NOTES AND CORRESPONDENCE

The Minor Stratospheric Warming of January 1989: Results from STRATAN, a Stratospheric–Tropospheric Data Assimilation System

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ABSTRACT

Using a stratospheric–tropospheric data assimilation system, referred to as STRATAN, a minor sudden stratospheric warming that occurred in January 1989 is investigated. The event had a maximum influence on the stratospheric circulation near 2 hPa. The zonal mean circulation reversed briefly in the polar region as the temperature increased 34 K in 3 days. The cause of the warming is shown to be the rapid development and subsequent movement of a warm anomaly, which initially developed in the midlatitudes. The development of the warm anomaly is caused by adiabatic descent, and the dissipation by radiative cooling. A brief comparison with the NMC analysis and temperature sounding data is also presented.

1. Introduction

In this paper we use an atmospheric analysis that is based on standard meteorological data assimilation techniques (Takano et al. 1987, Baker et al. 1987) to discuss the evolution of an intense minor warming in January 1989. This system, called STRATAN, uses a model as an analysis tool to provide a first-guess field for an objective analysis scheme. The scheme is a two-dimensional, multivariate, statistical objective analysis scheme (referred to as optimum interpolation or OI) that makes geostrophic adjustments to the wind field based on the temperature data. Wind measurements are used when available. The OI uses 18 vertical levels with the top at 0.4 hPa. A 19-level fourth-order quadratically conservative stratospheric–tropospheric general circulation model (Kalney et al. 1983) with a horizontal resolution of 4° latitude by 5° longitude and a top at 0.3 hPa is then run using the OI as initial conditions. The GCM is run for 6 h and adjusts the ageostrophic component of the wind field. The result is then used as the first-guess field for the new OI, which includes any current meteorological data. This assimilation process produces an analysis of temperatures, geopotential heights, winds (including vertical motions), and diabatic heating terms. Data sources for STRATAN include: rawinsondes, surface stations, satellites, buoys, rocketsondes, dropwindsondes, balloons, ships, and aircraft. STRATAN produces a meteorological dataset that is physically consistent and has a high density in both space and time. STRATAN data should, therefore, provide a better quality analysis to study stratospheric dynamics than conventional geopotential height data analysis or GCM model results that are initialized with actual data.

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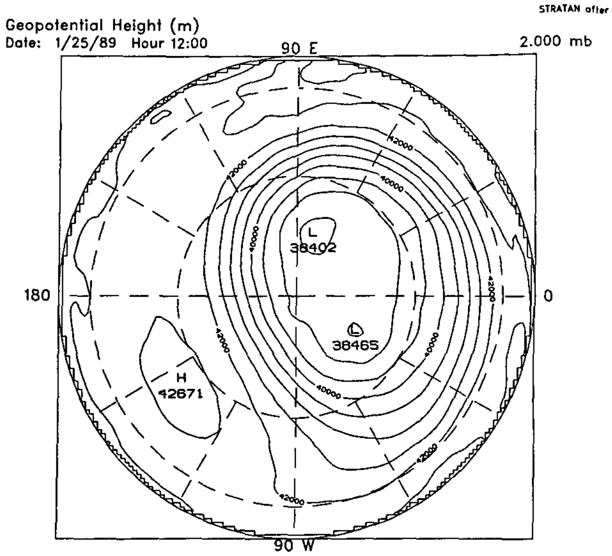
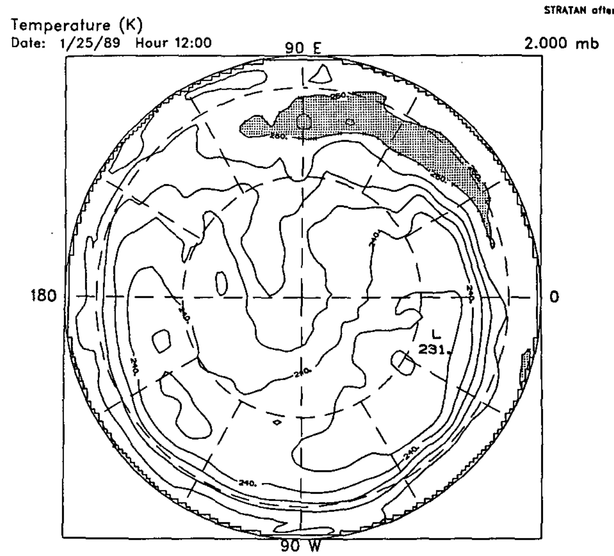




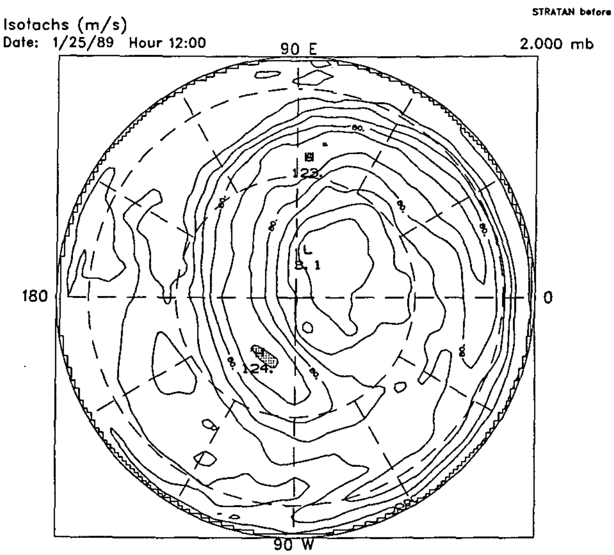


FIG. 1. Northern Hemispheric orthographic projection of STRATAN fields on 2-hPa surface for 1200 UTC 25 January 1989. (a) Geopotential heights (contour interval is 500 m), (b) temperature (above 260 K is shaded and the contour interval is 5 K), and (c) isotachs (above 120 m s^{-1} is shaded and the contour interval is 20 m s^{-1}). Greenwich meridian intersects right side of frame at the equator.

