A QUANTITATIVE STUDY OF THE "BRIGHT BAND" IN RADAR PRECIPITATION ECHOES

By Pauline M. Austin and Alan C. Bemis

Massachusetts Institute of Technology[1]

(Manuscript received 1 July 1949)

ABSTRACT

Observations and measurements of the bright band in radar echoes are presented. It is shown that the theory of coalescence and melting of snowflakes provides an adequate explanation of the phenomenon.

1. Introduction

When a radar system obtains echoes from an extensive volume of precipitation which is at or near the freezing level in the atmosphere, it is common to observe an intensification of the echoes from a horizontal layer at or just below the freezing level. The phenomenon is most clearly presented on radars which have an RHI (Range-Height Indicator) type of scope. Fig. 1 is a photograph of such a scope with dials showing the time, date, and direction in which the radar antenna is pointing. The vertical lines are 10-mile range markers. Thus the picture presents a vertical cross-section through the atmosphere extending 60 miles eastward from Cambridge and upward to approximately 35,000 ft. The bright spot in the lower middle of the picture occurs at one edge of the RHI scope and presents the echo from the ground and buildings in the immediate vicinity of the radar site. All of the fuzzy echo indicated on the lower part of the scope is from uniform, light precipitation. At about 7000 ft, one fifth of the distance from the bottom to the top of the vertical section, a horizontal "bright band" shows very clearly within the general precipitation echo.

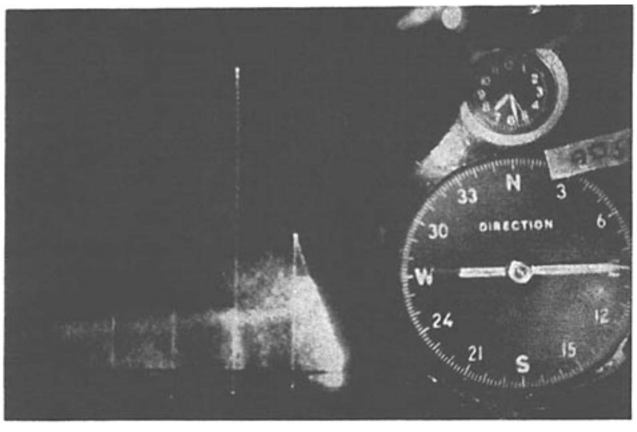

FIG. 1. Example of a well-defined bright band.

The "bright band" was first observed during wartime operations, and has been discussed in the literature by Ryde (1946), Bemis (1947), Byers and Coons (1947), Cunningham (1947), and Marshall *et al.* (1947). The purposes of this article are first, to define exactly what is meant by the term "bright band"; second, to present a complete explanation of the phenomenon; and third, to support the explanation with a few quantitative measurements.

2. Definitions and assumptions

Before proceeding further, several terms need definition. The 0C isothermal surface in the atmosphere will be referred to as the "freezing level," although the term does not accurately describe that surface. The letters "RHI" will ordinarily refer to the Range-Height Indicator or scope which, in the cases cited, presents the information from a TPS-10A radar. "Radar reflectivity" is a measure of the fraction of electromagnetic energy back-scattered toward the radar system by precipitation or other particles. The reflectivity per unit volume is defined as the sum of the scattering cross sections of the particles within a unit volume, and is therefore the effective scattering cross section of a unit volume of precipitation-filled air.

A most important assumption made throughout this discussion is that in all cases the precipitation rate is the same at all levels from a level just above the bright band to one just below. This assumption immediately places us at variance with Byers and Coons' (1947) explanation. However, it will be demonstrated that the assumption of a constant precipitation rate is in agreement with the observations in all those cases where a bright band was located at or below the freezing level. Stratified radar echoes from above the freezing level are usually associated with ice-crystal or snow clouds, are different in character, and have greater vertical extent. Fig. 2 shows a stratified echo about 3000 ft thick above a bright band development. At close ranges, saturation of the scope destroys all details. Between 20 and 30 miles range, however, the

---

[1] Published as a contribution from the Weather Radar Research project sponsored by the U. S. Army Signal Corps.

