The dry-line of Northern India and its role in cumulonimbus convection

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SUMMARY

An interesting feature of the flow pattern over India during the premonsoon months — a frontal surface separating a shallow, warm moist layer from a deeper, hot, very dry layer — is studied in some detail. In particular, its role in the development of severe cumulonimbus over north-east India is investigated. Routine observations are insufficient to resolve many of the features of this ‘dry-line’ and a simple dynamical model is constructed which suggests an interface in the form of a step with a height \( Z^* \) dependent upon the pressure gradient in the moist air \( A = \frac{1}{\rho} \frac{d\rho}{dx} \), surface drag coefficient \( C_D \) and Coriolis parameter \( f \), given by the expression \( Z^* = 4/15 A C_D f^2 \). An extension of the model to include a frictional stress variable with height suggests a circulation within the shallow moist layer which, together with solar heating, would likely be able to lead to cumulus and cumulonimbus development near the edge of the moist air; this is observed on satellite photographs.

1. INTRODUCTION

Cumulonimbus when accompanied by large hailstones, severe squalls or tornadoes are the most intense of meteorological phenomena, and yet they are one of the most difficult to forecast. The storms of north-east India are among the tallest and most severe anywhere in the world: heights of cumulonimbus tops over north-east India have been reliably measured and found to approach 20 km (Spavins 1970), and in 1969 one particularly severe thunderstorm accompanied by a tornado killed nearly 1,000 people and completely destroyed over 100,000 houses. Over north-east India and East Pakistan (now Bangladesh) in the two or three months before the onset of the monsoon widespread and severe thunderstorms present a serious forecasting problem: cumulonimbus convection of markedly different types, both in extent and severity, occur on days with closely similar synoptic patterns.

Ramaseswamy (1956) made a detailed study of storm situations over northern India considering both low- and upper-level flow patterns. He concluded that the occurrence of convective storms was related more closely to the 500 mb and 300 mb streamline patterns than to the sea-level flow patterns; most activity taking place in a region of upper divergence ahead of an upper-air trough or in the rear of a ridge, and, other things being equal, the stronger the jet stream aloft the more widespread and violent were the storms and squalls.

The present study describes the low-level synoptic flow pattern during the premonsoon months in a little more detail than previous studies, showing a definite structure, and suggests a mechanism by which cumulonimbus are initiated. On occasions of particularly severe thunderstorms, upper-level conditions are, of course, also important but will not be discussed in this paper.

2. LARGE-SCALE FLOW PATTERN DURING THE PREMONSOON PERIOD

During the premonsoon months (March to May) the flow pattern over northern India is a particularly interesting one. In the middle and upper troposphere a westerly jet stream lies south of the Tibetan Plateau while at low levels there is intense solar heating over the whole of India and an associated large-scale circulation: ascent over the Indian land-mass and descent over the surrounding seas. The general ascent over India is accompanied by a shallow inflow of warm moist air from the Bay of Bengal (sea-surface temperature about 27°C) all along the eastern coast (Fig. 1); this air has a depth of about a kilometre and high values of dew-point temperature (above 20°C). Over north-west and
Figure 1. Mean 500 mb (continuous) and surface (pecked) streamlines for April 1963. Wind discontinuities are shown by heavy pecked lines and land over 9,000 ft high is shaded.

Figure 2. The surface pressure (in mb) and streamline patterns at 1730 IST, 14 April 1969 over north-east India. Stations used in the analysis are indicated by dots. The position of Dacca is shown by the letter D.
north-central India the winds in the lower troposphere are from the arid regions to the west-north-west; consequently this air has properties markedly different from those of the moist low-level air farther east; it is hot and dry with dew-points as low as 0°C or even less, and has a lapse-rate near the dry-adiabatic up to about 3 km. Over north-east India the boundary between the dry north-westerlies and the moist southerlies or south-easterlies is striking because of their very different characteristics; we shall call this boundary the 'Indian dry-line'. The situation is similar to that of a large-scale sea-breeze front, the shallow moist layer undercutting the much deeper, hot, dry layer.

