Variability of sea ice emissivity estimated from airborne passive microwave measurements during FIRE SHEBA

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Abstract. Passive microwave radiometers with frequencies ranging from 37 GHz to 220 GHz were flown over the Surface Heat Budget of the Arctic (SHEBA) experimental site in May and July 1998. These measurements were motivated by the possibility of determining cloud liquid water path, ice water path, and precipitation over sea ice from these frequencies. The comprehensive cloud data set collected in this experiment offers a unique opportunity for improving and adapting passive microwave retrieval methods for application to arctic clouds. However, retrieval of cloud properties from a downward looking radiometer requires an estimate of the surface emissivity and its spectral, spatial, and temporal variation. In this study, brightness temperature measurements are used to calculate sea ice emissivity at each frequency using ancillary aircraft data to characterize the atmosphere and obtain surface temperature. Surface emissivities on clear sky days during the FIRE Arctic Cloud Experiment (FIRE ACE) aircraft campaign have been calculated and compared with previous estimates cited in the literature. Average emissivity at nadir for snow-covered sea ice in the experimental region during late May is estimated as 0.89 at 37 GHz, 0.74 at 89 GHz, 0.72 at 90 GHz, 0.73 at 150 GHz, and 0.84 at 220 GHz. In early July the average nadir emissivity for melting sea ice is 0.86 for 37 GHz and 0.84 for 90 GHz. Estimates of emissivity at 50° off nadir are compared with previous satellite and ground-based measurements of dual polarized emissivities at 37 GHz and 90 GHz. Significant variability exists in published emissivity values due to variations in dielectric and physical properties of snow and ice, but our results fall within previously observed ranges. Uncertainties in the emissivity calculations are estimated, and the accuracy required for use of surface emissivity estimates in cloud retrieval methods is discussed.

1. Introduction

Passive microwave measurements of sea ice have been available since the late 1970s and have been used extensively for monitoring various surface properties, including ice type, concentration, and extent. Limited efforts have been applied to microwave retrieval of cloud properties over sea ice for various reasons. First, the high emissivity and heterogeneity of the ice surface at microwave frequencies make separation of the atmospheric contribution more difficult than over a radiatively cold surface such as the ocean. Second, it has been widely assumed that in the absence of precipitation the atmosphere is effectively transparent for the commonly used frequencies (e.g., the 19 and 37 GHz channels of Special Sensor Microwave Imager (SSM/I)) when performing sea ice retrievals. Third, estimates of the surface emissivity are not produced routinely but are necessary for retrieving atmospheric properties from microwave measurements. Given a method for estimating the surface contribution, we can potentially develop retrieval algorithms for deriving cloud liquid and ice water path as well as rainfall and snowfall for application over sea ice.

Methods for estimating cloud liquid water path and rain rate rely on brightness temperature increases due to emission by cloud and rain droplets at SSM/I frequencies [Liu and Curry, 1993; Petty, 1994]. Although quantitative retrievals have been performed primarily over the ocean surface, clouds over sea ice are clearly visible in consecutive scenes of SSM/I imagery. Positive anomalies in 37 and 90 GHz brightness temperatures in the Arctic have been partially attributed to emission from liquid water clouds (G. Liu and J.A. Curry, Observations of microwave "hot spots" over the Arctic Ocean during winter, submitted to Journal of Geophysical Research, 2000). Cloud ice water path can be related to scattering-induced brightness temperature depressions at 150 and 220 GHz over the ocean [Deeter and Evans, 2000; Liu and Curry, 2000]. Wang et al. [this issue] estimate ice water path over sea ice from data at 183.3 ±7 and 340 GHz, but their retrievals are limited to high viewing angles where surface contributions are negligible.

Various researchers have addressed the issue of high background emissivity in the microwave portion of the spectrum with studies over land. Table 1 summarizes their efforts to estimate emissivity for a variety of surfaces. Land surface emissivity at SSM/I frequencies has been derived by Jones and Vonderhaar [1997] and Prigent et al. [1997]. Greenwald et al. [1997] went on to develop a method for measuring cloud liquid water path over land using the Jones and Vonderhaar emissivity estimates. Felde and Pickle [1995] retrieved emissivity at SSM/T2 frequencies via similar methods over various surfaces, including snow, bare soil, and vegetation. Estimates of microwave emissivity for snow-covered land and a variety of ice types in the Baltic Sea were made by Hewison and English [1999] using airborne radiometer data.

The variability in arctic sea ice and snow emissivity is well documented, and numerous researchers have related microwave signatures to physical properties of the medium.
emissivity results represent all seasons and ice types with most measurements at frequencies of 90 GHz and below. Large variations in emissivity with time, space, surface, and meteorological conditions have been documented throughout the past two decades [Comiso, 1983; Grenfell and Lohanick, 1985; Grenfell, 1992]. Comiso [1983, 1986] calculated emissivity from SMMR measurements over an annual cycle in large regions of the Arctic. Largest variations at this coarse resolution are observed during the onset of summer melt. Surface-based observations of emissivities reveal significant variability on a small spatial scale and provide a means for interpreting the variations in terms of observed physical properties of the surface. For example, emissivities during summer in the marginal ice zone (MIZ) are examined in conjunction with in situ measurements of snow and ice properties by Onstott et al. [1987]. Grenfell [1992] provides an analysis of regional variations in arctic multiyear ice signatures as related to seasonal changes in surface properties showing that even on a small scale (<100 m) there is significant inhomogeneity in the microwave signature.

The interaction of microwave radiation with snow cover has been specifically addressed by other researchers. Matzler [1987] summarizes the relationship of scattering to snow grain size and of absorption to liquid water content for natural snow covers. Barber et al. [1998] investigate the effects of snow cover on sea ice, demonstrating with laboratory data that for dry snow, snow grain size and volume scattering affect emission above 37 GHz.