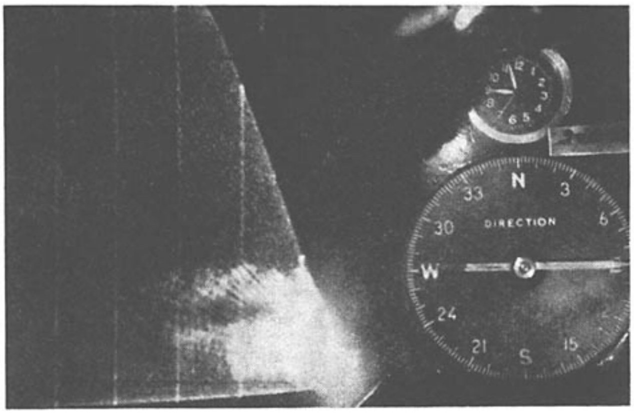

FIG. 2. Example of layer-type echo above bright band.

characteristic narrowness of the bright band is apparent. As one of the conclusions of this study it is suggested that the term "bright band" be reserved for those cases where the intensification of the radar echo is associated with melting snow.

3. Theory

The following explanation of the bright band is an extension of the theory first presented by Ryde (1946) and developed independently by Cunningham (1947). It ascribes the intensified radar echo to the coalescence and melting of snowflakes. Fig. 3 illustrates the contribution of three separate factors to the increased radar intensity in the bright band.

If it is assumed that the particles are all spherical and are of equal mass, the radar reflectivity per unit volume, η, is proportional to $mPv^{-1}(K^2/\rho^2)$,

where m = mass of each particle,
P = precipitation rate,
v = fall velocity of particles,
$K^2 = \left|\frac{n^2-1}{n^2+2}\right|^2$ where n is the complex index of refraction of the scattering material, and
ρ = density of scattering material.

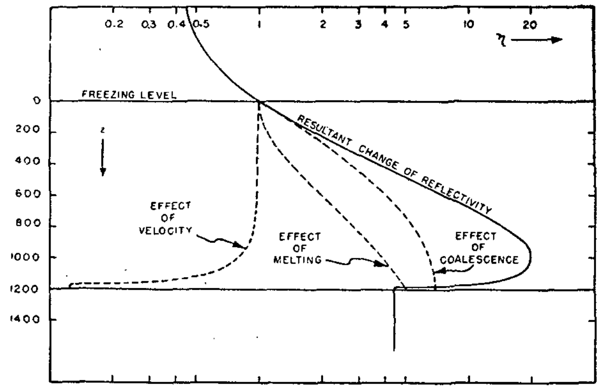

FIG. 3. Schematic diagram of the effect of three factors contributing to the bright band.

At microwave wavelengths, (K^2/ρ^2) is 0.94 for water but only 0.19 for ice. For a mixture of ice and water, where α is the fraction of water present,

$$(K^2/\rho^2) = 0.94\,\alpha + 0.19\,(1 - \alpha).$$

As explained above, it is assumed throughout this discussion that the precipitation rate, P, is constant with elevation within the bright band region and during any single observation. The three variables considered are then (K^2/ρ^2), v, and m.

The rate of melting, which affects the factor (K^2/ρ^2), has been calculated for a spherical snowflake (or cluster of snowflakes) of radius r which is falling through the air with a velocity v m sec^{-1}. The temperature of the environment is T degrees centigrade and that of the melting snowflake is zero. The heat transfer from the environment to the snowflake, in calories per second, is

$$h = 4\pi kcrT = 4\pi kcr\gamma vt$$

where k = heat conductivity of the air = 5.7×10^{-5} cal sec^{-1} deg^{-1} cm^{-1},
γ = lapse rate of temperature in the atmosphere in deg m^{-1}, and
t = time of fall from the freezing level.

