found in Section 2. This is followed by an overview of the general meteorological conditions during the ice drift that is provided as a background to the study, in Section 3. In section 4 the results of the analysis are presented and discussed. The report closes with a summary and some conclusions drawn from the main results in Section 5.

2 Method

2.1 The ASCOS experiment

Arctic Summer Cloud Ocean Study (ASCOS) expedition was deployed on the Swedish icebreaker Oden and took place in the central-Arctic in the late summer of 2008. As described in the ASCOS overview articles (Tjernström et al., 2011 and Tjernström et al., 2012 (manuscript in preparation)) the primary objective of the expedition was to enhance the understanding on processes important for the formation and life-cycle of low level clouds and the role these play in the surface energy budget of the high Arctic, during the transition from late summer to early fall. To achieve this objective ASCOS was designed as an interdisciplinary project with teams specializing in meteorology, oceanography, atmospheric gas-phase chemistry, aerosol chemistry and physics, and marine biology.

A detailed description on the route and sampling conditions encountered during the expeditions can be found in Tjernström et al., 2011. In brief, the Swedish icebreaker Oden departed Svalbard on 2 August heading north. 10 days later, Oden moored to a 3x6 km ice floe, which then drifted for nearly three weeks. Oden returned to Svalbard by 9 September. The ice drift track, from about N87°21’ W01°29’ to N87°09’ W11°01’, as well as the full route of the ASCOS, is illustrated in Fig. 1. ASCOS had a few brief open water and marginal ice zone research stations before reaching the ice floe and many observations systems were operational also during the transit to and from the ice drift. For this study, however, only data collected during the ice drift are used. The ice drift research station was in operation roughly between 00 UTC (Coordinated Universal Time) 13 August and 00 UTC 2 September, or DoY 226-246, where DoY is “day of year”, defined as DoY=0.00 at 1 January 00 UTC.

One of the main aims of the meteorological program of the ASCOS expedition was to provide scientific data on clouds and boundary-layer structure. During the ice drift different measurements systems were deployed both onboard Oden and on the ice floe. The vertical structure of the lower troposphere was continuously monitored at a high temporal resolution and numerous of meteorological data were collected.
2.2 Instrumentation

The remote sensing instruments used for this study include a scanning radiometer, a sodar and a wind-profiling radar. These instruments were all used on the ASCOS expedition for the continuous monitoring of temperature, wind and turbulence profiles. For cloud data we have also used measurements that were provided by a combination of different sensors, such as a cloud radar and ceilometer. A detailed description of the measurement systems used for this study is given below, and a summary is provided in Table 1. A complete discussion on the meteorological instrumentation used by ASCOS can be found in Tjernström et al. (2012, manuscript in preparation).

Temperature measurements: To provide high temporal resolution vertical temperature profile measurements a 60GHz scanning microwave radiometer was deployed on the roof of the starboard bridge-wing of Oden. As described in Tjernström et al., 2012 (manuscript in preparation) the instrument operates on a wavelength insensitive to water vapor and clouds, and passively sense atmospheric brightness temperature averaged over some distance away from the sensor. It scans continuously over all angles to the horizon, and boundary layer temperature profiles are retrieved from this information. The signal from the radiometer is unique for a given temperature profile, but the temperature profile is not unique for a given signal. This means that the radiometer needs a “first guess” temperature profile for the retrieval. This first guess was taken from the soundings made every 6th hour during the expedition. The scanning radiometer temperature profiles therefore provide additional information beyond simple interpolation of the sounding measurements for the time periods between the soundings and at high temporal resolution up to ~700 m. The instrument measures, however, temperature up to 1200 m, but the sensitivity is reduced with height. Also, because of the scanning measurement technique and the geometrical considerations, the vertical resolution of the temperature profiles degrades by height. Close to the surface it is theoretically ~10 m, degrading
to 100s of meters above 500 m. The retrieved temperature profiles therefore gradually adjust to interpolated sounding profiles with height, as the sensitivity and resolution degrades. The scanning radiometer was in operation during the entire expedition, but we only analyze the ice drift.

**Wind and turbulence measurements:** A phased-array Doppler sodar system was deployed in a noise abatement shield on the ice floe. The sodar instrument sends audible sound pulses in and off the vertical, retrieving both backscattered power from the emitted pulses and their Doppler frequency shift. The Doppler shift can be used to derive the three-dimensional wind vector, usually averaged over many pulses (10 minutes) to improve signal-to-noise characteristics. The backscattered power is sensitive to temperature fluctuations and is in itself an indirect and relative measurement of turbulence. However, weak turbulence in large stratification can provide the same backscatter as strong turbulence in weak stratification, and any turbulence in neutral conditions will theoretically give zero backscatter. The spread (spectrum) of the Doppler shift in a measurement volume also contains information of the turbulence, as it will represent the variance of the velocity. Although coming from the same instrument, the backscatter and wind-speed variance are in some sense independent observation. Note that the Doppler requires backscatter, but as long as some backscatter is present, a Doppler signal does not necessarily follow; this also depends on the signal quality and geometric considerations. For example, a good signal sensed in one direction but a poor signal in another within the same measurement cycle may provide a backscatter result, but no consistent Doppler results and, hence, no wind information. The sodar has a high vertical and temporal resolution, but as it relies on backscattered sound generated by turbulence, which is attenuated with distance, it is mostly restricted to measuring winds within the boundary layer (<600 m).

For monitoring winds higher up, a wind-profiling radar operating at a frequency of 449 MHz was deployed onboard *Oden*. The operational principle of the wind profiler is similar to that of the sodar, except it uses radar wavelengths sensitive to absolute humidity fluctuations. The choice of wavelength is optimized for dry Arctic conditions but the performance of this instrument is still limited by the relatively low absolute moisture in the cool Arctic and therefore these results are averaged over a longer time to improve signal-to-noise characteristics. Due to the speed of radar wavelength radiation, compared to sound waves, the vertical resolution of the radar wind measurements is also lower.

Both the sodar system and wind profiler radar were gradually deployed after the launch of the ice drift and did not become operational until after the first few days. The sodar, on the ice, also had to be taken off line before the end of the ice drift, to be brought back on board for the journey back. Data from the sodar and wind profiler are therefore only available between ∼18 August-1 September (DoY 231-246).

**Cloud monitoring:** A suite of remote sensing instruments were installed on-board *Oden* for monitoring of clouds and cloud properties. The center-piece was a millimeter cloud radar (MMCR), complemented by laser ceilometers and a dual wavelength microwave radiometer. As described in Tjernström et al. 2012 (*manuscript in preparation*) the MMCR (35-GHz) instrument measures backscattered energy and Doppler velocity spectra from the hydrometeors in each measuring volume. From these spectra, the total backscattered power, the mean Doppler velocity and the Doppler spectrum width can be derived. The cloud ceilometers were used to detect the height of the lowest liquid water cloud layer. Atmospheric brightness temperatures, from which perceptible water vapor and liquid water path are derived, were measured by the dual wavelength (24/31 GHz) microwave radiometer.

A combination of all these sensors provides data on different types of cloud microphysical properties. In this study we only use cloud data to evaluate the cloud boundaries and radar-reflectivity. Clouds were monitored during the entire ice drift period.
Table 1: Summary of the meteorological surface-based remote sensing instruments used in this study. Reference: Tjernström et al. 2012 (manuscript in preparation)

<table>
<thead>
<tr>
<th>Variables</th>
<th>Instrument</th>
<th>Temporal resolution</th>
<th>Vertical range/resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vertical temperature profiles</td>
<td>Scanning microwave radiometer (supported by soundings every 6th hours for initialization of algorithm)</td>
<td>5 min averages</td>
<td>30 m up to 1.2 km variable vertical resolution: O(10 m) at surface to O(100 m) at 1 km</td>
</tr>
<tr>
<td>Wind speed and direction, boundary layer structure</td>
<td>Sodar</td>
<td>10 min averages</td>
<td>30 m up to ~600 m 10 m vertical resolution.</td>
</tr>
<tr>
<td>Wind speed and direction</td>
<td>Wind profiler</td>
<td>30 min averages</td>
<td>144 m up to 3 km 30 m vertical resolution.</td>
</tr>
<tr>
<td>Cloud reflectivity and cloud boundaries</td>
<td>Doppler MilliMeter Cloud Radar, ceilometers</td>
<td>10 min averages</td>
<td>95 m up to 14.3 km ~45 m vertical resolution</td>
</tr>
</tbody>
</table>

2.3 Analysis

For the analysis in this study we use observations from the ice drift (DoY 225-246) only. Two different time periods within the ice drift as well as the total ice drift period have been analyzed. The periods, determined by differences in surface and background atmosphere, were selected for their relative length and that they have interesting but different characteristic that can be contrasted to each other. The meteorological setting during the ice drift, and the selection and description of the time periods is provided in section 3.

