6. POLAR LOWS

6.1 Characteristics of Polar Lows

Polar lows are intense lows that occur generally north of and/or west of the Arctic front, on the cold air side of the 500 mb polar jet and quite often in areas of pronounced low level cyclonic shear. They vary in size, from 100 n mi (160 km) to 700 n mi (≈ 1,000 km) in diameter. Surface wind speeds can reach hurricane force (60 kt plus). Polar lows are typified by an atmosphere exhibiting conditionally unstable conditions through the lower troposphere and midtroposphere. Vertical shear in the vicinity of polar lows is generally weak. The geostrophic wind accompanying polar vortices displays this characteristic and tends to back with height. Polar lows differ from midlatitude lows not only in size but often in the mechanism for development, the speed of development and often dissipation, and in their internal structure. The last attribute is still a subject of some speculation. For example, on the one hand Carleton (1985) concluded that ". . .a cold core and little vertical tilt distinguish the polar low in the mean." On the other hand, Shapiro et al. (1987) in their analysis "of the first research aircraft observations of the three-dimensional structure of a polar low" found that ". . .the polar low possessed a warm inner core, whose layer-averaged thickness temperature exceeded the exterior environmental temperature at 108 n mi (200 km) from the low center by about 9 °F (5 °C)." Temperature analyses at 1,000 ft (300 m), 850 mb, and the 580- to 1013-mb thickness, all indicated a warm core structure.

Rasmussen (1985) also noted that the polar low circulation is strongest near the surface and weakens upward, implying a warm core structure at lower levels. This type of structure agrees very well with the observation that polar lows develop quite frequently in reversed shear atmospheric conditions. In reversed shear conditions the thermal wind in the lower troposphere opposes the surface flow. Reversed shear can be established in a number of ways. Quite commonly it occurs when a cold pool of air moves to the south of a surface low.

In that type of configuration westerly low-level flow around the vortex is opposed by an easterly thermal wind component. The Norwegians (Midtbo et al., 1986) noted that "Investigations on the growth of baroclinic waves are generally dealing with occasions when the thermal wind and the progression of the disturbance are in the same direction. In case of a reversed-shear flow the baroclinic instability takes place in a different way. The wind at the steering level is northerly and directed nearly opposite to the thermal wind." They also noted that the surface wind speeds are generally stronger than expected when compared with ordinary unreversed situations.

Figure 6-1 shows an example of a reversed-shear configuration on 1 November 1985. The figure shows an analysis of the 1000-mb height contours (solid lines), indicating a low over the Norwegian Sea. The 1000- to 850-mb thickness analysis (dashed lines) reveals that the thermal wind west of the low center is southerly (parallel to the thickness lines, with cold air to the left). The thermal wind, then, opposes northerly low-level flow about the low center.

When surface observations are available in the region of polar low development, an analysis of the 3-hourly pressure change reveals falling isallobaric patterns ahead of the low and strong rises to the rear. Figure 6-2 shows such a pattern in the area where a polar low developed near 70°N 0° at 1200 GMT on 1 November 1985. Six hours later (Fig. 6-3), the pattern is even more pronounced as the pattern and storm are displaced southward. Note, as is typical in a polar low development, that the polar low is on the periphery or west of the analyzed low center.

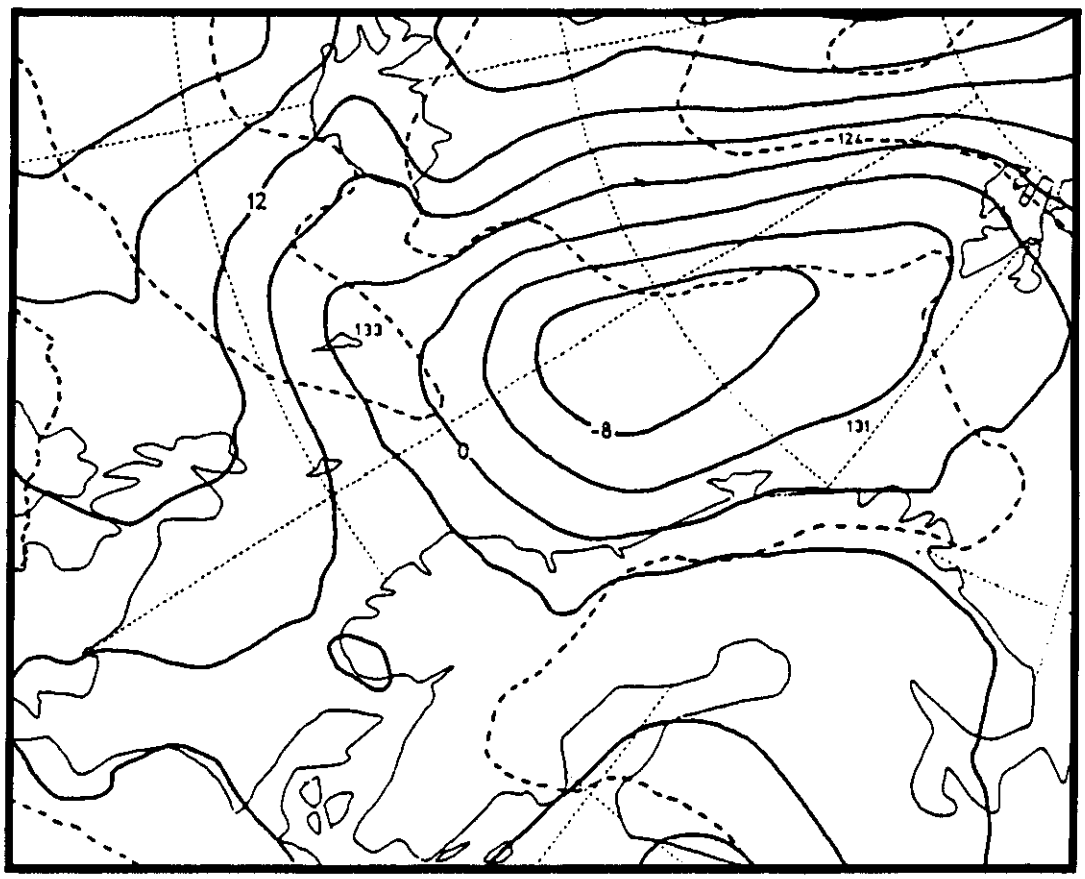

Figure 6-1. FNOC 1000-mb Analysis, 0000 GMT 1 November 1985. Solid lines are height contours (4 decameters) and dashed lines are 1000- to 850-mb thickness (4 decameters).

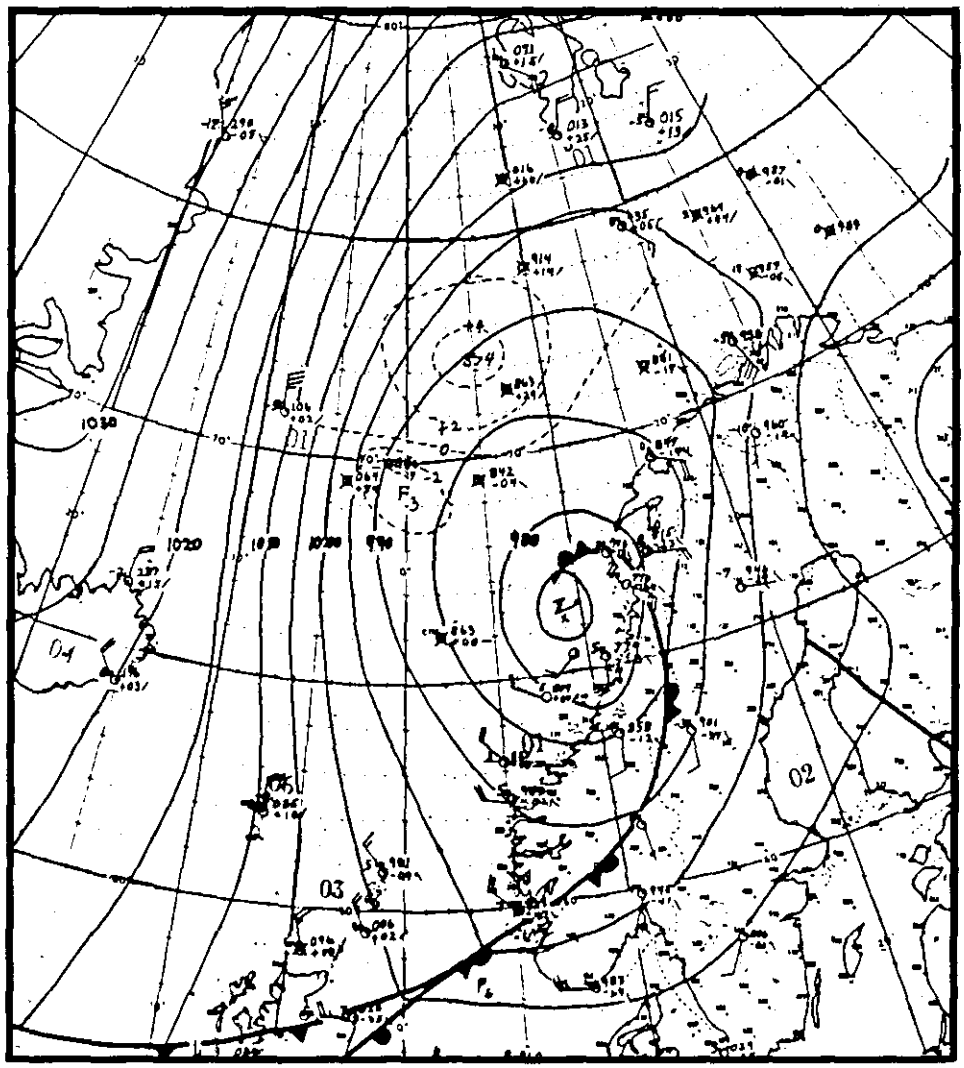

Figure 6-2. FNOC Surface Pressure Analysis, 1200 GMT 1 November 1985.

In the Norwegian report (Mitbo et al., 1986), criteria for selecting likely areas of polar low development were given. These are

1. Cold air advection must be present at the sea surface.
2. The thickness of the layer 850 to 500 mb must be below 13,000 ft (3,960 m). This limit should vary with the sea surface temperature: When the sea surface temperature is low, the limit should be lowered.
3. The curvature of the contour lines in the 500- to 700-mb levels must be cyclonic or zero.

Polar lows probably develop in different regions through either baroclinic instability or conditional instability of the second kind (CISK), and at times by a combination of these and other processes. In the "other" category is a third potential, described by Shapiro et al. (1987), in which essentially unmodified cold Arctic air moving southward over the ice, turned cyclonically, over the water, wrapping itself "about a warm(er) inner air mass, thereby isolating it from its warm air source. . . ."