The factor c represents an increase in the effective conductivity of the air because of the motion of the snowflake. This factor has been studied empirically by Houghton (1938) in connection with the evaporation of falling drops. The factor varies slowly with the fall velocity of the particle. It has a value slightly greater than two for v = 50 cm sec^{-1} and increases to a value of about three for v = 200 cm sec^{-1}.

The curve showing the effect of melting in fig. 3 was calculated for a snowflake aggregate 5 mm in radius which is falling with a velocity of one meter per second. It is assumed that the resultant raindrop is 1 mm in radius. Rough measurements of the size of snowflake clusters and of the drops they form upon melting showed that the values of 5 mm and 1 mm for radii are reasonable. The calculated distance of 1200 ft between the freezing level and the level at which melting is complete (the bottom of the bright band) is also reasonable when compared with the observed distances between the freezing level and the bright band. The effect of complete melting is an increase in the reflectivity by a factor of five because of the change in the value of (K^2/ρ^2).

The curve showing the effect of fall velocity has been sketched in qualitatively. The important characteristic of this curve is that the change in fall velocity is assumed to be slight during most of the melting process but a sharp increase occurs just as the melting is completed and the wet snowflake becomes a spherical raindrop. No measurements of the fall velocity of wet snowflakes are available. The assumptions are based upon observations of snowflakes and snowflake aggregates as they melt and upon the fact that a sharp decrease in intensity is observed at the base of the bright band. Snowflake clusters seem to maintain the same general diffuse shape during most of their melting period. Apparently they consist of a thin skeleton of ice which is covered by water; when the melting is nearly complete the surface tension of the water causes the ice skeleton to collapse and the melted snowflake assumes a spherical shape rather abruptly. At this time the fall velocity would be expected to increase rapidly, reaching its terminal velocity within a few seconds.

The effect of coalescence upon the fall velocity of the particles is believed to be slight. Nakaya and Terada (1935) and Hooper and Kippax (1947) have shown that large snowflakes do not fall much faster than small ones. Therefore, only a gradual increase in fall velocity is assumed in the upper part of the bright band.

It can be seen from fig. 3 that the increase in the fall velocity of the scattering particles is the only factor which contributes to the decrease in radar reflectivity below the bright band. Since we do observe a decrease, the effect of velocity change must be much as it is shown in fig. 3.

The mass of the particle, m, may in some cases remain constant during the melting time. However, it is observed that at temperatures near 0C falling snow usually contains many clusters or aggregates of individual snow crystals. This coalescence of snowflakes apparently begins at temperatures well below freezing and becomes more pronounced as the temperature increases and the snow begins to melt. The amount of coalescence varies greatly from storm to storm, though in most cases there seem to be at least a few large aggregates which contain between 50 and 100 individual snowflakes, or perhaps even more. Since no definite, quantitative information concerning the effect of coalescence is available, the coalescence curve in fig. 3 is only a rough estimate. It shows the effect of some coalescence above the freezing level. This coalescence continues below the freezing level and is of sufficient importance to cause the resultant reflectivity in the bright band to be twenty times as great as the reflectivity just above the bright band. Measurements of relative radar reflectivity in and above the bright band indicate that a factor in the vicinity of twenty is a common value for the increase in intensity, though it may be as small as five (no coalescence) or well above one hundred.

The combined effect of the three factors described above is shown by the solid curve in fig. 3. The possible contributions of two other factors have also been considered. It has been suggested that the difference in temperature and vapor pressure between the air adjacent to a melting snowflake, whose temperature is 0C, and the environment, which may be a few degrees warmer, would create around each flake a shell of air whose index of refraction is different from the surrounding atmosphere. Such a "shell" would act as a scattering particle and because of its large size, relative to the snowflakes or raindrops, might increase the radar reflectivity. However, the difference in index of refraction between such "shells" and the surrounding air is so slight that the scattering from them should be considerably less than that from the melting snowflakes themselves. Therefore any effect of this type is of minor importance.

