A climatology of supercooled layer clouds from lidar ceilometer data

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INTRODUCTION

It was demonstrated during the Cloud Lidar And Radar Experiment (CLARE'98) that supercooled water in the atmosphere can occur in the form of distinct layers several hundred metres thick that provide a strong signal for lidar, but because of the small droplet size are undetected by radar when ice is present (e.g. Hogan et al. 1999). The fact that these layers give such a strong echo at visible wavelengths implies that, when present, they are much more important in determining the radiative properties of the cloud than any ice that may be around. They will also be important in glaciation processes and therefore have an impact on cloud lifetime and precipitation formation. Currently, atmospheric forecast and climate models usually assume a simple ratio between ice and liquid water content that varies with temperature only; Fig. 1 shows this ratio for the ECMWF model. Clearly if such layers are common then there is a need to improve the representation of supercooled water in models.

Intensive observing periods, such as during CLARE'98, have served to demonstrate the existence of these layers and establish that supercooled liquid water is responsible. However, long-term measurements are required to determine their frequency and mean properties, ultimately what model climatologies would be tested against. In this paper a first attempt is made to characterise the frequency of supercooled layers using data taken from a 905 nm Vaisala CT75K lidar ceilometer, which has been operating almost continuously at Chilbolton, England since June 1996. Figure 2 shows a time-height section of lidar backscatter coefficient (β) from this instrument (resolution 30 metres and 30 seconds) through a typical layer with a temperature of around $-20°$C, on 15 October 1998 during

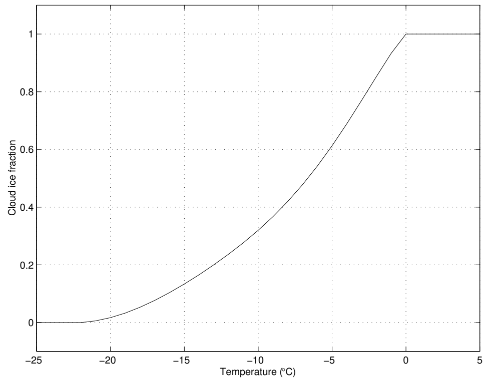

Fig. 1: The fraction of cloud water that is in the form of ice in the ECMWF model, as a function of temperature.

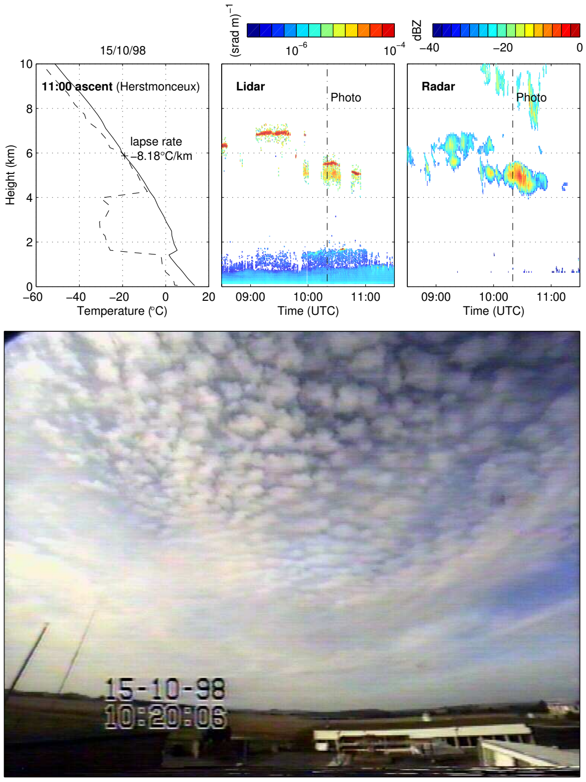

Fig. 2: Time-height section of lidar backscatter coefficient and 35 GHz radar reflectivity through a supercooled layer, with a simultaneous snap shot from the cloud camera. The radiosonde dry-bulb and dew-point temperature profile measured at Hersmonceux is also shown. The data were taken on 15 October 1998 at Chilbolton.