We have focused on different properties in determining the vertical and temporal structure of the boundary layer: temperature, clouds features, wind speed and direction, and turbulence. The analysis was divided into three sections: the vertical structure; the diurnal cycle; and turbulence variability.

As discussed earlier, temperature measurements were provided by the scanning radiometer. The location of the sensor on top of the Oden infrastructure interfered with the lowest altitudes of the observations and thus only data from the lowest altitude (30 m) was omitted. With the continuously scanning operation of this instrument, one can use the data either as discrete profiles with high temporal resolution or as time series of temperature at different heights. In this study, both concepts have been used.

Vertical temperature gradients were calculated from these profiles using discrete differences over intervals of 75 meters to minimize noise. For the clouds, data for cloud boundaries and radar reflectivity are used. These data were provided by the cloud monitoring system described in the previous subsection and were obtained from the ASCOS data base. The sodar and wind-profiler instruments provided wind speed and wind direction profiles. The sodar has a higher vertical resolution than the wind radar, but since the data is dependent on backscatter from turbulence, its maximum height is limited. Data from the wind profiler radar is therefore used for analyzing the upper part of the boundary layer. The wind measurements systems were deployed a few days into the ice drift, which limits their use for the first period of the ice drift. Both instruments, especially the sodar, also suffer from periods with missing data at different time intervals and becoming more severe with height.
To obtain an overview of the boundary layer conditions for the different periods and to provide a background for the rest of the study, the vertical temperature, temperature gradient and wind profiles are analyzed by studying probability density function (PDF) profiles.

To analyze the structures within the boundary layer the results were normalized. This can be done in different ways, but for this study the height to the inversion base was used as a proxy for boundary layer height and the vertical axis for each profile is scaled by this height. A main inversion practically always existed and was also provided from the ASCOS database. It was estimated using the method described in Tjernström and Graversen (2008), which is a refinement of the method used in Tjernström (2007). Many of the profiles contained more than one inversion and the main inversion is defined as that with the largest stability.

To study the diurnal cycle of the different parameters, data was high-pass-filtered in time to remove variability at time scales longer than 24 hours using a Fast-Fourier Transform (FFT) filter. This is done as an attempt to remove variability caused by weather systems and other day-to-day changes in the conditions. Before filtering, the data was detrended to remove the lowest order of variance. The reason for the detrending is to eliminate the edge effects from the FFT technique, which inherently assumes an indefinite time series, here implying periodic repetitions of a limited series of data. An over-all trend in the time series is interpreted by the FFT as a saw-tooth shaped signal, which shows up as noise at the lowest possible frequency. The high-frequency data, referred to as anomalies, were block-averaged in 1 h windows according to the local time of day. For this, all data had to be time adjusted to the local time. This was done after the filtering but before the averaging, using the sample times (in Coordinated Universal Time, UTC) and the ships position. This data was also provided from the ASCOS data base.

The diurnal cycle is defined using the median of the diurnal cycle of the anomalies. The median is used since it is more robust, being less sensitive to outliers in the data set. The statistical significance of the diurnal cycle is tested using a double-sided Student’s t test at the 95% significance level. This test implies that the null hypotheses (that there is no diurnal variation) can be rejected at the 95% level of certainty if the zero-line line cannot be drawn inside the significance interval. However, it is important to recall that a statistically significant variation can be presented without being relevant; there may not be any physical meaning in the variation or it may be too small. The opposite is also possible; a physical meaningful variation may not be statistically significant, because of under-sampling.

In the analysis of the diurnal cycle, data points that were missing had to be treated before using the FFT and this was done by simple linear interpolation. For calculating the diurnal cycle, interpolation intervals of 30 min was accepted for the temperature and sodar measurements and 1 h for wind-profiler. For a day to be used in the diurnal cycle analysis, a requirement of a minimum of 30-40% of present data for that day was set up. After the days were examined, all the height levels were also checked with a similar requirement. This means that days, and levels, with insufficient amount of data were discarded in the calculation of the median diurnal cycle. Further, for a level to be included in the presentation of the result, all block-hour averages for that level had to contain information corresponding to at least 2 days. If insufficient amount of data was used in any of the block-average hour calculations for one level, that level was discarded.

For the turbulence analysis, three different estimates of turbulence obtained were used. Temperatures from the scanning radiometer were used to calculate the instant temperature variance, $T'$$^2$ (where $T = \bar{T} + T'$, and $\bar{T}$ is the mean temperature and $T'$ is the fluctuating part of the temperature composite). Data at each height was detrended and high-pass filtered with a cut-off frequency corresponding to 1 h. The instantaneous variance is then formed by multiplying each sample by itself and averaging over 30 minutes.

Further, the sodar also provided two turbulence estimates. First, the backscattered power, which
itself is a measure of the temperature fluctuations. Second, the variance of the vertical wind vector 
($\sigma_w^2$), which was derived from the spectrum of the Doppler shift from each measurement volume. 
This is a product from the sodar software. These two are both measured with the same instrument, 
but are not directly dependent. One comes directly from the backscattered signal, while the other 
comes from the Doppler shift in the backscatter.

Except for a tethersonde turbulence instrument, that was semi-continuously profiling the lowest 600 m of the atmosphere, there are no real in-situ turbulence estimates above 30, this means that these three measurements can be seen as “turbulence proxies” and the time-height cross section of the median diurnal cycle was calculated for all three parameters and compared. To estimate the accuracy of these different estimates as turbulence measurements, the spatio-temporal correlation between the different turbulence proxies was also studied. As a final step in the turbulence analysis a Richardson number ($Ri$) analysis is carried out. Richardson number is defined by

$$Ri = \frac{g}{\theta_0} \frac{\partial \theta}{\partial z} \left( \frac{\partial U}{\partial z} \right)^2,$$

where $\theta$ is the potential temperature, $U$ is the scalar wind speed and $z$ is the height above surface. $Ri$ is the quota between buoyancy and wind shear and can be used to predict the occurrence of turbulence, but it does not by itself indicate the existence of turbulence. In the analysis $Ri$ was provided from the ASCOS database and is a composite of different profiles all based on remote sensing. Statistics of the previously discussed three turbulence estimates are studied within different Richardson number regimes. Since a low $Ri$ indicate stability or low turbulence, and vice versa, one would expect a distribution of the turbulence estimates that indicate different turbulent states.

3 Meteorological conditions during the ASCOS ice drift

The surface and cloud conditions can be used to illustrate the atmospheric conditions during the ASCOS ice drift. For the surface conditions we use times series of 10 individual surface temperature observations and their mean during the ice drift (Fig. 2).

Cloud conditions are illustrated using cloud radar reflectivity, proportional to the particle size to the sixth power, to identify vertical locations of cloud and precipitation hyrdrometeors. Fig.3 shows a time-height cross-section of the radar reflectivity time series (upper panel) as well as the lowest detected cloud-base and lowest cloud-top heights (lower panel), also during the ice drift. More on the details regarding the atmospheric conditions during ASCOS and contrasting to results from other expeditions can be found in Tjernström et al. 2011.

