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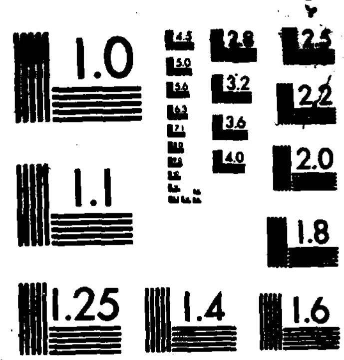

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VERTICAL PROFILES OF TEMPERATURE AND HUMIDITY BELOW 100 METERS

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Research Directorate
Research, Development, and Engineering Center

APRIL 1986

U.S. ARMY MISSILE COMMAND
Redstone Arsenal, Alabama 35898-5000

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Vertical gradients of temperature and humidity in the lowest 100 m of the atmosphere are extremely variable. The shape of a temperature profile depends upon atmospheric stability and upon radiational heating or cooling at the surface. These factors and the availability of moisture at the surface influence humidity profiles. Diurnal changes in the lowest few meters are particularly large when clouds are absent. This report also discusses effects of topography and intense frontal zones. Summaries are in two tables.

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I. INTRODUCTION

Temperature in the troposphere decreases with altitude in the U.S. Standard Atmosphere and in the U.S. Standard Atmosphere Supplements (Environmental Science Services Administration et al., 1966) for different latitudes and seasons. The constant rate of decrease of temperature with height in the troposphere of the U.S. Standard Atmosphere is 6.5 °C/km.

The rate at which temperature decreases with height is called the lapse rate. By convention the lapse rate is defined as the negative of the rate of change of temperature with height [1],[2]. The reader is cautioned that an occasional author or translator fails to follow this convention [3]. Because a negative change of temperature with increasing altitude is considered normal, an increase of temperature with height is called an inversion. The lapse rate in an inversion is negative.

The surface temperatures in standard weather data are measured by thermometers between 1 m and 2 m above the ground. In the United States these thermometers are in white instrument shelters which have louvered sides to provide adequate ventilation [4]. The shelters are mounted on a stand 4 ft (1.22 m) above the ground, and the thermometers are located at a central point within the shelter. This places the bottom of the thermometer 54 in (1.37 m) above the ground.

Actual vertical gradients of temperature near the surface are extremely variable and often deviate considerably from the Standard Atmosphere. The gradient is not constant with height in the real atmosphere, and diurnal changes are large near the surface. Godske et al. [5] have pointed out that during the daytime in summer the soil frequently becomes 20 °C warmer than the air temperature measured inside a radiation-shielded screen at 2 m. Desert surfaces can be 25 °C to 30 °C warmer than air between 1 m and 2 m when the sun is high and winds are light [6]. Temperature profiles over land normally have inversions in the lowest few meters at night. Extreme magnitudes of temperature gradients in the lowest 2 m are usually smaller at night than during the day.

Magnitudes of vertical temperature gradients between 2 m and 100 m are typically smaller than magnitudes below 2 m. This is especially true in the middle of the day when the temperature normally decreases with height, i.e., the lapse rate is positive and the vertical rate of change of temperature is negative. Lapse rates greater than the autoconvective lapse rate of 0.034 °C/m are not readily maintained for long periods throughout such a thick layer, and they are rare between 50 m and 100 m. On the other hand, there is no definite limit on the magnitude of the positive vertical changes with height in nocturnal inversions, and these may be larger than 0.10 °C/m throughout the lowest 100 m.

Available information on the vertical variation of absolute humidity indicates that it also has a diurnal variation. Absolute humidity usually decreases with height during the middle of the day, and it is not unusual for it to decrease at night. However, inversions are also frequent at night.

The remainder of this report discusses variations of temperature and humidity profiles with time of day, season, and geographical location.

II. TEMPERATURE PROFILES

This section contains information from a few long-term studies and from several studies with limited data. Although long-term investigations have been carried out at very few geographical locations, they are still quite useful. Lapse rates from extensive measurement programs are usually stratified according to such meteorological variables as time of day, cloud cover, wind speed, and stability. This makes it possible to use available meteorological data from other sites to infer frequency distributions of lapse rates at those locations.

Extensive measurements at Porton in southern England have been discussed by Geiger [3], Godske et al. [5], and Sutton [7]. The observational program included some measurements very near the surface. Geiger gave two-hourly monthly mean vertical temperature gradients for the height intervals 0.025 m to 0.30 m and 0.30 m to 1.20 m. The largest magnitudes occurred in June at noon when the mean vertical changes were -6.82 °C/m in the lower layer and -0.77 °C/m between 0.30 m and 1.20 m. Mean lapse rates between 0.025 m and 0.30 m were greater than 4.00 °C/m from 1000 hours through 1400 hours during May through August and also at 0800 hours in June and 1000 hours and 1200 hours in April. In December at noon the mean lapse rate in the layer from 0.025 m to 0.30 m was only 0.65 °C/m, which was less than one-tenth as large as the mean in June at noon.

The variations of the strength of inversions throughout the year and throughout the night at Porton were much smaller than fluctuations of temperature gradients during daylight hours. Largest mean inversions occurred in March at 2000 hours when the means were 2.58 °C/m from 0.025 m to 0.30 m and 0.58 °C/m from 0.30 m to 1.20 m. The smallest monthly mean temperature increases at this hour were 1.01 °C/m in May for the lower layer and 0.21 °C/m in July for 0.30 m to 1.20 m. The difference between the smallest and largest monthly means at 2000 hours in each layer was less than a factor of three. If one considers the hours between 2000 and 0200, the smallest monthly mean vertical gradient from 0.025 to 0.30 m was 0.54 °C/m at 0200 hours in July. This was slightly more than one-fifth as large as the 2.58 °C/m at 2000 hours in March. The pattern of the month-to-month variation of the vertical temperature gradient was irregular throughout the night. For example, the means in degrees Celsius per meter for the layer 0.025 m to 0.30 m at midnight for January through December were as follows: 1.09, 1.55, 1.97, 1.71, 0.89, 1.27, 0.83, 1.01, 1.05, 1.11, 1.35, and 1.61.

Godske et al. [5] considered the difference between clear and cloudy nights in June and December at Porton. Average vertical temperature gradients between 0.025 m and 0.30 m during the period between 2100 hours and 0300 hours in June were 3.01 °C/m for clear skies and 0.23 °C/m for cloudy skies. The gradients in December were 2.06 °C/m for clear skies and 0.29 °C/m for cloudy skies. Inversions on clear nights in the layer from 0.30 m to 1.20 m were also stronger in June, and those on cloudy nights were stronger in December. The vertical temperature gradients in this layer on clear nights were 0.60 °C/m in June and 0.52 °C/m in December. The very small gradients on cloudy nights were 0.035 °C/m in June and 0.051 °C/m in December.

Geiger [3] and Deacon [6] have provided graphical representations of the diurnal variation of temperature at Porton at the levels 1.2 m, 7.1 m, and 17.1 m. Vertical temperature gradients for clear and cloudy days in June and December could be computed by obtaining differences of temperatures from the graphs. The difference between temperatures at 1.2 m and 7.1 m near the middle of a clear June day was approximately 1.3 °C which corresponds to a lapse rate of 0.2 °C/m. The maximum lapse rate between 7.1 m and 17.1 m on clear June days was about 0.05 °C/m. The midday lapse rates on cloudy June days were 0.08 °C/m and 0.02 °C/m in the lower and upper layers, respectively. Maximum lapse rates during the day in December were about one-fourth as large as those in June on clear days and about one-half as large on cloudy days. On overcast nights in both June and December the atmosphere was essentially isothermal from 1.2 m to 17.1 m. On clear nights the rate of temperature increase between 1.2 m and 7.1 m could reach 0.2 °C/m in December and slightly larger values in June. The rate of change from 7.1 m to 17.1 m on clear nights was about 0.04 °C/m in December and was sometimes fifty percent larger in June.

