A Case Study of Jet Stream Clouds

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Abstract

Certain specific cloud forms are frequently associated with strong winds aloft. Investigation has shown that these clouds are usually observed in regions of strong horizontal and vertical wind shear, and in most cases in or above regions where negative absolute vorticity (dynamic instability) and low Richardson's numbers coincide. The small-scale details in jet stream clouds may be the result of turbulent motion in layers brought to or near saturation by large-scale vertical motions in the atmosphere. This paper contains a detailed study of a typical case.

Introduction

In an earlier article SCHAEFER (1953) presented pictures of four cloud types including specific forms of cirrus, cirrocumulus and altocumulus which are frequently associated with strong winds aloft. The use of cloud observations to locate wind maxima from the ground is limited, of course, to periods when lower clouds do not obscure the sky; however, high-flying aircraft do not experience this limitation and could undoubtedly make good use of such jet stream "indicators". One case wherein a commercial plane followed a cloud streak for 1,500 miles over the Atlantic has been brought to the writers' attention.

It seems unlikely that jet stream clouds are formed simply because the wind speed exceeds some critical value. We do know, however, that condensation may occur in the atmosphere if sufficient cooling takes place; this is synonymous in adiabatic processes to saying that ascending motion is required. It is well known that meridional circulations are associated with jetlets moving in a westerly current at a speed less than that of the maximum wind. Comparatively large-scale ascending motions are usually found, for instance, on the warm-air side of the jet stream entrance region.

The clouds which have come to be associated with jet streams are quite spectacular and the sky is frequently very chaotic in appearance. Reports from pilots indicate that the small-scale structure of these clouds might be due to turbulence elements. For this reason it was deemed desirable to study some of the more well-known atmospheric stability criteria in an attempt to find an explanation for the fine details observed in jet stream clouds.

Stability criteria

The jet stream has been defined as a narrow band of high wind speed embedded in a relatively quiescent surrounding atmosphere. Such a current is, therefore, characterized by strong wind shear, both horizontally and vertically, between the core and the surroundings. Several instability criteria which contain shearing terms (either horizontal or vertical) have been prominent in meteorological literature for the past 30 years. Three different criteria dealing with horizontal cyclonic, horizontal anticyclonic and vertical wind shear will be given particular consideration.

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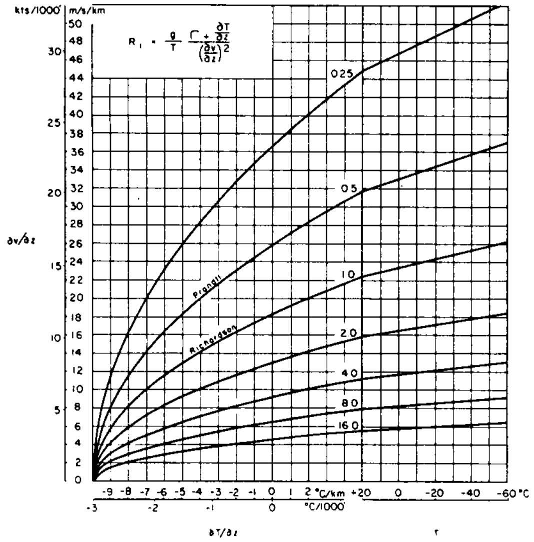

Fig 1. Graph of Richardson numbers (heavy solid lines) as function of lapse rate $(\partial T/\partial z)$, vertical wind shear $(\partial v/\partial z)$ and temperature (T).

Solberg (1939) has discussed the vanishing of the absolute vorticity in the case of transition to instability. The vertical component of the absolute vorticity in a zonal current (see Palmen 1948) is given by

$$\eta = -\frac{\partial u}{\partial y} + 2\,\omega \sin \phi + \frac{u}{R} \tan \phi \qquad (1)$$

Here η is the vertical component of the absolute vorticity, y the meridional coordinate (positive northward), ω the earth's angular velocity, ϕ the latitude, u the west wind speed and R the radius of the earth. If the sum of the relative vorticity and the earth's vorticity $(2\,\omega \sin \phi)$ is negative, the motion is dynamically unstable. It appears possible to realize this type of instability on the south side of strong jet streams. This possibility is therefore of interest in our study of jet stream clouds and will be treated in more detail later on.