To examine the dry-line in more detail we look at the synoptic pattern on the day of the tornadic thunderstorm mentioned in the introduction. The position of the dry-line and the associated pressure pattern and streamlines of the surface wind are shown in Fig. 2; the tornadic storm was occurring at Dacca at the time of this chart. The pressure pattern is similar to that associated with middle latitude warm fronts, with the hot dry north-westerlies representing the warm air mass. It should be emphasized, however, that the dry-line is a shallow phenomenon unlike middle latitude fronts.

3. The horizontal gradient of dew-point

The dry-line, though conveniently depicted on large-scale synoptic charts as a line, is really a transition zone; however, on occasions this zone is very narrow with a difference of mixing ratio of at least 12g/kg within 12 km, a very large gradient indeed. One such occasion is given in Fig. 3, which shows the pattern of screen-level dew-point observations at 1730 IST

Figure 3. The distribution of dew-point temperature measured at screen-level at 1730 IST, 21 May 1959. The position of Calcutta (C) is shown.

(Indian Standard Time), 21 May 1959, less than an hour before a severe tornadic thunderstorm reached Calcutta. (A similar large gradient of screen-level dew-point was observed on 14 April 1969, Fig. 2, but less data were available to the author for that day.) Routine 'surface' observations show that the width of the zone varies from day to day but that for about half of the premonsoon months, in periods of 4 or 5 days at a time, the zone is narrower than the spacing of the observing stations; at other times the zone is rather broad (200 to 300 km). The position of the dry-line, for convenience regarded as the position of the 15°C isopleth of surface dew-point, appears to change little for several days at a time; however, there is some diurnal motion: in general, the moist air is farther west in the morning than in the evening. The magnitude of the diurnal motion of the dry-line is difficult to assess but appears to be between about 50 and 100 km.
4. A vertical section

The moist layer over north-east India is often little more than a kilometre deep, and above it the flow is westerly and the air is dry, being derived from the deep dry-adiabatic layer to the west. Fig. 4 is a vertical section lying approximately west-north-west to east-south-east across northern India and contains isopleths of potential temperature and mixing ratio for 0530 IST, 21 April 1963. The night inversion in the temperature soundings has been omitted by ignoring screen-level values of dry-bulb temperature. Radiosonde ascents from five stations in northern India have been used in the analysis together with screen-level dew-points (0830 IST) from other observing stations. Also included in the section are components of the horizontal wind in the plane of the section, determined by pilot balloon ascents at stations close to the line of the section.

The moist layer is seen on the right of the section as a region of higher mixing ratio and lower dry-bulb potential temperature, and so is cooler but 'potentially' warmer than the drier air to the west. The wind component of the moist layer in the plane of the section is small and from the east-south-east, that is towards the dry-line. (At low levels in the dry air mass the component of the horizontal wind in the section is also from the east-south-east but this does not extend above 500 m and may be a night-time effect.) Above the moist layer, the dry air has a component from the west-north-west. In this section the interface between dry and moist air has been shown as a sloping surface, on the basis of the soundings made at Dacca and Calcutta (Figs. 5 and 6) together with the position of the interface at the ground. The top of the moist layer is very well-defined at both Dacca and Calcutta, at heights of about 1 km and 250 m respectively, but the precise shape of the sloping interface is uncertain. Subsidence is regarded as playing no important role; the pronounced inversion results from the large difference of potential temperature of the two air masses.
To the east of the dry-line friction at the ground causes a drain of momentum from only the moist layer since mixing is likely to be small across the pronounced inversion at the top of the layer. The angle of cross-isobar flow of the 'surface' wind is large in the moist air mass, approaching 60°, but is less (about 30°) in the dry air mass where there is convective mixing (during sunshine) up to about 3 km and the drain of momentum is thus distributed over a much deeper layer leading to a smaller angle of cross-isobar flow.