Throughout the ice drift we see a gradual decrease in temperature with time, along with the occurrence of brief colder periods lasting for up to several days (Fig. 2). The colder periods, seen around DoY 235 and after 245, are associated with periodic breaks or reductions in cloud cover, as been studied by Tjernström (2005) and Sedlar et al. (2011). Generally, except for the two brief cooler periods, the temperatures during ASCOS were in the -2°C to 0°C range, which is the interval between the freezing point of saline (ocean) and fresh (snow on top of the ice) water, and is typical for this season. During the first 9 days the temperatures were generally higher, often close to the upper limit of 0°C, while during the week that precedes the actual freeze-up, it varies around the lower limit, -2°C. The near surface conditions were also very moist, with relative humidity often close to 100%, and almost no cases where it dropped below 90 % (Tjernström et al. 2011).
Figure 2: Time series of 10 individual surface temperature observations (black dots), measured during the ASCOS ice drift, and their mean (solid blue line). Vertical dashed lines show demarcation between the periods discussed in the text.

Figure 3: Time-height cross section of radar reflectivity (dBZ$_E$) taken from the MMCR cloud radar (upper panel) and time series of lowest cloud base and lowest cloud top (lower panel), for the ice drift. Vertical dashed lines show demarcation between the periods discussed in the text.
As analyzed by Tjernström et al. (2011) clouds, usually with a very low cloud base, were present during most of the ice drift with the average cloud fraction was about 90%; or 80% considering only the boundary layer. Inspecting Fig. 3, the cloudy conditions are obvious but differences with time can also be seen. Early during the ice drift low clouds are common but deep frontal clouds, associated with synoptic weather systems, also occur at irregular intervals, most frequent up through DoY 229. Although higher clouds occur also later, the deep clouds disappear, except for one cloud system on DoY 236. Besides some infrequent high clouds, the ice drift after DoY 234 is dominated by lower clouds, mostly <1km. The heights to the lowest cloud base and top also reflect this; the lowest cloud base is often low (Fig. 3, lower panel). While the cloud top varies significantly at the beginning of the ice drift it is associated with the lower clouds through much of the rest of the time.

A dominating feature in ASCOS was a pronounced neutrally stratified layer in the lowest atmosphere up to ∼500 m accompanied by high relative humidity (generally >95%) up to 1 km (Tjernström et al. 2011). This means that specific humidity also increases across the boundary layer inversion, which is a common feature in the Arctic summer boundary layer (Tjernström et al. 2004a; Tjernström 2005). This also suggests the possibility that clouds may penetrate into the lower inversion, rather than being capped by it (Sedlar and Tjernström 2009; Sedlar et al. 2012).

The ice drift period can be split into four, or five, regimes using near-surface air temperature, surface energy budget, vertical structure and synoptic conditions (Sedlar et al. 2011, Tjernström et al. 2011), as indicators. The transitions between the different periods are marked in Fig. 2 and 3 by dashed vertical lines.

The initial period (DoY 225-234, hereafter referred to as period 1) may be split into two separate regimes (225-229.5 and 229.5 to 234, respectively) as in Tjernström et al. (2011), but in this study we will consider this as one period. During period 1, the temperature generally remained near freezing point of fresh water (0°C) and conditions were typical for the melt season. There was a significant excess of surface energy that could melt ice and snow at the surface (Sedlar et al. 2011). As seen in Fig. 2 and Fig. 3 the first part of the period 1 was more variable in temperature and was more synoptically active than the second part of the period. Period 1 can also be characterized by high cloud fraction within the lowest kilometer, but multiple layers of clouds, some being deep, were observed. Several weather systems affected the period, which explains the deep frontal cloud structures seen in the cloud radar. A weather system around 20 August (DoY 233), at the end of period 1, marked the end of the typical melt-season conditions.

The second period (DoY 234-236.4) experienced a temperature decrease down to -7°C and the cloud-radar reflectivity (Fig. 3) reveals mostly low level clouds and fog, with lower reflectivity than measured during period 1. This period also had relatively large cloud fraction at the lowest altitudes, but the clouds were mostly limited to below 400-500 m. The low clouds dissipated around DoY 235, with the only identifiable clouds being the upper-level optically thin cirrus from 5 km to up to around 9 km at its deepest. Melt ponds and also some of the open (saline) ocean water surface froze. Riming and frost deposit onto the surface which increased the surface albedo from <70 to >80% (Sedlar et al. 2011). At the end of the second period a weather system left a fresh layer of new snow cover at the surface which altered the surface albedo sufficiently so that the melt conditions seen in the first period could not reestablish. We will not discuss this period in detail here, since it too short for our analysis, but insofar as we calculate properties for the entire ice drift, data from this time will also be included.

During the third period (236.4-244.5, hereafter referred to as period 3) temperature increased and maxima is found near freezing pint of saline ocean water (-1.8°C). The period was dominated by a lower stratiform cloud layer and characterized by relatively steady conditions. Conditions were governed by a quasi-steady high-pressure system, and the persistent stratocumulus layer residing in the subsidence inversion (Tjernström et al. 2011). The period thus had high cloud fraction below 1 km, but at times multiple layers were present with ice-crystals falling form a liquid layer near the cloud top (3). During this period the cloud contributed to the maintenance of the surface energy
balance close to zero through a larger cloud-radiative forcing than in the other periods (Sedlar et al. 2011).

In the final period (DoY 244.5-246), the stratiform low cloud layer decreased in depth and reflectively, due to an over-all subsidence, and became tenuous and eventually absent from the radar-reflectivity record. The surface energy budget then became negative (Sedlar et al. 2011), and the temperature rapidly dropped to a minimum around -14°C. The ice drift was terminated shortly after this, at 00UTC on 2 September. Similar to the second period, this period is not included in the analysis here.

The entire ice drift took place during the transition of melt to freeze period. Period 1 is a typical melt season period. The start of the actual freeze in the last part of the ice drift, did not coincide with the end of the melt, and the almost one week long period in between (period 3) can be considered a transition period. During this time the persistent low cloud layer prevented the surface energy balance from becoming negative, by its surface radiative forcing (Sedlar et al. 2011). As soon as this cloud cover dissipated, near the end if the ice drift, the energy balance turned negative, the temperature dropped rapidly and the surface froze.

The analysis in this study has focused on the first and the third period (period 1 and period 3), but the total ice drift period has also been examined. The two longer periods (1 and 3) were chosen due to their duration; they are almost equally long and sufficiently long for an analysis of, for example, the diurnal cycle. They also show different meteorological characteristics that can be contrasted to each other.
4 Results & discussion

4.1 Vertical structure

The vertical structure of the lower atmosphere was evaluated from many different instruments during the ASCOS ice drift. Here we have focused on the vertical structures of the temperature profile, wind speed and wind direction and how they vary during different periods. Since clouds play a great role in the Arctic boundary layer, cloud top and base height are also studied.

Fig. 4 a-c shows the probability of temperature as function of altitude for the entire ice drift period and for period 1 and 3, respectively. Fig. 4 d-f, also shows the temperature profiles for the same periods, but with vertical axis normalized by the inversion base height, $z_i$, which serves as a proxy for the boundary layer depth. Further, each temperature profile is shifted by a mean temperature for the whole layer $(T_{z_i} + T_0)/2$, where $T_{z_i}$ is the temperature at the inversion base, $z_i$, and $T_0$, that near the surface) before the statistics is calculated. Note, that this temperature retains its unit, thus this is not a proper normalization.

![Figure 4](image)

**Figure 4**: Upper panels: Probability of temperature as function of altitude for the entire ice drift (a), period 1 (b) and period 3 (c). Note difference in frequency axis. Lower panels: Probability of temperature shifted by the mean temperature in the boundary layer, as function of the normalized height for the entire ice drift (d), period 1 (e) and period 3 (f).