CLARE'98. Small cells characteristic of altocumulus are apparent in the accompanying snap shot from the cloud camera, indicating that these layers are convective in nature. No layer is visible in the accompanying observations by the 35 GHz Rabelais radar at Chilbolton (resolution 75 metres and 30 seconds), since the echos at this frequency are dominated by the contribution from the larger ice particles falling beneath the glaciating water cloud. Ice cloud is visible in the photograph as much fainter but more homogeneous wisps beneath the liquid water cells. The cells in this case were around 500 m across, too small to be resolved by the 30 s resolution of the lidar given the typical wind speeds at this altitude. The radiosonde ascent shows that the layer was saturated with respect to liquid water and convectively unstable. Heymsfield et al. (1991) presented aircraft measurements of liquid water in two altocumulus clouds and demonstrated using a numerical model that the observations were consistent with radiatively-driven convective overturning.

There are of course a number of limitations in using a ground-based lidar ceilometer. Firstly, it measures only lidar backscatter coefficient (β), and strictly one requires depolarisation ratio as well in order to distinguish between liquid water and ice with confidence. However, one of the main reasons these layers are important is because of their high optical depth and the consequent effect on radiation, so it could be argued that the phase of a particular layer of high β is irrelevant. For glaciation the phase is obviously of primary importance. A second problem is that ground-based lidars suffer very strong attenuation by low-level liquid water clouds, so most of the time in mid-latitude maritime climates cannot even see up to the 0°C isotherm. Furthermore, the layers themselves attenuate the signal making it difficult to identify multiple supercooled layers.

On a more cautionary note, it should be recognised that there is a possibility that not all layers of high β correspond to the presence of liquid water; Thomas et al. (1990) reported observations of relatively high β in ice cloud, the magnitude of which was seen to fall rapidly as the lidar pointing angle was moved a little away from zenith. This was interpreted as being due to specular reflection from horizontally-aligned plate crystals. Throughout the period of ceilometer observation in this study the instrument was operating in a zenith-pointing configuration, so could be affected. It would seem fairly safe to assume that the layers observed on 20 October in CLARE'98 (see Hogan et al. 1999) were composed primarily of liquid water droplets because of the low lidar depolarisation and the in situ verification. The radiosonde profile in Fig. 2 strongly suggests that the layer in this example is composed of liquid water. One striking property of these layers is that they tend to completely extinguish the lidar signal (see Fig. 3 for an example), whereas specular reflection from plates is only an enhancement of the backscatter, and the extinction should remain largely unchanged. Certainly the clouds observed by Thomas et al. (1990) did not strongly attenuate the lidar signal. An apparent layer was observed on 21 October 1998 during CLARE'98 at a temperature of around −20°C that had a high depolarisation ratio (indicating ice crystals) and according to the in situ measurements did not contain significant liquid water. Clearly more work is required to establish whether layers composed only of ice are common, but for the remainder of this paper we shall assume that they are all supercooled liquid water.

METHOD

Layers can be identified easily by eye from time-height sections of β, so the first step is to automate the process of layer identification using a set of fixed rules. The data acquisition system from this commercial instrument outputs the height of the first cloud base (h) in addition to the β profile. It calculates h by performing a so-called Klett inversion of the β profile assuming a fixed extinction-to-backscatter ratio, and considers the slope, absolute value and historic observations at that height. Comparing h from this procedure with the β profile indicates that supercooled layers identified subjectively *always* coincide with the first cloud base, but that when no layer or sharp gradient in β is present, the first cloud base tends to occur in the thickest part of any cirrus cloud that is present.

