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TRMM NASA Grant NAG5-1586: Informal Status Report
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March 14, 1996

INTRODUCTION

Unfortunately Dr. Verner E. Suomi passed away on July 30 1995. Professor Suomi made numerous important contributions to the field of atmospheric sciences, including the TRMM program. We continue to work to fulfill Dr. Suomi's vision of extending, in space and time, the TRMM satellite observations of rainfall by using observations from geostationary satellites. Although geostationary satellites cannot probe the interiors of precipitating systems, they do observe their life-cycles. We have focused on using the geostationary observations to classify tropical cloud systems.

ACCOMPLISHMENTS AND UPCOMING WORK.

As noted in our RTOP report of 8 June 1995, we extended our original cloud classification scheme to deal with images from the GMS. We have also moved the computer programs off the McIDAS mainframe and onto a workstation to reduce costs. Testing of the algorithm revealed that the classification scheme was missing important cloud types when applied to the GMS data. Based on this result we decided to make extensive improvements to the classification scheme. These have been coded and in the next six months will be tested using training sets with the AIP/3 rainfall data developed.

We have further explored our observations of the high correlation between rain occurrence derived from SSMI and negative differences between the 11 and 6.7 μm brightness temperature differences (BT₁₁-BT_{6.7}). Using radiative transfer modeling and additional observations we have explained the occurrence of the negative BT₁₁-BT_{6.7} (Ackerman, 1995). However we have not been able to physically explain why the negative differences always occurred with an SSMI rainrate. This is being further investigated using GOES-8 and radar observations over the central US.

Our analysis of BT₁₁-BT_{6.7} led us to study negative differences in BT₁₁-BT₁₂. We have consistently observed negative differences over rapidly growing convective clouds. This study was presented at the 8th Conference on Satellite Meteorology (Moeller et al. 1996). We have not been able to explain these differences and are currently working with Dr. Bob Rabin of NOAA to explain the physical reasons for their occurrence. Negative differences in both BT₁₁-BT_{6.7} and BT₁₁-BT₁₂ are incorporated into the GOES cloud classification algorithm.

By the end of the summer the cloud classification will be trained to include rainfall estimates and ready for use with GMS and GOES data.

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GLOBAL SATELLITE OBSERVATIONS OF NEGATIVE BRIGHTNESS TEMPERATURE DIFFERENCES BETWEEN 11 AND 6.7 μm

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ABSTRACT

Global analysis of satellite spectral observations indicate the existence of negative brightness temperature differences between 11 and 6.7 μm (BT₁₁-BT_{6.7}) when cold scenes are viewed. Differences are typically greater than -5 K for the tropics and midlatitudes but can be smaller than -15 K over high-altitude polar regions during winter. In July, more than 60% of the observations over the Antarctic Plateau had BT₁₁-BT_{6.7} < -5 K. In January over Greenland, the frequency of occurrence is approximately 20%.

Three factors are investigated which may contribute to these observed negative brightness temperature differences: 1) calibration errors, 2) non-uniform scenes within the field of view and 3) physical properties of the observed phenomena. Calibration errors and non-uniform scenes may generate values of BT₁₁-BT_{6.7} that are less than zero; however, these differences are on the order of -2 K and therefore cannot fully explain the observations.

A doubling and adding radiative transfer model is used to investigate the physical explanations of the negative differences. Simulations of satellite spectral observations for thick clouds produce negative differences that are comparable to those observed in the tropics and midlatitudes. The magnitude of the differences are a function of cloud microphysics, cloud top pressure, view angle and the cloud optical thickness. The model simulations are also capable of producing large negative differences over high-altitude polar regions.

Distinguishing clear and cloudy regions from satellite infrared radiances is a challenging problem in polar winter conditions. Brightness temperature differences between 11 and 6.7 μm provide a technique to separate cold, optically thick clouds from clear-sky conditions when strong radiation inversions exist at the surface. The presence of a cloud inhibits the development of this inversion, and shields its detection using satellite radiance measurements. While physically reasonable, and in agreement with radiative transfer calculations, this technique has not been verified with ground nor aircraft observations. Further evidence that the large negative values of BT₁₁-BT_{6.7} are associated with surface inversions is presented by comparing the satellite observations with surface temperature measurements from an Antarctica automated weather station.

¹A portion of this work was completed during the author's visit to the Institute of Atmospheric Physics, DLR, Germany.

1. INTRODUCTION

Radiance measurements from satellites are now routinely used to monitor global climate parameters. Global analysis of cloud amounts, types, and optical properties using satellite measurements have been developed and are currently operational, e.g., the International Satellite Cloud Climatology Project (Rossow et al. 1988; Rossow and Garder, 1993) and CO₂ Slicing; (Wylie et al. 1994). Frey et al. (1995) developed an end-to-end algorithm which performs collocation and analysis of High-resolution Infrared Sounder/2 (HIRS/2) and Advanced Very High Resolution Radiometer (AVHRR) observations. This algorithm, referred to as Collocated HIRS/2 and AVHRR Products (CHAPS), was designed to operate on global observations and in a real-time, hands-off operational mode. Normally (see Section 2) the 11- μ m brightness temperature is greater than the 6.7- μ m brightness temperature; however, the CHAPS data indicates a large frequency of occurrence in which the 6.7- μ m brightness temperature is greater than the 11- μ m brightness temperature. This paper discusses the nature of these negative differences and provides explanations for their existence.

Section 2 presents the observations of negative brightness temperature differences over high-altitude, snow-covered regions during night and over tropical clouds with cold cloud tops. Section 3 explores these negative differences with radiative transfer calculations. The results are summarized in Section 4.

2. OBSERVATIONS OF NEGATIVE BRIGHTNESS TEMPERATURE DIFFERENCES BETWEEN 11 AND 6.7 μ m

The database used in this study consists of collocated observations from the AVHRR the HIRS/2 radiance measurements (both on the NOAA satellites), and National Centers for Environmental Prediction (NCEP) analysis fields. This collocation was accomplished by a global cloud-retrieval algorithm using the operational National Oceanic and Atmospheric Administration (NOAA) meteorological polar orbiting satellites. The CHAPS algorithm was run in real-time mode during July 1993 and January 1994 and January 1994 using observations made from the NOAA-12

spacecraft. The NOAA-12 satellite is in an approximately 102 minute sun-synchronous orbit with local observations made near 0730 and 1930 LST for most latitudes, but with more frequent sampling in the polar regions.

Discrimination between cloudy and clear scenes is essential to deriving accurate cloud products such as distributions of cloud altitude, cloud amount, and spectral cloud radiative forcing. The CHAPS clear-sky detection algorithm is presented in Frey et al (1995). These global, coincident observations from the AVHRR and the HIRS/2 provide a database to explore potential differences in the radiative characteristics of clouds in different regions of the world at different times of the year. This paper focuses on HIRS/2 observations of channels 8 (11 μm) and 12 (6.7 μm).

In clear-sky situations, the 6.7- μm radiation measured by satellite instruments is emitted by water vapor in the atmospheric layer between approximately 200 and 500 hPa (Soden and Bretherton 1993; Wu et al 1993) and has a brightness temperature (BT_{6.7}) related to the temperature and moisture in that layer. The 6.7- μm radiation emitted by the surface or low clouds is absorbed in the atmosphere and is not sensed by the satellite instruments. On the other hand, absorption by atmospheric gases is weak at 11- μm . Under clear sky conditions, the measured 11 μm radiation originates primarily at the surface, with a small contribution by the near-surface atmosphere. The region between 200 and 500 hPa is not sensed at 11 μm . Because the surface is normally warmer than the upper troposphere, 11- μm brightness temperatures (BT₁₁) are normally warmer than 6.7- μm brightness temperature; thus the difference, BT₁₁-BT_{6.7}, is normally greater than zero. The CHAPS global analyses of BT₁₁-BT_{6.7} show large negative differences (less than -10K) over the Antarctic Plateau and Greenland during their respective winters. Large negative differences may be indicative of clear-sky conditions and the existence of a strong low-level temperature inversions.

