Comparison of Cirrus Height and Optical Depth Derived from Satellite and Aircraft Measurements

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ABSTRACT

During the International Cirrus Experiment (ICE’89) simultaneous measurements of cirrus cloud-top height and optical depth by satellite and aircraft have been taken. Data from the Advanced Very High Resolution Radiometer (AVHRR) onboard the NOAA polar-orbiting meteorological satellite system have been used together with the algorithm package AVHRR processing scheme over clouds, land and ocean (APOLLO) to derive optical depth. NOAA High-Resolution Infrared Radiation Sounder (HIRS) data have been used together with a biscalp technique to derive cloud-top height. Also, the optical depth of some contrails could be estimated. Airborne measurements have been performed simultaneously by using the Airborne Lidar Experiment (ALEX), a backscatter lidar. Comparison of satellite data with airborne data showed agreement of the top heights to about 500 m and of the optical depths to about 30%. These uncertainties are within the limits obtained from error estimates.

1. Introduction

Clouds considerably influence the radiation interaction between space and atmosphere, and atmosphere and surface. They control the amount of energy that is available to ocean and land surfaces for evaporation and warming. It is well known (London 1957; Schneider 1972) that the annual mean net radiation flux at the top of the atmosphere (difference between absorbed and emitted radiation fluxes) is reduced in a cloudy atmosphere when compared to a clear atmosphere. This is generally supported by satellite observations (Ramanathan et al. 1989). However, the absolute values of the change of the net radiation budget vary, depending on the method applied, between $-27 \text{ W m}^{-2}$ (Ramanathan et al. 1989) and $-17 \text{ W m}^{-2}$ (Ardanuy et al. 1991). The radiation budget is determined by two counteracting effects: the backscattering of solar radiation to space and the thermal emission to space. If cloud coverage increases, the backscattering increases (albedo effect) and the thermal emission usually decreases (greenhouse effect). So clouds may either cool or warm the climate. Results of simulations of climate sensitivity and changes indicate the potential of changes in cloud amount and cloud height (Wetherald and Manabe 1988), cloud radiative properties (Roeckner et al. 1987), cloud condensation nuclei (Charlson et al. 1987), and the state of cloud water (Mitchell et al. 1989) as possible sources of climate feedbacks. High clouds usually warm the earth–atmosphere system; however, this depends strongly on their height, thickness, and microphysical composition (Stephens et al. 1990). Still in question is which optical properties (e.g., albedo and infrared emittance) shall be assigned to high clouds within the present climate as well as in a warmer climate, depending on meteorological field quantities like temperature, humidity, and wind. Improved field measurements and theoretical investigations are needed to solve this special parameterization problem of large-scale climate simulations.

An important test of the quality of a climate model and its results is the direct comparison of model results with observed clouds and their related optical properties. The latter can be obtained from aircraft and from satellite measurements that complement each other. A multispectral technique for determining cloud parameters has been in use at Deutsche Forschungsanstalt für Luft- und Raumfahrt (DLR) for a few years (Saunders and Kriebel 1988). This technique applies parameterized relations between radiative (reflectance) and derived microphysical (liquid water path) and optical properties (optical depth, infrared emittance). The accuracy of such derived quantities has not been finally assessed due to the variety of cloud types, ice clouds in particular. Further, a biscalp technique has been developed at DLR to determine the height of optically thin ice clouds (Pollinger and Wendling 1984). This paper validates these two techniques using airborne and surface lidar backscattering measurements coincident with NOAA-11 AVHRR and High-Resolution Infrared Radiation Sounder (HIRS) data. Section 2 discusses the synoptic conditions associated with the occurrence of cirrus clouds and contrails on 18 October 1989, during the International Cirrus Experiment.
(ICE) over the North Sea (Raschke et al. 1990). The retrieval techniques are described, which have been applied to the satellite and airborne data, and the results are presented in sections 3–5. Finally, satellite and airborne measurements are compared and discussed in section 6.

2. Synoptic situation over the North Sea on 18 October 1989

A low pressure system centered off the northwest coast of Ireland dominates the weather situation in the southern North Sea on 18 October 1989 (Fig. 1). Its cold front extends from north to south across the North Atlantic. In advance of the cold front, humid (80% relative humidity) and warm subtropical air is found near the surface within the warm sector. The surface visibility is accordingly reduced, especially in the ICE region where a ship (53°N, 5°E) reports only 4.4-km visibility and haze. Over France the altocumuli castellani clouds at medium level indicate a very unstable air mass there, because these clouds often develop into thunderstorms. Cirrus could be blown away with the southwesterly winds into the southern North Sea from the anvil of such a thunderstorm. This is one possible reason for the occurrence of cirrus without any clouds below in the ICE region.