Starting with the non-normalized plots, the probability plot for total ice drift period (Fig. 4a) show that the near surface temperature varied mostly between -4°C and 0°C. However at higher levels we see several of temperature regimes within the ice drift; there is a multi-modal distribution. At ∼300 m the distribution is clearly bi-modal, while at ∼600 m tri-modal. Assuming that 1.2 km
is above the boundary layer at all times, free tropospheric temperatures vary between -5 and 2°C. This variability reflects both that the ice drift took place during a transition from melting period to autumn freeze and also the variability with different air mass. Studying period 1 and period 3 (4, b-c), separately seems to untangle some of the multi-modal distributions and the variability over the total ice drift period can now be more easily understood. Some of the distributions of the statistical maxima in the ice drift plot can clearly be seen in the two sub-periods. Generally, the semi-linearity in temperature decrease with altitude seen in the sub-periods, below ~300 m for period 1 and below ~400 m for period 3, indicates a well-mixed lower layer. During the warmer period 1, there seem to be a rather weak capping inversion around ~400 m while for the cooler period 3 the height to the inversion is larger and more variable, found at ~600-900 m. Consequently, period 3 has generally a larger temperature change across the lower layer than period 1. A double structure can also be noticed in period 3, with a frequent second mixed-layer around 800-900 m, just below the capping inversion. This upper mixed layer is likely associated with the stratocumulus cloud layer that was present for most of the time (Fig. 3) during this period.

In the normalized plots (Fig. 4 d-f) the variability seen in the first plots is eliminated, or at least greatly reduced. Since the mean temperature in the boundary layer only shifts the temperature in these plots, the temperature difference across the layer is maintained. This “reduced” temperature is by definition zero at a mid-layer height, which explains the tendency for a “global” maximum at \( z/z_i = 0.5 \); an artifact of this normalization method. The layer, from the surface to the base of the main inversion \( (z/z_i < 1) \), will from here on loosely be referred to as the “boundary layer”. However, in the proper use of this nomenclature this whole layer would have to be turbulent. As will be examined and discussed later, this is not necessarily the case here, at least not at all times. Nevertheless, for simplicity the term “boundary layer” will be used, since it is this layer that facilitates the coupling between the free troposphere, always above \( z_i \), and the surface. In the normalized profile for the entire ice drift (Fig. 4 d) some variability remains, but a common structure for the entire period appears. This indicates that much of the variability seen in the non-normalized profile is due to either differences in actual temperature of the boundary layer or its depth or, most likely, both. However, the overall vertical structure is still similar: temperature is decreasing quasi-linearly with height to the top of the boundary layer, where a capping inversion is present and temperature increases with height.

Examining the normalized temperature distributions for period 1 and 3 separately, both similarities and slight differences appear. During the period 3 the temperature profile is more linear through the whole layer than in period 1, where the temperature profile appear more concave with height. In period 1 we find somewhat more isothermal conditions above \( z/z_i \approx 0.6-0.7 \). The differences in the shape of the temperature profile within the boundary layer indicate that the third period probably has a more vigorously mixed boundary layer. The capping inversion in the period 3 seems more marked and sometimes stronger than in period 1.

Further, the slightly wider distribution in temperature closer to the surface seen in period 3, at \( z/z_i \approx 0.1-0.3 \), indicates infrequently occurring low-level inversion. This suggest a transient layering of the boundary layer during period 3 into two or more sub-layers. This feature is not present during period 1.

Fig. 5, shows the vertical temperature gradient as a function of actual altitude (a-b) and as function of normalized altitude (c-d) for period 1 and 3, respectively. The approximate moist adiabatic lapse rate \( (0.65 \cdot 10^{-3} \ \degree C \ m^{-1}) \) is marked in the plots (dashed line), which is relevant because of the very high relative humidity and presence of clouds in the Arctic boundary layer. A temperature gradient profile close to this the moist adiabatic indicates well-mixed layer with near-neutral conditions, while a profile with larger negative gradient indicates a more unstable stratification.

The two upper panels of Fig. 5 show, as with the temperature profiles, that a well-mixed layer near the surface can be found in both periods, with temperature gradient near the moist adiabatic, with
generally slightly larger negative gradient in period 3 than in period 1. The depths of the mixed-layer, however, differ significantly for the two periods. In period 1 the mixed layer reaches about \( \sim 350 \) m while during the third period the corresponding this depth is almost \( 1 \) km. The multi-layer structure seen in the temperature profiles can also be noticed here. There are near-neutral layers both close to the surface and around \( \sim 900\) m, but with a wider PDF, with a layer in-between that has a more consistent lower stability.

Studying the temperature gradient against the normalized height (Fig. 5, lower panel), two relatively similar vertical structures appear. Period 3 has more consistently a larger negative temperature gradient, below the moist adiabat, while first period generally has a less negative gradient. This is consistent with the conclusion that period 3 had a more vigorous turbulent mixing than the first. Further, in period 3 there is a strong and very sharp temperature inversion while for period 1 there is a deeper and weaker layer of increased stability right at the base of the inversion.

**Figure 5**: Probability of the temperature gradient as function of height (upper panels) and as function of normalized height (lower panels) for period 1 (a and c) and period 3 (b and d). The moist adiabatic lapse rate is indicated by dashed line. In the normalized plots the height is scaled by the height to the capping inversion.
Because of an incomplete dataset with insufficiently data sampled during the first period, wind speed and wind direction are analyzed for period 3 only. Fig. 6 shows wind speed and wind direction statistics, respectively, as functions of the normalized height. Since sodar is dependent on backscatter, missing data at higher altitudes is a frequent problem and the wind radar is used for higher levels. This means that the profiles in the figure comes from two different dataset. The transition from sodar to wind-profiler data is marked with solid black line in the figures. While both instruments use Doppler techniques, they apply quite different technologies, with different averaging volumes and times. In the wind speed profile there is a slight mismatch at the interface in range between the two instruments. However, this seems not to be a problem in the wind direction statistics.

Wind speeds seem to be quite uniform with height, except very close to the surface, varying between 2-6 m/s, with maximum around 4-5 m s\(^{-1}\), through much of the boundary layer. The distribution is slightly wider within the boundary layer than above, with a maximum width in the upper half of the boundary layer, which could indicate occasional occurrences of a low level jet.

Winds aloft were predominantly from north-northwest at \(\sim 30^\circ\). A directional shift of about 50°, to almost easterly, can be seen within the boundary layer. This feature can also be seen in the second probability maxima around 220°, i.e. wind from southwest, which shifts to southerly winds aloft. An explanation of this wind veering aloft is yet to be found. There is no obvious sign of an Ekman spiral although there is a slight increase of the distribution width approaching the surface around the 0-90°maximum and slight gradients to lower values at the other two maxima in the pdf. The lack of an Ekman spiral is also an indication of a well-mixed layer during period 3, although it is well known that this feature also requires long periods with stationary conditions.

As clouds are frequent during both periods, and have large impact on the boundary layer conditions in the Arctic, we also examine the vertical distribution of the clouds for both period 1 and 3. Fig. 7 shows the statistics for lowest cloud base and top height for both periods as function of the normalized height.

\[\text{Figure 6: Probability of wind speed (left) and wind direction (right) as function of normalized height, during period 3. The solid black line indicates the transition from sodar data (lower layer) to wind-profiler data (upper layer). The height is scaled by the height to the capping inversion.}\]
Figure 7: Probability of the normalized lower cloud base height (blue solid) and lowest cloud-top height (black dashed) for period 1 (left) and period 3 (right), respectively. The height is scaled by the height to the capping inversion.

Generally, both periods feature a most commonly very low cloud base. However, period 3 also has a secondary maximum slightly above the middle of the boundary layer \((z/z_i \sim 0.5)\), indicating a multi-layer structure. This could imply that there are occasionally low-level clouds present that conceal the higher cloud base from the cloud-base lidar detector, which is rapidly attenuated by liquid-water clouds. This two-cloud layer separation can be seen in the cloud-radar reflectivity plot (Fig. 3) around DoY 239, where both a lower layer cloud and a higher stratocumulus cover can be seen. Note, however that this two-layer cloud structure could be present at other times also, but was not detected. When precipitation falls from the upper cloud layer, the cloud radar reflectivity will mask out any possible intermediate layer, since the larger-sized precipitation particles will be preferentially sampled. The cloud-base ceilometer will be attenuated by the lower cloud layer and will therefore not sample the upper cloud-layer base.

