UNIVERSITY OF ARIZONA

INSTITUTE OF ATMOSPHERIC PHYSICS

TECHNICAL REPORTS ON THE METEOROLOGY AND CLIMATOLOGY

OF ARID REGIONS

No. 3

SIGNIFICANCE OF DIFFERENT VERTICAL DISTRIBUTIONS OF WATER VAPOR IN ARID AND HUMID REGIONS

Horace R. Byers

March 15, 1957

TABLE OF CONTENTS

SIGNIFICANCE OF DIFFERENT VERTICAL DISTRIBUTIONS OF WATER VAPOR IN ARID AND HUMID REGIONS

ABSTRACT

The vertical distribution of water vapor can be expressed by an index coefficient which provides information about eddy and advective transports in a region or in an air mass. The relationship between evapotranspiration and eddy diffusivity of water vapor can be studied in this way. Striking differences in conditions between the arid Southwest and the remainder of the country are shown.

I. INTRODUCTION.

A number of writers, notably Reed (1933, 1939), Wexler and Namias (1938), Byers (1944), Jurwitz (1953), and Bryson and Lowry (1955) have pointed out that summer rains in the southwestern states develop with upper-air flow from the general region of the eastern Gulf of Mexico. In Southern Arizona the annual monsoon-like influx of this moist tropical air mass is a striking climatic feature. The monthly mean upper-air charts published regularly in the Monthly Weather Review often show, for the summer rainy season, flow directly from the southeast over Southern Arizona at 700 and 500 mb. At 850 mb a similar but less well-defined flow is indicated; at 300 mb the westerlies are dominant.

With the main water-vapor supply coming in aloft, over the mountains, one would expect that the vertical distribution of water vapor would be different from that found in moist tropical air masses in the eastern two-thirds of the country. One does not find that the water-vapor mixing ratio increases with height, however; apparently a certain amount of water is going into the air by evapotranspiration from the ground. A difference in vertical moisture gradient between the two areas is distinguishable.

II. EXPRESSION FOR VERTICAL DISTRIBUTION.

For comparison purposes, it is useful to consider the vertical distribution of water vapor to be the result of eddy transport. One can characterize the vertical distribution of water vapor by the quantity

$$-\frac{1}{q}\frac{dq}{dz}$$

where q is the specific humidity and z is the height. If this is designated by a reciprocal coefficient, λ^{-1}, the differential equation

$$\frac{dq}{q} = -\frac{dz}{\lambda} \tag{1}$$

may be written, which, integrated with λ constant, gives

$$\int_{z=0}^{z=z} \frac{dq}{q} = -\frac{1}{\lambda}\int_0^z dz \tag{2}$$

$$\ln\frac{q}{q_o} = -\frac{z}{\lambda}$$

$$q = q_o e^{-z/\lambda} \tag{3}$$

The coefficient λ has dimensions of length, being actually the height at which $q = \frac{1}{e}\, q_o$. It has a connotation of a characteristic mixing length or height for vertical distribution of water vapor by eddy diffusion. In the case of water vapor brought in strongly aloft by advection, as believed to be the case in Southern Arizona summer rains, this interpretation cannot be made.

If λ is constant in the vertical direction, the specific humidity decreases logarithmically with height. This is essentially the case in moist maritime tropical air of summer in the eastern United States, especially as exemplified by the mean sounding for thunderstorm days in Central Florida (Byers and Braham, 1949) where λ is 3.01 ± 0.32 km from the surface to 5 km as shown in Figure 1. This may represent an approximately steady-state distribution in actively stirred tropical air over a humid land surface.