Relatively few measurements of snow and sea ice have been conducted at high frequencies (i.e., above 90 GHz). As noted in Table 1, Hollinger et al. [1984] estimated emissivity for various ice types at 140 GHz during the autumn freeze-up period. Felde and Pickle [1995] and Bauer and Grody [1994] calculate the emissivity of snow-covered ground at SSM/T2 frequencies, but without coincident observations of snow physical characteristics, it is difficult to apply their results to other regions. Tait et al. [1999] discuss brightness temperature measurements at 89, 150, and 220 GHz over snow-covered ground but have not made quantitative estimates of surface emissivity. Hewison and English [1999] provide the most comprehensive report to date of variations in emissivity at high frequencies for snow and certain ice types.

Results of previous research demonstrate that it is difficult to characterize emissivity at a given frequency and polarization based simply on a knowledge of ice type, season, and/or location due to the high variability on a local scale. Since the surface contribution to the upwelling radiance is larger than the atmospheric contribution, even for a cloudy atmosphere, a realistic specification of the local surface emissivity is necessary for cloud retrievals. For example, emission by a liquid cloud may add ~10-20 K to the 90 GHz brightness temperature. Neglecting this contribution leads to a 5-10% error in surface emissivity. Therefore accurate estimates of surface emissivities derived for SHEBA are prerequisite for the development of cloud retrieval algorithms.

In this study we derive microwave surface emissivity from airborne radiometer data collected during the FIRE Arctic Clouds Experiment (FIRE ACE) and Surface Heat Budget of the Arctic (SHEBA) aircraft observing period [Curry et al., 2000; Perovich et al., 1999a]. During May, June, and July 1998, two aircraft carrying microwave radiometers overflew the SHEBA ice camp. Using these passive microwave measurements along with ancillary aircraft data, we are able to examine the variations in emissivity prior to the onset of melting when snow cover is still present and again during the summer melt season. The airborne sensors allow us to characterize the variability in emissivity on a much finer scale and over a wider range of frequencies than in previous satellite-based studies. Overlapping frequencies between the two sensors provide data for intercomparison. In situ data from the SHEBA ship are useful for relating emissivity values to observed surface conditions and for understanding differences in emissivity at varying frequencies.

2. Methodology

Surface emissivity can be derived from the equation of radiative transfer for microwave frequencies as demonstrated by Prigent et al. [1997]. Assuming a non-scattering atmosphere, the equation for the upwelling brightness temperature at a given frequency and polarization at altitude H above the surface is

\[ T_b = \varepsilon \sigma T_s \left( -\frac{\tau(0,H)}{\mu} + (1 - \varepsilon) \exp \left( -\frac{-\tau(0,H)}{\mu} \right) \int_0^H \theta(z) \kappa(z) \exp \left( -\frac{-\tau(z,H)}{\mu} \right) dz' \right) \]

\[ + \int_0^H \theta(z) \exp \left( -\frac{-\tau(z,H)}{\mu} \right) dz' \]

where \( T_s \) is brightness temperature measured by a radiometer, \( \varepsilon \) is surface emissivity, \( T_s \) is the physical temperature of the surface emitting layer, \( \tau \) is atmospheric opacity, \( H \) is height of the radiometer above the surface, \( \theta(z) \) is the atmospheric temperature as a function of height, \( \kappa(z) \) is the atmospheric absorption coefficient, \( \mu \) is the cosine of the viewing angle, and \( dz' = dz/\mu \) [Janssen, 1993]. The first term on the right-hand side represents surface emission, the second term is downwelling atmospheric emission which is reflected by the surface, and the third term is upwelling atmospheric emission.

Rearranging the above relationship to solve for surface emissivity, we obtain

\[ \varepsilon = \frac{T_b - T_{up} - \Gamma T_{down}}{\Gamma(T_s - T_{down})} \]  

(1)
where the following notation has been introduced:

\[ \Gamma = \exp \left( -\frac{\pi(0,H)}{\mu} \right) \]  

\[ T_{ap} = \frac{H}{\pi} \int_0^H T(z) \kappa(z) \exp \left( -\frac{\pi(z,H)}{\mu} \right) dz' \]  

\[ T_{down} = \frac{H}{\pi} \int_0^H T(z) \kappa(z) \exp \left( -\frac{\pi(0,z)}{\mu} \right) dz' \]  

Emissivity is determined by applying (1) through (4) at a given frequency and polarization using observations obtained from aircraft- and surface-based measurements. Each of these data sources is described in the following section, along with assumptions made in applying them for emissivity calculations.

3. Data

Aircraft measurements in the vicinity of the SHEBA ice camp were conducted in the spring and summer of 1998. Of particular interest for this work are flights by the National Center for Atmospheric Research (NCAR) C-130 aircraft and the NASA ER-2 aircraft, since both carried microwave radiometers. The NCAR C-130 performed eight research flights in May and eight in July of that year. The NASA ER-2 flew 11 missions from mid-May through early June. Review of the meteorological observations on each flight day revealed cloud-free conditions over the ice camp on May 20, May 24, and July 8. Both aircraft were operating in the region on May 20; the NCAR C-130 also flew on May 24 and July 8.

3.1. Sensor Descriptions

The Airborne Imaging Microwave Radiometer (AIMR) is a cross-track scanning system which flies on the NCAR C-130. Four channels measure upwelling radiation at two frequencies, 37 and 90 GHz, and two orthogonal polarizations which can be converted to horizontal and vertical components. The Airborne Imaging Microwave Radiometer views underlying scenes over an angular swath of 120°. Beam widths of 1° at 90 GHz and 2.8° at 37 GHz produce spatial resolutions of the order of 20 m to 300 m at typical flight altitudes and velocities. Corresponding swath widths are approximately 3 to 20 km. Detailed specifications for the AIMR can be found in the work of Collins et al. [1996].

Calibration of the AIMR is performed internally as the scanning mirror views two calibration loads after viewing the underlying scene. One load is heated to approximately 350 K, while the other is designed to float at ambient temperature, so two reference points are obtained relating voltage to radiance. Scene measurements are converted to radiance by interpolating between the reference points. Errors resulting from calibration uncertainties and polarization conversion are given by Collins et al. [1996]. Brightness temperature errors are estimated as less than 1 K for angles greater than 20°; at smaller angles the errors may increase to 1.5 K or more due to the polarization conversion technique. At angles near nadir, brightness temperature values are taken from the average of each orthogonal channel rather than from the polarization conversion algorithm to eliminate the increasing errors in the polarization retrieval. Error estimates for each of the channels are of the order of 0.5 K. It should be noted that these error estimates assume scene brightness temperatures fall between the hot and the cold load temperatures, which was not always the case in the FIRE SHEBA experiment. Because of the current mounting configuration on the NCAR C-130, the AIMR housing is actually considerably warmer than ambient temperature, causing the cold load to be significantly warmer than specified in the design of the system [Walther et al., 1999]. Typical cold load temperatures during SHEBA were 280 to 285 K (compared to ambient temperatures as low as 213 K at the highest flight altitudes). The uncertainties under these conditions have not been quantified.