Godske et al. [5] also included graphs of vertical variations of temperature in summer at Neiderling in Bavaria, Germany. The differences between the temperature at 0.05 m and 0.50 m during the middle of the day were approximately 5.2 °C for cloudless skies and 2.9 °C for overcast skies. The lapse rates were therefore 11.6 °C/m and 6.4 °C/m. Between 0.5 m and 1.5 m the temperature in the middle of the day decreased almost 1 °C with cloudless skies and about 1/3 °C with overcast skies. On cloudless summer nights the temperature increased approximately 1.8 °C from 0.05 m to 0.50 m and another 1.2 °C from 0.5 m to 1.5 m. On overcast summer nights in Neiderling the temperature was about 1/2 °C greater at 1.5 m than at 0.05 m.

The mean daily cycle of temperature in the lowest two meters at Belgrade, Yugoslavia, was presented by Godske et al. [5] to be representative of continental temperatures in summer. At about an hour after noon, the temperature difference between the ground surface and 0.4 m was 13.2 °C. The temperature difference between 0.4 m and 2.0 m was about 2/3 °C, and this gives a lapse rate of 0.4 °C/m. Temperature inversions existed on summer nights at Belgrade, and they were strongest near sunrise. At this time the rate of increase of temperature between the ground and 0.4 m was 1.2 °C/m, and the rate between 0.4 m and 2.0 m was 0.6° C/m.

Some measurements near the surface have been made in the United States. Geiger's [3] graph for clear spring days at Seabrook, New Jersey, showed that at 0.1 m the maximum temperature was reached a few minutes after the time of maximum radiation at noon. The maximum at 6.4 m was approximately 2.4 °C lower and was reached at about 1430 hours. Largest lapse rates occurred a few minutes after 1200 hours when the temperature at 0.1 m was 3.1 °C greater than the temperature at 6.4 m. This corresponds to an average lapse rate of 0.5 °C/m. Lapse rates in different layers deviated considerably from this average. Lapse rates at midday were approximately 5 °C/m between 0.1 m and 0.2 m, 1.7 °C/m between 0.4 m and 0.8 m, and 0.1 °C/m between 3.2 m and 6.4 m. The strongest inversions at night occurred between one and two hours before sunrise. The temperature difference between 6.4 m and 0.1 m was 2.5 °C at this time, and the rate of temperature with height was 0.4 °C/m. The magnitudes of gradients in the lowest meter were smaller at night than during the day. The magnitude of the temperature difference between 3.2 m and 6.4 m was about the same near the middle of the day as it was during the few hours before sunrise.

A field program at O'Neill, Nebraska, consisted of seven general observation periods from 1 August to 8 September 1953. These periods were chosen when the forecast was for clear nights and clear or partly cloudy days. Observations were terminated when conditions deviated too much from the ideal. The area within several kilometers of the site was flat, and the grass was short. Lettau and Davidson [8] gave the difference between temperatures at 0.82 m and 0.39 m at two-hourly intervals for each of seven observation periods. During each period, lapse rates were largest at 1235 hours when they were in the range 1.2 - 1.9 °C/m. The time of the strongest inversion varied from one night to the next and might occur from 2035 hours to 0435 hours. The largest nocturnal temperature difference between 0.82 m and 0.39 m was 0.476 °C. This corresponds to a vertical rate of change of 1.1 °C/m, and it occurred at 2035 hours.

Lettau and Davidson's [8] information for other levels illustrated the problem of variations among various instruments maintained by different groups. This was especially true if different types of means were taken or if the instrument lags were considerably different. For example, aspirated thermocouples from the Massachusetts Institute of Technology (MIT) and those from the University of California at Los Angeles (UCLA) gave temeratures at 8.0 m which typically differed by 0.3 - 1.0 °C. Furthermore, the magnitudes of vertical changes measured by the MIT and UCLA systems were not close, and sometimes even the signs were different. An example was 22 August 1953 at 0435 hours when the temperature between 4.0 m and 8.0 m increased 0.21 °C according to MIT thermocouples and decreased 0.45 °C according to UCLA thermocouples.

Very large vertical temperature gradients were found by Schlegel and Butch [9] in the Barrens region of central Pennsylvania. This is a low-lying area which has soil with low heat conductivity. Cold air drainage often combined with radiational cooling to produce minimum nocturnal surface temperatures which were much lower than those in the surrounding area. Minimum temperatures could be as much as 17 °C lower than those at State College about 8 km away. Schlegel and Butch [9] measured the largest vertical gradients in the Barrens at about an hour after sunset. Temperature differences of 4.4 °C between 0.2 m and 1.5 m were common. These large differences were most likely to develop on nights when frost occurred.

Minimum temperatures at night do not always occur at the surface. Geiger [3] and Deacon [6] discussed earlier measurements which showed that temperature profiles over level ground could have minima of a few centimeters or more above the ground on clear, calm nights. Minima 1-3 °C lower than surface temperatures were not unusual. Nocturnal profiles with minimum temperatures above the surface have been observed in India, Europe, and North and South America. Because the validity of some of the older measurements had been questioned, Oke [10] conducted a very thorough study in and near Hamilton, Ontario, (43° 17'N, 79° 56'W). Thermocouples were placed at 0.00, 0.01, 0.025, 0.05, 0.075, 0.10, 0.15, 0.20, 0.25, 0.50, and 1.00 m over grass, bare soil, and snow. Heights of elevated minima were from 0.01 to 0.50 m, and some could have been higher because there were no measurements between 0.50 m and 1.00 m. The probability of raised minima decreased rapidly as wind speed at 0.25 m approached and passed 1.00 m/sec.

Maximum temperatures are not always at the surface in the middle of the day. This was shown very clearly by Halberstam and Schieldge [11] who measured temperature and humidity over a field covered with melting snow where a highly stable layer formed in the lowest half-meter. The field was fairly flat for 4 km in all directions from the observation site at approximately 2100 m above mean sea level just south of Lee Vining, California. Measurements were taken on 15 - 19 March 1978 when there was a warming trend over most of the western third of the United States. During the middle of the day on 18 March the maximum near 0.5 m was very strong. At 1153 PST the temperature was 13.8 °C at 0.5 m, 11.5 °C at 2.0 m, and 10.8 °C at 0.25 m. At 1329 PST the temperature was 13.8 °C at 0.5 m, 11.6 °C at 2.0 m, and 11.6°C at 0.25 m. They did not take measurements right at the surface, but presumably this temperature was 0 °C.

Tait [12] developed an equation to describe vertical temperature differences near the surface during the day. Let t_1 and t_2 be temperatures in degrees Fahrenheit at height Z_1 and Z_2, respectively. Then $t_1 - t_2 = N (1.2 - 6.8 \sin \alpha) \log_{10}(Z_1/Z_2)$, where α is the elevation angle of the sun and N depends upon cloudiness. The value of N varies from 0.1 to 1.0 where the highest numbers are for the clearest skies. The above relationship is based upon observations during daylight hours in a three-month period in summer and early fall at Suffield, Alberta, in Canada. It is most reliable between 1.0 m and 10 m.

Fichtl and Nelson [13] measured temperatures on a 150-m tower at Kennedy Space Center in Florida. Measurements for 19 profiles were made in 1968 during the months January through June except for one set of measurements in October. All measurements were made under unstable conditions sometime between 0905 hours and 1537 hours EST. The average temperature at 18 m was 0.94 °C less than the average temperature at 3 m. The average lapse rate in the layer 3 - 18 m was 0.063 °C/m, and the range was 0.012 - 0.097 °C/m. Mean lapse rates were smaller for deeper layers. The mean between 3 m and 30 m was 0.049 °C/m, and the mean between 3 m and 60 m was 0.037 °C/m. The 19 profiles produced an average lapse rate of 0.023 °C/m for the entire layer between 3 m and 120 m. The range for this layer was 0.019 - 0.027 °C/m.