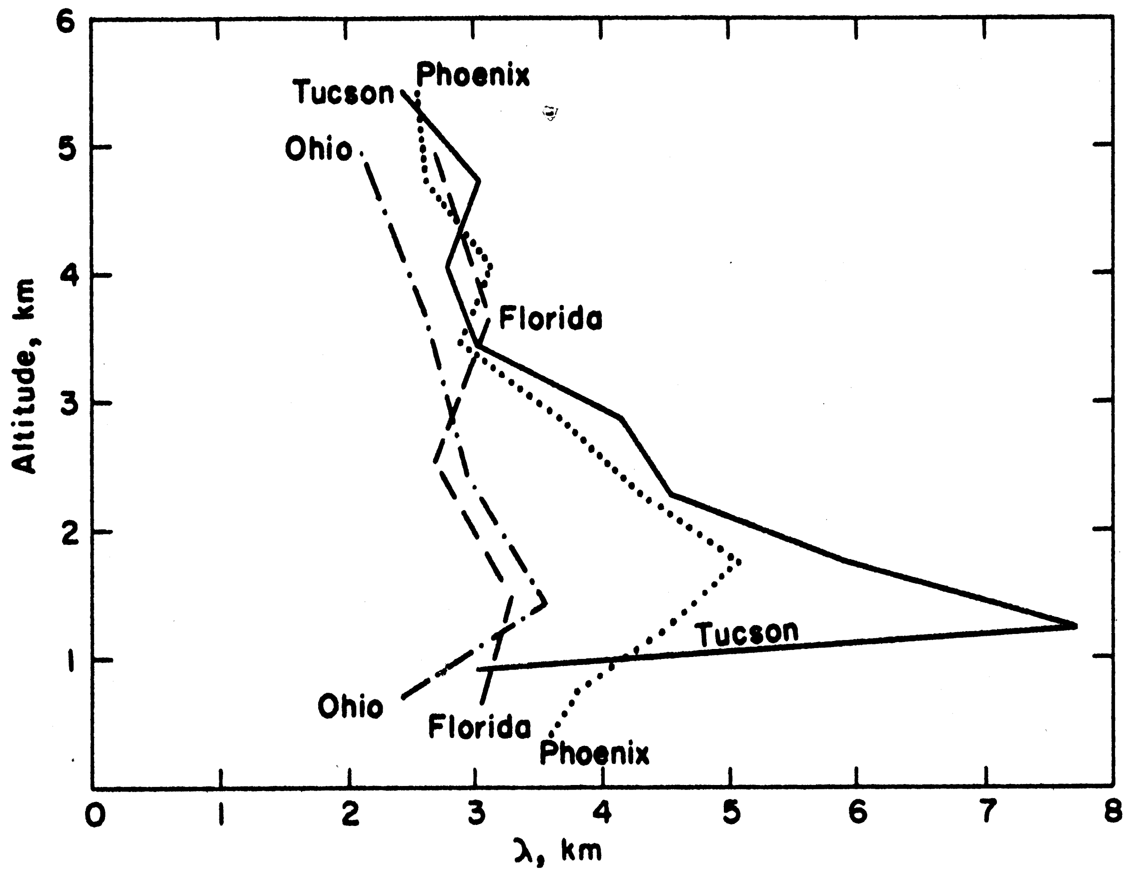
It is to be expected that λ is a measure of the eddy diffusivity and therefore of the mean upward transport of water vapor. It is related to the coefficient of eddy diffusivity D with which the upward mass transport across a cm^2 horizontal area in one second is expressed in primitive form by

$$T = -D\rho \frac{dq}{dz} \tag{4}$$

where ρ is the air density. The relationship is

$$\lambda = \frac{q\rho D}{T} = \frac{\rho_w D}{T} \tag{5}$$

The last expression comes from the definition $q = \rho_w/\rho$, where ρ_w is the water-vapor density. For a given D, λ^{-1} is the ratio of the mass of water vapor transported in unit time across a unit hori-



Phoenix
Tucson
Ohio
Florida
Tucson
Ohio
Florida
Phoenix
Altitude, km
λ, km
0
1
2
3
4
5
6
7
8

Figure 1

zontal area to the mass of vapor present in a unit volume at the point in question. If, for a given value of D, the vertical transport is directly proportional to the vapor density, then λ is constant. The Florida sounding in Figure 1 suggests these conditions, and it is reasonable to conclude that above the surface layers the eddy diffusivity D is constant in well-stirred tropical air.