2. Description

During late January 1989 there was a brief intense minor warming in the Northern Hemispheric upper stratosphere. The event was evident in the polar temperature and zonally averaged zonal wind field for only 3–4 days. The temperature at the north pole at 2 hPa rose 27 K in 2 days, from 251 K on 27 January to 278 K on 29 January. There was an associated brief reversal of the zonally averaged zonal wind above 2 hPa and north of 70°N on 29 January. A subsequent rapid cooling, to 257 K on 30 January, followed. The polar vortex during January 1989 was quite large and stable, with the coldest temperatures on record in the lower stratosphere (Nagatani et al. 1990).

The geopotential height field for 2 hPa on 25 January (Fig. 1a) shows the polar vortex displaced off the pole, roughly toward 30°E. Also of note are an anticyclonic circulation located near 56°N, 145°W and a small developing anticyclone near 34°N, 95°E. The corresponding temperature field at 2 hPa (Fig. 1b) indicates the development of a warm anomaly near 44°N, 90°E. The anomaly contains the warmest temperatures in the Northern Hemisphere. North of the anomaly is a tongue of warm temperatures across the pole, roughly oriented along the streamlines. An examination of the isotach analysis from the primitive equations (Fig. 1c) indicates that there is a large area of wind speeds greater than 100 m s^{-1} at 2 hPa. There are two local maxima in the jet that have speeds greater than 120 m s^{-1}. These maxima are located between the polar vortex and the centers of the anticyclones.

The anticyclonic circulation that was located near 34°N, 95°E on 25 January intensified on 26 January