There is another explanation for the existence of cirrus and why they remain in a moist layer around 9-km height. As confirmed by the lidar measurements, the cirrus clouds are just below the 300-hPa level (≈ 9.4 km). In Fig. 2, a weak high-level low is located over the English Channel (50°N, 1°E), while in 500 hPa there is none (not shown). Because this shallow upper low is slowly moving to the northeast, up to 4 K colder air (of −48°C removes air of −44°C) is advected in front of the trough in the upper-air region. Since there is warm-air advection near the surface, the whole air mass is strongly destabilized and the lapse rate increases remarkably. This means that once an ascent of an unsaturated layer is released, it will ascend until saturation is reached and cirrus spreads. Because it is only a shallow upper low, the upward motion is restricted to the upper area, and therefore, only cirrus clouds and no medium or low clouds occur. The upward motion is forced by positive vorticity advection (PVA) increasing with height. The region of PVA is obtained from the cyclonic curvature of the isohypses over France and the Netherlands, and the cyclonic wind shear in the southern North Sea. The shear vorticity is related to the rate of change of wind speed at the same wind direction as is reported from a ship east of England (53°N, 2°E) with 15 kt and from Hannover (52°N, 10°E) with 35 kt. Both effects force PVA, which in turn generates a divergence in the upper-air level and causes an area of upward air motion. Since the whole upper low pressure system moves slowly to the northeast over the southern North Sea, the occurrence of cirrus in this area is obvious.

Further, the vertical profile of the relative humidity over Norderney (54°N, 7°E) is explained by the foregoing discussion (Fig. 3). The humid surface layer with high moisture and low visibility is only 300 m thick. This information is used to complement surface lidar measurements (cf. section 6a). The layer above has relative humidities between 25% and 50%, followed by a humid layer between the 7.5- and 10.5-km heights with a maximum of more than 80% relative humidity over ice (dotted line). In this moist upper-air layer the cirrus clouds occur. In addition, the water vapor exhausted by aircraft can saturate the air, and contrails may form.

3. Airborne measurement of cirrus optical depth and height

Airborne lidar systems like ALEX (Airborne Lidar Experiment) (Mörl et al. 1981) allow the height of
even very thin ice clouds to be derived from range-resolved backscatter measurements along the flight path. From an inversion of the measured signal, the optical depth can also be estimated. But the usual inversion methods (e.g., Klett 1985; Ruppersberg and Renger 1992) are applicable in cirrus areas only with additional—but frequently not available—information about the extinction coefficient of at least one volume element with a sufficient lidar signal, as well as about the lidar ratio mentioned in the following. To overcome this problem, a special “shadow technique” was developed (Ruppersberg and Renger 1992), which is outlined in the following.

Height and contours of clouds are readily apparent in a color-coded presentation of the range-corrected counts \( D(x, r) \) depending on the distance \( r \) from the airborne lidar device, either above or below the position \( x \) of the flight path (cf. Fig. 6), where

\[
D(x, r) = C(x, r) \frac{r^2}{r^2_e}, \tag{1}
\]

and \( C(x, r) \) are the measured counts of the backscattered laser radiation from an interval \( \Delta r \) (in this case 7.5 m) at mean distance \( r \) from the lidar device; \( r_e \) is a fixed reference distance (e.g., 1000 m).

The optical depth \( \delta(x, \Delta r) \) of the considered interval \( \Delta r \) and, hence, the extinction coefficient \( \sigma_e(x, r) = \delta(x, \Delta r)/\Delta r \) at this point are related to \( D(x, r) \) by the well-known lidar equation, which can be written as

\[
\delta(x, \Delta r) = K\left(\frac{\sigma_e}{\beta}\right) \frac{D(x, r)}{\tau^2(x, r, r_L)[1 + Q_{ms}(x, r)]}. \tag{2}
\]