The Millimeter Imaging Radiometer (MIR) was flown on the NASA ER-2 during FIRE SHEBA. The MIR is a cross-track scanner measuring microwave radiation at several frequencies, including 89, 150, and 220 GHz. It scans over an angular swath of 100° and has a 3.5° beam width. Resulting swath width at the typical 20 km flight altitude of the ER-2 is approximately 48 km. The MIR calibration system is similar to that of AIMR with two internal loads viewed during each scan. Racette et al. [1996] give error estimates of better than 1 K for brightness temperatures between 240 and 300 K. Postflight calibration efforts have demonstrated errors of the order of 2 to 4 K for brightness temperatures below 100 K (lowest scene temperatures observed in SHEBA are about 200 K at 89 GHz). Separation of the vertically and horizontally polarized components is not possible with this instrument, so measurements at nadir are used for calculations of emissivity and comparison with AIMR estimates of emissivity. Specifications for both AIMR and MIR are summarized in Table 2.

A set of Heimann model KT19.85 pyrometers was flown on the NCAR C-130 for the purpose of remotely sensing surface (skin) temperature. These sensors measure radiation within the spectral range 9.6 to 11.5 μm over a 2° field of view. Absolute accuracy is ±0.5°C plus 0.7% of the difference between the sensor housing temperature and scene temperature. The accuracy of the measurement is also affected by absorption and

<table>
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<th>Table 2. Microwave Radiometer Specifications</th>
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<td><strong>AIMR</strong></td>
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<td>Radiometer type</td>
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<tr>
<td>Platform</td>
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<td>Altitude</td>
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* MIR also has channels at 183.3±1, 183.3±2, 183.3±7, and 340 GHz that are not used in this study.
emission of infrared radiation from atmospheric water vapor in the layer between the surface and the aircraft. Additional errors in surface temperature are introduced by assuming the surface is a blackbody. In reality, the spectral emissivity of ice and snow is slightly smaller than unity, resulting in some reflection of downwelling infrared radiation. Corrections for these effects are discussed in section 4.

3.2. Data Analysis

Equations (1) through (4) require input from several sources in order to estimate $e$, at a given frequency. Brightness temperatures $T_b$ at 37 and 90 GHz are obtained from the AIRM. The MIR provides $T_b$ at 89, 150, and 220 GHz. Straight and level flight segments over the SHEBA ice camp yield two-dimensional images of $T_b$ for analysis. Flight patterns for the C-130 tended to focus within 50 km of the ship. The ER-2 covered a wider region but also overflaw the SHEBA site. Figure 1 shows the portion of the C-130 and ER-2 flight tracks used for this study on May 20, 1998. On each C-130 flight, a series of 50 km segments oriented east-west and centered on the SHEBA ship are used for emissivity calculations. The ER-2 flight on May 20 included one pass over the SHEBA site, oriented approximately north-south, passing over the C-130 flight pattern.

Vertical profiles of temperature and humidity are needed to calculate the atmospheric upwelling and downwelling terms, $T_w$ and $T_{air}$. Soundings were routinely made by the C-130 at the completion of its flight pattern over the ice camp from the surface to about 6 km. Since the C-130 profiles do not extend up to the altitude of the ER-2 (20 km), a radiosonde observation from the SHEBA camp was used for that case. These profiles also serve as input to the model used for calculation of atmospheric absorption coefficient $\kappa$ at microwave frequencies following Liu [1998].

A measure of the physical temperature of the surface emitting layer is required for the emissivity calculation. Measurements from the KT19 infrared thermometer provide an estimate of skin temperature; however, radiation at microwave frequencies emanates from a layer of finite depth depending on frequency. Penetration depth at a given wavelength $\lambda$ can be calculated according to:

$$\delta_d = \frac{0.5}{\alpha}$$

(5)

where

$$\alpha = \frac{2\pi}{\lambda} \ln \sqrt{\varepsilon}.$$  

The dielectric constant $\varepsilon$ is calculated following parameterizations for various types of materials given by Ulaby et al. [1986]. Emitting layers in snow and sea ice may be several centimeters deep or more, so we need an effective temperature for the layer rather than just a skin temperature. Since this measurement is not available over the horizontal scale of the passive microwave measurements, we improvise by trying to relate skin temperature measurements to emitting layer temperature. Perovich et al. [1997] logged an annual cycle of sea ice temperature profiles in a multiyear ice floe in the Beaufort Sea. Their results show a nearly isothermal profile of sea ice temperatures to well below microwave penetration depths for the late May and early July periods of interest for this study. Similar profiles near the SHEBA site were measured in 1998 [Perovich et al., 1999b], so it seems reasonable to assume that the skin temperature is a good approximation for the temperature of the emitting layer at this time of year. The assumption is probably less valid when snow covers the sea ice, since the temperature profile through the snow layer cannot be assumed isothermal. A snow layer of about 35 cm was present in late May 1998 [Perovich et al., 1999b], but the differences between the atmospheric temperature and the ice temperature were not large, so the temperature gradient through the snow layer should be relatively small. Vertical profiles of the snow layer near the SHEBA site [Perovich et al., 1999b] show temperatures in late May of approximately 265 K with only a slight increase between the snow-ice interface and the snow-atmosphere interface. On the dates of interest here, the temperature difference appears to be about 1 K. Thus the error involved in using skin temperature to represent emitting layer temperature in these cases should be 1 K or less. We note that the vertical temperature gradient in the snow layer is quite variable, so this assumption is probably not valid in other seasons and under different meteorological conditions.