Figure 1 depicts the frequency of occurrence of HIRS/2 observations of $-5 \leq (\text{BT}_{11} - \text{BT}_{6.7}) \leq 0$ (Figure 1a) and $(\text{BT}_{11} - \text{BT}_{6.7}) < -5$ (Figure 1b) for July 1993 over the south pole on a $2.5^\circ \times 2.5^\circ$ latitude-longitude grid. Over the East Antarctica highlands, 60% of the observations

showed large negative differences (less than -5 K). The sharp boundary in the frequency of occurrence between east and west Antarctica is a result of the Transantarctica Mountains.

Figure 2 shows a frequency of occurrence of negative $BT_{11}-BT_{6.7}$ observed over the Arctic in January 1994. Figure 2a is of $0 \leq (BT_{11}-BT_{6.7}) \leq -5$ and 2b represents the number of occurrences of $(BT_{11}-BT_{6.7}) < -5$. More than 15% of the observations over the Greenland plateau have $BT_{11}-BT_{6.7} < -5$. Other regions of large frequency of occurrence include Siberia and the Canadian Northwest Territory. Figures 1 and 2 display a distinct bias towards land in the occurrence of negative brightness temperature differences.

In polar regions, strong surface radiation inversions can develop as a result of longwave energy loss at the surface due to clear-skies and a dry atmosphere. Figure 3 is a temperature (solid-line) and dew point temperature (dashed-line) profile measured over the South Pole at 0000 UTC on 13 September 1995 and illustrates this surface inversion. On this day the temperature inversion was approximately 20 K over the lowest 100 m of the atmosphere. The surface temperature was more than 25 K colder than the temperature at 600 hPa. Temperatures over Antarctica near the surface can reach 200 K (Stearns et al. 1993), while the middle troposphere is 235K. Under such conditions, satellite channels located in strong water vapor absorption bands, such as the 6.7- μm channel, will have a warmer equivalent brightness temperature than the 11- μm window channel. This brightness temperature difference between 11- and 6.7- μm may be very useful for detecting cloud-free conditions in the polar night. A doubling/adding model of the azimuthally averaged form of the infrared radiative transfer equation was used to simulate the brightness temperature differences at 11 and 6.7 μm using the temperature and moisture profile of Figure 3, and a spectrally varying surface emissivity (Warren 1982; Salisbury and D'Aria 1992). The simulated $BT_{11}-BT_{6.7}$ difference was -14 K.

As indicated in Figure 1, over the East Antarctica highlands, 60% of the observations showed large negative differences. Strong radiative cooling to the east of the Transantarctica Mountains is an important ingredient in the development of the katabatic winds that flow down through these mountains (Swithinbank 1973; Bromwich 1989; Breckenridge et al. 1993). Satellite

observations of BT₁₁-BT_{6,7} for winter highlands may aid in monitoring the development of this strong nocturnal inversion.

A box-plot (Tukey, 1977) of the BT₁₁-BT_{6,7} observations versus BT₁₁ observations, over intervals of 1K, over the south-pole region in July 1993 is shown in Figure 4. In this box-plot the squares represent median values and the rectangles denote the 75% and 25% quartiles. The whiskers represent the non-outlier maximum and minimum defined mathematically as the maximum and minimum observed values that lie within the range

$$q_{2.5} - 1.5(q_{range}) > \text{Observation} < q_{7.5} + 1.5(q_{range})$$

where $q_{2.5}$ and $q_{7.5}$ represent the 25% and 75% quartiles and $q_{range} = q_{7.5} - q_{2.5}$, is the inter-quartile range. Extreme values (asterisks) are cases satisfying:

$$q_{2.5} - 3.0(q_{range}) < \text{Observation} > q_{7.5} + 3.0(q_{range})$$

Outliers (circles) are denoted for those observations which are not extreme values and satisfy:

$$q_{2.5} - 1.5(q_{range}) < \text{Observation} > q_{7.5} + 1.5(q_{range})$$

See Figure 5 for a summary of the box-plot legend. As BT₁₁ gets colder, there is a decreasing trend in the median value of BT₁₁-BT_{6,7}. Note the number of outliers and extreme values that occur for BT₁₁ greater than 230 K. These outliers, which occur for the more negative values of BT₁₁-BT_{6,7}, could result from the presence of a thin cirrus overlying a strong surface inversion. The impact of thin cirrus on the negative brightness temperature differences is addressed in Section 2.4. In the 200 K to 210 K range BT₁₁ there are outliers of positive values of BT₁₁-BT_{6,7}. These outliers and extreme values are similar to the differences observed over tropical cold clouds, whose temperatures are similar to the clear-sky polar conditions discussed above.

Figure 6 depicts the frequency of occurrence of $-5 < \text{BT}_{11} - \text{BT}_{6,7} < 0$ within 50° of the equator, binned in a 2.5° latitude-longitude geographic grid, during January 1994 and July 1993. Unlike the polar winter cases, there are few values of $(\text{BT}_{11} - \text{BT}_{6,7}) < -5$. The occurrences of negative differences correlate highly with expected regions of convection. The Intertropical Convergence Zone is apparent in both months, as is the shift in its position to the summer hemisphere. The summertime peak in convection over South America and convection associated

with the southwest summer monsoon are also clearly visible in this figure. In addition, negative differences also appear in the mid-latitude storm tracks of the winter hemisphere.

There are three factors which may contribute to these observed negative brightness temperature differences: 1) Calibration errors, 2) non-uniform scenes within the field of view and 3) physical properties of the observed phenomena. Each of these factors is discussed below.

2.1 CALIBRATION ERRORS

The HIRS/2 instrument uses a rotating mirror to scan the earth by stepping at 1.8° intervals. To cover the 2240-km cross-track scan requires 56 steps of the mirror. The HIRS/2 enters calibration mode after 37 scans of the earth. Calibration is achieved by viewing space and a warm reference target (Lutz et al 1990). The noise equivalent differential radiance (NEAN) of the 11- and 6.7- μm channels are approximately 0.10 and 0.19 $\text{mW m}^{-2} \text{str}^{-1} \text{cm}^{-1}$, respectively. Error analysis conducted by Wu et al. (1993) suggests NEAN of individual observations of $\pm 0.2 \text{ K}$ and $\pm 0.92 \text{ K}$ in the BT₁₁ and BT_{6.7} observations, respectively. While absolute calibration errors can contribute to negative BT₁₁-BT_{6.7} observations, such errors cannot generate the -5 K differences shown in Figures 1b and 2b.

2.2 NON-UNIFORM FIELDS-OF-VIEW

The HIRS/2 instantaneous field-of-view (IFOV) is approximately 18 km in diameter at nadir and approximately 58 km cross-track by 30 km along-track at the end of the scan. CHAPS limits its analysis to $\pm 30^\circ$ of nadir. As the Planck function is non-linear, a non-uniform IFOV composed of two blackbodies with different temperatures can generate a spectral brightness temperature difference if the spectral channels are widely spaced. Consider an IFOV containing a scene with two clouds, one with a temperature of 190 K and the other at a temperature of 215 K. Assuming the clouds behave as blackbodies, the observed radiance is estimated by linearly averaging the radiance at each of these temperatures as a function of the fraction of cloud cover within the IFOV, that is

$$I_{measured}^{\lambda} = f I_{190}^{\lambda} + (1-f) I_{215}^{\lambda}$$

where I_{190}^{λ} and I_{215}^{λ} are the radiance at 190 K and 215 K, respectively at wavelength λ , f is the fraction of the field-of-view covered by the cloud with a temperature of 190 K, and $I_{measured}^{\lambda}$ is the measured radiance of the non-homogeneous IFOV. The measured radiance is then converted to a brightness temperature and the result shown in Figure 7. The solid and dashed lines represent the BT_{6.7} and BT₁₁ observations as a function of f . The difference is the dotted line whose magnitude is noted on the right-hand ordinate. It is possible to generate a BT₁₁-BT_{6.7} difference of -1.5 K based on scene non-uniformity; however, some of the observed differences over tropical convection are less than -3 K, in the case of polar winter scenes less than -15 K.