Hence we use h as the starting point for automatic layer identification, and do not attempt to identify more than one layer in each ray. Firstly, the height of the maximum β within 150 metres of h is found. Two tests are then applied that have been found to give best agreement with layers identified subjectively: a layer of supercooled water should have a value of β greater than 4×10^{-5} sterad^{-1} m^{-1} and this peak value should be at least 20 times greater than the value 300 m (10 range gates) above. An example of layer identification using this simple algorithm is shown in Fig. 3. An algorithm based on β alone was tried, but it was found that very cold layers could be missed while reflective clouds that were not layer-like in appearance, such as the lower parts of deep cirrus, tended to be included.

Radiosonde data was used to estimate the temperature at the altitude of the layer. The nearest operational upper-air station to Chilbolton is Herstmonceux, 125 km away, which carries out ascents every six hours. This station is used in preference to the so-called 'range' station at Larkhill, which is only 25 km away but does not perform regular ascents. Linear interpolation was performed in both time and height, but there is likely to be a residual error of several degrees in the derived temperature profile over Chilbolton.

RESULTS

The algorithm has been applied to all the ceilometer data taken at Chilbolton, from when the instrument was installed in the summer of 1996 until April 1999. Some data is missing, particularly in the first five months, but in total 2.47 million 30-second rays have been processed, equivalent to over 28 continuous months of observations.

We first consider the dataset as a whole to estimate the occurrence of supercooled layers as a function of temperature. The results are summarised in Fig. 4. Panel (a) shows the fraction of the dataset for which the instrument observed any cloud in each 5° temperature interval between −50°C and −5°C. Pixels were defined to be cloudy if the lidar backscatter coefficient was at least 2×10^{-7} sterad^{-1} m^{-1}. At temperatures warmer than −5°C the data were often contaminated by aerosol so are not shown. A method was devised to 'clean-up' the clear-air noise occasionally produced by this instrument. It can be seen that the occurrence of cloud in each 5° bin was less than 10% and decreased with decreasing temperature. This will be appreciably less than the true cloud occurrence, because of the problem of obscuration by lower level clouds at lidar wavelengths.

Panel (b) shows the fraction of clouds that contain a layer satisfying the definition given earlier, in each 5° interval. As one might expect, the fraction of clouds containing a supercooled layer decreases with temperature; 18.5% of clouds between −10°C and −15°C contain a supercooled layer, whereas between −30°C and −35°C the value is only 5.5%. The lower two panels depict similar information but in a cumulative sense. Panel (d) shows the fraction of observations with clouds colder than a given temperature that contained a layer colder than this temperature. We see that around 30% of