Errors in skin temperature measurements due to emission and absorption of infrared radiation by the atmosphere can be estimated and removed from the data prior to our emissivity calculations. The brightness temperature observed by a KT19 sensor is composed of radiance from (1) surface emission attenuated by the atmosphere, (2) surface reflectance of atmospheric emission attenuated by the atmosphere, and (3) emission from the atmosphere itself. Skin temperature is routinely derived from KT19 measurements by simply assuming it is equivalent to brightness temperature. This assumption is true only if the surface is a perfect emitter and the atmosphere does not absorb or emit any radiation. The true skin temperature can be derived by accounting for atmospheric effects and surface reflectance. Such corrections have been previously applied to infrared temperature measurements of sea ice by Steffen and Lewis [1988] and Muller et al. [1975]. Following their methods, we use a radiative transfer model [Key and Schweiger, 1998] to calculate correction factors for each of the days analyzed here. Temperature and humidity profiles from each of the aircraft were input into the

![Figure 1](https://example.com/image.png)  
**Figure 1.** Flight tracks of the NASA ER-2 and NCAR C-130 aircraft in the vicinity of the SHEBA ship on May 20, 1998.
model, and runs were performed over a range of hypothetical surface temperatures. The differences between ”top-of-atmosphere” infrared brightness temperature calculated by the model and the input surface temperature represent the measurement error. Figure 2 gives temperature errors as a function of KT19 brightness temperatures for typical flight altitudes.

The effect of varying infrared emissivity was also considered with these model runs. Infrared emissivities for snow and ice are discussed by Salisbury et al. [1994]. Over the spectral range of the KT19, they report dry snow emissivities ranging from 0.997 to 0.999. For ice (distilled water) they give values ranging from 0.97 to 0.995 for this spectral range. The actual emissivity depends on many factors (e.g., snow grain size, extent of melting) which cannot be characterized from the available data. Infrared surface temperature measurements were also made from towers at the SHEBA site; researchers report using an emissivity of 0.99 for those measurements [Claffey et al. 1999]. For consistency, we assume the same and perform model runs to estimate the errors induced by assuming an emissivity of unity versus 0.99. For the atmospheric conditions on the three days considered here, the effect on surface temperature of changing the emissivity to 0.99 is of the order of 0.2 to 0.3 K. We consider this amount to be within the error of the instrument and thus have not corrected the surface temperature data for this effect.

In calculating emissivity for the two-dimensional images produced by the scanning radiometer, surface temperatures were not available for each pixel since KT19 is a nadir-viewing sensor. In the case of the AIMR it is assumed that the nadir surface temperature at a given point along the flight track represents the surface temperature for each pixel in the cross-track scan, provided the pixel does not contain a lead. Statistics compiled from the C-130 flight tracks in the SHEBA region show little variation in surface temperature (standard deviations of the order of 0.2-0.3 K) on a given day at the resolution of the KT19, so this assumption seems reasonable. Lead surface temperature statistics were also compiled. Again, small variations are observed over the region, so a constant lead temperature is assumed for each day analyzed. For the MIR data set, coincident surface temperature measurements were not available. In addition, the resolution of the MIR does not allow for differentiation of most leads. Thus an average surface temperature derived from KT19 measurements over a 50 x 50 km square centered on the SHEBA ship is applied to a MIR image covering a similar area.

3.3. Error Analysis

Uncertainties in our calculation of $\varepsilon$, arise from errors in the measurement of $T_s$ and $T_a$, from the assumption that the microwave emitting layer temperature is equivalent to skin temperature, from atmospheric corrections applied to $T_a$, and from numerical estimates of the atmospheric contribution to the microwave signal.

Assuming $\varepsilon$, values of the order of 0.75, the $T_s$ error ranges given in section 3.1 translate into $\varepsilon$, error bars of $\pm 0.005$ for MIR and $\pm 0.01$ for AIMR. It is possible that $T_a$ errors for one or both sensors are larger than published, since the brightness temperatures in this environment tend to fall below the range of the calibration load temperatures for both sensors. The 1 K estimated error for MIR applies to the temperature range 240 - 300 K, whereas brightness temperatures measured in SHEBA were of the order of 200 K. (Error estimates for MIR brightness

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**Figure 2.** Surface temperature corrections as a function of flight altitude and surface temperature as measured by the KT19 radiometer: (a) corrections for May 20, 1998; (b) May 24, 1998; (c) July 8, 1998.
temperature of 100 K or lower are 2 - 4 K, according to Racette et al. (1996)). Calibration errors are possibly larger than estimated for the AIMR since scene brightness temperatures are nearly 100 K below the temperature of the cold load, but quantitative estimates of error bars under these conditions are not yet available (C. Walther, personal communication, 2000).

Uncertainties in $T_e$ arise from instrument errors, from our atmospheric correction procedure, and from our assumption that the IR surface emissivity is unity. Sensitivity analyses with a radiative transfer model have shown that the uncertainty introduced by the surface emissivity assumption is of the order of 0.2-0.3 K. Instrument errors are approximately 0.5-0.7 K. The atmospheric correction terms applied to $T_e$ measurements are typically 0.5 K or less, and uncertainties arising from model approximations of the correction terms are smaller than instrument errors. Uncertainty is also introduced by assuming an isothermal snow layer when there may be a slight temperature gradient over the microwave emitting layer. As noted in section 3.2, the maximum temperature difference across the snow layer is 1 K, so the isothermal assumption potentially introduces the largest contribution to the total error. If we assume a 1 K error in the specification of $T_s$, the corresponding uncertainty in $\varepsilon$ is 0.003 when $\varepsilon$ is of the order of 0.75.

Finally, the numerical estimate of atmospheric contributions to the brightness temperature contains errors due to errors in input data, uncertainty in the calculation of the absorption coefficients, and approximation of the integral terms. We have not quantified these errors but note that the integrations specified in (2)-(4) are performed over a 4 km layer for AIMR calculations and over a 20 km layer for MIR calculations. Thus the errors introduced may be considerably different for each data set. For the data we analyzed, the atmospheric terms in (1) are significantly smaller than $T_e$ and $T_s$, so we expect that most of the uncertainty arises from errors in $T_e$ and $T_s$.

