IMPROVEMENTS TO 3I RETRIEVALS OVER SEA ICE AND APPLICATIONS TO ESTIMATING ARCTIC ENERGY FLUXES

Jennifer A. Francis
Polar Science Center, Applied Physics Laboratory
University of Washington
Seattle, Washington

1. INTRODUCTION

Numerous climate studies suggest that polar regions play a disproportionately important role in the global climate system: they may be very sensitive to changes (e.g., Wetherald and Manabe, 1988), and complex air-sea-ice feedback mechanisms may hold the key to climate stability (e.g., Thorndike, 1992). Recent observational studies of temperature trends in polar regions, however, are contradictory. Hansen and Lebedeff (1987) found an increase in Arctic air temperatures since 1880, while Stone et al. (1992) found no significant trend between 1958 and 1986. Because of our incomplete knowledge of air-sea-ice interactions, they may be poorly represented in general circulation and global climate models (Battisti, 1992). Better representations are essential if an accurate assessment and prediction of climate change is to be made (Ingram et al., 1989).

The historical dearth of polar data is the underlying reason for our relative lack of knowledge about these processes. There is little hope that we will ever have a better conventional observing network in polar regions because of their extreme climate and inaccessibility; thus we must turn to satellites and models as tools to gain a better understanding of interactions involving sea ice. Data sets from two Arctic field programs have recently become available and are valuable for comparison with and validation of satellite retrievals and model results.

This study focuses on the application of TOVS data processed by the Improved Initialization Inversion (3I) algorithm to studies of the Arctic energy budget. Unfortunately, the potential of TOVS-derived information for investigating air-sea-ice interaction and polar climate has not been realized owing to unacceptably large errors in retrievals over snow- and ice-covered regions. The primary reason for these errors is that 3I does not adequately account for the unique physical characteristics of snow and sea ice. Products from 3I have been used successfully, however, in studying high-latitude weather phenomena over the unfrozen sea (e.g., Claud et al., 1992a, b).

The objectives of the present investigation are to identify sea-ice-related deficiencies in TOVS-derived products in the Arctic during the cold season and to study methods of reducing these errors. The improved retrievals are then applied to the estimation of several components of the Arctic energy budget, including surface turbulent fluxes and the horizontal advection of sensible heat and water vapor from lower latitudes into and within the Arctic Basin.

2. DATA DESCRIPTION

2.1. Ground-truth in situ data

Data from two field experiments are used to validate retrievals from TOVS. The Coordinated Eastern Arctic EXperiment (CEAREX) drift phase was an ambitious program conducted near the Svalbard archipelago from September to January 1988/89. The research vessel Polarbjørn was driven into the pack ice northeast of Svalbard (~ 82°N, 30°E) and allowed to drift southward until it broke out in early January 1989. Throughout this period rawinsondes were launched at least every 12 hours, and a complete suite of meteorological and oceanographic measurements was obtained.
During March and April 1992, the Lead Experiment (LeadEx) field program was conducted in the Beaufort Sea northeast of Barrow, Alaska (LeadEx Group, 1993). The experiment consisted of an ice camp positioned near 73°N 146°W for six weeks and research flights by several aircraft. Data from two of the aircraft, a Convair C-131 from the University of Washington and a Twin Otter, are particularly useful for validation of TOVS retrievals. These transects mitigate some of the usual spatial incompatibility between in situ measurements and 100 km x 100 km TOVS pixels. Both aircraft were equipped with a downward-looking Heimann KT-19 temperature sensor, which measures infrared radiances in a 2° cone between 8 and 14 μm and converts this information to temperature. Data from the KT-19 provided high-resolution (= 10 m) measurements of ice surface temperature (IST).

2.2. Satellite data
TOVS Level 1-b radiances for the region north of 70°N were purchased from the Satellite Data Services Division (SDSD) of the National Oceanographic and Atmospheric Administration (NOAA) and from the National Center for Atmospheric Research (NCAR). Orbits corresponding to the CEAREX period were from the NOAA-10 polar orbiting satellite, and to LeadEx from NOAA-11. The TOVS radiances were processed with the Improved Initialization Inversion (3I) algorithm developed by the Atmospheric Radiation Analysis group at the Laboratoire de Météorologie Dynamique in Palaiseau, France (Chedin et al., 1985).