Cloud tops varied significantly during period 1 because of the more variable weather conditions and often occurring deeper clouds, also consistent with the cloud-radar reflectivity seen in Fig. 3. During period 3 there is a distinct cloud top maximum around the top of the boundary layer, up to 1.5 times its height. This indicates that the boundary layer is capped by a single layer stratocumulus and also that the top often penetrated into the capping inversion. This is in contrast to similar clouds in the sub-tropic marine boundary layer, where the stratocumulus usually is capped by the inversion.
4.2 Diurnal cycle

In this section the results from the diurnal variability analysis for the ASCOS ice drift period are presented. The analysis focuses on period 1 and period 3 separately. The diurnal variability in temperature and stability, is a leading mode of variability in boundary layers. This occurs in response to larger solar heating of the surface in the middle of the day, leading to higher temperatures and a reduced stability near the surface. However, in the Arctic there are several of processes that could lead to a reduction of this cycle. First, due to the northerly location the sun is lower on the horizon and the solar zenith angle varies less than at more southerly latitudes and, second, during the period of melting of snow and ice, the excess heating from of solar radiation during the day tends to increase the melt rather than heat the surface. The magnitude of any diurnal cycle and height to which it may reach may indicate the strength of the vertical mixing in the boundary layer. If a diurnal cycle exist, but decays with height this indicate a warming that is inhibited with height due to the influence of stability.

Fig. 8 shows the median diurnal cycle of temperature anomaly (solid line) and the corresponding 95%-significance interval (dashed lines) at 75 m above the surface, for period 1 and 2, respectively, as an illustration. If a zero line can be drawn within the significance interval, the variability is said to not be statistically significant. If not, the so-called “null hypothesis”, that there is no diurnal variation, can be rejected at the 95% level.

Both periods obviously show a significant diurnal cycle at this height, but it is smaller and less significant during the first period. This is, however, expected since the near-surface temperature is often close to zero during this melting period (Fig. 2), especially during the second half. Generally, both periods experience a cooling from around midnight or somewhat earlier. In the third period temperature increases earlier (already at ∼06 Local Time (LT)) than in the first (∼10 LT). A decrease in positive anomaly, after the highest temperatures have been reached, can be found in both periods. Again, this happens earlier and faster in period 3, where there is even a cooler period between ∼13-16 LT.
To examine the temporal structure of the whole depth of the boundary layer, the median temperature anomaly is calculated as a function of height. Fig. 9 shows time-height cross section of the diurnal temperature anomaly for the two periods. The black solid lines illustrate the median cloud base and top height, measured and calculated independently from the cloud radar and ceilometer cloud-base data, for each period, respectively.

![Figure 9: Time-height cross section of the diurnal cycle of the median temperature anomaly profile for period 1 (left) and period 3 (right). The black lines indicate the diurnal cycle of the median cloud top and cloud base height.](image)

The diurnal cycle in temperature anomaly showed in the previous figure (Fig. 8) can now be seen up to ∼600 m in both periods, with some slight variations at the higher levels. The lowest temperatures occur in the early morning, followed by a period of positive anomaly. The onset of the warming is generally earlier in period 3 as seen in Fig. 8. The complex pattern with a shallow local cooling ∼13-16 LT in period 3 seem to be nicely coincident with a perturbation in the median cloud-base height.

In period 1 there is a shift in the cooling-warming pattern with height to earlier in the day above 600 - 800 m. Above 1 km there is an almost complete reversal in this pattern with warming ∼03-13 LT and cooling from ∼15 LT. Also in period 3, a complete reversal in the diurnal cycle can be found at around 1 km, with cooling during day (09-20 LT) and highest temperatures 21-06 LT. Here, again, the anomaly cycle is nicely coincidence with the estimate of the median cloud top. It is important to keep in mind that the accuracy of the temperature measurements made with the scanning radiometer at the highest levels are questionable. The vertical resolution and sensor sensitivity of the scanning radiometer measurements is gradually reduced with height and the results should not be over-interpreted. While, the reversal in the temperature cycle for period 1 and period 3 are both statistically significant, based on careful inspection of the data we believe that temperatures above ∼1km are likely mostly due to linear interpolation of the 6-hourly soundings. Still there is no reason
to disbelieve the latter.

Due to the large variability in cloud top height during period 1, caused by the passing of deep frontal clouds associated with passing weather systems, the median cloud-top cycle for this period is not statistically significant. However, it illustrates the large diurnal variation in cloud conditions. Note, that since this variability is due to passing weather systems, a much larger number of days than available would be required for stable statistics. The cloud base during period 1 is basically always below 100 m. For period 3, the diurnal changes in cloud base and top heights are all significant except for the cloud top changes from ∼17-23 LT. It should also be noted here that there is also a considerable variability in the clouds from day to day during period 3 (see Fig. 3). In terms of the absolute magnitude, the cloud base varies much less than the height to the cloud top. The diurnal variability in cloud top and base height during period 3 will be further discussed later.

Performing the normalization of the vertical axis and recalculating the diurnal cycle in the normalized height frame the picture changes somewhat. In Fig. 10 the time-normalized height cross section of the diurnal temperature variability is shown. Note that depending on the depth of the boundary-layer, observations from the lowest measurement height may temporarily be located above the lowest (normalized) heights in this analysis. This happens most often during period 1, leading to a reduction of normalized data at the very lowest levels.

![Figure 10](image-url)  
*Figure 10: Time-height cross section of the diurnal cycle of the median temperature anomaly profile for period 1 (left) and period 3 (right). The vertical axis was normalized by the inversion base height for each temperature profile used in the analysis.*
Period 1 now show a coherent diurnal cycle throughout the entire boundary layer depth, with higher temperatures from 10-11 LT and lower temperatures from early morning to the onset of the warming. This pattern is slightly delayed with height. The reversed pattern seen in Fig. 9 at higher altitudes is, however, not present in this plot. There are some possible reasons for this. First, the reversed pattern may be higher up than 1.5 times the boundary-layer depth and, second, there is no reason to believe that such a pattern, basically appearing in the free troposphere, should scale with boundary-layer depth. In this sense one may say that this normalization highlights processes in the boundary-layer, which it is designed to do, while suppressing information elsewhere in the vertical.

In contrast to period 1, period 3 has its phase-shifted signal within the boundary layer. In the lower half there is a slight positive temperature anomaly from ∼08 LT through the afternoon and into the evening, with a maximum around 21LT, and a relative cooling from around midnight to 08LT. Embedded into this, there is a local cooling at lower levels during midday hours (12-14 LT), but the amplitude is very small. In the upper 40 % of the boundary layer, in contrast, there is a cooling from 03 LT to 19 LT, with higher temperatures overnight, similar to the pattern in the 600-800 to 1km layer seen in the non-normalized profile (Fig. 9). Above the inversion ($z/z_i > 1$) the diurnal cycle pattern is complex and difficult to interpret. This is probably partly due to the low resolution of the scanning radiometer at these heights, which makes these results inaccurate, but also to the normalization.

To examine the variability in stability, the diurnal cycle in the temperature gradient anomaly is also analyzed, as function of normalized height (Fig. 11). Note that a positive anomaly implies a more stable stratification than the average, while a negative anomaly implies a larger instability. The interpretation of the results is quite difficult to carry out as the temperature gradient appears to be highly variable on shorter time scales, and many features seen in the figures are not significant. Nevertheless, in period 3 there seem to be a mostly negative anomaly in temperature gradient from midnight to local noon in the lower part of the boundary layer, and a more positive anomaly from noon to night. In the upper 40% of the boundary layer there is a mostly negative anomaly from ∼03 LT at the top of the boundary layer, and ∼07 LT further down, until around 18 LT. This layer is the same as layer that is cooling during the day in period 3 (Fig. 10). Thus, it seems as if there is a cooling at the top and warming below, which increases instability, or decreases stability, preferentially midday than compared to midnight where less unstable conditions are found.