The system parameter \( K \) contains all geometric, optical, and electronic parameters of the lidar system that affect the signal. It is difficult to measure and has no long-term stability for mobile systems. The backscatter coefficient \( \beta \) is the value of the scattering function of the considered volume for the signal scattering opposite to the direction of the incident laser beam; \( \sigma_e \) is the extinction coefficient of the same volume. The lidar signal is caused by backscattering and, along with the primary laser pulse, is diminished on its way through the atmosphere by extinction (= scattering + absorption). Whereas \( \beta \) and \( \sigma_e \) in the troposphere and at wavelengths between the visible and the near infrared vary by more than 6 powers of 10 each, the so-called lidar ratio \( \sigma_e/\beta \) varies by less than 2 powers of 10. In pure molecular (so-called Rayleigh) atmospheres the variability is \( 8\pi/3 = 8.38 \); in all other cases, average values are used that, of course, may considerably differ from actual values: 14 is typical for cirrus, 20 for water clouds, 23 (50) for fresh (aged) volcanic dust, and 30 for maritime, 48 for tropospheric, 50 for rural, and 67 for urban aerosol (Pinnick et al. 1983; Dubinsky et al. 1985; Jursa 1985; Kaestner and Quenzel 1987; d’Almeida et al. 1991). Thus, maritime aerosol, for example, causes about 2.1 times the extinction per unit of backscattering of a typical cirrus.

The direct transmittance \( \tau(x, r, r_L) \) is the ratio of the directional laser energy (i.e., without the scattered nondirectional part) at the point \( x, r \) within the considered interval \( \Delta r \) to that laser energy that has entered the measurement path at the point \( x, r_L \) close to the lidar device. The ratio of multiple to single scattering of the received radiation is \( Q_{ms} \). In cirrus clouds, this ratio cannot be neglected, because of the extremely steep forward scattering of the ice crystals. In this paper, the error in \( Q_{ms} \) is estimated by computing the \( Q_{ms} \) with different scattering functions for ice particles obtained from the literature (Wendel et al. 1976) and by taking its variance. Multiple scattering is approximated by taking into account secondary scattering only.

Assuming the product \( K(\sigma_e/\beta) \) being determined explicitly or implicitly by additional information (e.g., on the lidar ratio and the extinction coefficient of a volume element with sufficient lidar signal), and \( Q_{ms} \) being estimated as mentioned, Eq. (2) can be solved by successive application of Eqs. (2) and (3), starting at the distance \( r_0 = \Delta r/2 \) and \( \tau(x, r_0, r_L) = 1 \):

\[
\tau(x, r + \Delta r, r_L) = \tau(x, r, r_L) \exp[-\delta(x, \Delta r)]. \tag{3}
\]

But, as mentioned at the beginning of this section, such additional information frequently is not available in cirrus areas. A solution to this problem is given by the “shadow technique.” Assume the measured area to be divided into two horizontal layers. The first, close to the flight path, may have an optical thickness of about 0.1 and may be inhomogeneous in the horizontal \( x \) and the vertical \( r \) directions. The same is valid for the second layer, but the inhomogeneities in the second layer shall not be correlated with those in the first; the layers, for example, shall not include the same convection cell. Nevertheless, the measured lidar counts \( C(x, r) \) and \( D(x, r) \) (Eq. (1)) are contracorrelated in the \( x \) direction, because large optical thicknesses in the first layer diminish the signals in the second, and vice versa. Thus, the product \( K(\sigma_e/\beta) \) can be estimated.
and the lidar system is calibrated by solving Eq. (2) with this product increasing from small values until reaching the point at which the contracorrelation turns into correlation. As $K$ is constant during the measurement, we get the $x$ variations of the otherwise unknown lidar ratio $\sigma_a/\beta$ in the shading area, additionally to the optical thicknesses and the extinction coefficients, but to get these parameters within the shaded (or any other) area we have to assume its $\sigma_a/\beta$ ratios in relation to those within the shading area.