Another criterion for dynamic turbulence, this time in the zone of cyclonic shear to the north of a wind maximum, has been given by Arakawa (1951). According to his theory the criterical cyclonic shear, again for zonal flow, is given by

$$-\frac{\partial u}{\partial y} = \omega \sin \phi + \frac{2u}{R} \tan \phi \qquad (2)$$

This critical value is often realized in a narrow region (usually the upper frontal zone) just north of well-defined jet streams. Most of the cases of clear-air turbulence (eg. Hislop 1951, Berggren 1952 and Bannon 1952) have, in fact, been from this zone of strong cyclonic shear. The reason why clear-air turbulence and not turbulent-type clouds are usually found here will also be discussed later.

Richardson (1920) has developed a criterion for the increase or decrease of atmospheric turbulence where there is wind shear in the vertical. He postulates that turbulent flow will continue if the rate of supply of energy by eddy stresses is equal to or greater than the work done to maintain the turbulence against any stabilizing forces. The Richardson number is given by

$$R_i = \frac{g}{T}\,\frac{\dfrac{\partial T}{\partial z} + \Gamma}{\left(\dfrac{\partial \bar{v}_x}{\partial z}\right)^2 + \left(\dfrac{\partial \bar{v}_y}{\partial z}\right)^2} \qquad (3)$$

Here g is the acceleration of gravity, T the absolute temperature, $\partial T/\partial z$ the observed lapse rate and Γ the adiabatic lapse rate. The components of the vertical wind shear $\partial \bar{v}_x/\partial z$ and $\partial \bar{v}_y/\partial z$, should be evaluated from vector differences. In this paper the customary approximation

$$\left(\frac{\partial \bar{v}_x}{\partial z}\right)^2 + \left(\frac{\partial \bar{v}_y}{\partial z}\right)^2 \approx \left(\frac{\partial v}{\partial z}\right)^2 \qquad (4)$$

will be used throughout. By computing R_i from vertical cross sections and only in situations where the wind veers but slightly with height, it is assumed that this approximation is not too restrictive. The difficulty of assigning a critical value to the Richardson number is encountered here; different investigators have found values ranging from $^1/_4$ to unity. In figure 1 is presented a graph which permits rapid computation of Richardson numbers from vertical cross sections (dependent, o course, upon our assumption in equation 4). By entering the left hand portion of the graph with lapse rate and vertical wind shear as arguments, one immediately obtains R_i at $T = 20$ C. By first following a constant R_i to the vertical line corresponding to this temperature and then moving horizontally to the right, the corrected R_i at any other temperature

a b

Fig. 2. Cloud photographs taken at Schenectady, New York on 2 March 1953 (a) toward east at 1230 EST and (b) toward northeast at 1300 EST.

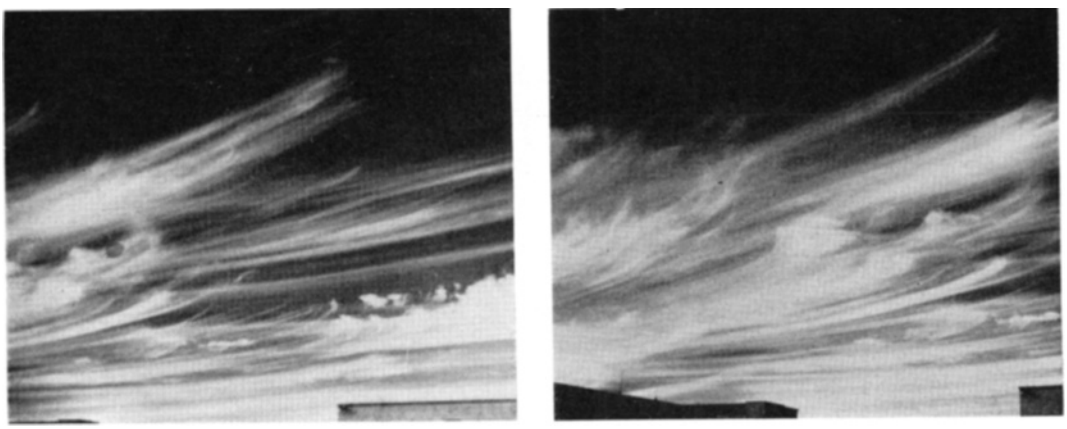

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Fig. 3. Cloud photographs taken at Schenectady, New York on 2 March 1953 at 1645 EST (a) toward north and (b) toward northwest.

