RADIOMETEOROLOGICAL ANALYSIS OF TORNADOES
BY RADAR PRECIPITATION ECHOES

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RADIOMETEOROLOGICAL ANALYSIS OF TORNADOES

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PREFACE

Research on tornado indentification and tracking by radar at the Oklahoma Institute of Technology is in the beginning process of development, because the radar set has been in operation only since last October. Conclusions can not yet be drawn, because observational data obtained thus far are not complete enough to yield a statistical result.

Since the problem as a whole is a radiometeorological one, the investigators of the problem must have a through understanding of both fundamental meteorology and the field of radar wave propagation in radio engineering. With this in mind, the author has tried to present the most important basic information and discussion in both phases, so that those who continue the work may have a proper guide in comprehending the problem.

Suggestions have been made to improve the present sferic tornado detection system and radar tornado detection system.

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CHAPTER I

RADIOMETEOROLOGY

With the recent advance of radio technique the science of radiometeorology can be generally classified into two phases: namely, the sferic observation, and microwave propagation and its storm observation.

Sferics is a contraction of the word atmospherics meaning natural electrical phenomena in the atmosphere detected by radio methods. Sferics have variously been known as clicks, grinders, sizzles, strays, parasites, and other names. They comprise natural static which interferes especially with amplitude-modulated radio wave reception. Sudden electrical discharges resulting in redistribution of charges within and between clouds, between clouds and air space above or below, and between clouds and earth, gives rise to electrostatic induction fields and to radiation fields. It is the last which principally forms sferics at a distance, and which is of particular concern in radio-meteorology. Sferics, in radiometeorology, are significant due to their origin in relatively intense convection intimately involving water vapour. Sferics are positively correlated with cyclones and thunderstorms, and lightning seems necessarily to be the source of sferics.

The observation of sferics has taken several forms; display of the wave form of individual sferics, measurement of intensity and rate of occurrence, and observation of bearings of sources of sferics. For more than twenty years, the radio direction finder with cathode ray tube indication, first described by R. A. Watson-Watt[1] and J. F. Herd, has been used in the study of the direction of arrival of sferics. The equipment uses a pair of crossed-loop receiving aerials connected to twin balanced amplifiers, the outputs from which are connected to the two pairs of plates of the cathode ray tube. The arrival of a sferic causes a line or narrow ellipse to be traced on the screen of that tube in a direction corresponding to that of the source of disturbance at which the sferic originated. A tube with a long after glow flourescent screen is used to enable the observer to make accurate readings on flashes of very short duration.

Research on tornado detection by sferic direction finder has been and is being carried on here at Oklahoma Institute of Technology, and recently the United States Army Signal Corps AN/GRD-1A, static direction finder, has been put into operation to strengthen this phase of tornado indentification.

Reasearch on tornado sferics by waveform analysis has been underway at O. I. T. since 1947. The equipment has been modified and redesigned[2] according to the analysis of experimental results and information from other workers. A vertical whip

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[1]R. A. Watson-Watt, "The Directional Recording of Atmospherics," <u>Journal Institution of Electrical Engineers</u>, LXIV (1926), p. 596.

[2]P. N. Hess, <u>Installation and Operation of Electronic Sferic Detection Equipment</u>, Master of Science Thesis, Oklahoma Agricultural and Mechanical College (1950).

antenna is used to pick up the incoming sferics and is by-passed to the trigger circuit and high-gain vedio amplifier through the antenna cathode follower which was inserted to match the high impedance of the relatively short antenna. The amplifier has been designed to have linear amplification over a broad range of frequency around 400 kilocycles, which is the probable range of tornado aferic frequency. The output of the amplifier is connected directly to the vertical input of the cathode-ray oscilloscope.. The main purpose of the trigger circuit is to "trigger" the oscilloscope into operation and to actuate the camera in front of the scope. It has been called to the attention of the author that the time delay in the triggering response of the trigger circuit, now in use at O. I. I., by incoming sferics is of the order of 500 microseconds[3] which is comparable to the total duration of a lightning discharge. Hagenguth[4] has shown that the lightning stroke of ordinary thunderstorms may consist of a number of current peaks and continuous current flow. The statistical evidence shows that the duration of the current peaks is of the order of 20 to 40 microseconds, while the total duration of the stroke varies between 0.0006 sec. and 0.35 sec. The longest duration measured is about 1.5 sec. Fifty per cent of the strokes have two or more current peaks, while a maximum of 42 current peaks has been measured.. The time interval between successive current peaks

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[3] Ibid. p. 18.
[4] J. H. Hagenguth, "The Lightning Discharge," Compendium of Meteorology (1951), p. 140.

varies between 0.02 sec. and 0.09 sec., with a maximum of 0.5 sec. between successive discharges. Ligada[5] has also shown by radar echo that the duration of a lightning stroke of ordinary thunderstorm is about 0.02 sec. It is therefore clear from above statistics that the duration of tornado sferics is also very short, possibly a little longer than that of ordinary thunderstorm sferics, and accordingly that period of time in actuating the trigger circuit into operation is considerable in comparison with the duration of a whole sferic. This means the initial display of sferic wareforms will always be missed. It is quite possible that this very initial part of the sferic waveform is the part which will enable the differentiation of tornadoes from ordinary thunderstorms as well as the differentiation of an incipient tornado from a well-developed tornado. In order to have the full advantage of complete display of sferic waveforms, the author strongly recommends the re-design of the trigger circuit with an almost instant response.

The results of the preliminary investigation of the sferics show that electric discharges from tornadoes are much different from those observed during ordinary thunderstorms. The rate of occurrence of sferics during the progress of a tornado was greater than that of a thunderstorm and the amplitudes of sferics were on the average considerably greater. Furthermore, it was noticed that the tornado sferics contained higher frequency components than those of ordinary thunderstorms. The waveforms

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[5]M. G. Ligada, "Lightning Detection by Radar," Bulletin of the American Meteorological Society, XXXI (1950), pp. 279-283.

of the observed tornado sferics are now being studied by use of a sferic waveform analyzer and will be grouped with future results to yeild statistical information.

The study of the propagation of microwaves in the troposphere undoubtedly leads radio engineers to feel the necessity of thorough understanding of meteorological phenomena, and at the same time, through the use of radar for precipitation detection, the science of meteorology has acquired an entire new and unique method of weather observation. As a result of that interdependence of these two sciences, the science of radiometeorology becomes more important than it has ever been.

Radar shows graphic, dynamic, and up-to-date depiction of precipitation formations of all types, and these in several dimensions. Techniques for analysis of radar precipitation echo signals have not yet been completely developed, and for that reason it appears that radar has vast potentialities both in the fields of physical meteorological research and weather observation and forcasting, as well as other closely allied activities.

Fundamentally, radar is a kind of radio instrument which produces powerfully beamed microwaves propagating in the lowest layer of the atmosphere--the troposphere. It is therefore the intention of the author to present first a brief and basic study of both the troposphere and the propagation of microwaves in that layer of atmosphere. Following these basic presentations it will be expedient to bring some discussions and investigations on the physical characteristics of tornadoes and the feasibility of using radar to detect tornadoes--the most violent yet least extensive of all storms.

However, it should be clearly borne in mind that this system of tornado detection by sferic analysis and radar precipitation echo analysis can not be successfully operated without constant and continuous information about weather changes. When the results of sferics and radar precipitation echoes are analyzed, they must be coordinated with synoptic analysis to give the exact and right interpretations. This requires close cooperation between meteorologists and radio engineers. Unlike the ordinary daily weather forcast, reliable specific tornado forecasts can not be made at the present time. At best, conditions may be found which are favorable for the development of tornadoes over a wide area. Although the technique of forecasting the exact place and time a tornado will strike has not yet been perfected, Lloyd[6] and others realized that, once a tornado had formed, its future course could be predicted with reasonable accuracy from a knowledge of winds aloft in the warm air mass.

For more than fifty years the United States Weather Bureau has been issuing warnings of severe local storms within the next 24 hours without mentioning tornadoes specifically. In some midwestern cities, the Bureau has considered plans for the detection and tracking of tornadoes by telephone, by short-wave radio sets with independent power supplies, and on radarscope. However, recently the Bureau has been successful in locating in advance the general and approximate area of a tornado, such

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[6] J. R. Lloyd, "The Development and Trajectories of Tornadoes," Monthly Weather Review, LXX, (1942), pp. 65-75.

as the one on March 21, 1952 in Arkansas and neighboring states. Due to the lack of certainity as to the exact time and location of tornadoes, no effective and definite precautions can be taken. Because of uncertainty, some false warnings have been given by the Bureau in tornado forecasts.

It is the purpose of the research at Oklahoma Institute of Technology to cross-check the exact time and location of tornado ocurrence within the range of O. I. T.'s detection system. Once a tornado-like thunderstorm is tracked by radar, its path of attack can be easily traced from radarscopes. To distinguish a tornado from a severe thunderstorm, as well as an incipient tornado from a well-developed tornado, by radar precipitation echoes it is necessary to obtain statistical results from the analysis of precipitation echo pictures and to have a complete understanding of the theory of formation of tornadoes. Tepper[7] has thrown some light on radar and synoptic analysis of tornadoes with his theory of the formation of of tornadoes by pressure jump lines.

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[7]Morris Tepper, "Radar and Synoptic Analysis of a Tornado Situation," Monthly Weather Review, LXXVIII (September, 1950), pp. 170-176.

CHAPTER II

A STUDY OF ATMOSPHERE AND ITS VARIOUS DISTURBANCES

The atmosphere surrounding the earth is a mass or body of gases, chiefly a mixture of nitrogen and oxygen. By volume, the average proportion of these gases at sea level is about 78 per cent nitrogen and 21 per cent oxygen. The balance of 1 per cent is divided among several gases, including argon, carbon dioxide, helium, hydrogen, and neon. Water vapour, another important constituent of the atmosphere, is always present although the amount varies greatly from time to time, depending upon the temperature and location. It is because of the variability in the amount of water vapour , which may consist of only a trace or as much as over 4 per cent by volume, that the composition of the atmosphere is given for dry air. Dust consists of particles of all kinds of matter distributed by wind and is always present in the atmosphere in variable amounts.

The atmosphere extends upward around the surface of the earth for many hundreds of miles, although its exact limits are unknown. The air has weight and therefore exerts pressure, equally in all directions and not merely downward. The atmospheric pressure at any height is proportional to the weight of the air above that height and therefore atmospheric pressure decreases with altitude. In dynamic units, atmospheric pressure is expressed in millibars. A millibar is one thousandth of a bar. A bar is equal to 1,000,000 dynes per square centimeter. One bar, or 1,000 mb., equals 29.53 inches. One inch equals 33.864 mb. Sea-level pressure of 76 cm., equal to 29.92 inches, is equivalent to 1,013.25 mb. Assuming that the density of the atmosphere at sea level equals 100 per cent, at 19,000 feet it is 50 per cent, and at 39,000 feet it is 25 per cent. In other words, one-half of the mass of the atmosphere is below 19,000 feet and three-quarters is below 39,000 feet.

Conventionally the atmosphere may be divided into three main layers--troposphere, stratosphere, and ionosphere. The name troposphere has been applied to the lowest layer, extending to a height of from 5 to 8 miles. In the troposphere the temperature decreases at the lapse rate of about 3.5^{o} F. per thousand feet of elevation. The term "lapse rate," or vertical temperature gradient, is the change of temperature with altitude ($-dt/dh$). The troposphere was so named because of this regular change of temperature. This lowest layer of air contains the clouds, practically all the water vapour, and 80 per cent of the total mass of air. All the atmospheric disturbances like rain, storms, etc. occur in this layer. And it is this layer in which modern radiomen have become interested due to the rapid development of the knowledge of microwave propagation during World War II.

The second layer of the air is called the stratosphere because it has a nearly uniform temperature and freedom from storms. The upper limit of the troposphere is named tropopause, since the temperature no longer decreases regularly above that point. The stratosphere ranges from about 8 miles to 50 miles above the surface of the earth. This layer is of much interest in aviation and in long-distance ballistics because it is free from wind and moisture.

Above the second layer lies the ionosphere. The ionosphere became known because of its effect on long radio waves early in broadcasting history. Radio waves progressing from their source toward outer space are apparently refracted and reflected back to earth by this high-conductive ionized layer. The height of the reflection layer varies: the zone at about 70 miles, called the Kennelly-Heaviside (E) layer, reflects the radio waves between 300 and 400 meters long; and the Appleton (F) layer, about 140 miles in elevation, reflects the short radio waves. The reflecting layers may vary considerably at different hours of the day or night. Some of the shorter radio waves may pass out through the ionosphere into outer space and not be reflected at all.

Flohn and Penndorf[1] recently stressed the necessity of suitable nomenclature for atmospheric strata as well as a clear definition of the boundaries. The atmosphere is divided into an inner and an outer atmosphere; from the latter particles may escape. The inner atmosphere is divided into three spheres--troposphere, stratosphere, and ionosphere--with each sphere in turn being subdivided into 3 or 4 layers. The stratification of the atmosphere is based upon the thermal structure of the

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[1]H. Flohn and R. Penndorf, "The Stratification of the Atmosphere," Bulletin of the American Meteorological Society, XXXI (March and April, 1950), pp. 71-78.

atmosphere. Boundaries of each layer are fixed by a sudden change of lapse rate. The following analysis will be confined to the study of the troposphere since it is in this layer that all the weather changes and atmospheric disturbances occur and it is in this layer that the microwave propagation plays an important part.

The heating and cooling of the atmosphere are the basic causes of weather changes. The sun is virtually the only source from which the atmosphere receives heat. Some heat is received also from the interior of the earth and from the moon and stars, but the amount is so small, compared with heat from the sun, that it is considered inconsequential as far as the weather is concerned. The heat emitted by the sun is called solar radiation or insolation, although insolation is often used with a more restricted meaning to indicate the solar radiation received by the earth. Accordingly, insolation varies principally with (1) the distance of the earth from the sun, (2) the directness and duration of the sun's rays, and (3) the amount of water vapour and dust in the atmosphere.

Since the sun is the only source of heat, it is not, perhaps, obvious why the temperature of the atmosphere should rapidly decrease with increase of height, except near the surface of the earth at the time and place of a temperature inversion. The average lapse rate (vertical temperature gradient) is about 3.5° F. per 1,000 feet for moist air and about 5.5° F. per 1,000 feet for dry air within the troposphere. Essentially, however, the phenomenon is due to the fact that the atmosphere is heated more by radiation from the earth, which is heated by the sun, than by radiation from the sun. Most of the solar radiation passes through the atmosphere without being absorbed because of its extremely short wave-lengths. The earth's surface is heated and in turn heat energy is radiated by the earth at a lower temperature in the form of longer wave-length radiation, of which the atmosphere is far more absorptive than it is of comparatively short wave-lengths solar radiation. Part of this heat is also absorbed by the water vapour in the atmosphere. In addition to radiation, the transmission of heat to the atmosphere by the earth also takes the form of usual conduction and convection processes. It is shown[2] that considering the average cloudiness to be 52 per cent, and considering average conditions over the entire earth, one may state that, of the total energy incident at the top of the atmosphere, 42 per cent is reflected back to space (including that reflected diffusely by means of scattering process), 11 per cent is absorbed selectively by water vapour in the atmosphere, 4 per cent is absorbed in the atmosphere by permanent gases, dust, and clouds, and 43 per cent is absorbed at the surface of the earth. It is important to notice the comparatively large percentage of absorption by water vapour in the atmosphere, although the average percentage of water vapour content, by volume, is about 2 per cent. This also gives a hint as to why the amount of water vapour present in the atmosphere greatly influences the propagation of microwaves, and why attenuation increases as water

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[2] H. B. Byres, General Meteorology (1944), p. 24.

vapour condenses to form water droplets of larger diameter, and as the relative wave length of propagation decreases. When solar radiation or light passes through a medium containing particles (including molecules), of a diameter less than the wave length of the light, a portion of the light is scattered in all directions. This is sometimes called Rayleigh scattering, after Lord Rayleigh, who studied it in detail. It is effective only for the short wave lengths of light, since it is proportional to λ^4. Blue light is scattered more easily than red, and the blue color of the sky is attributed to this effect. The sunlight which starts out as white reaches the earth with reddish tinge. This is especially noticeable at sunset when the light passes through its longest path of the atmosphere, and it is explained by the fact that blue light has been scattered by the atmosphere and only the reddish portions reach the earth directly. Particles larger in diameter than the wavelengths of the light do not produce scattering. Thus cloud particles do not change the color of light because of their relatively large size. Fine dust particles often produce scattering.