4. Results

Surface emissivity at AIMR and MIR frequencies has been calculated over an area centered on the SHEBA site as shown in Figure 1. Clear-sky days are required for these calculations; flight days satisfying this requirement were May 20, May 24, and July 8 for the C-130 and May 20 for the ER-2. In late May the multiyear ice floe on which the ship was located was covered by a layer of dry snow about 35 cm deep. Corrected surface temperature measured from the aircraft and averaged over a 50 x 50 km region centered on the ship was 263.9 K. Measurements from sensors mounted on towers near the SHEBA ship, as described by Caffrey et al. [1999], give temperatures ranging from 263.6 K to 264.4 K at the approximate time of the C-130 overflight.

Melting began near the end of May, so the snow layer is assumed to consist of dry snow on May 20 and 24. Numerous leads were present in the vicinity of the ship, but they were largely frozen with almost no open water observed from the aircraft. Conditions had changed drastically by early July when the second series of flights began. No snow was present on the ice during the July 8 flight. There were numerous open leads, and meltponds had formed on the ice. Corrected surface temperatures, averaged over the 50 x 50 km area, were 272.6 K compared with surface-based measurements ranging from 272.8 to 273.5 K.

Results given in section 4.1 use $\varepsilon$, at nadir since estimates from both AIMR and MIR are discussed (recall that MIR does not separate horizontal and vertical components). In section 4.2 we discuss angular variations in $\varepsilon$, and show polarization differences from AIMR data.

4.1. Spatial, Temporal, and Spectral Variations

Variations in 37 and 90 GHz emissivities at nadir along the C-130 flight track on May 20 are plotted in Figure 3. Spatial resolution of the AIMR at this altitude (4000 m) is about 70 m for the 90 GHz channels and 200 m for 37 GHz, so leads of at least those widths are resolved in these measurements. The inhomogeneity of the surface is apparent in the range of $\varepsilon$, values that occur over small distances. At 37 GHz, $\varepsilon$, at nadir ranges from 0.8 to 0.97, while at 90 GHz, the values span a larger range from 0.65 to 0.95. Most of the variation at 90 GHz is attributable to leads; the standard deviation of $\varepsilon$ exclusive of leads is of the order of 0.035.

The spectral difference in $\varepsilon$, may be explained by considering the observed snow layer covering the ice together with microwave penetration depths. Table 3 gives penetration depths for differing materials as defined in (5). Estimated penetration depths for dry snow assume an average snow density of 0.32 g cm$^{-3}$ (M. Sturm, personal communication, 2000). Penetration depths for multiyear ice (assuming a salinity of 0.7 parts per thousand (ppt) and density of 0.77 g cm$^{-3}$) and for first-year ice (assuming a salinity of 4 ppt and density of 0.88 g cm$^{-3}$) are shown for comparison. It can be seen from the first line in Table 3 that the snow layer has little effect on emission at 37 GHz, since penetration depth for dry snow exceeds the measured depth of 35 cm. Thus we conclude that the emitted surface radiation at 37 GHz is coming primarily from the sea ice beneath the snow, though we would expect some loss due to scattering in the snow layer (the transmissivity of the observed snow layer is 74% at 37 GHz). At 90 GHz and above, the theoretical penetration depths suggest that the snow layer has a much larger effect (in fact, Greenfell [1992] showed that a 20 cm layer of fine-grained winter snow is actually opaque at 90 GHz).

![Figure 3. Sea ice emissivity at nadir along a straight and level flight track as derived from AIMR measurements on May 20, 1998. Solid line is 37 GHz emissivity; dashed line is 90 GHz emissivity.](image-url)
Table 3. Microwave Penetration Depths in centimeters at -10°C

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<th>37 GHz</th>
<th>89-90 GHz</th>
<th>150 GHz</th>
<th>220 GHz</th>
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<tr>
<td>Dry snow</td>
<td>120</td>
<td>28</td>
<td>13</td>
<td>7</td>
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<tr>
<td>Multiyear ice</td>
<td>7</td>
<td>4</td>
<td>2</td>
<td>1.5</td>
</tr>
<tr>
<td>First-year ice</td>
<td>1.4</td>
<td>0.94</td>
<td>0.75</td>
<td>0.61</td>
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The enhanced attenuation is due to volume scattering by the snow particles which becomes more significant with increasing frequency, as demonstrated by Hallikainen et al. [1987]. They also observed increasing losses at 90 GHz due to surface scattering in cases where the surface snow layer is hard. Snow column observations from the SHEBA site (M. Sturm, personal communication, 2000) describe the surface snow layer as a wind slab of moderate hardness and density. Thus surface scattering losses may add to the extinction at 90 GHz. Similar spectral gradients from a surface covered by dry snow have been noted by Mastler [1987] and Onstott et al. [1987].

Signatures from new ice within leads are apparent in the trace, especially at 90 GHz, where \( \varepsilon \) values briefly jump to 0.8 and above as would be expected for first year ice (FYI). The leads are more easily detected at 90 GHz than at 37 GHz for two reasons. First, the spatial resolution of the AIMR is higher at 90 GHz, so smaller leads are not resolved in the 37 GHz measurements. New ice in leads is apparent at 37 GHz, although the difference between \( \varepsilon \) for leads versus that for MYI is smaller than at 90 GHz. The reason for the large difference at 90 GHz may be that the snow layer on leads is thinner than on MYI. If the snow layer on a lead were less than the penetration depth at 90 GHz, then emission from the new ice below would contribute to the signal. No observations of snow cover on the new ice in leads are available, but it seems plausible that there would be less snow on leads due to heating from below the thinner layer of ice, which enhances sublimation [Schramm et al., 1997], and perhaps due to wind blowing snow off the smooth ice.

By July the ice surface has changed dramatically since melting has occurred and the snow cover is gone. A similar flight pattern consisting of 50 km segments centered on the ship was flown by the C-130 on July 8. Variability in \( \varepsilon \), along one flight segment is shown in Figure 4. In contrast to the measurements in May, \( \varepsilon \) values at the two frequencies are now quite similar. At surface temperatures near 0°C, significant amounts of liquid water are observed on the surface. Under these conditions, penetration depths are comparable to the wavelength (less than 1 cm) and radiative properties are determined by the shallow top layer [Onstott et al. 1987]. Thus each channel is viewing the same dielectric material in this case. In general, the MYI \( \varepsilon \) values are more homogeneous than during May, ranging from 0.85 to 0.93 at nadir. Open water in the leads is radiometrically cold with emissivities as low as 0.65. At both frequencies the standard deviation exclusive of leads is smaller (~0.025) than in May.