A second factor which might affect the radar reflectivity is the non-spherical shape of the snowflakes or clusters. Calculation of the electric moment of spheroids (see Gans, 1912) indicates that this factor should be small when the particles are composed of ice, but may be appreciable if water is the scattering matter. A disc-shaped water particle might return several times as much power as a spherical one of equal mass if it is sufficiently flattened and favorably oriented with respect to the polarization of the beam. If it is unfavorably oriented, the scattering cross section may be only a fraction of that of the spherical drop. However, the theory cannot be applied directly to melting snowflakes because they have a lacy structure and do not necessarily melt into a solid disc of water. It is hoped that simultaneous quantitative measurements on two radars whose beams are polarized in different directions may be made to yield information concerning this effect. However, consideration of this effect has not been included in the present study.

4. Quantities measured and estimated accuracy of measurements

In studying the bright band, an attempt was made to measure the following quantities by means of radar: (1) height of the center of the bright band; (2) vertical thickness of the bright band; and (3) radar reflectivity in the bright band, η_{BB}, above it, η_A, and below it, η_B. The meteorological quantities measured were the height of the freezing level and the lapse rate of temperature in a layer 2 or 3 thousand feet thick, just below the freezing level. Selected cases are listed in table 1 and the accuracy of the various measurements is discussed below.

The height of the center of the bright band was rounded off to the nearest 500-foot level since the uncertainty of the observations is about 500 ft.

It was not possible to make good measurements of the thickness of the bright band because the resolving power of the TPS-10 is not great enough. The shape of the RHI presentation and the ground clutter at

TABLE 1. Summary of bright band measurements on 24 selected days.

(1) Date 1948	(2) Eastern Standard Time	(3) Alt. of freezing level	(4) Alt. of bright band on RHI	(5) Alt. of bright band from radar meas.	(6) Altitude difference ΔZ	(7) η_B $\times 10^5$	(8) $\frac{\eta_{BB}}{\eta_B}$	(9) $\frac{\eta_A}{\eta_B}$	(10) Thickness of bright band	(11) Lapse rate °C/1000′
5/20	1200	2500	1500	—	1000	—	—	—	Vertical thickness of bright band ≥ 1000 ft at 10 miles range	2.5
6/24	1930	14000	(13000)	—	1000	—	—	—		2.5
4/6	1330	8500	(8000)	8000	500	8.0	3.5	0.17		2.0
4/15	1100	7800	(7000)	—	800	—	—	—		2.0
5/7	1830	10600	9500	9500	1100	6.0	6.0	0.32		2.0
5/25	0900	10000	9500	—	500	—	—	—		2.0
6/23	1530	12600	(12000)	12000	600	50	2	1.0		2.0
10/11	2200	10200	9000	9000	1200	1.9	2.5	0.79		2.0
11/10	2200	13600	13000	—	600	—	—	—		2.0
11/19	2200	11600	11000	11000	600	500	10	0.08		2.0
5/22	0830	6700	5500	—	1200	—	—	—		1.5
6/13	1100	11000	(11000)	—	0	—	—	—		1.5
4/1	1600	9500	(9500)	9500	0	1.2	11	1.0	Vertical thickness of bright band <1000 ft at 10 miles range	2.5
7/1	0930	13400	(13000)	12500	900	9.5	12	0.7		2.0
4/12	1100	11400	(11000)	—	400	—	—	—		1.5
4/14	2030	10400	(9000)	9000	1400	38	9	0.40		1.5
5/5	2230	6400	4000	—	2400	—	—	—		1.5
5/14	0030	10800	11000	—	−200	—	—	—		1.5
11/4	0200	10100	9500	9500	600	1.5	6	0.25		1.5
11/4	1400	10200	9000	9000	1200	38	4	0.13		1.5
11/22	1600	4600	3500	—	1100	—	—	—		1.5
11/27	0930	8500	8000	8000	500	78	10	0.06		1.0
5/13	1400	11200	10000	10000	1200	19	8	very small		0.5
5/25	2030	—	9500	9500	—	7.5	11	0.13		—