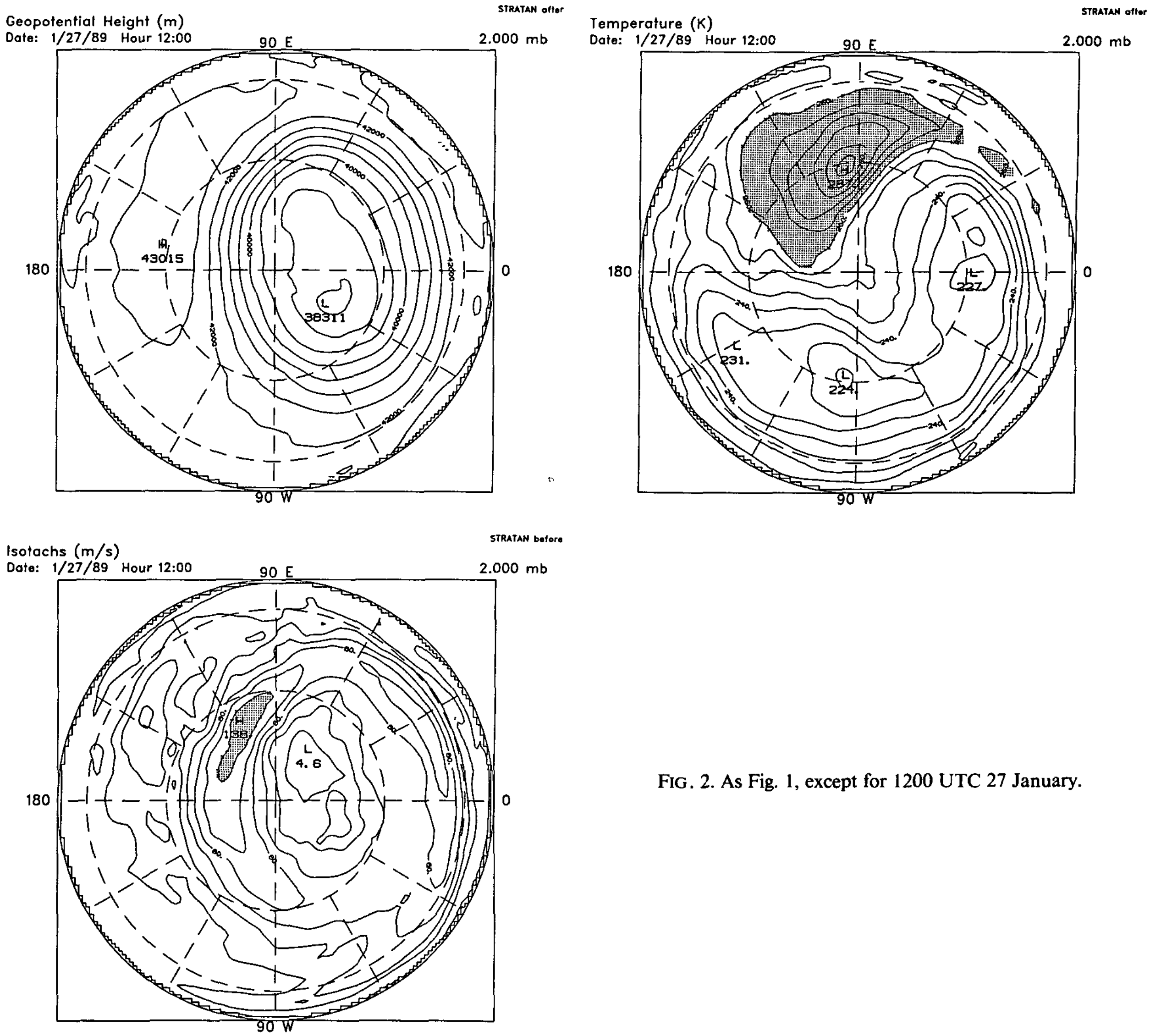


FIG. 2. As Fig. 1, except for 1200 UTC 27 January.

and moved to near 46°N, 120°E, as the other anticyclonic circulation weakened. The warm anomaly increased significantly on 26 January, to a maximum temperature of 275 K. In response to the anticyclonic development, the maximum wind speed in the jet stream increased to 132 m s^{-1}. On 27 January, the geopotential height field at 2 hPa (Fig. 2a) still showed a large polar vortex displaced from the pole. The developing anticyclone continued to intensify, with the central heights increasing approximately 300 m. The decaying anticyclone appears to retrograde and combine with the developing anticyclone between 26 and 27 January. The temperature analysis (Fig. 2b) shows that the warm anomaly continued its explosive development, with the maximum temperature now at 287 K, as it migrated northward. The isotach analysis (Fig. 2c) shows that the jet streak between the vortex and the anticyclone intensified to 138 m s^{-1}.

The anticyclone continued to develop through 28 January, becoming more circular in shape. The 29 January geopotential height analysis at 2 hPa (Fig. 3a) shows the anticyclone centered east of the date line. By 30 January the polar vortex had become more circular, and the anticyclone weaker and elongated. The warm anomaly maintained its strength and size between 27 and 28 January, but by 29 January (Fig. 3b) the anomaly had begun to decay, as the maximum temperature decreased to 278 K, and migrated to the pole. The decay by 30 January was dramatic, as the maximum temperature decreased to 260 K. The isotach analysis showed that the jet maximum on 28 January was 148 m s^{-1}, and the jet axis continued to move

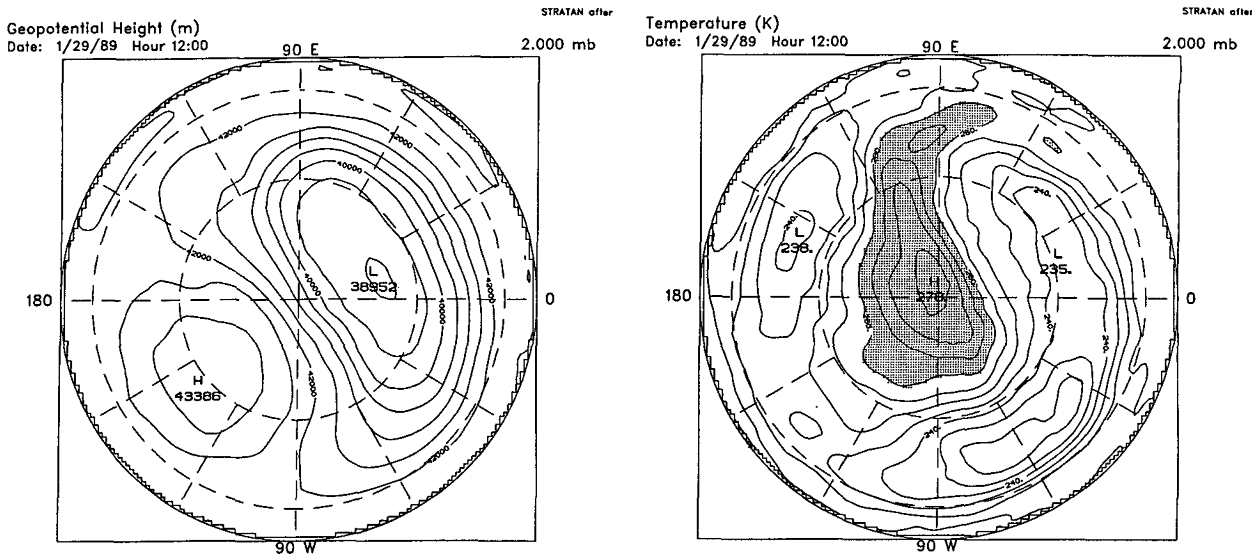


FIG. 3. As Fig. 1, except for 1200 UTC 29 January with no isotach analysis.

poleward. By 29 January the maximum speed had dropped to 139 m s^{-1}, and by 30 January it had dropped to 125 m s^{-1}.