It is necessary to resume discussion of the heating and cooling of the atmosphere and its consequent change of temperature pressure, and water vapour content or humidity. There is great difference in the effects of heating land and sea surfaces. For the same amount of insolation, land surfaces become much warmer than sea surfaces and also lose their heat much more readily. Sea surfaces warm very slowly because convection currents immediately begin and bring in cooler waters. Other factors are involved such as the reflection of heat from water surfaces. Together they cause the difference between the climates of inland and maritime regions. The sea changes extremely little in temperature between day and night as compared with the land. There are three reasons for this:

1. A large amount, about 40 per cent, of the insolation received is reflected back from the surface and thus lost as far as heating the water is concerned. The remaining 60 per cent of the insolation, which is absorbed, is transmitted to considerable depth. Thus it is not a thin surface layer but a layer of considerable depth and great mass that is involved.
2. A considerable amount of evaporation takes place from the surface of the ocean, and insolation is used, therefore, in causing this change of state rather than in causing a rise in the temperature of the water.
3. It requires a larger amount of heat to raise the temperature of a given quantity of water than any other known substance. Since its specific heat is greater than that of any other substance. For this reason a very small rise in temperature takes place as the result of absorbing a considerable amount of insolation.

Therefore, the temperature of the ocean rises very little during the day and falls but little during the night. Similarly, seasonal changes in water temperatures are less than those in ground temperatures.

Dry ground, on the other hand, reflects but a small per-cent of the insolation that falls upon it, transmits practically none, and thus absorbs nearly all of the insolation. The rise of temperature of dry ground is very great whenever it is exposed to the sun's rays. There are three reasons for this:

1. It absorbs nearly all of the insolation without reflection.
2. Being opaque and not subject to mixing, the heating takes place in a thin layer.
3. Its specific heat is much less than that of water.

Since the changes in temperature of the surface layers of the atmosphere depend mainly on the changes in the temperature of the earth's surface, it is obvious from the foregoing that the atmosphere over the ocean will have a less daily, as well as less seasonal, range in temperature than the atmosphere over the land.

The atmosphere is called a standard atmosphere, when it is stable or in stable equilibrium. In standard atmosphere, temperature, pressure, and humidity (water vapour concentration) are all virtually decreasing linearly with increasing altitude. It is important to notice that vapour pressure is directly proportional to the amount of water vapour in the atmosphere and is therefore also decreasing with increasing height.

A reversal of the usual decrease in temperature with altitude is known as temperature inversion, which is one of the most important characteristics of air masses. Listed according to the processes that cause them, the types of temperature inversions are due to:

1. Radiation or contact cooling at the surface
2. Convective turbulence and subsidence
3. Frontal inversions

Owing to cooling by radiation looses at the earth's surface or contact of originally warm air with a cold surface, the vertical temperature lapse rate often is modified with the few hundred feet and the temperatures aloft remain moderately high.

Whenever turbulence lasts long enough, it results in a thorough mixing of the atmosphere through the layers where turbulence exists. There is always a limiting height above which the turbulent or convective mixing does not penetrate, and it is at this altitude that temperature inversions are produced. Subsidence,[3] or slow sinking of the air in a high--pressure area, accounts for many temperature inversions. The process of subsidence makes air layers more stable than they were at their original higher levels, and in its slow movement downward the air is heated adiabatically due to compression. The air near the ground does not participate in the subsidence, because only the slightest degree of turbulence, which is nearly always present at low levels, can completely counteract the slow sinking. The lower atmosphere, then, acts as a sort of shielding layer against subsidence and shows no appreciable temperature increases, while above the temperature at any level within the range of subsidence will slowly rise.

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[3] Ibid., pp. 179-180, p. 254.

The frontal inversion occurs whenever a warm front is overrunning a cold front or a cold front is underrunning a warm front. However, it is expedient to point out here that the ordinary inversions in an air mass have a rapid decrease in moisture content accompanying the temperature rise, whereas the frontal types usually show an increasing specific humidity in the inversion. Radiometeorologically speaking, frontal inversions are not always the inversions which produce radio ducts in microwave propagation.

The horizontal movement of air is always the result of horizontal differences in pressure which, in turn, are primarily due to temperature differences. Horizontal pressure differences cause winds, which tend to blow from areas of high pressure to regions of low pressure. The greater the difference in pressure between two areas, the greater is the velocity and force of the wind. The change in pressure per unit of horizontal distance is called the barometric or pressure gradient. A low pressure area develops when a part of the surface of the earth becomes warmer than the surrounding regions. Heated air increases in volume and, as its expansion laterally is very limited, most of its expansion occurs vertically. The height of the expanded column of air is thereby increased and this mass of air flows outward in all sides. Consequently, the amount of air over the heated surface is decreased and has less pressure. A high pressure area develops from the overflow aloft from low pressure areas or from the cooling of a part of the surface of the earth. As cool air contracts, the air from aloft flows downward to the surface of the earth, increasing the total mass of air above the cold surface; and temperature thereby becomes greater. Therefore, the unequal heating and cooling of large land and water surfaces are responsible for the development of these high and low pressure areas. The forces acting on moving air are not only due to the pressure gradient but also to the deflection caused by the rotation of the earth, and accordingly, air does not flow at right angles to the isobars on account of this Coriolis force, which in the Northern Hemisphere pulls air around toward the right.

Studies show that areas of low pressure are always surrounded by areas of high pressure, and therefore air tends to flow toward this center of low from regions of higher pressure near it. As previously mentioned, the moving air is deflected by the rotation of the earth to the right in the Northern Hemisphere, and consequently, air moving in toward a low thus tends to spiral in a counterclockwise direction. At areas of high pressure, air spirals outward, clockwise in the Northern Hemisphere.

Studies also indicate that a low develops usually near the zone of contact between two moving air masses--one consisting of warm tropical air and the other of cold air. This low is therefore a cyclonic center. For the origins and classification of air masses, the reader should refer to any book on general meteorlogy. In North America, cP air masses originate in Canada and tend to move southward east of the Rocky Mountains, bringing low temperatures to the eastern states along a lengthy advancing edge of cold air which is termed the polar front. Air reaching the interior of the United States in winter may also come from the North Pacific (mP), bringing comparatively mild winter weather to the Pacific Northwest. Marine-tropical air (mT) enters the continent primarily from the Gulf of Mexico, but occasionally it originates near the Gulf of California and moves inland. This air mass usually is moist and warm, and it requires little contact with cP or mP air to cool it sufficiently to produce clouds and rain. Marine-tropical air from the Gulf of Mexico is a principal source of summer rainstorms in the eastern part of the United States.

If a cold air mass is relatively stable and a warm air mass advances and overrides the cold mass, the advancing line or zone of discontinuity is called a warm front. This is usually accompanied by high light clouds--cirrus and cirro-stratus--and falling pressure. Later, lower and denser clouds with some rain or snow may follow before the storm ends. On the other hand, if a mass of cold air advances at ground level and acts as a wedge to lift warmer moist air to higher altitudes, the contact zone is called a cold front; the resultant cooling of the warm air because of increase of altitude will bring storm conditions along the advancing cold front. Generally the storms generated by the cold front are severe, with linear arrangement--the squall line--and heavy dark clouds of the thunderhead type. Atmospheric pressure rises suddenly with the arrival of the cold front, and clearing and relatively cold weather may be expected after the storm passes by.

From the above discussion it can be seen that both warm fronts and cold fronts produce cyclones. The condensation of moisture in the cyclone releases latent heat which expands the air and reduces the pressure, letting more air in from all directions toward the resulting low pressure area. This process of engulfment of a mass of warm air is termed occlusion: the warm air of the low is lifted above ground level, and an occluded front is formed.

Almost all the storm result from the general circular way in which winds blow around the center of low-pressure areas, which are zones of contact between two moving air masses of contrasting characteristics as described. Such storms are called cyclones and should not be confused with tornadoes, the destructive storms that will be discussed fully in the next chapter. Cyclones, often called "lows," appear on weather maps as oval or elongated areas with air pressure lower than that of surrounding regions. On the weather maps of the United States the opposite areas of high pressure are labeled "highs" and are technically called anticyclones. Cyclonic disturbances vary tremendously in size. In general, however, they are very large and may have a width of 500 to 1,000 miles or more; but, since they are confined to the troposphere, they are only 5 to 8 miles in height. The storms may move thousands of miles before wearing themselves out and disintegrating.

In the zone of contact between two different air masses a sudden fall in temperature may result in heavy precipitation. This also means that there is a steepening of the temperature lapse rate in very moist air. Under such changes are produced squalls, thunderstorms, and sometimes destructive tornadoes. The thunderstorms thus produced are commonly called frontal thunderstorms which are of mechanical type because the movement of undercutting or overrunning air masses produces the necessary instability. The other kind of thunderstorms is called con--vectional, or heat or a local thunderstorm, which is the result of the overheating of the lower layers of the atmosphere when the air is relatively calm and has a high moisture content. Whenever the air in the lower layers is warmer and has higher moisture content than the upper layers, convectional instability develops. Showers are usually convectional storms of less degree of convection with no lightning. Thunderstorms occur nearly everyday in the tropics. In the latitudes, they are common to the interior of continents in summer. The condensation of moisture and the formation and separation of raindrops in thunderstorms develops static electricity. When sufficient difference of the electric potential has developed, lightning passes between two oppositely charged clouds, or from a cloud to the earth. The high temperature of the electric flash heats the air through which it passes, causing a terrific expansion of the gases, which produces the noise called thunder.

Storms in the tropical regions are called tropical cyclones. Tropical cyclones of certain degrees of intensity are known as hurricanes in the West Indies and the eastern North Pacific; as typhoons in the Western North Pacific and the China Sea. Tropical cyclones differ in many respects from extratropical cyclones. Besides being smaller and more intense, they are also distinguished by their greater symmetry. They are much less frequent than extratropical cyclones and are entirely phenomena of the oceans during warmer parts of the year. Another distinguishing characteristic is the central calm or "eye," which generally has an area of 5 to 30 nautical miles in diameter and has extremely low pressure. The process bringing about the formation of tropical cyclones is not definitely known, and the general prevailing interpretations are: (1) the classical thermodynamic convection theory, and (2) the frontal theory. The hypothesis of Sawyer[4] and that of Riehl[5] are the latest treatments in this field. However, it is fortunate that the tracking of hurricanes by radar precipitation echoes has been successfully carried out by the Florida Engineering and Industrial Experiment Station at the University of Florida. Tornadoes have been reported in a few instances with hurricanes in the Bahamas and in Cuba and fairly frequently in Florida.

The author feels it is expedient to introduce briefly the formation and classification of clouds. Clouds are formed and rain may result, when the condensation of the moisture in the air is brought about, not by contact with cold solid objects at the surface of the earth but by great masses of cold air high above the earth. Fog is basically a cloud which touches earth. A cloud is not a durable phenomena. It simply marks a place in the atmosphere where the condensation of water vapour

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[4]J. S. Sawyer, "Notes on the Theory of Tropical Cyclones," Quart J. Royal Meteorological Society, LXXIII (1947), pp. 101-126.

[5]H. Riehl and R. L. Shafer, "The Recurvature of Tropical Storms," Journal of Meteorology, I (1944), pp. 42-54.

is going on. The principle methods of cloud formation may be classified according to the following cooling processes:

1. The direct ascent of air by convection or because of topography
2. The gradual ascent of air over a large area, as in front of the warm sector of a cyclone where the advancing mass of warm air forms a sloping surface in rising over the heavier colder air.
3. The mixing of two unsaturated air masses of different temperatures, which result in a mixture that is more saturated at the mean temperature

The forms of clouds are generally classified into three formations: (1) Cirrus--detached, delicate, fibrous with a generally white silky appearance and composed of ice crystals. These are about 30,000 feet in elevation. (2) Stratus--a uniform low layer, from a few hundred to a few thousand feet in elevation, generally grey in color. (3) Cumulus--thick clouds with vertical or convectire development, with their upper surfaces dome-shaped with round proturbances and with bases nearly horizontal. When the word "nimbus" is placed before or after such a cloud name, it means that rain is actually falling.

Clouds may be grouped according to their respective altitudes: (1) upper clouds (mean lower level, 20,000 feet)--cirrus, cirrocumulus, cirrostratus. (2) Middle clouds (mean lower level, 6500 feet)--altocumulus, altostratus, (3) low clouds (mean lower level, 400 feet)--stratocumulus, stratus, nimbostratus, (4) convection clouds (mean upper level, that of cirrus; mean lower level, 1600 feet)--cumulus, cumulonimbus.

Cumulonimbus clouds are common to the rear of a cyclone and are associated with most summer showers and all thunderstorms. These clouds have strong upward air currects and are of such great vertical extent that their tops have outgrowth of cirrus type clouds. The characteristic funnel-shaped clouds of tornadoes strech down toward the earth from cumulonimbus cloud bases. Cumulonimbus clouds are always present before the formation of tornadoes. The characteristics of tornado formations will be fully discussed in the next chapter.

CHAPTER III

METEOROLOGICAL CHARACTERISTICS OF TORNADOES

A tornado, which is sometimes called a "twister," is the most violent local vortex in the atmosphere. It has an intense spiral motion around a vertical or inclined axis and is a funnel-shaped cloud with violent upward spiraling winds of incredible violence. Its lower part is often characterized by a narrow pendant cloud extending from a cumulonimbus cloud base to or nearly to the ground. Tornadoes over the water are called waterspouts, because they draw water into their funnels instead of dust and debris as in the case of tornadoes. The tornado is the least extensive yet the most violent of all storms.

Tornadoes occur on all continents, but are rare except in Australia[1] and the United States. Tornadoes in the United States are commonly experienced in central level states. Tornadoes may occur at any time, and the number of tornadoes in any one year may vary considerably from the annual average. The seasons, listed in the order of tornado frequency for the United States, are late spring, early summer, late summer, autumn, and winter. May has the greatest number; December, the least. Over 80 percent of reported tornadoes occur between noon and 9 p.m.

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[1] E. M. Brooks, "Tornadoes and Related Phenomena," Compendium of Meteorology (1951), p. 673.

