Characteristics of Satellite-Derived Clear-Sky Atmospheric Temperature Inversion Strength in the Arctic, 1980–96

YINGHUI LIU
Cooperative Institute for Meteorological Satellite Studies, University of Wisconsin—Madison, Madison, Wisconsin

JEFFREY R. KEY
Office of Research and Applications, NOAA/NESDIS, Madison, Wisconsin

AXEL SCHWEIGER
Polar Science Center, University of Washington, Seattle, Washington

JENNIFER FRANCIS
Rutgers–The State University of New Jersey, New Brunswick, New Jersey

(Manuscript received 8 September 2005, in final form 21 February 2006)

ABSTRACT

The low-level atmospheric temperature inversion is a dominant feature of the Arctic atmosphere throughout most of the year. Meteorological stations that provide radiosonde data are sparsely distributed across the Arctic, and therefore provide little information on the spatial distribution of temperature inversions. Satellite-borne sensors provide an opportunity to fill the observational gap. In this study, a 17-yr time series, 1980–96, of clear-sky temperature inversion strength during the cold season is derived from High Resolution Infrared Radiation Sounder (HIRS) data using a two-channel statistical method. The satellite-derived clear-sky inversion strength monthly mean and trends agree well with radiosonde data. Both increasing and decreasing trends are found in the cold season for different areas. It is shown that there is a strong coupling between changes in surface temperature and changes in inversion strength, but that trends in some areas may be a result of advection aloft rather than warming or cooling at the surface.

1. Introduction

Low-level atmospheric temperature inversions are nearly ubiquitous at high latitudes during the polar winter, and are the dominant feature of the atmospheric temperature field in the Arctic (Curry et al. 1996). They may result from radiative cooling, warm air advection over a cooler surface layer, subsidence, cloud processes, surface melt, and topography (Vowinkel and Orvig 1970; Maykut and Church 1973; Busch et al. 1982; Curry 1983; Kahl 1990; Serreze et al. 1992). Temperature inversions influence the magnitude of heat and moisture fluxes through openings in the sea ice (leads). The depth of vertical mixing in the boundary layer, aerosol transport, photochemical destruction of boundary layer ozone at Arctic sunrise, surface wind velocity, and lead formation (Andreas 1980; Andreas and Murphy 1986; Bridgman et al. 1989; Barrie et al. 1988; Barry and Miles 1988). Knowledge of inversion characteristics is therefore needed for process studies and modeling, for example, for simulating the movement of sea ice (Overland 1985; Hibler and Bryan 1987). A number of recent studies have investigated the characteristics of polar temperature inversions based on radiosonde data (Bradley et al. 1992; Kahl 1990; Kahl et al. 1992a; Serreze et al. 1992). However, temperature inversion information is spatially incomplete, being limited to point measurements at coastal and interior meteorological stations.

Amplified warming in the northern high latitudes is a pervasive feature of general circulation model simula-
tions with enhanced greenhouse forcing. This amplified warming is partly because of the temperature–albedo feedback associated with the retreat of snow and sea ice (Houghton et al. 2001). In addition, the breakdown of the shallow but steep near-surface temperature inversions in the polar regions is another possible factor contributing to polar amplification (Chapman and Walsh 1993). Studies of trends in temperature inversion characteristics are therefore needed to improve our understanding of polar amplification, and some work has been done in this area. The decreasing trend in mid-
winter surface-based inversion depths along a transect from Alaska to the Canadian high Arctic for the period 1966 to 1990 was investigated by Bradley et al. (1993), Walden et al. (1996), and Bradley et al. (1996). A long-term increasing trend of low-level temperature inversion strength was found over the Arctic Ocean from 1950 to 1990 based on radiosonde data from Russian drifting stations, and dropsonde data from U.S. Air Force “Ptarmigan” weather reconnaissance aircraft (Kahl and Martinez 1996). However, these studies were limited in their spatial extent and did not include the 1990s, a period in which rapid changes have been observed.