It is interesting to note the proximity of the center of the warm anomaly to the maximum in the jet. The center of the warm anomaly is always positioned at the entrance region of jet maximum. Fairlie et al. (1990), in a modeling study of the major warming during the winter of 1984/85, identified the entrance region of the jet as an area of strong descent and therefore adiabatic warming. One of the advantages of using the STRATAN data is that the vertical velocity field is produced at the time of analysis. The anomaly has been defined to be the region delineated by the 260-K contour on the 2-hPa surface, the shaded region in Figs. 1b, 2b, and 3b. This criterion was chosen because just preceding this event there were no temperatures this far north of the tropical regions. From Figs. 1b, 2b, and 3b, it is easy to see the explosive development and movement of the warm anomaly. Spatially averaged vertical motions within the anomaly for 1200 UTC on each day are listed in Table 1. If these motions are representative of the entire day, the adiabatic warming of the air subject to these average vertical motions for 24 h is large enough to account for the observed increase in temperature. The vertical motions also imply extensive vertical transport at this level, with the average vertical motion during the peak development indicating a displacement of approximately 2 km day^{-1}.

A horizontal cross section of the vertical motion at 2 hPa on 27 January (Fig. 4) showed that downward motion does not exist everywhere within the warm anomaly. The downward motion (shaded) is concentrated on the upstream side of the anomaly, with upward motion on the downstream side. Figure 4 also indicates that the downward motion extends far upstream, and upward motion far downstream, from the anomaly. The strongest downward motion and nearly strongest upward motion are adjacent to the center of the anomaly. The adiabatic temperature changes associated with this pattern imply that the maximum temperature increase is upstream of the anomaly and that the maximum temperature decrease is on the downstream side.

The rapid decrease in the maximum temperature of the warm anomaly, 18 K between 29 and 30 January, is close to the longwave cooling rate calculated by the STRATAN radiation package (Rosenfield et al. 1987). The calculated longwave radiational cooling rate for 29 January (Fig. 5) indicates that there were radiational

TABLE 1. Area greater than 260 K averaged vertical motions and implied vertical displacement and adiabatic warming from motion.

Date	T(max) (K)	Area > 260 K (10^{13} m^2)	Vertical velocity (nb s^{-1})	$\Delta(Z)$ (m)	$\Delta(T)_{calc}$ (K)	$\Delta(\max T)_{real}$ (K)
24	264	0.3	0.9	−306	3.0	3
25	266	1.0	3.7	−1216	11.9	9
26	275	1.9	4.0	−1327	12.9	11
27	288	2.4	7.3	−2449	23.9	1
28	287	2.5	7.2	−2464	24.1	−10
29	277	2.0	2.5	−852	8.3	−16

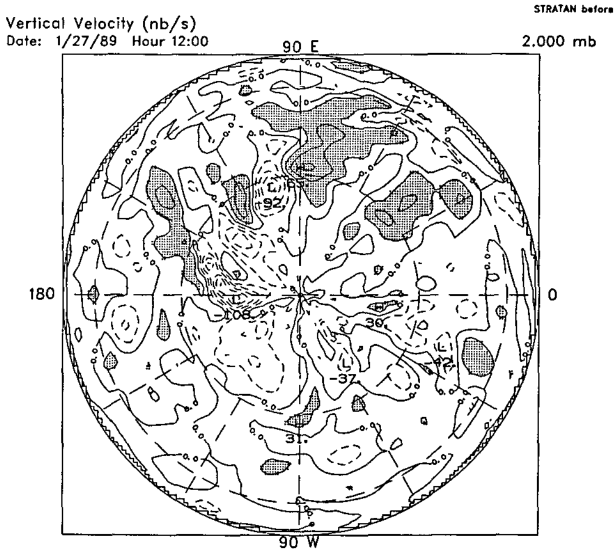


FIG. 4. Northern Hemispheric orthographic projection of vertical velocity (nb s^{-1}) on 2-hPa surface for 27 January 1989. Above 20 nb s^{-1} is shaded and the contour interval is 20 nb s^{-1}.

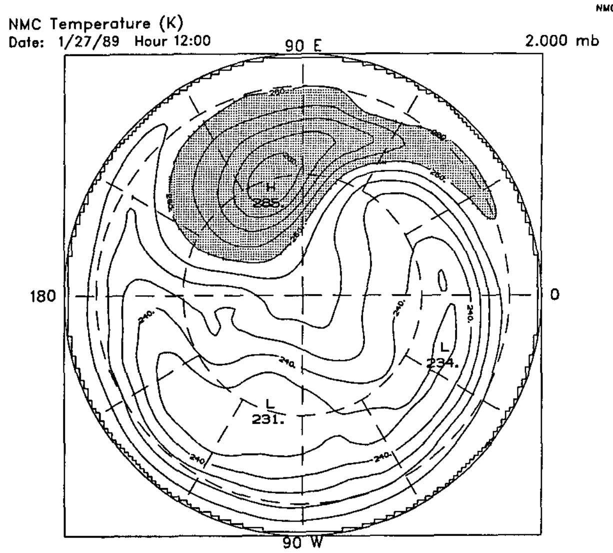


FIG. 6. Northern Hemispheric orthographic projection of NMC temperature analysis on 2-hPa surface for 27 January 1989. Above 260 K is shaded and the contour interval is 5 K.

cooling rates of up to 13 K day^{-1} in the center of the warm anomaly. The vertical motions after 28 January become less organized as the strong upward and downward motion centers associated with the anomaly dissipate.