2.3 PHYSICAL PROPERTIES OF THE OBSERVED PHENOMENA

Absolute calibration errors and non-uniform FOVs do not exclude a natural cause for the observed negative BT₁₁-BT_{6.7} quantities. Figure 8 is a box-plot of the brightness temperature difference versus BT₁₁ for observations made in January 1994 and July 1993 and bounded by 50°N and 50°S latitudes. The relationship between the median value of BT₁₁-BT_{6.7} versus BT₁₁ is similar for the two months. January displays greater variability about this relationship, particularly in the non-outlier maximum and minimum values. For BT₁₁ less than approximately 206K, the median values of BT₁₁-BT_{6.7} are negative and, except for the few values of BT₁₁=186 K, remain negative. Over this geographic regions there are very few values of BT₁₁-BT_{6.7}<-5. The extreme values and outliers observed for BT₁₁ < 200K over the Antarctic (Figure 4) are uncommon observations in the tropical convective regions of the earth. This suggests that the extreme and outlying values of Figure 4 are thick, cold clouds. This is supported by a scatter diagram of BT₁₁-BT_{6.7} versus BT₁₁ (not shown) for the two geographic regions. Thus, at cold temperatures in the polar regions, the BT₁₁-BT_{6.7} versus BT₁₁ analysis may distinguish clear-sky inversion cases from cold clouds. This is investigated in the following section using theoretical simulations.

2.4 SIMULATIONS OF DIFFERENCES OVER STRONG NOCTURNAL INVERSIONS

The large negative differences of BT₁₁-BT_{6.7} observed over the polar high-altitude plateaus in winter are consistent with theory and physical expectations. Detecting the presence of this surface inversion is important, as the energy exchanges at a snow-ice surface are dominated by radiative processes (Maykut 1986; Gow and Tucker 1990). During winter conditions, the magnitude of the longwave net radiation balance is largely driven by the presence or absence of clouds. For cloud-free conditions the net infrared flux is large and is directed away from the surface, thus causing the cooling of the lower layers of the atmospheres. This radiative cooling plays an important role in the formation of cold air masses which may develop into downslope winds (foehns and boras).

In this section, model simulations were used to investigate if the negative brightness temperature difference can be observed in the presence of cirrus clouds. The temperature and moisture profiles of Figure 3 were used in a sensitivity analysis. The effects of the presence of high cloud on the simulated satellite observed radiances were investigated assuming cirrus clouds situated between 250 and 300 hPa.

This sensitivity analysis was performed using the azimuthally averaged form of the infrared radiative transfer equation:

$$\mu \frac{dI(\tau, \mu)}{d\tau} = I(\tau, \mu) - (1 - \omega_0) B(T) - \frac{\omega_0}{2} \int_{-1}^1 P(\tau, \mu, \mu') I(\tau, \mu') d\mu' \quad (1)$$

where $B(T)$ is the Planck function at temperature T , ω_0 is the single scattering albedo, $P(\tau, \mu, \mu')$ is the phase function and τ is optical depth. The radiative transfer equation was solved using a doubling and adding model with a spectral resolution of 20 cm^{-1} . Spectral absorption by gases was assigned according to Stackhouse et al. (1991). As this study aims to assess the potential impact of clouds obscuring the temperature inversion, Mie theory was used to calculate the single scattering properties of the cloud. The particle-size distribution was assumed to follow a gamma distribution with an effective radius of $12 \text{ }\mu\text{m}$. The surface IR spectral reflectance follows that of Warren (1982) and Salisbury and D'Aria (1992) and was assumed to be independent of angle. If

no cloud is present, the $BT_{11}-BT_{6.7}$ difference is -14 K. Varying the surface spectral emissivity changes the difference by approximately ± 0.5 K. Figure 9 depicts the theoretical simulation of $BT_{11}-BT_{6.7}$ versus cloud ice water path (IWP) for a nadir view. This simulation indicates that clouds with an IWP of greater than approximately 20 g m^{-2} will obscure a surface temperature inversion, if one exists. For thick clouds, the brightness temperature differences are significantly less than those associated with nocturnal inversions. The sensitivity of $BT_{11}-BT_{6.7}$ to cloud microphysical and macrophysical properties is explored in Section 2.6.

2.5 COMPARISON WITH SURFACE METEOROLOGICAL OBSERVATIONS

The model simulations of the winter Antarctic atmosphere are consistent with the HIRS/2 observations. While these simulations are convincing, validation that the differences result from the presence of a temperature inversion require additional measurements of the state of the atmosphere. Further evidence is presented by comparing the satellite observations with temperature measurements from an automated weather station (AWS). Stearns et al. (1993) discuss the instrumentation and data processing procedures of the Antarctic AWS measurements. Figure 10 compares the satellite measurements with the AWS observations for July 1993. The solid line is the 3 hourly averaged air temperature measured at Dome C (74.50°S, 123°E). The 30°C temperature range reflects changes in cloud conditions. The warmer air temperatures are a result of increased downward longwave radiative flux due to the presence of cloud (Sinkula, 1993). The open circles represent the BT_{11} measured by the HIRS/2 within a 0.5° latitude/longitude grid centered on Dome C. The satellite measured brightness temperatures capture the daily changes in the temperature measured at Dome C. There are many instances during July 1993 when the BT_{11} was less than the AWS measured air temperature, indicating the existence of a surface temperature inversion. The larger the difference between the AWS air temperature and the BT_{11} , the steeper the surface temperature inversion. The filled circles in Figure 10 represent observations of $BT_{11}-BT_{6.7}$, the scale of which is on the right-hand axis. For this location and month, $BT_{11}-BT_{6.7}$ less

than -10°C occur when the AWS air temperature is less than -70°C , an indication of clear-sky conditions and the presence of surface radiation inversion (Sinkula, 1993).

2.6 SIMULATIONS OF DIFFERENCES OVER COLD TROPICAL CLOUDS

Observations at 11 and $6.7\ \mu\text{m}$ have been used to study high clouds (e.g., Szejwach 1982, Pollinger and Wendling 1984, Liou et al. 1990). This section focuses on simulating the observed negative differences between the two spectral brightness temperatures. The BT_{11} and $\text{BT}_{6.7}$ observations were simulated using the doubling/adding model discussed above. The gas distributions and temperature profile of a standard tropical atmosphere were assumed and an optically thick cloud (LWP of $40\ \text{g m}^{-3}$) was placed at various levels in the atmosphere. A range of effective particle radii was assumed along with different viewing angles consistent with the CHAPS observations (Frey et al. 1995). Figure 11 depicts the range of the simulated $\text{BT}_{11}-\text{BT}_{6.7}$ as a function of the BT_{11} values. For cold clouds ($\text{BT}_{11} < 210\ \text{K}$) negative values are produced, in good agreement with the observations (e.g., Figure 8). Negative values arise due to non-linearity of the Planck function and the presence of water vapor above the cloud tops. While the emissivity of the cloud is a function of particle size, the spectral differences in the emissivity between $6.7\text{-}\mu\text{m}$ and $11\text{-}\mu\text{m}$ are small; however, gray clouds will generate negative differences of several tenths of a degree due to the non-linear relationship between the observed radiance and brightness temperature. In addition, the amount of radiation reflected at cloud top is greater at $6.7\ \mu\text{m}$ due to the larger downward radiance incident on cloud top. As the cloud top is placed lower in the atmosphere, water-vapor absorption above the cloud decreases the $6.7\text{-}\mu\text{m}$ radiance, with only a small effect on the $11\text{-}\mu\text{m}$ observations. Thus, the negative differences are reduced and eventually become positive, as in the clear sky condition. The range in $\text{BT}_{11}-\text{BT}_{6.7}$ for a given BT_{11} results from varying cloud microphysical properties, view angles and water vapor structure above the cloud top.

If a physical mechanism is responsible for the observed differences over cold clouds, then other satellite observations should also measure negative values of $\text{BT}_{11}-\text{BT}_{6.7}$. Negative values

have also been observed with the GOES-7 VAS (Visible Infrared Spin Scan Radiometer (VISSR) Atmospheric Sounder) in the Multi-Spectral Imaging (MSI) mode. Calibration of the VAS instrument is discussed by Menzel et al (1981). Random noise of individual observations at a scene temperature of 230 K are ± 1.2 K and ± 0.3 K in the BT_{6,7} and BT₁₁ observations, respectively (Menzel, 1995). Eight cloud systems, including their environments, in the eastern tropical Pacific (10°S-20°N; 120°W-150°W) between 1700 and 1900 UTC were sampled using the GOES-7 VAS 11- and 6.7- μ m channels during the fall of 1992. A histogram of BT₁₁-BT_{6,7} values as measured by the VAS is given in Figure 12. Negative values are observed by the VAS that are consistent with the HIRS/2 observations and theoretical simulations.