Temperature inversion information can be directly derived from radiosonde data, but such radiosonde data are sparsely distributed in the Arctic. Generally, accurate temperature inversion information cannot be obtained from gridded meteorological fields owing to the limited vertical resolution. Satellite-borne sensors provide an opportunity to monitor the clear-sky temperature inversions in the polar regions. Liu and Key (2003, hereafter LK03) developed an empirical algorithm to detect and estimate the characteristics of clear-sky, low-level temperature inversions using data from the Moderate Resolution Imaging Spectroradiometer (MODIS) on the Terra and Aqua satellites. However, MODIS does not yet provide a long enough record for climatological analyses. The TIROS-N Operational Vertical Sounder (TOVS) has observed the earth’s surface and atmosphere since 1979 with spectral channels similar to those used in the MODIS algorithm, and therefore provides an opportunity to study changes in temperature inversions over the last two decades. In this paper we adapt the empirical MODIS method to the High Resolution Infrared Radiation Sounder (HIRS), which is part of the TOVS instrument, for the period 1980–96. Of primary interest are the spatial distribution and trends in inversion strength (INVST), defined as the temperature difference between the top of the inversion and the surface. Results are restricted to clear-sky conditions during the cold season (November through March) in the Arctic.

2. Data

The radiosonde data in this study are from Historical Arctic Radiosonde Archive (HARA) (Kahl et al. 1992b), which comprises over 1.5 million vertical soundings of temperature, pressure, humidity, and wind, representing all available radiosonde ascents from Arctic land stations north of 65°N from the beginning of the record through 1996. Radiosonde data from the series of former Soviet Union drifting ice stations during the period 1954–90 are also included. All the soundings are processed with quality controls using the method described by Serreze et al. (1992). Twice-
daily sounding data from 1980 to 1996 from 61 land stations (Fig. 1) and the drifting ice stations are used.

There are many different definitions of the inversion layer in the literature (Bilbello 1966; Maxwell 1982; Kahl 1990; Serreze et al. 1992; Bradley et al. 1992). In this study, for each sounding that contains an inversion, the inversion base is defined by the station elevation and the inversion top is the atmospheric level with the maximum temperature between the surface and the 700-hPa level. Isothermal layers at the base, top, or embedded within an inversion layer are included as a part of an inversion, as are thin layers with a negative lapse rate, provided that they are not more than 100 m thick. With these criteria, both surface-based inversions (Bradley et al. 1992) and elevated inversions (Serreze et al. 1992) are included in this study, as long as the surface is colder than the maximum temperature below 700 hPa. The inversion strength is defined as the temperature difference between the inversion top and the surface; inversion depth is defined as the altitude difference between the inversion top and the surface. We note that inversion strength can also be defined as a ratio of the temperature and altitude differences across the inversion, but we use the simpler definition to be consistent with the studies cited above. Because inversion depth is more difficult to retrieve than inversion strength (LK03), especially with a relatively low spatial resolution sensor such as HIRS, only inversion strength is considered here.

HIRS brightness temperatures (BT) at 7.3 μm and 11 μm are used in the clear-sky inversion strength retrieval. The HIRS data used in this study include NOAA-6 (1979–82), NOAA-7 (1983–84), NOAA-9 (1985–86), NOAA-10 (1987–16 September 1991), NOAA-11 (17 September 1991–1994), and NOAA-12 (1995–96) data. The spatial resolution of the original HIRS brightness temperature data is 17 km at nadir. Cloud detection tests from the 3I algorithm (Francis 1994; Stubenrauch et al. 1999) are applied to distinguish clear from cloudy scenes.
3. Theoretical basis and method