is given. This figure clearly illustrates the importance of the vertical wind shear. It is self-evident that one should expect to find low Richardson numbers in the vicinity of jet streams; it is impossible to have a real jet without having both horizontal and vertical wind shear. The distribution of R_i with respect to well-defined wind maxima will be discussed for a typical case.

Through the use of the stability criteria which have been summarized above one might, therefore, hope to arrive at an explanation for the "turbulent motion" cloud types which have been observed in connection with jet streams. At least these three criteria, all based upon dynamic principles, offer a starting point in our investigation.

Turbulence and cloud formation

Obviously instability and the resulting turbulent motion do not necessarily mean that clouds will form. One needs only to look at the numerous reports of "clear-air turbulence" to realize that such is the case. Several factors seem to be of importance in this connection. The data collected by Barrett and others (1949) using the dew-point hygrometer demonstrate the complexity of the moisture

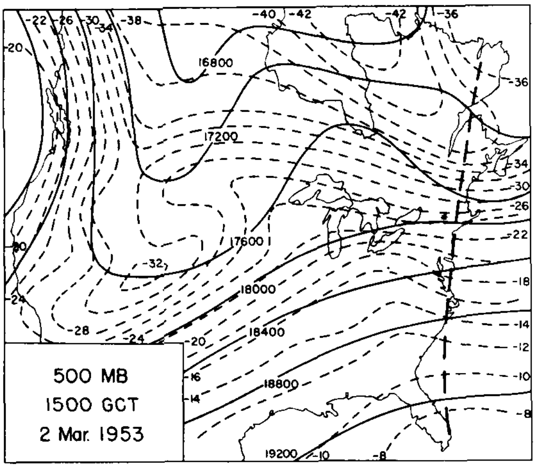

Fig. 4. 500-mb chart, 1500 GCT, 2 March 1953. Heavy solid lines, height contours at intervals of 400 feet. Thin dashed lines, isotherms at intervals of 4° Centigrade. Heavy dashed line, location of cross section. Star shows location of Schenectady, New York.

distribution at high levels in the atmosphere. Frequently the air is so dry that turbulent ascent clearly is unable to bring about condensation.

Vuorela (1953) has recently compleetd a study of the dew-point differences around jet streams over western Europe. His cross sections indicate that there is usually dry air on the cyclonic shear side of the jet stream in the region where clear-air turbulence is most frequently reported. There are exceptions, of course, and he has found cases where a narrow tongue of moist air extends vertically along the warm-air side of the polar front into the zone of strongest cyclonic wind shear. A few cases of jet stream clouds have also been observed here, but they are exceptions to the general rule.

Schaefer has noted that when jet stream clouds are present the sky to the northward is usually clear, and the visibility in this direction is extremely good. This suggests that the lateral circulation about the jet stream is thermally direct in these conditions. It has been shown by Riehl et al. (1952) that a lateral circulation of this sense should be found in the entrance zone of moving jetlets. Since the clouds discussed here are frequently visible from the ground, we can also conclude that the overall ascending motion cannot have persisted long enough to bring about condensation in a deep atmospheric layer. If, however, one or several thin layers were brought close to saturation by some large-scale ascending motion, condensation could then occur in the turbulent wave crests.