In comparing meteorological situations, one might conclude that an unusually high value of λ means

1) that water vapor is being brought in horizontally by advection aloft in addition to that which is evaporating from the surface and/or

2) that the surface is dry or has a low rate of evapotranspiration (T is small) and/or

3) that the eddy diffusivity, D, has a large value.

A low positive value could be interpreted as indicating

4) that the lower moist air is brought in under dry air aloft (e.g. Gulf air under the dry westerlies) and/or

5) that the rate of evapotranspiration is high and/or

6) that the atmosphere is thermally quite stable, perhaps with a temperature inversion ($\left|\frac{-1}{q}\frac{dq}{dz}\right|$ large, D small).

Negative values would signify increasing specific humidity with height and downward transport by any eddies which might exist.

One might expect that, in summer, conditions 1), 2) and 3) would be present in Arizona, that 4) and 5) would typify the humid eastern half of the country and 6) would be characteristic of the Pacific Coast.

III. COMPARISON BETWEEN DIFFERENT REGIONS.

Values of λ plotted in Figure 1 were obtained by computing $-\frac{1}{q}\frac{dq}{dz}$ through 50-mb intervals in the mean soundings. The soundings are those taken on 39 thunderstorm days in the summer of 1946 near St. Cloud, Fla., 30 thunderstorm days in the summer of 1947 near Wilmington, Ohio (Byers and Braham, 1949), 79 rain days in July and August, 1949 - 1954 at Tucson, Arizona, and 53 rain days for the same months and years at Phoenix, Arizona.

Except for some effects near the surface, the values of λ up to 3 km are considerably higher in Arizona than in Florida and Ohio. The two eastern stations are somewhat similar up to that height. Phoenix and Tucson are quite similar, as would be expected. Above 3.5 km the Florida and Arizona curves are very close, but Ohio shows lower values.

The comparison of the Arizona curves with those for the more humid eastern locations suggests the effects of upper-air advection of water vapor from the southeast. This upper moist air is in typical tropical equilibrium above 3 km. In the lowest level, near the ground, the lower values of λ at Phoenix and Tucson could be explained by the fact that the soundings are made at 2000^h local time when some water vapor might be expected to have collected near the ground after the daytime heat convection has subsided.

The Florida data, as explained before, probably represent the true steady-state equilibrium for summer tropical air. The soundings there, as well as in Ohio, were nearly all taken in the afternoon. The low values in Ohio at the upper levels suggest mixing with the dry upper westerlies common to that area but not found at this height in the other locations. The lower values next to the ground in Ohio may represent greater evapotranspiration there than in Florida. The striking differences in the local terrain, soil, and vegetation cover between the two places could account for this. The terrain conditions are described by Byers and Braham (1949).

The study represented in this paper was begun with the rain conditions at the different locations in mind. It became apparent that it would be useful to determine the extent to which the differences found on rain and thunderstorm days were true of the entire summer season. Mean monthly soundings for the summer season in the years 1946-1955 at selected U.S. stations already tabulated by the Institute of Atmospheric Physics were used to compute the values of λ. The combined results for June, July and August of this ten-year period are shown in Table 1.

Most striking in this table is the systematic difference between the values of λ in the southern Mountain States and in the Central, Southern and Pacific Coast States. Comparing the mean values between 800 and 600 mb at the stations listed in the first six columns with the mean of all the others, one finds a λ of 4.6 for the former against 2.5 for the latter. The transition to the

