551·515·4 : 551·54I

Pressure fluctuations associated with thunderstorms*

G. C. ASNANI

Meteorological Office, Poona

ABSTRACT. Based on certain plausible assumptions, a formula is derived to give a quantitative estimate of quasi-static perturbation pressure associated with the thunderstorm "high". This shows a theoretical possibility although not a theoretical necessity of a severe hailstorm being accompanied by a pressure "dip". A few actual cases are quoted in qualitative support of this theoretical result.

1. Introduction

It is well known that the atmospheric pressure as registered by a barograph at the ground undergoes rapid fluctuations in the neighbourhood of a thunderstorm. For example, during the growing stage of a thundercloud, there is a feeble pressure "low" in the vicinity of the cloud. During the precipitating stage, there is a relatively strong pressure "high". There have been numerous qualitative discussions of these pressure fluctuations in the past, particularly in respect of the thunderstorm "high", in the building of which several static and dynamic factors have been supposed to play a part. In this paper, we present detailed *quantitative* estimates of different contributions to this high, taking into account temperature fluctuations and also changes in the quantity of raindrops and hailstones. The computed values are found to be in broad agreement with observations. One interesting result following from these quantitative estimates is the possibility of severe hailstorms being accompanied by a fall of pressure. This theoretical result is shown to be supported by some barographic records analysed by the author.

2. Computation for thunderstorm high

The computations are presented in the following paragraphs.

2.1. *Specification of initial conditions*

It is well known that during the growing stage of a convective cloud, hydrometeors are held inside it by updrafts. With passage of time, the size and the amount of hydrometeors increase to a stage where these cannot be supported by the updrafts. Then the hydrometeors descend relative to the ground. This descent initiates cooling of the cloud air, reduction in the intensity of updrafts and the gradual replacement of updrafts by downdrafts. As an idealised case, we shall define the instant of time $t=0$ as the moment when the hydrometeors start descending relative to the ground. We shall futher make the following assumptions about the physical conditions at $t=0$.

(*i*) Under the combined influence of buoyancy and frictional drag, the hydrometeors are moving with their respective "terminal" velocities relative to their surrounding air inside the cloud. This assumption appears to be justified in view of the short interval of time taken by hydrometeors to attain their respective terminal velocities in the atmosphere.

(*ii*) The air of the updrafts inside the cloud is assumed to be moving with a "limiting" velocity relative to the surrounding air under the combined influence of buoyancy and frictional drag. This limiting velocity of the updrafts is analogous to the terminal velocity of the hydrometeors. This assumption of limiting velocity appears to be justified by the findings of Scorer and Ludlam (1953), who through semi-empirical and semi-theoretical reasoning came to the important conclusion that in a convective cloud, an air

*Subject matter of this paper formed part of Ph.D. thesis submitted by the author to the University of Bombay in February 1959

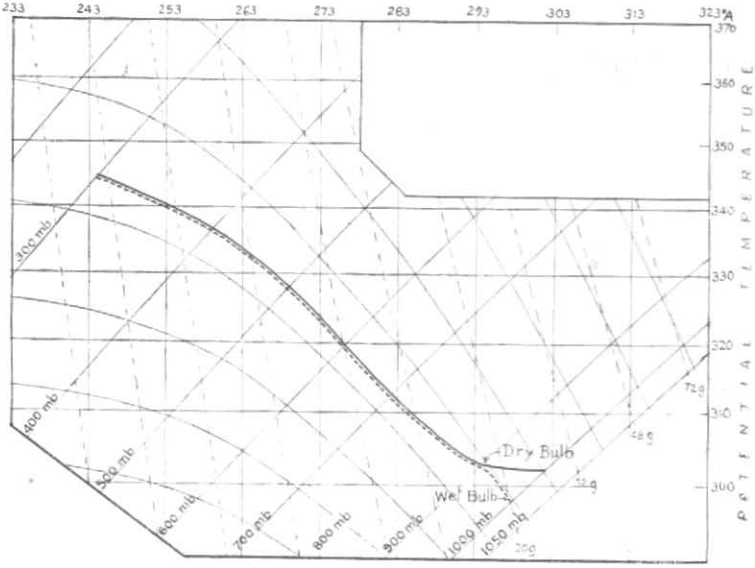


Fig. 1

bubble rising up under the combined influence of buoyancy and frictional drag, has a constant velocity over nearly all its life.