Godske et al. [5] have discussed data from Potsdam, Germany. Temperature differences between 34 m and 2 m were measured near the beginning of the twentieth century. Inversions at Potsdam began an hour or two before sunset and lasted until a couple of hours after sunrise throughout the year. Inversions were stronger in summer than in winter. Strongest inversions occurred in the second half of August at about 2000 hours when the gradient between 2 m and 34 m was 0.044 °C/m. Vertical gradients greater than 0.03 °C/m occurred at some time between sunset and midnight from June through September. Mean temperatures between 2 m and 34 m decreased with height between two hours after sunrise and two hours before sunset throughout the year at Potsdam at the turn of the century. Largest decreases were at noon during summer. Lapse rates at noon reached 0.040 °C/m in June and July and were at least 0.030 °C/m from April through September.

More recent measurements in a rural area in the United States also showed strong seasonal variations. Takle [14] and Takle et al. [15] measured differences of temperature between 2 m and 32 m on a tower near Ames, Iowa. The period July 1964 - June 1970 provided a set of data relatively free of equipment failure. Inversions were most likely to form at sunset during the cooler half of the year. Inversions usually began to develop before sunset in the warmer half of the year. This was most pronounced in August when inversions were most likely to begin almost two hours before sunset. Inversions also persisted longer after sunrise in summer. In January the atmosphere between 2 m and 32 m developed a superadiabatic (unstable) lapse rate 60 minutes after sunrise. In July an average of 116 minutes after sunrise was required for the atmosphere to become unstable.

Measurements were made during the three-year period July 1945 - June 1948 in Sussex, England, at Rye where little urbanization existed [3] [16]. In the mean data, inversions occurred throughout the night and temperatures decreased with altitude in the middle of the day in the layer between 15.2 m and 47.2 m and in the layer between 47.2 m and 106.7 m. Inversions existed at all hours in January means for the layer between 1.1 m and 15.2 m, but other months followed the expected pattern where temperature decreased with height near the middle of the day. In late spring and in summer, mean lapse rates in the middle of the day were near 0.06 °C/m in this lowest layer and near 0.02 °C/m in the layer between 15.2 m and 47.2 m. Lapse rates at noon were approximately 0.009 °C/m from March through June in the layer from 47.2 m to 106.7 m. In July lapse rates in this higher layer were very small in the middle of the day, and this was attributed to a sea breeze effect. Above the lowest couple of meters, magnitudes of vertical gradients in inversions at night are often much stronger than magnitudes of lapse rates in the middle of the day, and this is reflected in the mean values at many locations. For example, the largest mean two-hourly lapse rate at Rye between 1.1 m and 15.2 m was 0.063 °C/m at noon in August, but during some hours of the night in spring and fall the vertical gradient was greater than 0.080 °C/m.

Geiger [3] showed the annual frequency distribution of lapse rate in each layer at Rye when all hours were combined. In the layer between 1.1 m and 15.2 m, 10 percent of the profiles had rates of temperature increase with height greater than 0.108 °C/m, and 10 percent had rates of decrease larger than 0.050 °C/m. The median in this low layer was a slight inversion, and 45 percent had negative vertical temperature changes with height. Temperature decreased with height 60 percent of the time between 15.2 m and 47.2 m, and 27 percent of vertical rates of change were between 0.00 and -0.01 °C/m. Temperature decreased with height 68 percent of the time between 47.2 m and 106.7 m, and 60 percent of vertical rates of change were between 0.00 and -0.01 °C/m. The largest positive vertical change was 0.136 °C/m in this layer, and the most negative value was -0.042 °C/m. The corresponding values for the layer between 1.1 m and 15.2 m were 0.534 °C/m and -0.205 °C/m. Inversions can become very strong because an increase of temperature with height is a stable stratification which inhibits extensive vertical mixing of air parcels. Comparable decreases can occur only very near the surface with intense solar heating.

DeMarais [16] compared lapse rates from Rye with those from the commercial and business district of Louisville, Kentucky, for the period 23 August 1957 - 15 July 1958. Measurements were available for the layer 51.8 - 18.3 m at Louisville and 47.2 - 15.2 m at Rye. Mean rates of decrease of temperature with height were comparable at the two stations during the day, but profiles were very different at night. Inversions occurred throughout the night during the entire year in the layer 47.2 - 15.2 m at Rye. Strongest mean monthly inversions at Rye occurred in October from 0400 hours to 0600 hours when the temperature profile showed a vertical rate of change of 0.04 °C/m. Mean temperatures at Louisville showed only two inversions in a two-hourly semimonthly plot for the layer 51.8 - 18.3 m. Weak inversions (<0.004 °C/m) occurred at Louisville at 2200 hours in the first part of October and at 0200 hours in the second half of April. The layer is isothermal at a few nocturnal hours between October and April. Mean nocturnal lapse rates are rather large in summer. During the first half of July the mean temperature at Louisville in the layer 51.8 - 18.3 m decreased faster than 0.010 °C/m at all hours. Mean lapse rates near the middle of the day in summer were in the range 0.018 - 0.028 °C/m.

Neighboring urban and rural vertical temperature profiles were compared by Duckworth and Sandberg [17]. Simultaneous soundings in urbanized locations and at more open sites in the San Francisco Bay area were taken during the period July 1951 - June 1953. Soundings from the time period 2000 - 0100 PST showed that 30 of 32 in the less urbanized areas had inversions. Simultaneous profiles in built-up areas had inversions below roof level in only seven cases, and isothermal conditions existed in seven more cases. Temperature decreased at least up to the rooftops in the 18 additional profiles. In most cases the urban and rural soundings crossed each other somewhere below 300 ft (91.44 m), and the urbanized area was actually cooler at the higher levels.

More recently Bowling [18] documented large urban effects on lapse rates in Alaska. Three pairs of soundings from tethered balloons in and near Fairbanks, Alaska, showed large differences near the surface. Soundings for 22 and 23 December 1981 had inversions in both urban and rural areas, but the rural inversions were much stronger. The rural atmosphere was about 5 °C colder at the surface on 22 December but was warmer above 45 m where the soundings crossed. Rural temperatures were 7 °C lower at the surface on 23 December. The two soundings were close above 60 m and actually crossed at about 105 m. On 15 December the rural sounding below 35 m had an inversion of 0.2 °C/m while the urban atmosphere was nearly isothermal in this lower layer. The temperature increased by about 5 °C between 35 m and 110 m on both soundings which were nearly identical in this upper layer on 15 December 1981.

Another recent investigation [19] showed that vertical temperature gradients at two different rural sites in the same general geographical region could differ substantially. Both sites were in the eastern piedmont of the Carolinas, and both were in clearings surrounded by forest. One site was in a valley by a lake, and the other was 175 km north on a low hilltop. Sensors to measure temperature difference were mounted at 11.0 m and 59.9 m at both sites. Routine on-site maintenance and calibration were strictly maintained from 1976 through 1982. There was a correlation of 0.71 in the daily site-to-site vertical temperature difference between 11.0 m and 59.9 m at dawn.

Considerable differences between the sites existed in spite of this high correlation. Predawn inversions with rates of change of temperature greater than 0.05 °C/m occurred 29 percent of the time on the hilltop and 10 percent of the time near the lake. Changes greater than 0.12 °C/m existed before dawn seven percent of the time on the hilltop and less than one percent of the time near the lake.

Observed seasonal variations at both sites in the Carolinas were unlike those in several earlier studies at other locations. For example, Takle et al. [15] found that the strongest inversions in a rural area near Ames, Iowa, occurred in summer. Inversions greater than 0.12 °C/m between 11 m and 60 m were absent in July and August at the two locations in the Carolinas. These strong inversions were most likely to occur in the months September through December on in March and April. Moderate inversions from 0.015 to 0.030 °C/m were most common in summer.

Hansen [20] has presented 147 twenty-minute profiles of wind and potential temperature on a 62-meter research tower at White Sands Missile Range (WSMR), New Mexico, which is close to El Paso, Texas. The climate is semiarid, and the surface is rough and nonhomogeneous. Profiles were from the months January, February, March, May, June, August, and November 1962. Hansen considered the 147 profiles to be of research quality and to be representative. The procedure for relating the vertical gradient of potential temperature to the vertical gradient of temperature is given in the appendix. Many persons use potential temperature instead of temperature in theoretical studies of the atmospheric boundary layer, and so the conversion equation in the appendix should be quite useful for readers who wish to do further searches on their own.