Under clear conditions, temperature inversion strength can be estimated using brightness temperatures of absorbing and nonabsorbing thermal infrared channels. The peaks of the weighting functions for the 7.3- and 11-μm channels are approximately 650 hPa, and the surface, respectively, as shown in Fig. 2. The brightness temperature of the window channel at 11 μm, BT$_{11}$, is most sensitive to the temperature of the surface. The 7.3-μm water vapor channel brightness temperature, BT$_{7.3}$, has the greatest contribution from the layer around 650 hPa. The magnitude of the brightness temperature difference (BTD) between the 7.3- and 11-μm channels, BT$_{7.3}$-BT$_{11}$, will therefore be related to the temperature difference of the inversion top and the surface; that is, the inversion strength. Here, BT$_{13.3}$ - BT$_{11}$ and BT$_{13.6}$ - BT$_{11}$ are also related to inversion strength, where weighting functions for the 13.3- and 13.6-μm carbon dioxide channels peak even lower than that of 7.3-μm channel. However, BT$_{7.3}$ - BT$_{11}$ is more effective than BT$_{13.3}$ - BT$_{11}$ and BT$_{13.6}$ - BT$_{11}$ to retrieve inversion strength due to the larger surface contribution at 13.3 and 13.6 μm than that at 7.3 μm (Liu and Key 2003).

a. Method

The inversion strength retrieval algorithm described by LK03 is based on collocated radiosonde and satellite data, and a similar approach is used in this study. Radiosonde data from 61 stations in the Arctic and HIRS brightness temperature data from 1980 though 1996 are used to construct the dataset of matched clear-sky pairs.

![Fig. 1. Locations of the weather stations used in this study. Stations shown as squares were used for validation.](image-url)
of the HIRS pixels closest in time and space to radiosonde observations. HIRS spots must be within 75 km and one hour of the radiosonde observation to be used. The clear-sky determination is based on the cloud detection steps from the three algorithm. Figure 3 shows the relationship between inversion strength from radiosonde data and BT$\_{7.3}/H_{11002}$BT$\_{11}$ for the cases in all seasons and for the cold season, defined as November through March. When BT$\_{7.3}/H_{11002}$BT$\_{11}$ is larger than –10 K in all seasons (Fig. 3a), the inversion strength is linearly related to BT$\_{7.3} - BT\_{11}$. When BT$\_{7.3} - BT\_{11}$ is less than –10 K, there is no linear relationship. During the cold season (Fig. 3b), most cases have BT$\_{7.3} - BT\_{11}$ larger than –10 K, which exhibits a good linear relationship between inversion strength and BT$\_{7.3} - BT\_{11}$.

The cases with BT$\_{7.3} - BT\_{11}$ less than –10 K occur primarily in the warmer months, possibly because of the larger variability in water vapor amount and vertical distribution, and weaker inversion strength during that part of the year. Figure 4 shows the average specific humidity profiles in the cold and warm seasons. In the cold season, the water vapor content of the atmospheric column is low, so that the peak of the 7.3-μm weighting function is near the inversion top. During the warm season, the larger water vapor content raises the peak of the weighting function such that the channel is less sensitive to changes in inversion strength. It is for these reasons that we focus here on the inversion strength during the colder months.

Inversion strength in the cold season can be estimated by a linear combination of BT$\_{11}$, BT$\_{7.3} - BT\_{11}$, and (BT$\_{7.3} - BT\_{11}$)$^2$, with the coefficients determined by linear regression when BT$\_{7.3} - BT\_{11}$ is larger than –10 K. The equation used to retrieve inversion strength is

$$\text{INVST} = a_0 + a_1(BT_{11}) + a_2(BT_{7.3} - BT_{11})$$

$$+ a_3(BT_{7.3} - BT_{11})(\sec \theta - 1),$$

where the coefficients $a_0$, $a_1$, $a_2$, and $a_3$ are determined through multiple regression with INVST from radiosonde data, and $\theta$ is the sensor view angle. When BT$\_{7.3} - BT\_{11}$ is less than –30 K, the estimated inversion strength is defined to be 0 K. The BTs from two channels are used to derive the inversion strength, so we refer to this as the two-channel method.