4. SUMMARY

Global observations of cold scenes by the HIRS/2 demonstrate that valid 6.7- μ m brightness temperature measurements may be greater than the corresponding window channel values. Similar observations were seen in a limited set of GOES-7 VAS observations. At issue is the physical explanation of these observations and potential applications to remote sensing the environment.

Calibration errors and non-uniform fields-of-view are capable of generating negative values of BT₁₁-BT_{6,7}; however, the magnitude is not comparable to the observations reported here. Radiative transfer model studies indicate that the negative differences are a result of a strong surface temperature inversions or cold clouds high in the troposphere.

A doubling and adding radiative transfer model is used to investigate the physical explanations of the negative differences. Simulations of the satellite spectral observations for thick clouds produce negative differences that are comparable to those observed in the tropics and midlatitudes. The magnitude of the differences are a function of cloud microphysics, cloud top pressure, view angle and the cloud optical thickness. The model simulations are also capable of producing large negative differences over high-altitude polar regions in the presence of a surface inversion.

The brightness temperature differences between 11- and 6.7- μm can be used to detect the existence of strong surface inversions. Clouds inhibit the formation of the inversion and obscure the inversion from satellite detection if the IWP is greater than approximately 20 g m^{-2} . In July, more than 60% of the observations over the Antarctic Plateau had $\text{BT}_{11}-\text{BT}_{6.7} < -5 \text{ K}$. In January over Greenland, the frequency of occurrence is approximately 20%. Further evidence is presented by comparing the satellite observations with temperature measurements from an automated weather station (AWS) Dome C (74.50°S, 123°E). The satellite measured BT_{11} captured the daily changes in the air temperature measured at Dome C. There were many instances during July 1993 when the BT_{11} was less than the AWS measured air temperature, indicating the existence of a surface temperature inversion. For this location and month, $\text{BT}_{11}-\text{BT}_{6.7}$ less than -10°C occurred when the AWS air temperature is less than -70°C -- a surface temperature indicative of clear-sky conditions and the presence of surface radiation inversion.

The analysis presented demonstrates that negative differences are also observed over high clouds with temperatures colder than approximately 210 K. There is, however, much scatter in the observations and whether this variability is associated with specific cloud types and atmospheric conditions requires a more detailed analysis.

Clouds in the polar regions, where the cloud temperature is often nearly the same, or even warmer than the surface temperature, are difficult to detect using passive radiometric measurements from satellites. During the polar winter, cloud detection techniques must revert to infrared (IR) algorithms only, increasing the difficulties of detection. A method to distinguish clear-sky conditions over high-altitude polar regions during nighttime may be feasible using brightness temperature differences between 11- μm and 6.7- μm observations. This clear-sky detection method is based on the occurrence of strong surface radiative inversions that develop in winter conditions.

ACKNOWLEDGMENTS

This work was supported by the NOAA Climate and Global Change Program under grant NA26GP0234-01, NASA FIRE Program under Grant NAS1-1177 and the NASA TRMM Grant NAG5-1586. The author is indebted to programming efforts of R. A. Frey and discussions with C. Moeller.

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6. LIST OF FIGURES

- Figure 1. Frequency of occurrence, as a function of $2.5^\circ \times 2.5^\circ$ latitude-longitude grid, of HIRS/2 observations of $0 \geq (BT_{11} - BT_{6,7}) \geq -5$ (Figure 1a) and $(BT_{11} - BT_{6,7}) < -5$ (Figure 1b) for July 1993 over the south pole. Percentages are listed in the legend.
- Figure 2. Frequency of occurrence, as a function of $2.5^\circ \times 2.5^\circ$ latitude-longitude grid, of HIRS/2 observations of $0 \geq (BT_{11} - BT_{6,7}) \geq -5$ (Figure 2a) and $(BT_{11} - BT_{6,7}) < -5$ (Figure 2b) for January 1994 over the north pole. Percentages are listed in the legend.
- Figure 3. Temperature (solid-line) and dew point temperature (dashed-line) measured over the south pole on 00 UTC on September 13, 1995.
- Figure 4. Box-plot of the negative $BT_{11} - BT_{6,7}$ observations versus BT_{11} observations, over intervals of 1K, over the south-pole region in July 1993. Squares represent median values, the rectangles denote the 75% and 25% quartiles. The whiskers represent the non-outlier maximum and minimum. Outliers and extreme values are denoted by circles and asterisk respectively. See Figure 5 for a schematic representation.
- Figure 5. Schematic diagram of a box-plot.
- Figure 6. Frequency of occurrence of $-5 < BT_{11} - BT_{6,7} < 0$ within 50° of the equator, binned in a $2.5^\circ \times 2.5^\circ$ latitude-longitude geographic grid, during January 1994 and July 1993.
- Figure 7. Effect of cloud variability in the FOV on the brightness temperatures of 11- and 6.7- μm . Two clouds in the FOV are assumed, one with a temperature of 190 K and the second having a temperature of 215 K. f represent the 190 K cloud fraction. The solid and dashed lines represent the $BT_{6,7}$ and BT_{11} observations as a function of f . The difference between the two brightness temperatures is the dotted line whose scale is the right-hand ordinate.

Figure 8. Box-plot of the negative $BT_{11}-BT_{6,7}$ observations versus BT_{11} observations, over intervals of 1 K, observed in January 1994 (a) and July 1993 (b) for measurements bounded by 50°N and 50°S latitudes.

Figure 9. Theoretical simulation of $BT_{11}-BT_{6,7}$ versus cloud ice water path for clouds overlying a strong nocturnal inversion typical of the Antarctic Plateau.

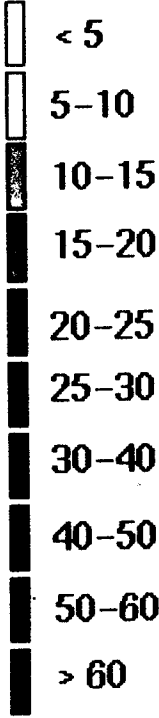
Figure 10. Comparison of satellite measurements with the surface observations over Antarctica during July 1993. The solid line is the 3 hourly averaged air temperature measured at Dome C (74.50°S , 123°E). The open circles represent the BT_{11} measured by the HIRS/2 within a 0.5° latitude/longitude grid centered on Dome C. The filled circles represent observations of $BT_{11}-BT_{6,7}$, the scale of which is on the right-hand axis.

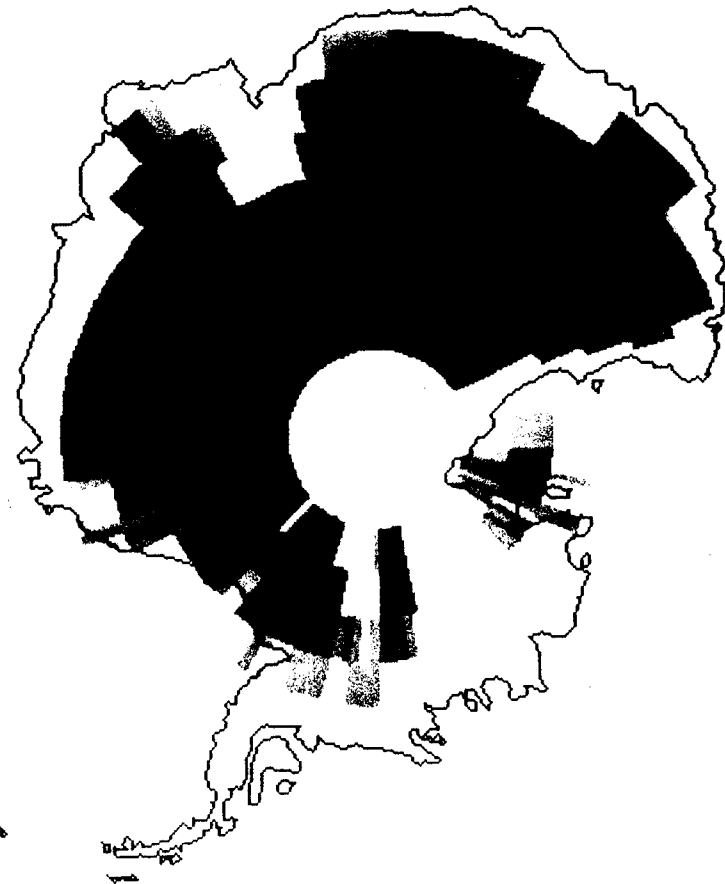
Figure 11. Range of simulated $BT_{11}-BT_{6,7}$ as a function of BT_{11} assuming a standard tropical atmosphere and various cloud types and viewing angles.