Hansen included 74 profiles for winter in his report. Temperatures in the early afternoon typically decreased rapidly with lapse rates from 0.04 °C/m to 0.06 °C/m in the layer between 1.15 m and 61.10 m. Nocturnal increases of temperature with height in this layer in winter sometimes reached 0.06 °C/m, but inversions were quite weak on other nights. A larger part of the vertical temperature change was in the lower portion of the layer. Lapse rates during the early afternoon in winter were usually between 0.07 °C/m and 0.11 °C/m in the layer between 1.15 m and 30.65 m. At night, vertical changes in this layer could be very small in winter or might reach 0.11 °C/m.

Only 36 profiles for spring were included in Hansen's report. The lapse rates during daylight were about the same in spring as in winter. The strongest inversion in this limited amount of data for spring was only about one-third as strong as the strongest inversion in winter.

Eighteen profiles were given for summer at WSMR. Lapse rates during daylight hours in summer were comparable to those in winter and spring. The inversion on one summer night was much stronger than any in the other seasons. At 2220 hours, 5 June 1962, the temperature increased at a rate of 0.12 °C/m as an average for the entire layer between 1.15 m and 61.10 m and at a rate of 0.24 °C/m between 1.15 m and 30.65 m. The vertical changes at 2145 hours and 2200 hours were almost as large.

All of the 19 profiles for fall in Hansen's report were for daylight hours on 1 November 1962. The largest lapse rate in the layer between 1.15 m and 61.10 m was 0.05 °C/m which occurred at 1204 hours.

In a later study Hansen [21] examined generalized growth and predictions of nocturnal inversions based on 217 micrometeorological profiles and 305 soundings from different locations. Inversions typically existed very near the surface from about one hour before sunset to one hour after sunrise. Different types of averaging were used, and all types produced graphs where inversions reached 100 m in less than one hour after sunset. An additional one to two hours was required for the top of the inversion to reach 200 m. The maximum height of about 500 m was reached after an additional ten or more hours.

Gradients over the Egyptian desert regions can also be very strong. Geiger [3] showed mean profiles above desert soil in Egypt during July and January. Cloudless skies exist almost all the time in July in Egypt. In the middle of the day in July temperature decreased at a rate of 0.12 °C/m between 1.1 m and 16.2 m and at a rate of 0.018 °C/m between 16.2 m and 61.0 m. Strongest inversions in July occurred near sunrise when the rate of increase was 0.076 °C/m between 1.1 m and 16.2 m and 0.022 °C/m between 16.2 m and 61.0 m. Inversions were stronger in January than in July. In the lower layer in January near sunrise the rate of temperature increase with height was 0.099 °C/m, and it was 0.027 °C/m between 16.2 m and 61.0 m. Lapse rates were smaller in January than in July in the middle of the day. In January the mean temperature in the Egyptian desert in the middle of the day decreased at a rate of 0.073 °C/m between 1.1 m and 16.2 m and at a rate of 0.016 °C/m between 16.2 m and 61.0 m.

Shirvaikar et al. [22] computed mean temperature profiles for different Pasquill stability categories at Tarapur. Pasquill categories A through G represent conditions from very unstable to very stable. The site is 1.5 km from the Arabian Sea and 100 km from Bombay. Data are for the layer 6.4 - 120 m during January, April, July, and October. The largest mean rate of increase was 0.037 °C/m in January for Pasquill category G. The largest mean decrease with height occurred with category C in July when the mean lapse rate was 0.016 °C/m. In October and January the mean temperature increased with height at a rate of 0.004 °C/m with Pasquill's neutral category D. In April and July mean temperatures decreased with height with category D, and the lapse rates were 0.007 °C/m and 0.010 °C/m, respectively. Shirvaikar et al. examined mean lapse rates in the lower layer 6.4 - 30 m associated with the Pasquill category D for January. Temperatures decreased with height between 1000 hours and 1800 hours, and inversions existed during the remaining hours.

Kloppel et al. [23] [24] studied nocturnal temperature inversions on a 300-m tower at the southeastern edge of Hamburg, Germany. At the end of clear and calm nights, it was common for the temperature between the instrument shelter and 100 m to increase at an average rate of 0.05 °C/m. The latter study focused on inversions which reached at least 250 m and followed a special process of dissolution. Thirty percent of the selected cases had a rate of change of at least 0.10 °C/m in the layer between 50 m and 110 m according to Kloppel's [24] Figure 7. In fact, ten percent had mean values of at least 0.10 °C/m through the entire layer between 50 m and 250 m.

Kottmeier [25] examined special data from 11 very stable nights at Gartow (53°4'N, 11°26'E) and Sprakensehl (52°48'N, 10°32'E) in northern Germany. Under these conditions the wind speed in the layer above the planetary boundary layer acted as an external boundary condition. The geostrophic wind could be used as a first approximation to this wind, but fluctuating ageostrophic components sometimes produced mesoscale effects. Both the height of the boundary layer and the depth of the inversion increased as the wind speed increased. Wittich and Roth [26] made a detailed examination of one clear and stable night in October at Gartow. In the early evening the temperature increased 4 °C in the lowest 90 m and was nearly isothermal in the next 90 m. The depth of the inversion increased, and the shape of the profile changed as the night went on. By about 0300 hours the profile had developed so that the layer below 100 m had a mean inversion of 0.03 °C/m, and the temperature increased from 100 m to 130 m at a rate of 0.08 °C/m. The geostrophic wind speed at the top of the boundary layer on this night was 10.9 m/sec.

This can be compared with a recent theoretical study of the influence of geostrophic wind on nocturnal atmospheric cooling by Estournel and Guedalia [27]. They discussed two simulations where all parameters in their one-dimensional model were the same except geostrophic wind speed. The height of the top of the inversion increased steadily throughout the night when the wind speed was 3 m/sec, and steepest temperature gradients remained near the surface. This case with low wind speed strongly resembled the results obtained by Knighting [28] who considered only radiational effects on nocturnal temperature profiles. When Estournel and Guedalia used a geostrophic wind speed of 10 m/sec, the thickness of the inversion grew faster in the early evening, and growth leveled off at the end of the night. The profile changed shape during the night so that steepest temperature gradients were close to the top of the inversion near sunrise.

These theoretically predicted nocturnal changes of temperature profiles agree well with observations which were made by Van Ulden and Holtslag [29]. They measured temperatures and winds on a 200-m tower at Cabauw in a flat area in the center of the Netherlands. Measurements were made on two clear nights within a month of the summer solstice. On one night the wind speed at 200 m was only 1 m/sec, and the strength of the inversion near the surface increased through the night until near sunrise. The magnitude of the vertical temperature gradient decreased monotonically with height throughout the night. On the other night when the wind was 10 m/sec at 200 m, the strength of the inversion in the lowest 20 m did not increase after 2200 hours. By 0330 hours, the magnitude of the temperature gradient between 80 m and 120 m was three times as strong as the magnitude of the gradient in the lowest 40 m, and the gradient was very small between 40 m and 80 m.

Lazar's [30] study in the eastern Alps revealed multilayered inversions which frequently contained three main sections. Thicknesses of inversions could be 800 m or more, and the intensity sometimes reached 20 °C when snow was on the ground. Lazar presented several representative profiles, and the typical increase of temperature in the lowest 100 m above ground was 3 °C to 6 °C. Free-air measurements were compared with corresponding elevations at surface stations. Stations on mountain ridges showed small differences of temperature, but differences could be several degrees in basins.