Hard-copy visible and thermal images from the Advanced Very High-Resolution Radiometer (AVHRR) and from Operational Linescan System (OLS) were obtained for subjective validation of cloud fraction retrievals.

3. PROBLEMS AND IMPROVEMENTS OVER SEA ICE
While TOVS has been used successfully in a number of applications, including investigations in the Greenland/Norwegian Sea and Antarctica (e.g., Claud et al., 1991, 1992a, 1992b; Heinemann, 1989), the potential of this instrument has been virtually unrealized in sea-ice-covered regions. This underutilization is primarily due to generally poor performance of TOVS retrievals in these areas. One of the goals in this study is to identify the sources of retrieval errors and investigate methods to remedy them.

Inadequacies in the retrieved products are identified by comparing them with ‘ground-truth’ rawinsondings and surface observations obtained during CEAREX and with aircraft measurements during LeadEx. Results show obvious systematic discrepancies that can be attributed to the fact that certain properties of sea ice and snow had not been accounted for in the algorithm. Following is a discussion of these problems and the approaches taken to ameliorate them.

Figure 1 compares two TOVS-derived temperature profiles with collocated radicinodings from the CEAREX drift phase. Figure 1a is a typical example of a case that is clear according to surface observers, but for which 3I retrieves a 25% cloud cover. The initial clear/cloudy determination is made in the beginning of the algorithm, and separate processing paths are followed by clear and cloudy boxes. In this case, the initial diagnosis is incorrect, which contributes to large errors in the retrieved temperature profile. In particular, note that the surface-based inversion is not captured, and the snow/ice surface temperature retrieval is 9 K too warm. Figure 1b illustrates another common situation: clouds are detected by both 3I and the surface observer, but the satellite-retrieved surface temperature is again much too warm and the low-level temperature inversion is not captured.

Surface temperature, boundary-layer stratification, and cloud conditions are crucial pieces
of information for diagnosing surface energy fluxes; thus these problems must be corrected before TOVS retrievals in can be used for this purpose.

3.1. Snow/ice surface microwave emissivity

As a first step toward improving 3I/TOVS retrievals over sea ice, a method to discriminate between open water and sea ice is required. Since the sea ice has a higher emissivity in MSU channel 1 (hereafter MSU1) than does open water, this channel can be used to identify sea ice. The 3I algorithm computes the surface emissivity at this frequency with two techniques (Claud et al., 1987): the first is a regression with MSU1 and MSU2, and the second is a regression with MSU1 and an infrared window channel (HIRS8 in day, HIRS18 at night). Initial emissivity estimates are systematically too low at view angles greater than 25° from nadir; at the swath edges where sea ice is known to exist, values are close to those expected for open water ( = 0.7). A correction for this problem is obtained by comparing the MSU1 brightness temperature (TB) of an ice-covered area viewed at nadir with the TB of the same area viewed at a different angle in another orbit within a few hours of the first. Figure 2 shows the relationship of (MSU1\textsubscript{nadir} − MSU1\textsubscript{θ}) versus θ, where θ is the view angle. The view-angle dependence revealed in this plot is consistent with model results (S. Warren, personal communication). Although MSU1 TBs in the Level-1b data were previously corrected to nadir over land and water, the dependence of snow/ice emissivity on view angle is not taken into account. Errors in TB are approximately 8% at the swath edge (i.e., = 18 K too low), which translates to emissivities that are 10% to 15% too small. A least-squares fit to these points is used to obtain an empirical correction factor (MSU1\textsubscript{cor}) to the MSU1 TBs and is shown as a solid line in Figure 2:

\[
\text{MSU1}_{\text{cor}} = -0.0765148 + 0.0034240^2.
\]  

More realistic estimates now allow the differentiation of ice-covered and open-water HIRS spots, which is needed for later modifications to cloud detection tests and surface temperature estimates.

3.2. Cloud Detection

As discussed previously, one of the most important preliminary steps in the retrieval algorithm is determining whether a HIRS spot is clear or cloudy. The validation of cloud retrievals over snow and ice, however, is a difficult problem. In addition to the incompatibility between surface-based and satellite observations, even a trained eye often cannot identify clouds, especially low stratus and thin cirrus, because of the low contrast in albedo and temperature between clouds and the frozen surface. Nevertheless, visible and infrared images from AVHRR, in addition to surface observations, are used to evaluate the success of the clear/cloud diagnosis. Although the problem of cloud detection over sea ice has certainly not been solved completely in this study, errors in the 3I procedure are reduced by accounting for certain characteristics of sea ice. Table 1 and the following discussion summarize changes to the cloud tests. For a complete description of the original cloud tests see Wahiche et al. (1986) and Chedin et al. (1985).

