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POSSIBLE MODIFICATION OF LOW-LEVEL CONVERGENCE PATTERNS AND VERTICAL VELOCITIES OVER THE RAJASTHAN DESERT

PALEOCLIMATOLOGY
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VI. Derivation of the Governing Equations

At the start of this investigation, the author hoped to estimate the effect of removing certain fractions of the dust on the diurnal variation of the thermal wind from basic thermodynamics and radiation physics. It soon became apparent that such a study far exceed the limitations of the author's time and background. Therefore certain rather brutal assumptions will be applied in order to get at least a qualitative idea of the effect of increasing the diurnal forcing function. It will then remain for others to refine these ideas when the data and investigational tools become available.

Assume that the temperature at any time and height may be described as

$$(6.1) \quad T(z,t) = T_0 + \Gamma z + \Delta T_0 (1 - z/H) \cos \omega t$$

where T_0 is an avera e surface temperature, Γ is a mean lapse rate taken as $-6.5°$ C/km, ΔT_0 is the amplitude of the surface temperature wave, and H is a height taken as that where frictional influences are no longer important. Eq. 6.1 describes an atmosphere with a where the temperature oscillates about its mean with an amplitude decreasing with height. This is only an approximation to the atmosphere's actual behavior. The lapse rate in the lower few meters approximates a linear decrease. Noting this fact, the value of ΔT_0 inserted should be somewhat less than that measured at the surface.

Differentiating (6-1) with respect to x,

$$(6.2) \quad \frac{\partial T}{\partial x} = \frac{\partial T}{\partial z}\frac{\partial z}{\partial x} = \frac{\partial T}{\partial z} \times (\text{SLOPE of GROUND})$$

The thermal wind equation is:

$$(6.3) \quad \frac{\partial V}{\partial z} = \frac{g}{f\bar{T}}\frac{\partial T}{\partial x} = \frac{g S}{f\bar{T}}\frac{\partial T}{\partial z}$$

where g is the gravitational acceleration, f the coriolis parameter, S the slope of the ground surface, $\bar{T}$ is taken as a mean temperature.,

As in Section V., the geostrophic wind V is described as:

$$(6.4) \quad V = \overline{V_g} + z\overline{V_g'} + V_{1,0}(1 - z/H)\cos \omega t$$

The thermal wind is then:

$$(6.5)\quad \frac{\partial V}{\partial z} = \overline{V_g'} - \frac{V_{1,0}}{H} \cos \omega t$$

From 6.1:

$$(6.6)\quad \frac{\partial T}{\partial z} = \Gamma - \frac{\Delta T_0}{H} \cos \omega t$$

Combining (6.5), (6.6) and (6.3) taking Γ =-6.5°C/km, $\bar{T}$ =300°K gives V_g'=-0.0021 sec^{-1} and V $V_{1,0}$ = .327 $\Delta T_0 \cong \Delta T_0/3$. When ΔT_0 is in °C, $V_{1,0}$ will be in M Sec^{-1}.

Recalling the brutality of the assumptions involved, increasing ΔT_0, the amplitude of the diurnal wave, will increase the magnitude of the variation in the thermal wind.

The equations of Section V show that the low level wind distributions depend on geographic latitude through the coriolis parameter. Knowing this distribution, the low-level convergence and the average vertical velocities over the area can be derived from the equation of continuity.

For incompressible flow, the vertical velocity W at height H can be found by integrating the equation of continuity (Haltiner and Martin, 1957)

$$(6.7)\quad w = -\nabla_H \cdot \int_0^H \mathbb{V}\, dz$$

Integrating over the area A

$$\iint_A w\, dA = -\iint_A \left[\nabla_H \cdot \int_0^H \mathbb{V} dz \right] dA$$

Applying Gauss' Theorem, the average vertical velocity $\overline{w}$ over the area can be found as

$$(6.8)\quad \bar{w} A = \oint_C \left[\int_0^H V_n\, dz \right] dR$$

where V_n is the velocity normal to the boundary of the area and R is the distance around the perimeter of the area.

VII. Computations and Results

The outline on Fig. 1 shows the area for which the computations were carried out. The eastern and western boundaries are oriented NE-SW and span 24° N to 29°N. The northern and southern boundaries are 483 km (300 mi) long. Wind speeds at twenty levels to H=1.63 km were computed from the thermo-tidal wind model for each 1° latitude 24° 29°N. The height of 1.63 km corresponds to $z = 2\pi$ in Eq. 5 for K=2.2 m^2/sec and f computed for 26.5N. At this height the amplitude of oscillations are negligable. The geostrophic components were chosen as $\overline{Vg}$=21m/sec, $\overline{Vg'}$=-0.0021/sec. Fig. 4, 5 show typical wind spirals for different values of the thermal wind.

The Fortran program by which these spirals were computed oriented the coordinate system so that the v component is towards the northeast and the u 90° to the right. This is called the geostropic system. A coordinate system rotated 45° to the right will give the winds in the geographic system.

The outline of the area under consideration is given in Fig. 6 with the orientation of the normal components on each boundary shown. The effect of the volume advection through segment AB will tend to cancel that through segment CD and is thus neglected.