Borghi and Giuliacci [31] compared temperatures from Monte Bisbino at an elevation of 1322 m and Grigna Settentrionale at 2406 m with temperatures at these altitudes above Milano-Linate in northern Italy. Both of the higher stations are between 50 km and 60 km from Milano-Linate, which has a surface elevation of only 103 m. Observations were made on 49 clear days in June, July, and August 1972. The mean temperature at Monte Bisbino was 1.4 °C less than at Milano-Linate at 0000 GMT and 3.8 °C more at 1200 GMT. The mean temperature at Grigna Settentrionale was 1.0 °C less than at Milano-Linate at 0000 GMT and 2.5 °C more at 1200 GMT.

In the United States Mahrt [32] examined fast-response aircraft soundings of a very stable boundary layer in central Oklahoma at 0530-0640 LST on 5 May 1979. Analysis of the turbulence structure showed that the principal source of turbulence was shear generation near the top of the surface inversion layer. Downward flux of heat in the turbulent shear zone warmed the upper part of the inversion layer. Much of the turbulence exhibited statistical properties consistent with shear instability on horizontal scales of 300 - 400 m. Some turbulence also existed on smaller scales which could have been even less than a horizontal scale of 100 m. Earlier studies by Andre and Mahrt [33] and Garratt and Brost [34] showed that effects of clear air radiative cooling predominated over turbulent heat flux when shear was only modest near the top of inversions.

Significant horizontal inhomogeneity often exists in locations with uneven terrain. Radiational cooling on clear nights causes drainage winds and pooling of cold air in valleys. Lenschow et al. [35] showed that abrupt transitions from nocturnal to daytime conditions in summer were frequent at sites in valleys and were uncommon on plateaus. Data were from the National Hail Research Experiment mesonet in northeastern Colorado and from Haswell in southeastern Colorado. Temperature changes of 12 °C within 30 minutes were observed even in valleys as shallow as 50 m and with slopes of only 0.007. Analysis indicated that a well-mixed boundary layer was advected into the valley from upstream slopes or plateaus.

Whiteman's [36] observations from very deep valleys in western Colorado showed somewhat less rapid temperature changes in most cases. Direct comparison with measurements by others is difficult because of very different data collection and analysis procedures.

Thompson [37] studied data from the summer and fall of 1957 in Red Butte Canyon in Wasatch Range of northern Utah. Rapid cooling near sunset was common. On the night 24-25 September 1957 conditions were considered typical. At each position of a 32-m tower the temperature dropped 10 °C in 40 minutes, and then the rate became quite slow. The inversion was quite strong near the surface. In the hour beginning at 1705 MST, the vertical temperature increase was at an average rate of 0.35 °C/m between 2 m and 16 m. After sunrise the next morning the temperature rose at only about 40 percent of the rate of fall the previous evening.

Nocturnal temperatures in another mountainous area in the United States were studied more recently by Orgill and Schreck [38]. Observations were made along a slope in the Geysers Geothermal Resource area of California on several nights in July 1979 and September 1980. Skies during experimental periods in these months were clear 88 percent of the time. Nocturnal cooling in July 1979 caused temperatures to be less than 20 °C in pockets of air over topographic depressions while air was 25 - 27 °C over steeply sloping areas. The inversion structure usually consisted of a surface stable layer 50 - 200 m thick with a quasi-isothermal layer above and an upper inversion layer beginning at 400 m above the slope. Orgill and Schreck [38] have given a table of potential temperature gradients at a station half-way up the slope in the lowest 100 m above ground for the experimental periods of September 1980. If these are converted to temperature gradients according to the procedure in Appendix A, it is seen that gradients are quite steep. For example, the mean temperature between the surface and 100 m increased by at least 0.08 °C/m from 1900 hours to 0600 hours on the night of 24 - 25 September 1980. The nocturnal vertical temperature gradient during the experimental periods in September 1980 had a maximum magnitude of 0.11 °C/m in the lowest 100 m, and most were greater than 0.05 °C/m.

Special measurements on a 300 m tower were made at the Boulder Atmospheric Observatory on 87 nights during the period 25 March - 22 June 1981. The site is on the high plains of Colorado approximately 30 km north of Denver and 30 km east of the foothills of the Rocky Mountains. Multiple inversions developed on 35 of these nights according to Xing-Sheng et al. [39]. They discussed the profile for the 10-minute period 0250-0300 MST, 24 April 1981, in detail. The temperature increased by 6 °C in a strong surface inversion between 2 m and 22 m. This corresponds to a rate of 0.3 °C/m in this low layer. The temperature increased only about 0.2 °C between 22 m and 50 m, and another 1.5 °C increase took place between 50 m and 100 m. Four additional abrupt changes in vertical temperature gradient occurred between 100 m and 300 m. Nai-Ping et al. [40] examined data from the night 26 - 27 March 1981 during a cold frontal passage at the Boulder Atmospheric Observatory. The vertical temperature difference between 10 m and 300 m stayed near 6 °C throughout the night from 2100 hours to 0530 hours. The difference between 10 m and 100 m decreased from about 4 °C to about 1 °C while the difference between 100 m and 300 m increased from slightly more than 2 °C to about 5 °C. The largest portion of the change occurred after midnight when a low-level jet was established between 100 m and 150 m. Counter-gradient heat fluxes existed in the vicinity of the low-level jet and near the top of the inversion.

Shapiro [41] documented an even more interesting frontal passage at the Boulder Atmospheric Observatory during the afternoon of 24 March 1982. Measurements from the meteorological research tower indicated that the frontal zone was much more intense than it appeared on a conventional map or even on a special mesoscale analysis. Large horizontal gradients of temperature and wind were concentrated within a distance of less than 200 m when the front passed the tower at 2127 GMT (1427 MST). The temperature dropped 6 °C in 10 s during frontal passage. Shapiro et al. [42] later examined a front which was 900 m wide, and the temperature dropped 8 °C in one minute upon passage at 2000 GMT (1300 MST) 19 September 1983. At levels above 100 m on the tower, the temperature rose 3 °C a few minutes later and then the atmosphere became very unstable and cooled with turbulent undulations of 2 °C amplitude. At such times these temporal changes may be as important as the instantaneous vertical temperature profile. Furthermore, some instantaneous profiles showed extreme gradients because the frontal boundary reached different levels at slightly different times.

Extremely rapid changes of temperature and lapse rate have also been documented along the southern coast of New South Wales in eastern Australia. These rapid changes were associated with an intense type of cold front called a southerly burster. These fronts occurred in the warm part of the year, and orographic effects appeared to play a role in the genesis of all cases [43]. In the period 1974 - 1983, southerly bursters occurred 95 times at the Sydney Airport, and 62 percent occurred between 1300 and 2100 hours local time. All were in the months September through March.

Colquhoun et al. [43] examined thermograph traces at 10 m and lapse rates betweeen 10 m and 110 m for two southerly bursters. At the time of frontal passage on 20 November 1973 the lapse rate increased from 0.01 °C/m to the very unstable value of .06 °C/m in approximately 3 minutes. This was followed by a general downward trend with large fluctuations, and an inversion of about 0.03 °C/m was reached in 45 minutes. This was followed by an increase with small fluctuations until approximately 90 minutes after frontal passage when the lapse rate leveled off at about the value that it had just before frontal passage. On 25 November 1982 the lapse rate decreased from 0.020 °C/m to 0.005 °C/m during the hour before frontal passage. When the front passed, the lapse rate increased to 0.074 °C/m within 10 minutes. The lapse rate decreased to about 0.030 °C/m during the next 10 minutes. The lapse rate decreased slowly and gradually with very small fluctuations after that, and no inversions occurred.

It remains to be seen if such rapid changes of lapse rate at frontal surfaces are common throughout the world. Older studies of frontal zones did not provide such good resolution in both space and time. For example, Brundidge [44] analyzed the mesoscale structure of 11 nocturnal cold frontal passages in the fall and winter of 1961-62 at Cedar Hill, Texas. Observations were from a 1420-ft (433-m) tower and eight 55-ft (17-m)towers. Brundidge's data were reported at 10-minute intervals and represented averages over all but the last 20 seconds of the preceding 10-minute period. Brundidge expected superadiabatic layers near the ground behind the front, but these were found in only one of the 11 cases. It is possible that large lapse rates which existed less than three minutes were lost in the averaging process.