(*iii*) Simple consideration of continuity suggests that for compensation of updrafts inside the cloud, there must be a downdraft in the air outside the cloud. It is assumed here that in the horizontal plane, the area of the compensating downdraft is very much more than that of the updraft so that the air outside the cloud can be treated as practically at rest along the vertical.

The combined effect of these three assumptions is the reduction of the pressure problem at $t=0$ to one of hydrostatics, *i.e.*, the pressure at the ground surface is given by the weight of air *and* of the hydrometeors inside the vertical column of unit horizontal cross-section, the column extending from the ground to the top of the atmosphere. Other specifications of the initial conditions are:

(*a*) Base of cloud is at 900 mb, (*b*) Top of cloud is at 300 mb, and (*c*) Pressure-temperature distribution inside the cloud and below its base is as shown in Fig. 1.

For our purposes, perturbation pressure at time t is denoted by $\triangle p_0$ and it is the change of barometric pressure since the instant of time $t=0$. It is assumed that we have, by suitable techniques, eliminated pressure tendency arising out of changes in the synoptic system or out of the migration of diurnal and semi-diurnal pressure waves. Also, the cloud of our problem is supposed to consist of a single cell far removed from other clouds such that the pressure effects of other clouds become negligible in the neighbourhood of our cloud.

2·2. *Perturbation pressure arising out of dynamic factors*

The perturbation pressure arises from two sets of factors, quasi-static and dynamic. There appear to be two principal dynamic factors:

(*i*) Vertical retardation of the downdraft due to the ground acting as an obstacle in the vertical flow; and (*ii*) Horizontal acceleration and retardation due to buildings, trees and other ground structures acting as obstacles in the horizontal flow.

Of these two dynamic factors, the first has been the subject of several discussions and even of some theoretical controversy (*Cf.* Schaffer 1947, Mull and Rao 1950). While admitting that the observational material available upto date is inadequate to solve this controversy, we would like to treat this problem of contribution of vertical accelerations by dividing the entire vertical column from cloud top to the ground into two parts. We assume that in the upper part with its base at say p_1 mb, the individual downdrafts are more or less having "limiting" velocities under the combined operation of buoyancy and frictional drag, hence their pressure effect would be taken into account through quasi-static computations. In the lower part also, *i.e.*, below the level p_1, the individual downdrafts would have similar "limiting" velocities and would thus be amenable to hydrostatic treatment but for the fact that the ground acts as a barrier. Hence in this lower layer, there is, apart from the hydrostatic pressure perturbation, an extra perturbation pressure, the value of which near the ground is approximately given by $\frac{1}{2}\sigma w_1^2$ where σ is the mean density of air between the level p_1 and the ground; w_1 is the magnitude of downward vertical velocity at the level p_1. For determining the values of p_1, and w_1, we have the relevant data of the Thunderstorm Project (1949) which, though inadequate for a firm conclusion, do suggest that we shall not be far out in our computations if we take p_1 as 900 mb and w_1 to be of the order of 5 metres/sec. We then have $\sigma \doteqdot 1050 \times 10^{-6}$ gm cm^{-3} and $\frac{1}{2}\sigma w_1^2 = 0{\cdot}13$ mb. Even if w_1 were 10 metres/sec, which value seems to occur perhaps only rarely, the value of $\frac{1}{2}\sigma w_1^2$ would be only $0{\cdot}52$ mb. The bulk of perturbation pressure in thunderstorm high appears to be contributed by other factors.