Table 1

Mean Seasonal Values of λ (km)
June, July, August
1946-55, inclusive

Pressure interval (mb)	Phoenix	Las Vegas	Ely	Grand Junction	El Paso	Albuquerque	San Antonio	Oklahoma City	Dodge City	Omaha	Nashville	Brownsville	Lake Charles	Miami	Santa Maria	Tatoosh I.
1000-950												2.2	2.2	2.6	1.3*†	2.3
950-900	10.0						5.7	4.5		4.8	3.4	1.8	2.7	2.9	4.8	2.8
900-850	7.2	4.9					3.8	3.7	3.5	3.3	3.8	1.9	2.6	2.8	3.6	3.2
850-800	6.0	5.8			6.1		2.6	2.9	3.6	2.7	2.8	2.3	2.4	2.5	10.8*	2.7
800-750	5.2	7.2	3.1	3.2	5.7	4.6	1.9	2.2	2.9	2.3	2.4	2.6	2.7	2.6	2.3	2.8
750-700	3.8	6.4	5.1	6.8	6.3	5.0	2.0	2.2	2.7	2.6	2.3	3.1	2.8	2.7	2.2	2.4
700-650	2.8	4.6	5.2	4.6	4.3	5.5	2.4	2.5	2.3	2.5	2.8	2.8	2.8	3.0	2.6	2.7
650-600	2.4	2.5	3.7	3.9	3.1	4.4	2.3	2.7	1.9	2.3	2.3	2.5	2.7	3.3	2.7	2.4
600-550	2.5	2.3	3.2	2.8	2.6	3.8	2.3	2.4	2.6	1.9	2.3	2.5	2.7	2.6	2.8	2.1
550-500	2.1	1.7	2.0	2.1	1.9	2.4	2.0	2.1	1.8	1.8	1.7	2.1	2.1	2.5	4.0*	2.5

*Contains values of $\Delta\varphi$ near 0. †Contains negative values.

east occurs in the 475 miles between Grand Junction and Dodge City, between Albuquerque and Oklahoma City and the somewhat greater distance between El Paso and San Antonio, while on the west the transition is sharp between Las Vegas and Santa Maria (approximately 275 mi.). The belt of high λ-values occupies more than a third of the longitudinal span of the country.

It may be somewhat surprising to note that the mean summer values at comparable stations are very close to those for rain or thunderstorm days plotted in Figure 1. At the upper levels the whole summer mean is a slight amount lower than on rain or thunderstorm days and in the low levels it is greater in the mean than on selected days, but this is true only at inland stations. The low-level difference between Nashville of Table 1 and Wilmington, Ohio of Figure 1, for example, must be due to the fact that the soundings at the latter station were taken in the afternoon while at the standard stations they are at 0300 Greenwich time. Since the soundings in the table were taken in a variety of air masses, it is suggested that the diurnal changes in λ in the low levels in summer are roughly the same as the changes between one air mass and another.

To get an idea of the differences between air masses, the values of λ were computed from the tables of Showalter (1939) for the three principal summer air masses at Omaha - maritime tropical, superior and continental polar. The data are shown in Table 2. The data are for 1936, an exceptionally dry summer.

Table 2

Values of λ in Summer Air Masses at Omaha

Height, m	mT air	S air	cP air
500-1000	5.0	2.7	3.0
1000-1500	5.0	3.1	3.3
1500-2000	4.8	1.6	3.6
2000-2500	4.0	3.0	3.6
2500-3000	3.4	2.9	
3000-4000	3.1	2.5	
4000-5000	3.7	2.0	

Three other plots based on mean soundings are shown in Figure 2. They are taken from Showalter's (1939) tables of air-mass properties. The mean curve for maritime tropical air in summer at Pensacola, Fla. shows very little systematic variation with height. The values of λ run slightly higher than in the St. Cloud soundings, possibly because some of the data here may represent a steady state with respect to the ocean surface (Pensacola is on the northern Gulf of Mexico). The mean data for the very dry superior air mass show the characteristic low values to be expected when dry air comes in aloft. There were no cases of this air mass below 1000 m during Showalter's period of study. The El Paso summer mT air, also plotted in this figure, has values comparable to the soundings for summer rain days at Tucson (Figure 1).