![Figure 2](image2.png)

**Fig. 2.** Weighting function for the 6.7-, 7.3-, 11-, 13.3-, and 13.6-μm channels based on a subarctic winter standard atmosphere profile.

![Figure 3](image3.png)

**Fig. 3.** Relationships between inversion strength from radiosonde data and BT$\_{7.3} - BT\_{11}$ for (a) all seasons and (b) for the cold season. Cases with inversion strength less than 0 K are not considered to be inversions.
In the cold season there are a few cases with $BT_{7.3} - BT_{11}$ between $-30$ and $-10$ K, where the inversion strength is $0$ K for some situations and larger than $0$ K for others. The estimated inversion strength is therefore defined to linearly increase from $0$ to $2$ K when $BT_{7.3} - BT_{11}$ increases from $-30$ to $-10$ K. Radio-sonde data in all the matched cases are low-elevation meteorological stations, so the retrieval equation is most applicable to low-elevation areas.

In LK03, inversion strength regression equations for both high-elevation surfaces (higher than $2800$ m) and low-elevation surfaces (lower than $500$ m) were derived. In this work, only equations for low-elevation surface were derived. If the MODIS equation for a low-elevation surface is used to derive inversion strength over a high-elevation surface, the retrieved inversion strength will have a lower-than-actual value when the retrieved value is less than $17$ K, and a higher-than-actual value when the retrieved value is larger than $17$ K. If the same relationship applies to the HIRS retrievals, then estimated inversion strength over Greenland (high elevation) will be lower than the actual inversion strength, because the monthly mean retrieved inversion strengths over Greenland are between $10$ K and $17$ K. The negative bias of the retrieved monthly mean inversion strength over Greenland is between $-2.5$ K and $-1.0$ K. The retrieval bias for areas with surface elevations between $500$ m and $2800$ m is also negative, but smaller than that over Greenland. For these reasons, inversion strength retrievals over high-elevation surfaces are not included in this analysis (white areas in Figs. 7, 8, 9, and 10).

The 17-yr HIRS record comes from six different satellites: NOAA-6, -7, -9, -10, -11, and -12. Intersatellite biases result from orbital drift, sensor degradation, differences in spectral response functions, changes in calibration procedure, and deficiencies in the calibration procedure for each satellite. Orbital drift results in sampling at somewhat different times of the day at the beginning and end of a satellite’s lifetime, which would have the most pronounced impact on retrievals where the diurnal cycle is significant. The diurnal cycle of inversion strength in the Arctic during the cold season is, however, small. More than 90% of the monthly mean inversion strengths from 26 Arctic stations during the cold season at 0000 UTC are within $2.0$ K of the monthly mean at 1200 UTC. Because the diurnal cycle of the inversion strength is small, we do not expect orbital drift to significantly affect the inversion strength trends.

TOVS intersatellite calibration is essential for climate change studies. The level 1b data were calibrated using the standard procedure described in the NOAA Polar Orbiter user’s guide (Kidwell 1998). In addition, empirical corrections were applied to the calibrated brightness temperatures in the TOVS Path-P product generation. These corrections (also known as DELTAs) were applied to $NOAA-10$, -11, and -12, and the deltas for $NOAA-10$ were applied to $NOAA-6$ through -9 (Schweiger et al. 2002; Chen et al. 2002). However, some intersatellite differences remain after these calibration steps (Chen et al. 2002). To minimize intersatellite calibration differences, we determine separate sets of inversion strength regression coefficients for each satellite or pair of satellites from 1980–96: one for $NOAA-6$ (1980–82), one for $NOAA-7$ (1983–84) and $NOAA-9$ (1985–86), one for $NOAA-10$ (1987–16 September 1991), and one for $NOAA-11$ (17 September 1991–1994) and $NOAA-12$ (1995–96). Four sets of regression coefficients are therefore derived.
are over 18,000 potential radiosonde–satellite matched pairs for each year over the 61 stations, the actual numbers of matched pairs that can be used to derive each retrieval equation are 223, 597, 404, and 441, respectively, due to cloudiness and space and time distance constraints.