Heavy rain and lightning discharges always accompany a tornado, but are not likely within the tornado itself. Hail is also common. However, Brooks[2] reported that although most of the tornadoes ~~were~~ are accompanied by thunderstorms, hail, and rain, some ~~occurred~~ occur without electrical activity or precipitation. The direction of travel of tornadoes is roughly parallel to that of the center of a cyclonic storm, hence usually northeastward in the United States. Brooks[3] mentioned that the diameter of a tornado, or width of a tornado path, which averages close to 250 yds., varies from zero up to about 1 or 2 miles. The length of a tornado path averages about 4½ miles, and ranges from only about 100 ft. up to 300 miles. The translation speed of a given tornado is not uniform. Also the average translation speeds of different tornadoes vary, ranging from nearly zero to nearly 150 mph., with a mean close to 35 mph. The average tornado duration is computed to be ½ min. at a single point and 8 min. on a path along the ground. An extreme case of duration of over 7 hours along the ground has been observed (Illinois, May 26, 1917). The pressure and winds within a tornado must be determined largely by indirect methods, since direct methods are almost impossible up to the present. These methods include a detailed analysis of the tornado damage and a theoretical study of the relation between wind and atmospheric pressure. Calculation of wind based on structural failure and impacts of flying

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[2]E. M. Brooks, "Some Characteristics of Tornadoes in 1948 near St. Louis," Bulletin of American Meteorological Society, (1948), p. 520.

[3]Brooks, op. cit. p. 676.

objects yield values ranging from less than 100 mph. to one case of more than 300 mph.--enough to demolish the strongest buildings. Theoretical calculation between wind and atmospheric pressure in a frictionless vortex surrounded by a windless environment shows that a pressure of 900 mb., or a decrease of about 100 mb. would give a wind of about 300 mph.

The air preceding a tornado is usually warm and humid, whereas the air following it is usually cool and dry. The phenomena indicate clearly that there must be a cold front either aloft or behind the local warm and humid air before the occurrence of a tornado. Both from theory and observation it was found that the presence of upper cold front air on top of the moist and warm air is one of the most important meteorological requirements for the formation of a tornado. Different and sometimes contrasting hypotheses on the generation and maintenance of tornadoes have been proposed; however, meteorologists are almost unanimous in agreeing that tornadoes are a result of excessive instability and therefore of steep lapse rates in the atmosphere. Tornadoes are usually in advance of the squall line or cold front, because the cold air comes in aloft well ahead of surface cold air and occasionally helps to cause such disturbance and local instability that a tornado develops. (A statistical study by Gaigeror[4] of 92 tornadoes in the United States has shown that the vast majority of tornadoes occur at

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[4] S. S. Gaigeror, "Synoptic Conditions Accompanying Tornadoes in the United States during 1884," Bulletin of the American Meteorological Society, XXI (1940), pp. 229-236.

some distance ahead of the cold front. There were 72 cases out of the 92 in which the tornadoes preceded the cold front, and the average distance before the front was 165 miles. There were 10 cases where the tornadoes occurred also in the warm sector, but nearer to the warm front. The remaining 10 cases were not in the warm sector--occurring either before a warm front, before an occluded front or behind a cold front (5, 2, and 3 cases, respectively). Apparently in the latter cases the tornadoes formed in the overrunning tropical air and broke through the frontal inversion or the instability layer. Only warm sectors of genuine maritime-tropical air from the Gulf of Mexico are capable of producing tornadoes.

From Gaigeror's statistics the author feels that tornadoes can be divided according to their origin of formation into two groups: namely, the pre-cold frontal type and the instability type. The pre-cold frontal type of tornado is one in which strong cold-air advection is found aloft before its occurrence. Under certain conditions, cold air masses, usually of Pacific origin or from Western Canada, become heated so much in low layers over the western plateaus that they ride up over the maritime-tropical air of the Mississippi Valley. Although this cold dry air is heated up in the low layers, it preserves its low temperature at greater heights, and thus a steep temperature lapse rate is developed. Apparently, as this air moves aloft over the maritime-tropical air, nothing much happens at first, because of the slight temperature inversion or stable layer separating it from the air below. However, some cloud formation in the maritime-tropical air may release enough heat through condensation so that it will be able to penetrate into the upper layer. There the lapse rate is so steep that an ascending parcel or cloud mass would be greatly accelerated upward, resulting in the explosive type of convection that apparently is necessary for tornado development. Showalter[5] has shown by his typical aerological soundings of the maritime air in which tornadoes develop that it has high relative humidity usually up to only about 1-3 km. A thin stable layer, which may be an inversion, separates the maritime air from the dry cold air aloft, which is characterized by a steep lapse rate. This thin layer is convectively unstable because of the rapid decrease of humidity with height. Once convection is started the moist and warm maritime-tropical air becomes the ideal source of energy supply to develop more severe convection which may finally result in tornado formations. Convectional cells are common in thunderstorms. The horizontal and vertical dimensions of the updraft or convective column in the cloud vary from 20,000 ft. to 60,000 ft., with an average dimension of about 40,000 ft. Maximum updraft velocities of 84 ft. per second have been measured. Some meteorologists contend that the most violent thunderstorms have the greatest vertical development. Others specify excessive vertical draft velocities as the criteria of

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[5]A. K. Showalter, "The Tornado--An Analysis of Antecedent Meteorological Conditions," Preliminary Report on Tornadoes, U. S. Weather Bureau, Washington, D. C. (1943), p. 139.

violence approaching tornadic conditions. Nunn[6] mentioned that high vertical growth rates might be an exclusive feature of tornadic cells. The vertical and horizontal development of a convective cell can be observed by RHI and PPI scope respectively.

Instability lines[7], which are nonfrontal squall lines, are sometimes accompanied by tornadoes and by a presumably extreme vertical instability through a relatively deep layer of the atmosphere. A well-developed instability line is marked by squalls or thunderstorms along a line that is usually several hundred miles in length, and in a typical case 50 to 300 miles ahead of a surface cold front. Lloyd[8], in his studies of tornadoes, attributed at least the tornado-producing squall-line condition to an upper cold front. Advection of colder air aloft and the resulting decrease of temperature, as would be required by existence of an upper cold front, is observed in many cases. But cooling aloft also takes place, in some instances, ahead of squall conditions, indicating that the degree of instability near the leading edge of the cooling aloft is not always sufficient for convective activity. Back from the leading edge, temperature aloft may be lower, and surface temperature higher, so that at some point the critical lapse rate necessary for convection may be realized. The exact distance between the squall condition and the leading edge of the colder air aloft is difficult to

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[6] W. M. Nunn, Radar Analysis of Tornadoes, Master of Science Thesis, Oklahoma Agricultural and Mechanical College (1952), p. 82.

[7] J. R. Tulks, "The Instability Line," Compendium of Meteorology (1951), p. 647.

[8] J. R. Lloyd, "The Development of Tractories and Tornadoes," Monthly Weather Review, LXX (1942), pp. 65-75.

determine by observation, but appears to vary from a few miles to hundreds of miles. Where the leading edge of advection of colder air aloft is very close to the squall condition, and there is evidence of discontinuity in density as required for a front aloft, the condition is not properly classified as "nonfrontal." In many instances, however, the existence of a discontinuity aloft along or near the squall line is difficult or impossible to determine from synoptic data, and the only practical recourse for the synoptic meteorologist is to follow the instability line. Figure 1 represents diagrammatically a typical well-marked instability line in the warm sector of an extratropical cyclone.

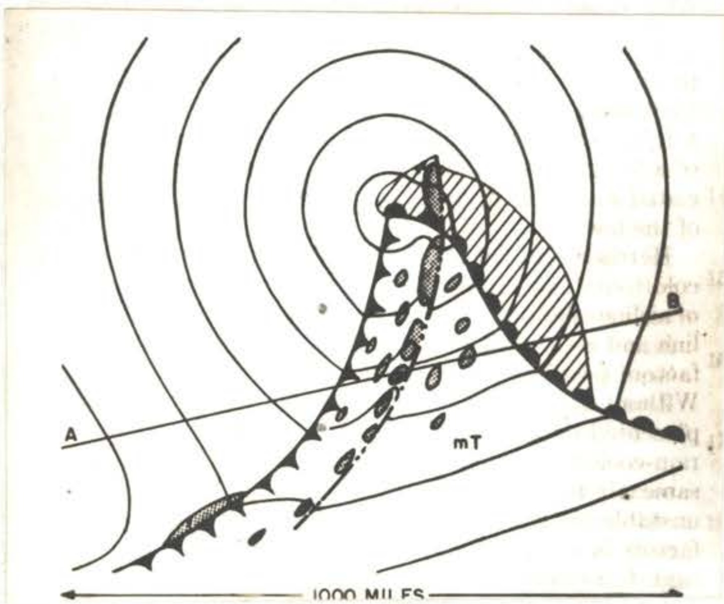

Fig. 1 - Model of cyclone with instability line in warm sector. Shading indicates area of active squalls, and hatching shows warm front precipitation.

Squalls and thunderstorms (shaded area) are shown as a nearly solid band along the instability line, at scattered points elsewhere in the warm sector, and within the area of warm-front precipitation (hatched). In this type of low, and at this stage of development, precipitation is rarely observed behind the portion of the cold front following the instability line.

The physical structure and mechanics of the instability line are not well known, and, therefore, future progress toward a better understanding in this field will require both the collection of more adequate detailed observational data and the analysis of existing and future data. More observational data for upper levels in the immediate vicinity of tornadoes is especially needed; this is difficult to obtain but will be very important because of the frequent association of tornado conditions with the instability line.

Tepper[9, 10] has recently introduced his hypothesis of "pressure jump line" in connection with the instability-type of squall line and its possible relation to tornado development. He proposed that a squall line may be explained as a pressure jump induced by accelerations along a cold front and moving along an inversion in the warm sector as a gravitation wave aloft (following a suggestion by Freeman[11]). Generally this type of squall line roughly parallels the cold front, lies to the east of the cold front, and moves more rapidly than the cold front. He specified that the cold front acts as a piston and suggested that the initial impetus to the formation of the squall

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[9] Morris Tepper, "A Proposed Mechanism of Squall Line: The Pressure Jump Line," Journal of Meteorology, VII (Feb. 1950), pp. 21-29.

[10] Morris Tepper, "On the Origin of Tornadoes," Bulletin American Meteorological Society, XXXI (Nov. 1950), pp. 311-314.

[11] J. C. Freeman, Jr., "An Anology Between the Equatorial Easterlies and Supersonic Gas Flow," Journal of Meteorology, V (1948), pp. 139-146.

line is furnished by acceleration of the cold front. As a result of these accelerations, the pressure jump is formed and moves out ahead of the cold front. When the cold front begins to decelerate, a rarefaction wave is formed behind the pressure jump. Thus, the mechanism for the eventual dissipation of the pressure jump phenomena is inherent in this pattern, because the rarefaction wave, moving faster than the jump, eventually overtakes it and destroys it. He did not explain the mechanism that induces the acceleration along the cold front, which in turn induces the pressure jump, or squall line.

Tepper also speculated that tornadoes usually form in the zone of intersection of two unequal pressure jump lines, one oriented generally N-S or NW-SE (i.e. roughly parallel to the cold front) and the other oriented generally E-W or NE-SW. He cited the discussion of interaction by Courant and Friedrichs[12] to lend support to this conjecture. When two shock waves intersect in the two-dimensional flow of a compressible fluid, a vortex sheet is formed. This vortex sheet has been found experimentally in gas flows and in flow of water in a channel. Such a vertical vortex sheet should form behind the intersection of two pressure jumps.

Tepper[13] also analyzed a tornado situation by radar pictures with his "pressure jump line" hypothesis and synoptic data.

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[12] R. Courant and K. O. Friedrichs, Supersonic Flow and Shock Waves, Interscience Publishers, New York, p. 177.

[13] Morris Tepper, "Radar and Synoptic Analysis of a Tornado Situation," Monthly Weather Review, LXXVIII (Sept. 1950), pp. 170-176.

The precipitation echo pictures taken during the period preceding and following a tornado suggest the possibility that the "pressure jump line" theory may be pointing in the correct direction. Upon careful investigation, he found that all available synoptic data were compatible with the conclusion that two pressure jump lines were present. These two pressure jump lines were considered to have produced the pattern of precipitation echoes indicated on the radar scope. Furthermore, the occurrence of a tornado in the intersection of two pressure jump lines is in accordance with the tornado hypothesis suggested by Tepper. However, it should be kept in mind that these echoes are merely pictures of the portion of the precipitation area "seen" by the radar set and do not represent the pressure jump lines themselves. The precipitation pattern results from the forced lifting produced by the pressure jump line and consequently would normally lag behind the pressure jump line in both time and space. It seems important to mention here that this hypothesis deals perhaps with only one of several mechanisms which may produce tornadoes; therefore, it may not work in all cases to which it is applied. The research of radar tornado detection at O. I. T. is to obtain enough observational data to be grouped and analyzed to yield statistical results to corroborate the present and future hypotheses and to suggest new hypotheses in this phase of tornado formation and tracking.

Observational researches and theoretical developments in other phases must be directed to secure complete understanding of all tornado developments and, therefore, an effective forecasting and tracking. system Brooks[14] suggested that since the tornado is a local circulation, the meteorological data ought to be gathered over micronetworks which could detect the development of a small secondary cyclone and could locate accurately squalls, thunderstorms, and hailstorms which might indicate probabilities of tornadic situations. After a tornado occurs, careful surveys should be made of the damage to determine winds and atmospheric pressure drops. A standardized questionnaire should be used in personal interviews and should be published in local newspapers with a request for replies. Copies of local photographs should be obtained for analysis--the best photographs are those which include the top of the cumulonimbus cloud formation above the pendant cloud rather than the pendant alone. A more complete knowledge of detailed analysis of weather within limited regions of probable tornado occurrence is needed for the understanding of the nature and cause of tornadoes. Such knowledge will put tornado forecasting and tracking on a more accurate and reliable basis.

Finally it is important to mention the empirical method of forecasting tornado development by Fawbush and Millar.[15] After a lengthy investigation of a large number of synoptic situations in the United States prior to 1949, they found that a tornado situation developed, when, and only when, the synoptic situation was characterized by the following conditions:

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[14]Brooks, op. cit. p. 679.

[15]E. J. Fawbush and R. C. Millar, "An Empirical Method of Forecasting Tornado Development," Bulletin of the American Meteorological Society, XXXII (Jan. 1951), pp. 1-9.

(1) A layer of moist air near the earth's surface must be surmounted by a deep layer of dry air.

(2) Horizontal moisture distribution within the moist layer must exhibit a distinct maximum along a relatively narrow band (i.e., a moisture wedge or ridge).

(3) The horizontal distribution of winds aloft must exhibit a maximum speed along a relatively narrow band at some level between 10,000 ft. and 20,000 ft. with the maximum speed exceeding 35 knots.

(4) The vertical projection of the axis of wind maximum must intersect the axis of the moisture ridge.

(5) The temperature distribution of the air column as a whole must be such as to indicate conditional instability.

(6) The moist layer must be subjected to appreciable lifting.

They further stated that the above conditions must be satisfied simultaneously at the time of the first appearance of tornadoes and similar storms.