Figure 12. Histogram of $BT_{11}-BT_{6,7}$ values as observed by the GOES-7 VAS instrument.

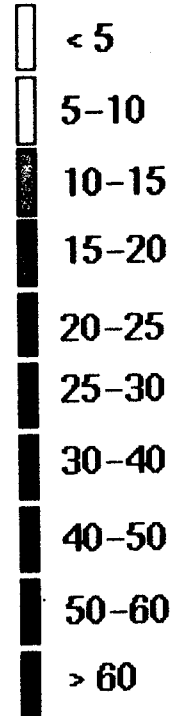


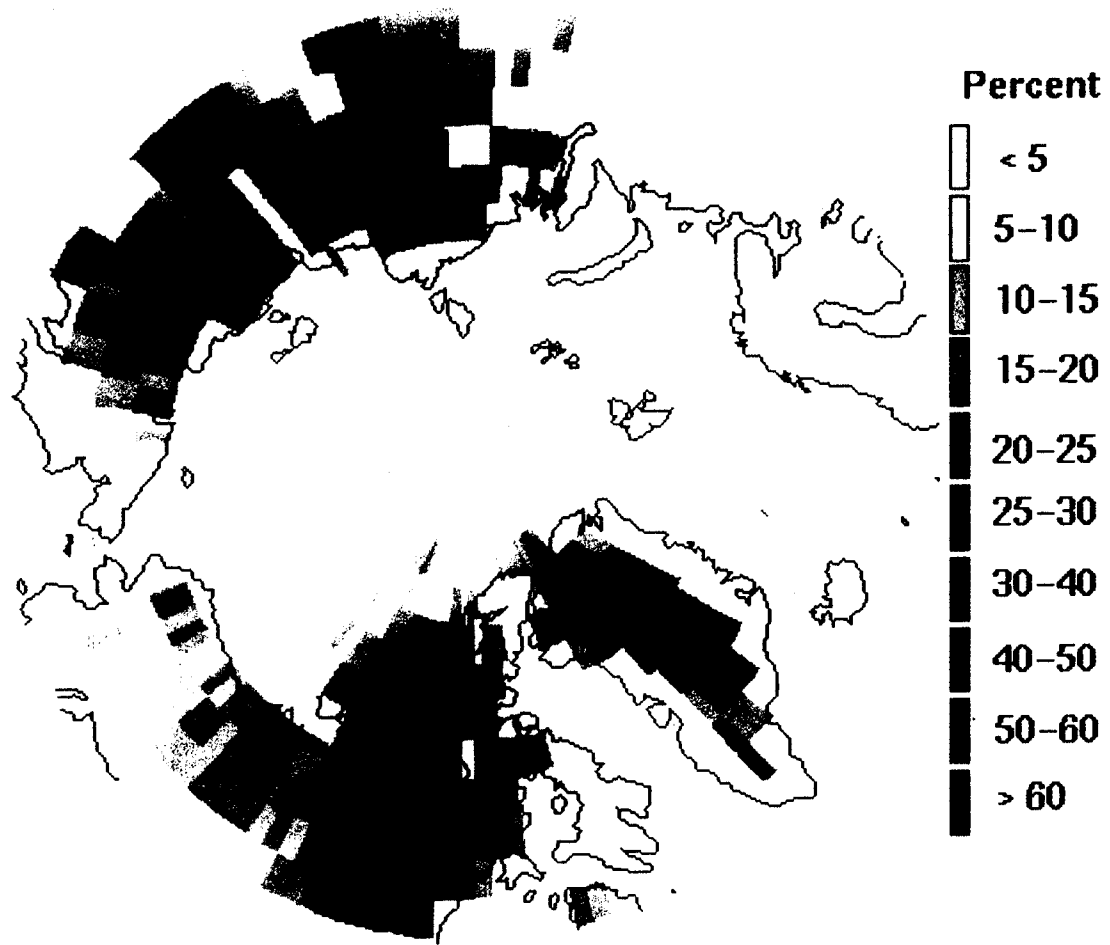
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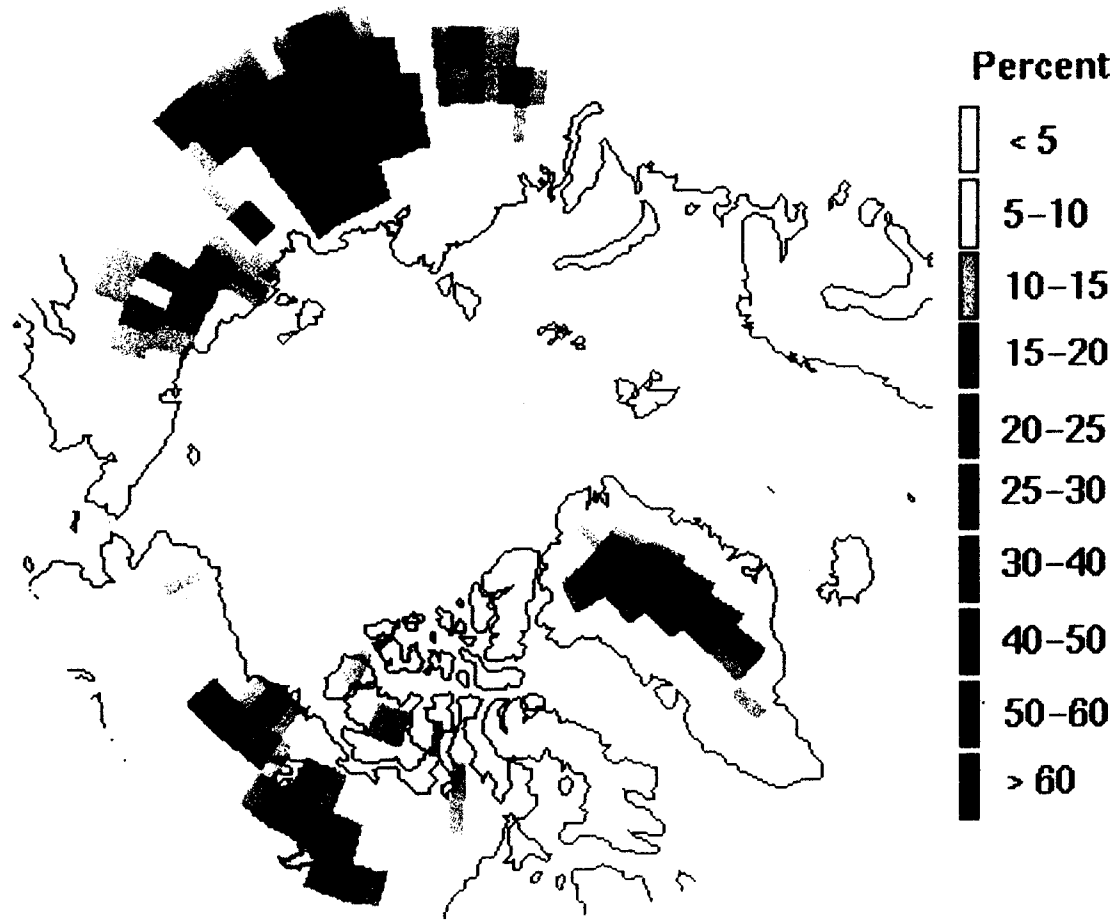