$$\left.\begin{aligned}\eta_{BB} &= \text{radar reflectivity in the bright band}\\ \eta_A &= \text{radar reflectivity above the bright band}\\ \eta_B &= \text{radar reflectivity below the bright band}\end{aligned}\right\}\text{in mm}^2\,\text{m}^{-3}.$$

close ranges render it difficult to make observations at ranges any nearer than 10 miles. At this range, the apparent thickness of the bright band was often the same as the beam width in the vertical direction (650 ft) or only slightly thicker (Hooper and Kippax, 1947, found an average thickness of 750 ft.). However, in order to deduce the effect of meteorological conditions upon the thickness of the bright band, the cases studied were divided into two groups: (1) the "narrow" ones, which appeared to be less than 1000 ft thick at a range of 10 miles; and (2) the "wide" ones which seemed to be 1000 feet or more in thickness at 10 miles.

The radar reflectivity was measured by an instrument called a Pulse Integrator attached to the SCR/615-B radar (Weather Radar Research, 1946). The beam width on this radar is 4 deg, about six times as great as that of the TPS-10 in the vertical direction. The observations were made by taking readings at a number of elevation angles, keeping the range constant. The range chosen was usually between 10 and 15 miles, though in some cases it was necessary to use larger ranges. Because of the width of the beam (3700 ft at 10 miles), the reflectivity measured was usually a composite one since only a fraction of the beam was filled by the bright band, another fraction by the rain below it and the remaining part by the snow above. In deducing the reflectivity within the bright band from this composite picture, it was assumed that the thickness of the bright band was 1000 ft. The height of the bright band was assumed to be that measured on the RHI scope, though in a few cases this had to be adjusted somewhat in order to make the measurements at different elevation angles appear consistent.

The accuracy of these reflectivity measurements is difficult to estimate, but a few remarks concerning it should be made. It has not been possible to obtain a standard suitable for checking the absolute values of the Pulse Integrator measurements. The most optimistic estimate is an uncertainty of 3 db, which is equivalent to a factor of two, but the uncertainty may be as large as a factor of ten. Relative measurements, however, are much more reliable. Comparison of readings made at one sitting should be accurate to 0.5 db (10 per cent) while comparison of measurements taken on different days may be in error by about 1 db. In summary, while the absolute values of η_B, column (7) table 1, are dubious, the comparative values are an indication of the relative intensities of precipitation on the different days. The measurements used in obtaining the ratios given in columns (8) and (9) are accurate within about 10 per cent, but some further uncertainty is introduced in calculating the fraction of the beam filled by the bright band. In making this calculation, it was assumed that the bright band was 1000 ft thick in every case. Therefore, if the bright band was narrower than 1000 ft, the relative intensity (η_{BB}/η_B) was actually greater than the calculated value, while the relative intensity was less than the calculated value for the cases where the bright band was greater than 1000 ft in thickness. Moreover, the reflectivity above the bright band usually shows a continuous decrease with increasing elevation so that η_A must be considered as the average reflectivity for several thousand feet above the bright band. In considering all these factors, it appears that the values of η_{BB}/η_B and η_A/η_B should be accurate only to about 30 per cent.

The meteorological data were obtained by interpolation between radiosonde ascents at Nantucket, Portland, and Albany. The times of observation of the bright band were chosen as close as possible to the times of the ascents and cases were eliminated where fronts in the immediate vicinity or other meteorological irregularities made interpolation especially untrustworthy. However, it is believed that the values of the heights of the freezing level may be in error by plus or minus 500 ft. The lapse rates were estimated to the nearest 0.5C/1000 ft.