5. A DYNAMICAL MODEL

To obtain an indication of the likely slope of the interface between the two airstreams we consider a simple dynamical model of the dry-line which includes friction, pressure gradients and Coriolis accelerations. The moist inflow is represented by a wedge of cool
air with the y-axis along the wedge; below the interface \((z_i(x, t))\) we assume that the pressure gradient in the y-direction \(\partial p/\partial y = 0\), but \(\partial p/\partial x \neq 0\) and is independent of height. We base these approximations on observation (e.g. Fig. 2). Above the interface we assume the air to be stationary and neglect any effect which its presence has on the motion. Further, we assume a stress at \(z = 0\) (the lower boundary) given by \(\tau_0 = \rho C_D v_0^2\) (\(\rho\), air density; \(C_D\), drag coefficient; \(v_0\), wind speed at \(z = 0\)), zero stress at the interface \(z = z_i\) and a linear change of stress with height. With these assumptions we expect to model all but the constant flux layer, occupying the lowest 10 m, say. A linear change of stress with height implies perfectly efficient mixing within the moist air (an infinite eddy transfer coefficient for momentum). The assumption of no mixing with the air above the interface is likely to be a good approximation during sunshine while convection is confined to the moist air and before cumulonimbus have developed. The model is summarized in Fig. 7.

![Figure 7. A summary of the model.](image)

We neglect accelerations of the fluid so that the horizontal momentum equation becomes:

\[
f k \wedge v + \frac{1}{\rho} \nabla p = \frac{1}{\rho} \frac{\partial \tau}{\partial z}
\]

where \(1/\rho \partial \tau/\partial z = (-C_D v_0 |v_0|)/z_i\) and \(k\) is a unit vector in the z-direction. Thus at all times we assume a balance between the friction force, Coriolis force and pressure gradient force (Fig. 8).

![Figure 8. The balance of forces for air parcels.](image)
First we note that since $\partial \tau / \partial z$ and $V_p$ are independent of height, if we assume $\rho$ is constant then $v$ is independent of height and equals $v_0$. Resolving along and perpendicularly to the direction of the wind we get

$$A \sin \alpha = \frac{C_D v^2}{z_I}$$

$$A \cos \alpha = f u$$

where $\alpha$ is the angle of cross-isobar flow, measured in an anticlockwise direction, and $A = 1/\rho \partial p/\partial x$. Eqs. (1) and (2) can readily be solved for $v$ and for $z_I < 5 \text{ km}$ we can expand the solution to obtain

$$v^2 = \frac{A z_I}{C_D} - \frac{1}{2} \left( \frac{f z_I}{C_D} \right)^2 + O(f^3).$$

To find the change in shape of the interface with time we consider the continuity of mass in a column. Thus

$$w_t = -\frac{\partial}{\partial x} (z_I v \sin \alpha)$$

where $w_t$ is the vertical velocity of the interface. But $z_I v \sin \alpha$ is a known function of $z_I$ alone and hence

$$w_t \equiv \left( \frac{\partial z_I}{\partial t} \right)_x = F(z_I) \left( \frac{\partial z_I}{\partial x} \right)_t$$

where

$$F(z_I) = \frac{d}{dz_I} (z_I v \sin \alpha).$$

Again the expression for $F(z_I)$ can be expanded for $z_I < 5 \text{ km}$ to give

$$F(z_I) = \left( \frac{A}{C_D} \right)^4 \left( \frac{3}{2} z_I^4 - \frac{15}{8} \frac{f^2 z_I^4}{A C_D} \right) + O(f^3).$$

Eq. (4) is a highly non-linear equation whose general solution is unknown, but particular solutions can be found using the "method of characteristics" (Landau and Lifshitz 1959) as follows

$$z_I = z_I(x, t)$$

so

$$dz_I = \left( \frac{\partial z_I}{\partial x} \right)_t dx + \left( \frac{\partial z_I}{\partial t} \right)_x dt$$

and, following isopleths of $z_I$,

$$\left( \frac{\partial x}{\partial t} \right)_{z_I} = -\left( \frac{\partial z_I}{\partial x} \right)_t \left( \frac{\partial z_I}{\partial t} \right)_x.$$