Figure 5 shows the distribution of \( \varepsilon \) at nadir for the May 20 and July 8 cases. (Results on May 24 are similar to those on May 20). The 37 GHz distributions for May show a broad range of emissivity values. The high end of the distribution includes values for leads, but we also see similarly high values in other areas, so there is not a distinct separation between leads and the surrounding ice. Single peaks centered at about 0.73 characterize the 90 GHz distributions on May 20 and May 24. A smaller number of pixels are spread over the 0.8 to 0.95 range. These represent leads detected at this frequency. The July plots (Figure 5b) show the similarity of the 37 and 90 GHz distributions at this time of year. In this case, the lead pixels are those with \( \varepsilon \) values below 0.8.

Distributions of \( \varepsilon \) at nadir for MIR channels on May 20 are shown in Figure 6. The range of values observed is smaller than that seen by AIMR; this is probably a result of the larger footprint of MIR and its inability to resolve individual leads. Mean values of \( \varepsilon \) at nadir are 0.74, 0.73, and 0.84 for the 89 GHz, 150 GHz, and 220 GHz channels, respectively. Few measurements of snow and sea ice at the higher frequencies have been published, so it is difficult to find emissivity values in similar conditions for comparison. Hollinger et al. [1984] estimated emissivity for multiyear sea ice during autumn as 0.67 and 0.68 at 90 and 140 GHz, respectively. Emissivity values for snow-covered ground from SSM/T2 (91 and 150 GHz) have been reported by Felde and Pickle [1995] who calculated values of 0.86 and 0.85, respectively. Bauer and Grody [1994] give values for snow-covered ground at the same frequencies as 0.77 and 0.76. Without knowing the snow characteristics in either study, it is not meaningful to compare our absolute values with theirs. We note, however, that the spectral gradients for the SSM/T2 retrievals over snow are negative and of similar magnitude as in our snow-covered cases. Hewison and English [1999] calculated emissivity at 91 and 157 GHz for a variety of snow and ice types. They report a positive gradient in three of their cases (lake ice plus snow, deep dry snow, and fast ice). Tait et al. [1999] analyze MIR measurements at 89, 150, and 220 GHz over snow-covered ground. They note higher brightness temperatures at 220 GHz.
efficiency approaches an asymptotic value as the size parameter becomes greater than 1. Here \( r \) is the radius of the scattering particles and \( \lambda \) is wavelength. Thus while emission is increasing throughout the spectral range, scattering increases with frequency only until the size parameter reaches the Mie-scattering range.

Estimates of \( \chi \) can be derived from in situ snow cover observations at SHEBA. Wavelengths corresponding to AIMR and MIR frequencies range from 1.4 mm (220 GHz) to 8.1 mm (37 GHz). Measurements of the snowpack vertical structure revealed eight layers with distinct properties (M. Sturm and J. Holmgren, personal communication, 2000). Estimated grain sizes of snow within those layers range from 0.6 mm to 6 mm in diameter. The layers comprising the upper 15 cm of the snowpack, which interact most strongly with higher frequencies, contain grain sizes of 0.8 mm to 3 mm in diameter (0.4 to 1.5 mm radii). Considering those dimensions together with the wavelengths, resulting size parameters are of the order of 1.2 - 4.7 at 150 GHz and 4.8 - 6.7 at 220 GHz. Comparing these estimates with scattering efficiency curves given by Ulaby et al. [1981, 1986], we surmise that there is no longer an increase in scattering with frequency above about 150 GHz. The continued increase in emission with frequency may then explain the increase in emissivity between 150 and 220 GHz seen in Figure 7.

The MIR and AIMR share a nearly common frequency (89 and 90 GHz, respectively), so it is useful to compare the emissivity calculations from both sensors. Because the aircraft did not fly exactly coincident flight paths and were separated in time by a few hours, we have not attempted to make a pixel-by-pixel comparison. We can compare the results in a statistical sense however, since both aircraft flew in the vicinity of the SHEBA site on May 20. The mean \( \epsilon_S \) at nadir in the 50 x 50 km box surrounding the ship is 0.72, as calculated from AIMR measurements at 90 GHz. Along the 150 km linear track where the ER-2 passed over the ship, the mean \( \epsilon_S \) from MIR at 89 GHz is 0.74. As shown in section 3.3, most of this difference can be explained by uncertainties in the measurement of \( T_s \) and \( T_R \).

**Figure 5.** Distributions of sea ice emissivity at nadir for AIMR frequencies along a C-130 flight segment: (a) May 20, 1998; (b) July 8, 1998.

compared to the other frequencies but have not separated the surface and atmospheric contributions.

The spectral variation between 37 and 220 GHz, as shown in Figure 7, may be explained by considering the relative importance of scattering and emission processes at each frequency. Emission by snow increases with frequency, as governed by the Planck function, thereby raising the emissivity. However, extinction by snow at these frequencies is typically dominated by scattering processes [Hallikainen et al., 1987] which act to decrease the emissivity. According to Mie theory as summarized by Ulaby et al. [1981], the scattering efficiency factor increases, while the size parameter,

\[ \chi = \frac{2\pi r}{\lambda}, \]

remains in the Rayleigh-scattering range, but the scattering ef-

**Figure 6.** Distributions of sea ice emissivity at nadir for MIR frequencies along the ER-2 flight track on May 20, 1998.
4.2. Viewing Angle and Polarization Variations