Expressing (6.8) finite difference terms, and assuming the values of U for 25°N and 28°N hold as average for segments OA and DE respectively:

$$(7.1) \quad \overline{w} A = \Delta z R_1 \Sigma (V_{24°} - V_{29°}) + \Delta z R_2 \Sigma (U_{25°} - U_{28°})$$

The two terms on the right hand side express the contributions to the net divergence by advection through the north and south boundaries and the easterly and westerly boundaries respectively. Table I gives the results of these calculations.

The diurnal variation of vertical velocity is shown in Fig. 7 for three assumed values of the amplitude of the thermal oscillation.

Figure 5: Wind spirals showing effect of increased thermal wind

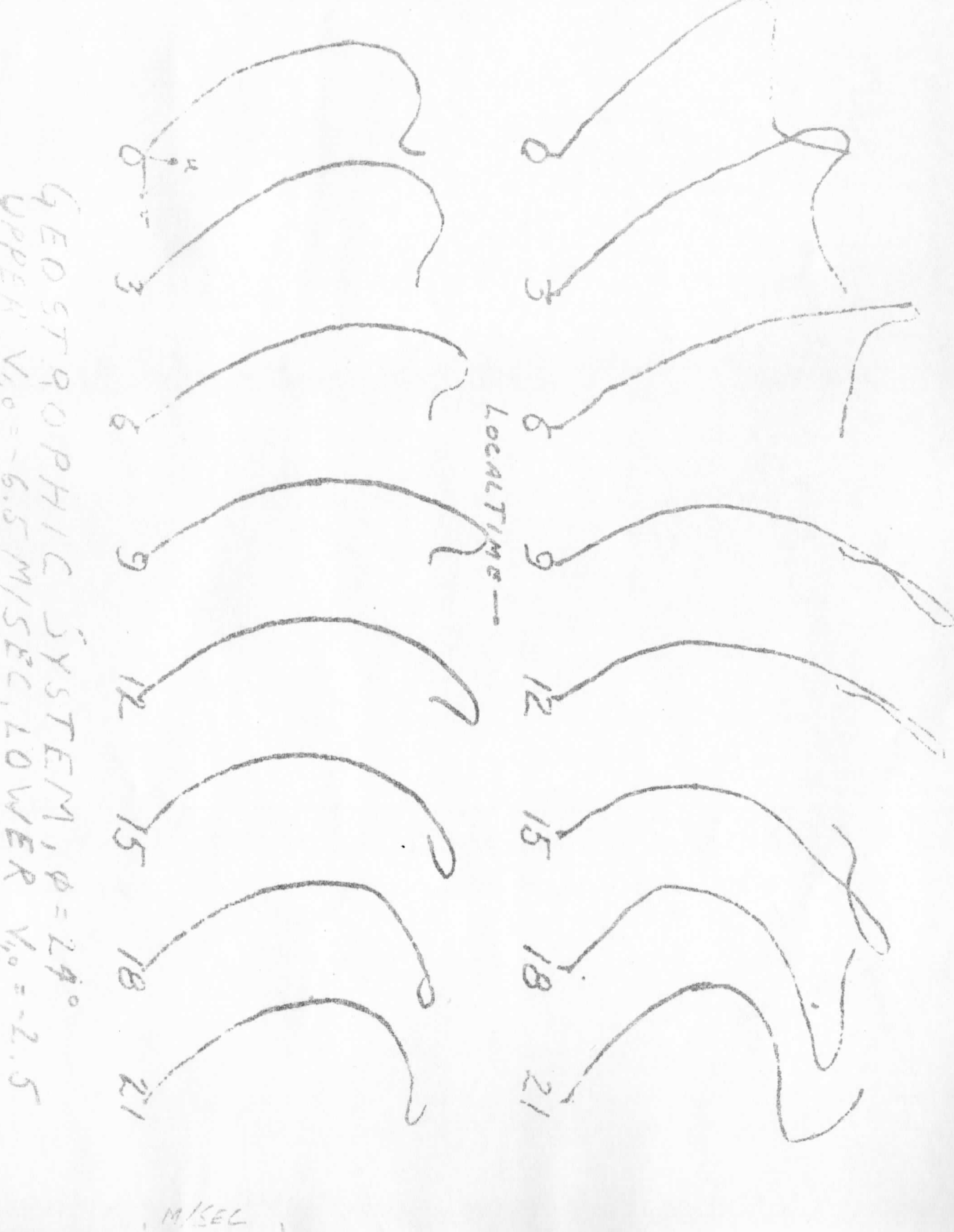

FIGURE 4: Wind spirals showing effect of latitude

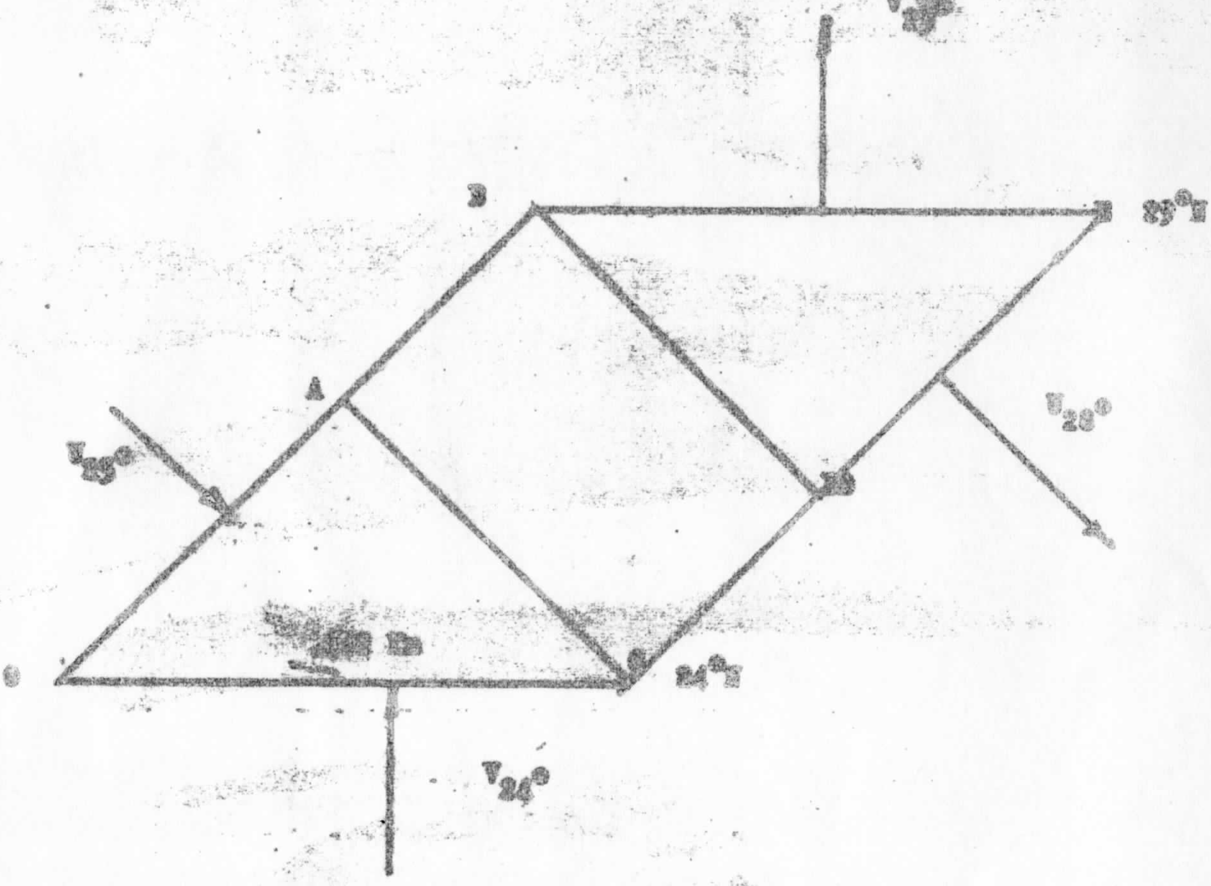

Figure 6; Orientation of area under consideration

TABLE 2

CONTRIBUTIONS TO HORIZONTAL DIVERGENCE

UNITS - KM^3/SEC

	V_1 = -2.5 M/sec		V_1 = -4.5		V_1 = -6.5	
TIME	U	V	U	V	U	V
0	+2.10	-6.84	+3.78	-19.90	+5.46	-18.09
3	+1.45	-7.15	+2.80	-16.14	+3.77	-18.60
6	-.05	-3.32	-.09	-7.88	-.13	-8.62
9	-1.52	+2.46	-2.68	+4.98	-3.96	+6.01
12	-2.10	+6.79	-3.68	+19.90	-5.47	+17.67
15	-1.45	+7.15	-2.61	+16.11	-3.77	+17.53
18	+.05	+3.32	+.09	+7.89	+.13	+8.63
21	+1.52	-2.46	+2.74	-4.95	+3.95	-6.39