In period 1 the diurnal cycle in temperature gradient is more difficult to interpret. Roughly, there seem to be an increase in stability from midnight to afternoon (∼16 LT), followed by a period of increased instability, in the lower half of the boundary layer. A reversed, somewhat similar but weak pattern can be noticed in the upper 50% of the layer. Again, the complex pattern in variability above the inversion, during both periods, has very low significance and is difficult to interpret.
The wind speed and direction are similarly analyzed for their diurnal cycles. As discussed in the method section, the analysis of wind data was performed only for period 3, due to insufficient amounts of data during the first period. Fig. 12 shows the diurnal wind speed and wind direction anomaly as function of the normalized height, respectively. Note that data comes from two different sensors, sodar for the lower levels and wind radar aloft, the interface distinguished by a solid black line.

There is a clear diurnal cycle in wind speed, consistent throughout the boundary layer in both the sodar and radar wind data, which are entirely independent measurements. The lowest wind speeds are found around local noon and higher winds from 15 to 19-20 LT. While the maximum is equally strong in both sensors, the minimum is weaker in the sodar results than for the radar winds. Similarly for the wind direction, the anomaly is positive during the second half of the day for the upper part of the boundary layer. This means that the boundary layer winds are closer to geostrophic wind direction, assuming that there is a counter-clockwise turning of the wind in the boundary-layer. However, the positive anomaly in wind direction has an earlier onset and is ending later than for the wind speed. The signal is also stronger in the radar and shifts significantly in time with height. Above the boundary layer there is a reversed directional variability cycle, with positive anomaly in the morning to afternoon and negative from afternoon all though the night.

In a steady-state framework, the wind at the interface to the free troposphere has to relax to the geostrophic wind. In a neutral boundary layer the direction deviates gradually more and more with decreasing height; this the classical so-called “Ekman spiral”. However, in a more unstable layer, with more vertical mixing, the wind direction is expected to become more uniform with height through the bulk of the layer, while the total angle to the geostrophic wind is generally smaller. A veering of the wind in time, to closer to the geostrophic direction, through a bulk of the boundary layer is
therefore an indication of a more turbulent boundary layer, where momentum flux from the upper parts of the boundary-layer is more effectively transported downwards.

Figure 12: Time-height cross section of the diurnal cycle of the median wind direction anomaly profile (left) and wind speed anomaly profile (right) for period 3. Vertical axis is normalized by the inversion base height for each wind measurement profile used in the analysis. The black line show the demarcation between sodar (lower layer) and wind profiler (higher layer) data used in the analysis.

There is not always a very good match in the diurnal analysis for the two instruments. However, in the lower part of the boundary layer (sodar data) there seem to be a directional variation that is somewhat close to the inertial period at this latitude (around 12 h), with negative veering from midnight to early morning and positive around noon, then again becoming negative in the afternoon to late evening and slightly positive from late evening to midnight. For the wind speed, the variability above the inversion also seems to be on a shorter time scale. Inertial oscillations in the wind can be triggered by alterations in the momentum budget due to stability changes. More analysis is required, using winds in the overlap between the instruments, to determine if differences across the interface between the sensors are a reflection of a true vertical variability or an artifact of using two different sensors.

There is clearly a diurnal cycle in the parameters that have been analyzed. Recalling the difference in mean cloud structure from Fig. 9, and noting the differential temperature cycle in the vertical, a relevant question is if the diurnal variability can be related to processes that are due to the clouds. Because of the more well defined cloud cover and a better data coverage during period 3, we will further focus of this period.
Fig. 13 show the median structure of the cloud boundaries (solid black lines), the inversion base height (red dashed line) and the cloud radar reflectivity (as seen in Fig. 3). Intense and deep vertically coherent echoes in the cloud radar data are mostly related to precipitation. As discussed earlier, the cloud base and top heights are significant except for the cloud top changes in the late afternoon into the evening.

Figure 13: A representation of the diurnal cycle in the cloud layer during period 3 showing the median cloud base and top height (black solid lines) and median inversion base height (red dashed line) and the median cloud-radar reflectivity (shading). Reflectivity is sensitive to particle size to the 6th power. Brighter contours typically represent ice crystals that are typically larger than liquid cloud droplets.

The cloud base is statistically higher between 11 and 16 LT and lower after that until 19 LT. At the same time the cloud top is on average lower, such that the over-all thickness is at a minimum from 11 to 16 LT. At this time the clouds seem to more often than average extent into the lower parts on the inversion, while condensate seems to more often fall out of the clouds (more radar echos below the cloud base) in the afternoon though early morning.

It should be noted, however, that it is not possible to say whether the variability in cloud base is due to an actual change in the base of the cloud or if, for example, a higher measured cloud base is in fact the result of a lower cloud breaking up, exposing a higher cloud layer to the instrument. Recall, that there was an occasional dual-layer structure of clouds during this period (Fig. 7). This means that a variability in cloud base could also indicate the presence of a lower cloud layer rather than one cloud layer that varies in thickness during the day. In that case, the median diurnal cycle of cloud boundary would imply that the double structure, with two cloud layers, is more seldom occurring during the middle of the day when the cloud base is higher.

Several processes may contribute to the formation or dissipation of a lower cloud layer. First, the depth of the entire layer is much deeper than what surface generated turbulence is expected to maintain, a lower layer of clouds may appear at the top of the lower mixed layer, somewhat independent of the top cloud layer. Second, precipitation from the top cloud layer may have two effects on a potential lower cloud layer. Evaporation of precipitation particles may add moisture to the lower layer, thus enhancing lower clouds. However, precipitation was observed to be mostly frozen, so that if temperatures are sufficiently low, this might instead have a dissipating effect on a lower cloud mostly consisting of super-cooled droplets, as frozen particles would be in an overstaurauded environment.
inside a liquid water cloud.

Since the cloud top generates turbulence, analyzing the diurnal cycle within the cloud is of interest. A new temperature time series was therefore generated by interpolation to be located at 100 m below the cloud top and its diurnal cycle was analyzed (Fig. 13). It seems like the cloud top is warmer than average from ~21 LT through the night until local noon, especially around 08-11 LT. It is then colder than average from local noon until night. This cycle is statistically significant and has rather large amplitude, about 2°C. Note, again, that the temperature measurements at this height are not well resolved by the scanning radiometer, especially early in the period when the cloud top was often above 1km. Also note that, as seen in Fig. 13, the cloud top is occasionally extending into the lower inversion so that at temperature time series constructed from a fixed altitude below the cloud top may be at different heights within a normalized boundary layer depth, and sometimes even inside the inversion. A proper explanation of this will require more analysis and remains to be found.

Figure 14: The diurnal temperature anomaly at a distance of 100 m from the cloud top for period 3. A new temperature time series was generated by interpolation to be located at 100 m below the cloud top. The plot show the median value (solid black) and the 95% confidence interval (blue dashed).
4.3 Turbulent variability

During ASCOS direct measurements of turbulence were conducted quasi-continuously in the surface layer (up to 30 m) and also with a tethersonde instrument package that was flown up and down through much of the cloud layer. Except for the tethersonde, however, there were no real in-situ turbulence observations above 30 m. In this section we will present results from an analysis of other sources of indirect turbulence information from the remote sensing instruments.

Four different turbulence estimates are used for this analysis, all based on surface based remote sensing instruments:

- Temperature variance obtained by high-pass filtering with 1h cutoff of the temperature provided by the scanning radiometer at different heights,
- backscatter intensity measured by sodar,
- volume averaged vertical wind-variance from the sodar, and
- a Richardson (Ri) number analysis from a composite of different profiles

As a background for the turbulence analysis, the power spectrum from the detrended scanning radiometer temperature time series is first studied. The power spectrum can indicate occurrence of high-frequency turbulence in the boundary layer as well as variability on longer scales. Fig. 15 show the logarithm of the temperature power spectrum (in color) as a function of height and frequency (x- and y-axes, respectively), for period 1 and period 3, respectively. As a reference, the frequency corresponding to diurnal variability (24 h) is marked with a dashed line in the figure.