Extreme temperature gradients can also occur in association with strong winds which occasionally occur in the lee of most mountain ranges. These winds are called foehns in the Alps in Europe and chinooks east of the Rocky Mountains in North America. Temperature changes of 11 - 22 °C in 15 minutes are not surprising [1]. Hamann [45] described one unusual case where the change of temperature was 27.2 °C in two minutes on 22 January 1943 at Spearfish in the Black Hills region of western South Dakota. Mathai et al. [46] made temperature measurements when a chinook passed a meteorological research tower at Calgary, Canada. A strong temperature inversion began to develop on the evening of 17 February 1979. By 0100 MST the temperature at 90.4 m was more than 11 °C greater than the temperature at 9.1 m. The temperature difference between the two levels remained larger than 8 °C until 0800 MST.

Chinooks are expected to be more frequent than once per year in parts of Montana, Wyoming, and Colorado [47] and in Alberta, Canada [48]. These winds seldom occur in summer (June, July, and August).

A similar type of wind is called a Santa Ana in southern California [1]. Sommers [49] examined closely-spaced data in a canyon at 900 m and on a nearby ridge at 1625 m. In one carefully documented case the lapse rate remained essentially adiabatic (0.01 °C/m) for 48 hours.

Much of the foregoing discussion concerns meteorological conditions which are unusual because investigators have preferred to collect data from exceptionally well-developed cases. This information can be put in perspective by considering some thresholds which have been developed to screen large amounts of data. The Tennessee Valley Authority (TVA) collects about 2.5 million observations per year from sensors on several towers in the Tennessee Valley. TVA uses computerized data processing techniques to handle this large mass of data. Flagging criteria identify extreme conditions which must be evaluated carefully by a qualified meteorologist [50].

TVA measures vertical temperature differences in the following three layers: (1) 110 m - 10 m, (2) 110 m - 61 m, and (3) 61 m - 10 m. Lapse rates exceeding the autoconvective rate (decrease with height greater than 0.034 °C/m) are flagged. Flagging also occurs if the decrease is greater than 0.01 °C/m between 1800 hours and 0500 hours. Vertical gradients are flagged if the temperature increases with height at a rate greater than 0.01 °C/m between 1000 hours and 1700 hours. Inversions greater than 0.15 °C/m are flagged at any hour.

The TVA screening program also flags vertical gradients when unusual changes occur within an hour. All three hours are flagged whenever there is a double switch of sign in consecutive hours (plus to minus to plus, or minus to plus to minus). If the absolute value of the change between two consecutive hours is greater than 0.015 °C/m between 0900 hours and 1700 hours local standard time, inclusive, both hourly values are flagged. Between 1700 hours and 0900 hours, flagging occurs when the absolute value of the change between two consecutive hours is greater than 0.04 °C/m.

Brief comments concerning temperatures over the ocean will be given here, and the interested reader may consult Godske et al. [5, pp. 107-108 and 488-490] for more information. Measurements from the Meteor expedition in the tropical Atlantic showed that the sea surface temperature had a very small diurnal oscillation with an amplitude less than 0.2 °C. The amplitude of the oscillation of air temperature was only about 0.5 °C. No inversions existed at night. The mean lapse rate throughout the lowest 200 m was superadiabatic (greater than 0.01 °C/m) throughout the 24 hours. Large decreases of temperature with height in the lowest 200 m existed even beneath trade-wind inversions. Air at 6 m was approximately 0.5 °C colder than the sea surface temperature at night and within $\pm$0.3 °C of the sea surface temperature in the middle part of the day.

Recent observations by Sethu Raman et al. [51] were made off the eastern coast of the United States during the period 9-13 November 1983. The boat moved around off the North Carolina coast within the longitudes 75 - 78°W and the latitudes 33 - 36°N. When flow was southerly with a long path over water, the temperature of air at 10 m was within ± 3.0 °C of the sea surface temperature. When northwesterly flow was associated with a cold air outbreak, the temperature difference between the sea surface and 10 m reached 11 °C in the middle of the afternoon of 12 November 1983. The difference remained greater than 9 °C for several hours. This shows that near coasts extremely unstable conditions can exist very near the water surface.

III. HUMIDITY PROFILES

Humidity profiles are affected by the factors which influence temperature profiles and by the availability of moisture at the surface. It follows that diurnal changes of humidity are large in the first few meters above the ground. Minimum water vapor in the early morning coincides with minimum temperature near the surface. The decrease of stability after sunrise causes evaporation and upward transfer of water vapor from the surface. Initially the water vapor accumulates in a shallow layer near the ground, and a maximum exists between 0800 and 1200 hours. Continued surface heating causes the lapse rate to become so steep that vertical exchange extends through a much deeper layer while the ground becomes drier. A second minimum dew point occurs in the afternoon at about 1400-1500 hours. Convection diminishes as surface heating decreases, and a second maximum dew point occurs within three to four hours after sunset. During the night there is a flux of water vapor to the ground, and the dew point decreases until sunrise. A pattern similar to this idealized double oscillation is most likely to be followed on clear summer days. This has shown up clearly in data from clear July days at Quickborn [3] in Holstein, West Germany.

Mean two-hourly vertical gradients of absolute humidity were recorded for the three-year period July 1945 - June 1948 at Rye in southern England about 5 km from the coast [3]. Mean absolute humidities in the layer 1.1 - 15.2 m decreased with height from 1000 hours to 1400 hours in every month. Mean absolute humidity inversions occurred in this layer during the nocturnal period 2200-0400 hours in every month except June. The two-hourly observations of vertical change between 1.1 m an 15.2 m showed double maxima only in May, June, and August. The most rapid mean decrease with height was 0.061 $(g/m^3)/m$, which may be written as 0.061 g/m^4, and it occurred at 0800 hours in June. The strongest mean inversion was 0.029 g/m^4 at 0400 hours in July. Magnitudes of vertical gradients of absolute humidity were generally largest in summer and smallest in winter. No rate of decrease was greater than 0.010 g/m^4 from December through March. Sometime during the day the rate of decrease with height exceeded 0.040 g/m^4 from May through August. Diurnal variations of vertical gradients were smallest in winter and largest in summer.

Geiger [3] included a similar table for the layer 47.2 - 106.7 m at Rye. Magnitudes of vertical gradients were much smaller in this higher layer than in the lower layer, and inversions were particularly weak. No mean inversions occurred in January or in July through November in the higher layer. The strongest inversion had a rate of increase with height of only 0.0015 g/m^4 and occurred at 1800 hours in May. The strongest inversions in April and June were slightly weaker, and those in February, March, and December were much weaker. The largest rate of decrease with height in the layer 47.2 - 106.7 m was 0.0072 g/m^4 at 1200 hours and 1400 hours in October. Vertical gradients larger than 0.0030 g/m^4 occur only in August, September, and October at Rye.

Annual frequency distributions for all hours showed that humidity inversions occurred about 40 percent of the time in each layer, and absolute humidity decreased with height 60 percent of the time. Maximum inversion rates over the three years were 0.22 g/m^4 for 1.1 - 15.2 m, 0.11 g/m^4 for 15.2 - 47.2 m, and 0.06 g/m^4 for 47.2 - 106.7 m. Maximum rates of decrease were 0.41 g/m^4 for 1.1 - 15.2 m, 0.13 g/m^4 for 15.2 - 47.2 m, and 0.08 g/m^4 for 47.2 - 106.7 m. In the layer 1.1 - 15.2 m, 10 percent of the vertical rates of change with height were less than -0.05 g/m^4, and 10 percent were greater than 0.03 g/m^4. In the layer 15.2 - 47.2 m, 80 percent were between -0.017 g/m^4 and 0.011 g/m^4. The corresponding values for 47.2 - 106.7 m were -0.008 g/m^4 and 0.005 g/m^4.