POSITIVE VALUES INDICATE NET INFLOW

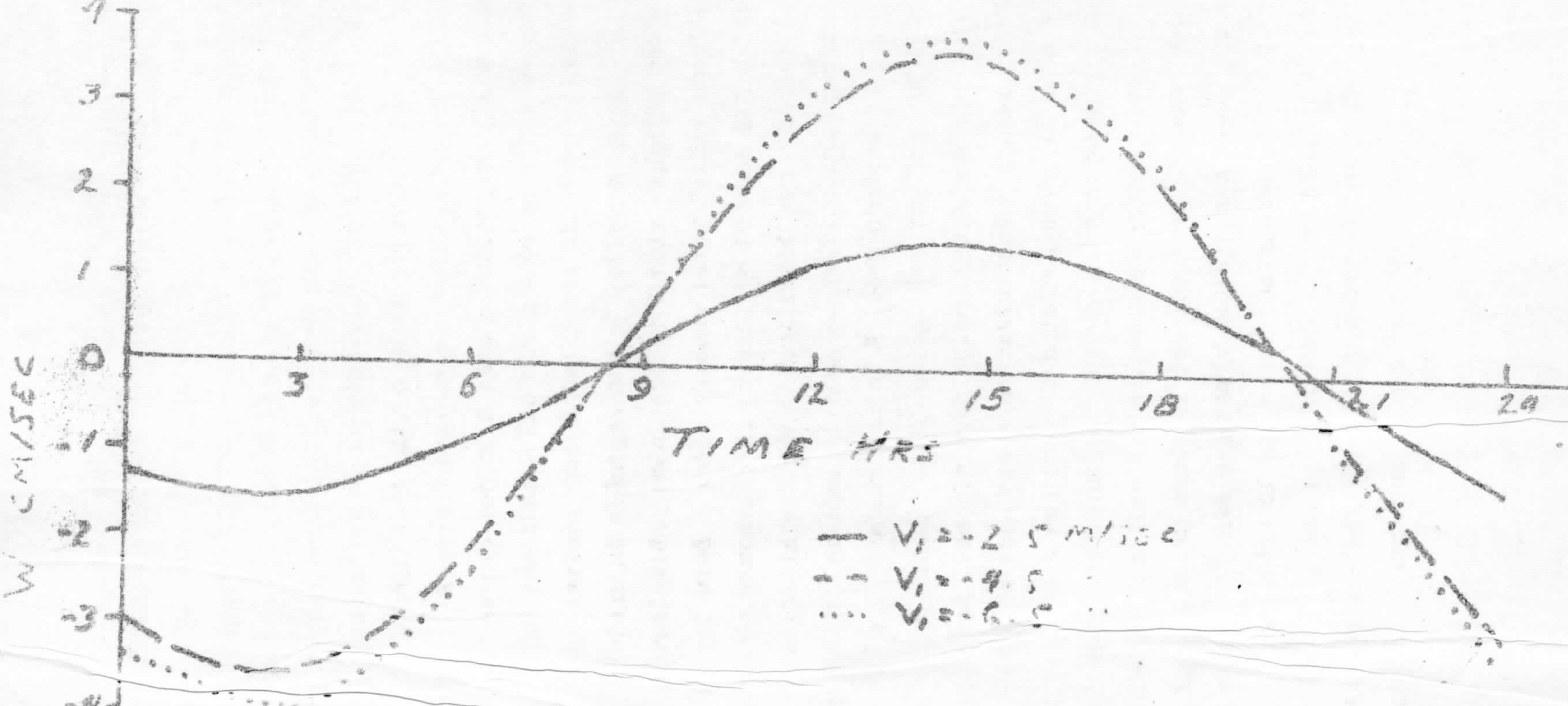

FIG. 7 DIURNAL VARIATION OF VERTICAL VELOCITY

VIII. Discussion

The results of the previous section indicate that increasing the diurnal temperature range will increase the amplitude of the vertical velocity oscillation. This increase will enhance the possibility of rainfall during the day. These conclusions are actually only qualitative since there are a number of rather powerful restrictions which must be considered.

The assumption of an oscillating component of the lapse rate which decreases linearly with height is only a gross approximation of the actual state of the atmosphere. A second approximation might be an exponential decay. In addition the assumption of a simple cosine oscillation for this component is not valid in practice. The minimum temperature occurs sometime shortly before dawn, the maximum near local noon; the temperature wave described above has a minimum at midnight and a maximum at noon.

The form of the thermal-tidal model employed only explains the main features of the low-level wind a while already being fairly complicated mathematically. The eddy viscosity, K, assumed to be constant actually varies with time and height in a fairly complicated way. Staley (1956) has worked with the similar problem of heat conduction, and states, "With increasing functional elaboration of K (eddy conductivity), solutions have become more and more complex, culminating, for the case of K varying sinusoidally with time and linearly with height, in a mathematical extravaganza...(p. 14)" He then gives an intimidating summary of the solution. Thus one sees that not only does introduction of some variable eddy viscosity complicate the solution, but that the temperature solution is actually much more complex than the simple form offered here.

Sangster (1958) in a similar study of low level convergence over the Great Plains finds that with southerly geostrophic flow, the advection through the east and west sides of his areas contribute little to the net devergence. From Table I it can be seen that if the area considered in this study were oriented so that the side boundaries were also north-south that the avection through them would exactly cancel.

Comparing the amplitudes of the waves in Fig. 7 it can be seen that increasing the amplitude of the thermal wind from 4.5 to 6.5 m/sec increases the amplitude of the vertical velocity variation an insignificant amount. It would appear, then, that the vertical velocity response to an increase in the thermal wind is greatest for initial increases, and then approaches some maximum value, thus a favorable omen for climatic modification.

To conclude, it appears that if the temperature range can be increased by stabilization of the surface of the Rajasthan Desert, the diurnal variation of the vertical velocity over the area will be increased leading to greater instability during the day. Before a more quantitative determination of the magnitude of the effects can be made the actual form of the temperature lapse rate must be made and applied to more sophisticated forms of the thermo-tidal wind model.

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POSSIBLE MODIFICATION OF LOW-LEVEL CONVERGENCE PATTERNS
AND VERTICAL VELOCITIES OVER THE RAJASTHAN DESERT

Once in every ten or twenty yards, some grey-green plant, deep rooted and too thorny for even camels to eat, tenaceously and with a kind of vegetable ferocity struggles for life. And at longer intervals, draining the moisture from a rood of land, there rise, here and there, the little stunted trees of the desert. From close at hand the sparseness of their distantly scattered growth is manifest. But seen in depth down the long perspective of receeding distance, they seem like the in fact remotely scattered stars of the Milky Way--numerous and densely packed. Close at hand the desert is only rarely flecked by shade; but the further distances seem hedged with a dense dark growth of trees. The foreground is always desert, but on every horizon there is the semblance of shadowing forest. The train rolls on, and the forests remain for ever on the horizon; around one is always and only the desert.