During the ICE mission 216 on 18 October 1989, backscattered signals were received from cirrus clouds with the upward-pointing airborne lidar system ALEX. Occasionally, these signals were shaded by a contrail underneath. Using the shadow technique, the product $K(\sigma_a/\beta)$ and the spatial variations of the lidar ratio $\sigma_a/\beta$ within the shading contrail could be determined with an accuracy of $\pm 20\%$. That means that the corresponding products of each $x$ interval vary within this range, depending on the position and width of the $x$ interval, on different assumptions concerning $Q_{\text{ms}}$, and on some other assumptions of minor importance. For the extinction coefficients and optical depths of the investigated cirrus (cf. Fig. 6), which were measured about 1 h prior to the NOAA (National Oceanic and Atmospheric Administration) overpass, a $\sigma_a/\beta$ ratio two times higher than that of the contrail flanks was used. This is equivalent to a reasonable variability between an average value of the $\sigma_a/\beta$ ratio of 14 in ice clouds and a maximum value of 28. The latter has been used in this paper to match the measurements. However, these data depend only a little on this ratio because the optical depth of the investigated cirrus was very low, so that the haze underneath contributed more to the NOAA AVHRR signal than the cirrus itself. This will be shown from the analysis in section 6a.

The haze extinction coefficients below the flight level had to be extrapolated from the layer between the aircraft and the cirrus, because downward lidar measurements had not been performed. For this extrapolation, a constant ratio of aerosol to molecule extinction was assumed. This seems to be an acceptable approximation outside thermal inversions (Johnson et al. 1979). For the haze, the value for maritime aerosol, which is 30, was taken. This corresponds to a $\sigma_a/\beta$ ratio 2.1 times greater than that of the contrail flanks. The haze extinction coefficients were extrapolated from the laser wavelength 1064 nm to the center wavelength of the NOAA AVHRR channel 1, which is 630 nm. Fortunately, the resulting haze optical depth could be compared with surface-based measurements of the optical depth (cf. section 6a).

4. Satellite measurement of the optical depth

The optical depth relates micro- and macrophysical properties of a cloud. Amount, size distribution, and index of refraction of the cloud droplets bring about the optical depth that in turn determines transmission and extinction properties. Thus, reflectance and emittance are related to optical depth and to the liquid or ice water path. This opens the possibility to derive the optical depth in the solar spectral range from measurements of the reflectance. The validation of such parameterized relations can be made either by statistical techniques using routine surface observations resulting in rather coarse spatial resolution or by dedicated measurements such as aircraft measurements, for example. A few case studies using airborne data to validate parameterized relations between cloud reflectance and optical depth have been published. Kriebel et al. (1989) have used a simple relation for water clouds but with a correction for ice clouds. More recently, Minnis et al. (1990) have used radiative transfer calculations to relate reflectance to optical depth and have also obtained good agreement. In this paper we will present another comparison of satellite-derived optical depth with airborne data.

The algorithm package for detection, classification, and analysis of clouds, which is applied in the case study reported here, called APOLLO (AVHRR processing scheme over clouds, land and ocean), has been developed at the Hooke Institute in Oxford, United Kingdom (Saunders and Kriebel 1988), and is maintained at DLR in Oberpfaffenhofen, Germany. It uses all five channels of AVHRR 2. To find out whether a pixel is cloud free or not, five threshold tests are applied. These are a reflectance threshold test, a temperature threshold test, a channel 2 to channel 1 reflectance ratio test, a channel 4-minus-channel 5 threshold test, and a channel 4 spatial coherence test. A pixel is called cloud-free only if none of the tests detects clouds. Further, each cloudy pixel is tested whether it is fully cloudy or not. This is determined by applying the channel 2 to channel 1 reflectance ratio test a second time and assigning a narrow interval to fully cloudy pixels as well as by using the spatial coherence test a second time with a different threshold. This divides the pixels into three groups called cloud free, fully cloudy, and partially cloudy. If snow or ice are to be expected, a modified procedure is used that increases the number of distinct groups to eight (cf. Gesell 1989). Channel 3 reflectance is used to determine the phase, together with the difference of channel 1 and channel 2 reflectances. The fully cloudy pixels are then classified into low, medium, high, and ice clouds using a temperature threshold. Ice clouds are defined as dark (thin) and cold (high); that is, only cirrus clouds are classified as ice clouds. For all fully cloudy pixels, the reflectance of the cloud in AVHRR channel 1 is calculated, that is, in the visible spectral range. This calculation is not trivial (cf. Kriebel et al. 1989), particularly because of the anisotropy correction required and the contribution of the surface to the measured signal. This reflectance is related to the optical depth by an empirical parameterization scheme for water clouds (Stephens 1978),
which has been modified for ice clouds. The reflectance of ice clouds is usually higher than that of water clouds at the same optical depth. An appropriate conversion factor has been determined by Platt et al. (1980). Contrails are identified as clouds, but not as fully cloudy pixels, because they are usually narrower than about two pixels.