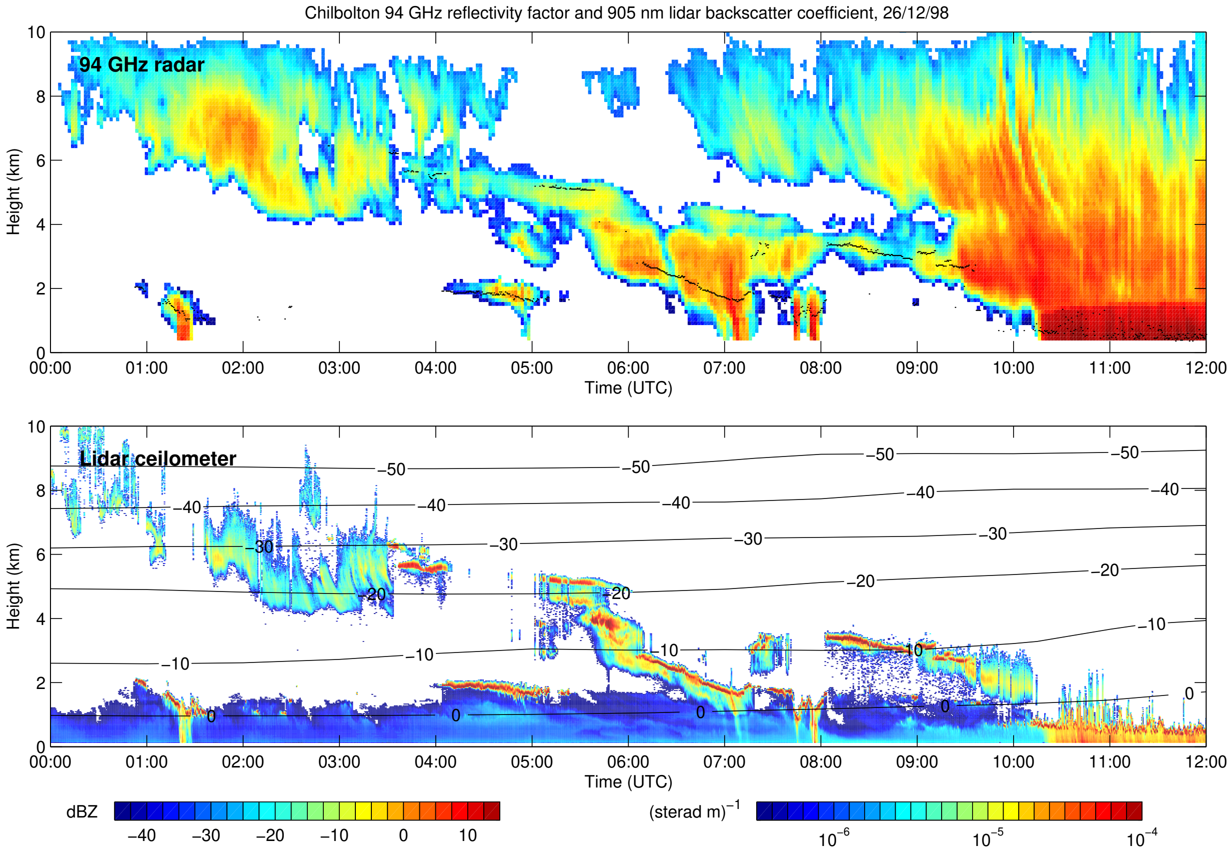

Fig. 3: Example of the objective layer-identification scheme. The top panel shows a time-height section of radar reflectivity from the 94 GHz Galileo radar at Chilbolton (resolution 2 minutes and 120 metres). Superimposed are black points indicating the presence of a layer, derived from the ceilometer data shown in the lower panel (resolution 30 seconds and 30 metres). Also shown is temperature according to the ECMWF model.

the time that cloud colder than −10°C was observed, a layer was observed in it, falling to 20% when considering only clouds colder than −20°C.

Figure 5 shows the mean layer duration and horizontal extent as a function of temperature. Horizontal extent was calculated from layer duration using the wind speed at that height as given by the interpolated radiosonde profile. Because of the frequent temporary obscuration of the layers by passing low level cumulus, layers were deemed continuous in this analysis if any gaps in them lasted no longer than 10 minutes. We see that at −5°C the average layer persisted for over half an hour, with the average duration falling steadily with decreasing temperature. Typical horizontal extents were between 20 and 30 km, although because of obscuration this is likely to be a considerable underestimate. In a few individual cases, layers associated with the tops of altocumulus were observed to persist for up to 9 hours.

It is apparent from Fig. 4b that the fraction of clouds containing a layer does not fall quite to zero at −40°C, where in theory no supercooled water should exist. There are a number of reasons for this; the most likely possibility is that the algorithm is wrongly identifying a very few high clouds as being supercooled layers when they are entirely composed of ice. Visual examination of the ceilometer data on such occasions suggests these events are aircraft contrails, which due to the large numbers of aerosols present tend to consist of high concentrations of very small ice crystals, so can understandably be mistaken for layers of liquid water. Indeed, Fig. 5 indicates that clouds identified as supercooled layers that are colder than −40°C persist on average for only three minutes. It is also possible that the temperature calculated by interpolating radiosonde profiles could be in error by in excess of 5°. In any case, layers colder than −40°C were diagnosed for only 4.9 hours of the 28 months of observations, corresponding to only 0.024% of the dataset.