...Aldous Huxley, Jesting Pilate

I. Introduction

Rajasthan lies in the northwest part of the Indian Subcontinent and comprises what was formerly a member of princely states under the division called Rajputana. Roughly 60% of the land area lies to the northwest of the Aravalli Hills, which form a rough climatic divide running NE-SW. The area to the northwest of the Aravallis is characterised by low rainfall, great annual and diurnal temperature ranges, and high potential evaporation. (Pramanik, 1952a,b). The dryest part of this area is the Thar or Great Indian Desert with an annual precipitation of less than 5" and generally unpleasant climate (Kendrew, 1942, p.121).[1] The land to the east and south of

1. It is interesting here to note the subjective descriptions of the same area by two authors with different backgrounds. Kendrew, an Englishman writing during the last days of the British Raj, says, "Work must be suspended during the hottest hours, and any activity out of doors is impossible as long as the sun is above the horizon. It is inadvisable to venture into the open without taking careful precautions against the sun, for the heat and the glare, both direct and reflected, are intense." (1942, p. 121). Later (pp. 145-148) he quotes an even more horrendous account by one Rev. J. M. Merk,

of the Aravallis is higher, more fertile, and receives more rainfall than that to the west and northwest. (Pramanik, 1952a).

The area lying to the northwest of the Aravallis is variously referred to as the Rajasthan or Rajputana Desert, and, depending on the definition used, occupies one-quarter to one-half a million square miles. Bryson and Baerreis (N. D.) place the boundaries as roughly Delhi to Karachi, the Rann of Kutch to the Himalayas. (See Fig. 1). Some authors (e.g. Pruthi and Bhatia, 1952) feel that the designation desert is a misnomer since after the few rains that do fall there is abundant growth of annual vegetation the density of growth depending on the amount of precipation.

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a resident of the Punjab, an account which Kendrew refers to as "a good description of the weather at this (July) time."

In contrast, Agharkar (1952) refers to the area as " This dry, hot and unproductive desert, which is renowned for its bracing and healthy climate....(p. 245)"