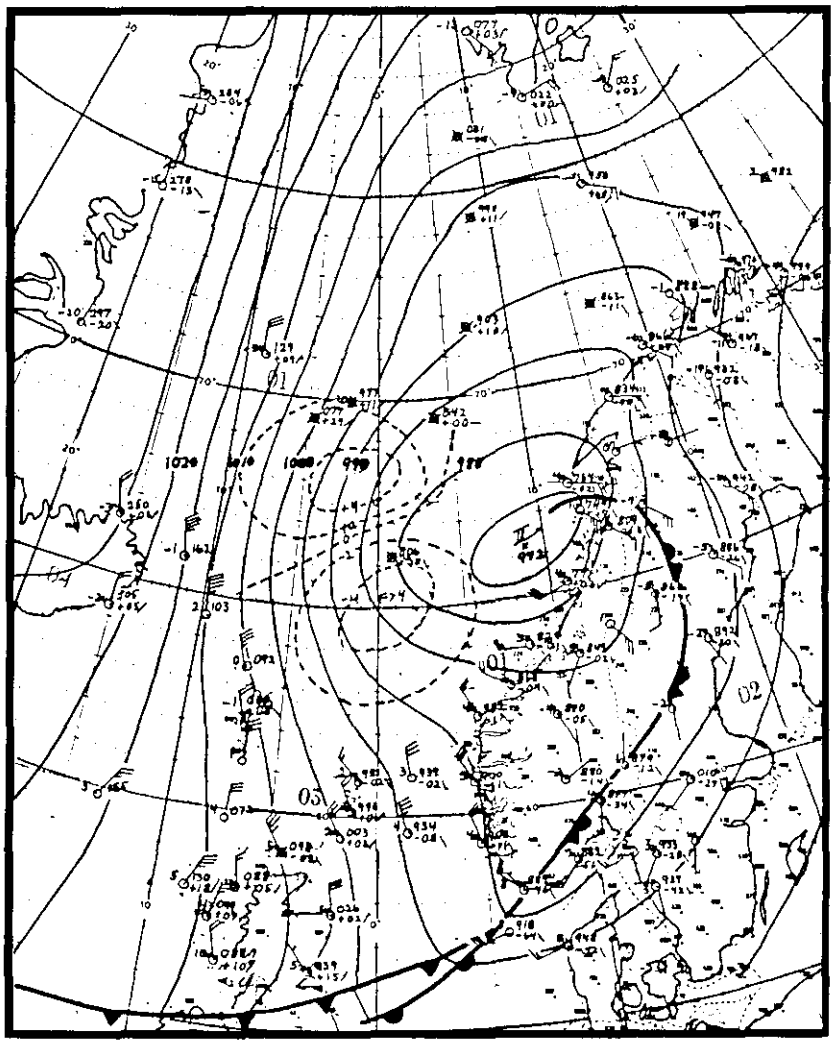

Figure 6-3. FNOC Surface Pressure Analysis, 1800 GMT 1 November 1985.

This process was first described by Bergeron (1928) who termed the process "a seclusion." The importance of the Shapiro study is that the classic definition of a polar low, being one that necessarily develops in a homogeneous cold polar air stream behind the Arctic front, is shown to be incorrect, at least in some instances.

Another important mode of development, believed to be of primary importance in the Bear Island region, was advection of an upper level low across the ice edge in the region of pronounced sea surface temperature gradient (Rasmussen, 1985). The Naval Environmental Prediction Research Facility (NEPRF) hypothesized that the upper cold core low caused the formation of the polar low of the same horizontal scale through the initiation of convection, which transformed the initial cold core system to a low level warm core system by release of latent heat and further development through the CISK mechanism.

Polar lows are difficult phenomena to forecast and seldom are forecast at all by numerical methods. Their life cycle can be completed in as little as 6 to 12 hours, or as long as 2 to 3 days; wind speeds can gust to 115 kt ($\approx$ 60 m/s), and seas can build to 50 ft (15 m) or higher. Structural icing is common and very hazardous for ships because of below freezing ambient air temperature in high winds and resulting spray. Sudden unexpected encounters with polar lows have cost the Norwegians an average loss of three ships a year over a timespan of many years because of these effects. Satellite images every few hours are an essential requirement in detecting polar lows and monitoring their evolution.

6.2 Polar Low Studies

In view of the importance of satellite imagery in studying polar lows, in this section a classic example of polar low development is shown for the area near Iceland. Additional examples over Greenland and the Norwegian Sea support the general concept. Finally, a schematic model is developed to indicate the precursor surface and 500-mb flow conditions associated with polar low development, especially applied to the Greenland Sea region.

6.2.1 Polar Low Evolution Near Iceland, 13 December 1986

Very seldom are conventional observations adequate to define the polar low in its initial development. A series of IR NOAA images on 13 December (Figs. 6-4(a) through 6-4(f)) reveal the complete evolution of a polar low within a period of approximately 14 hours in the area near Iceland. Evolution during this time starts from its inception as a short-wave trough (Fig. 6-4(a)), through cloud cluster development (Figs. 6-4(b) through 6-4(d)) to the beginning of a cloud vortex (Fig. 6-4(e)), and final encirclement of a spiral band to define a nearly cloud-free eye (Fig. 6-4(f)).

Synoptic charts (not shown) reveal that the polar low develops within a massive low pressure system covering the entire North Atlantic between Greenland and Norway, with a center near Iceland. The polar low does not develop near the center of low pressure but rather as an offset feature moving in the cyclonic flow around the periphery of this center. As the main low pressure center moves westward from near Iceland toward the coast of Greenland, the polar low is steered northward from its position in Fig. 6-4(b) toward the west coast of Iceland (Fig. 6-4(c)), and then further northward through the Denmark Strait (Figs. 6-4(d) through 6-4(f)). Because the low is an embedded feature in a larger scale circulation that has a strong pressure gradient extending outward for hundreds of miles, assessment of the intensity of the feature in itself is difficult. Similar appearing vortices passing Keflavik have produced sustained winds of over 50 kt (personal communication, CDR R.T. Pearson, Navy Oceanography Facility, Keflavik, Iceland, 1987).

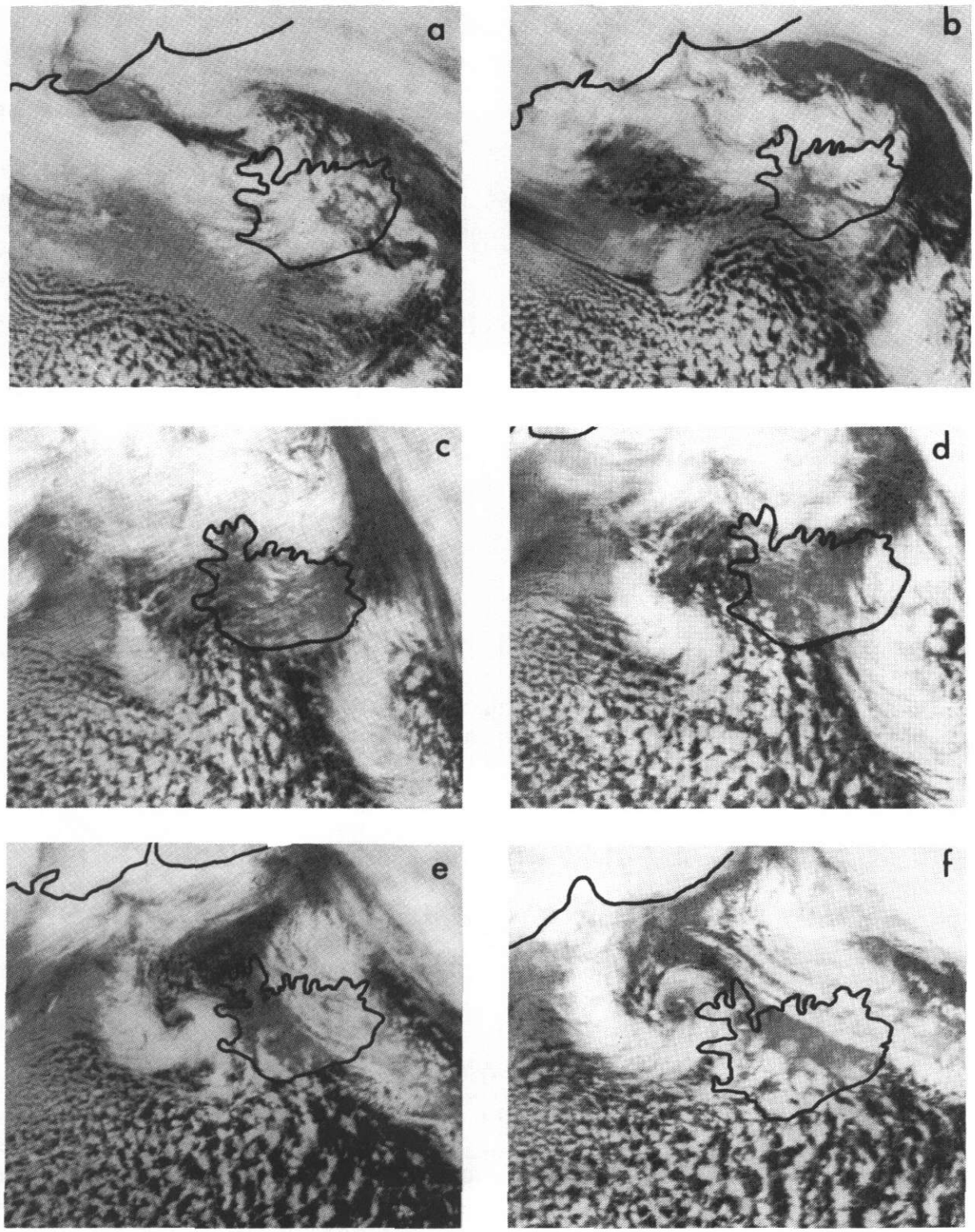

Figure 6-4. IR NOAA 9 and 10 Imagery for 13 December 1986: (a) 0603 GMT, (b) 1051 GMT, (c) 1420 GMT, (d) 1601 GMT, (e) 1904 GMT, (f) 2045 GMT.

Importantly, the small polar low develops within a general low pressure region adjacent to an unusually enhanced field of open-celled cumulonimbus. The 500-mb data (not shown) reveals a cold trough aloft superimposed over the enhanced convection. The trough is preceded by a warm ridge so that upper level winds turn anticyclonically downstream from the trough. This pattern appears favorable for polar low development. From a satellite perspective, the enhanced convection is the clue that an upper cold trough is in the area. This information, combined with other favorable factors, serves as advance notice of potential polar low production.

6.2.2 Polar Low Evolution, 26–27 February 1984

On 26 February the FNOC surface analysis for 0000 GMT (Fig. 6-5) reveals a synoptic-scale low west-northwest of Iceland. An IR DMSP mosaic over the region at about 0200 GMT (Fig. 6-6) fails to reveal a distinct vortex associated with this low pressure region. Readily apparent in the DMSP data, however, is the large field of enhanced, open-celled convection in the region of cyclonic flow southwest of Iceland. Aloft at 500 mb (Fig. 6-7), a cold trough extends over the field of enhanced convection.