Measurements very near the surface during clear winter weather in 1933-1937 at Poona, India, showed another interesting result [3]. The mean vapor pressure at sunrise was 7.5 mm-Hg at 0.008 m and at 0.15 m. However, the means of measurements at sunrise were 7.4 mm-Hg at both 0.025 m and 0.075 m, and the minimum was not at the surface. Vapor pressure at sunrise increased monotonically from 7.5 mm-Hg at 0.15 m to 8.8 mm-Hg at 3.05 m. Near noon the maximum was 10.0 mm-Hg at 0.008 m and the decrease was monotonic to 8.3 mm-Hg at 3.05 m.

More recently, Klug and Webs [52] examined vertical distributions of specific humidity in the Federal Republic of Germany. Information was obtained from 1241 soundings taken irregularly during a four-year period in various regions of the country. In 72 percent of the cases the change between the surface and 100 m was within ± 0.5 g/kg. The unit g/kg indicates the number of grams of water vapor per kilogram of moist air. (The density of air near the surface is about 1.226 kg/m^3.) Only 8 percent showed an inversion stronger than 0.5 g/kg, while 20 percent of the time the decrease between the surface and 100 m was at least 0.5 g/kg.

Klug and Webs also examined profiles as a function of stability. Under very stable conditions, 27 percent of the profiles had inversions where specific humidity increased at least 0.5 g/kg between the surface and 100 m, and 35 percent decreased at least this much. Under very unstable conditions not one humidity inversion between the surface and 100 m was as strong as 0.5 g/kg, but the rate of decrease with height was greater than (0.5 g/kg)/100 m in 45 percent of the very unstable conditions.

Frequency of different types of humidity profile also had a diurnal variation in the German data. Between the surface and 100 m, humidities increased by at least 0.5 g/kg in 16 percent of the soundings at night, 10 percent in the morning, 5 percent during the day, and 6 percent in the evening. The rate of decrease with height in the lowest 100 m was more than (0.5 g/kg)/100 m for 42 percent of the soundings in the evening and for 14 - 16 percent during the other parts of the day.

Inversions in temperature and humidity do not necessarily occur simultaneously even though both are more probable at night than during the day. A good example was given by Xing-Sheng et al. [39] in their profile for the 10-minute period 0250-0300 MST, 24 April 1981, at Boulder Atmospheric Observatory. The temperature at 22 m was 6 °C greater than the temperature at 2 m. The specific humidity at this time did not show an inversion at all, and in fact, it decreased 0.3 g/kg between 2 m and 22 m. The temperature increased approximately 1.7 °C between 22 m and 100 m while the specific humidity decreased by approximately 0.1 g/kg. Of course, this example only showed what could happen in a mountainous region.

Sasano [53] examined a limited amount of simultaneous temperature and humidity data from flat terrain on the Kanto Plain about 60 km northeast of Tokyo. Sasano's Figure 9 contained mean vertical profiles of specific humidity and potential temperature for the period 0801-0812 Japanese Standard Time on a calm and sunny November morning. (The appendix of this report shows how to convert a change of potential temperature to a change of temperature). The temperature profile had a very weak inversion in the lowest 100 m where the average rate of increase of temperature with height was only 0.002 °C/m. The specific humidity was 4.8 g/kg at the surface instrument shelter and decreased to 3.8 g/kg at 100 m. Thus, the specific humidity decreased at a rate of (1.0 g/kg)/100 m in a nearly isothermal atmospheric layer. It can be concluded that humidity and temperature profiles can have considerably different shapes in both mountainous and level areas.

IV. SUMMARY AND CONCLUSIONS

The vertical gradients of temperature and humidity in the lowest 100 m of the atmosphere are extremely variable. The shape of the temperature profile depends upon atmospheric stability and upon radiational cooling or heating at the surface. Humidity profiles are influenced not only by these factors but also by the availability of moisture at the surface. Diurnal changes of both elements in the first few meters above the ground are particularly large on clear days because radiational changes are greater when clouds are absent. Tables 1 and 2 summarize the data.

Gradients in the lowest half-meter of the atmosphere are often very strong in clear weather. The temperature of a desert surface at low latitudes when the sun is high and winds are light is often more than 10 °C greater than the air temperature at 0.50 m. The difference between the surface temperature and air temperature at 0.50 m in middle latitudes may also be more than 10 °C on clear summer days. Such large differences are especially likely to occur if the surface is something like tar macadam. The temperature difference between 0.05 m and 0.50 m above a grass-covered area often reaches more than 5 °C on clear summer days and 3 °C on cloudy summer days. It is unusual for nocturnal inversions to have vertical temperature changes greater than 2 °C between the surface and 0.50 m.

Minimum temperatures at night and maxima during the day do not always occur at the surface. Nocturnal minima over level ground on clear and calm nights can exist at heights from a few centimeters to at least half a meter. These minima may be 1 - 3 °C lower than the surface temperature. Maxima above the surface in the middle of the day are even more rare. These are most likely to exist above melting snow. The temperature at 0.50 m may be as much a 3 °C greater than the temperature at 0.25 m and 2 °C greater than the temperature at 2.0 m.

Temperatures usually change at a slower rate between 0.5 m and 1.5 m. In the middle of summer days, lapse rates are often in the range 1.0 - 2.0 °C/m on clear days and 0.2 - 0.6 °C/m on cloudy days. Average lapse rates near 1 m in the middle of the day in December are often less than one-tenth as large as those in July. On clear nights the rate of increase of temperature with height near 1 m is usually within a few tenths of a degree of 1.0 °C/m. Averages for clear nights are usually somewhat larger in June than in December, but seasonal variations are smaller than they are below 0.5 m. Average inversions near 1 m are stronger in December than in June on cloudy nights, but both usually show vertical rates of change less than 0.10 °C/m.

Average vertical rates of change of temperature between 1 m and 15 m are smaller than those near the surface. At midday, maximum lapse rates for the layer 1 - 15 m are not greater than 0.12 °C/m at most sites. Nocturnal inversions in the layer 1 - 15 m may have vertical temperature changes as large as 0.20 °C/m in some places. However, values of 0.10 °C/m or less are more common at most sites.

In the daytime, temperature gradients generally get weaker at higher levels. An average lapse rate as large as 0.030 °C/m for the entire layer between 2 m and 100 m is unusual.

The magnitude of the mean vertical temperature gradient through the layer from 2 m to 100 m is frequently much larger at night than during the day. At the end of clear and calm nights it is common for temperatures to increase with height at an average rate of at least 0.05 °C/m. Nocturnal inversions are not necessarily stronger near the surface than at higher levels. Steepest temperature gradients may actually occur at the top of an inversion if the wind speed near the top is greater than 10 m/s. It is possible for the thickness of an inversion to be several hundred meters. It is unusual for the magnitude of the average gradient between 2 m and 100 m to be larger than 0.15 °C/m.

A limited amount of information about humidity profiles in the lowest 100 meters is available. Indications from available data agree that inversions of absolute humidity exist about 40 percent of the time, and the amount of atmospheric water vapor decreases with altitude 60 percent of the time. Profiles with decreases of 1.7 g/m^3 or more between the surface and 100 m occur about 10 percent of the time. Inversions with increases greater than 1.0 g/m^3 in the lowest 100 m occur about 10 percent of the time at Rye in England. Strong inversions are much less likely to occur in the Federal Republic of Germany than at Rye. Inversions are most likely to occur at night. Humdidity and temperature inversions do not necessarily occur simultaneously.

Table 1. Summary of studies of vertical variation of temperature expressed in units of degrees Celsius per meter.