The uncertainty in optical depth estimates using AVHRR data and APOLLO has several sources. The calibration of the AVHRR is neither very accurate nor very stable (Holben et al. 1990), which leads to ±10% uncertainty in the retrieved optical depth. Also, the anisotropy correction causes about 10% uncertainty because the correction factors from Taylor and Stowe (1984) are zonal averages and do not represent their individual variability. The uncertainties due to the incorrect ozone transmittance and incorrect surface albedos of water and vegetated land surfaces are small. The application of the Stephens parameterization to optical depths less than 1 causes another ±15% uncertainty, because of the necessity to extrapolate an adjustment factor that describes the backscattering coefficient of the phase function. The conversion factor required for using the water cloud parameterization (Stephens 1978) for ice clouds depends on solar elevation and particle spectrum, is not well known, and gives rise to another 20% uncertainty (Platt et al. 1980). Errors in navigation can be neglected compared to the previous uncertainties. The cloud displacement during the time between satellite measurement and validation can be accounted for by means of radiosondes to determine the wind speed in the cirrus height. All these uncertainties result in an rms error of about ±30%. This is the uncertainty of the optical depth of the whole atmospheric column from the surface to the top of the atmosphere, which includes clouds, aerosols, and molecules. This has to be considered for any comparison of satellite-derived optical depths with airborne lidar measurements.

5. Satellite measurement of the cloud height

Height, temperature, and emissivity of cirrus clouds are of climatic relevance (Schneider 1972). Therefore, satellite-based remote sensing of these quantities is essential for climate research. Height and temperature determination from satellite in comparison to airborne lidar measurements has been addressed by Pollinger and Wendling (1984), and in a number of recent papers related to the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE), as, for example, by Minnis et al. (1990) and Minnis et al. (1992).

The determination of the temperature of semitransparent cirrus clouds is difficult because the radiation measured at the satellite contains both the thermal emission of the cloud and the radiance from below the cloud, which is transmitted through the cloud. Therefore, it is not possible to invert a radiance measured at the satellite to obtain the temperature of a semitransparent cirrus.

In the following a method is presented to determine the temperature of thin cirrus clouds. It is based on the exploitation of the information from two spectral channels in the thermal infrared range for two contiguous pixels within the cirrus cloud. Thus, no clear-sky radiances have to be used, which is the main difference to the CO2 slicing technique proposed by Smith and Platt (1978) or Wylie and Menzel (1989). The idea of the bispectral approach was first published by Cayla (1978) and was extended and improved by Pollinger and Wendling (1984), and Meerkötter and Wendling (1990).

The algorithm to determine the temperature is derived from four equations that result from the radiances measured at the satellite from two pixels in two spectral channels. Each of the equations can be written in a simplified form as

\[ L = L^* \tau + (1 - \tau)B, \]

if the emission of the atmosphere above the cirrus and the reflection by ice crystals are neglected. Here \( L^* \) is the emission of surface and atmosphere below the cloud, \( \tau \) is the transmission of the cloud, and \( B \) is the Planck emission according to the cloud temperature. After rearranging,

\[ B_2(T) = \frac{L^*_2 L_2^2 - L^*_1 L_1^2}{L_1^2 L_2^2 - f(L_1^2 - L_2^2)}. \]

Here \( B_2 \) is the Planck radiance that depends on the unknown cloud temperature \( T \), and \( L^* \) are the radiances measured at the satellite, where the lower index in Eq. (6) refers to wavelength, and the upper to the pixel. The factor \( f \) results from the relation of the Planck functions in both spectral channels.

There are three essential conditions to derive Eq. (6). First, the cirrus cloud must totally fill the satellite pixels and the transmission must be different in the two pixels, whereas the cloud height, that is, the cloud temperature, is the same. Second, the two spectral channels have to be chosen such that the ratio of the spectral cloud transmittance in the two pixels is the same for both spectral channels. This can be checked by numerical simulation. Third, the upward radiance at the cloud base is the same in both pixels, which implies that horizontal homogeneous and optically thick water clouds below the cirrus will not affect the solution.