![Figure 15: Logarithm of the temperature power spectra from the detrended scanning radiometer temperature time series (color) as function of height (y-axis) and frequency (x-axis), for period 1 (left) and period 3 (right). Vertical dashed line indicates the frequency corresponding to 24 hours. Note that the spectral power as a function of frequency is here multiplied by the frequency itself, as is common in meteorological applications.](image)

Analyzing these plots it is, again, important to recall that the sensitivity of the scanning radiometer declines with height. This means that a decreasing spectral power with height at the higher frequencies is expected. In period 1 it is clear that the high frequency spectral power decreases with height, which then is either because the turbulence declines with height or because the sensitivity of the instrument does (or both). However, in period 3 the high-frequency spectral power does not
continuously decrease with height. Instead it declines with height above 400 m and reaches a minimum between 700-800 m. Above this it again increases, within the 800 m – 1000 m layer; in fact the maximum power at the highest frequencies in this layer is higher than close to the surface. This increase in high-frequency spectral power is roughly coincident with the height of the mean cloud top layer during this period (Fig. 13). We can conceptually speculate that there is a partial layering of the boundary layer in period 3, where the lower layer is in contact with the surface and is mainly driven by shear-generated turbulence and is affected by the weak diurnal cycle in solar radiation. Aloft, in the cloud layer, turbulence is generated by buoyancy from cloud-top cooling. These two layers may sometimes be connected and sometimes not. With this conceptual model, we may also explain the decrease with height in high-frequency variance in the lower layer (below 400 m), in both period 1 and 3. Assuming that the sensitivity in scanning radiometer is approximately constant here, the decrease would then be an indication of decreasing importance of surface based turbulence with height.

Studying other variability scales a maximum around the diurnal variability in the same layer where minimum high-frequency power is measured, is found in period 3. In the 0-700 m layer, longer time scales dominate. Although highly speculative, the diurnal variation in the upper parts of the boundary layer in period 3 may be a sign that cloud-generated turbulence goes through a diurnal cycle and when it is strong it penetrates down to the top of the surface-based layer. In period 1 entire the boundary layer, seem to be generally dominated by low frequencies. These low frequency maxima are likely related to the very active synoptic weather systems during this period.

With idea of a two-layer turbulence structure in period 3 we have used the different estimates of turbulence to examine how they, separately, may indicate the existence of turbulence during this period. An analysis to determine if these estimates are accurate as turbulence measurement has also been performed. The results of these analyses are presented below.

The median diurnal variation as function of height for temperature variance, wind-speed variance and backscatter are shown in Fig. 16. These “turbulence proxies” are all directly derived from observations. Due to insufficiently amount of sampled data at higher altitudes, the wind-speed variance diurnal cycle can only be analyzed below 400 m.

![Figure 16](image_url): Time-height cross sections of the median diurnal cycle of three turbulence estimates: Temperature variance (a), vertical wind speed variance (b) and sodar backscatter (c), analyzed for period 3.
Both temperature and wind variances have a maxima below 200 m and roughly between ∼09 and 21 LT. The temperature variance increase seem to start earlier in the morning and also show some abrupt minima, at noon and at ∼18-20, while the wind speed variance maximum is more localized, between 09 and 21 LT but during this period extends gradually upward. We could here speculate and interpret this as a surface-based boundary layer with a diurnal cycle with the strongest turbulence during day and into the afternoon.

The layer of enhanced backscatter is also thicker during this period, but the magnitude has a very weak – if any – signal; close to the surface the median backscatter is nearly constant. In this perspective the temperature variance and backscatter do not show any signal above ∼500 m. Note that for temperature variance, data is available up to 1.2 km but only the lower 600 m is shown here. Analyzing the logarithm of the temperature variance up to 1.2 km, there is some weak signal at higher levels (not shown). This may be an indicator that there is a signal there, which is eroded by the averaging in the analysis of the diurnal cycle but is present in the spectral analysis seen in Fig. 15. The wind variance also seems to decrease drastically with height, but since there is a lack of data above 400 m no conclusions can be drawn.

To study the accuracy of these turbulence signals the spatio-temporal covariation between the three measurements are studied. In Fig. 17 scatter plots for the different combination of the data are shown. Correlation coefficients are calculated for the 30 min mean of the turbulence estimates, in the 0-600 m layer. Taking the logarithm of both measurements, there is some low correlation (0.48) between the of sodar backscatter and temperature variance. However, calculating the correlation for different heights and layers the skill is lower. It is plausible that the low signals in the upper layers, found in both data sets, contribute to the correlation. The correlation between temperature and wind variance is essentially zero, also when calculating this for different levels and layers. Backscatter and wind variance have some correlation (0.44) in 0-600 layer. A similar correlation between the variables can also be found in other layers as well, but any conclusions about this would need further analysis since these two measurements might not be all that independent after all.

**Figure 17:** Scatter plots illustrating the spatio-temporal correlation in the 0-600 m layer between the three turbulence estimates; temperature variance, vertical wind variance and sodar backscatter. The correlation between (a) temperature variance ($\log_{10}$) and backscatter ($\log_{10}$) was 0.48, (b) temperature and wind variance 0.03, and (c) wind variance and backscatter ($\log_{10}$) 0.44. The analysis is made for period 3.
As a final step in this turbulence analysis we also examined different Richardson number regimes and the probability of the previously discussed turbulence estimates within these regimes. \( Ri \) is a parameter that can be used to predict or analyze the occurrence of turbulence, but it must be born in mind that it does not by itself indicate the existence of turbulence. However, it can be a useful quantity to intuitively study different turbulent regimes in the boundary layer. It is also important to keep in mind that since the wind shear appears squared in the denominator in the expression for \( Ri \), these estimates are difficult to obtain, even from normally behaved wind speed profiles.

When interpolating these turbulence proxies to the \( Ri \) analysis stencil, one would expect that the distributions of any, or all, would be a function of the \( Ri \) interval in each regime; peak at a higher value for lower \( Ri \). Fig. 18 show the statistics of temperature variance, wind variance and backscatter within four different \( Ri \) regimes: \( Ri < 0 \) (unstable), \( 0 < Ri < 0.25 \) (weakly stable), \( 0.25 < Ri < 1 \) (moderately stable) and \( Ri > 1 \) (very stable). Only the results in the 0-200 m layer is showed, since this is where the signals are strongest (Fig. 16).

The statistical distribution for different \( Ri \) are virtually indistinguishable for temperature and wind speed variance. Some differences in the distributions for different \( Ri \)-regimes is found in the backscatter data. However, this outcome is unexpected; the lowest \( Ri \) should give the statistically highest backscatter and the result instead shows the statistically highest backscatter for the \( 0 < Ri < 0.25 \) regime while the distribution for \( Ri < 0 \) is approximately the same as for \( Ri > 0 \). It is known that low \( Ri \) may be generated by turbulence through mixing but can remain low long after turbulence has died. This means that a near-neutral, but non turbulent, layer can still give a low \( Ri \). From this data analysis it is impossible to determine if these results are because of low \( Ri \) does not necessarily predict turbulence or because all, or any, of the estimates and methods to detect or analyze turbulence fail for some other instrument related reason.

Further, with no obvious correlation between the three turbulence proxies (Fig. 17) it is difficult to say if any of these measurements can be used as estimates for boundary layer turbulence structure. Any conclusions based on these parameters should be qualified by great scepticism.
5 Summary and conclusions

In this study we have focused the analysis on the vertical and temporal structure of the summer Arctic boundary layer, using data from the ice-drift of the Arctic Summer Cloud Ocean Study (ASCOS) field campaign. ASCOS was a 40-day expedition to the high Arctic, north of 87°, during the summer of 2008, as a part of the International Polar Year. The ASCOS expedition took place from 2 August through 6 September, 2008, while the ice drift was deployed 12 August through 1 September.

Since transport processes in the boundary layer, related to turbulence, leave a mark in the vertical structure, analyzing this structure should lead to a further conceptual understanding of these processes that govern the interactions and, hence, ultimately to better descriptions in models. The vertical structure of the Arctic boundary layer was examined by studying the statistics of different parameters important for the boundary layer; temperature, wind and boundary layer clouds. A normalization of the height was done by scaling the altitude by the height to the lowest inversion, which serves as a proxy for boundary layer height.