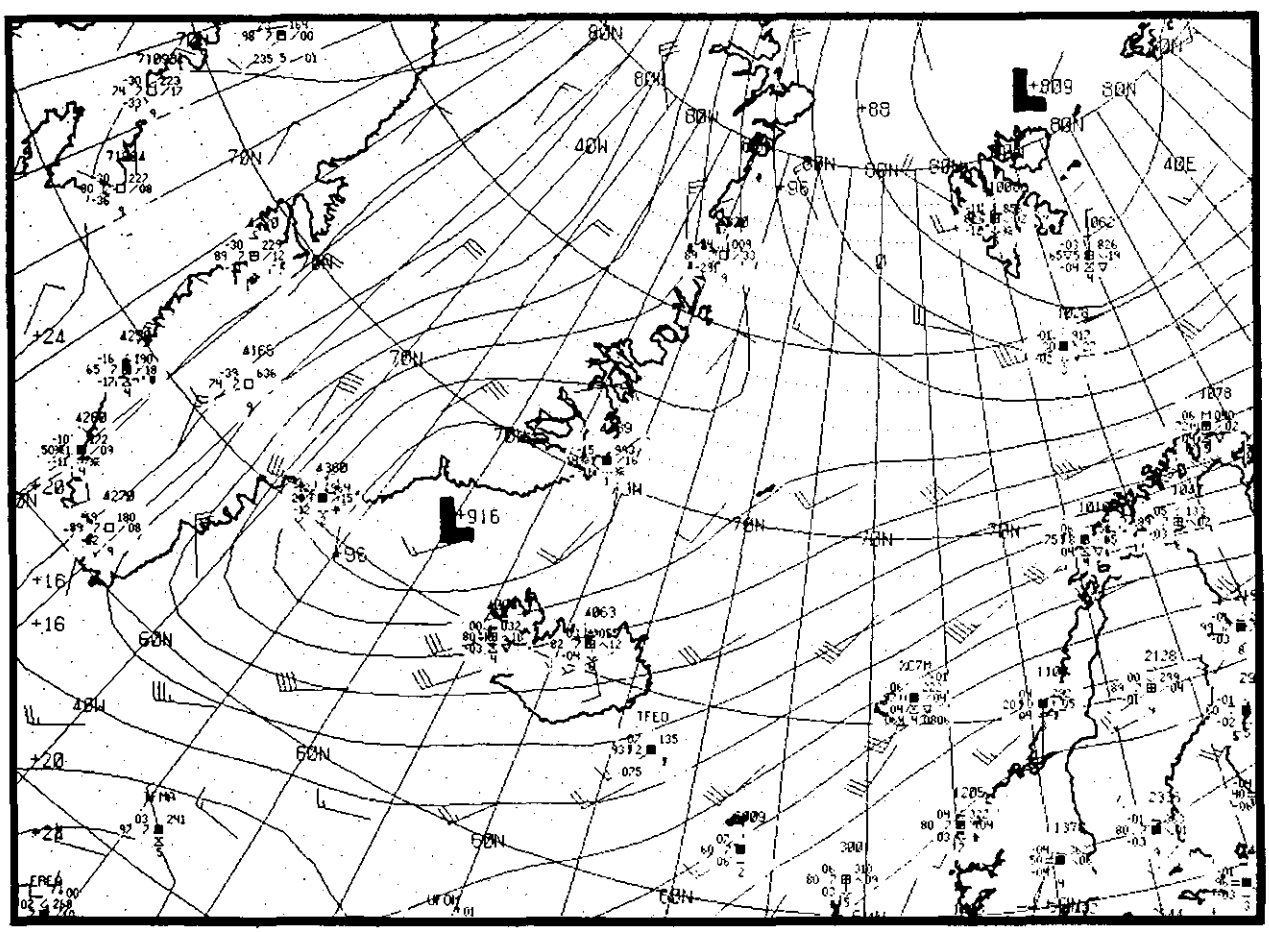

Figure 6-5. FNOC Surface Analysis, 0000 GMT 26 February 1984.

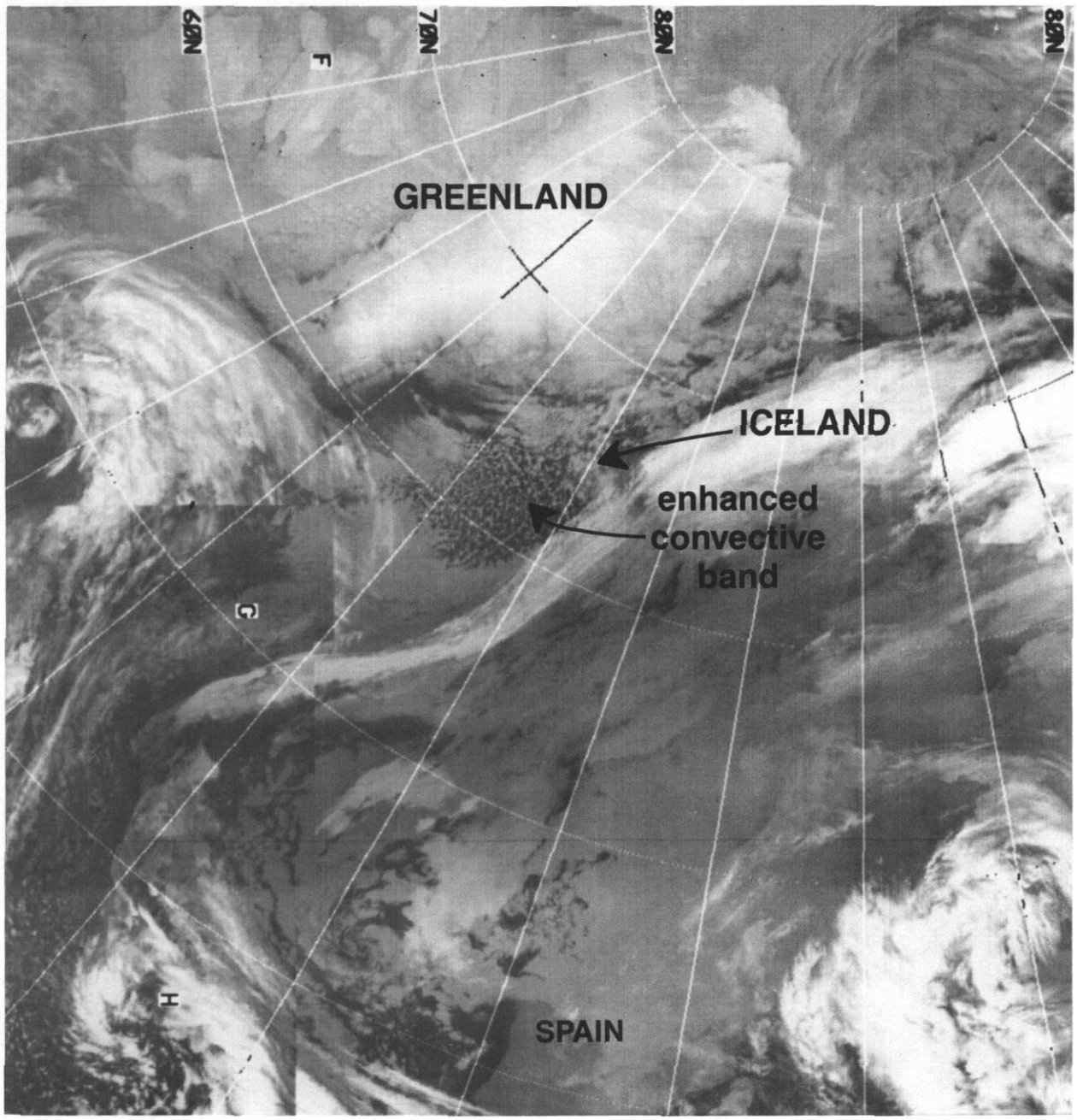

Figure 6-6. DMSP Imagery, 0200 GMT 26 February 1984.

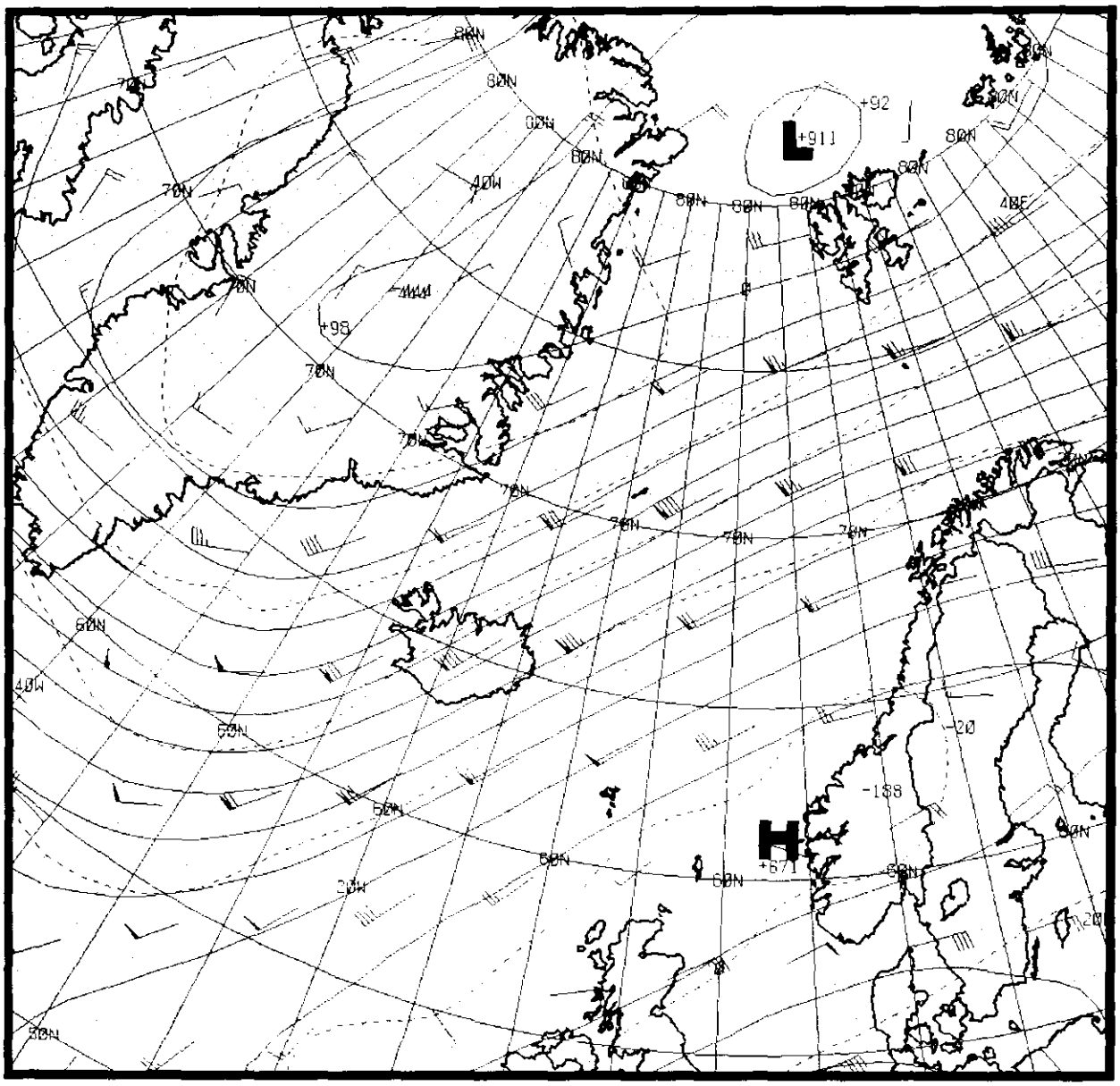

Figure 6-7. FNOC 500-mb Analysis, 0000 GMT 26 February 1984.