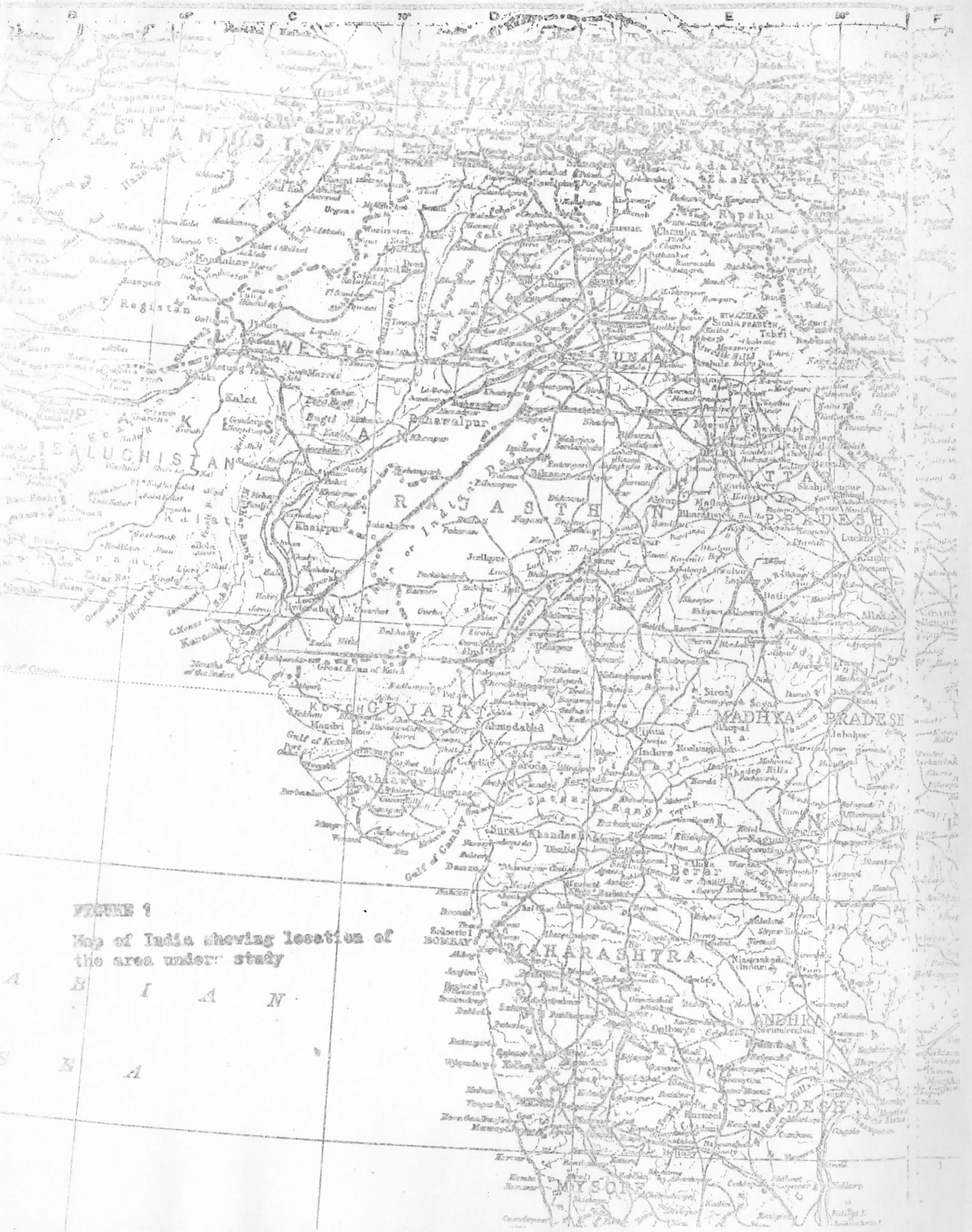


FIGURE 1

Map of India showing location of the area under study

II. The Present Climate of Rajasthan.

The location of Rajasthan would ordinarily lead one to believe that the area should be characterized a wet winter and dry summer, while actually the reverse is the case. In addition, the contrast between the rainfall patterns of northwest and northeast India presents a problem, especially considering the similarity of the climatic controls which influence the two areas (See Trewartha, 1961, pp. 166-168).

The northwestern part of India is the center of a strong surface low pressure area, is invaded by equatorial air from the Arabian Sea, at least in the surface layers. This air would appear to be trapped by highlands, producing convergence and ascent. All of these factors should tend to produce greater rainfall than is now experienced.

Counteracting these influences, however, is the fact that the area is capped by anticyclonic air aloft, with associated subsidence. Thus the humid monsoon air enters the area through a shallow layer only, and has little chance for vertical development. It seems then that if there were less subsidence over the area, the moist air layer would be deeper and the summer rainfall would be greater.

Bryson and Baerreis (N.D.) describe the blanket of dust which covers much of the tropics from North Africa to Burma, and note that the layer is thickest and most dense over Rajasthan. Bryson et al. (1963) have measured the radiation divergence over northern India, and have calculated that the effect of the dust layer is to increase subsidence over the area by as much as 50% over that which would occur in a dust free atmosphere.

III. The Past Climate of Rajasthan

There is evidence that the climatic of Rajasthan was formerly more conducive to agricultural production and human settlement (See Bryson and Baerreis, N.D. pp. 7-14., also Sankhalia, 1952). Archaeological evidence, somewhat conflicting, indicates that the Harappan civilizati n flourished in this area at abo t 1500-2000 B. C. The evidence suggests that the climate during this period was wetter and allowed extensive settlements in the valleys of the ancient Saravati and Drashadvati rivers (present Chaggar and Chautang rivers), which now flow only during the monsoon season.

The terminal date for the Harappan culture is uncertain, but the evidence suggests that it was followed by the Painted Grey Ware culture. It appears that the climate during the period in which this culture predominated was dry r. The few sites found are on the river banks while the Harappan sites w re farther from the rivers over looking the valleys. This might indicate that Harappans farmed the flood plains of the rivers, while the Painted Grey Ware People had to settle near the river in order to utilize the last of the disappearing water suppl . The dates of occupation by this culture are also uncertain, but may be bracketed as roughly 1000-270 B. C. , indicating a break in occupation between the end of Harappan and the arrival of the Painted Grey Ware peoples.

After the time of the Painted Grey Ware culture, it seems as if the climate again became f vorable for extensive habitation. This is the period of the Rangmahal culture. The arrival of Rangmahal culture may have been contemporaneous with the end of the Painted Grey Ware culture or there may have been some break b tween the two. However, the Rangmahal Culture flourished in the early centuries A. D. with extensive occupation, some sites having mud walls around them and evidenceof brick houses. This culture lasted until 700 or 800 A.D. when the population became nomadic, possibly as a result of a return to dryer conditions.

In summary, the archeological evidence suggests fairly wet conditions about 2000-1500 B.C., dryer conditions between about 1000-270 B.C., a return to wetter conditions from roughly 200-700 A.D. and a final return to dry climate.

The archeological evidence leaves the question of the cause of the dessication of Rajasthan open to question. This dessication may have been caused by natural climatic shifts, cultural practices of land use, or a combination of the two. Bryson and Baerreis (N.D.) note that large changes are not needed to explain the deterioration of the climate.

> Indeed, the evidence is that the plants that grew in the region of Harappan occupation are largely found in the fringes of the area today and the same can be said of the animals. At the present time, the desert is said to be advancing into the arable lands at the rate of half a mile a year, indicating that the ecological balance of man, plant, and climate is imperfect. This does not seem to be a recent development, yet the climate around the world has changed within the past hundred years. If one assumes that the advance has averaged a quarter to half a mile per year for some time, and extrapolates backward, the appearance of a widening fringe of desert about the Thar Desert core would date to 400 to 1000 years ago, at the end of Rangmahal time or later. (Bryson and Baerreis, p. 14)

IV. Possible Climatic Modification.

Although, as indicated above, the causes of dessication are open to question, it appears that the sequence might have been somewhat as follows: (Hora, 19 2) Natural climatic changes brought warmer and dryer conditions, at the same time, man cut down forests for domestic purposes faster than it could be naturally replentished. Grazing by herds of the now nomadic peoples further reduced the vegetative cover. All of these factors increased the amount of dust which could be raised by the wind. Seth (1961) notes the reports of dust storms in southwest Rajasthan by the seventh century A.D. The addition of dust to the atmosphere increases the radiation divergence in the mid-troposphere, which inturn means diabatic cooling and increased subsidence and, thus, less rainfall. Whatever the triggering mechanism might have been, once the process was started, a positive feed-back, along with land misuse by humans, accentuated the dry conditions and the creep of the desert into previously arable lands.