The temporal variability within the boundary layer is studied to examine how different processes interact on different time scales. A diurnal cycle indicates an impact of solar radiation, while variability on short time scale is indicative of turbulence processes. The magnitude of any diurnal cycle and height to which it may reach also indicate the strength of vertical mixing. The diurnal variability of temperature, wind and cloud-boundaries was analyzed.

Also, with turbulence as a key process in understanding the vertical and temporal structure of the boundary layer, a turbulence analysis was also carried out. Four different indirect estimates of turbulence, all based on surface based remote sensing, were analyzed: 1) temperature variance, 2) backscatter intensity from the sodar, 3) vertical wind variance and 4) Richardson ($R_i$) number analysis. The intervariability of the three first estimates was studied by examining their diurnal cycle and the spatio-temporal correlation. Different $R_i$ regimes were then compared to the statistics of the three other turbulence estimates to study the assumed relationship between $R_i$ and turbulent state.

The analysis focused on two separate periods during the ASCOS ice-drift. Period 1, first ∼9 days of the ice-drift and period 3, ∼8 days almost at the end of the ice drift. These periods were chosen for their duration, they were almost equally long and sufficiently long for a diurnal cycle analysis, and they showed different meteorological characteristics that could be contrasted. Period 1 had near-surface temperature close to 0°C, characteristics of the melting season. This period was also characterized by very variable synoptic conditions. Period 3, on the other hand, featured more steady conditions with a persistent stratocumulus in a subsidence inversion, sometimes with multiple layers of clouds. The surface temperature was around −2 °C; this period was in the transition from the melt season to fall freeze-up. Due to logistical reasons some instruments did not become operational until after the first few days of the ice drift. Hence, insufficient data limits the wind and turbulence analysis of period 1.

Some of the main results of the analysis are summarized below:

- The vertical structure exhibits a well-mixed boundary layer during both periods, but indicated a more vigorously mixing and thus generally lower stability in period 3 than period 1. Period 3 also features a double - or layering - structure, with a second well-mixed layer close to the inversion. This second mixed layer is likely associated with the stratocumulus cloud layer.
- The boundary layer appears shallower in period 1 than in period 3 and the capping inversion is also weaker. The sharper inversion in period 3 is probably due to a combination of both the subsidence encountered during this period and enhanced mixing generated in the stratocumulus layer. During period 2, turbulent mixing seems to come mostly from the surface, thus the
efficiency of the mixing decreases with height.

• Analysis of cloud boundaries revealed that in period 3, the boundary layer is capped a single stratocumulus layer, the top of which often penetrated into the capping inversion. This is in contrast to similar clouds in for example the sub-tropic marine boundary layer, where the stratocumulus clouds are capped by the inversion. Further, this cloud structure implies that there must be substantial moisture above the inversion base, which is also in contrast to subtropical stratocumulus.

• There is a clear diurnal cycle found in the variables analyzed. The patterns become more apparent when the vertical axis is normalized by the inversion base height. This means that boundary layer processes control the cycle. The diurnal cycle can be said to be a function of the location within the boundary layer, and not the actual height above surface.

• For period 1 a significant diurnal cycle in temperature anomaly, with warming from mid-morning and cooling from night, was found. This cycle was uniform with height throughout the boundary layer, with a slight delay with height. This is contrary to expectations, since period 1 appeared to be less unstable and thus less well-mixed than period 3.

• Period 3 had a reversed pattern in diurnal cycle in temperature anomaly in the upper levels of the boundary layer compared to lower levels, with an upper level cooling midday and warming during afternoon and night. This upper-level reversed pattern was coincident with a less stable layer during midday, found in the diurnal cycle of the temperature gradient.

• Wind speed (only analyzed for period 3) also showed a distinct diurnal cycle uniform with height with higher winds in the afternoon. An inertia periodicity in the diurnal cycle of the wind direction anomaly was found in the lowest layers. This should be further examined and supported by other measurements.

• In power spectra of temperature for period 3 we found an indication of turbulence at higher levels, which could be related to the upper level cloud-induced turbulence. However, in the turbulence analysis we did not find any signs of this turbulence at these higher levels, and for all three turbulence estimates analyzed the signal decreased significantly with height. Neither backscatter nor measured wind-speed variance from the sodar reached above 600 m, and also the temperature measurements, used for the temperature variances profiles, has degrading vertical resolution with height. To further examine the turbulent state at higher altitudes (and higher temporal resolution) other measurement systems must be deployed.

• It is difficult to conclude if any of the turbulence measurements analysis can be used to determine boundary layer turbulence structure, and they should therefore be qualified by skepticism. No obvious correlation between the three turbulence proxies was found, even though they showed some resemblance in their diurnal cycle close to the surface. Further, the Richardson number analysis did not show a discrepancy between the occurrences of the different turbulence estimates for different $\text{Ri}$ number regimes. From this analysis it is therefore impossible to determine if these results are because $\text{Ri}$ is not a good prediction of turbulence or because all, or any, of the turbulence estimates to detect or analyze turbulence fail.
A diurnal cycle of some form is clearly found in all the parameters examined. In this study we have not examined the diurnal cycle of the surface net radiation or sensible or heat-flux anomaly. However, one can expect that the lower portion of the boundary layer is driven by these factors even though they are small. The mechanisms that account for the upper level cycle cannot be explained and understood from these results alone, especially since the turbulence analysis in this study did not give any verifying results of upper level turbulence. In a previous study by Tjernström (2007) data from Arctic Ocean experiment 2001 (AOE-2001) was used to examine the diurnal cycle of summertime central Arctic cloud-capped boundary layer. Results showed that near surface parameters were relatively steady, but a statistically significant diurnal cycle was found in the vertical. The diurnal cycle in temperature and temperature gradient anomaly found in period 3 is quite similar to that of previous study results, with a midday cooling layer above the warming lower layer, which enhances the instability in the upper part of the boundary layer. Vertical variability in wind speed and wind direction were not studied in Tjernström (2007). They hypothesized that this pattern was related to the coupling between the upper cloud layer and the lower surface layer. During midmorning the upper layer cool and the lower layer warm sufficiently so that the stability between the two layers becomes small enough to allow cloud-topped induced turbulent to mix down to the surface. In our analysis this could then explain the distinct diurnal cycle in the wind and wind speed, all throughout the boundary layer, found in period 3. Since this would require a well-mixed boundary layer, coupling between cloud and surface turbulence is a plausible explanation model. Since period 1 encountered many deep frontal clouds, the same hypothesis cannot be used as the only explanation of the vertical structure.

One apparent difference between period 3 in this study and the analysis in Tjernström (2007) is the boundary-layer depth, which is much larger here than for the AOE-2001 data. Thus, there is a possibility that an intermediate layer, between the top of the surface based boundary-layer and the turbulent layer might exist, while in AOE-2001 the top of the somewhat more stable surface-based layer was often found at the height of the cloud base. This intermediate layer is possibly sometimes mixed by turbulent eddies from the clouds and sometimes not, and thus, it would be intermittently turbulent.

Further analysis would be required to fully understand the differences that appear between the two periods. To compare the cloud-capped boundary layer structures in period 1 and 3 and the effects of low Arctic clouds on the diurnal cycle periods with frontal clouds should be eliminated from the analysis, as was done in Tjernström (2007). For a more exhausting study the layered structure period 3, turbulence estimates from the Doppler vertical motions from the cloud radar should be employed, supported by the more infrequent turbulence observations with the tetheredsonde turbulence instruments.

Finally, it is important to keep in mind that this study is performed using process-level data sampled only for a short period of time, during one summer in the Arctic region. Thus the question of how representative these results are for climatological Arctic processes must always be borne in mind. Several of other studies on the Arctic summer boundary layer have been done, and comparison and validation of the results should be made. However, further measurement campaigns and data analyses, such like this, are of great importance to enhance our understanding, and the certainty of previous results, of the Arctic climate and its small-scale processes that are somewhat unique to the region.
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