By 2330 GMT DMSP data (Fig. 6-8) shows an east-west oriented band of heavy convection north of Iceland, with cyclonically curved bands in the cellular cloud field leading up to the band passing over Iceland. The surface analysis for 27 February at 0000 GMT (not shown) reveals an elongated trough leading through the banded cloudiness toward Bear Island. The 500-mb analysis for the 27th at 0000 GMT (Fig. 6-9) shows the cold trough aloft with flow turning anticyclonically over the banded structure.

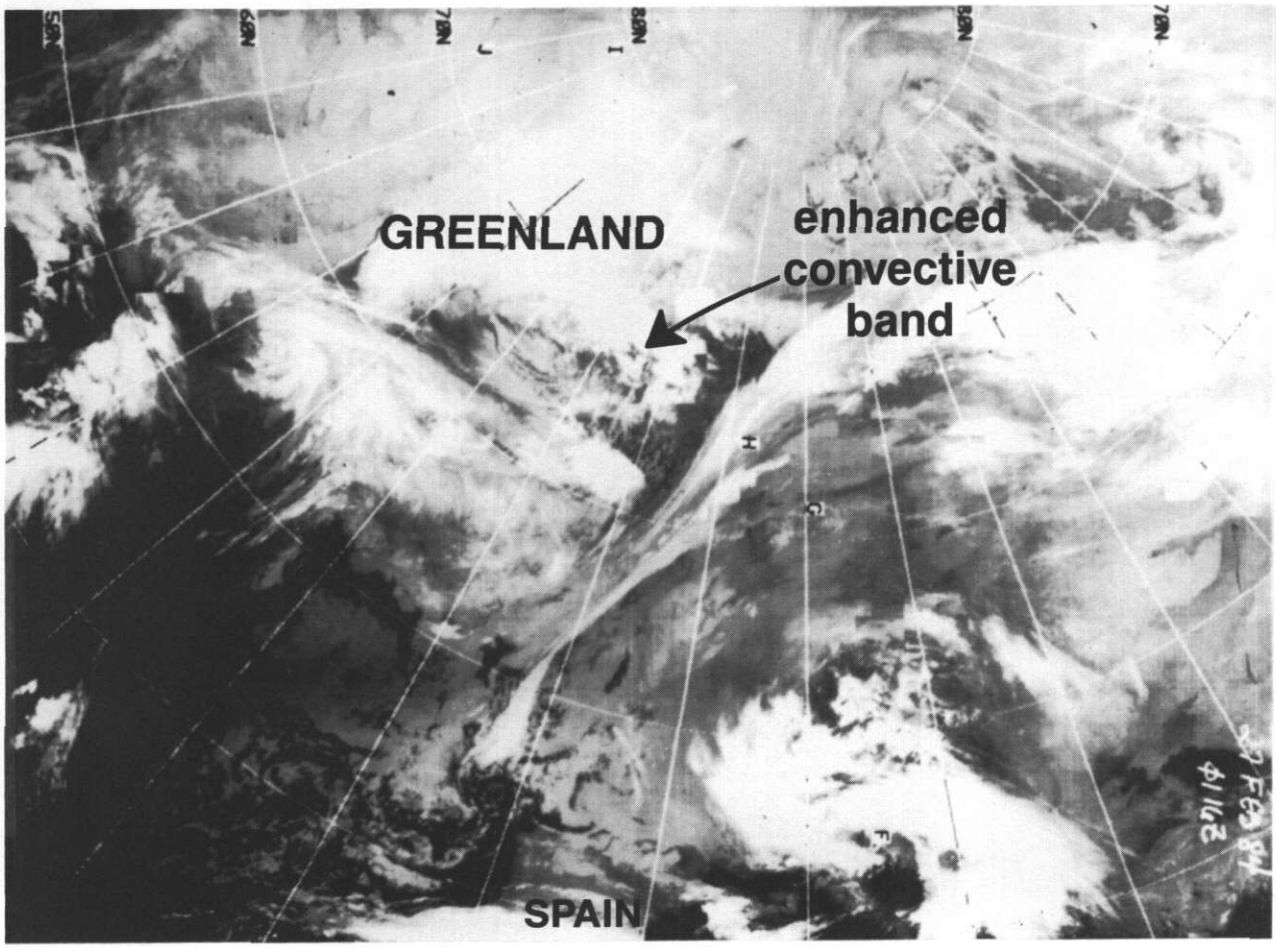

Figure 6-8. DMSP Imagery, 2330 GMT 26 February 1984.

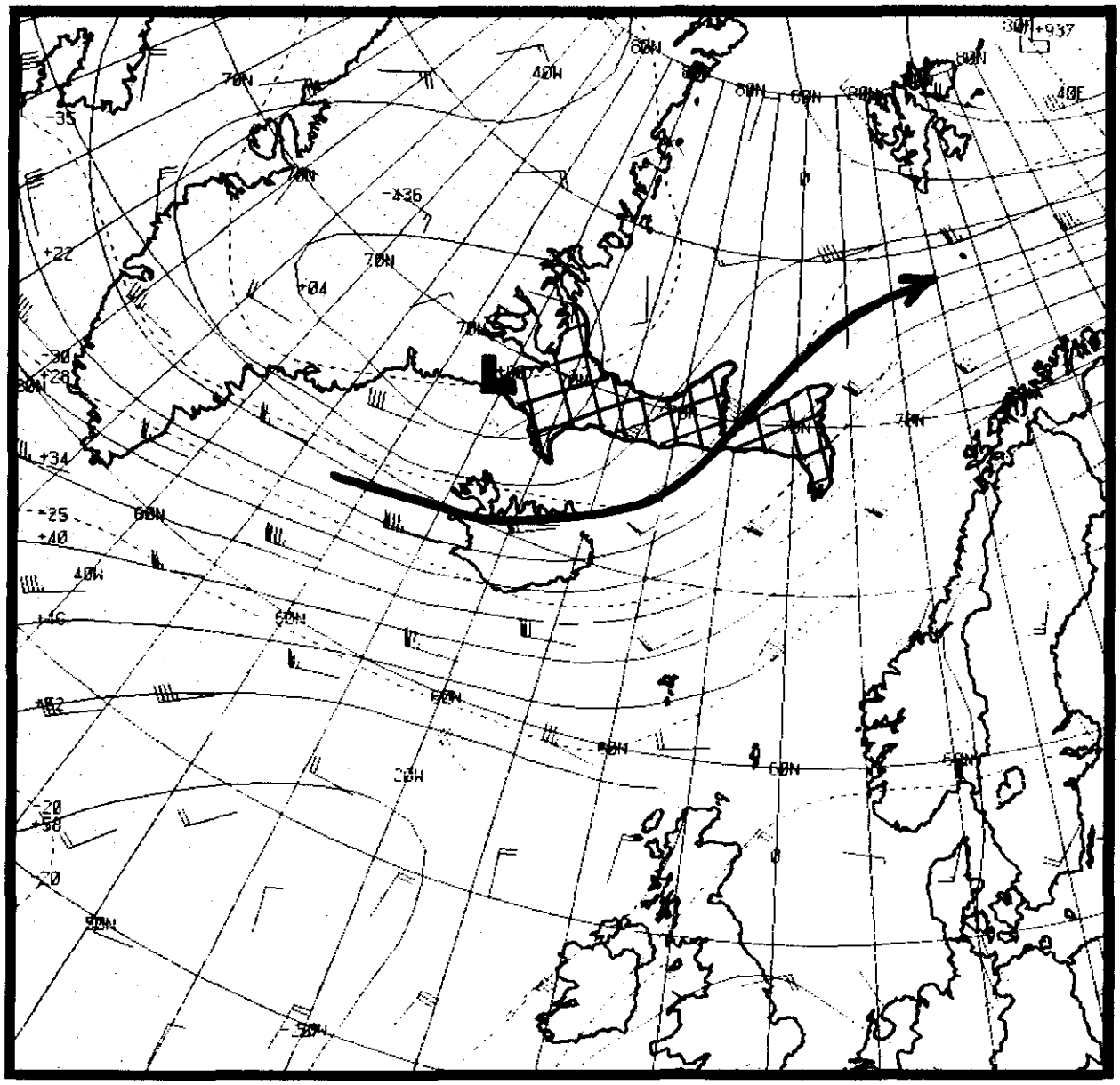

Figure 6-9. FNOC 500-mb Analysis, 0000 GMT 27 February 1984.

Further development of the banded structure revealed by DMSP data near 0700 GMT on the 27th (Fig. 6-10) shows an apparent grouping of at least three distinct cloud clusters. These structures coalesce into a significant polar low, as revealed in NOAA-7 data at 1340 GMT on the same day (Fig. 6-11). This storm was penetrated by a P-3 research aircraft near the time the NOAA data was acquired. Surface winds up to 65 kt (33 m/s) were measured as the storm reached peak intensity (Shapiro, 1986). Again, an important observation is the presence of heavy convection in the storm's southern quadrants associated with a cold trough aloft.

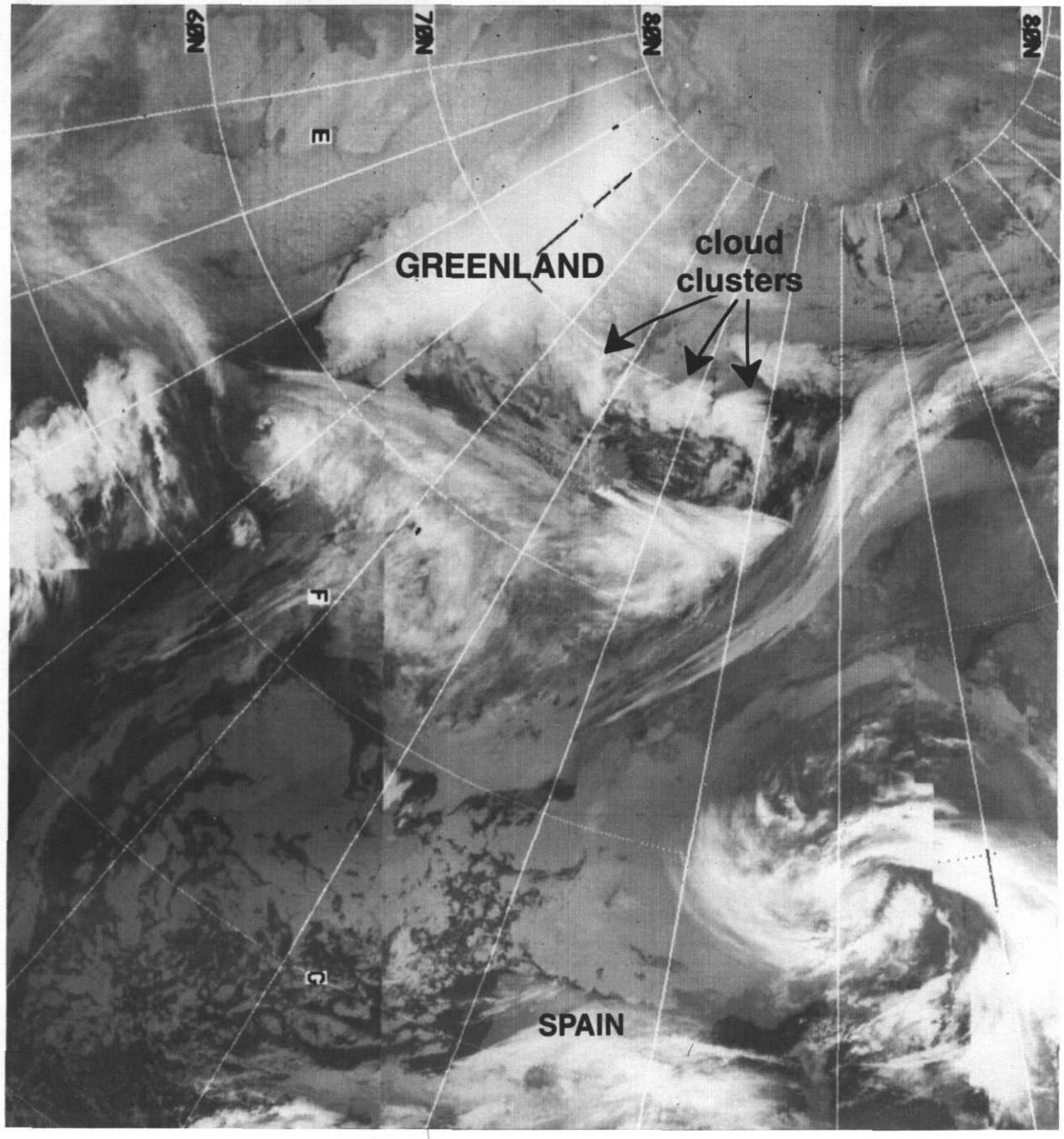

Figure 6-10. IR DMSP Imagery, 0700 GMT 27 February 1984.