Flight observations in jet streams

During the past two years some extremely interesting first-hand observations of the physical nature of jet streams have been described by a senior pilot of the British Overseas Air Corporation. In two papers, Captain Bernard Frost (1953, 1954) has described his procedure for locating a jet stream maxima in the region between 18,000—25,000 feet. The first of his papers describes his temperature gradient method for approaching a jet stream. In his words, "The most valuable instrument used in locating the jet stream proved to be the outside air thermometer. — — — It was found that once in a strong wind at a certain level by flying exactly downwind the temperature remained unaltered and the wind speed remained unchanged. Further, an amazingly accurate guide for calculation of wind strength on either side of the jet stream (within altitude limits normally flown, viz., 15,000—25,000 feet) was that the wind *decreased some 8 knots for every degree Centrigrade drop in temperature on the Polar (or cold) side and it decreased some 16 knots for every degree Centigrade rise in temperature on the Equatorial (or warm) side.*" Frost's first paper is illustrated with pictures of jet stream cloud types which are classic forms according to Schaefer. In his second paper, Captain Frost describes further and more detailed observations with special reference to cloud forms observed. His earlier observations and conclusions supplemented by those of his second paper are in complete agreement with the observations reported by Schaefer.

By private communication a detailed critique of Frost's original paper was presented by a group of Pan American pilots and navigators. Their experiences are also in complete agreement.

The case of 1—3 March 1953

On the morning of March 1, 1953 the sky over Schenectady, New York was completely cloudless with a deep blue color; the visibility was unlimited. Such clearness of air in eastern New York occurs only a few times a year.

c d
e f
g h

Fig 2 (cont.)

c Wave cloud at cirrus level, 1230 EST, March 2, 1953.

d Ripples in cirrus at 1231 EST, March 2, 1953.

e Ripples and billows at cirrus level, 1232 EST, March 2, 1953.

f Waves and banded cirrus, 1300 EST, March 2, 1953.

g Hole developing in banded cirrus, 1301 EST, March 2, 1953.

h Edge of large wave crest in cirrus, 1302 EST, March 2, 1953.

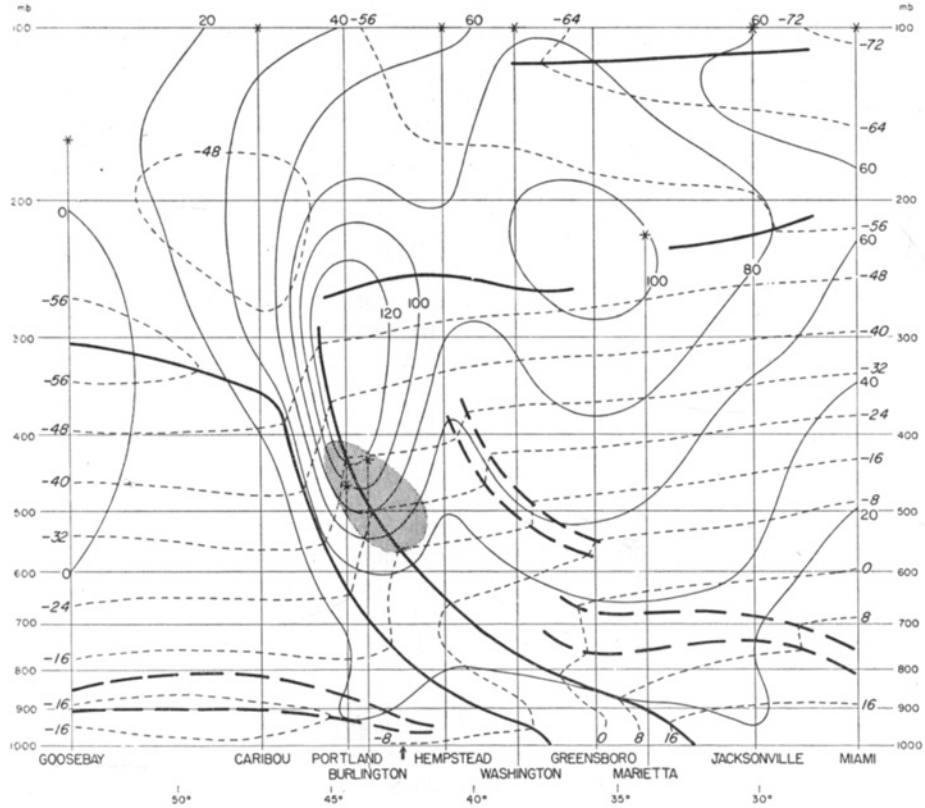

Fig. 5. Vertical cross section from Goose Bay, Labrador to Miami, Florida at 1500 GCT, 2 March 1953, along heavy dashed line in figure 4. Thin dashed lines, temperature (deg. C); thin solid lines, isotachs of observed wind component (knots) normal to section; heavy lines, frontal boundaires and tropopauses; stars, highest level to which wind observations were available; arrow, location of Schenectady. Region where Richardson numbers < unity shown by stippling.