The Adjacent Spots Test (#7 in Table 1) and the Maximum Value Test (#9) require a relatively homogeneous surface temperature within a 3I box. These tests look for significant differences between adjacent spots in a HIRS window channel; the colder spot is declared cloudy. In sea-ice-covered regions, cracks and areas of thin ice are common and can cause large variations in surface temperature. Warm thin-ice-covered areas would be declared clear by these tests, and cold thick-ice-covered areas would be declared cloudy, even though both areas might actually be clear. Because of the new surface-type algorithm, these tests are now bypassed in sea-ice-covered regions, which results in a decrease in retrieved cloud cover over sea ice when leads are present, especially in mar-
original ice zones.

The MSU2 – HIRS11 Test (#10) did not exist in the original 3I package and exploits differences in TB between MSU2 and HIRS11 (corrected to nadir). These channels have similar sensitivity functions that peak near 700 mb; however, HIRS11 is affected by cloud droplets in the polar troposphere while MSU2 is almost completely unaffected. In clear sky the difference between these channels is about 8 K (MSU2 is colder because the atmosphere radiates less energy at this frequency); thus if HIRS11 – MSU2 < 7, the spot is declared cloudy. Normally HIRS11 is used in the retrieval of water vapor, but because the polar atmosphere is extremely cold and dry, this channel can be combined with MSU2 to detect thin mid-level clouds which are often overlooked by other tests, especially in winter when these clouds are most common and difficult to detect.

Finally, changes to the Window Channels Test (#3) and the HIRS8 – HIRS19 Test (#6) exploit the more subtle differences in TB between cloudy and clear scenes over ice-covered regions, and allow for clouds that are warmer than the surface.

Figure 3 compares CEAREX observations of cloud fraction with final TOVS/3I retrievals before and after the cloud tests were modified. A recent revision of the TIGR library (Achard, 1991), which added more polar profiles to the first-guess library, may also contribute to the improvement. The unmodified algorithm missed existing clouds in some cases and detected nonexistent clouds in others.

3.3. Snow/ice surface temperature

Good estimates of ISTs are crucial for calculating the surface radiation balance, for determining the stratification in the near-surface layer, and for estimating surface turbulent fluxes of latent and sensible heat. As illustrated in Fig. 1, however, surface temperatures retrieved by the original algorithm over sea ice are frequently 5 to 10 K too warm. An investigation of this problem reveals that these errors are probably caused by overestimates of water vapor content in the frigid, dry polar atmosphere. The 3I algorithm retrieves surface temperature and water vapor content simultaneously (Tahani, 1991) when the effective cloud fraction is less than 90%. Radiance at the top of the atmosphere are computed from retrieved temperature profiles and first-guess water vapor content, the results are compared with observed radiances, and if discrepancies exist the water vapor content and surface temperature are adjusted accordingly. If water vapor in middle and upper levels is overestimated, upwelling radiances computed for the window channels will be too low, and the algorithm compensates by increasing its retrieved surface temperature. To obtain more accurate estimates of IST, alternative retrieval methods were pursued.

3.3.1. Clear-sky cases

Because of the extremely low moisture content of the winter Arctic atmosphere, surface-emitted energy in the infrared window channels is attenuated very little as it passes through the cloud-free atmosphere. Consequently, instead of the original method used to retrieve surface temperature described above, which often produces ISTs that are much too warm, a simpler method is employed for clear boxes. Comparisons with CEAREX surface temperature measurements indicate that the brightness temperatures of HIRS18 (4.0 μm) and HIRS19 (3.7 μm) are good predictors of IST when there is no insolation. This result is consistent with IST retrievals from AVHRR imagery (Y. Yu, personal communication). While clear-sky errors using the original algorithm are consistently larger than 8 K, the rms error using this method is 2.2 K. Retrieved temperatures are still warmer than CEAREX observations, but this would be expected since the 3I footprint contains areas of leads and thin ice, while the in situ data are from a multiyear ice floe.