Bryson and Bae r is suggest that the process c n be reversed by stabilizing of the surface with vegatative cover, possibly grass. This cut off the source of dust and eventually decrease the subsidence leading to increased summer rainfall. This increased rainfall would in turn keep the grass growing and reverse the growth of the desert.

They also list a number of further possible consequences of decreasing the dust blanket. One of these would be an increase of the diurnal forcing function. An increase of the forcing function woild lead to an increased diurnal temperature range at the surface. The purpose of this paper, after these somewhat lengthy preliminaries, is to investigate the effect of possible increa e of diurnal temperature range on the low level flow after Lettau's (1964) thermo-tidal wind spiral model.

V. Lettau's Termo-Tidal Wind Spiral Model

Detailed wind profile measurements in the lowest 1.5 km. of the atmospheric boundary layer (Lettau and Davidson, 1957) documented the nocturnal low-level jet over the Great Plains of the United States. This jet stream is characterized b a level a few hundred meters above the surface where the wind speed is considerably super-geostrophic. Pitchford and London (1962) have shown that there seems to be a direct relationship between occurences of the low-level jet and nocturnal thunderstorms over the Great Plains.

Lettau (1964, 1966) has attempted to explain the low-level jet as a result of the diurnal oscillation of the thermal wind caused by heating of a sloping surface. Heat diffusion is normal to the earth's surface. Thus for a gently sloping surface the isotherms will be parallel to this surface. Consequently, a horizontal temperature gradient is established. This horizontal temperature gradient induces a vertical variation in the geostrophic wind, or what is known as a "thermal-wind." The diurnal oscillation will induce a diurnal oscillation in the horizontal temperature gradient, i. e. a diurnally oscillating thermal wind. Introduction of this oscillating thermal wind into the equations of motion predict the main features of the nocturnal low-level jet over the Great Plains.

Assuming that inertia accelerations are unimportant and that an apparent eddy viscosity, K, can be defined which is independent of time and height, the horizontal equations of motion are:

$$\frac{\partial u}{\partial t} - (v - V)f - K\frac{\partial^2 u}{\partial z^2} = 0$$

$$\frac{\partial v}{\partial t} + (u - U)f - K\frac{\partial^2 v}{\partial z^2} = 0$$

where v is parallel to the direction of the geostrophic wind, u is perpendicular to the right of v, V is the forcing function or geostrophic wind assumed variable with time and height, U is assumed zero, and f is the coriolis parameter ($= 2\omega \sin(\text{lat})$)

The following gives the conditions and one solution for the Thermo Tidal flow.

MODEL OF THERMO–TIDAL WINDS

Forcing Function: $V(t, z) = \overline{V}(z) + V_1(z) \cos \omega t; \qquad U(t, z) = 0$

Response Function: $u(t, z) = \overline{u}(z) + u_1(z) \cos \omega t + u_2(z) \sin \omega t,$

$$v(t, z) = \overline{v}(z) + v_1(z) \cos \omega t + v_2(z) \sin \omega t$$

Boundary Conditions: $u(t, 0) = v(t, 0) = 0; \qquad V(t, 0) = \overline{V}_0 + V_{1,0} \cos \omega t$

$$u(t, \infty) = 0; \qquad v(t, \infty) = V(t, \infty)$$

OLUTION FOR HARMONICS OF FREQUENCY ω

ith $V_1 = V_{1,0} - z V_1'$, Lettau–MacKay (1965)

$$v_1 = (a + b) V_1 - V_{1,0} (a e^{-\alpha} \cos\alpha + b e^{-\beta} \cos \beta)$$

$$v_2 = V_{1,0} (a e^{-\alpha} \sin \alpha - b e^{-\beta} \sin \beta)$$

$$u_1 = - (\omega/f) v_2 - V_{1,0} (\tfrac{1}{2} e^{-\alpha} \sin\alpha + \tfrac{1}{2} e^{-\beta} \sin\beta)$$

$$u_2 = (\omega/f) v_1 - V_{1,0} (\tfrac{1}{2} e^{-\alpha} \cos\alpha - \tfrac{1}{2} e^{-\beta} \cos\beta),$$

where $a = \dfrac{f/2}{f - \omega}$, $b = \dfrac{f/2}{f + \omega}$; $\alpha = \eta/\sqrt{2a}$; $\beta = \eta/\sqrt{2b}$; $\eta = z\sqrt{f/2K_{Ek}}$.