After obtaining the retrieval equation coefficients for each satellite or pair of satellites from 1980 through 1996, the initial HIRS clear-sky BTs are converted to inversion strength based on the inversion strength retrieval equations. The estimated inversion strength is then mapped to the 100 × 100 km Equal-Area Scalable Earth Grid (EASE-Grid) based on the longitude and latitude of the original TOVS data. For each day, a composite map of clear-sky inversion strength over the region north of 60°N is created. The monthly mean inversion strength for each grid point is calculated as the mean of all the daily composite inversion strengths in a month, which include inversion strength values of zero. The seasonal mean inversion strength is derived as the mean of monthly means in that season. The monthly trend of inversion strength for each grid point is derived using linear least squares regression based on the monthly mean inversion strength for the 17-yr period 1980–96. The statistical significance level is determined with an F test. It is important to note that the number of daily retrievals in each month is relatively evenly distributed across the first, second, and third 10-day periods from 1980 to 1996. This is important because trends in cloud amount could, in theory, result in nonuniform sampling of clear-sky pixels in different months and years, and artificially introduce trends in clear-sky inversion strength.

b. Validation

To evaluate the regression equations, cross validation is performed. For each of 100 iterations, 70% of all the radiosonde and satellite data matched pairs for a satellite or satellite pair were randomly selected to determine the regression coefficients. The new equation was then applied to the remaining 30% of the matched pairs to determine a bias and root-mean-square-error (RMSE) of the inversion strength retrieval. The mean bias and RMSE were then calculated. The magnitude of the mean biases for each satellite were less than 0.1 K, and the mean RMSE were 2.7, 2.8, 2.8, and 2.7 K, respectively, for the different satellite or satellite pairs.

The radiosonde data from drifting ice stations ended at 1990, so these data were not used to derive the coefficients of the retrieval equations. However, these data were used to show that the regression equations derived from station data can be applied over other areas. The radiosonde data from drifting stations were matched with satellite data. The regression equations were then applied to the satellite data to derive the two-channel inversion strength, and compared to the station-based inversion strength to calculate the bias and RMSE. A total of 223 radiosonde and satellite data matched pairs were collected from the drifting stations from 1980 to 1990. The mean bias and RMSE were 0.25 and 2.5 K.

Radiosonde data at 11 stations (shown as squares in Fig. 1) were also used to validate the inversion strength monthly means and trends from the two-channel method. For this validation exercise the radiosonde data from the 11 stations were not used to derive the regression equation coefficients. Each of these 11 stations has at least 50% of all possible 0000 and 1200 UTC soundings in each month. The inversion strength monthly means and trends from radiosonde data, as well as the two-channel retrievals at these 11 stations are compared in Figs. 5 and 6. The monthly mean is the mean of all the individual means for that month over the period 1980–96, and the monthly trend is derived using linear least squares regression based on the monthly means. Considering the different sky conditions, the monthly means of two-channel inversion strength agree with those from station data very well in the cold season, with a bias around 0.3 K. Trends from the two-channel method agree with the trends from station data reasonably well (Fig. 6).

In the remainder of the paper, the coefficients of the equations used to calculate the monthly mean and trend of the inversion strength were derived based on all the available radiosonde and satellite data matched pairs.