Regarding the contribution made by the second dynamic factors, *viz.*, the horizontal acceleration and retardation due to the ground structures, we may state that the theoretical and the experimental work on aerofoils suggests that the perturbation pressure will vary from one group of obstructions to another. It is, therefore, difficult to make a correct estimate of these perturbations, although the formula

$$\text{perturbation pressure} = \tfrac{1}{2}\rho_0(v_\beta^2 - v_\alpha^2)$$

appears to give a useful approximation; here v_α is the actual horizontal velocity and v_β is the horizontal velocity which would have prevailed if the flow had been un-obstructed; ρ_0 is the air density near the ground.

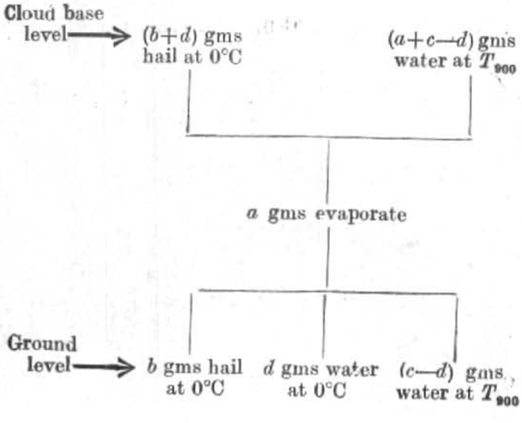
2·3. *Perturbation pressure arising out of quasi-static factors*

During their descent relative to the ground the hydrometeors absorb heat from the surrounding air which undergoes a series of thermodynamic changes. We make the following assumptions and specifications about the changes during the time interval t —

(*i*) Conditions above the cloud top remain unchanged.

(*ii*) Virtual temperature of the cloud column (300-900 mb) changes by an average amount $\triangle T$, the average being reckoned with reference to log p scale.

(*iii*) $(b+d)$ gms of hail at 0°C and $(a+c-d)$ gms of water at T_{900} (*i.e.*, temperature at 900-mb level) leave 1 sq. cm of the cloud base. These hydrometeors evaporate, melt or otherwise get heated according to the scheme shown below.



We admit that there is scope for some alteration in the assumed values of hydrometeor temperatures. But such possible alterations are insignificant for our computations and more so for our conclusions in this paper.

The perturbation pressure arising out of quasi-static factors will consist of contributions made by the following changes occurring during the time interval t—

(*i*) Change in the virtual temperature of the cloud column 300—900 mb; (*ii*) Change in the temperature of the column below the 900-mb level, (*iii*) Change in the humidity of column below the 900-mb level due to evaporation of a gms of water, and (*iv*) Loss and re-distribution of hydrometeors in the entire vertical column.

We shall denote contributions of these four factors by $\triangle_1 p_o$, $\triangle_2 p_o$, $\triangle_3 p_o$ and $\triangle_4 p_0$ respectively. Then

$$\triangle_s p_o \equiv \triangle_1 p_o + \triangle_2 p_o + \triangle_3 p_o + \triangle_4 p_o$$

2.3.1. *Value of* $\triangle_1 p_o$

As a result of the change $\triangle T$ in the virtual temperature, the thickness of the column 300-900 mb changes by $\left(\frac{R}{g}\triangle T \log_e \frac{900}{300}\right)$.

Since 300-mb level remains unchanged, the altitude of the 900-mb surface changes by $\left(-\frac{R}{g}\triangle T \log_e \frac{900}{300}\right)$. As an approximation, we may take this to be also the change in the altitude of the 1000-mb surface due to this factor alone. This approximation is equivalent to assuming that pressure-temperature distribution in the layer 900-1000 mb does not change due to the operation of this influence.