5. Results

To obtain information concerning the frequency of appearance of the bright band, all of the radar photographs for the year 1948 were examined. Photographs of precipitation were taken on 115 days, and these may be divided into the following groups:

- (a) Snowing on the ground (no bright band) — 30 days
- (b) Layer aloft (but no bright band as here defined) — 11 days
- (c) Raining on the ground — 74 days
  - (1) Bright band observed — 38 days
  - (2) Thunderstorms — 21 days
  - (3) Low-level rain (no bright band) — 9 days
  - (4) No RHI photographs taken — 6 days

In the case of the thunderstorms or convective showers, it was found that a bright band does not appear in the developing phase, but usually can be observed during the dissipating stages. Byers and Coons (1947) were the first to make this observation. When the bright band first appears in a thunderstorm, it is very hazy and several thousand feet thick. Then it gradually becomes narrower and better defined. There was no attempt to make quantitative measurements on the bright bands in dissipating thunderstorms because they were continually changing in thickness and usually lasted only a few minutes.

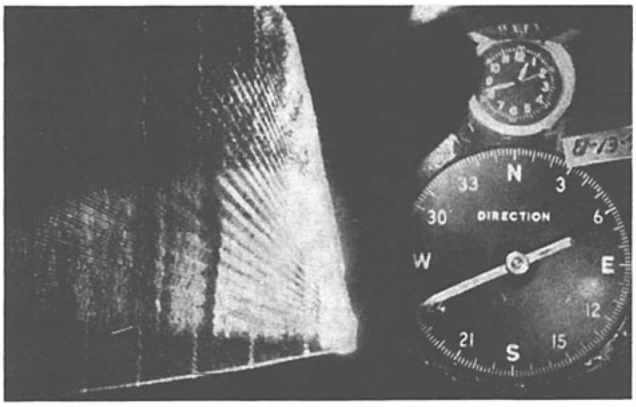

FIG. 4. Example of bright band in the rear of a thunderstorm.

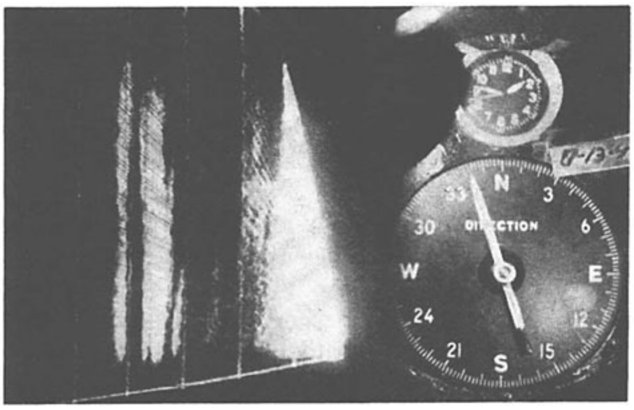

FIG. 5. Example of bright band development near an active thunderstorm.

Figs. 4 and 5 are two photographs of thunderstorms taken on the same day, 13 July 1948. Fig. 4, taken at 1242 EST, shows a group of showers in which the bright band is fairly well established although it is broad and fuzzy. The snow echoes above the bright band are relatively faint and cannot be seen at ranges greater than 30 miles. The maximum range at which the rain below the bright band can be detected is about 65 miles. This difference in maximum detectable ranges indicates that the reflectivity of the rain is about five times that of the snow. A photograph such as this one, especially if there happened to be no showers at close ranges, might suggest that sometimes there is no precipitation above the bright band. It is observed, however, that whenever the bright band can be seen at close ranges, there is always some indication of precipitation above it. Fig. 5 was taken about an hour later and shows a shower extending to 20 miles in which a wide bright band appears. Two tall thunderstorms appear at ranges between 30 and 45 miles. Although they extended well above the freezing level, no indication of a bright band could be found at any gain setting. These are typical echoes from active thunderstorm cells.

In the cases labelled "low-level rain," the RHI photographs showed no bright band but a uniform top to the precipitation echoes at a level either in the vicinity of or slightly below the freezing level. Although these cases may demonstrate some bright band characteristics, they were not considered in this study.