CHAPTER IV

TROPOSPHERIC MICROWAVE PROPAGATION

With the introduction of centimeter electromagnetic waves, entirely new meteorological effects on propagation of radio waves have come to light. The interdependence of radio wave propagation and meteorology is illustrated by the phenomena of (1) radio refraction and (2) scattering and attenuation by suspended water droplets in the lower atmosphere. The first of these is the variable non-rectilinear propagation of centimetric waves through lower atmosphere, the range of such waves over the earth's surface often extending the geometrical horizon of the transmitter. Ionospheric refraction plays no part in this propagation, for, in the wave-lengths concerned, penetration of the ionosphere is almost quite complete. The second major phenomenon is the scattering and attenuation of the shorter of these waves by suspended water drops or snow crystals when present in sufficient density in or beneath a cloud.

The refraction phenomena of radio waves can be basically understood by a comparison with those of light rays from the sun, the astronomical refraction in meteorological optics, since light is essentially electromagnetic waves of extremely short wave-lengths. As it was mentioned in the first chapter, under standard conditions the earth is surrounded by the atmosphere, the density of which decreases linearly with altitude. Thus light rays from the sun, when penetrating the atmosphere, suffer from the law of refraction (Snell's Law) which states that tangential components of light rays remain constant and are independent of height. The light ray from the sun describes, instead of a straight line, a curve which has for its equation

$$nh(\sin i) = k = \text{constant},$$

where h is the distance from point P to the earth's center, as shown in Figure 1, i the angle of incidence, and n the refractive index of light which is a function of h.

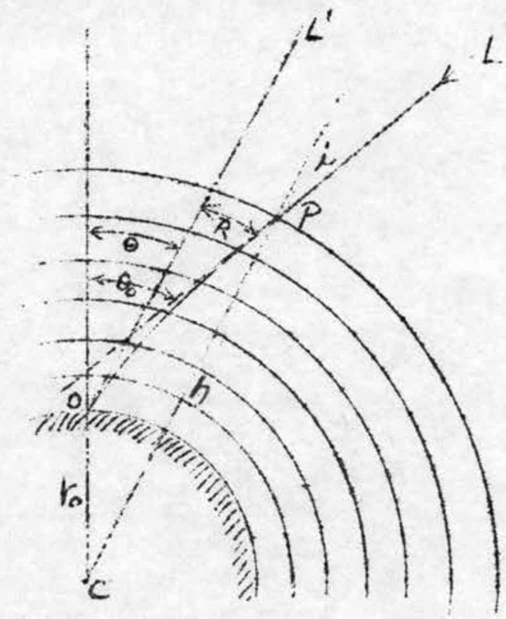

Fig. 1. Schematic diagram of astronomical refraction.

In Figure 1, r_o = CO is the earth's radius; an observer at O sees a light ray L, that enters the atmosphere at P with the angle of incidence i, at the apparent bearing O_o, instead of the true bearing of O. The difference $\theta - \theta_o = R$ is the astronomical refraction. Therefore, there is some downward bending of light rays in the atmosphere due to the normal decrease of density of air with altitude. It is well known that even an ordinary optical search-light beam, straight though it may appear to the eye, has in fact a downward curvature throughout the atmosphere. It has been shown that a suitable allowance to make for this curvature of light rays is one-fifth of that of the earth.

In the same way, radio waves, whether long or short, undergo refraction when propagating in the atmosphere. It might be wondered why the importance of radio wave refraction in the lowest layer of the atmosphere should only have been recognized in recent years, inspite of the fact that radio wave propagation has been studied for many decades. Part of the reason for this is that such tropospheric refraction as occurs at dekameter and longer wave-lengths is apt to be completely overshadowed by ionospheric refraction. Whereas, with the development of meter and shorter wave-lengths, with which ionospheric refraction plays little part, tropospheric refraction is more easily recognized and studied. But the main reason is that tropospheric refraction itself depends on wave-lengths, being of comparatively little importance at dekameter and longer wave-lengths, but increasing in importance as the wave-length is reduced from meters, through decimeters to centimeters. This is not because the refractive index of the atmosphere at radio wave-length depends on the wave-length to any great extent, but because the longer wave-lengths respond to such a rough average of the atmospheric gradient that fine atmospheric structure close to the earth's surface is largely ignored and can only be fully appreciated at sufficiently short wave-lengths.

With the above introduction it seems necessary to present briefly the phenomena of dekameter or longer wave propagation in order to provide a clear insight of the difference between broadcasting wave and radar wave propagation. The successful reception of long wave signals transmitted across the Atlantic by Marconi, in 1901, needed some explanation in that these waves, if similar to light, should have been unable to follow the curvature of the earth. Heaviside and Kennelly offered the suggestion that an ionized layer in the upper atmosphere could serve as a reflecting surface which would confine the radiation to the earth. This layer has been given the name "Kennelly-Heaviside layer." Subsequent investigation has led to considerable modification in the original hypothêsis. The existance of not one, but of several ionized layers, as described in Chapter I, has been demonstrated by experiments. Instead of the wave's being reflected from the conducting layer, as light from the surface of a mirror, it enters the mediam and is bent back to earth again by ionospheric refraction in the ionosphere. However, it is convenient for purposes of calculation to regard the process as one of simple reflection, by which the wave travels with the velocity of light to the hypothetical reflecting plane where it is then reflected and returned at the same velocity. The height of this plane is called the virtual height of the layer, which is seen from Figure 2 to be somewhat greater than the actual height. When waves enter the ionized medium or other medium of different refractive index, the signal or group velocity of the waves is retarded

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[1] R. I. Sarbacher, Hyper and Ultrahigh Frequency Engineering (1943), pp. 132-137.

below the velocity of light. This group velocity is the rate at which the energy of the wave is travelling. The phase velocity of the wave is the rate at which the phase changes along the path of the wave and can be much greater than the velocity of light, since it is equal to the velocity of light divided by the refractive index of the medium. This speeding up of the phase velocity causes the wave path to be bent back to the earth. The amount of bending experienced will depend

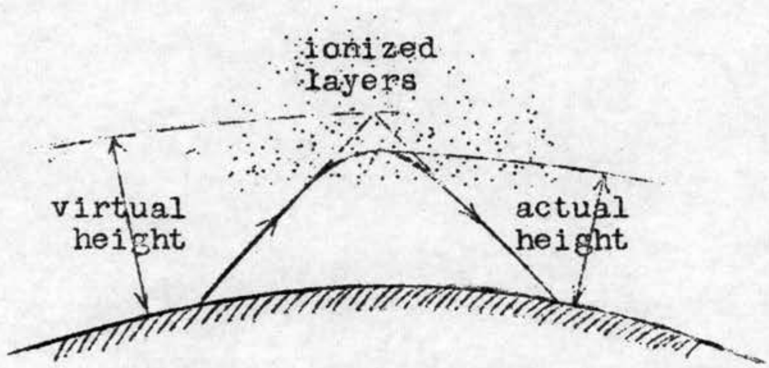

Fig. 2. Refraction of a radio wave by layers in the ionosphere

upon the degree of ionization and its gradient, as well as upon the direction and frequency of incident waves. With a given density of ionization, the amount of bending will diminish with increase of frequency, so that at some critical value of the latter, the wave will penetrate the lowest ionized layer and be reflected from some more intensely ionized layer above the first. Wave lengths below 8 or 10 meters will usually not be bent sufficiently to return to the earth, and hence are not useful for long-distance communication.

In ordinary radio terminology, long-wave transmission is of ground-wave transmission and short-wave transmission sky-wave transmission. The radio waves with which this chapter deals are essentially those short enough to penetrate the ionosphere at all times. Broadly, this means meter waves and shorter ones, though both physical characteristics and their operational consequences have led to a concentration of interest and effert in the centimeter band. The problem is therefore, in ordinary radio terminology, that of ground-wave transmission, now customary called microwave transmission or propagation.

Propagation of microwaves in the troposphere is materially influenced by the distributions of temperature, pressure, and water vapour. The cause of bending or refraction, is found in the manner of the variation of the refractive index of the atmosphere with altitude. Under all conditions met with in the troposphere, the refractive index of the atmosphere for radio waves is a few hundred parts in a million greater than unity. The index of refraction n is given by the well known empirical equation

$$n = 1 + \frac{79}{T}\left(P + \frac{4800e}{T}\right)\times 10^{-6},$$

which is usually written in the form

$$(n - 1)\times 10^{6} = \frac{79}{T}\left(P + \frac{4800e}{T}\right),$$

where T is the absolute temperature, P and e are the total pressure and water vapour pressure, respectively, in millibars, at height h above sea level. The refractive index of dry air is practically the same for radio waves as for light waves and expressed by the equation

$$(n - 1)\times 10^{6} = 79\frac{P}{T}.$$

On the other hand, the refractive index of water vapour, which is always present to some extent in the atmosphere, is different for light waves and radio waves. This arises from the fact that the water vapour molecule has a permanent dipole moment which has different responses to electric forces of different frequency. At the relatively low frequencies used in radio, the water vapour molecules not only acquire electric polarization, but also orient themselves sufficiently rapidly enough to follow the electric field changes. However, in the case of the high optical frequencies, electronic polarization alone occurs. The result is that the dielectric constant, and thus the refractive index (square root of dielectric constant), of the water vapour is greater for radio than for optical frequency.

Usually the atmospheric refractive index is pictured as spherically stratified similarly to the stratification of the atmosphere, and under standard atmosphere, n can be also expressed as a function of height h alone, since T, P, and e are themselves functions of h.

Although the refractive index of the atmosphere differs little from unity, and the changes (n - 1) with height are very small, they are nevertheless sufficient to bring about remarkable effects in radio transmission phenomena, since directional deviations of the order of a fraction of a degree are of great practical importance in centimeter wave radar propagation.

This atmospheric refraction of radio waves can be studied more in detail by use of Snell's law, $n_1 \text{ Sin } \theta_1 = n_2 \text{ Sin } \theta_2$; i.e., $n \text{ Sin } \theta = \text{constant}$.

According to Fig. 3, Snell's law may be written:

$$n_1 \cos \alpha_1 = n_2 \cos \alpha_2 \tag{1}$$

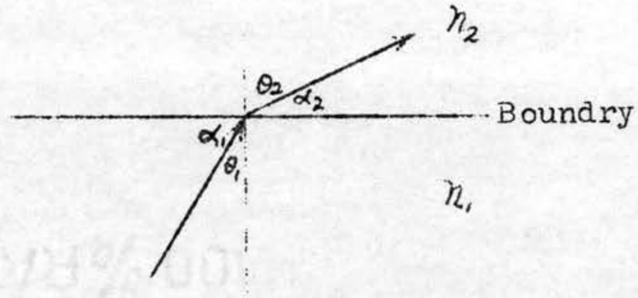

Fig. 3. Refraction at Boundary

Physically speaking, radio energy radiates in the form of spherical wave fronts. The velocity at any point on the wave front is given by

$$V = \frac{C}{n} = \frac{3 \times 10^8}{n} \text{ meters per second}$$

since n decreases with height, the upper portions of the wave front move with higher velocity than the lower portions, and the wave paths as represented by the rays are therefore carried downward.

It can thus be imagined that the lower atmosphere is made up of many layers, each of a somewhat different refractive index n. Such a concept will lead to successive refractions as in Fig. 4.

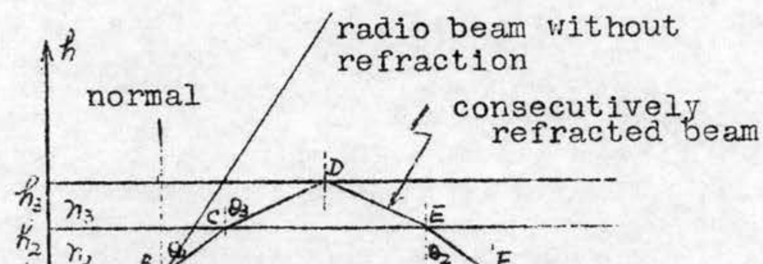

Fig. 4. How a radio beam becomes a curved beam by means of consecutive refraction.

Because air is denser at lower levels than at higher levels, the incident ray AB making angle θ_1 with the normal to the h_1 level will be refracted away from the normal by an angle θ_2, causing a transmitted beam BC, etc., until at D, horizontal gliding (at grazing angle) would take place. However, according to Gans, total reflection takes place at this point. Thereafter, the ray again goes through consecutive refractions. It can be shown[2] mathematically by the rate of change of phase velocity due to the change of refractive index, that, if P is allowed to be the radius of curvature of the refracted beam,

$$\frac{1}{p} = -\frac{1dn}{ndh} \tag{2}$$

This expression gives the path relation in terms of the radius of curvature p, if the index of refraction n is known at any point in the path, as in this case of linear variation of n.

The above presentation is based upon the fact that the surfaces of a constant refractive index are planes. In reality, the surfaces of a constant refractive index are not planes, but are concentric spheres about the earth's center. In this case Snell's Law assumes a slightly different form. Instead of using angles referred to the plane surfaces, it is now necessary to refer the angles to horizontal planes tangent to the sphere about the center of the earth (see Figure 5). The new equation shown by Humphreys[3] takes the form,

$$nr \cos \alpha = an_0 \cos \alpha_0. \tag{3}$$

where r and a are values of the radius vector from the center of the earth to a point in the atmosphere and to the earth's surface, respectively. α now stands for the angle formed by the

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[2]August Hund, Short-wave Radiation Phenomena II (1952), pp. 981-983.

[3]W. J. Humphreys, Physics of the Air (1940), p. 457.

ray with a plane normal to the radius vector. α_0 and n_0 are the values of α and n at the ground surface. If h is the height above the ground surface, so that $r = h + a$, the equation (3) may be written as: $n(1 + h/a) \cos \alpha = n_0 \cos \alpha_0$. (4)

h/a is a very small quantity, and n differs from unity by only a few parts in 10,000. Under these conditions $n(1 + h/a)$ may be replaced by $n + h/a$ with negligible error. The quantity $n + h/a$ is called the modified refractive index N. The above equation (4) then becomes: $(n + h/a) \cos \alpha = n_0 \cos \alpha_0$. (5)

It is customary to use, instead of $n + h/a$, the symbol M defined as follows for numerical convenience:

$$M = (n + h/a - 1) \times 10^6 \qquad (6)$$

and is called refractive modulus.[4] At the surface of the ground, M reduces to: $M_0 = (n_0 - 1) \times 10^6$. (Equation 6a). Values of M, in units of MU (M unit), for the atmosphere lie in the range of 200 to 500. It should be noted that whereas n usually decreases with height, the refractive modulus M, except for limited regions in non-standard atmosphere, increases with height.

Consequently the vertical gradient of M for the standard atmosphere is: $dM/dh = (dn/dh + 1/a) \times 10^6$. (7)

In northern temperate latitudes, the rate of decrease with height of the refractive index dn/dh has been found, on the average,

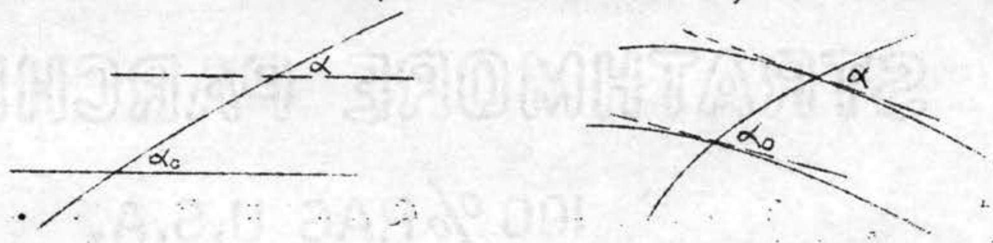

Fig. 5. Refraction through plane and curved layers

[4] D. E. Kerr, Propagation of Short Radio Waves, (1951), p. 191.

to be

$$\frac{dn}{dh}\times10^6 = -\frac{1}{4a}\times10^6 = -0.039 \text{ MU per meter} \qquad (8)$$

using the value for radius of the earth $a = 6.37\times10^6$ meters.