Equation (6) cannot be solved directly, because the factor \( f \) depends on the cloud temperature. It is a transcendental equation that has to be solved iteratively and by using a start cloud temperature. In the first iteration, Eq. (6) gives a cloud temperature that usually differs considerably from the actual cloud temperature. By means of this temperature and the temperature profile of the actual atmosphere above the cirrus that may be
given by radiosonde data, a first iteration cloud height can be determined. Starting at this height, the spectral emission and transmission of the atmosphere from above this height up to the satellite are computed by means of a radiative transfer code that is based on matrix operator theory as described by Plass et al. (1973). The values of emission and transmission are then taken to eliminate the influence of this atmospheric layer on the measured radiances. This correction of the measured radiances leads to an improved cloud temperature and height after the next iteration step because the masking influence of the high atmosphere on the satellite measurement is reduced. This iteration will be performed until the convergence is sufficient.

This iteration technique produces a solution, that is, a cloud temperature, by using channels that have the maximum of the spectral weighting function near the cirrus and near the surface. Presently, HIRS channels on the polar-orbiting satellites of the NOAA series offer the best possibilities for applying this technique because HIRS contains several channels in the absorption bands of CO$_2$ and H$_2$O, as well as in the transparent atmospheric window regions.

From numerical simulations, the optimum channel combinations in the CO$_2$ absorption band were found as channels 4 (14.2 µm) and 7 (13.4 µm) and channels 5 (14.0 µm) and 7. Since water vapor absorption is weak in these channels and its concentration is usually low above the cirrus level, a priori information about the water vapor profile is of less relevance for atmospheric correction. If the humidity profile is known, channels 8 (11.1 µm) and 12 (6.7 µm) can also be used. Since only the profile above the cirrus cloud has to be taken into account for atmospheric correction, uncertainties of 3 K, for example, result in relatively small uncertainties of about 1 K for the cirrus temperature. This corresponds to an uncertainty in cloud height determination on the order of 0.1 km. Compared to an error estimation given by Wielicki and Coakley (1981) for the CO$_2$ slicing technique, the bi-spectral method seems to be less sensitive to errors introduced by uncertain temperature and humidity profiles. The additional error arising from radiometric noise will be discussed in the next section.

The horizontal resolution of the cirrus temperature that can be obtained if this technique is applied to the HIRS channels corresponds to an area of two HIRS pixels. This gives a horizontal resolution of the cirrus temperature of 30–50 km. Inhomogeneities within the pixel of optical depth, temperature, and cloud height are not resolved. A comparison with airborne measurements along a flight path just through one HIRS pixel, as was done on 18 October 1989, has to be interpreted in view of such inhomogeneities (section 6b). The atmospheric correction of the radiances was performed with a radiosonde profile obtained at the same location as the HIRS pixels but with a 1-h delay. However, this delay should not have caused dramatic errors, because no significant changes of the temperature profile were to be expected in the atmosphere above the cirrus.

6. Results and comparison

With the measurements performed during ICE on 18 October 1989 we were able to compare the optical depth and the height of a cirrus cloud derived from satellite data directly with airborne data. The main difficulty in obtaining such data is the simultaneity of the measurements of satellite and aircraft at a suitable cloud. Simultaneity, according to our experience, means a time difference of at least less than half an hour, better less than 10 min. Additionally, radiosonde data and surface-based lidar measurements were available.

a. Optical depth

A validation of the cloud optical depth derived from satellite data with a parameterization scheme embedded in APOLLO is indispensable because both the conversion of the cloud reflectance and the parameterization scheme rely on datasets that are valid only within certain limits.

Figure 4 (AVHRR channel 1 reflectance) shows a section of the German Bight with the flight route of the ALEX drawn in. The flight route crosses a cirrus field from north to south, which seems to be more dense in the north of the islands than in the south. The reflectance along the flight path varies only little, that is, from 8% to 14%. Some contrails that cross the flight path from southwest to northeast show up remarkably bright. The temperature image of the same area (Fig. 5, AVHRR channel 4 temperatures) shows that the temperatures are lower to the north of the islands where the cirrus clouds seem to be brighter in the visible channel, that is, optically thicker. The temperature data along the flight path of about 275 K confirm that over clouds with low optical depths the emitted radiation is determined mainly by the temperature of the underlying sea surface of about 285 K.