A summary of the measurements made on the 38 days when a persistent bright band was observed is presented in table 1. This table includes data taken on 24 days, the other 14 days having been eliminated either because the meteorological data were incomplete or because the height indication on the radar was not properly calibrated.

The times in column (2) of table 1 are the times of the radar observations; the radiosonde ascents were made at approximately 1030 and 2230 EST each day.

In column (4) the heights in parentheses correspond to days when the RHI calibration is somewhat doubtful. The altitudes given in column (5) are calculated from measurements made on the Pulse Integrator on the 615-B radar. These serve as a check on the values in column (4) which were measured on the RHI scope.

The quantity ΔZ in column (6) is the distance between the freezing level and the center of the bright band. It should be an indication of the average distance the snowflakes fall during melting. Therefore, one might expect that ΔZ would be greater for heavy precipitation which contains large drops than for light rain. Also it might be expected that ΔZ would be smaller when the lapse rate is steeper. However, there was no significant correlation between ΔZ and any of the other quantities measured; probably this lack of correlation was due to the fact that the uncertainty in measuring the heights of the freezing level and of the bright band is of the same order of magnitude as ΔZ itself. The average value of ΔZ is 830 ft. This is considerably greater than the average value of 310 ft obtained by Hooper and Kippax (1947).

As seen in column (10), the bright bands are divided into two groups, the "wide" ones and the "narrow" ones. Within each group the cases are arranged in order of decreasing lapse rate, as indicated by column (11). Comparison of the lapse rates in the "wide" group with those in the "narrow" group shows that the bright band tends to be narrow and well defined in the more stable situations and is wide and fuzzy when the lapse rate is steep. In very unstable situations, where convective showers and thunderstorms occur, the bright band does not appear at all or else is extremely wide. Moreover, comparison of the relative intensities in the two groups (η_{BB}/η_B, column (8), table 1) shows that the narrow bright bands are usually more intense than the wide ones. The observations suggest that the stronger vertical currents associated with unstable situations prevent the establishment of a precise and well-defined melting level such as can be observed in a more stable and stratified atmosphere.

In order to consider the relation between the stability and the amount of coalescence, η_{BB}/η_A and γ are shown in table 2. This table suggests that there is a greater amount of coalescence with the more stable lapse rates.

The measurements of the thickness of the bright band and its distance below the freezing level which were made by Hooper and Kippax (1947) were obtained by pointing the antenna vertically upwards. The echoes were then observed on an R-scope, which presents the intensity of the received signal as a function of range. Fig. 6 shows a photograph of an observation of this type on the new Signal Corps radar which was especially developed for weather purposes. In this case the resolving power of the instrument depends upon the length of the pulse instead of the beam width. For the case illustrated, this is 0.5 microseconds which corresponds to a wave train about 500 ft in length. In fig. 6 the range markers are at one mile intervals, the total range being five miles. The rain echoes extend from the ground up to 8000 ft where a sharp rise in signal intensity indicates the position of the bright band. Above this level the intensity falls to a value somewhat less than that corresponding to the rain. The apparent thickness of the bright band is about 500 ft. It is interesting to note the difference in character between the echoes from the rain below the bright band and the snow above it. The rain echoes are of a very fine lacy structure and fluctuate very rapidly, while the snow echoes show a much coarser structure and the fluctuations are relatively slow. The explanation for this phenomenon

TABLE 2. Comparison of η_{BB}/η_A and γ.

Date 1948	Eastern Standard Time	η_{BB}/η_A	γ
5/13	1400	large	0.5
11/27	0930	170	1.0
11/4	1400	31	1.5
4/14	2030	23	1.5
11/4	0200	25	1.5
11/19	2200	100	2.0
4/6	1330	21	2.0
5/7	1830	18	2.0
7/1	0930	16	2.0
10/11	2200	3	2.0
6/23	1530	2	2.0
4/1	1600	11	2.5

η_{BB} = radar reflectivity in the bright band.
η_A = radar reflectivity above the bright band.
γ = lapse rate in °C/1000 ft.