It is appropriate now to introduce the concept of equivalent earth radius, which was first used by Eckersley[5] in radio wave application. It has been shown by equation (2) that the curvature of a refracted beam is related to the gradient of n by

$$\frac{1}{p} = -\frac{1}{n}\frac{dn}{dh} \qquad (2)$$

since n is almost equal to unity, the above equation takes the form

$$\frac{1}{p} = -\frac{dn}{dh} \qquad (9)$$

$1/a$ is the curvature of the earth. The algebraic sum of these two curvatures is the curvature of the ray relative to that of earth. The net result is this: if the earth is replaced by an equivalent earth with an enlarged radius equal to ka the rays may be drawn as straight lines. Then

$$\frac{1}{a} - \frac{1}{p} = \frac{1}{ka}, \qquad (10)$$

and, introducing equation (9),

$$k = \frac{1}{1 - \frac{a}{p}} = \frac{1}{1 + a\frac{dn}{dh}} \qquad (11)$$

For standard atmosphere we have $\frac{dn}{dh} = -\frac{1}{4a}$ as shown in equation (8) and from equation (11) $k = 4/3$. To state the result in

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[5]T. L. Eckersley, "Ultra-Short-Wave Refraction and Diffraction," Journal of Institution of Electrical Engineers, LXXX (1937), pp. 296-300.

another way: using the equivalent earth with radius equal to $4a/3$ corresponds to replacing the actual atmosphere, in which the index n decreases with height, by a homogeneous atmosphere with an equivalent index n'' which is independent of height. This transformation of coordinates greatly facilitates the calculation and interpretation of coverage of radio waves in the standard atmosphere, as in the case of radar coverage.

More generally, if the rate of change of n with height differs from the value $-\frac{1}{4a}10^6$ MU per meter given above, the equivalent earth radius departs from the value $4a/3$. In general the equivalent earth radius is ka. k may be expressed in terms of $\frac{dh}{dM}$ by substituting equation (7) for (11) and takes the form

$$k = \frac{1}{a}\frac{dh}{dM}10^6 \qquad (12)$$

This shows that k, under the standard condition of atmosphere, is proportional to the slope of M curve.

Since the change of earth's radius takes care of the variation of refractive index and substitutes a homogeneous for the actual atmosphere, it follows that in a diagram in which the earth is given a radius ka, the radiation propagates along straight lines. The difference is schematically shown in Figure 6 by a single radar lobe.

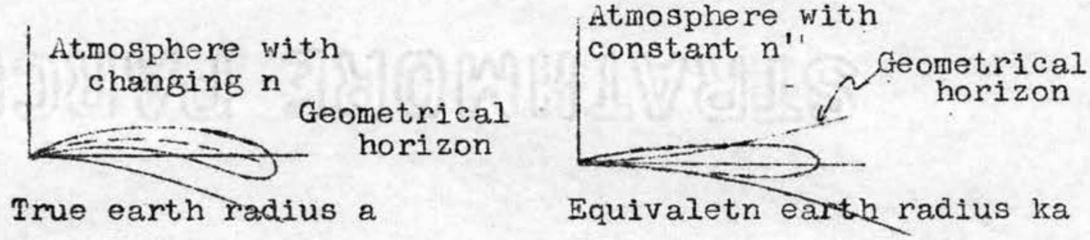

Fig. 6. Curvature of the lobes as affected by method of representation.

In figure 7, which shows the ture geometrical conditions, the radio horizon appears extended as compared to the geometrical horizon, because of the curvature of the rays.

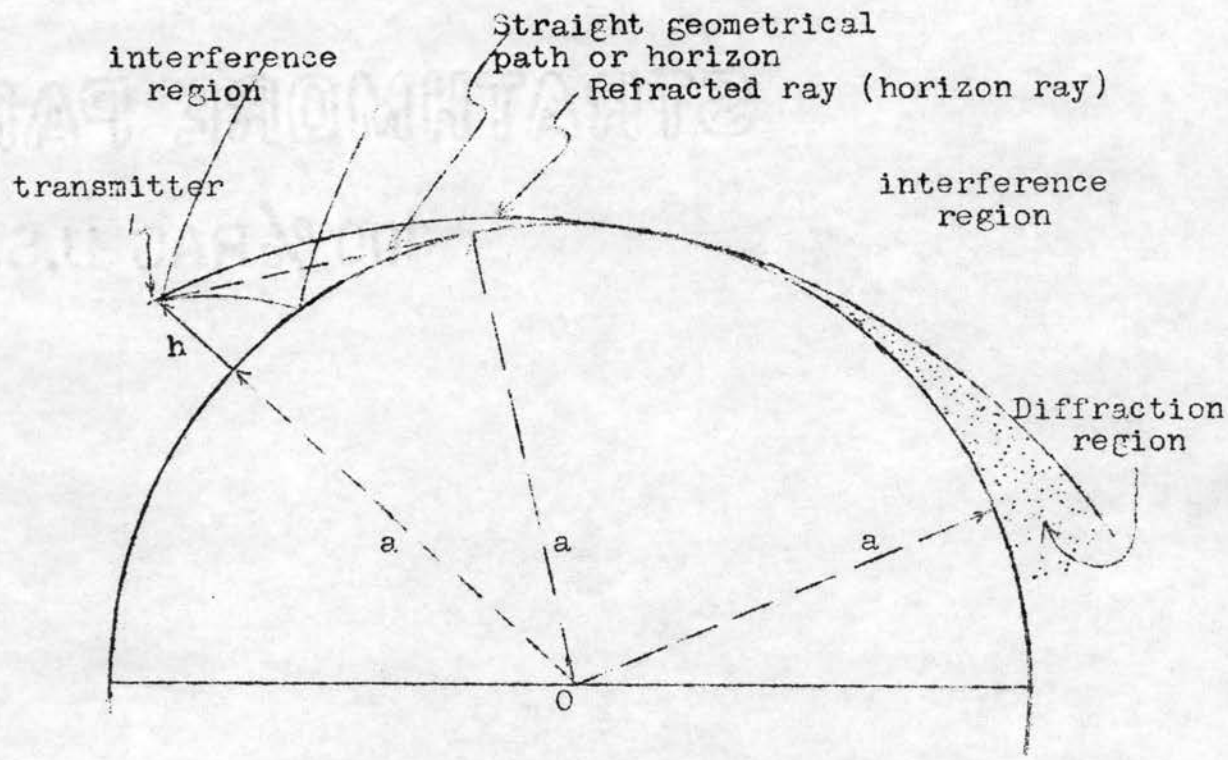

Fig. 7. Ray curvature over the earth of radius a in an actual atmosphere.

In Figure 8 the rays have been straightened out due to increased earth radius, but a line that was straight in Figure 10 appears curved in Figure 8.

The foregoing discussion has shown that a linear decrease of refractive index n can be converted into an increase of the earth's radius. The reverse process is equally feasible: to eliminate the earth's curvature by using a modified refractive index curve. This is generally a procedure which involves no assumption about the variation of refractive index with height. From equation (5) it is seen that the effect of the earth's curvature is equivalent to that of a refractive index increasing

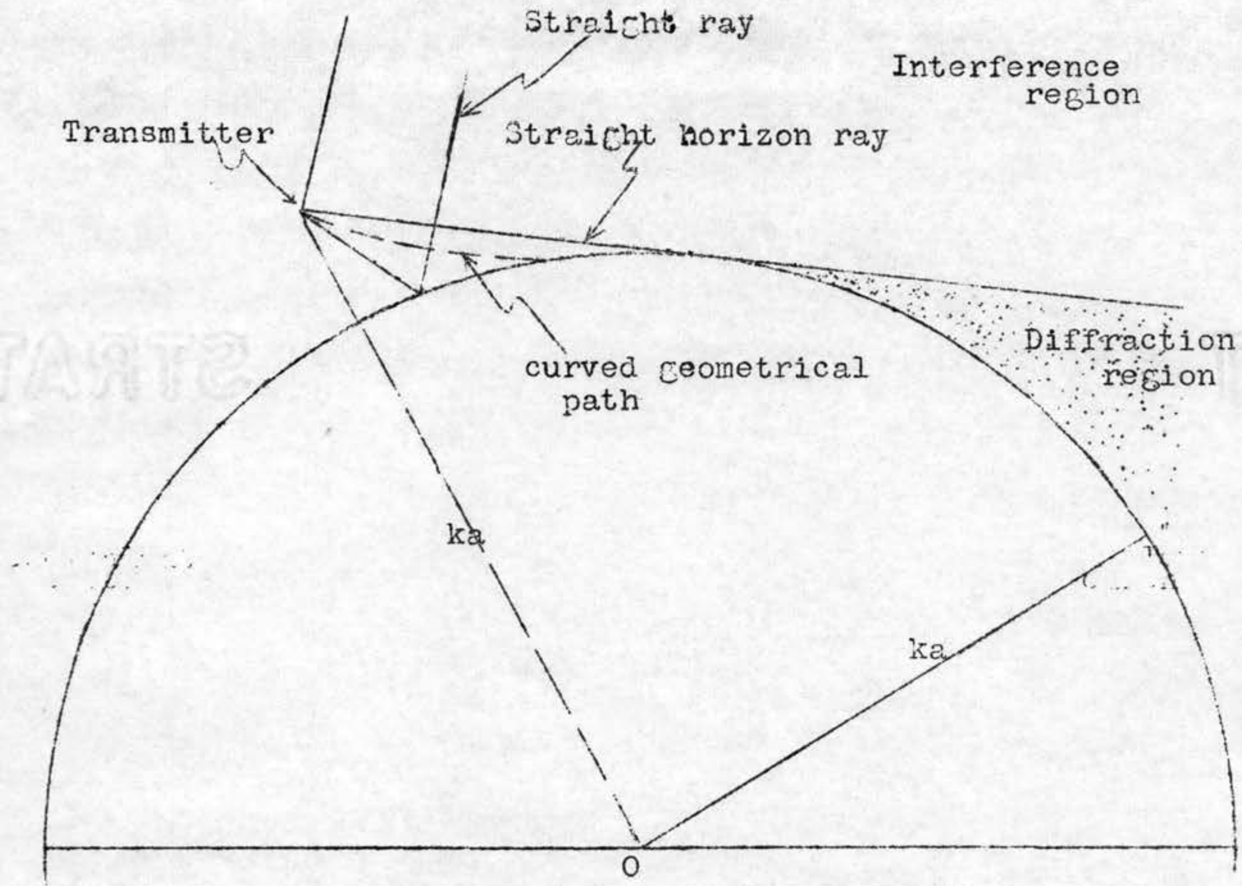

Fig. 8. Rays in homogeneous atmosphere. Equivalent radius ka. linearly with height at the rate of 1/a. Here the earth is flattened, thus eliminating the curvature effect by adding to the refractive index the term h/a. In other words, the angle between a ray and the horizontal over a curved surface of earth is the same as the angle between a ray and the horizontal over a flat surface of earth when the refractive index n has been replaced by n + h/a. If M increases linearly with height, which is the case for standard atmosphere, the rays appear curved upwards on a flat earth diagram as shown in Figure 9.

From Figure 9a it can be seen that a ray which emerges horizontally from the transmitter, will appear inclined at angle α_0 with the horizontal. It is necessary now to attempt tracing

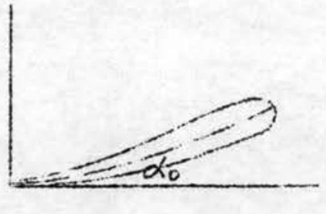

Fig. 9a
Flat earth

$k = \infty$

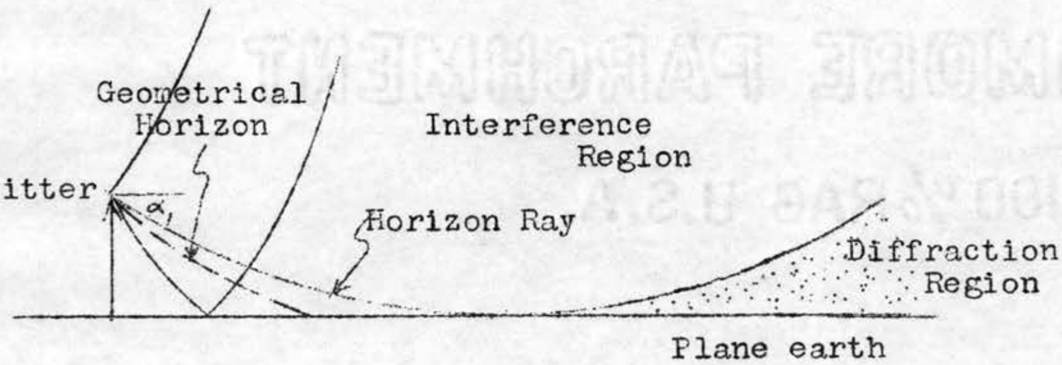

Figure 9b. Rays in plane earth diagram

the paths of rays emitted by a transmitter at various angles with the horizon, and show how their passages through the atmosphere are controlled by the variations of the modified refractive index. Returning to equation (5) we have

$$(n + h/a)\cos\alpha = n_0 \cos\alpha_0. \qquad (5)$$

For small angles $\cos\alpha$ may be replaced by $1 - \frac{\alpha_0}{2}$. Then equation (5) takes the form

$$n - n_0 + \frac{h}{a} = \frac{1}{2}(n\alpha^2 - n_0\alpha_0^2 - \frac{h}{a}\alpha^2) \qquad (13)$$

since $h/a \ll 1$, $\alpha \ll 1$, and n and n_0 are practically equal to 1, equation (13) reduces to the form

$$(n + h/a - 1) - (n_0 - 1) = \frac{1}{2}(\alpha^2 - \alpha_0^2). \qquad (14)$$

Substituting (6) and (6a) into (14), we have

$$(1/2)(\alpha^2 - \alpha_0^2) = (M - M_0) \times 10^{-6} \qquad (15)$$

Equation (15) gives a solution to the ray path since α is the angle which the ray makes with the horizontal, it is equal to $\frac{dh}{dx}$, the slope of the ray. Solving equation (15) for ,

$$\alpha = \frac{dh}{dx} = \sqrt{\alpha_0^2 + 2(M - M_0)10^{-6}}$$

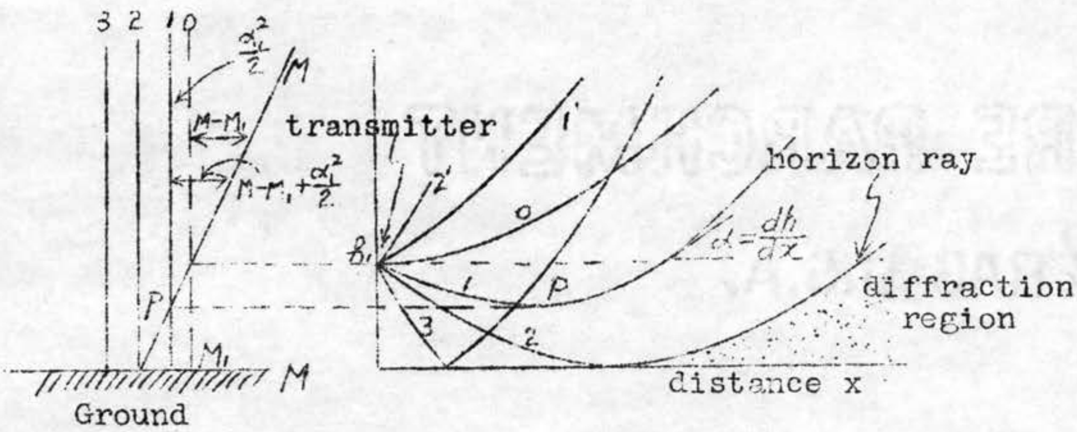

Fig. 10. Rays in standard atmosphere (plane earth).