Using Eq. (4) we obtain

$$\left( \frac{\partial x}{\partial t} \right)_{z_I} = -F(z_I)$$

$(\partial x/\partial t)_{z_I}$ is the speed in the $x$-direction of isopleths of $z_I$. The variation of $F(z_I)$ with $z_I$ (using the complete solution) is shown in Fig. 9 for various values of $A$. There will be a steepening of the slope of the interface for values of $z_I$ where $\partial / \partial z_I [-(\partial x/\partial t)_{z_I}]$ is positive and a flattening for values of $z_I$ where this quantity is negative. This means that our wedge of cool air will tend to form into a 'step' of height $z^*$, the value of $z_I$ for which $-(\partial x/\partial t)_{z_I}$ is a maximum. The height of the step as a function of time is shown in Fig. 10 for an interface with an initial slope of 1/100, a drag coefficient of $6 \times 10^{-3}$ (typical of rather open country), a Coriolis parameter of $5 \times 10^{-3} \text{ s}^{-1}$ and the range of values of pressure gradient indicated. $F(z_I) = -(\partial x/\partial t)_{z_I}$ is a maximum with respect to $z_I$ at $z_I \approx 4/15 AC_D f^2$. This indicates the dependence of the equilibrium step height on the pressure gradient, drag
Figure 9. The variation of $(\partial x/\partial t)_H$ with $z_t$ for six values of pressure gradient (in mb/100 km) and the values of $f$ and $C_d$ of $5 \times 10^{-3}$ s$^{-1}$ and $6 \times 10^{-3}$ respectively.

Figure 10. The height of the step as a function of time for two values of pressure gradient.

coefficient and Coriolis parameter. For a pressure gradient of 1 mb/100 km, about the observed value on 14 April 1969 (Fig. 2), a step height of about 1 km is indicated, a typical depth of the moist layer. Substituting for $z_t$ in Eqs. (3) and (1) we find

$$v^2 = \frac{A^2}{f^2} \left( 1 - \frac{2}{15} + \ldots \right)$$

$$\sin \alpha = 1 - \frac{2}{15} + \ldots$$

giving a wind speed about half of the geostrophic value and an angle of cross-isobar flow of about 60° which agrees well with observations. Note that the ratio $v/v_g$ (where $v_g$ is the geostrophic wind speed) and $\alpha$ are independent of all the parameters of the motion.

We note that the flux of moist air into north-east India across the coast is proportional to $A^2$ since both the depth of the moist layer and the inflow speed are proportional to the pressure gradient. This does not necessarily mean that large pressure gradients are favourable for the development of severe storms since other factors are important.

The assumption that fluid accelerations $(Dv/Dt)$ are negligible is not always valid at the developing step face where accelerations are of the same magnitude as the other terms in the momentum equation while the step is small; however, these accelerations become small
(＜10 per cent of the other terms in the momentum equation) before the step has reached half its equilibrium height \((z_t^*)\) and should not therefore invalidate the solution.

The dry-line phenomenon over north-east India is basically a large-scale sea-breeze front in which the slopes are reported to be about 45° (Simpson 1969), compared with a slope of about 1° for a frontal surface of the middle latitude type. This is not inconsistent with the dry-line model which indicates a steep interface. It is salutory to note that \(f\), usually ignored in sea-breeze statics, is here seen to have a decisive effect on the structure: if we put \(f = 0\) the solution becomes one of a developing step with the height increasing indefinitely (as \(t^2\)); it is the inclusion of the Coriolis acceleration which gives rise to a characteristic depth \(z_t^*\).

![Figure 11. The dry-line in the form of a step.](image)

We have assumed in the model that the air above the interface is stationary and have deduced the speed of advance of the moist air relative to the dry air mass (given by \(u(z_t^*)\)). At the expense of confusing the friction law a little we can assume that the treatment is valid for axes moving with the wind speed at low levels within the dry air mass and deduce that this would reduce the speed of advance of the moist air by the same value, normally about 5 m s\(^{-1}\).

The form of the variation of \(\tau\) with height which we have employed, that of constant \(\partial\tau/\partial z\), is deficient in that in practice \(\partial\tau/\partial z\) is larger near the ground than higher up. We consider a dry-line in the form of a step of height \(z_t^*\) in a co-ordinate system moving in the negative \(x\)-direction with the speed, in the \(x\)-direction, of the moist air and assume that there is no flow through the boundaries \(x = 0\) and \(z = z_t^*\) (see Fig. 11). We are interested in the transverse circulation which results when \(\partial\tau/\partial z\) is not constant with height. If we call this perturbation speed in the \(x\)-direction \(u'\) then we find

\[
u' = -\frac{1}{\rho f} \left( \frac{\partial \tau}{\partial z} - \frac{\partial \tau}{\partial z} \right)
\]

where

\[
\frac{\partial \tau}{\partial z} = -\frac{\tau_0}{z_t^*} \text{ since } \tau_{t_0} = 0.
\]

So where \(|\partial\tau/\partial z| > |\partial\tau/\partial z|\) (i.e. at lower levels) there is a flow towards the surface dry-line and where \(|\partial\tau/\partial z| < |\partial\tau/\partial z|\) there is a flow away from the surface dry-line, implying a circulation in the moist air of the form shown in Fig. 12. We can expect this transverse circulation to exist in addition to any mean motion of the moist layer, and such a circulation in the moist air is suggested by the cross-section shown in Fig. 4.