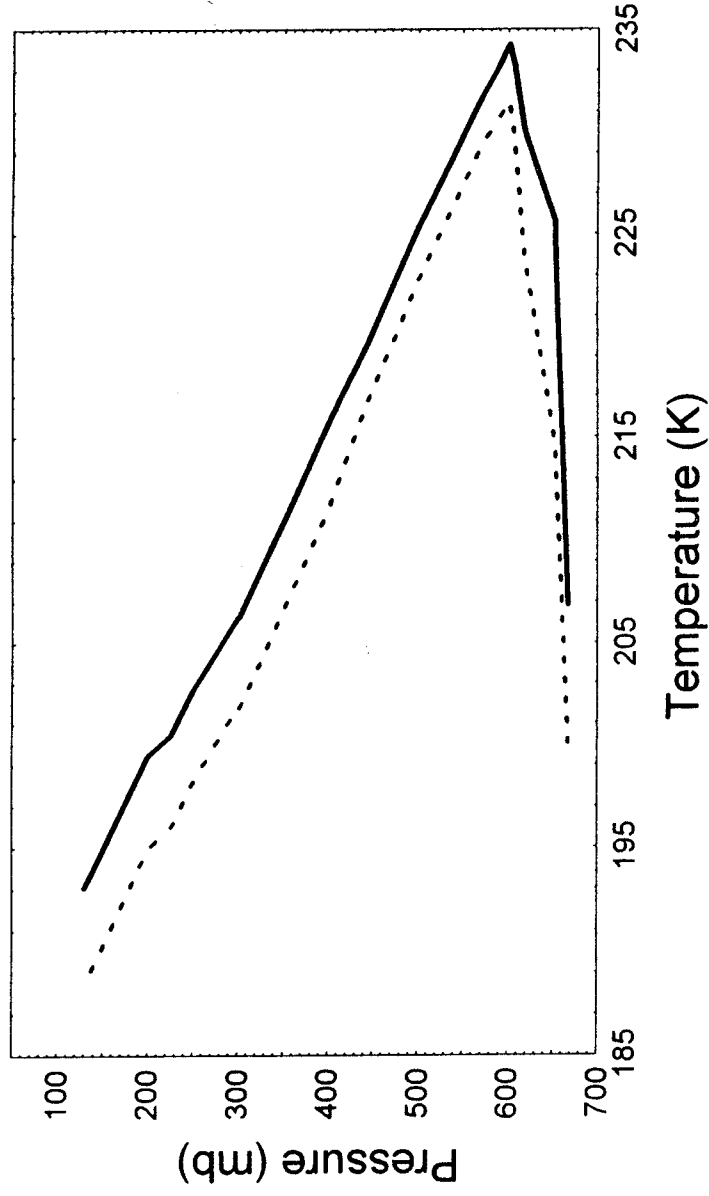
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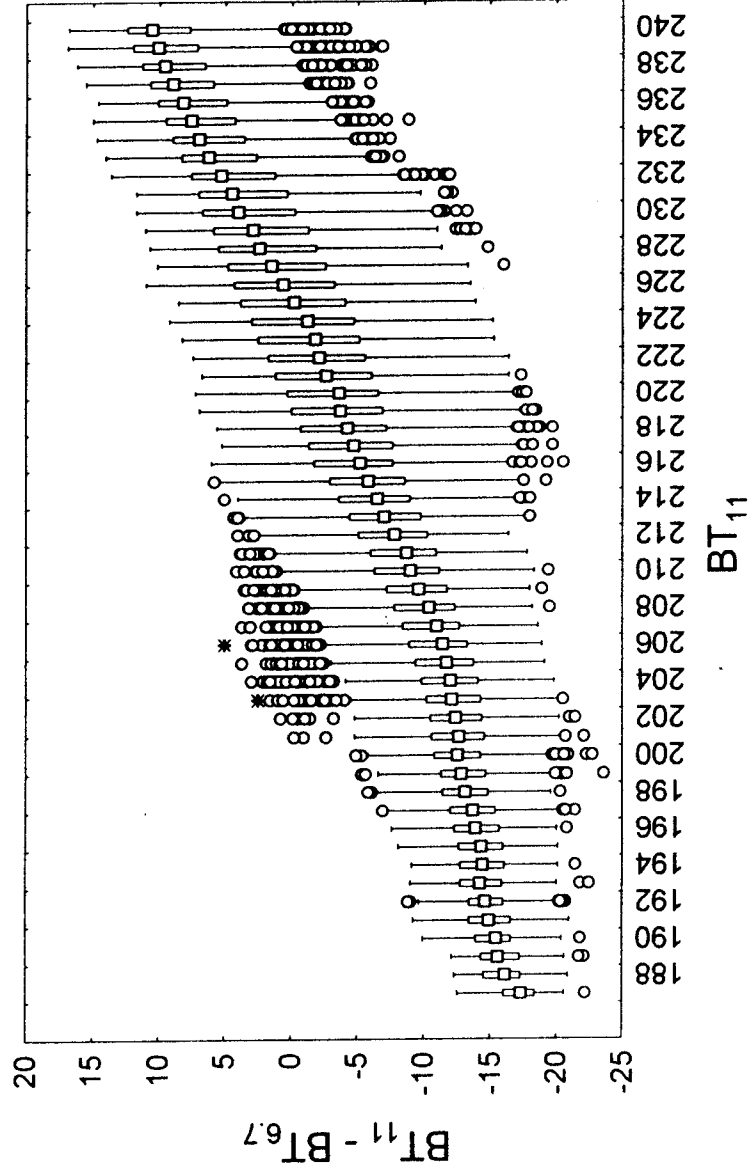


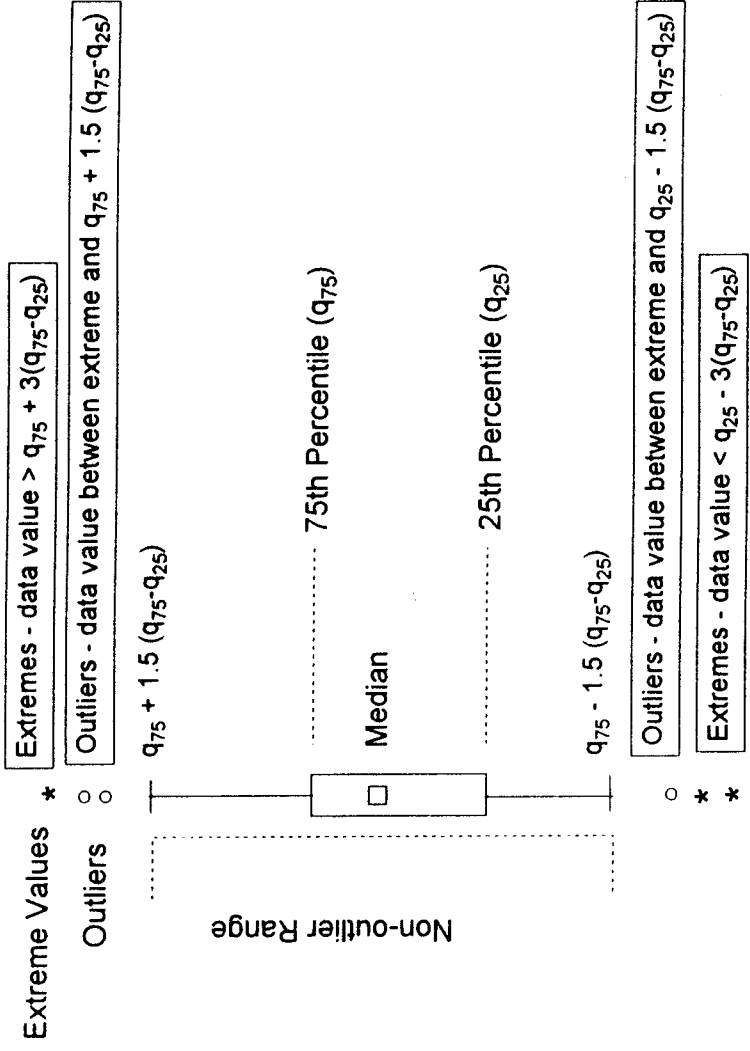


South Pole (September 13, 1995 0000 UTC)



July 1993 (South Pole Region)