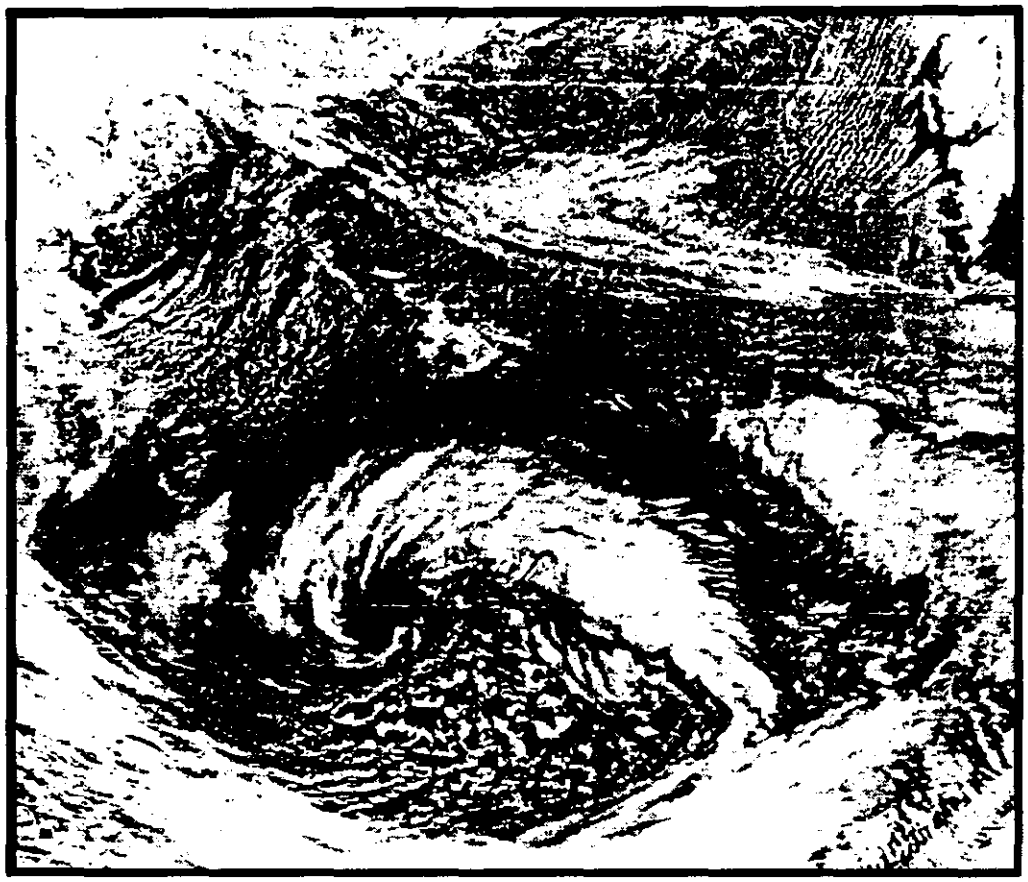

Figure 6-11. NOAA-7 Imagery, 1340 GMT 27 February 1984.

Figure 6-12 shows the FNOC 500-mb analysis for 1200 GMT on this date with the outline of storm cloudiness superimposed. These data verify the location of the cold trough that traveled with the moving disturbance. The surface analysis, Fig. 6-13, fails (as is common in these circumstances) to locate the center of the major development that had reached hurricane force by this time.

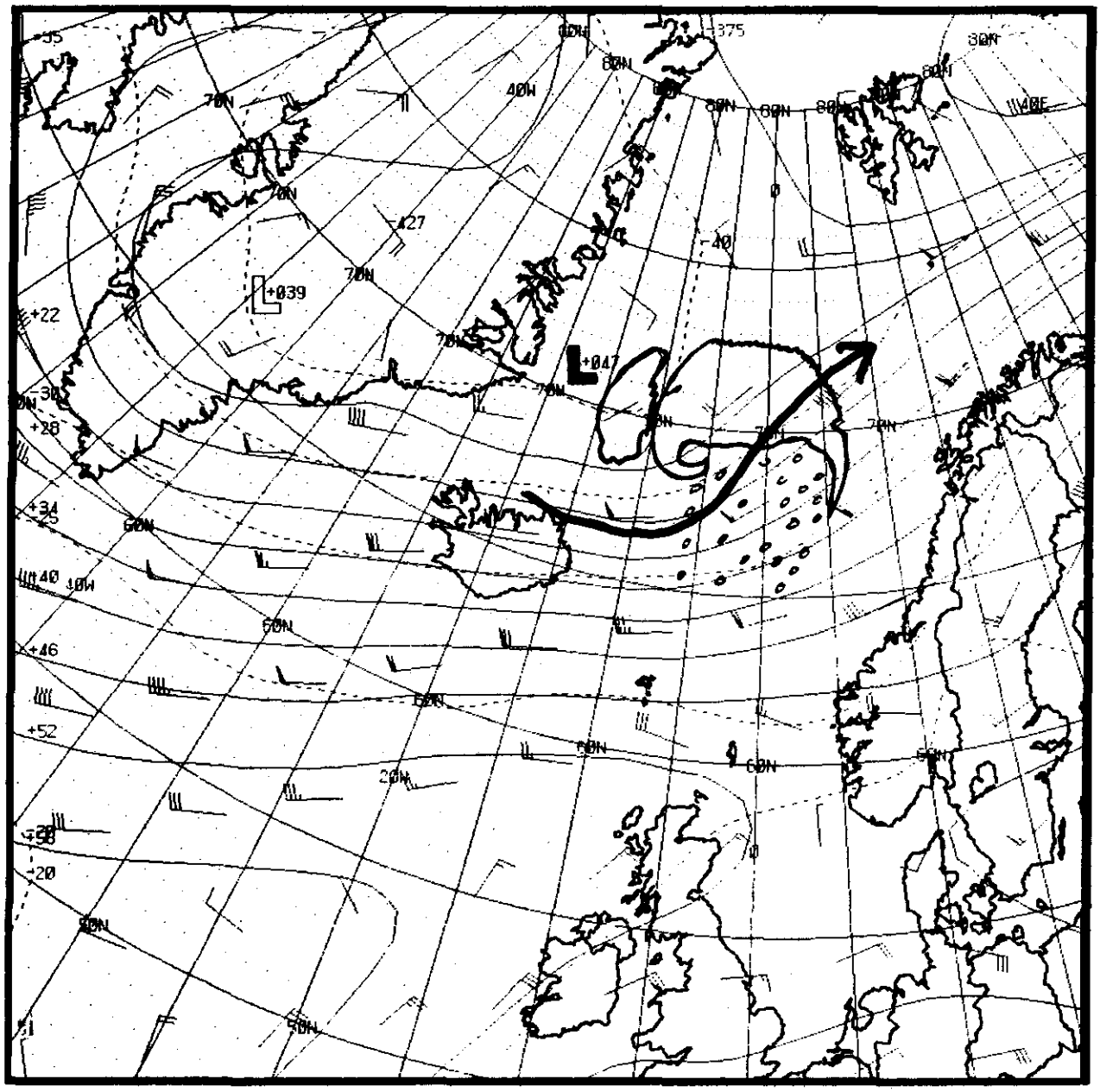

Figure 6-12. FNOC 500-mb Analysis, 1200 GMT 27 February 1984.

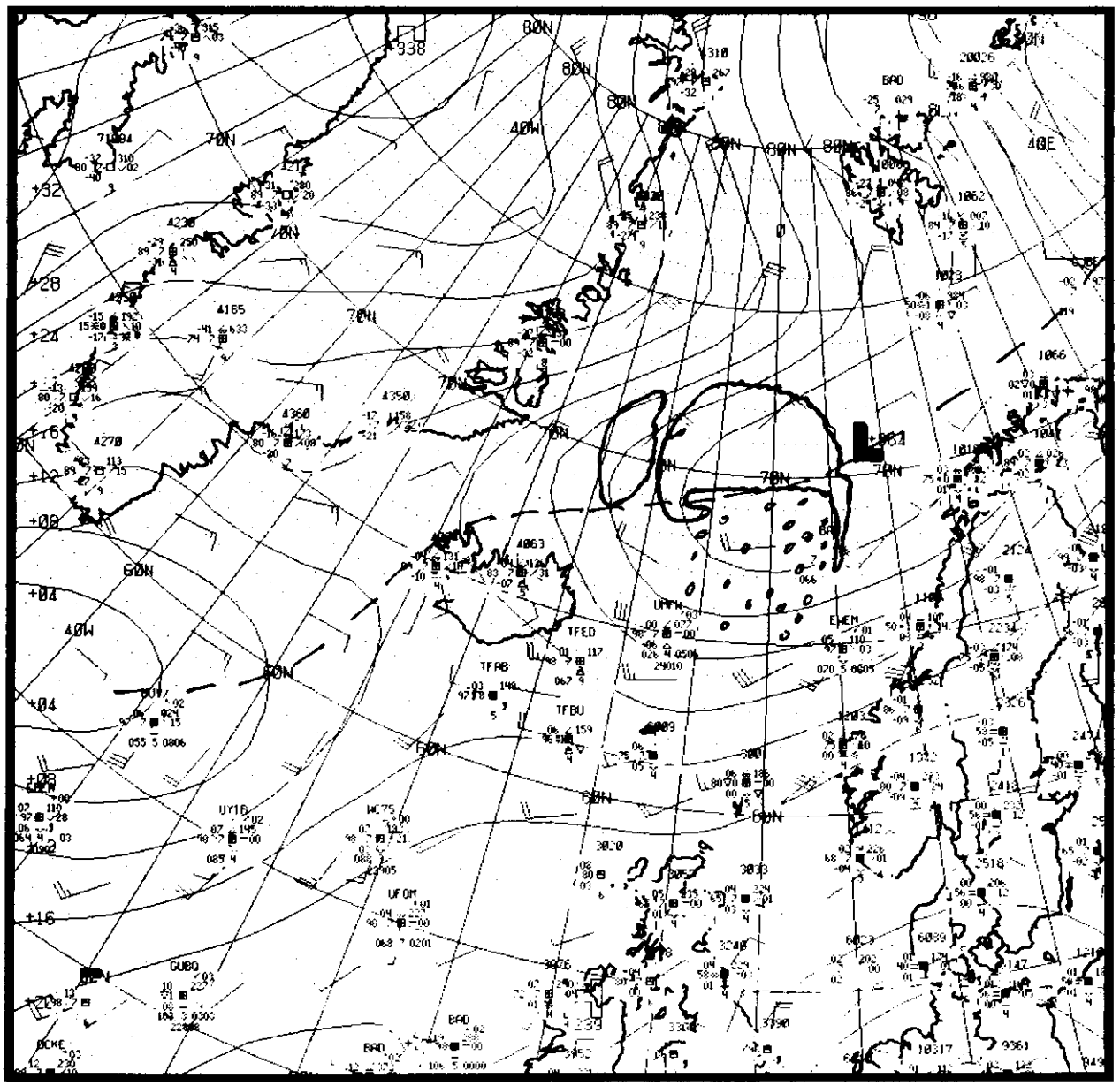

Figure 6-13. FNOC Surface Analysis, 1200 GMT 27 February 1984.