An attempt was made to estimate the optical depth of these layers by performing a simple Klett-type inversion on each profile to remove the effects of attenuation. An extinction-to-backscatter ratio suitable to liquid water of 15 sterad was employed. However, all gate-by-gate procedures for correcting attenuated backscatter profiles are potentially unstable and very sensitive to both instrument calibration and the chosen extinction-to-backscatter ratio, and indeed our retrieved optical depths calculated by this method were often impossibly large. The mean optical depth of those layers for which the procedure did not explode was around 0.2, but given the problems with this technique it is doubtful that this value is accurate.

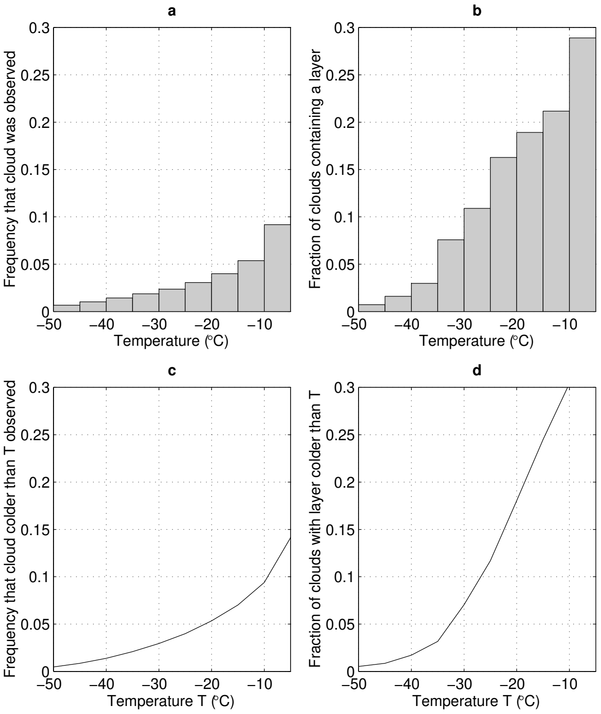

Fig. 4: Statistics from the full 31-month dataset: **(a)** Fraction of observations in which cloud was seen in each 5° temperature range; **(b)** Fraction of clouds that contain a layer in each 5° temperature range; **(c)** Fraction of observations in which a cloud colder than a given temperature is observed; **(d)** Fraction of clouds colder than a given temperature that contain a layer.

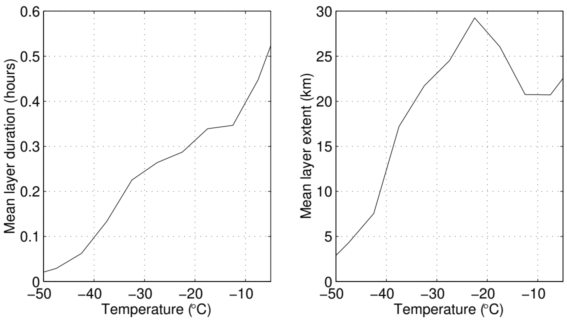

Fig. 5: Mean duration and horizontal extent of individual layers versus temperature.

The apparent physical thickness of the high β region was typically around 150 metres (5 range gates), but because of the strong attenuation the true thickness is likely to be greater.

We next divided the dataset into months to look for any seasonal or longer-term trend. A few months had too little time in which the ceilometer was operating to produce robust statistics, so have been rejected from this analysis. The remaining 31 months all have data equivalent to more than 15 continuous days of observations, and the average is equivalent to 27.7 continuous days. Figure 6 shows the fraction of clouds in three different 10° temperature intervals that contain a supercooled layer, as a function of time. We use 10° rather than 5° intervals in an attempt to reduce scatter. No robust seasonal or other