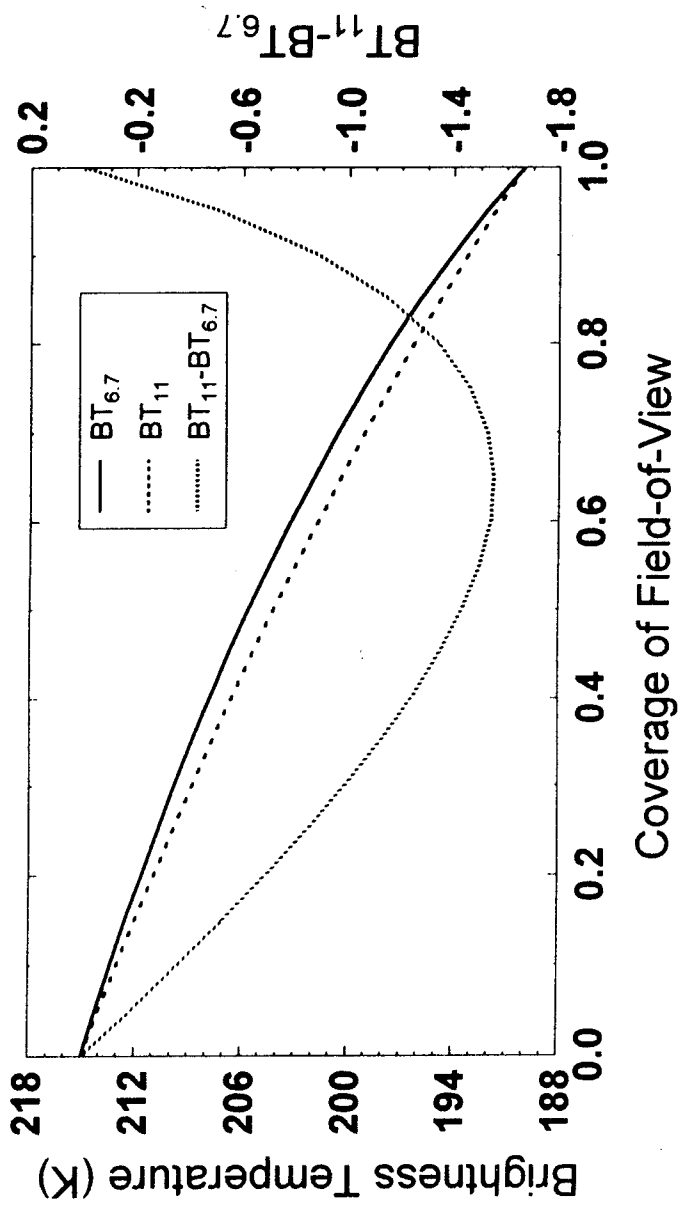
July

January

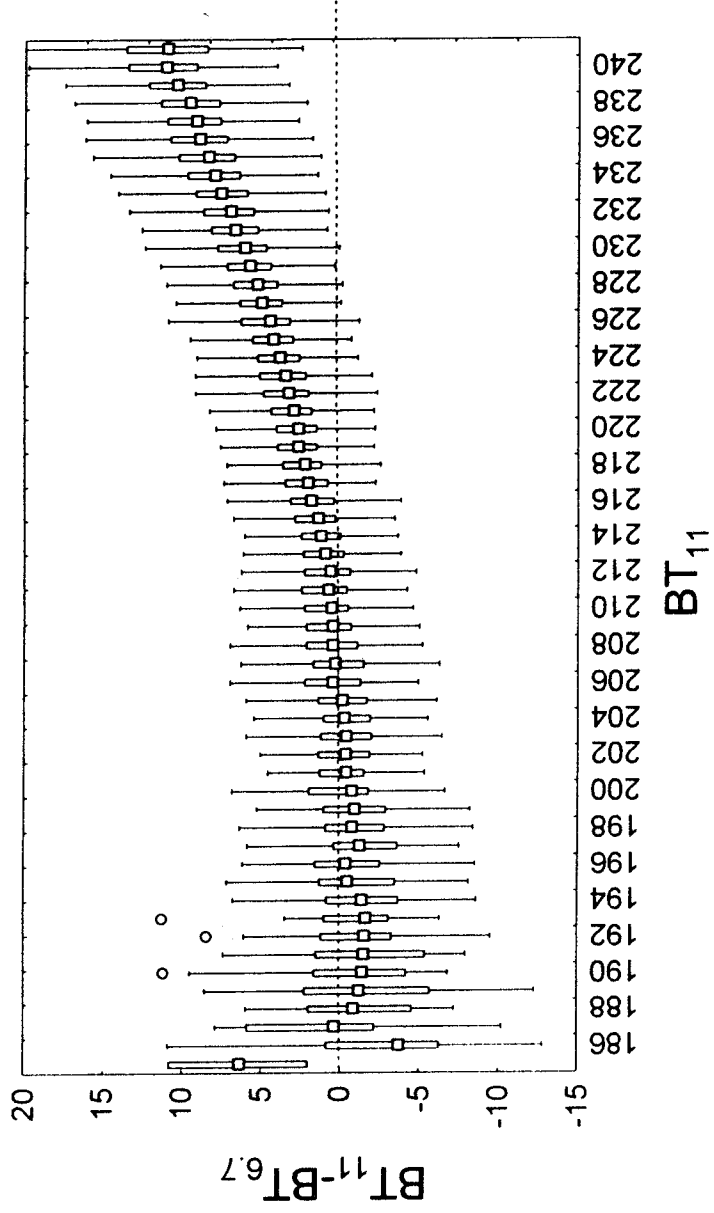


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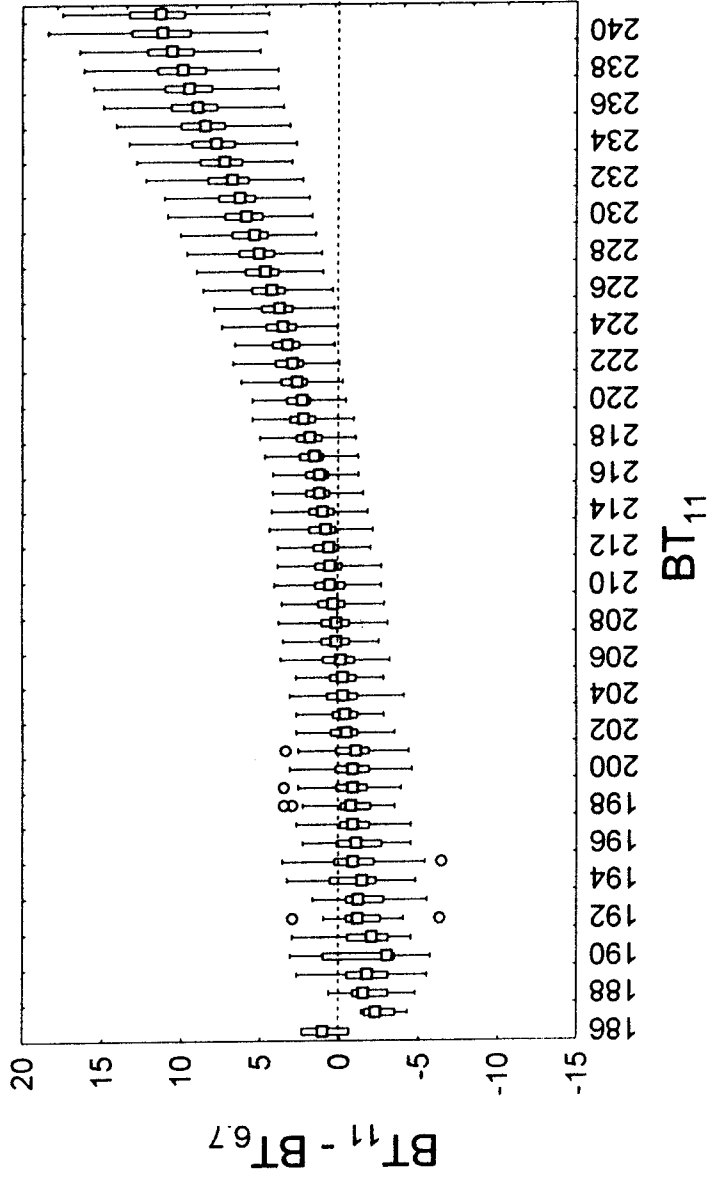
 5-10%