6.2.3 Polar Low Evolution, 13–14 February 1984

Figures 6-14 and 6-15 on 13 and 14 February 1984, respectively, illustrate how effectively enhanced cumulonimbus convection in open-celled areas can be used to track general areas where polar low development is anticipated. This convection occurred in the northernmost cloud cluster of Fig. 6-15, as documented in another P-3 aircraft flight into the storm area on 14 February (Shapiro, 1986). In this example, as in the others, a cold trough aloft triggers deep convection in the general region of surface low pressure. The cloud band north of the convection is associated with an asymptote of convergence in surface streamline flow (not shown). At 500 mb (not shown) the flow tends to turn anticyclonically over the overcast cloud band; therefore, a pattern of low-level convergence and upper level divergence develops very similar to the warm core structure of tropical cyclones that polar lows often resemble.

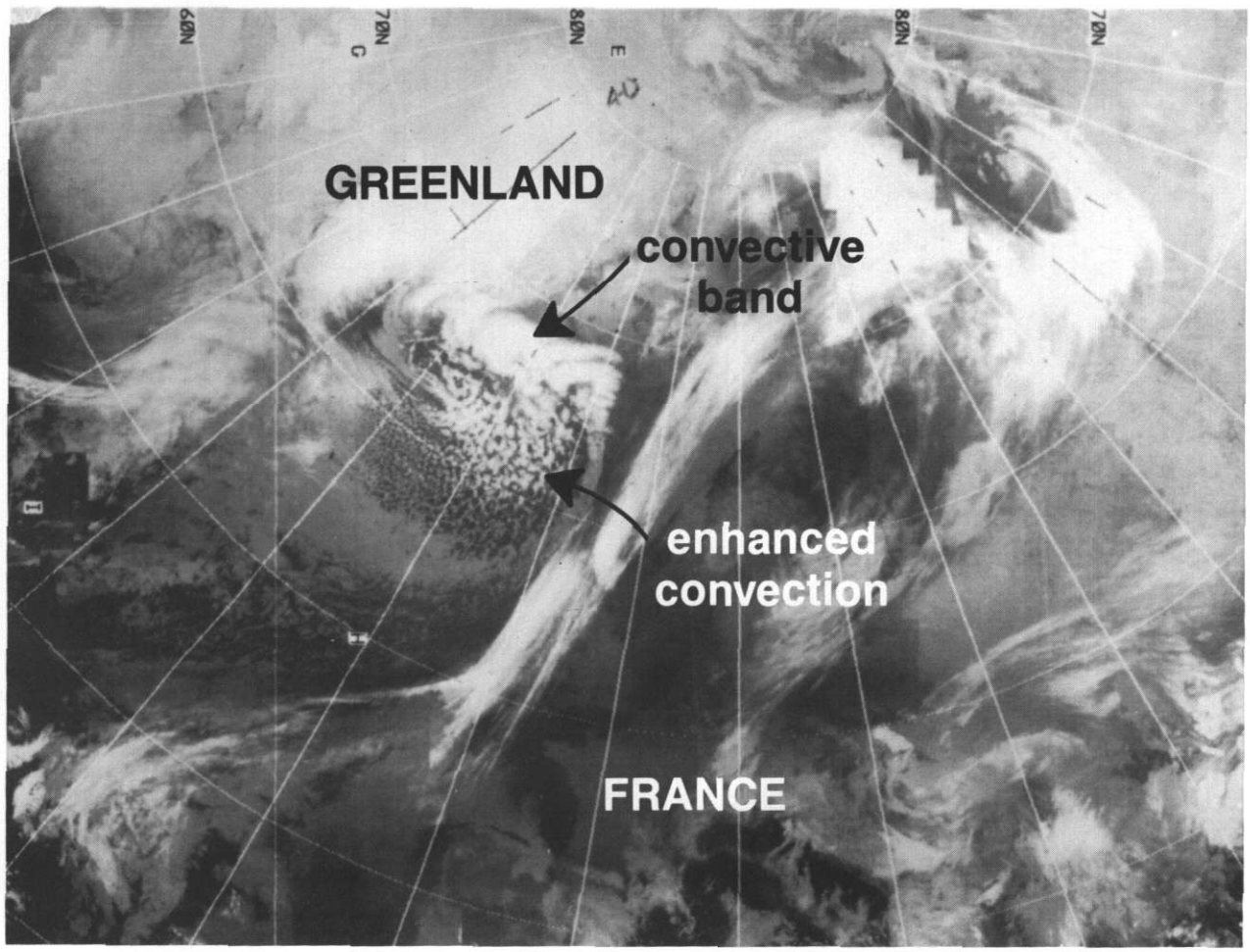

Figure 6-14. DMSP Imagery near Iceland, 0040 GMT 13 February 1984.

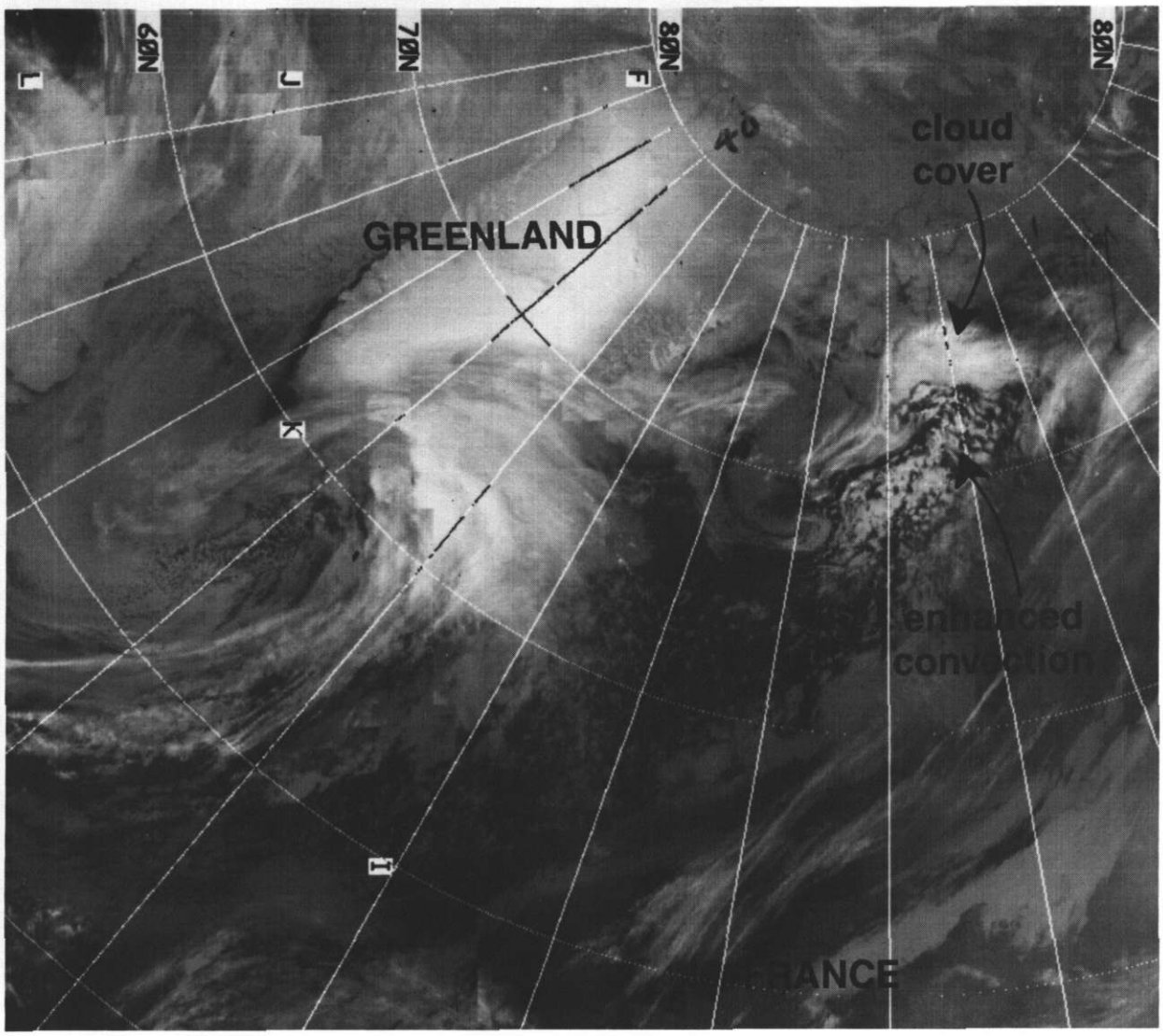

Figure 6-15. DMSP Imagery near Norway, 0234 GMT 14 February 1984.

6.2.4 Polar Low Development Near Bear Island, 12-14 December 1984

The Bear Island region is another area favored for polar low development. The climatologically favored easterly flow past Spitsbergen produces inverted troughs (similar to easterly waves) that move across the Fram Strait between Spitsbergen and Greenland. Vortices develop along and especially at or near the base of the trough axis. Figure 6-16 shows an example collected during MIZEX of 1987. Here fresh southeasterly winds flow off ice from the area just south of Spitsbergen and converge along a distinct, nearly north-

south-oriented cloud band. Winds are much weaker west of the cloud band and turn cyclonically in that region in the manner of an inverted trough. A cloud vortex is apparent at the southern end of the cloud band, and a second, even better developed vortex appears just southeast of the first. Open-celled cloud fields are in the region, but no evidence exists of deep convection that would signal the presence of an upper level cold low or trough. These vortices fail, therefore, to develop further and soon dissipate.