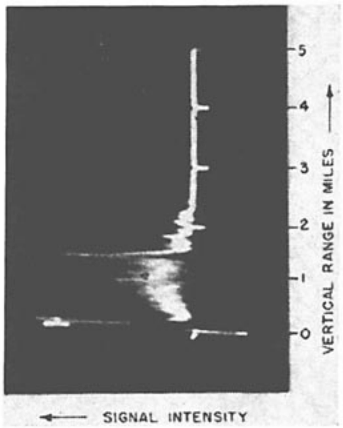

FIG. 6. Bright band as presented on R-scope when the antenna is pointing vertically.

may lie in the fact that large differences in fall velocity exist between raindrops of different sizes so that a rapidly changing interference pattern is produced. However, snowflakes do not have such a wide range of fall velocities. Therefore their interference pattern changes more slowly.

6. Conclusions

A general conclusion from this study is that the bright band associated with the freezing level is caused by the coalescence and melting of snowflakes and not by the formation of any new particles. This melting theory is shown to be an adequate explanation. As proof of its generality we have the fact that a bright band may always be observed when precipitation falls through the freezing level except when strong convective action disorders the melting process. This conclusion is in agreement with Ryde (1946) and in disagreement with Byers and Coons (1947). "The idea of drop formation in the colloidally unstable layer of heterogeneous ice-water mixture" which the latter authors associate with the bright band should occur at temperatures several degrees below freezing and not at or below the freezing level. It is suggested that the term "bright band" (or "bright line") be reserved, by definition, for intensification of the radar echo due to coalescence and melting of snowflakes. Because the bright band, as defined above, may usually be distinguished from other layer-type radar echoes, its existence and location is of considerable meteorological importance. It indicates snow above and rain below, an observation which should assist in analysis of aircraft icing conditions aloft. Its thickness bears a roughly inverse relationship to stability. Its location gives an approximate but continuous measure of the altitude of the freezing level. The coalescence processes which occur within the bright band region have considerable effect on the precipitation reaching the ground since they transform many small particles to a few large ones. It seems desirable therefore to attempt more accurate radar and meteorological measurements in order to permit a more detailed analysis of the phenomenon and its relation to meteorological problems.

REFERENCES

Bemis, A. C., 1947: Weather radar research at M.I.T. *Bull. Amer. Meteor. Soc.*, **28**, 115–117.

Byers, H. R., and R. D. Coons, 1947: The 'bright line' in radar cloud echoes and its probable explanation. *J. Meteor.*, **4**, 75–81.

Cunningham, R. M., 1947: A different explanation of the 'bright line'. *J. Meteor.*, **4**, 163.

Gans, R., 1912: Uber die Form ultramikroskopischer Goldteilchen. *Ann. Physik*, **37**, 881–900.

Hooper, J. E. N., and A. A. Kippax, 1947: Interim report on measurements of radar echo intensities from rain and snow. *TRE Report T-2082.*

Houghton, H. G., 1938: A method for the local dissipation of natural fog by means of sprayed calcium chloride solution. *Pap. phys. Ocean. Meteor., Mass. Inst. Tech. and Woods Hole ocean. Instn.*, **6**, No. 3.

Marshall, J. S., R. C. Langille, and W. McK. Palmer, 1947: Measurement of rainfall by radar. *J. Meteor.*, **4**, 186–192.

Nakaya, U., and T. Terada, Jr., 1935: Simultaneous observations of the mass, falling velocity and form of individual snow crystals, *J. Faculty of Science*, Hokkaido Imperial University, Series II, **1**, No. 7.

Ryde, J. W., 1946: The attenuation and radar echoes produced at centimetre wave-lengths by various meteorological phenomena. Report of a conference on *Meteorological factors in radio-wave propagation* held on 8 April 1946 at The Royal Institution, London. Published by The Physical Society.

Weather Radar Research, 1946: *First Technical Report under Signal Corps Project*, Dept. of Meteor., Mass. Inst. of Tech., Cambridge.