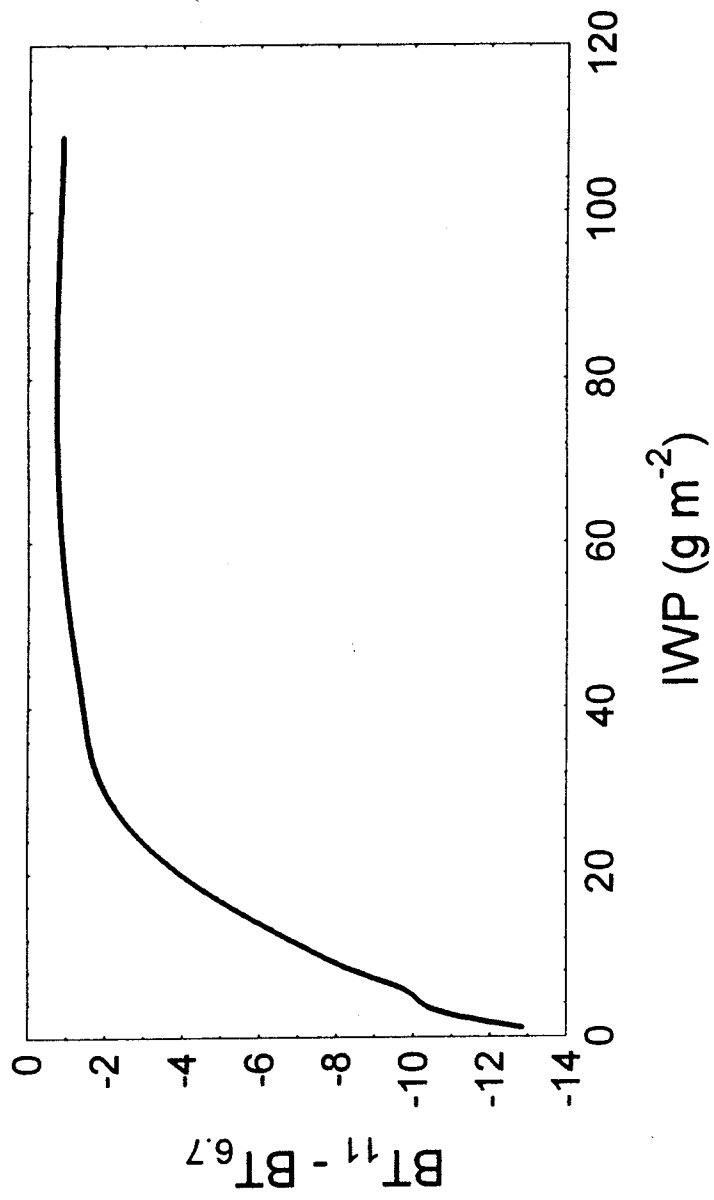
January 1994 (50N to 50S Latitude)



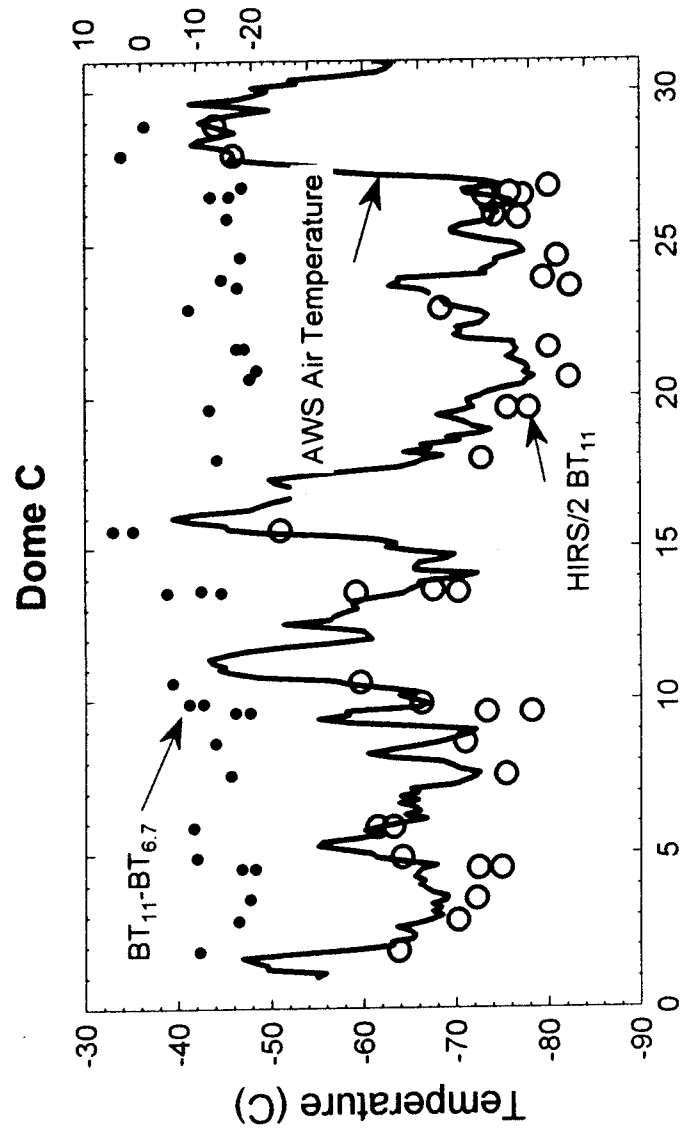
July 1993 (50N to 50S)



Simulations over Antarctica

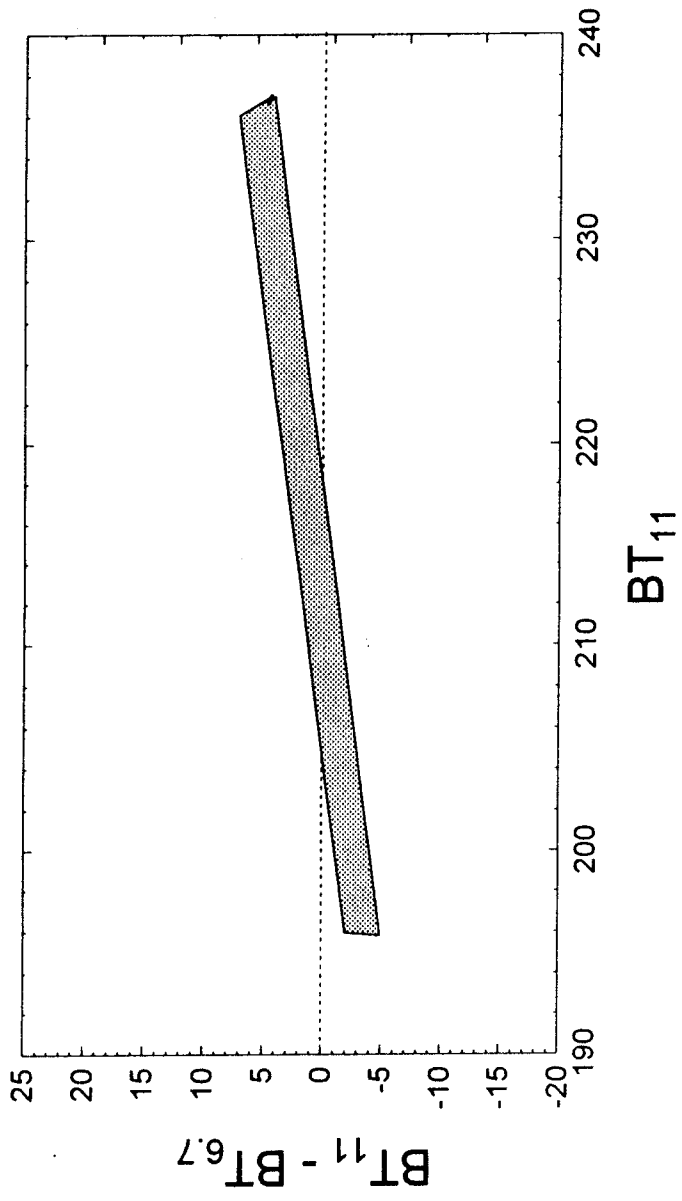


BT₁₁-BT_{6.7} (C)



July 1993

Simulations



GOES-7 VAS Observations