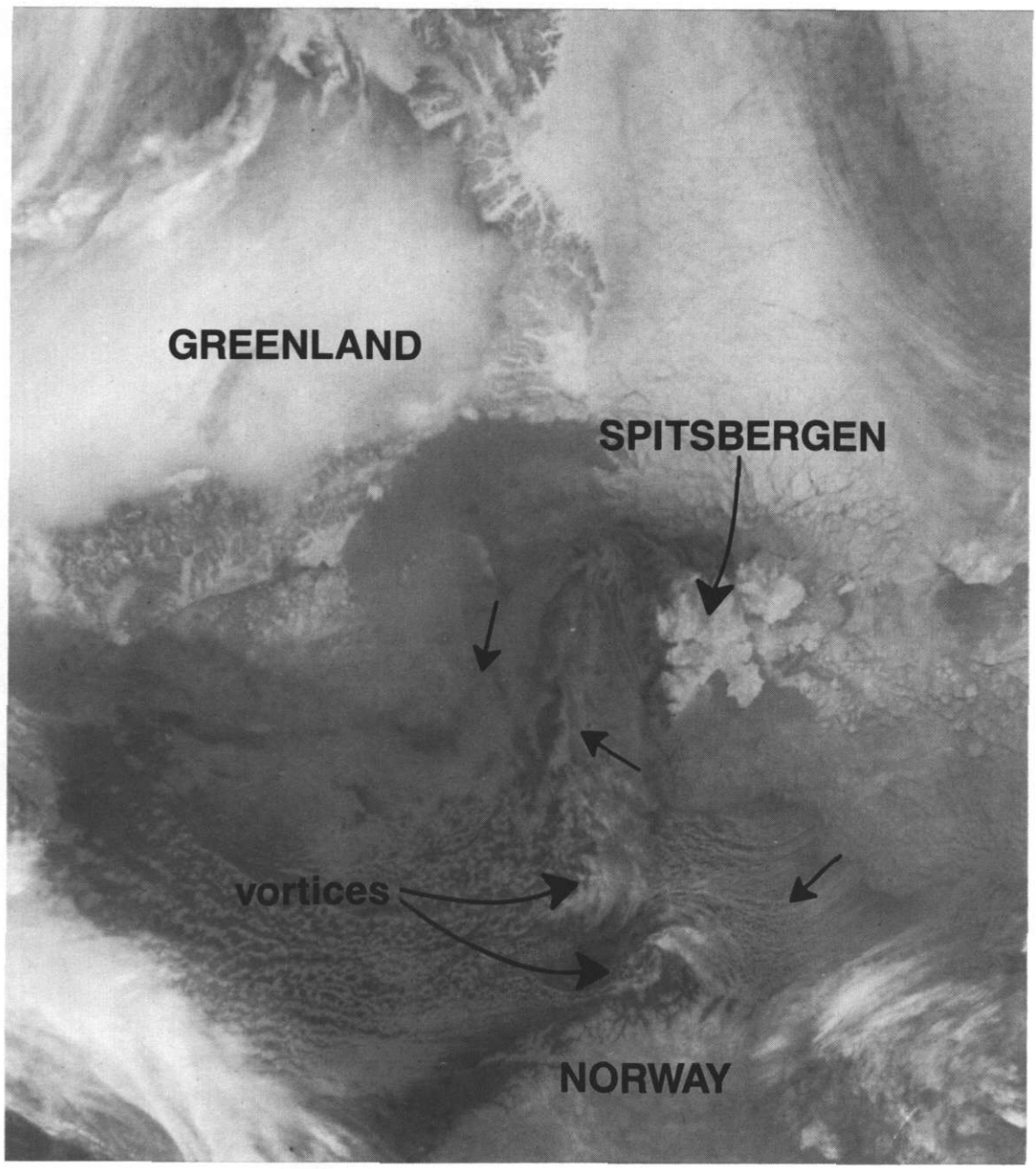

Figure 6-16. DMSP Imagery, 0337 GMT 26 March 1987.

Figure 6-17 shows a similar-appearing vortex at the end of the inverted trough just to the west of Bear Island, indicated by an X (Spitsbergen is marked with an S), on the image. In this example deep convection with abundant cumulonimbus development is evident south of the vortex.

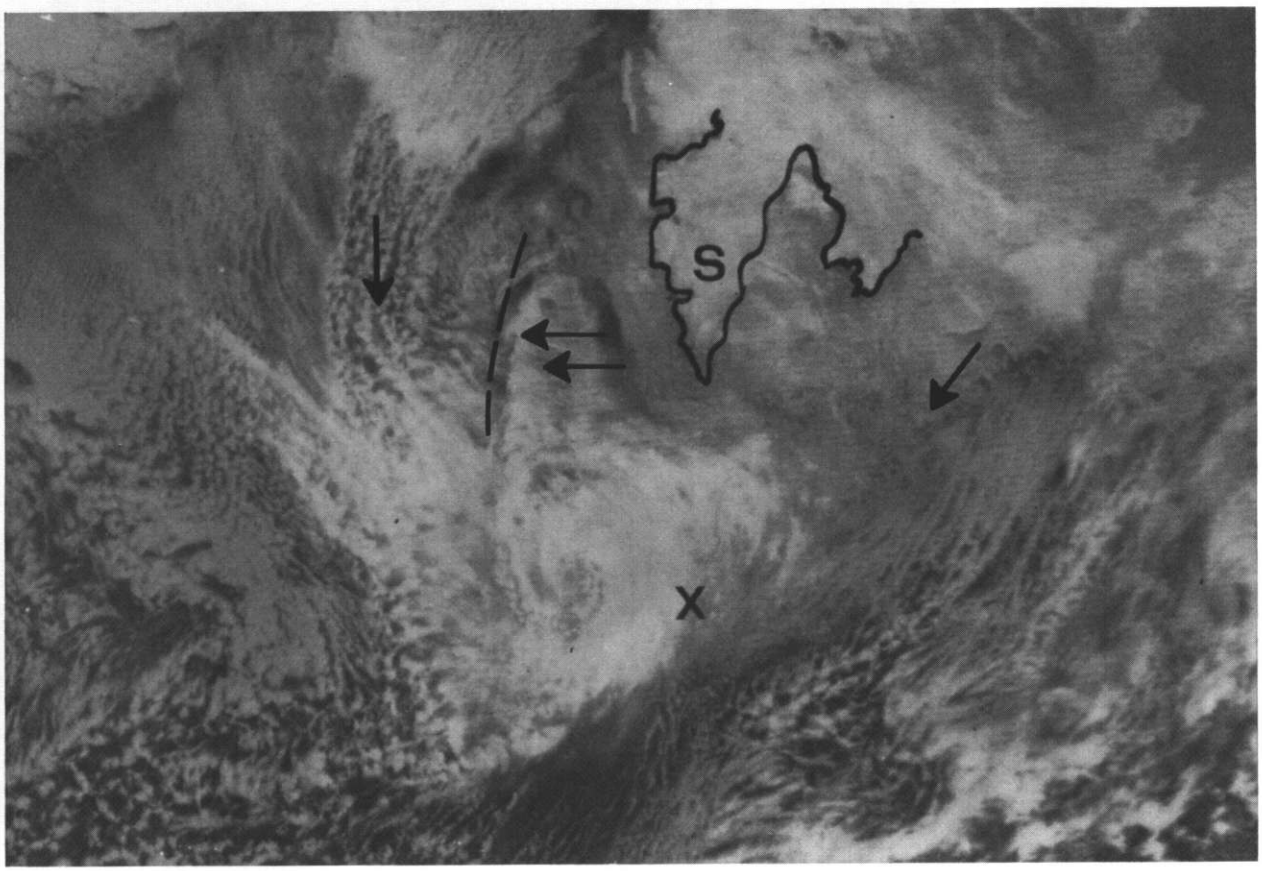

Figure 6-17. IR NOAA 7 Imagery, 1257 GMT 12 December 1982 (Courtesy of University of Dundee).

The 500-mb data (not shown) indicate a deep cold trough over the region with temperatures below −40 °F (−40 °C). Within about 12 hours the storm deepens to an intense-appearing vortex with a clear eye and within 24 hours changes from cold core to warm core structure (Rasmussen, 1985). A NOAA-7 image of the storm at 1245 GMT on 13 December is shown in Fig. 6-18. Characteristic very heavy cumulonimbus activity is evident in all of the southern directions from the storm center.

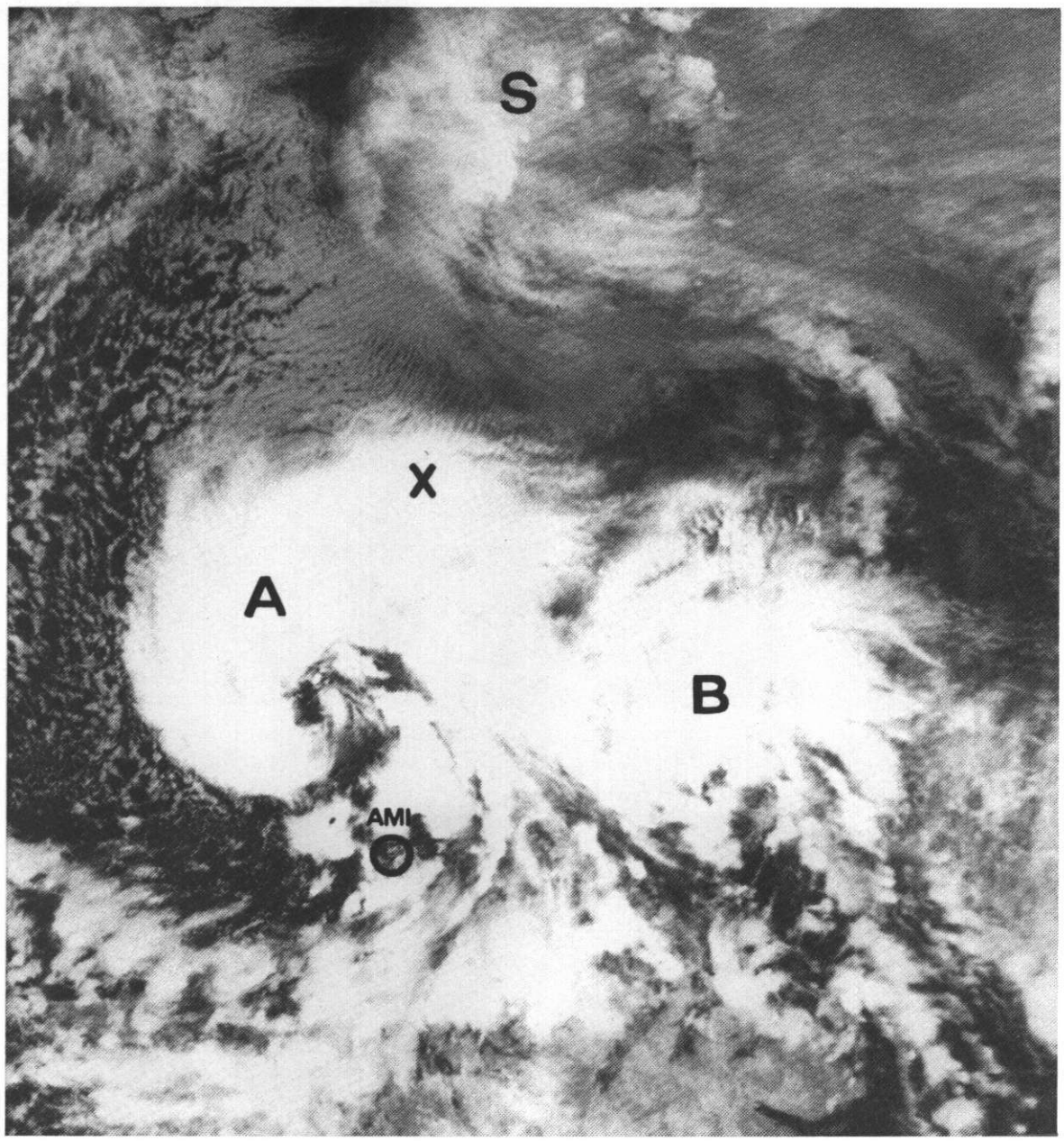

Figure 6-18. IR NOAA 7 Imagery, 1245 GMT 13 December 1982 (Courtesy of University of Dundee).