The atmospheric electricity as measured by a point discharge collector was above 0.04 uA throughout the day. In late afternoon a few long lines of altocumulus appeared in the west, illuminated by the setting sun. At 1600 EST the maximum wind speed over Albany, New York was reported to be 220 knots.

On March 2, 1953 a most spectacular display of jet stream clouds was visible from Schenectady. Four large masses of cirrus containing ripples, waves and billows passed over during the day. The following account is taken from the observer's (Schaefer) notes: "These four cloud masses were apparently in the crests of traveling waves having a wave length of approximately 300 miles. Between crests occurred periods in which the sky was nearly cloudless. Such waves have been observed a number of times in association with high-velocity jet streams." In figures 2a through 2h are shown photographs taken shortly after noon on the 2nd of March, while figures 3a and 3b show two photographs taken at 1645 EST on the same day. Note that the earlier pictures show cloudless skies on the left side of the jet stream looking downstream. The atmospheric electricity was well above the level required to class it as Jet Stream Atmospherics.

In figure 4 is shown the 500 mb chart over North America at 1500 GCT, 2 March 1953. From the concentration of isotherms at this level, it is possible to locate a jet stream which extends from the Gulf of Alaska south of a deep trough over the southwestern United States and thence northeastward. There is also evidence of a second isotherm concentration

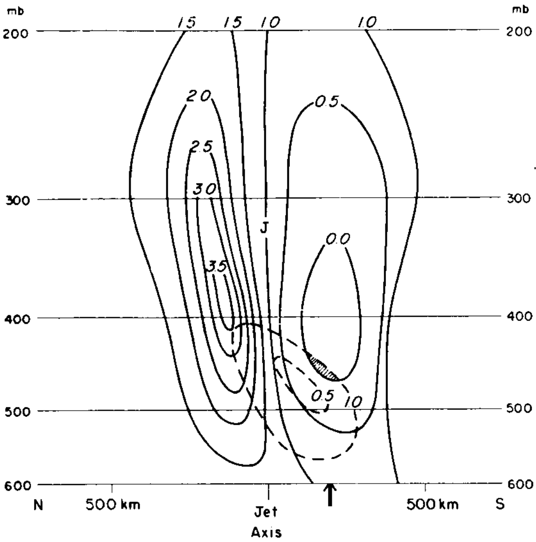

Fig. 6. Vorticity analysis in vicinity of polar-frontal jet in figure 5. Solid lines, component of absolute vorticity normal to constant pressure surfaces (units are multiples of $f = 2\omega \sin \phi$); dashed lines, isolines for Richardson numbers of 1.0 and 0.5. Cross hatching indicates zone where $R_i < 1$ and absolute vorticity < 0 coincide. Location of Schenectady shown by arrow.

extending across southern Canada toward the New England states. The location of Schenectady is shown on the chart by a star.

Figure 5 is a vertical cross section at 1500 GCT along the heavy dashed line in the preceding figure. All stations used in this cross section were projected onto this line using the direction of the 300 mb streamlines; the highest level to which observed winds were available at each station is shown by a small star and the location of Schenectady is depicted by the arrow at the bottom of the section. The isotach analysis (thin solid lines) was based primarily upon observed winds; however, geostrophic winds were used to complete the pattern at high levels over the polar-front jet stream. These isotachs give the wind speed in knots normal to the cross section. We see that there were actually two distinct wind maxima over the Atlantic coast at this time; furthermore, the polar front is strongly defined from the surface up to almost 300 mb. The clouds which were observed over Schenectady at this time were definitely to the south of the polar-frontal jet stream. This means that we can, in this case, neglect the criterion given by Arakawa which is applicable only to strong cyclonic shear. A cursory glance at the figure reveals, however, that there is a zone of strong horizontal wind shear to the south side of this jet stream; in addition, there is also the strong vertical shear which has been shown to be so important in Richardson's criterion.