![Figure 12. (a) The profile of stress with height. (b) The transverse circulation in the moist air.](image)
A transverse circulation has obvious importance in the maintenance of a 'sharp' dry-line: since very large gradients of surface dew-point are observed a local circulation is implied producing convergence near the ground, opposing horizontal diffusion due to convection.

A transverse circulation would play a more important role in producing favourable conditions for convection to begin near the dry-line. During sunshine one would expect air in the lowest 100 m to be about 2 or 3°C warmer and have a mixing ratio of about 3 or 4 g/kg more than air at 500 m under conditions found near the dry-line. The transverse frictionally-driven circulation described above would reinforce local convection near the edge of the moist air by lifting this warmer, moister air near the ground in the ascending branch of the circulation. ESSA satellite pictures for the premonsoon months show, on many occasions, cumulus and cumulonimbus clouds in a narrow band along a considerable length of the dry-line; Fig. 13 is typical. The photograph was taken at about local noon and shows cumulus and cumulonimbus along a line over northern India near 80°E and stretching from the Himalayan foothills to near latitude 20°N. This line coincides almost precisely with the position of the dry-line on the same evening, implying that near the dry-line conditions are particularly favourable for cumulus and cumulonimbus convection.

Although the shape of the interface approximates to a step, farther east of the dry-line the interface slopes gently upwards (at less than 1 in 100) giving greater depths of the moist layer there. This results in a more rapid warming of the moist layer near the surface dry-line than well to the east: consider a sounding made at Calcutta (Fig. 6) which shows the shallow layer of moist air with the overlying deep layer of air of much higher potential temperature. We are interested in the modification, by local convection, of the lower troposphere, leading eventually to cumulonimbus convection. An upper bound to the sensible heat input required before deep convection can begin locally can be estimated by assuming that convection will be dry up to the base of the inversion and that deep convection will begin shortly after \( \theta \) in the local convective layer reaches the value at the top of the inversion. The mixing ratio will also be assumed constant throughout this layer of local dry convection, i.e. we assume efficient mixing within the moist layer. We must consider the virtual temperature profile, in view of the large water content of the lower layer, and this is included in Fig. 6 with the shaded area corresponding to the required sensible heat input. We anticipate that a further small sensible heat input will cause a relatively large increase in the height of the convective layer and that cloudy convection will ensue.

To determine the sensible heat input discussed above we calculate:

\[
Q = \frac{c_p}{g} \int \Delta T \, dp
\]

where \( \Delta T \) is the increase in temperature of a layer of thickness \( dp \) and the other symbols have their usual meaning. From such a calculation we find that 45 cal cm\(^{-2}\) is required, a rather small value which, in the absence of large amounts of cloud would be surpassed before local midday, despite the large fraction of the energy input used for evaporation. In practice we expect some mixing down of the potentially warm air above the inversion by thermals which rise beyond their equilibrium level (Ball 1960). Although this process represents a heat transfer through the inversion, the calculation of the sensible heat input is unaffected provided the layer above the inversion has a dry adiabatic lapse rate. (This means that any air mixed down ends up at its original potential temperature, and so does not affect the sensible heat computation.)

At Dacca (for position see Fig. 2), which is farther from the surface dry-line, the sensible heat input required to raise \( \theta \) in the moist surface layer to the value at the base of the inversion is about 160 cal cm\(^{-2}\), which represents perhaps more than a whole day's sensible heat input (taking into account the infra-red loss from the moist layer, about 0.15 cal cm\(^{-2}\) min\(^{-1}\), reflection of solar radiation from clouds and ground, the absorption of the atmosphere above and the evaporation of water at the surface). We can thus expect
penetration of the inversion to occur first near the surface dry-line, where the moist layer is relatively shallow, especially in view of the local vertical motion there (the transverse circulation). Once convection has penetrated the inversion, the cumulus can develop into cumulonimbus which move eastwards, with the moist layer feeding the updraught and the dry layer of low $\theta_e$ feeding the downdraught. This appears to have been the sequence of events on 21 April 1963 (the day for which the vertical section was constructed). Fig. 14 shows the position of the dry-line and the subsequent 24-hour rainfall pattern. The latter is consistent with shower formation near the dry-line and a subsequent eastward movement and development, with over 2 cm of precipitation along the coastal region of East Pakistan near Chittagong.