Figure 10 illustrates the M curve with a slope ka = 4/3 a, which is the standard refraction as shown in equation (12). Let the subscript 1 stand for the transmitter level (of height h_1). Pass a vertical line through the corresponding point M_1 of the M curve. Lay off the distance $\alpha_1{}^2/2$ to the left of M_1 for a particular ray, 1, which emerges from the transmitter at angle 2, with the horizontal. α should be measured in 10^{-6} radian. The distance between M and 1 at any height h then is equal to the slope of the ray at height h. Hence, ray 1 starting downward from the transmitter is bent more and more toward the horizontal as h decreases. At point P this ray becomes horizontal and from there on increases in slope with increasing height. Ray 1' starting upward from the transmitter at the same angle $\bar{\alpha}_1$ continues to curve upward more and more rapidly as h increases. Ray 0 starting horizontally is curved upward as shown in Figure 9a. Ray 2 is the horizon ray which represents the limit to which rays can be directed by refraction. Beyond this lies the diffraction region where ray tracing can not be used. Ray 3 is reflected from the ground and in crossing some of the other rays produces the phenomena of interference. It must be emphasized that Figure 10 is a plane earth diagram in which the ordinary downward curvature of the earth has been eliminated and replaced by an additional upward curvature of the rays.

The preceding study of the propagation of radio waves in standard atmosphere, usually called orthodox propagation, serves basically as a background for that of the propagation of radio waves in non-standard atmosphere. This latter is usually called unorthodox or "anomalous" propagation, which deals with radiation in "radio ducts" caused by non-linear variation of the of the modified refractive index. Non-standard propagation takes place whenever the rate of variation of the refractive index in the lower atmosphere deviates considerably from the "standard" linear slope. The variation might consist either in a deviation from linearity, which is the most common case, or in a linear slope in the lowest layer that is widely different from the value assumed for standard atmosphere. The refractive index is a function of temperature, pressure, and the partial pressure of water vapour as given by the equation in the beginning of this chapter. The dependence of the refractive index on pressure leads to a regular decrease with height, but the change of barometric pressure with the weather produces only an insignificant effect on the gradient. The variations of refractive index in the lower atmosphere therefore are mainly due to rapid changes of the temperature and moisture with height. Temperature may sometimes increase with height for a few hundred or thousand feet above the ground and then, at greater heights, begin to decrease again. This vertical increase of temperature is called temperature inversion as explained in Chapter I. This temperature inversion must be very pronounced in order, by itself, to produce a duct. In practice, a temperature inversion contributes to duct formation when accompanied by a sufficiently strong moisture lapse, which is found when a layer of moist air near the ground is superimposed by very dry air. It is seen that such a strong moisture lapse is produced due to a rapid decrease of moisture over a short vertical distance. However, the value of the refractive index depends more particularly on the manner in which the moisture content varies with height near the surface of earth, and to a lesser extent on the distribution of temperature with height. The moisture content of the atmosphere is small at low temperatures of arctic regions and increases considerably with the higher temperatures of the tropics. Ordinary inversions in the atmosphere have a rapid decrease in moisture content accompanying the temperature rise. Typical experimental sounding curves[6] of temperature, and mixing ratio and calculated refractive modulus M show clearly the inter-relation among them. Schematic diagrams in Figure 11 below in the case of a superstandard surface layer (to be explained later) with a surface duct would give a

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[6]Kerr, op. cit., p. 330, pp. 364-367.

general idea of the profiles of the curves.

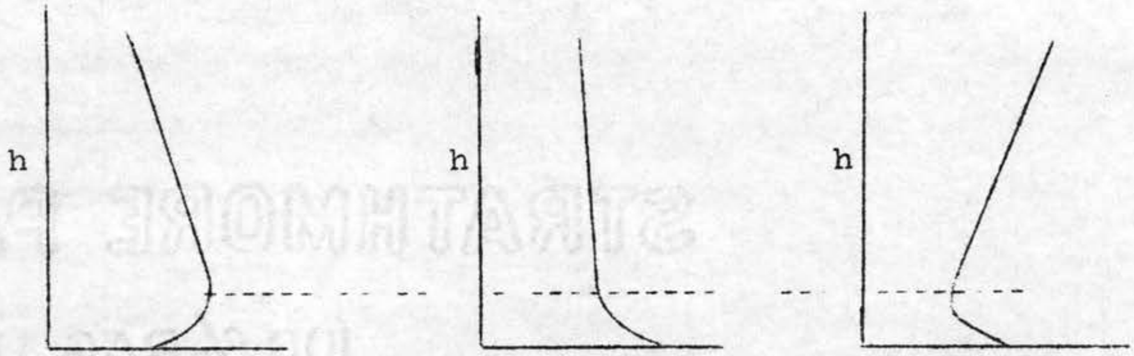

Fig. 11. Temperature, humidity, and refractive modulus profiles for a surface duct.

Two important parameters[7] in radiometeorology of duct formation are: (1) the temperature excess of the appropriate air mass in relation to the earth's surface, and (2) the humidity deficit of the appropriate air mass in relation to the earth's surface. The first is the excess of the potential temperature of the air mass over the value corresponding to the earth's surface, and the second is the deficit of the specific humidity of the air mass below that corresponding to the earth's surface. They are schematically shown in Figure 12.

Explanation must be made concerning the term "temperature inversion." It does not mean that a thermometer in the upper air at a height of, say 2,000 feet must register a temperature greater than that of the earth's surface. The reason for this is that, even for a dry, well-mixed (standard) atmosphere, there

[7]H. G. Booker, "Elements of Radiometeorology: How Weather and Climate Cause Unorthodox Radar Vision Beyond Geometrical Horizon" (Abstract), Institution of Electrical Engineers Journal, XCIII (1946), p. 461.

is a decrease of temperature in ascending through atmosphere, amounting to a drop of 5° F. per 1,000 feet of ascent. For moist air the decrease amounts to about 3.5° F. per 1,000 feet.

(a) Potential Temperature (b) Specific Humidity

Fig. 12. Modification of an air mass by surface conditions

This is associated with the reduction of density with altitude. This natural decrease of temperature must be allowed for when deciding whether the upper air is exceptionally warm. Thus the temperature of the upper atmosphere must be compared with the temperature corrected to the surface, on the assumption that temperature increases by 5° F. per 1,000 feet of descent. This corrected temperature of the upper atmosphere is known as its "potential temperature," and its excess over the surface temperature is known as the "temperature excess" (Figure 12-a).

The curves[8] in Figure 13 show the idealized modefied refractive index profiles.

A linear M-profile having standard slope is called a standard M-profile (B), calculated by using an earth's radius equal to 4a/3.a Because an M-profile seldom has the standard slope at

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[8] Kerr, op. cit. p. 14.

all heights, it is convenient to divide the atmosphere at a given time into layers, such that within each layer the gradient of M is substantially standard, or entirely greater than, or entirely smaller than, the standard value. Layers in which the gradient is greater than standard are called sub-standard, as in the presence of such layers of sufficient depth the performance

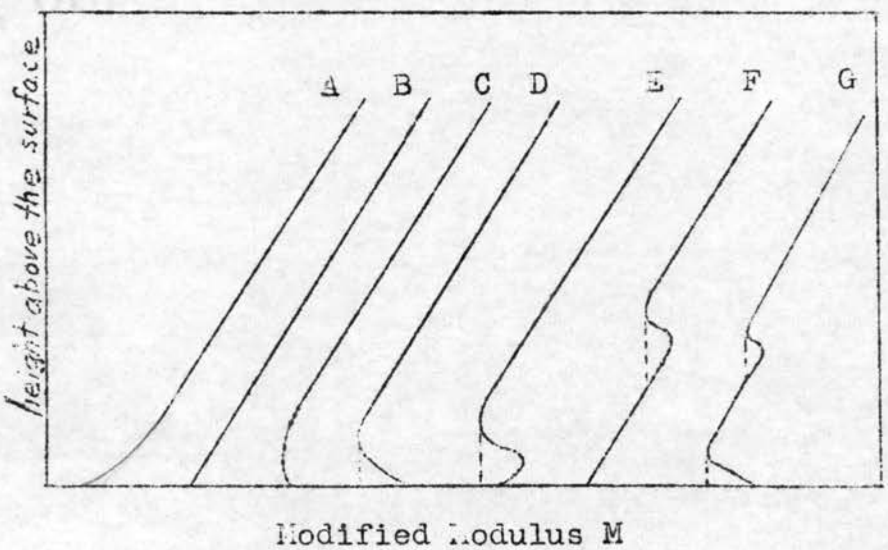

Fig. 13. Idealized modified modulus profiles

of radar and comunication facilities is generally poorer than under standard conditions. Similarly, layers for which dM/dh is algebraically less than standard are called superstandard because in their pressence performance is generally enchanced. Finally, layers in which the gradient is essentially standard are called standard layers. Note that all profiles in Figure 13 assume a standard gradient at sufficient height. Profile A shows a substandard surface layer, whereas profile C shows a superstandard layer that is not an M inversion. Curves D through G exhibit various kinds of M inversions in both surface and elevated ducts.

The causes of temperature inversions are mainly the results of advection, radiation, and subsidence as explained in Chapter I. For more detailed understanding of these causes, the reader should refer to Burrows and Atwood.[9]

When a duct is produced by subsidence inversion, the distribution of potential temperature and specific humidity with height is frequently of the type shown in Figure 14, and this leads to an elevated superstandard layer instead of one resting on the surface of the earth.

(a) potential temperature (b) specific humidity

Fig. 14. Profiles of P. T. and S. H. associated with an elevated duct

In such a case, the degree of refraction experienced in communicating between points on the earth's surface depends on: (1) the temperature excess and humidity deficit of the air mass above the layer in comparison with that below the layer, (2) the precise profile of the refractive index in the layer, and (3), vitally, the height of the layer above the surface of the earth.

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[9] C. R. Burrows and S. S. Attwood, Radio Wave Propagation (1949), pp. 39-44, and pp. 152-159.

In presence of a duct, the height of the transmitter or receiver has dominating influence upon the propagation of radio waves. Let a ray be considered emanating horizontally from a radio transmitter as the height of transmitter is varied. Let the transmitter be many thousands of feet up in the air at T_1 (Figure 15), where a ray has a radius of curvature of about $4a/3$. If the transmitter is gradually brought down to the

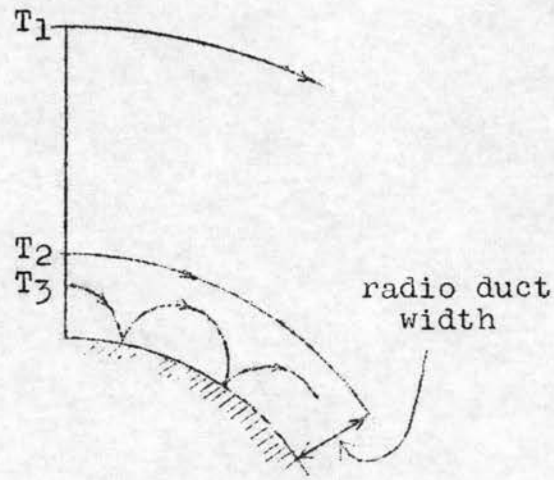

Fig. 15. Effect of radio duct on radar rays

surface of the earth, the ray is then subject to more and more downward curvature, which tends to make the ray follow the earth's curvature. When the transmitter is brought down to the top of the radio duct at T_2, the ray curvature becomes equal to the curvature of the earth. As the transmitter is brought beneath the top of the duct, say at T_3, the ray is bent downward to such an extent that it hits the ground and suffers successive reflections from it. The ray is in fact trapped within the duct. Such trapping causes radar to receive echoes from targets beyond the geometrical horizon. It is readily seen that in addition to the meteorological parameters (such as temperature excess, humidity deficit, and wind speed, which control the distribution of the radio refractive index, and therefore propagation, in the lower atmosphere), there are nonmeteorological parameters such as wave lengths, and the height of transmitter and receiver which also have a big influence upon radio propagation.

The ray tracing for a superstandard M-profile is shown in Figure 16. The method of ray tracing is the same as that for the standard M-profile as explained for tracing in Figure 10.

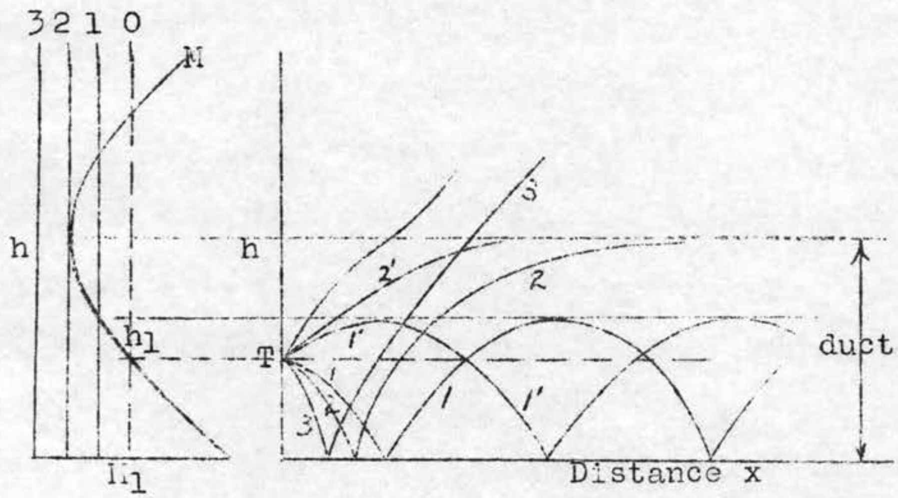

Fig. 16. Rays with a ground-based duct

Burrows[10] also made a ray tracing in an elevated duct.

It should be noticed that the effects of nonstandard or guided propagation are negligible when the angle of elevation of the ray is over 1°.[11]

In concluding this phase of microwave propagation study with respect to weather, the following general description is necessary. Broadly speaking, orthodox propagation is associated

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[10] Ibid. p. 13.

[11] Janp, 101, "Variation in Radar Coverage" (June, 1944).

with poor weather, whereas guided propagation is associated with fine weather. Weather that is cold, rough, and stormy, rainy or cloudy, usually involves a situation in which the lower atmosphere is quite well stirred up. Consequently there is no sharp discontinuity, and therefore no important downward bending. On the other hand, in weather that is fine, clear, and settled and anticyclonic, air in the upper atmosphere is descending (subsiding) and bringing potentially warm, dry air down to within a few thousand feet of the earth's surface. This situation is favorable for duct formation and there for unorthodox propagation.