3. Comparison

In this section brief comparisons are made of the STRATAN data to the National Meteorological Center (NMC) daily analysis and satellite vertical temperature profile data during the warming. The STRATAN data has the advantage of being of finer time resolution than the NMC data, which is available only once per day. STRATAN also produces a physically consistent three-dimensional wind field, whereas the NMC data has only geopotential height data available. For study of this event, the finer time resolution was useful since the event appeared and disappeared in only a few days.

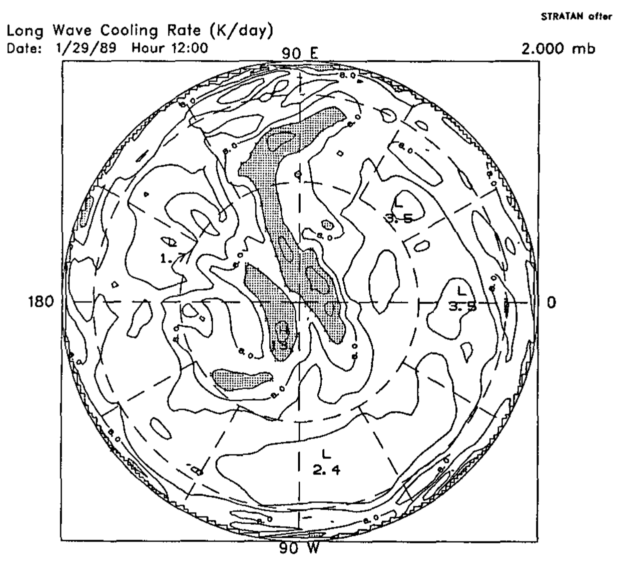


FIG. 5. Northern Hemispheric orthographic projection of longwave cooling rate on 2-hPa surface for 29 January 1989. Above 10 K day^{-1} is shaded and the contour interval is 2 K day^{-1}.

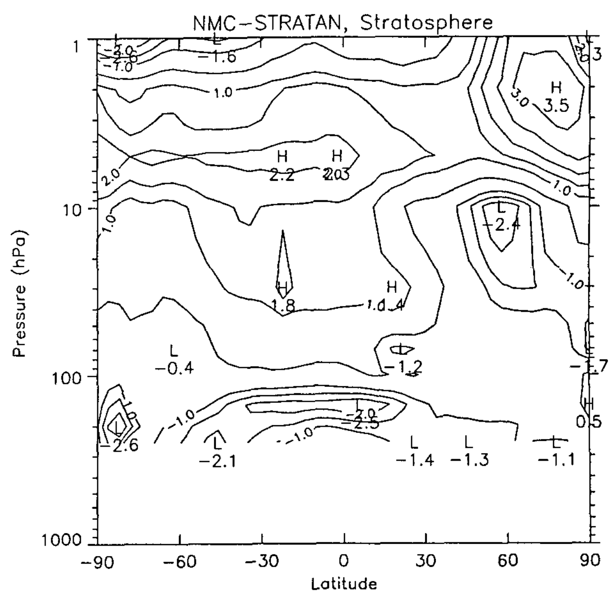
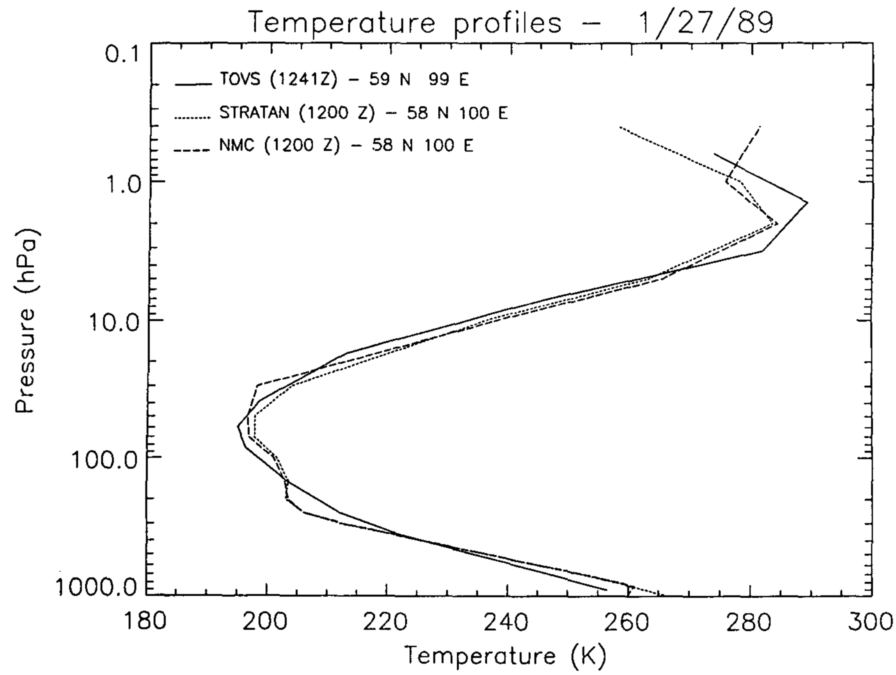


FIG. 7. Difference of NMC analysis and STRATAN analysis of January 1989 average zonal mean temperatures (K) in the stratosphere.