4. Climatology of inversion strength

a. Spatial characteristics

Figure 7 shows the spatial distribution of monthly mean, clear-sky two-channel inversion strength in the Arctic for November–March, and for winter (December–February; DJF) averaged over the period 1980–96. The spatial distributions have similar patterns from November to March, but with different magnitudes. The lowest monthly mean two-channel inversion strength is over Greenland, Iceland, and Norwegian (GIN) Seas, Barents Sea, and northern Europe, which is attributable to the turbulent mixing over open water and high cloud cover in this region during the winter time (Serreze et al. 1992). The inversion strength increases eastward, followed by a decrease over Alaska. Over land, the strong monthly mean inversion strength can be seen over northern Russia and northern Canada, with the largest values near several Russian river valleys owing...
to strong radiative cooling under clear conditions. Over the Arctic Ocean, the largest values are over the pack ice north of Greenland and the Canadian Archipelago; the lowest values are in the Kara, Laptev, and Chukchi Seas. Temporally, the monthly mean inversion strength over the ocean is largest in February (~16 K) and smallest in November (~12 K) over the pack ice north of Canada. Similarly, the strongest inversions over land
occur in January and February, and the weakest in November and March.

b. Trends

The monthly trend of the clear-sky inversion strength is shown in Fig. 7. The spatial distribution of monthly trends is similar in December, January, and February. The winter average trend shows decreases in inversion strength over the Chukchi Seas, with an average around \(-0.13\) K yr\(^{-1}\). The inversion strength also decreases over northern Europe with average rate around \(-0.13\) K yr\(^{-1}\). Inversion strength increases over north central Russia at rate around \(0.10\) K yr\(^{-1}\), and increases in northeastern Russia, and also between Severaya Zemlya and North Pole at the rate of \(0.13\) K yr\(^{-1}\). All the changes are statistically significant at the 90% or higher confidence level based on the F test. In March, the inversion strength decreases significantly over the Laptev, Chukchi, and Beaufort Seas at a rate over \(-0.10\) K yr\(^{-1}\); and decreases in regions surrounding

Fig. 7. (top) Monthly mean clear-sky inversion strength (K), and (bottom) monthly trend of clear-sky inversion strength (K yr\(^{-1}\)) in November–March, and winter (DJF), 1980–96, using the two-channel statistical method. A trend with a confidence level larger than 90% based on the F test is indicated with +.
Novaya Zemlya and north central Russia at a rate over −0.10 K yr$^{-1}$. The decreasing rate over northern Europe is more than −0.05 K yr$^{-1}$. In November, there has been a significant decrease in inversion strength over the Chukchi and Beaufort Seas at a rate over −0.10 K yr$^{-1}$.

Kahl and Martinez (1996) found significant increases in inversion strength over the Arctic Ocean during winter and autumn from 1950 through 1990. Based on their Fig. 6, the inversion strength increase occurred primarily between 1950 and the late 1970s. After 1980, they did not find any significant increase of inversion strength. That result is consistent with the present overall (Fig. 7), where much of the Arctic Ocean shows no trend. However, the inversion strength over some areas exhibits a strengthening trend and others show a weakening trend from 1980 through 1996.

c. Discussion

Given the strong coupling of surface temperature and inversion strength by means of radiative cooling, trends in surface temperature should be correlated with trends in inversion strength. Figure 8 shows the monthly surface skin temperature trend based on the TOVS Path-P surface skin temperature retrievals in the cold season, averaged over the period 1980–96. In the cold season, areas with decreasing trends in inversion strength are generally those areas that show increasing surface skin temperature trends (e.g., northern Europe in winter). Similarly, areas with increasing inversion strength trends are generally areas with decreasing surface skin temperature trends (e.g., north central Russia in winter).

The correlation coefficient between the monthly mean surface skin temperature anomalies and monthly mean two-channel inversion strength anomalies for November to March over the period 1980 to 1996 is shown in Fig. 9. The negative correlation coefficients are less than −0.6 over northern Europe, north central Russia, Alaska, and part of northeastern Russia, which means the inversion strength trend over these regions is closely related to the surface skin temperature trend. This is, in fact, the case for most of the Arctic. However, the correlation coefficient is near zero from the Canadian Archipelago across the central Arctic Ocean and through the East Siberian Sea into Siberia. For most of this area the surface temperature trend is near zero (Fig. 8), but some portions exhibit statistically significant positive or negative inversion strength trends (Fig. 7). It is possible that in these areas the trend in inversion strength trend may be more a function of changes in heat advection into or out of the Arctic than changes in surface temperature. For example, the inversion strengths decrease over the East Siberian Sea, but the surface skin temperature shows little or no trend. The weakening of the inversion in that area may result from cold air advection and a decrease in the temperature of the atmosphere aloft, which effectively decrease the inversion strength.