It is necessary now continue discussion of microwave propagation due to the phenomena of diffraction. The term "diffraction" is a product of physical optics. Formations of image out of line of sight are usually referred to as diffraction phenomena. It is well-known from optical experiments, as well as confirmed by theory, that a screen with a narrow aperture (slit) will pass, for instance, monochromatic light to another screen behind the first one in such a way that instead of a sharp image of the opening, either a blurred wider image or several bright and dark images of the slit appear. Similar pattern are observed when two or more narrow apertures are used with a common light source. Phenomena of diffraction in optics are divided into two classes: namely, the Fresnel diffraction and the Framhofer diffraction. Fresnel diffraction theory is generally used in treatment of radiowave diffraction phenomena. Diffraction is a phenomenon accompanying all forms of wave motion, its effect being more marked as the wavelength relative to the obstacle dimensions increases. It deals with variations from straight-line wave courses when partially cut off by an obstacle, such as when an electromagnetic wave passes near edges of an opening (e.g. wedge shaped mountains) or a hole that may cause wave interference. Wave propagation behind the horizon (in geometrical shadow region of the earth) may be partially due to diffraction. It is often thought that short electromagnetic waves travel according to geometrical optics (nothing below the line of sight) but somewhat below this line if atmospheric refraction exists, experiments show that waves may reach much farther into geometrical region than refraction can count for. Such diffraction regions are clearly shown in Figures 7, 8, and 9. Experimentally it has been shown, and is theoretically true, that the field strength in the diffraction region declines rapidly with decreasing height to a minimum at ground level; the rate of decrease is larger for the higher frequencies. Neither the direct ray nor the reflected rays can penetrate into this region, which therefore receives radiation entirely by diffraction of the energy around the earth's curvature. In applying the theory of diffraction to the propagation of radio waves around the earth's curvature, it is usually simplified by treating the problem as diffraction around straight edge. Burrows[12] made a brief study of this. Eckersley[13], in 1937, had already discussed the diffraction phenomena of

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[12] Burrows, op. cit., 461-470.
[13] Eckersley, op. cit.,

ultra-short-wave propagation. The diffracted field below the horizon, in fact, takes the form of a partially-guided wave whose track-width, calculated by Eckersley(1938), extends from the earth's surface up to a height which is about 35 feet at a wave-length of 10 cm., 750 feet at 10m., and 16,000 feet at 1 km.

Booker[14] treated the phenomena of refraction and diffraction of the tropospheric propagation in terms of the mode theory of a series of characteristic E or H waves similar to those that can travel between parallel metal plates or sheets. But, whereas the track-widths of all modes occurring between the parallel metal sheets are the same and equal to the distance between the sheets, those in the atmosphere have track-widths that increase with the order of the wave. The lower edge of the track often, but not always, coincide with the surface of the earth. The height of the upper edge of track depends on the distribution of the refractive index with height and also upon the order involved. He considered that the fields in the diffraction regions were due to the leakage radiation from the top of the meteorological guide. He also mentioned that propagaton in the troposphere appears to involve two separate problems: (1) the effect of atmospheric refraction, and (2) the effect of diffraction round the curved surface of the earth. He tried to explain these two problems by reducing them into either one of diffraction or one of refraction. However, the writer of this thesis

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[14]H. G. Booker, "The Mode Theory of Tropospheric Refraction and Its Relation to Wave Guides and Diffraction," Meteorological Factors in Radio-Wave Propagation, pp. 80-121. (Repart of a conference at the Royal Institution, London, April 8, 1946.)

does not think it reasonable, because these two problems are essentially different in their fundamental nature.

Booker,[15] Straiton,[16] and Crain[17] recently used the theory of radio scattering in the troposphere to explain the existance of field strength for which refraction or diffraction can account. The theory of scattering by turbulent medium is applied to the scattering of radio waves in the troposphere. The atmosphere is a nonhomogenous medium, although mean values of the refractive index usually have the same horizontal homogeneity. The non-homogeneities are produced and supported by turbulent motion. Using atmospheric nonhomogeneities in the refractive index, the scattered power was shown to be a function of the intensity and scale of the existing turbulent variations. The atmosphere is pictured as a turbulent medium in which random fluctuation of the refractive index occur within distances of a few centimeters to a few meters, depending on the stability of the atmosphere. From their results, they found that radio field intensities observed at distances considerably beyond the line of sight have regularly been found to be much higher than would be expected on the basis of conventional refractional theory of the tropo-

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[15] H. G. Booker, "A Theory of Radio Scattering in the Troposphere," Proceedings Institute Radio Engineers (Apr., 1950), pp. 401-412.

[16] A. W. Straiton, "A Study of Tropospheric Scattering of Radio Waves," Proceedings Institute Radio Engineers (June, 1951), p. 463.

[17] C. M. Crain, "Measurement of Parameters Involved in the Theory of Radio Scattering in the Troposphere," Proceedings Institute Radio Engineers (January, 1952), p. 50.

spheric radio waves. They concluded that a radio wave beyond the horizon of the transmitter source is made up of a component due to refraction and a component due to scattering, the latter often being as strong as, or stronger than, the refraction component, depending upon the point to be considered. Turbulence results from frictional forces generated as the result of surface drag and wind shear--mechanical turbulence--and the bouyancy forces derived from the heat received at the earth's surface by solar radiation--thermal turbulence. In a typical model of turbulence, eddies set up as a result of mechanical or thermal forces produce the observed fluctuations at a point through a mixing process.

When an electromagnetic wave passes over an object having dielectric properties differing from those of the surrounding medium, some of its energy is absorbed by the object and appears as heat, while some is scattered in all directions without change of wave-length. Both phenomena are entirely negligible at wave-lengths greater than about 10 cm., but as the wave-length decreases, the scattering and absoption becomes important, until at wave-lengths around one centimeter they place a limitation on transmission over appreciable distances through rain. Scattering will be briefly discussed in the next chapter. The only gases[18] in the atmosphere that need consideration for absorption here are water vapour and oxygen. The reaction occurs in this case because the molecular structure of these gases is such that the individual molecules behave like dipoles; that is, they possess

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[18] Kerr, op. cit. p. 26.

permanent dipole moments, which may be either of electric or magnetic type. These dipole moments furnish the mechanism by which the electric or magnetic field of a passing wave reacts with the molecules, causing them to rotate end over end as to oscillates in many other possible ways. The molecule of water vapour has a dielectric moment which interacts with the electric field of the radiation and has a resonance wave length near 1.33 cm. Oxygen, on the other hand, is paramagnetic and has a molecule with a magnetic dipole moment which interacts with the magnetic field of the radiation and has resonance wave lengths of 0.5 cm. and 0.25 cm. Nitrogen and other rare gases are without effect. Water vapour and oxygen together cause gaseous absorption to become noticeable at a wavelength as small as 1.5 cm. and very pronounced below 1 mm. The resulting attenuations produced by the presence of oxygen and water vapour have been studied by Van Vleck (1942, 1945). From his work it appears that the effects are not large. Thus, over the centimeter band (3 to 10) one may expect attenuations of the order of 0.01 dk/km to be produced by the oxygen and water vapour content of the atmosphere. The attenuation is just beginning to be apparent around a wave length of 10 cm.

Anderson[19] obtained data showing that the average attenuation of 1.25 cm. radiation due to rainfall is 0.37 db/mile/mm./hr. (the theoretical calculation made by J. W. Ryde yields an average of 0.25 db/mile/mm./hr.). Drop size measurements were

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[19]L. J. Anderson, "Attenuation of 1.25 cm. Radiation Through Rain," Proceedings of Institute of Radio Engineers, XXXV (1947), p. 351.

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also made but no conclusions could be drawn because of the wide scatter in the drop-size distribution obtained. He also showed that if the maximum distance over which communication can normally be established is 100 miles, light to moderate rain (about 1 - 4 mm./hr.) will reduce the range to about 10 miles. The radiated power must be increased by 10^{20} (200db) times in order to re-establish communication over 100-mile path.

Roberston[20] studied the effect of rain upon 1 - 4 cm. waves with experimental observations. At a wavelength of 1.09 cm. the waves are appreciably attenuated, even by light rain. Attenuations in excess of 25 db/mile have been observed in rain of cloudburst proportions. The attenuation of waves somewhat larger than 3 cm. is slight for moderate and light rainfall. During a cloudburst, however, the attenuation may approach 5 db/mile. The results are tabulated below.

Wavelength	db/mile at Light rain (1 mm./hr.)	db/mile at Moderate rain (4 mm./hr.)	db/mile at heavy rain (15 mm./hr.)	db/mile at cloud burst (100 mm./hr.)
= 3.20 cm.	?	<0.5	<1	4 to 5
= 1.09 cm.	<1	1 to 2	4 to 10	30 to 40

Studies and researches in the field of microwave propagation are still in the beginning processes of development. Much work has to be done in order to have complete understanding of every phase of microwave propagation. Later studies on millimeter radar wave propagation will undoubtly bring new information, as well as new problems, in this field.

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[20] S. D. Roberston, "The Effect of Rain Upon the Propagation of Waves in the 1 and 3 Centimeter Regions, "Journal of the Institute of Electrical Engineers, XXXIV (Apr., 1946), p. 178.

CHAPTER V

RADAR STORM OBSERVATION

The principles of radar and its operation will not be presented here since the reader may refer to any of the textbooks such as <u>The Principles of Radar</u> by the staff of the Massachusetts Institute of Technology, <u>The Principles and Practice of Radar</u> by H. E. Penrose, etc. The theory for the application of radar to storm detection will be briefly reviewed in order to give a comprehensive insight of the object in mind, but for detailed theoretical treatment in this connection, the reader should refer to the original papers by G. Mie and J. W. Ryde and the book <u>Propagation of Short Waves</u> by D. E. Kerr.

Observation and detection of precipitation by radar is possible because of the scattering of high frequency radio waves by precipitation forms. The basic theory of scattering of electro-magnetic waves by dielectric spheres was first published by Mie (1908). Ryde made the application of the theory to the detection of precipitation (1946). Both observational experience and theoretical treatment indicate that radar echoes from precipitation are accounted for by scattering of water drops or precipitation in frozen form. In a first approximation, this scattering is calculated by the well-known Rayliegh Law, when the drops are sufficiently small compared with wavelengths, and in this case the amount of radio energy scattered is inversely proportional to the fourth power of the wavelength and directly to the summation of the sixth powers of the radii of the droplets contained in unit volume of the precipitation illuminated, and inversely as the square of the range (assuming total interception of the radar beam by precipitation area). Spatial non-uniformity of precipitation and lack of information as to the distribution of drop size have introduced so many complications that no strict quantitative analysis has yet been made.

Ryde[1, 2] has shown that, if P_r is the received echo power,

$$P_r = \frac{P_0 GANS(1-\cos\theta)W}{4R^2}$$

where P_0 = transmitter peak power

A = effective area of the aperture of aerial

G = Gain of aerial

θ = half angular aerial beam width

R = distance of scattering object from the radar set

N = no. of scattering drops or water particles per unit volum of air

S = scattering function of each drop in the direction of aerial

W = pulse width

For drops which are small compared with an incident radar wavelength λ, the value of NS is given by:

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[1] J. W. Ryde, "The Attenuation and Radar Echoes Produced at Centimeter Wavelengths by Various Meteorological Phenomena," Meteorological Factors in Radio Wave Propagation, (Physical Society Report, London, 1946).

[2] J. W. Ryde, "The Attenuation of Centimeter Radio Waves and Echo Intensities Resulting from Atmospheric Phenomena," Journal Institution of Electrical Engineers, XCIX (1946), Part IIIA, p. 101.

$$NS = \frac{\Sigma N(D)D^6}{\lambda^4} f(k)_\lambda$$

where D = drop diameter

k = dielectric constant of the drop at wavelength λ

The average cloud is composed of water droplets of radii lying between 8 to 30 microns (one micron = 10^{-6} meter), there being about 200 droplets per c.c. Rain, of course, comprises drops of various diameters ranging from 0.01 cm. to 0.08 cm. in heavy rain. Some knowledge of the drop size distribution is therefore essential to evaluate the expression $\Sigma N(D)D^6$. Laws and Parsons[3] have compiled the average drop size distribution with precipitation rate.

The above mathematical relationships were based upon and developed by assuming negligible attenuation of the wave in its passage to the precipitation area. In practical experiments and observations it is found, as discussed in the preceding chapter, that under certain circumstances atmospheric attenuation may be appreciable. The degree of attenuation varies with the wavelength of electromagnetic wave and the nature and size of attenuating particles. From the foregoing equations, it can be seen that in the absence of rain attenuation the received power for 3.2 cm. waves is approximately 100 times greater than for 10 cm. waves. When observational comparisons were made on the basis of equal radar characteristics for both the 3.2-cm. set and the 10 cm. set, it was found due to different degrees of

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[3] J. O. Laws and D. A. Parsons, "The Relation of Raindrop Size to Intensity," Transactions of the American Geophysical Union, XXIV (1943), Part II, p. 452.

attenuation, that the 10-cm. wave is more suitable for storm detection through appreciable distances of heavy rain while the 3.2-cm. wave is better for that of moderate and light rain.

Not all kinds of clouds can be detected by radars of cm. wavelengths, but cumulonimbus clouds always produce radar echoes because the water droplets in these clouds increase in size to such a dimension as to cause effective scattering effect on cm. waves. Cumulonimbus clouds, although they occur in packs, are not densely packed, there being quite relatively channels among droplets in the clouds. These channels can be traced and closely traveled in by radar waves. Precipitation forms such as rain, snow, hail, sleet, etc. always give strong radar echoes.

The information presently available in the field of radio-meteorology by means of radar observation is as follows:

1. Instantaneous location of all precipitation over several thousand square miles horizontally and all altitudes from a single radar observing station.

2. Direction and velocity of precipitation movement.

3. Qualitative information concerning intensity of precipitation.

4. Heights of cloud bases and tops.

5. Approximate height of freezing level.

6. Information as to whether or not certain storms are thunderstorms.

7. Position, direction, and speed of hurricanes.

8. Distribution of fall-velocity for precipitation particles at any level above the radar.

9. Qualitative information regarding water-vapour and temperature distribution in the vertical.

10. Growth rate of convective cells.

Not all of the information contained in the list above is yet obtained to a useful degree of accuracy. Considerable effort is being made to improve the accuracy of such measurements.

Precipitation may be divided into various types, using basic causes as the means of classification, as follows:[4]

A. Frontal precipitation

 (1) Cold front

 (2) Warm front

B. Instability showers and thunderstorms

C. Hurricanes and typhoons

This classification is convenient for the purpose of description, but many variations and combinations occur, and in practice such large variations in degree occur that identification is not always easy. The same problems are encountered in radar observation of precipitation, and often echoes are observed that do not fit well into this simplified classification.

A. Frontal precipitation. (1) Cold front. Perhaps the most striking and easily understood of all radarscope display is that of echo signals from the squalls associated with an active cold front. The photograph of PPI scope in Fig. 1 shows a typical cold front.

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[4]M. G. H. Ligda, "Radar Storm Observation," Compendium of Meteorology, 1951, pp. 1271-1276.