EQATIONS 5

Nocturnal Low-level Jet on East-facing Slope at 42.5° N

24-hr. Means, and First Harmonics: Geostrophic System

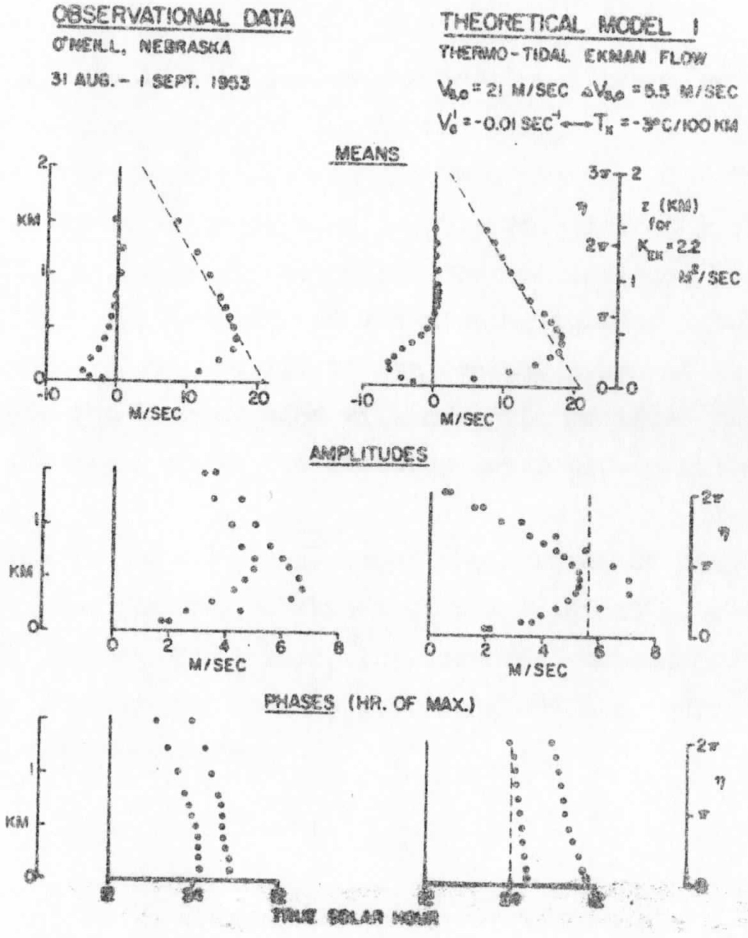


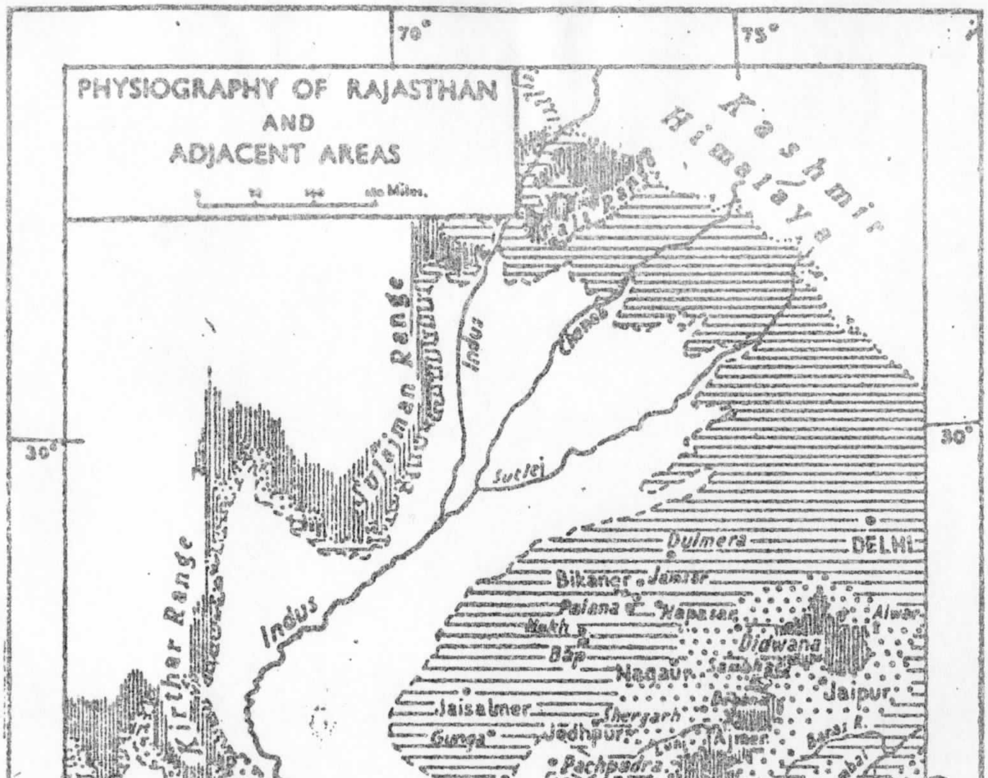
FIGURE 2

Fig. 2. show comparison between calculated and observed parameters for an east-west facing slope at 42.5 N. It can be seen that the main features of the flow are explained. Further refinements of the model are under way to allow for variations of K with time and height.

As stated above, one of the consequences of removing dust from the atmosphere would be to increase the diurnal range of temperature, and the diurnal variation of the thermal wind.

In order for the thermal wind to have its maximum effect on the low level winds, it must blow roughly parallel to geostrophic wind (the wind velocity at the height where frictional influence is negligable). The geometry of the sloping surface requires that the thermal wind blow parallel to the contour lines of the ground surface. Thus the thermal wind will have its maximum effect when the prevailing winds above the friction layer blow parallel to the contour lines.

Reference to Fig. 3 reveals that the surface of much of Rajasthan slopes towards the northwest with a gradient of roughly 1:1000. Thus during the summer, the southwest monsoon winds blow roughly parallel to the contour lines and the thermal wind will have its maxumum effect.



P. K. Ghosh, G.S.I.

Figure 3