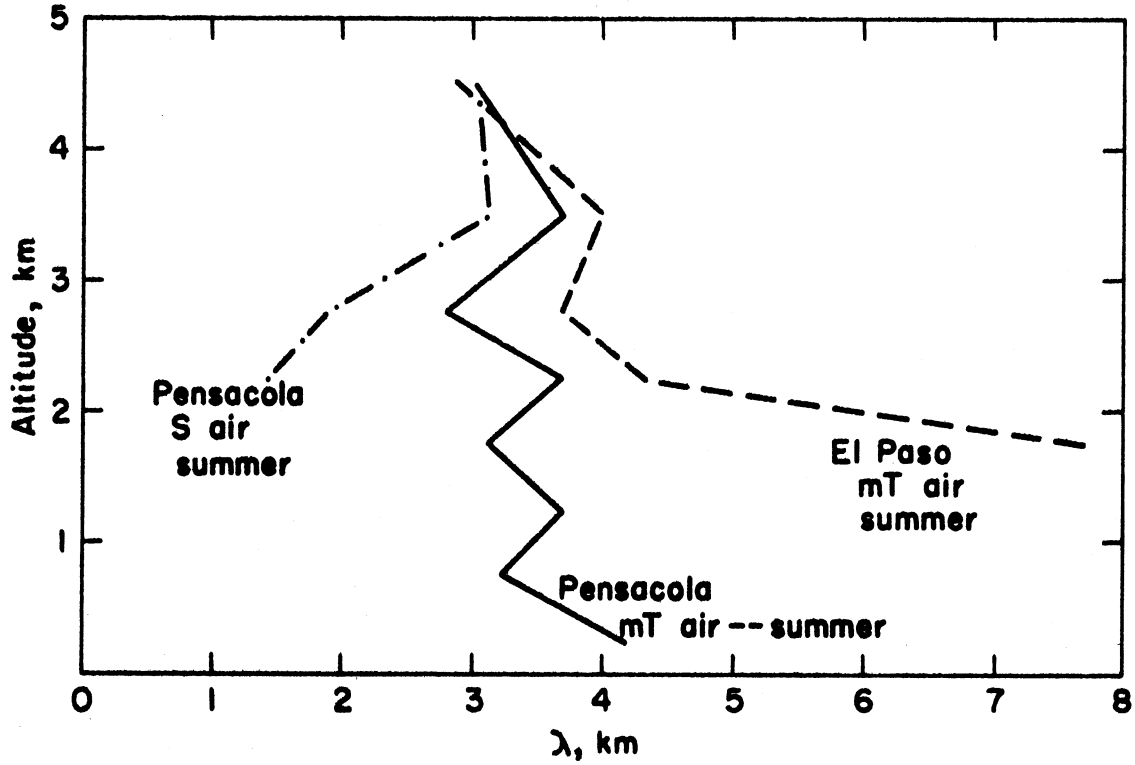

Altitude, km
λ, km
0
1
2
3
4
5
6
7
8
Pensacola
S air
summer
El Paso
mT air
summer
Pensacola
mT air -- summer

Figure 2

One should be cautioned against rigorous comparison between Figure 1 and Figure 2, since the data were assembled in two entirely different ways. Showalter collected the data by levels, but the curves in Figure 1 are based on all levels for each day considered. For example, Showalter has 67 observations of mT air at Pensacola at 500 m, 41 at 5000 m, and for superior air, 1 observation at 1000 m, 27 at 5000 m.

IV. RELATION TO EVAPOTRANSPIRATION.

At some level near the surface, the upward transport of water vapor, T, would be equivalent to the evapotranspiration from the surface. Suitable measurements of evapotranspiration for conditions similar to those of the soundings used here have not been made. For comparison purposes between regions, relative values can be obtained by making some simplifying assumptions in the equation

$$T = \frac{\rho_w D}{\lambda} \tag{6}$$

We might assume that for afternoon conditions, the value of λ in Arizona in the low levels is twice the value of that quantity in Ohio. It is known that the water-vapor content near the ground is about two-thirds that of Ohio. If the eddy diffusivity D is the same in the two places, then T, or the evapotranspiration, in Arizona is one-third that of Ohio.

Benton and Dominitz (1954) have calculated evapotranspiration in various parts of North America from the mass balance of water vapor in the atmosphere. For the North Central United States, the evapotranspiration averages about 5.6 inches of water per month during the summer months. If under these mean conditions it is also true that the evapotranspiration rate of Arizona at this season is one-third that of Ohio, it would amount to 1.87 inches per month.