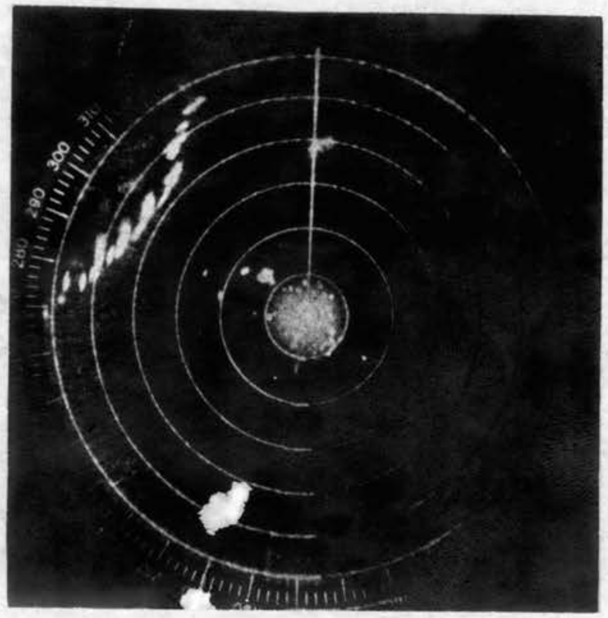

Fig. 1. Photograph of PPI scope showing approaching cold front, 20-miles markers.

This generally appears as a band of well-defined echoes on the plan position indicator. The echoes may break up and join several times in their movement across the scope. Radar observations of the approach of a cold front show the following sequence of events: Scattered storm-echo signals are first detected at maximum (100-200 miles) ranges depending upon the activity of the front. These echoes are caused by hydrometeors in the upper portion of the tallest cumulonimbus clouds along the front, and generally lie in an arc which may be closely identified with the portion of the front as reported by surface observation stations. As the front approaches, the radar detects precipitation at successively lower levels and the original cells appear to increase in size and intensity. When the nearest portion of the front is about 50 miles away, a large portion of it appears to be a solid line of precipitation if the antenna elevation angle is kept at or near zero degree. The inexperienced observer will sometimes conclude that the front is actually intensifying, whereas the radar is simply detecting rain at lower levels which is almost invariably more widespread. This trend continues until the front passes over the radar, at which time echoes from the more distant storms along the front may disappear from the scope entirely because of rain attenuation. At this time the precipitation may appear to be almost evenly distributed around the radar for a distance of many miles. After the frontal passage, the above sequence of events is reversed until the front passes beyond the maximum range of detection or dissipates. The height of the cold front precipitation echo is predominantly greater than that of the warm front due to the greater vertical development of convective cells.

(2) Warm front. Interpretation of radar displays resulting from warm-front precipitation is considerably more difficult than the interpretation of echoes from cold-front precipitation. This is a result of larger area irregularly covered by the precipitation and the possible instantaneous variation of conditions. The precipitation causing the echo signals exhibits varying degrees of convective activity depending upon the stability and convective stability conditions in the air masses involved. When conditions are stable in both air masses, the return appears as shown in Fig. 2. During more unstable warm frontal conditions, the PPI scope will show stronger echoes from

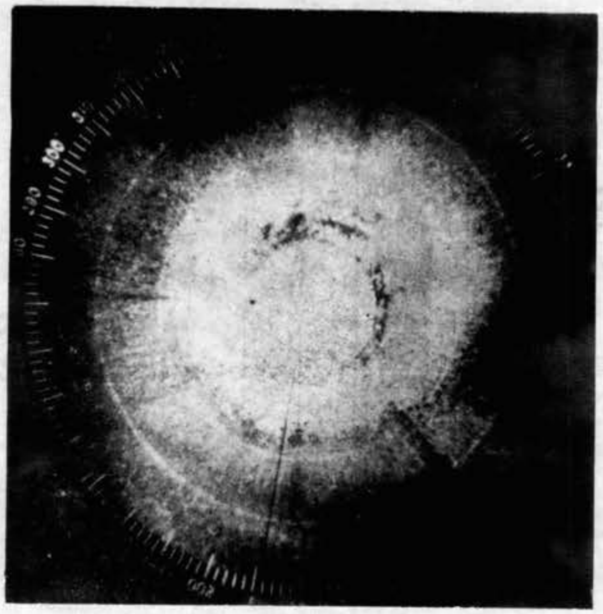

Fig. 2. Stable warm-frontal precipitation echoes on a PPI scope, 5-mile markers.

individual storm cells. Absence of uniform or systematic patterns is frequently observed.

B. Instability showers and thunderstorms. Instability showers, being largely convective in nature, generally show somewhat finer detail in echo signals on the PPI scope, as shown in Fig. 3, than does warm-frontal precipitation. There is an even greater lack of symmetrical arrangement of the convective cells.

The appearance of thunderstorm echo signals on PPI is similar to that shown in Fig. 4. The difference between the thunderstorm and the showers appears to be a matter of degree of convection and the dividing line is not sharply defined. These storms, because of their great height due to strong convection, are detectable at greater distances than any other type of precipitation. Using radar it is possible to determine when a

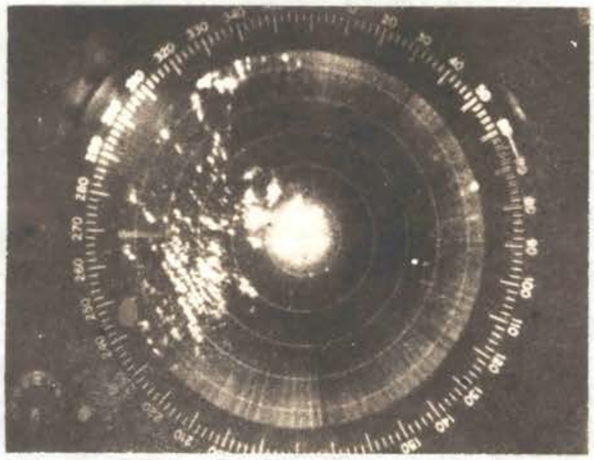

Fig. 3. Photography of PPI scope showing rain showers, 10-mile markers.

given convective cell reaches the thunderstorm stage. The best vertical representation of thunderstorm convection echo signals is found on RHI scope.

Fig. 4. Photography of PPI scope showing thunderstorm at close range, 5-mile markers.

C. Hurricanes or typhoons. Radar detection of hurricanes and typhoons is even more dramatic than radar detection of cold-front squall lines. A typical typhoon is shown in Fig. 5. The "eye" of the storm and the rotary motion of the winds are discernible from precipitation echoes.

Fig. 5. Photograph of PPI showing a typical typhoon.

No other storm has been known to produce such a distinctive echo for any length of time as the tropical hurricane. By the time a hurricane is within range of land-based radars, its direction and speed are usually well known. However, these radars are useful for precise determination of the storm's position at any instant, and provid very valuable up-to-the-minute data on the storm. It has been found by experience that the best wave length for use in hurricane detection is about 10 cm., at least from the standpoint of rain attenuation, as mentioned above. The importance of radar as a research tool for studying the structure of hurricanes should not be overlooked. It may be possible to measure wind speed in different portions of hurricanes by means of radar, and to obtain clues concerning the structure of these severe storms.

One of the most interesting properties of the echoes observed under conditions of widespread precipitation is pronounced stratification.[5] Sometimes well-defined, nearly continuous horizontal layers are indicated for a considerable distance, whereas, at other times multi-layers of limited extent are observed. If a radar with PPI is pointed directly upward at an echoing layer, the echo appears as a circular ring. If the antenna is tilted downward, the diameter of the ring increases in the expected manner with the angle, confirming the existence of a layer. A typical example of a layer-type echo is shown in Fig. 6.

Observation on radar echoes from stratified precipitation indicates a horizontal band which produces a stronger return than the parts above and below it. This has also been observed[6] in connection with thunderstorms after convective activity ceased. It is descriptively designated as the "bright band." While certain detailed processes of bright-band formation are the cause of dispute, all investigators agree that it is in some way connected with the change of state which occurs at the freezing level. Two theories[7] have been advanced to explain

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[5] D. E. Kerr, Propagation of Short Radio Waves, 1951, p. 634.

[6] A. C. Bemis, "A Qualitative Study of the 'Bright Band' in Radar Precipitation Echoes," Journal of Meteorology, VII (1950), pp. 145-151.

[7] M. G. H. Ligda, op. cit. p. 1274.

this phenomena:

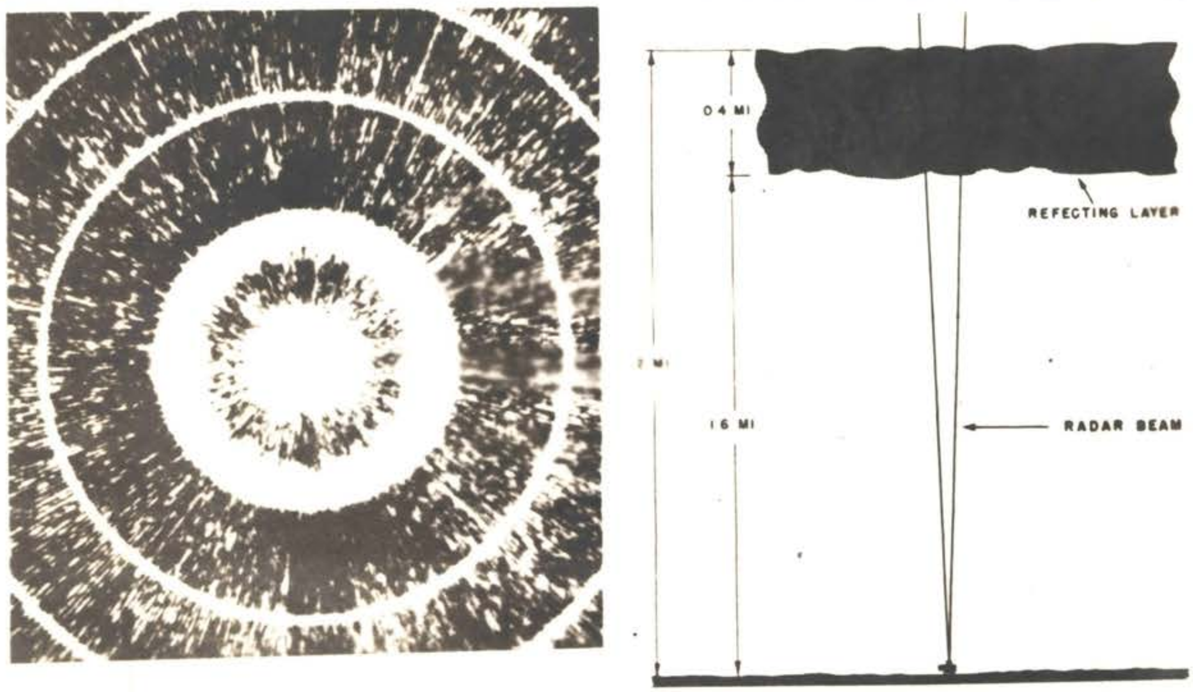

Fig. 6. Photograph of PPI scope showing layer-type echo, 1-mile markers. Antenna directed straight up.

1. The region of strong echo is due to the drop formation in the colloidally unstable layer of heterogeneous ice-water mixture. Any precipitation detected above that altitude would therefore probably be caused by convective transport.

2. Snow particles, too small to give more than a weak echo, fall to the zero degree isotherm, and melt at or just below this level. While melting, they have the low fall velocity of snow-flakes, but high reflectivity of water. Coalescence, often observed near melting temperatures, also serves to increase the reflectivity. After the snowflakes have completely melted, the fall velocity increases and the drops become widely spaced. The result is a region of weaker echoes below the level of melting.

Mather[8] made a study of the dimensions of precipitation areas as they appear on the radar scope. Measurements of vertical thickness and horizontal diameter of these areas are made for both cold- and warm-frontal weather situations. It is found that convective cells associated with cold fronts are taller in vertical extent and smaller horizontally than those associated with warm fronts. There is a small seasonal variation. Echo dimensions are also compared with stability of the air. As the air becomes more unstable, the precipitation echoes become larger both vertically and horizontally. Hilst[9] made a study of the initial growth of thunderstorm cells, and his results gave a general idea as to the rate of growth of convective cells.

By means of radar, regions of more intense precipitation in a given storm can be located with fair accuracy by reduction of receiver "gain" to a point where the strongest echo signal shows on the PPI scope. By plotting the outline of the storm between successive equal gain reductions, the storm echo signals are reduced to a number of contours, each contour representing a level of equal echo-signal power. In general it is found that the strongest echo signals, which are closely associated with more intense rainfall, are near the center of storm-cell echo signals.

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[8] J. R. Mather, "An Investigation of the Dimensions of Precipitation Echoes by Radar," Bulletin American Meteorological Society, XXX (1949), p. 271.

[9] G. R. Hilst, "Radar Measurements of the Initial Growth of Thunderstorm Precipitation Cells," Bulletin American Meteorological Society, XXXI (1950), pp. 95-99.

Radars operating near 1 cm. (K-band) have not been developed beyond the laboratory stage. However they are promising in the field of detection of fine clouds and fogs. Such equipment may eventually prove valuable for reporting the rate of approach or development of coastal fog or other phenomena.

Because of the comparative recency of the application of radar as a meteorological observation instrument, problems concerning detection of some meteorological phenomena are still unanswered for the present. Eventually it may be expected that description of radar detection of such phenomena as tornadoes, waterspouts, dust storms and other atmospheric anomalies will be available. The usefulness of radar for storm detection can be greatly amplified by improving equipment and observational techniques. On account of the ease of radarscope interpretation in locating the precipitation areas, television may be an excellent medium to convey the image of the PPI scope to the citizens for their instant information of weather changes.

Attempts[10, 11] have been made to measure the rainfall and raindrop size by means of radar, but due to the nonuniformity of drop dimensions, precise results cannot be obtained yet.

The detection of tornadoes by means of radar is still in experimental stages, and therefore detailed analysis on statistical thunderstorm and tornado precipitation echo pictures, both of

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[10] P. M. Austin, "Measurement of Approximate Raindrop Size by Microwave Attenuation," Journal of Meteorology, IV (1947), pp. 121-124.

[11] J. S. Marshall, "Measurement of Rainfall by Radar," Journal of Meteorology, IV (1947), p. 186.

RHI type for vertical growth and decay rate study and PPI type for horizontal study, must be made in order to yield results which will support or give hints to the theory of tornado formation as well as to supply precise information for predicting and identification. A word might be said here that it is not likely possible to see the tornado funnel echoes on the radarscope since the funnel itself is a whirling air mass which does not contain enough water content to yield a precipitation echo. The interpretation of precipitation echoes associated with tornado situations depends mainly on the synoptic data available and a proper hypothesis of tornado formation. These problems remain in the realm of meteorologists. It is necessary again to emphasize the importance of close cooperation between radio engineers and meteorologists. The radar pictures made by Tepper[12] of a tornado situation indicate a band of precipitation echoes corresponding in both time and place with the tornado. It was found upon careful observation that all available synoptic data were compatible with the conclusion that the tornado situation was associated with two intersecting pressure jump lines. These pressure jump lines were considered to have produced the pattern of precipitation echoes indicated on the radarscope. Furthermore, the occurrence of a tornado in the intersection of two pressure jump lines is in accordance with the tornado hypothesis suggested by Tepper. The precipitation echo pictures obtained

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[12]Morris Tepper, "Radar and Synoptic Analysis of a Tornado Situation," Monthly Weather Review, LXXVIII (1950), pp. 170-176.

at O.I.T. are still not complete enough to yield conclusions, and will not be discussed in this thesis.

The author strongly recommends that an additional large number of observation stations be set up within, and outside of, the range of the O.I.T. radar observation system so that enough synoptic data may be available for precise interpretation of precipitation echoes. The results from the sferic detection system and the direction finder system will also supply valuable data for the interpretation of precipitation echoes associated with tornadoes. With integration of these systems of tornado detection, research work on the identification and tracking of tornadoes has a promising future.

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