Reference	$\Delta T/\Delta z$ ($^oC/m$)	layer (m)	Comments
Bowling (1986)	0.2	sfc-35	Fairbanks, 15 Dec 81, urban
	0.0	sfc-35	Fairbanks, 15 Dec 81, rural
	0.07	35-110	Fairbanks, 15 Dec 81, urban and rural
Riordan et al. (1986)			Eastern piedmont of Carolinas in hour before sunrise
	> 0.05	11-60	29 % of mornings at hilltop site
	> 0.05	11-60	10 % of mornings at site near lake
SethuRaman et al. (1986)			Over ocean near eastern coast of the United States
	-0.3 to 0.3	0.0-10.0	Most of the time
	-1.1 to -0.9	0.0-10.0	Several hours following cold air outbreak
Colquhoun et al. (1985)			New South Wales in eastern Australia
	-0.06	10-110	Following southerly burster passage 20 Nov 73
	-0.074	10-110	Following southerly burster passage 25 Nov 82
Wittich and Roth (1984)			Gartow in northern Germany, one clear stable night
	0.03	sfc-110	October at 0300 hours
	0.08	110-130	October at 0300 hours
Nai-Ping et al. (1983)			Boulder Atmospheric Observatory in Colorado
	0.04	10-100	2100 MST, 26 March 1981
	0.01	100-300	2100 MST, 26 March 1981
	0.01	10-100	0530 MST, 27 March 1981
	0.025	100-300	0530 MST, 27 March 1981
Xing-Sheng et al. (1983)			Boulder Atmospheric Observatory in Colorado
	0.3	2-22	0250-0300 MST, 24 April 1981
	0.007	22-50	0250-0300 MST, 24 April 1981
	0.03	50-100	0250-0300 MST, 24 April 1981

Table 1 (Continued)

Reference	$\Delta T / \Delta z$ ($^{o}C/m$)	layer (m)	Comments
Halberstam and Schieldge (1981)	12.0	0.25–0.50	Near Lee Vining, CA, over melting snow at midday
	-1.5	0.50–2.0	
Lazar (1981)	0.03 to 0.06	sfc–100	Eastern Alps, typical inversions on clear, calm night
Klöppel (1980) and Klöppel et al. (1978)	0.05	sfc–100	Hamburg. Typical at end of clear, calm night
Mathai et al. (1980)	0.10 to 0.14	9.1–90.4	Inversion during chinook passage at Calgary, Canada
Schlegel and Butch (1980)	3.4	0.2–1.5	Shortly after sunset in Barrens region, PA
Reynolds and Pittman (1978)			Flagging criteria for extreme conditions
	<-0.034	10–110	Any hour
	<-0.010	10–110	1800–0500
	>0.010	10–110	1000–1700
	>0.15	10–110	Any hour
Fichtl and Nelson (1970)			Kennedy Space Center, unstable daytime conditions
	-0.063	3–18	Mean
	-0.012 to -0.097	3–18	Range
	-0.023	3–120	Mean
	-0.019 to -0.027	3–120	Range
Deacon (1969)			Porton, England
	-0.2	1.2–7.1	Clear June midday
	-0.08	1.2–7.1	Cloudy June midday
	-0.05	7.1–17.1	Clear June midday
	-0.02	7.1–17.1	Cloudy June midday
Thompson (1967)	0.35	2–16	Red Butte Canyon, Utah, 1705–1805, 24 Sep 57

Table 1 (Continued)

Reference	$\Delta T/\Delta z$ ($^{o}C/m$)	layer (m)	Comments
Hansen (1966)			WSMR, semiarid climate, inhomogeneous surface
	-0.04 to -0.06	1.15-61.10	Typical midday range
	0.12	1.15-61.10	Strongest inversion in data
Geiger (1965)			Porton, England
	-6.82	0.025-0.30	June mean at noon
	-0.77	0.30-1.20	June mean at noon
	-0.65	0.025-0.30	December mean at noon
			Egypt
	-0.12	1.1-16.2	Mean July midday
	-0.018	16.2-61.0	Mean July midday
	0.076	1.1-16.2	Mean July near sunrise
	0.022	16.2-61.0	Mean July near sunrise
	-0.073	1.1-16.2	Mean January midday
	-0.016	16.2-61.0	Mean January midday
	0.099	1.1-16.2	Mean January near sunrise
	0.027	16.2-61.0	Mean January near sunrise
Godske et al. (1957)			Porton, England, at night, 2100-0300
	3.01	0.025-0.30	Clear June
	0.23	0.025-0.30	Cloudy June
	2.06	0.025-0.30	Clear December
	0.29	0.025-0.30	Cloudy December
	0.60	0.30-1.20	Clear June
	0.035	0.30-1.20	Cloudy June
	0.52	0.30-1.20	Clear December
	0.051	0.30-1.20	Cloudy December
Lettau and Davidson (1957)			O'Neill, Nebraska, August - September 1953
	-1.2 to -1.9	0.39-0.82	Middle part of clear or partly cloudy day
	1.1	0.39-0.82	Strongest inversion on clear night

Table 2. Summary of studies of vertical variation of water vapor expressed in units of specific humidity (g kg^{-1} m^{-1}).

Reference	$\Delta q/\Delta z$ (g/kg)/m	layer (m)	Comments
Sasano (1985)	-0.010	sfc-100	Kanto Plain in Japan, 0801-0812, a Nov. morning
Xing-Sheng et al. (1983)			Boulder Atmospheric Observatory, Colorado, USA
	-0.015	2-22	0250-0300 MST, 24 April 1981
	-0.0013	22-100	0250-0300 MST, 24 April 1981
Klug and Webs (1981)			Federal Republic of Germany
	<-0.005	sfc-100	20 percent of time
	> 0.005	sfc-100	8 percent of time
Geiger (1965)			Quickborn, Holstein, Federal Republic of Germany
	-0.060	2-13	*Mean for clear July weather at 0400 hours*
	0.052	2-13	Mean for clear July weather at 1200 hours
			Rye, England
	-0.33 to 0.18	1.1-15.2	Range
	-0.04	1.1-15.2	10^{th} percentile
	0.02	1.1-15.2	90^{th} percentile
	-0.06 to 0.04	47.2-106.7	Range
	-0.007	47.2-106.7	10^{th} percentile
	0.004	47.2-106.7	90^{th} percentile
			Poona, India
	0.37	0.15-3.05	Clear winter weather near sunrise
	-0.20	0.15-3.05	Clear winter weather at midday
Lettau and Davidson (1957)			O'Neill, Nebraska, Aug-Sep 1953, clear
	-0.49	0.39-0.82	10^{th} percentile
	0.11	0.39-0.82	90^{th} percentile

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APPENDIX

Potential temperature is often used instead of temperature in atmospheric boundary layer studies. The potential temperature, θ, is related to the temperature, T, according to the following equation:

$$\frac{T}{\theta} = \left(\frac{p}{p_o}\right)^{R/C_p} \tag{1}$$

Both temperature and potential temperature must be measured on an absolute scale. The pressure p and the reference pressure p_o must be in the same units, and p_o is normally assumed to be 1000 mb. The superscript is the ratio of the specific gas constant, R, to the specific heat at constant pressure, c_p. Logarithmic differentiation of Equation 1 with respect to height, Z, yields

$$\frac{1}{T}\frac{\partial T}{\partial Z} = \frac{1}{\theta}\frac{\partial \theta}{\partial Z} + \frac{R}{C_p p}\frac{\partial p}{\partial Z} \tag{2}$$

The rate of change of pressure with height is given by the hydrostatic equation:

$$\frac{\partial p}{\partial Z} = -\rho g \tag{3}$$

where the Greek letter ρ is density and g is the acceleration due to gravity. If Equation 2 is multiplied by T and substitution is made from the hydrostatic equation, the result is

$$\frac{\partial T}{\partial Z} = \frac{T}{\theta}\frac{\partial \theta}{\partial Z} - \frac{g}{C_p}\left(\frac{\rho R T}{p}\right) \tag{4}$$

The equation of state

$$p = \rho R T \tag{5}$$

may be used to simplify the last term in Equation 4. Near the surface of the earth, the ratio T/θ is approximately unity and Equation 4 becomes

$$\frac{\partial T}{\partial Z} = \frac{\partial \theta}{\partial Z} - \frac{g}{C_p} \tag{6}$$

The ratio g/c_p is approximately 10 °C/km or 0.01 °C/m. This is often called the adiabatic lapse rate.

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