AIMR images of emissivity on May 20, 1998 (Plate 1) show
the two-dimensional variation in $\varepsilon$ over an area of about 17 by
50 km. The flight track is along the vertical axis of these im-
ages which correspond to the east-west segment shown passing
over the ship in Figure 1. As noted when comparing the
nadir traces, $\varepsilon_0$ at 37 GHz and $\varepsilon_0$ at 90 GHz are substantially
different, due to the snow cover and differing penetration depths.
(Note that these images are not horizontally or vertically pola-
rized components of $\varepsilon$, but instead are averages of the raw
channels at each frequency). FYI within leads is easily detected
in the 90 GHz image as a high-emissivity feature while the
snow cover over the surrounding ice pack masks any surface de-
tail at this frequency. Many of the same leads visible at 90 GHz
are apparent in the 37 GHz image, but substantial variation in
$\varepsilon_0$ over the ice pack is also noted. Given that ground-truth data
covering the entire region are not available, we cannot be cer-
tain about the reason for such large variations. One possibility
is a difference in salinity (brine content) of the sea ice. For ex-
ample, the high emissivity region in the bottom left of the 37
GHz image (orange shades indicating emissivities of 0.9-0.95)
may consist of FYI, while the surrounding areas (green and yel-
low shades) are MYI. A more shallow snow layer resulting in
less volume scattering might also contribute to a higher emis-
sivity, although that explanation seems less likely, given the
sharp boundaries observed. Furthermore, emissivity values of
0.9 and greater are characteristic of FYI according to Eppler et
al. [1992].

Estimates of $\varepsilon_0$, given in the literature, are generally at view-
ing angles 50° off nadir to correspond with satellite microwave
radiometers. Horizontally and vertically polarized emissivities
at this angle have been averaged along a 50 km flight segment
for comparison with previous estimates at similar frequencies
and times of year, as described in Table 1.

Our results for late May are given in Table 4. AIMR esti-
mates at 37 GHz for both polarizations in May are signifi-
cantly higher than the MYI averages compiled by Eppler et al.
but fall within the range given by Comiso. As noted above, our
averages may have included substantial areas of FYI; if that is
the case, we would expect our estimates to be higher than the
Eppler MYI values. At 90 GHz our averages are slightly higher
than Eppler et al. at both polarizations. However, our values
include leads which have a higher emissivity than the surround-
ing ice pack. Omitting leads from our averages would reduce the
90 GHz emissivity by about 0.01, bringing our values somewhat
closer.

Table 5 contains AIMR estimates of emissivity for July
compared with historical estimates. AIMR 37 GHz estimates
are lower than Eppler’s averages and the measurements reported
by Onstott et al. They are near the lower end of Comiso’s
range. Our values are also lower than published averages at 90
GHz. Again, inclusion of leads and melt ponds in our averages
may account for some of the difference. Melt pond fraction has
been estimated on July 8 by Tschudi et al. [this issue] as
24.6%. They also estimate the fraction of open water, includ-
ing leads and holes in melt ponds as 51.1%, so a significant frac-
tion of the surface could be expected to have a reduced emis-
sivity compared to the MYI.

Figure 8 shows the variation with angle of horizontal and
vertical emissivities, $\varepsilon_0$ and $\varepsilon_0$, at each frequency for the
May and July cases. In the snow-covered situation in May, $\varepsilon_0$ shows
little dependence on angle while $\varepsilon_0$ decreases with angle (Fig-
ures 8a, 8b). Matzler [1987] finds similar signatures for dry
winter snow. The polarization contrast, $(\varepsilon_0-\varepsilon_0)/\varepsilon_0$, is smaller at
90 GHz than at 37 GHz as would be expected due to increased vol-
ume scattering by the snow layer. Scattering by MYI would
also tend to depolarize emissions at both frequencies compared
to a specular surface. The summer melt conditions in July (Fig-
ures 8c, 8d) show $\varepsilon_0$ increasing and $\varepsilon_0$ decreasing with angle at
37 GHz. The 90 GHz signature is more similar to the spring
time situation with smaller polarization contrast and fairly
constant values of $\varepsilon_0$ with angle. The average polarization con-
trast of 0.04 at this frequency compares to that for a specular
surface (open lead pixels in the same image) of 0.17, showing
the depolarization by scattering from a rough surface.

4.3. Application to Cloud Retrievals

The high variability in microwave emissivity on the scale
of airborne microwave measurements (100 m to 1km) has been
demonstrated in these results, and even smaller-scale varia-
tions have been documented by other researchers. Sub-pixel-

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**Table 4. Sea Ice Emissivity at 50° for Spring Conditions**

<table>
<thead>
<tr>
<th></th>
<th>37 GHz (H)</th>
<th>37 GHz (V)</th>
<th>90 GHz (H)</th>
<th>90 GHz (V)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AIMR (late May)</td>
<td>0.84</td>
<td>0.89</td>
<td>0.68</td>
<td>0.71</td>
</tr>
<tr>
<td>Eppler (dry MYI)</td>
<td>0.706 (0.079)*</td>
<td>0.764 (0.115)</td>
<td>0.650 (0.011)</td>
<td>0.680 (0.105)</td>
</tr>
<tr>
<td>Comiso (MYI, late May)</td>
<td>0.75 (0.088)</td>
<td>0.77 (0.90)</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>

*Standard deviations given in parentheses.
Table 5. Sea Ice Emissivity at 50° for Summer Melt Conditions

<table>
<thead>
<tr>
<th></th>
<th>37 GHz (H)</th>
<th>37 GHz (V)</th>
<th>90 GHz (H)</th>
<th>90 GHz (V)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AIMR (early July)</td>
<td>0.83</td>
<td>0.89</td>
<td>0.80</td>
<td>0.84</td>
</tr>
<tr>
<td>Eppler (summer melt)</td>
<td>0.93 (0.010)*</td>
<td>0.97 (0.015)</td>
<td>0.92 (0.010)</td>
<td>0.95 (0.020)</td>
</tr>
<tr>
<td>Comiso (MYI, July)</td>
<td>0.82-0.92</td>
<td>0.89-0.92</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Onstott (MYI, July)</td>
<td>0.90</td>
<td>0.96</td>
<td>0.94</td>
<td>0.95</td>
</tr>
</tbody>
</table>

*Standard deviations given in parentheses.