The red curve in Fig. 7 shows the total optical depth along the flight path that has been determined from NOAA AVHRR satellite data. The values vary from 0.22 to 0.42. Each point corresponds to a pixel of about 1 km$^2$; the whole curve comprises a 60-km flight path. First, the optical depth increases, corresponding to the visual impression of Fig. 4, with thicker cirrus clouds north of the islands. Then the optical depth drops to about 0.24, where it remains with the exception of two significant maxima that result from contrails.

The lidar measurement was upward from 3.55-km flight altitude. The lidar plot (Fig. 6) showed strong backscattering from 1217 to 1221 UTC (north of the islands), followed by low backscatter with the exception of the striking but small-scale backscatter from three
Fig. 4. AVHRR channel 1 reflectance image of the southern North Sea with the aircraft flight route from 1217 to 1228 UTC as indicated. Satellite overpass time is 1228 UTC. Also indicated are the two HIRS pixels 1903 and 1904.

Fig. 5. Same as Fig. 4 but AVHRR channel 4 temperatures (K).

Fig. 6. Extinction coefficients $\sigma_r$ and meteorological range $V_N$ versus flight position $x$ or time $t$. Part of ICE mission 216, flight altitude is 3.55 km. Lidar shots are averaged over 300 m horizontally and over 97.5 m vertically. The homogeneous layer below 4.5 km indicates air containing even more aerosol below the aircraft.

Fig. 7. Optical depths derived from AVHRR data (red) and lidar data (blue) on 18 October 1989, along the flight path as indicated in Fig. 4 from 1217 to 1227 UTC.
contrails at approximately 1221, 1223, and 1226 UTC. A further contrail at 1219 UTC, which is visible in Fig. 6, is already so weak that its backscatter signal can hardly be discerned in Fig. 7. The lidar plot and the relative course of the satellite-derived optical depth along the flight path agree well. The lower boundary of the thicker cirrus clouds can be well identified. It increases within 27 km from 8.2 to 8.8 km. The geometrical thickness of the cirrus is not so well defined; however, it is at least 600 m.

The optical depth resulting from these backscatter signals is shown as the blue curve in Fig. 7. The optical depth from the earth surface up to the flight altitude (dashed line) is extrapolated from the lidar measurement of the optical depth from the flight altitude to the cloud bottom (dotted line). The resulting optical depth that lies between 0.21 and 0.22 is lower than that estimated from simultaneous surface-based measurements of visibility and from surface-based lidar measurements of the extinction coefficient profile (Ansmann 1991), which is 0.26 ± 0.04. The reason probably is that the aerosol concentration in the boundary layer is in fact higher than that obtained from the downward extrapolation. The lidar-derived curve in Fig. 7 comprises 10 min of flight (60 km). Each point of the curve represents an average over five lidar shots, and hence, corresponds to 0.5-km distance, or twice the resolution of the satellite data. The NOAA satellite overpasses the airborne lidar at 1226 UTC (54-km distance to the beginning of the data curve), where the agreement of both data curves is best.

The contrail at 26 km cannot be identified from the satellite optical depth curve alone because the optical depth previous to the contrail is high as well. The lidar data do not show these enlarged optical depths between 1220 and 1222 UTC. A speculative interpretation is that due to the convergence of the wind field near the coastline the boundary layer is lifted, which results in an increased aerosol content and, hence, optical depth below the aircraft. In spite of this discrepancy, the average of both curves agrees to ±10% rms. This is significantly better than the expected ±30% uncertainty of the satellite retrieval. From the lidar data, it follows that the cirrus clouds have optical depths from 0.01 to 0.1, while contrails cause a local increase from 0.1 to 0.25. The large differences in the optical effects of the contrails are possibly due to either the different shape and size of the scattering particles that are indeed unknown or to a different age of the contrails. Also, different numbers of engines per aircraft may cause different optical depths of the resulting contrails.

b. **Cloud height**

Figure 8 shows the iteration result as obtained from HIRS radiances in channels 5 and 7 for the pixels 1903 and 1904 of a scan along the coast of the North Sea at 1226 UTC 18 October 1989 (Fig. 4). The curve in Fig. 8 is calculated with an atmospheric correction based upon the Norderney radiosonde profile at 1332 UTC. Dashed lines represent temperatures at the lowest cloud base and highest cirrus cloud top, which have been derived from lidar measurements onboard the aircraft during the flight across pixel 1903 from 1220 to 1223 UTC. The cirrus height is related to temperature by means of the radiosonde profile (cf. Fig. 3). As can be seen in Fig. 8, the iteration curve ends inside the maximum temperature range. Assuming the temperature at cirrus top between 230 and 235 K, however, the solution deviates by about 4 K. This discrepancy mainly results from the influence of radiometric noise and to some degree from an uncertain atmospheric correction.