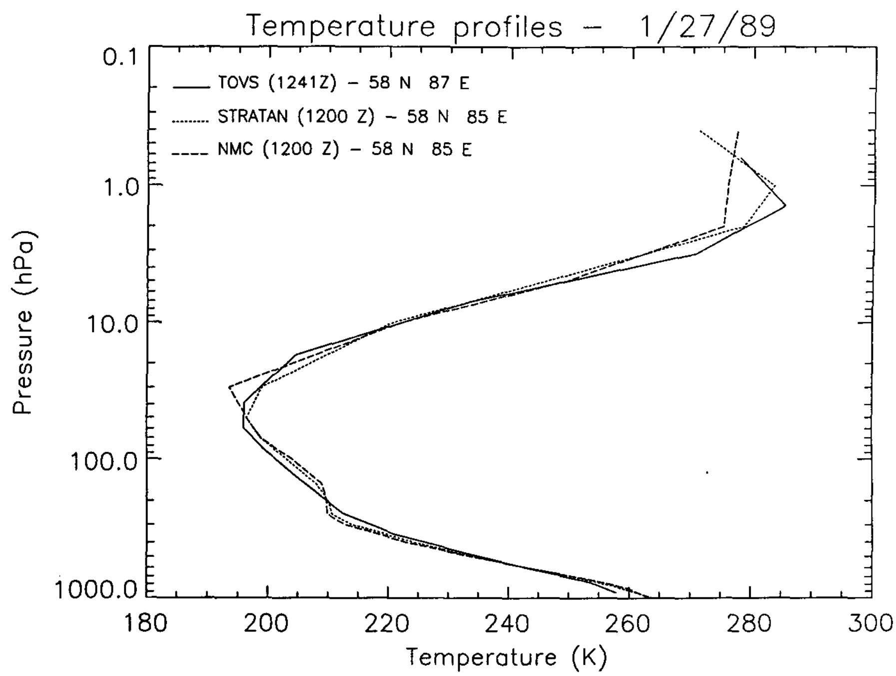


FIG. 8. Vertical temperature profiles (hPa vs K) from TOVS satellite (solid), STRATAN (short dash), and NMC analysis (long dash) for (a) 27 January at 59°N, 99°E, (b) 27 January at 58°N, 87°E, and (c) 29 January at 85°N, 143°E. Times and location of the STRATAN and NMC profiles are indicated in the legend.

Data at 6-h intervals indicated that the rapid temperature increase was uniform throughout the 2-day period, and helped in tracking the development of the anticyclone.

The differences between STRATAN and NMC temperature analyses are in general small. When the two analyses are compared with the high-resolution Meteorological Measurement System (MMS) data taken on ER-2 in the lower stratosphere over the North Atlantic and Scandinavia, STRATAN agrees more closely (Rood et al. 1990). The improvement over the NMC analysis is small and is most apparent at low

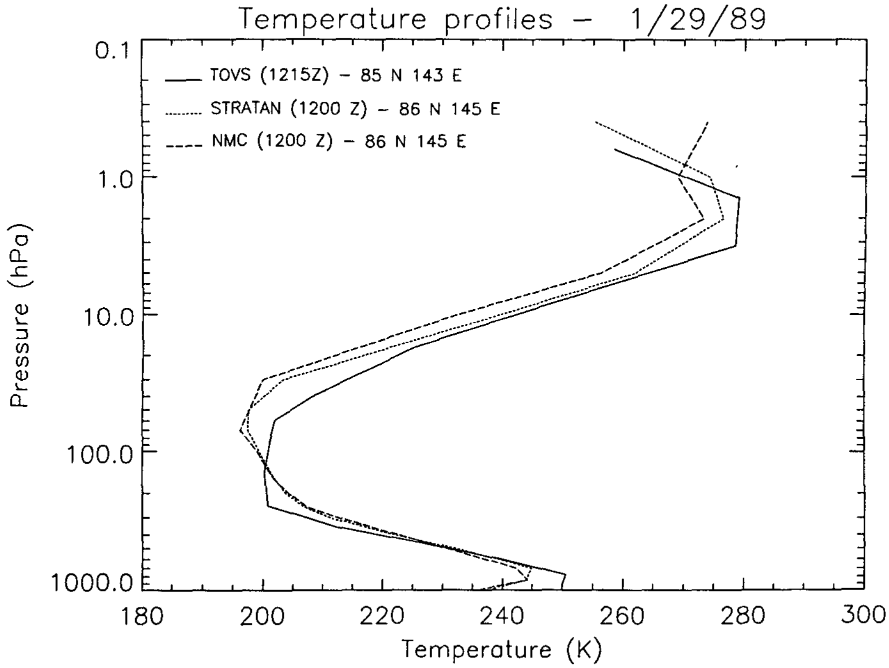


FIG. 8. (*Continued*)

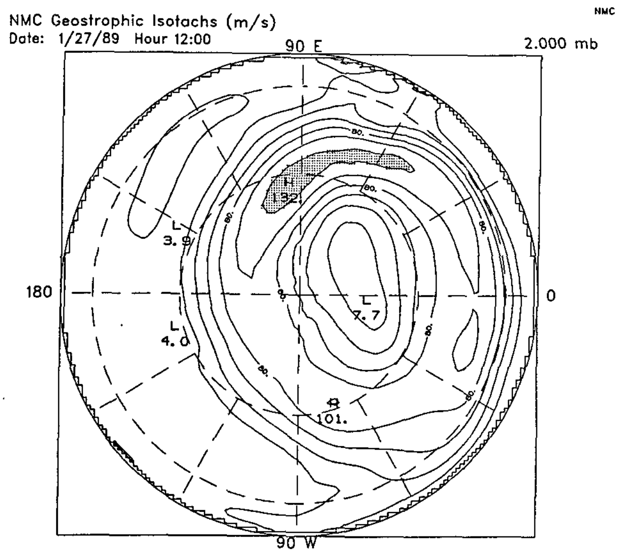
temperatures. The major differences between the analyses in the upper stratosphere are the physical constraints provided by the assimilation procedure, since both analyses use the *TIROS-N* Observational Vertical Sounder (TOVS) satellite data and the National Environmental Satellite Data and Information Service (NESDIS) retrievals.