During daylight, surface temperature retrieval is more difficult because HIRS18 and HIRS19, which are the most insensitive to water vapor of all the HIRS window channels, are contaminated
by reflected solar radiation. Fortunately, HIRS8 (11.1 μm), which is also a window channel but is more sensitive to water vapor, can be used successfully in the dry polar atmosphere. After the HIRS8 TB is corrected for the surface emissivity and water vapor (Wahiche et al., 1986), it is used as a direct estimate of surface temperature. This method was validated by comparison with aircraft-based measurements from the LeadEx program, and results in rms errors of 1.7 K.

3.3. ii. Cloudy cases

Most retrievals of surface temperature from satellite measurements use IR and near-IR window channels. Because clouds are efficient emitters at these wavelengths, surface temperatures usually cannot be retrieved when cloud fractions are high. Because the Arctic is relatively cloudy, a method to estimate surface temperature in overcast conditions would be extremely valuable. The original 31 algorithm estimates surface temperature if the retrieved cloud fraction is below 90% (although in the plotting software a limit of 60% is recommended). As shown in Fig. 1, however, IST estimates are often significantly too warm. A method that employs the TOVS microwave channels is appealing because they are relatively insensitive to clouds. The MSU1 (window channel) brightness temperature is affected by a variety of surface characteristics besides surface temperature; separating the temperature signal from the other effects is impossible without detailed knowledge of the surface type. The application of MSU2 data, however, has produced encouraging results.

Comparisons with CEAREX surface temperature measurements suggest that MSU2 TBs vary linearly with ISTs when MSU2 TB < 236 K and when there is little insolation. This relationship holds in both clear and cloudy conditions, providing an estimate of IST when clouds prevent the use of IR window channels. Equation (2) is the regression used to predict IST from MSU2:

\[ \text{IST} = -3.548369 \text{ MSU2} + 0.019686 \text{ MSU2}^2, \quad \text{MSU2} < 236 \text{K}. \]  

(2)

This method is applied in cloudy scenes only; the IR window channels provide better estimates if cloud fraction is below 20%.

Results of Wexler (1936) and Overland and Guest (1991) may provide a physical explanation for this relationship. Overland and Guest (1991) analyzed clear-sky, snow-surface temperatures during winter from CEAREX and found that ISTs over thick ice are controlled by the downwelling IR radiation. This flux, in turn, is governed primarily by the temperature of the ‘radiation boundary layer,’ the level of maximum air temperature at the top of the temperature inversion, which is typically found between 0.5 and 2 km. Since a large fraction of the MSU2 signal emanates from this layer, the TB is physically related to the IST. Under cloudy skies, snow/ice surface temperatures are relatively warm (Colony, 1992) because of increased downward infrared fluxes from the high-emissivity clouds. In the Arctic winter, clouds generally occur in regions of warm advection ahead of cyclonic disturbances; thus the air is also warmer, which is detected by the MSU2 TB. The relationship in Eqn. 2 breaks down for MSU2 TB > 236 K, probably because the transfer of sensible heat counteracts warming due to increased downward infrared fluxes. In daylight, MSU2 cannot be used for IST estimates because the surface is warmed by insolation while the atmosphere is nearly transparent to it. Consequently, the coupling of downwelling infrared flux to snow/ice surface temperature is not maintained.

Finally, retrieved surface temperatures of unfrozen ocean in high latitudes are often below the freezing point of seawater (= 271.2 K). Because the 31 algorithm now distinguishes between seawater-covered and open water areas, this problem is solved by setting open-water temperatures to 271.2 K if the retrieval is below the freezing point. Similarly, the retrieved IST is not allowed to exceed 273.2 K.

These modifications to the surface temperature retrieval algorithm have reduced the rms
error for all CEAREX collocations (clear and cloudy) from 7.5 K (8 collocations) to 0.9 K (10 collocations). The rms error during LeadEx is 1.7 K (12 collocations).

4. APPLICATIONS TO ESTIMATING ENERGY FLUXES IN THE ARCTIC BASIN

TOVS provides information about the surface and atmosphere in the central Arctic where there are few conventional data. The resolution of TOVS products is also compatible with that of GCMs (general circulation models and global climate models), which rely on bulk parameterization schemes to simulate processes that occur on small scales. Information from TOVS, therefore, may be valuable for studies of synoptic–scale energy transfer processes, for validating GCM simulations, and ultimately for providing improved model parameterizations. A discussion of two promising applications is presented here: the exchange of turbulent energy in the lower troposphere, and the advection of heat and moisture from lower latitudes into the Arctic Basin.