Benton and Estoque (1954) have demonstrated from upper-air data that in summer through the atmosphere over the North American continent more water vapor flows out than in, and have concluded that at that season the continent is a source of water vapor. From a comparison with precipitation data, they show that, averaged over the continent, evapotranspiration exceeds precipitation during summer. The same is shown for smaller areas in the eastern United States by Benton and Dominitz (1954). It is impossible to use their technique for an area as small as the State of Arizona, but if, as indicated above, the evapotranspiration in Southern Arizona is about one-third that of the North Central States, it would average about the same or slightly less than the normal July and August precipitation.

The results of Benton and associates suggest that in summer the moisture enters the continent from the Gulf of Mexico and tropical Atlantic, and mixes with the dry air of the upper westerlies, of which the superior air is a striking example. Over the humid parts of the continent, ρ_w and D are relatively large while λ is small, so T, the evapotranspiration, is great. In the arid regions, although D may be about the same, or possibly slightly higher, λ is considerably higher while ρ_w is considerably less than in humid areas, so T is smaller.

As a check on the various quantities, the Benton and Dominitz value of evapotranspiration for the North Central States is used in conjunction with the mean Ohio thunderstorm soundings to obtain D at 950 mb. The result is

$$D = \frac{T\lambda}{C_w} = 6.45 \times 10^{-4}\ cm^2\ sec^{-1}$$

On the basis of the type of data used here there is no way of determining whether or not this is an appropriate value, but the order of magnitude is reasonable (Sutton 1951).

V. CONCLUSIONS.

The coefficient λ, which is a measure of the vertical distribution of water vapor, is useful for making comparisons between different geographical regions and between air masses. From it, inferences may be drawn concerning some of the predominant meteorological processes which characterize a region or a particular air mass. From the water vapor measurements in aerological soundings, one can obtain D/T, the ratio of the eddy diffusivity to the evapotranspiration, if it is assumed that the water-vapor distribution is determined by vertical transports alone. However, water-vapor advection probably is important. It would be desirable in a continuing phase of this study to determine D under selected conditions so that the relationship to evapotranspiration may be determined more fully.

Regional comparisons show that the average λ for the summer months in the southern Mountain States is nearly twice the average for that season in the Central, Southern and Pacific Coast States. The comparative values are in keeping with reasonable values of D and T. It is to be expected that arid regions should have lower values of T and perhaps higher daytime values of D than humid regions. Water-vapor advection above the 800-mb level in the Mountain States is believed to have a marked effect on the data.

ACKNOWLEDGEMENT.

The mean data for Arizona rain days were computed by Mr. Robert B. Des Jardins of the Institute of Atmospheric Physics.

REFERENCES

Benton, G. S. and J. Dominitz (1954): The use of atmospheric data in the evaluation of evapotranspiration. Johns Hopkins Univ. Dept. Civ. Engrng. Sci. Rept. No. 3.

Benton, G. S. and M. A. Estoque, 1954: Water-vapor transfer over the North American continent. J. Meteor., 11, 462-477.

Bryson, R. A. and W. P. Lowry, 1955: Synoptic climatology of the Arizona summer precipitation singularity. Bull. Am. Meteor. Soc., 36, 329-339.

Byers, H. R., 1944: General meteorology, New York, McGraw-Hill, pp. 274-275, 395.

Jurwitz, L., 1953: Arizona's two-season rainfall pattern. Weatherwise, 6, 96-99.

Reed, T. R., 1933: The North American high-level anticyclone. Mon. Wea. Rev., 61, 321-325.

____, 1939: Thermal aspects of the high-level anticyclone. Mon. Wea. Rev., 67, 201-204.

Showalter, A. K., 1939: Further studies of American air-mass properties. Mon. Wea. Rev., 67, 204-218.

Sutton, O. G., (1951): Atmospheric turbulence and diffusion. Am. Met. Soc. Compendium of Met. (T. F. Malone, ed.), Boston, Am. Met. Soc., p. 494.

Wexler, H. and J. Namias, 1938: Mean monthly isentropic charts and their relation to departures of summer rainfall. Trans. Am. Geophys. Union, 19, 164-170.