scale features, such as melt ponds, are often not resolved when viewing the surface from aircraft altitudes. At satellite resolutions, most leads will not be resolved. Here we consider the question of how well a quantity averaged over a pixel is represented by a single measurement of the quantity at the resolution of that pixel. The 50 x 50 km region analyzed in this study is of the same scale as a single satellite pixel. The AIMR provides about 2500 pixels at each viewing angle over this region. The mean emissivities we have presented in previous sections are derived from averages over those 2500 AIMR pixels. However, a satellite measurement would provide a single measurement of brightness temperature over the entire area. We simulate the satellite measurement by averaging brightness temperature and surface temperature measurements over the 50 x 50 km area and then use those averages to calculate a single emissivity for the area. The result should be representative of the value a satellite measurement would provide for a single pixel. Comparison of this result with the pixel-by-pixel calculations of emissivity show that the average values are quite similar (within about 1.5%). Thus we conclude that a coarse

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Figure 8. Variation of emissivity with viewing angle averaged along a 50 km flight segment near the SHEBA ship: (a) 37 GHz on May 20, 1998; (b) 90 GHz on May 20, 1998; (c) 37 GHz on July 8, 1998; (d) 90 GHz on July 8, 1998.
measurement of $T_s$ can still be used to approximate the aggregate value of emissivity even though sub-pixel-scale variability is not resolved.

The accuracy of surface emissivity estimates must also be considered when they are to be incorporated in cloud retrieval schemes. Over sea ice the surface contribution to the upwelling brightness temperature is easily the largest term, even in the case of thick liquid water clouds. A comparison from a summertime FIRE SHEBA flight (July 18, 1998) shows a $\Delta T_s$ of 10-20 K for a multilayer liquid water cloud compared to a clear-sky measurement of the surface on that day. A $T_s$ variation of 10 K would be equivalent to an uncertainty of 0.05 in $\varepsilon$. Therefore our emissivity uncertainty must be substantially less than 0.05 or the effect of the cloud on $T_s$ will be lost. In cirrus clouds, where $T_s$ is reduced due to scattering by ice crystals, Wang et al. [this issue] report $\Delta T_s$ of 13-38 K in storm related cirrus clouds and cite smaller depressions for arctic cirrus clouds, so accurate measurements of $\varepsilon$ are important in that case also.

An understanding of the temporal variations in emissivity over a given area is also useful when performing cloud retrievals. A clear-sky estimate of surface emissivity, such as those derived here, is needed to determine the contribution of a cloud to $T_s$. Alternatively, surface measurements could be made from an aircraft flying below the cloud base, but given the low-level nature of arctic clouds, this is not usually possible. Therefore surface emissivity measurements will not usually coincide in time with measurements of overlying clouds. The persistence of surface features and their associated emissivity is difficult to assess with the small data set considered, but we can draw some conclusions from these cases and comparisons with previous studies. For example, the two cases considered in May, which were separated by four days, gave virtually the same results. In the absence of precipitation, and with surface temperatures remaining approximately constant, we would not expect to see a drastic change over this time frame. Thus it would be reasonable to assume our emissivity estimates would be sufficient for a later cloud retrieval under those conditions. Similarly, the melt conditions observed during our July case appeared to persist for several days as surface temperatures remained near the freezing point, so we suspect the emissivity remained fairly constant.

Finally, the information contained in spectral gradients of emissivity might be useful in cloud retrieval algorithms. For example, the negative spectral gradient between 37 and 90 GHz

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**Plate 1.** Images of sea ice emissivity from AIMR at 37 and 90 GHz on May 20, 1998. The flight track is along the vertical axis of each image and passes over the SHEBA ship. Approximate dimensions of the images are 17 km by 50 km.
emissivities in May is similar to that observed by other researchers in snow-covered conditions. During the FIRE ACE flights we sometimes observed clouds that were visible at 90 GHz but appeared transparent at 37 GHz. Thus it might be reasonable to derive a 37 GHz emissivity from such images and use an assumed spectral gradient to derive the corresponding 90 GHz emissivity to perform cloud retrievals at the higher frequency. We have also noted that the gradient is very small during summer melt conditions, so again it might be possible to derive a 90 GHz emissivity from the 37 GHz value and apply that to cloud retrievals. Making such assumptions, however, would require some ancillary information about the surface conditions as well as a more quantitative assessment of the spectral gradient and its stability in time.

5. Summary and Conclusions

Emissivities of sea ice at a range of microwave frequencies have been calculated from airborne radiometer data. The method applied accounts for emission by the surface, atmospheric upwelling, and reflection of downwelling atmospheric radiation. Infrared surface temperatures and vertical profiles of atmospheric temperature and humidity are used as input. Emissivities at a range of frequencies have been calculated around the SHEBA site for clear-sky days during the FIRE SHEBA aircraft campaign in the spring and summer of 1998.

The spectral variations observed between 37 and 90 GHz in May are characteristic of snow-covered sea ice where volume-scattering increases with frequency thereby reducing the emissivity. The smaller spectral gradient between 37 and 90 GHz in July is indicative of significantly smaller penetration depths at each frequency due to the presence of liquid water on the surface. Data at 150 and 220 GHz in May show that the sign of the gradient is reversed at higher frequencies for snow-covered sea ice. We demonstrate that the relative values of snow grain size and wavelength result in size parameters suggestive of a transition from Rayleigh scattering to the Mie-scattering regime between 150 and 220 GHz.

Sources of uncertainty in our method are considered and found to produce errors in emissivity of the order 0.01-0.02. This estimate is substantiated by a comparison of 89 and 90 GHz emissivities from the AISR and MIR sensors. The change in brightness temperature produced by clouds observed in FIRE SHEBA is equivalent to a 0.05 variation in emissivity, so retrieval of cloud properties will require surface emissivity estimates with uncertainties smaller than 0.05.

Acknowledgments. This research was supported by the NSF SHEBA program, the NASA FIRE program, and a NASA Earth System Science Fellowship. We thank J. Wang for making the MIR data available to us and C. Walther of the NCAR Remote Sensing Facility for assistance with the AIRS data. Discussions with G. Liu and J. Maslanik were helpful in conducting this research. Comments provided by an anonymous reviewer on an earlier version of this paper were also helpful.

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(Received March 31, 2000; revised July 31, 2000; accepted August 9, 2000.)