4.1. Turbulent energy exchange

Two of the key variables required for estimating surface heat and momentum transfer are surface temperature and the stratification of the planetary boundary layer (PBL). As shown in Fig. 1, however, retrievals from the original 3I algorithm often do not capture the low-level temperature inversions that are nearly ubiquitous over ice and snow in polar regions (except during the melt season). Because of improvements to surface temperature estimates and cloud detection methods over sea ice, inversions are now retrieved more reliably, and an estimate of the bulk stratification of the planetary boundary layer is possible.

Using measurements from CEAREX and Soviet drifting ice stations, Overland and Davidson (1992) developed simple empirical relationships between the bulk stratification of the PBL (the difference in potential temperature between the 900 mb level and the surface $\Delta \Theta_{900}$) and two important PBL parameters: the geostrophic drag coefficient $C_g$ and the turning angle between the geostrophic wind and surface stress $\alpha$ (counterclockwise in the northern hemisphere):

\[ C_g = 0.037 - 8.3 \times 10^{-3} \left( \frac{N_{900}}{\bar{N}_{900}} \right)^4 \]  \hspace{1cm} (3)

\[ \alpha = 11.7 + 13.1 \left( \frac{N_{900}}{\bar{N}_{900}} \right)^2 \]  \hspace{1cm} (4)

where

\[ N_{900} = \left( \frac{g \Delta \Theta_{900}}{\Theta_{900} h_{900}} \right)^{1/2} \]  \hspace{1cm} (5)

and

\[ \bar{N}_{900} = 0.024 \text{ s}^{-1}. \]  \hspace{1cm} (6)

Here, $N_{900}$ is the Brunt-Väisälä frequency for the layer between the surface and 900 mb, $g$ is gravity, $h_{900}$ is the height of the 900-mb surface, and $\Theta_{900}$ is the median of $\Theta$. The TOVS/3I retrievals can provide estimates of $\Delta \Theta_{900}$, from which $C_g$ and $\alpha$ can be calculated. The 3I-derived stratification parameter shown in Fig. 4a agrees well with those computed from radiosonde data. Figure 4b is an example of the $C_g$ field derived from TOVS/3I retrievals for 5 November 1988 at 1200Z. Values have been interpolated to the NMC octagon grid to allow further calculations by incorporating NMC analyses. In the central Arctic Basin, where the lower atmosphere is very stably stratified because of a strong high pressure system, $C_g$ is small and $\alpha$ is large, as would be expected. The geostrophic drag coefficient increases toward the marginal ice zone where boundary layer stability decreases.

By combining the geostrophic wind $G$ derived from NMC surface pressure fields with
retrievals from TOVS/3I, estimates of the air-ice stress $\tau$ and the 10-m wind speed $U_{10}$ can be obtained. The product of $C_g$ and $G$ defines the friction velocity $u_*$, which can then be combined with air density $\rho$ to compute the air-ice stress: $\tau = \rho u_*^2$. Overland (1992) showed that if average, rather than actual, values of $C_g$ are used to compute $\tau$, results may be in error by a factor of four in highly stable conditions. One can also estimate $U_{10}$ using the neutral drag coefficient $C_{D0}$, which has an average winter value of $2.5 \times 10^{-3}$ (Overland and Guest (1991)): \[ U_{10} = \frac{C_g}{C_{D0}} G. \] (8)

Air-ice stress is an important parameter for understanding and modeling ice motion, and $U_{10}$ is required to compute surface turbulent fluxes. Figure 4c shows an example calculation of $U_{10}$. A comparison of $U_{10}$ estimates with observations during CEAREX shows errors in wind direction of approximately 20° ($\alpha$ too small), which is comparable to the accuracy of measurements from standard wind vanes. The wind speed, however, is well predicted in some situations and underestimated by as much as 50% in others. Investigation of the poorly predicted cases reveals that most of the error is not due to inaccurate estimates of the stratification, but rather arises from poor surface pressure analyses.