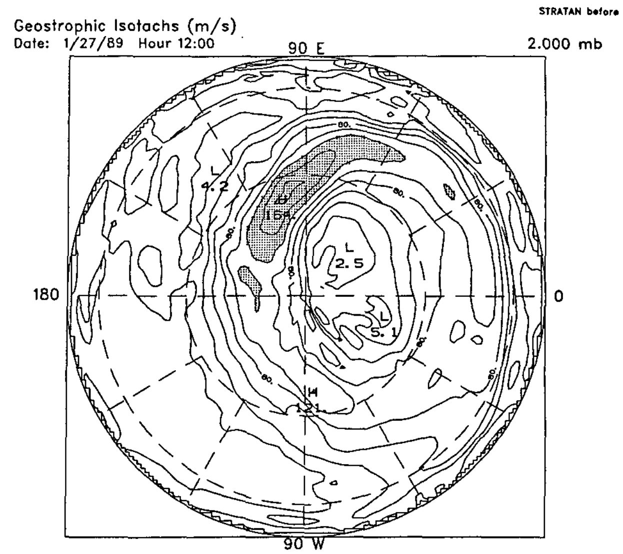
As the warming began to develop on 25 January, the two temperature analyses agree to within a couple of kelvins over most of the Northern Hemisphere at 2 hPa. A comparison of the NMC (Fig. 6) and STRATAN (Fig. 2b) temperature analyses for 27 January indicates that the warm anomaly is depicted similarly, with STRATAN being slightly warmer. Throughout the warming, the two analyses represent the warm anomaly with similar characteristics. STRATAN is noticeably colder in the cold belt at 60°N between 0° and 120°W. This difference is the most systematic bias between the analyses. The horizontal structure is also much smoother in the NMC analysis.

The difference between the NMC and STRATAN zonal mean temperature fields averaged for the month of January (Fig. 7) shows a systematic bias at the tropical tropopause (see Rood et al. 1990, for a discussion of tropopause problems). The NMC analysis is warmer in the upper stratosphere of the tropics and Southern Hemisphere. In the Northern Hemisphere, the NMC analysis is colder below 10 hPa and warmer above. The assimilation procedure used in STRATAN incorporates a general circulation model so that vertical continuity is maintained through the representation of the Rossby waves. Therefore, systematic differences between the height fields and the temperature might be present.

Comparisons of vertical temperature profile data from the TOVS to profiles extracted from STRATAN near the center of the warm anomaly show good agreement (NMC profiles are also shown for comparison). The TOVS profiles indicate that within the warm anomaly there is a layer, between 1 and 5 hPa, that is warmer than STRATAN. The spatial distribution of differences within the warm anomaly indicates that the largest differences are clustered on the anticyclonic shear side of the jet. A comparison of profiles on the anticyclonic shear side of the jet (Fig. 8a) and cyclonic shear side of the jet (Fig. 8b) (see Fig. 6 for jet axis location) on 27 January demonstrates an apparent bias in the STRATAN. The profiles on the cyclonic side agree quite well, but on the anticyclonic side there is a cool bias in STRATAN above 5 hPa. As the warm anomaly moves north, the TOVS data become sparse. Only one TOVS profile could be found north of 85°N on 29 January (Fig. 8c), when the warm anomaly was centered on the North Pole. The TOVS profile (Fig. 8c), which is on the anticyclonic shear side of the jet, is again warmer than the STRATAN profile above 5 hPa, but STRATAN in this case is warmer above 1 hPa.

A comparison of STRATAN wind speeds to the geostrophic wind speed derived from NMC height fields (Fig. 9a) at 2 hPa on 27 January (compare to Fig. 2c) shows that the largest winds in the jet core are similarly represented in both magnitude and position (approximately 60°–70°N, 90°–120°E). Both upstream and





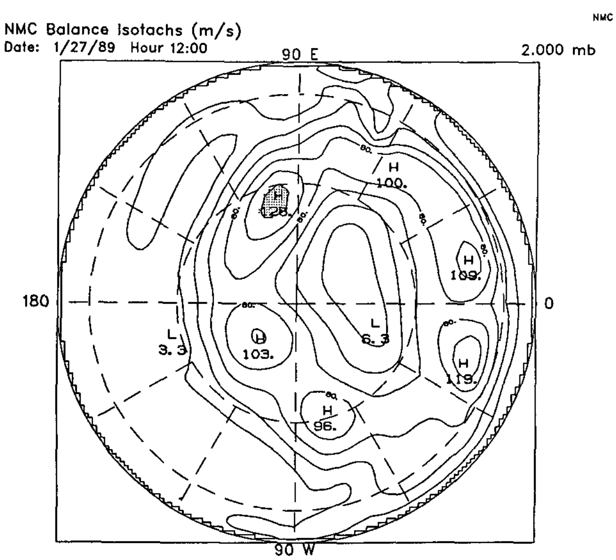


FIG. 9. Northern Hemispheric orthographic projection of fields on 2-hPa surface for 1200 UTC 27 January 1989. (a) NMC geostrophic isotachs, (b) STRATAN geostrophic isotachs, and (c) NMC balance isotachs. Above 120 m s^{-1} is shaded and the contour interval is 20 m s^{-1}.