![Figure 14. Isopleths of rainfall (in cm) for the 24 hours ending at 0830 IST, 22 April 1963. The position of the surface dry-line (taken as the position of the 15°C isopleth of screen-level dew-point) at 0830 IST, 21 April 1963 is shown as a heavy pecked line.](image)

6. **Comparison with the dry-line of the United States**

Under suitable large-scale conditions a dry-line or dew-point front is found over the middle-west states of Texas, Oklahoma, Kansas and Nebraska. Severe thunderstorms often begin near the dry-line, but the mechanism of their initiation is not well understood. The dry-line marks the boundary between a deep (about 4 km) dry air mass flowing from the south-west (across the arid western states) from the Pacific, and a moist airstream (about 2 km deep) flowing northwards from the Gulf of Mexico. Small-scale convection in the moist layer is suppressed by a potentially warm layer above flowing from the Mexican Plateau. The position of the dry-line is controlled largely by topographical features: the dry-line is usually found in the lee of the northern highlands of Mexico and may remain quasi-stationary for several days. Although very large gradients of moisture content occur near the dry-line, very little density contrast exists (Fig. 15). The interface is normally very steep or even vertical at low levels but a tilting to the east is usual at higher levels, as is the case over north-east India.

With the Indian dry-line, though topography must play some role, no obvious control is evident since the line is often found over the Gangetic Plain where even low hills are absent. Another important difference is in the density difference across the line; over north-east India, at least before solar heating begins, the virtual temperature in the moist layer is about 5°C lower than in the dry layer.
The relationship predicted by the model between the depth of the moist layer, the pressure gradient in the moist air, the drag coefficient and the Coriolis parameter appears not to hold closely for the United States dry-line. Values of the parameters estimated from observational data on three days on which the dry-line was well-defined are consistent only if the value of the drag coefficient is set at about $2.5 \times 10^{-2}$, probably too large by a factor of three or four. This discrepancy may mean that the depth of the moist layer is determined by topographic features to some extent, or that the observational data used are unrepresentative (e.g. due to topography).

7. FURTHER OBSERVATIONS

The Indian dry-line is not routinely recognized and marked on surface weather charts except as a pressure trough even though it is among the most significant features of the flow pattern over north-east India during the premonsoon months. The dry-line and its structure should be recognized as important as an indication of severe cumulonimbus development. A well-defined dry-line is not always apparent and on such days topography probably plays an important role in cumulonimbus development, but days on which intense convection occurs are characterized by a narrow transition zone with a position well to the east of the mean for the time of year. Further investigation of the precise role of the dry-line in cumulonimbus development is important. More detailed observations are required, but a rather simple observational programme would be sufficient to provide useful information and to test some of the results of the model. In particular, the time development of the transverse circulation and its relationship to the development of cumulonimbus convection near the edge of the moist air, and the predicted dependence of the depth of the moist air on the pressure gradient could readily be investigated. India has a good network of pilot balloon stations - over 50 throughout the country - and over north-east India six are particularly well-positioned for studies of the dry-line, namely:
INDIAN DRY-LINE

42498 Bhagalpur  42704 Asansol  42809 Calcutta/Dum Dum
42591 Gaya  42798 Jamshedpur  42886 Jharsuguda

When the dry-line lies over this area a programme involving hourly pilot balloon observations from these six stations, together with screen-level observations of temperature, dew-point, pressure and wind at other stations throughout the area would provide valuable data. By making use of satellite pictures and visual observations of cumulus development, especially if supplemented by reconnaissance with a light aircraft, one could hope to determine the time development of the circulation within the moist air, its relation to cumulus development and the motion of the dry-line.

A complete understanding of the role of the dry-line in cumulonimbus development should greatly help forecasting over north-east India.

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