Then

$$\frac{\triangle_1 p_o}{1000} = -\frac{1}{T_{1000}} \triangle T \log_e \frac{900}{300} \text{ mb.}$$

$$\text{Or } \triangle_1 p_o = -\frac{1000}{T_{1000}} (1 \cdot 10 \triangle T) \text{ mb.}$$

2.3.2. *Value of* $\triangle_2 p_0$

The layer from the cloud base to the ground loses (583 a) calories of heat to allow a gms of water originally at T_{900} to get evaporated at approximately T_{950} and also loses another (80 d) calories of heat to allow d gms of hail originally at 0°C to get converted into an equal amount of water at 0°C. The heat capacity of this layer being about 25 calories, its temperature changes by

$$-(23 \cdot 3\, a + 3 \cdot 2d)\ ^\circ\text{C}$$

$$\therefore \triangle_2 p_o = \frac{1000}{T_{1000}} (2 \cdot 45\, a + 0 \cdot 34\, d) \text{ mb.}$$

2.3.3. *Value of* $\triangle_3 p_o$

$\triangle_3 p_o$ results from a loss of $(0 \cdot 6\, a)$ gms of mass at a mean level of 950 mb. Then

$$\triangle_3 p_o = -\frac{1000}{T_{1000}} \frac{T_{950}}{950} (0{\cdot}98)(0{\cdot}6\, a) \text{ mb.}$$

$$\text{Or } \triangle_3 p_o = -\frac{1000}{T_{1000}} (0 \cdot 18\, a) \text{ mb.}$$

2.3.4. *Value of* $\triangle_4 p_o$

During the time-interval t, among other things, the following two phenomena take place—

(*i*) The air column loses $(a+b+c)$ gms of suspended hydrometeors.

(*ii*) There is re-distribution of hydrometeor mass inside the vertical column. For example, at time $t=0$, there were no hydrometeors below the cloud base. At time t, this layer contains those hydrometeors which left the cloud base after time-instant $t=0$ but have not yet reached the ground. Also the larger and heavier hydrometeors have descended through greater height than the smaller and slower ones.

Thus, during the time-interval t, there has been loss of hydrometeor mass as well as downward displacement of the remaining hydrometeor mass inside the air column under examination. Under quasi-static conditions, both these factors have their influence on the barometric pressure at the ground. We could make a correct estimate of the pressure effect at the ground, if we knew two things—

(*i*) the mean level at which the loss of hydrometeor mass can be considered to have taken place, and

(*ii*) the resulting change in the pressure-temperature distribution from this level to the ground. Correct computation is, therefore, a fairly complicated issue and we are not in a position to make a precise estimate. However, we suggest a very approximate method of computation.

Elementary considerations suggest that the loss of hydrometeor mass takes place in the topmost layer of the cloud. However, the density of hydrometeor mass is known to be nearly the maximum in the vicinity of the freezing level. Hence the loss may be considered to take place in the uppermost parts of the cloud but nearer to the freezing level. With these considerations in view, we take the effect to be equivalent to the loss of $(a+b+c)$ gms of hydrometeor mass at 500-mb level, closer to but above the freezing level. We also assume that pressure-temperature distribution does not change below this level due to this influence.

Then,

$$\triangle_4 p_o = -\frac{1000}{T_{1000}} \frac{T_{500}}{500} (0\cdot98)(a+b+c) \text{ mb.}$$

Or

$$\triangle_4 p_o = -\frac{1000}{T_{1000}} (0\cdot54)(a+b+c) \text{ mb.}$$

3. Discussion

Combining the quasi-static and dynamic factors and taking $T_{1000} \doteq 300$, we get

$$\triangle p_o = (-3\cdot7\triangle T + 5\cdot8a - 1\cdot8b - 1\cdot8c + 1\cdot1d) + \tfrac{1}{2}\left\{\sigma w_1^2 + \rho_o(v_\beta^2 - v_\alpha^2)\right\}.$$

This theoretical formula cannot pretend to explain all the types of pressure fluctuations observed on the barographs. Our statement of assumptions at every stage emphasizes the limitations of our computations. However, we shall employ this formula for two purposes—

(*i*) to see if the order of magnitude of the computed perturbation pressure compares favourably with the observed magnitudes of thunderstorm pressures, and (*ii*) to study perturbation pressure type in the case of severe hailstorms.