The relationship between changes in inversion
strength, surface temperature, and large-scale circulation are illustrated in Fig. 10, which shows the correlation between the cold season monthly mean anomalies of surface skin temperature and the Arctic Oscillation (AO) index, and between the inversion strength anomalies and the AO index. The correlation between the surface temperature and AO anomalies is positive in northern Europe and northern Russia but negative over the Canadian Archipelago and Alaska. This is very similar to the results given by Wang and Key (2003). The inversion strength and AO index correlation is negative over northern Europe, north central Russia and the East Siberian Sea, and positive over the Canadian Archipelago. Over the East Siberian Sea the correlation between the AO and surface temperature is near zero or negative, but the correlation coefficient between the AO and inversion strength is significantly negative. As described above, this implies a change in the temperature of the atmosphere aloft as a result of changes in large-scale circulation.

5. Summary

A 17-yr time series of clear-sky temperature inversion strength for the months of November through March in the Arctic is derived from HIRS data using a two-channel empirical method. Employing a different set of regression coefficients for each satellite or pair of satellites from 1980 through 1996 alleviates any inter-satellite calibration problems. Cross validation with test samples and the use of independent drifting station radiosonde data show the applicability of the retrieval equations across the Arctic.

Fig. 9. Correlation coefficient between the monthly mean surface skin temperature anomalies and the monthly mean two-channel statistical inversion strength anomalies over the period 1980–96. A correlation coefficient with a confidence level larger than 95% based on the F test is indicated with +.
For the two-channel monthly mean inversion strength, the weakest temperature inversions occur over the Norwegian and Barents Seas and over northern Europe. Inversions are strongest over the pack ice and in several river valleys in the Eurasian Arctic. Over the ocean, inversions are strongest in February and weakest in November. Over land, inversions are strongest in January and weakest in March. There is a significant decreasing trend in inversion strength over the East Siberian and Chukchi Seas. Over north central Russia, there is an increasing trend in winter, but a decreasing trend in March. Over Alaska there is a significant increasing trend in November and March.

An analysis of the correlation between surface temperature and inversion strength trends, and between these two parameters and the Arctic Oscillation index, demonstrates the strong coupling between changes in surface temperature and changes in inversion strength. This is not surprising given that the primary control over surface-based inversions in the polar regions is radiation cooling. However, the analysis revealed that in some areas, trends in inversion strength are poorly correlated with trends in surface temperature, but more highly correlated with changes in large-scale circulation. Changes in inversion strength in areas such as the East Siberian Sea, for example, may therefore be a result of warm or cold air advection aloft rather than warming or cooling at the surface.

Acknowledgments. The authors wish to thank the anonymous reviewers for their constructive comments and suggestions. This research was supported by NOAA and NSF Grants OPP-0240827 and OPP-0230317. The views, opinions, and findings contained in this report are those of the authors and should not be construed as an official National Oceanic and Atmospheric Administration or U.S. government position, policy, or decision.

REFERENCES


Fig. 10. Correlation between the monthly AO index anomalies and (left) the monthly mean surface skin temperature anomalies, and (right) monthly mean two-channel statistical inversion strength anomalies (right), 1980–96. A correlation coefficient with a confidence level larger than 95% based on the F test is indicated with +.


Stubenrauch, C. J., A. Chedin, R. Armante, and N. A. Scott, 1999: Clouds as seen by satellite sounders (3I) and imagers (ISCCP). Part II: A new approach for cloud parameter determination in the 3I algorithms. J. Climate, 12, 2214–2223.