4.2 Horizontal heat and moisture advection

The tropics absorb more solar radiation than they radiate to space as terrestrial radiation, and thus are the source region for the global heat engine. The polar regions constitute the heat sink of this cycle, since they have an annual net energy deficit. The surplus of heat in the tropics is transported poleward by the atmosphere and ocean, where it is then radiated to space to maintain the earth’s radiative equilibrium. The horizontal advection of heat into the polar regions from lower latitudes, therefore, is an important component of the global energy cycle. Oort (1983) compiled atmospheric circulation statistics, from which estimates of the advection of heat and moisture into high latitudes were estimated by Oort (1974) and Nakamura and Oort (1988). The rawinsondings from the Arctic used to generate these statistics, however, were from land-based stations only, and very few were poleward of 80°N. For the area north of 70°N, Nakamura and Oort (1988) calculated that 98% (69%) of the energy lost to space annually (in winter) is transported by the atmosphere from lower latitudes in the form of sensible heat and water vapor. Transient eddies effect most of the transport, but the longitudinal variation in this flux is not known. TOVS retrievals at a horizontal resolution of 100 km x 100 km provide a means to estimate the horizontal advection of heat and moisture at synoptic time and space scales.

Products from TOVS/3I include vertical profiles of temperature, from which the geometric distance between two pressure levels $Z_T$ can be computed. Since this distance, or thickness, is directly proportional to the mean temperature of the layer, horizontal gradients in thickness $\nabla Z_T$ can be combined with the geostrophic wind vector $G$ to estimate the horizontal advection of sensible heat: \[ \frac{\Delta T}{\Delta t} = \frac{g}{R_d} \ln \left( \frac{p_1}{p_2} \right) (G \cdot \nabla Z_T), \] (9)

where $\Delta T$ is the change in mean-layer temperature during time interval $\Delta t$, $R_d$ is the dry gas constant, and $p_1$ and $p_2$ are the pressure levels bounding the layer of interest ($p_1 > p_2$).

Figure 4d is an example of the horizontal advection in the layer between 1000 mb and 700 mb, computed from TOVS-derived thicknesses and geostrophic winds calculated from the NMC surface pressure field at 1200 Z on 5 November 1988. In the area of maximum advection located north of Siberia, the thickness values are 20 to 30 m higher at 1200 Z the next day (not shown).
Horizontal advection of water vapor can be computed in a similar manner by combining retrieved layer-average water vapor content with geostrophic winds. While absolute values of retrieved water vapor content are apparently too large for polar airmasses, the gradients in water vapor appear reasonable (C. Claud, personal communication). Because it is the gradients, not the actual values, that are required to compute advection, this method may yield useful estimates of vapor transport. Validation of this quantity is difficult, however, since local sinks and sources of water vapor due to condensation, precipitation, and evaporation occur simultaneously, and moisture measurements from radiosondes are unreliable in extremely cold, dry conditions.

Weekly and monthly average estimates of advection should provide a better understanding of the primary sources of energy for the Arctic atmosphere and perhaps highlight regions that should be monitored for signs of change.

5. SUMMARY

Polar environments are some of the least understood components of the global climate puzzle because, until the advent of satellites, there has been little information with which to study them. The retrieval of geophysical variables from satellite measurements in high latitudes, however, presents unique challenges. This study focuses on problems associated with retrieving information from the TOVS instrument in the Arctic Basin during the cold season (no surface melting). Comparisons of products from the Improved Initialization Inversion (3I) algorithm with in situ data from field experiments in the Arctic Basin reveal significant and systematic errors in estimates of variables needed to study the surface energy balance. Microwave emissivity of the ice surface, for example, is as much as 15% too low because the dependence of the 50 GHz brightness temperature on satellite view angle over snow and sea ice had not been accounted for in the algorithm. Determination of surface type (open water or sea ice) is now possible after including this effect in the processing. Moreover, ice surface temperature estimates are often 5 to 15 K too warm, scenes are systematically misdiagnosed as clear or cloudy, and low-level temperature inversions are often not captured. Improvements to the ice surface temperature retrieval have reduced rms errors from approximately 7 K to 2 K; modifications to cloud detection tests have improved results over sea ice, particularly where it is inhomogeneous; and more accurate surface temperatures have improved estimates of the low-level stratification.

Applications of improved retrievals to studies of the Arctic energy budget are encouraging. Preliminary calculations of air–ice stress, 10-m wind speed, and horizontal advection of heat and moisture are presented. These applications will be expanded to estimates of weekly and monthly averages on regional and basin-wide scales, as well as to the calculation of surface radiation fluxes.