The stippled area in figure 5 is the region in which equation 3 yields Richardson numbers less than unity. Figure 6 shows this same region of low R_i superimposed upon an analysis of the vorticity distribution around the polar-front jet stream. The vorticity was determined from figure 5 and represents the component of the absolute vorticity vertical to isobaric surfaces; the isopleths are labeled in multiples of $f = 2\omega \sin \phi$. Notice that the minimum Richardson number is less than 0.5, and that negative absolute vorticity is indicated in a zone to the south of the jet axis. It is of particular interest that directly over Schenectady (located by arrow) there is a zone where negative vorticity and $R_i < 1$ actually coincide (hatched area).

On March 3, 1953 the sky was nearly cloudless at sunrise with the air remaining dry and very clear. The sun was veiled with a dense haze of ice crystals. At 0815 EST thin lines of cirrus appeared in the west with masses of altocumulus billows following. This lower cloud deck was at 8,000 feet and the ice-crystal haze was between 30,000 and 38,000 feet. Rapid cloud movements continued into the evening with heavy sleet and freezing rain falling at Schenectady between 2200 and 2300 EST.

Conclusions

In the case studied here, Schenectady was clearly in the jet stream entrance zone and such is frequently the case when jet clouds are visible from the ground. It seems likely, therefore, that the ascending motion on the anti-cyclonic-shear side of the entrance due to the lateral circulation about the jet is important in transporting moisture to high levels in the atmosphere.

Examination of a number of jet stream cases has shown that most wind maxima associated with a strong polar front have an area of low Richardson's numbers in approximately the same location as shown in figure 5. Values as low as 0.15 have been computed. Furthermore, synoptic experience indicates that the absolute vorticity is frequently negative on the right-hand side (looking downstream) of these jet streams. The fact that this area of dynamic instability overlaps the region of low Richardson's numbers in the majority of cases where jet stream clouds are observed may be significant. Turbulent motion could apparently account for the small-scale details inherent in the macro-structure, provided the air in a shallow layer is at or near the saturation point.

Lapse-time photographs of jet stream clouds occasionally reveal a rolling motion on a somewhat larger scale. Motion about horizontal axes parallel to narrow oceanic currents has already been postulated by ROSSBY (1954). It seems possible that this phenomenon may be detected in the atmosphere through the use of the lapse-time camera. It is hoped that the movies now being taken in the western United States will provide sufficient data for further investigation into the fine structure of jet streams.

REFERENCES

ARAKAWA, H., 1951: Possible heavy turbulent exchange between the extratropical tropospheric air and the polar stratospheric air. *Tellus*, **3**, pp. 208—211.

BARRETT, E. W., HERNDON, LEE R., and CARTER, HOWARD J., 1949: A preliminary note on the measurement of water vapor content in the middle stratosphere. *J. Meteor.*, **6**, pp. 367—368.

FROST, B., December 1953: Flying in jet stream winds. *Shell Aviation News*, London, No. 186.

— September 1954: More about the Jet Stream. *Shell Aviation News*, London, No. 195.

PALMEN, E., 1948: On the distribution of temperature and wind in the upper westerlies. *J. Meteor.*, **5**, pp. 20—27.

RICHARDSON, L. F., 1920: The supply of energy from and to atmospheric eddies. *Proc. Roy. Soc.*, (A), **97**, pp. 354—373.

RIEHL, H., et al., 1952: Forecasting in middle latitude. *American Meteorological Society Monograph*, No. 5, 80 pp.

ROSSBY, C.-G., 1953: A comparison of current patterns in the atmosphere and in the ocean basins. *Proc. U.G.G.I., 9th Gen. Assembly, Brussels, Autumn 1951*.

SCHAEFER, V. J., 1953: Cloud forms of the jet stream. *Tellus*, **5**, pp. 27—31.

SOLBERG, H., 1936: Le mouvement d'inertie de l'atmosphère stable et son rôle dans la théorie des cyclones. *Procès-verbaux de l'assoc. de météor., Un. géod. géophys. Inst. Edinbourg*, pp. 66—82.

VUORELA, LAURI A., 1953: On the air flow connected with the invasion of upper tropical air over Northwestern Europe. *Geophysica*, **4**, No. 3, pp. 105—130