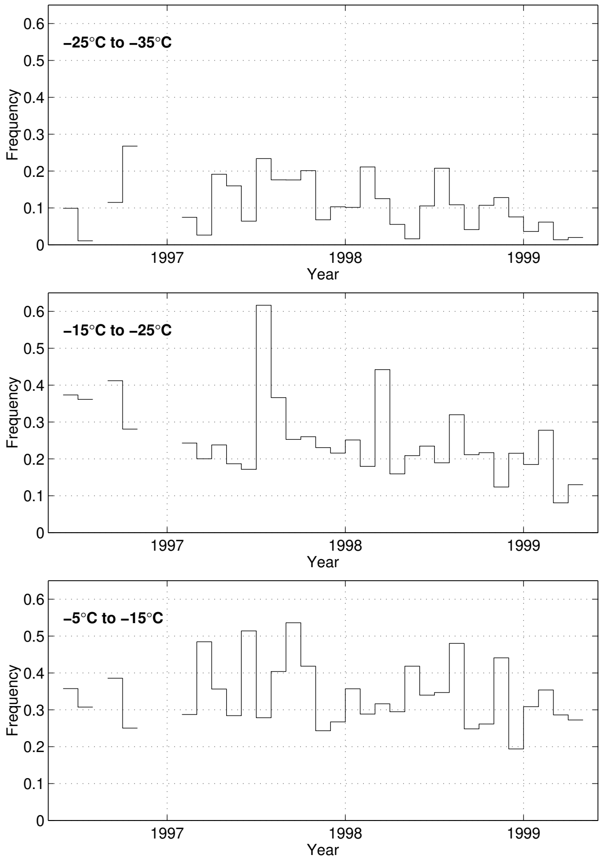

Fig. 6: Frequency of supercooled layers in three different temperature ranges, for individual months.

trend is obvious, but in any case the dataset can only really be considered continuous from February 1997, and it appears that 2 years is not sufficient to reveal any trend if one exists.

THE SUN-CLIMATE LINK

On a more speculative note, there is currently an ongoing controversy over the possible role of the sun in climate change, and a possible mechanism that has been suggested relates to the 11 year cycle in sunspots and solar-wind intensity moderating the flux of high-energy galactic cosmic rays at the top of the atmosphere. This much is well established, but the more speculative part is the suggestion that these cosmic rays can initiate glaciation, which in turn could be important in the development of weather systems. Observational support for this the sun-climate link has always rested on alleged correlations between sunspot activity and, for instance, rainfall over a particular location, but observational evidence to support particular mechanisms is conspicuously absent. Datasets such as in this paper could be used to support or refute the glaciation link because it is exactly the clouds that are claimed to be affected by cosmic rays that are being measured.

During the period of this dataset, solar activity was increasing, and therefore cosmic-ray flux decreasing. Hence if cosmic rays can cause supercooled water to glaciate and disperse, then their occurrence should increase through the data set. Certainly from this relatively short time series no such trend is evident (if anything there seems to be a slight trend in the opposite direction). In the $-25°C$ to $-15°C$ region where one might expect any signal to be strongest (because it is colder than the ice multiplication region and warmer than the homogeneous nucleation region), the frequency of occurrence is around 0.2, and there are some obvious 'spikes' corresponding to September 1996, July 1997 and March 1998. In all probability these are no more than statistical noise.

CONCLUSIONS AND FUTURE WORK

A first attempt has been made to characterise the frequency of supercooled layer clouds as a function of temperature. More work is required to verify that the majority of these layers indeed consist of supercooled water, but it is found that they occur surprisingly frequently; 30% of the time that the lidar sees cloud colder than $-10°C$ it also sees a layer colder than this. Given that they are much more radiatively important than any ice at the same altitude, and their role in glaciation and precipitation processes, it is important that some attempt is made to represent them properly in forecast and climate models. It would appear that a simple fixed ratio between ice and liquid water as a function of temperature is too crude to simulate the radiative properties of sub-freezing clouds. It would be useful to investigate with existing ceilometer data whether the occurrence of supercooled layers can be correlated with any large scale model field that could be used as the basis for a parameterisation. Much would also be learned by combining β with simultaneous measurements of radar reflectivity and lidar depolarisation.

The dataset was clearly not long enough for any seasonal or interannual trends to be evident, so it would be interesting to repeat the procedure for a longer period and at different sites.

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