Acknowledgments

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### Table 1: Modifications to cloud detection tests in 3I for polar regions.

<table>
<thead>
<tr>
<th>Cloud Test in 3I</th>
<th>Original Test</th>
<th>Modifications to cloud test for polar regions</th>
</tr>
</thead>
<tbody>
<tr>
<td>3: Window Channels Test</td>
<td>Spot declared cloud if:</td>
<td>Spot declared cloudy if:</td>
</tr>
<tr>
<td></td>
<td>- day:</td>
<td>- Sea ice day:</td>
</tr>
<tr>
<td></td>
<td>[HIRS18c - HIRS8c] &gt; 12.0</td>
<td>HIRS19 - HIRS18 &gt; 6.0</td>
</tr>
<tr>
<td></td>
<td>- night:</td>
<td>- Sea ice night:</td>
</tr>
<tr>
<td></td>
<td>HIRS18c - HIRS8c &gt; 2.5 or HIRS8c - HIRS18c &gt; 4.0 or HIRS19c - HIRS18c &gt; 3.0 or HIRS18c - HIRS19c &gt; 4.0</td>
<td>HIRS19 - HIRS8c &lt; -1.0 or HIRS19 - HIRS8c &gt; 2.0</td>
</tr>
<tr>
<td>6: HIRS8 - HIRS19 Test</td>
<td>Spot declared cloudy if:</td>
<td>Spot declared cloudy if:</td>
</tr>
<tr>
<td></td>
<td>- night only:</td>
<td>- Sea ice, night only:</td>
</tr>
<tr>
<td></td>
<td>HIRS8 - HIRS19 &gt; -0.5</td>
<td>HIRS8 - HIRS19 &lt; -0.5 or HIRS8 - HIRS19 &lt; -4.0</td>
</tr>
<tr>
<td>7: Adjacent Spots Test</td>
<td>Compares IR window channel TBs in two adjacent spots – if ΔTB &gt; 1.5 K, spot declared cloudy.</td>
<td>Test not performed over sea ice because leads may cause difference in TBs that is due to surface temperature differences rather than cloud cover differences.</td>
</tr>
<tr>
<td>9: Maximum Value Test</td>
<td>Compares IR window channel TB of each HIRS spot in a box to the max. TB in the entire box – if ΔTB &gt; 4.0, declared cloudy.</td>
<td>Test not performed over sea ice for same reason given in #7.</td>
</tr>
<tr>
<td>10: MSU2 - HIRS11 Test</td>
<td>Did not exist in original 3I</td>
<td>Spot declared cloudy if:</td>
</tr>
<tr>
<td></td>
<td></td>
<td>- Airmass type 4 or 5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>HIRS11 - MSU2 &lt; 7</td>
</tr>
</tbody>
</table>
Fig. 1: Two examples of temperature profiles retrieved using the original 3I algorithm (dashed) compared with radiosondings from CEAREX (solid). The observed and retrieved profiles are less than 30 minutes apart in time and less than 50 km apart in distance. For the situation in orbit 11086 (5 November 1988), 3I retrieved a 25% cloud cover, whereas surface observers reported clear skies. For orbit 11114 (7 November 1988), both 3I and surface observers reported overcast conditions.

Fig. 2: Error (%) in MSU1 brightness temperature (TB) versus sensor view angle $\theta$. Symbols are the difference between the TB of a sea-ice location viewed at nadir ($\theta = 0^\circ$) and at $\theta > 0^\circ$ in another orbit within a few hours. The curve is a least-squares fit to the observations.

Fig. 3: Comparison of cloud fraction observed during CEAREX (solid), by the original 3I algorithm (short dash), and by the modified 3I algorithm (long dash).
Fig. 4: Examples of (a) the stratification parameter $\Delta \Theta_{900}$ (K), (b) the geostrophic drag coefficient $C_g \times 10^5$, (c) the 10-m wind speed $U_{10}$ (m s$^{-1}$), and (d) horizontal advection of sensible heat (K day$^{-1}$) on 5 November 1988 at 1200 Z. Retrievals from several orbits within 4 hours of 12 Z have been interpolated to the NMC octagon grid.
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J R Eyre

European Centre for Medium-range Weather Forecasts
Shinfield Park, Reading, RG2 9AX, U.K.

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