The 500-mb analysis at this time is shown in Fig. 6-19. The southeast wind direction at Bear Island suggests outflow characteristic of a warm core system. Winds immediately above 500 mb turn to the southwest, indicating even more strongly the outflow characteristic.

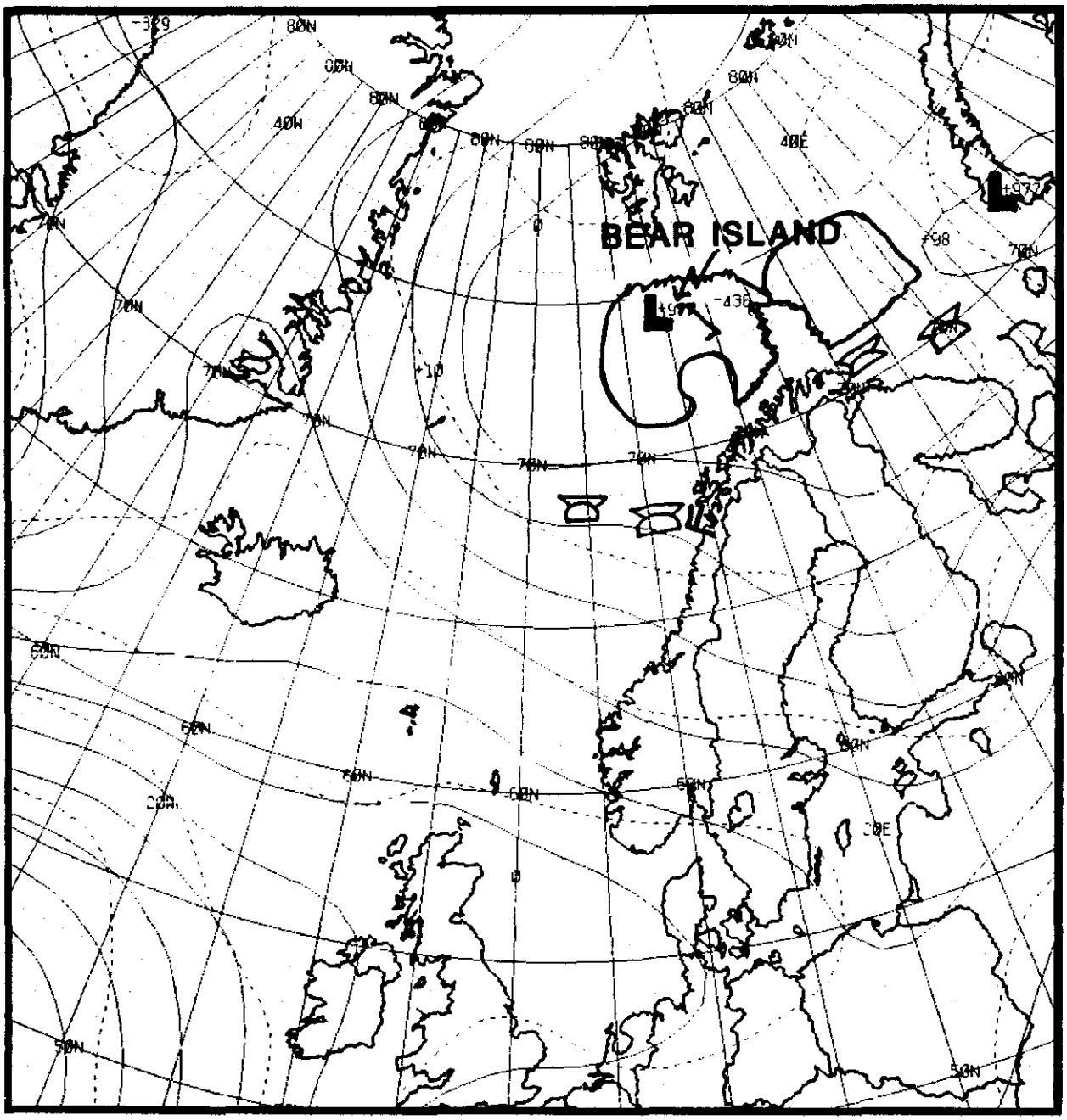

Figure 6-19. FNOC 500-mb Analysis, 1200 GMT 13 December 1982.

6.3 Precursor Surface and 500-mb Flow Patterns

6.3.1 Precursor Surface Conditions

The examples in the preceding section document the importance of an upper cold low or trough relative to polar low development. They also suggest general flow patterns in relation to satellite observed cloud development that are favorable for polar low evolution. Figure 6-20 shows the frequently observed surface pressure pattern associated with polar low development in the region near Iceland. The overcast cloud band may contain one or more vorticity centers from which the major low will develop. The band generally appears offset from the main low pressure center in a region of weak vertical shear. A streamline analysis of the area would reveal an asymptote of convergence in the area occupied by the band. Businger (1985) noted the typical ridging over northwestern Greenland that preceded polar low development. The effect of the ridging, in conjunction with the general trough over Greenland and the Norwegian Sea, results in a northerly flow component of very cold air along the east coast of Greenland, over the MIZ. When this cold air is eventually advected over the warm Gulf Stream Current, which flows past Iceland, an enormous exchange of heat and moisture can contribute to storm intensification.

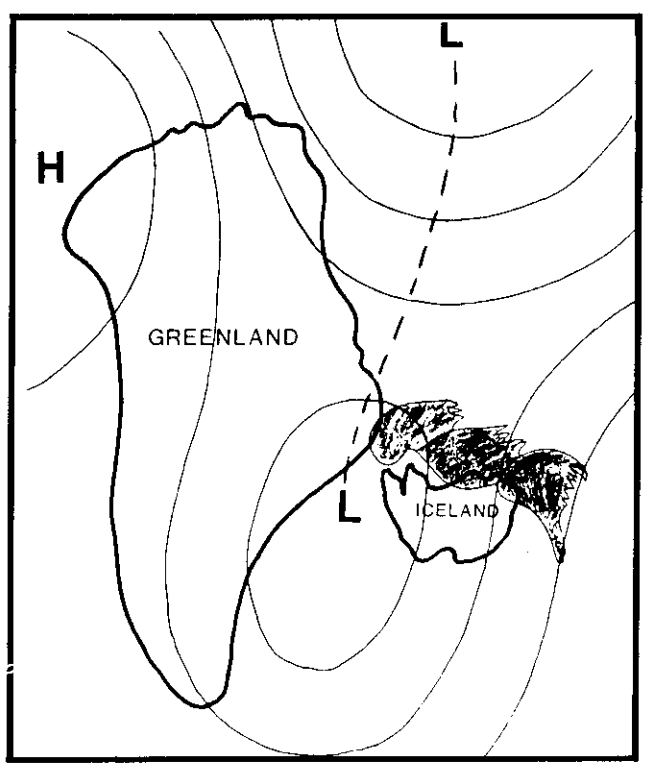

Figure 6-20. Schematic of Surface Conditions Favorable for Polar Lows.

6.3.2 Precursor 500-mb Flow Patterns

Figure 6-21 shows the typical 500-mb flow pattern frequently noted in polar low development. The cold upper level trough triggers deep open-celled convection. Release of latent heat in the convective cloud band may assist in building the warm ridge sometimes observed over this feature. In this configuration the thermal wind component, V_T acts to accelerate the wind flow over the convective cloud band, thereby enhancing upper level divergence (Fig. 6-21). The upper level trough and developing polar low move together with time in mutual interaction.

In conclusion, intense polar low developments can be distinguished from weaker nondeveloping vortices, noted in Arctic regions, by the presence of deep open-celled convection adjacent to banded overcast cloudiness. The deep open-celled convection indicates cyclonic low-level flow with a cold low or cold trough aloft. Good quality, high resolution satellite data are required at frequent intervals to monitor the progress of such developments.

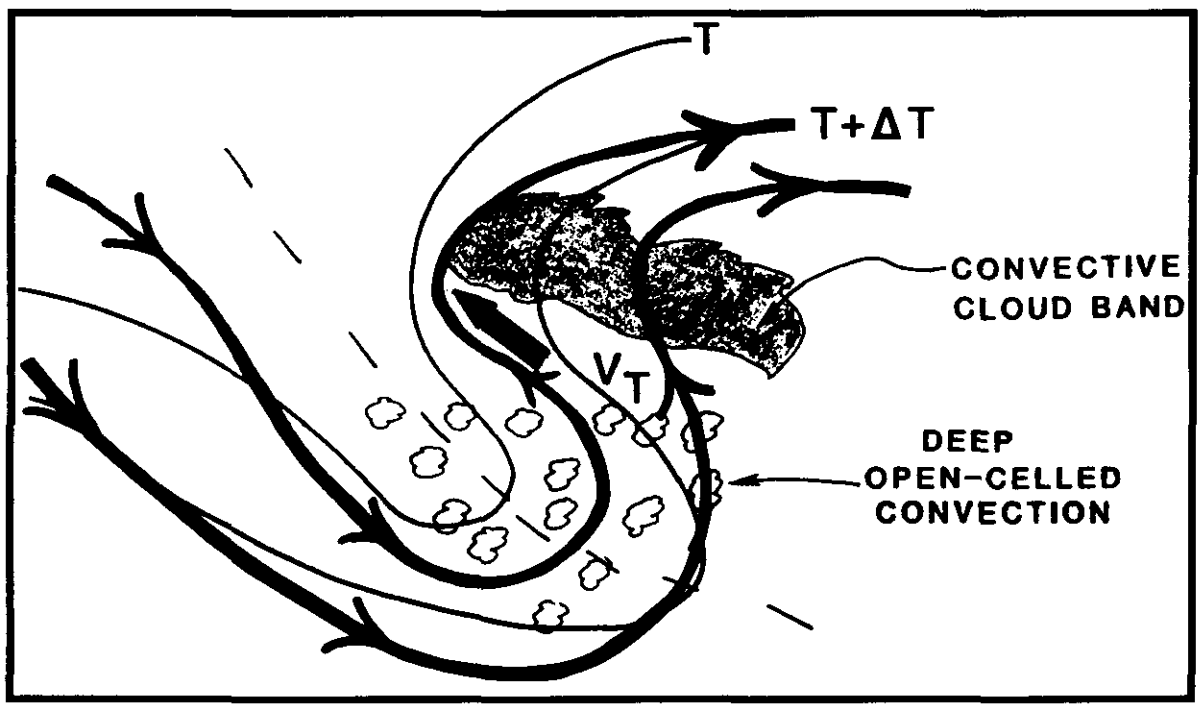

Figure 6-21. Schematic of 500-mb Conditions Favorable for Polar Lows.