The signal to noise ratio is determined by the difference of cloud optical depth in both satellite pixels. A cirrus cloud appearing different in both satellite pixels gives large signal-to-noise ratios, whereas under more homogeneous conditions the uncertainty increases. Although measurements of the actual microphysical composition of the cirrus are not available, an error estimate based upon realistic assumptions may be given. The uncertainty, for example, reaches ±1.5 K for an optically thin cirrus layer of 1-km depth consisting of particles with \( r = 40 \mu m \) mode radius and different particle concentrations of \( N_1 = 0.01 \text{ cm}^{-3} \) and \( N_2 = 0.05 \text{ cm}^{-3} \) in the two satellite pixels, respectively. The particle concentration \( N = 0.01 \text{ cm}^{-3} \) corresponds to an optical depth of \( \delta = 0.11 \) in the HIRS channels, which is slightly reduced to \( \delta = 0.10 \) at visible wavelengths. A more homogeneous cirrus with \( N_1 = 0.01 \text{ cm}^{-3} \) and \( N_2 = 0.02 \text{ cm}^{-3} \) leads to an uncertainty of about ±5 K; for \( N_1 = 0.1 \text{ cm}^{-3} \) and \( N_2 = 0.2 \text{ cm}^{-3} \), it is reduced to ±0.5 K. Thus, differences between aircraft- and satellite-derived temperatures on the order of ±5 K are explainable for optically thin
cirrus clouds, as observed on 18 October 1989. An additional uncertainty of about 1 K may be caused by the atmospheric correction procedure.

Although the error estimate suffers from the lack of actual microphysical data and insufficient statistics, temperature deviations between the aircraft and satellite measurements can be explained. This case study illustrates that the accuracy of the bispectral method is within acceptable limits, especially for cirrus clouds with an optical depth larger than \( \delta = 0.1 \). Averaging over a number of solutions will reduce the uncertainty caused by radiometric noise for cirrus clouds of extremely small optical depth. Thus, for the determination of cirrus, temperature uncertainties on the order of \( \pm 1-5 \) K should be expected, which usually correspond to an error of \( \pm 0.2-0.8 \) km in height. This compares well with the accuracy of \( \pm 0.6 \) km, which Minnis et al. (1990) estimate for cirrus height determination during FIRE by using a technique based on visible and infrared satellite data.

7. Conclusions

The comparisons of section 6 demonstrate that cirrus optical depth and height derived from satellite measurements compare well with the estimated range of uncertainty. This was shown at rather thin cirrus clouds with optical depths in the range of 0.1. Because the agreement is close to the range of uncertainty of the lidar measurements, the spatial inhomogeneity of the observed cirrus fields seems not to disturb the satellite measurement significantly.

The validation of the satellite-derived data by means of aircraft lidar measurements has been successfully performed with cloud height and, within the accuracy limits of the airborne lidar data, with optical depth. However, the quantity of the data is not sufficient to finally assess and to generalize the quality of the derived optical depths and heights from satellite data. To improve this situation, measurements from the European Lidar Airborne Campaign, conducted by the European Space Agency, are investigated, where the cirrus itself produces the shadow required for calibration because of the downward-looking lidar, and therefore, no assumptions on the aforementioned quotients are necessary. Future campaigns like the European Cloud and Radiation Experiment will be used, too. The same datasets are also appropriate to further establish the quality of the cloud height determination procedure.

The lidar system was calibrated by means of the shadow technique, which relies on the reduced backscatter due to an extinction layer in front of it. The largest uncertainty of the lidar data is the quotient of the lidar ratios of the cirrus and the underlying contrail where the latter has produced the shadow on the cirrus. However, future experiments will allow for increasing the experience with this kind of calibration technique, or avoid the need for such assumptions on quotients of the lidar ratios.

In summary, the comparison between cloud parameters derived from satellite and from airborne measurements shows that the techniques presented provide quantitative information on the physical properties of clouds in a quality that corresponds at least to the estimated range of uncertainty.

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