downstream of the jet core, however, the geostrophic winds are significantly larger than the STRATAN winds. For instance, downstream, between 270° and 300°E at 60°N, the NMC geostrophic winds are approximately 100 m s^{-1}, while the STRATAN winds are closer to 80 m s^{-1}. Similarly, upstream of the jet core the geostrophic winds are greater than 120 m s^{-1} throughout a large region where the STRATAN winds are less than 100 m s^{-1}. In general, the NMC geostrophic winds are larger than the primitive equation winds from STRATAN in the inflow and outflow region of the jet, which is consistent with the results of Boville (1987) and Elson (1986).

In the case of NMC geostrophic winds (Fig. 9a), the absolute maximum wind in the jet core is about the same as that in STRATAN. More generally, the geostrophic winds are larger than the primitive equation winds in the core, not just in the inflow and outflow regions (see Rood et al. 1989, Fig. 1). STRATAN also produces a height analysis from which the geostrophic wind can be calculated (Fig. 9b). The maximum geostrophic wind speed in the STRATAN jet core is approximately 30 m s^{-1} larger than the NMC or STRATAN analysis. As mentioned above, systematic differences with respect to the NMC height and temperature fields develop above 10 hPa, and in this instance, the height gradients are significantly tighter in STRATAN. Three factors suggest that the assimilation height fields are a more accurate representation of the atmosphere: 1) the assimilation procedure provides physical continuity with the lower stratosphere, where the data are more dense and of higher quality; 2) investigation of the curvature effects in the comparison of geostrophic winds (e.g., Holton 1979, 62–64) to what are believed to be the actual winds (e.g., MMS, model experiments) makes more sense using the STRATAN height fields;

and 3) the quality of constituent transport calculations with STRATAN winds (Rood et al. 1989, 1990).

In the lower stratosphere, the nonlinear balanced winds (Randel 1987) and STRATAN are in very good agreement throughout the middle latitudes. There are, of course, major differences between STRATAN and all balanced wind estimates in the subtropics and tropics. In the upper stratosphere, straightforward calculations of nonlinear balanced winds consistently show a noisy structure that is not realistic (e.g., Fig. 9c). Therefore, while the nonlinear balanced winds may provide a good estimate of winds in certain space and time domains, some algorithm and data problems remain for upper stratospheric applications.

Finally, comparisons of STRATAN (Fig. 2b) with NMC (Fig. 6) temperatures and STRATAN (Fig. 2c) with NMC (Fig. 9a) winds show differences in the horizontal structure. As discussed in Rood et al. (1989, 1990), the assimilation procedure controls noise through the elimination of nonphysical processes. It also generates noise with the data insertion and the violation of balance between the momentum and thermal fields (e.g., gravity waves). Some of this generated noise is subsequently controlled by the GCM forecast. The assimilation procedure also has the capacity to more realistically represent the horizontal structure because the data is treated in a truly asynoptic manner, and the dynamical model can transmit information into regions where there are no data.

An assessment of the reality of the horizontal structure in the STRATAN field is a difficult task. At high altitudes, such as those reported here, there is undoubtedly small-scale structure from gravity waves that is not real. Furthermore, there are bull's-eye structures, most prevalent at the pole, which are generated during data insertion. On the other hand, some of the synoptic-scale structure and structure forced by the vertical continuity of the Rossby waves is likely to be real. A current assessment of the data quality is that below 5 hPa, the day-to-day fields from STRATAN are accurate. Above 5 hPa, the planetary- and synoptic-scale structure and the time-averaged fields are good. However, at these altitudes numerous small-scale short-lived features are always present in the analysis. These features are analysis artifacts.

4. Summary

An intense minor warming confined to the upper stratosphere has been described using analyses from the data assimilation system STRATAN. The warming develops with a burst of downward motion in the middle latitudes and then propagates poleward. The warming then dissipates by radiative cooling. To our knowledge, this is the first diagnostic study of a warming with a consistent assimilated dataset.

The STRATAN data are compared to the conventional NMC analyses. The temperature analyses in the upper stratosphere show some systematic differences, with STRATAN being colder in the midlatitude cold belt. The biggest differences, however, are in the wind analyses, where the inadequacies of geostrophic and balanced wind estimates are most evident. It is argued that STRATAN wind and height fields should be more representative of the atmosphere because of the internal physical constraints provided by the assimilation procedure.

A comparison of the STRATAN temperature profiles to TOVS temperature profiles shows that the TOVS profiles within the anomaly are warmer on the anticyclonic shear side of the jet than in STRATAN. The differences are less than 5 K and are isolated to between 1 and 5 hPa, very close to the jet axis.

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