3.1. *Magnitude of perturbation pressure*

For a moderate thunderstorm without hail, we may take

$a=0\cdot1$ gm, $b=0$, $c=3$gm (*i.e.*, 3 cm of rain),

$d=0$, $\triangle T=-2$°C, and $\tfrac{1}{2}\sigma w_1^2 = 0\cdot2$ mb.

The value of a is taken for the conditions shown in Fig. 1, where this much evaporation would suffice to produce, through evaporational cooling below the cloud, nearly saturated conditions in this layer. The value of $\triangle T$ has been taken, keeping in view the comments of Ludlam and Scorer (1953) on the estimates of $\triangle T$ made earlier by the Thunderstorm Project (1949). The value of $\tfrac{1}{2}\sigma \dot{w}_1^2$ as $0\cdot2$ mb seems adequate in view of what was stated in Section 2·2 above and also in view of the author's finding that a downdraft of 5 metres per second is quite capable of producing

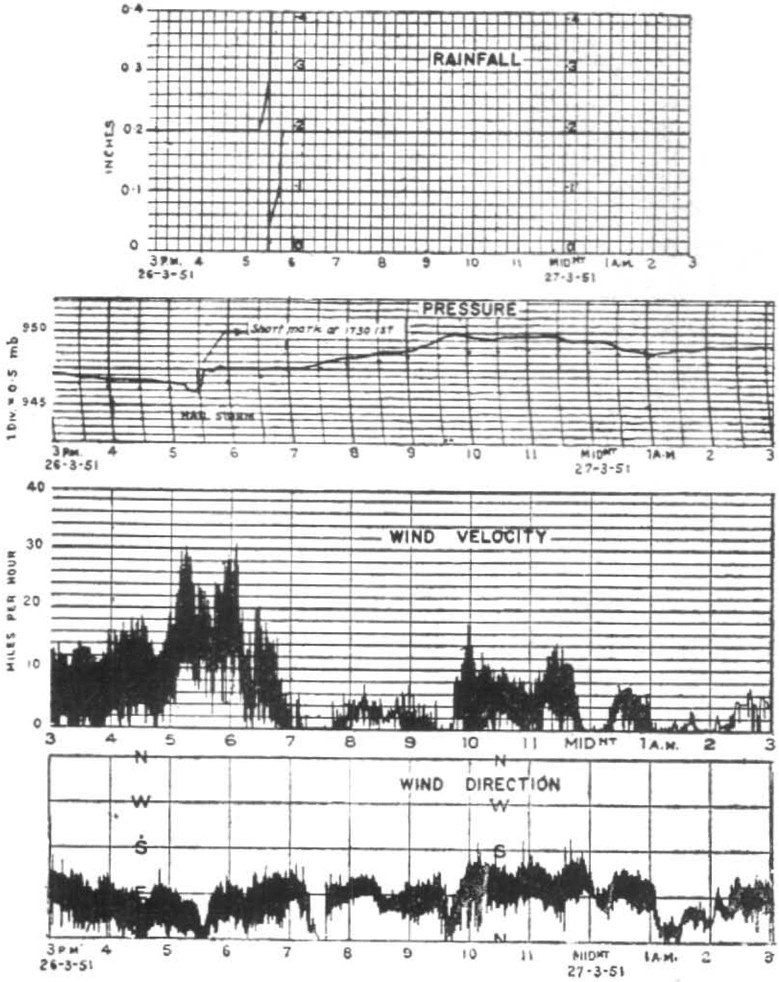


Fig. 2. Autographic records of Begumpet (Lat. 17°26′ N, Long. 78°27′E)

Between 1715 and 1745 IST there was a severe thunderstorm accompanied by rain and hail. The barograph traced a pressure dip.

(Reproduced from page 131, *Indian J. Met. Geophys.*, **9**, 1958) horizontal squall speeds of 50 kt or more. In individual severe thunderstorms, however, the value of $\frac{1}{2}\sigma w_1^2$ may be exceeding 0·2 mb. Regarding the contribution made by the other dynamic factor, we are not quite in a position to give a reliable estimate of

$$\left\{\tfrac{1}{2}\rho_\circ \; (v_\beta^{\;2} - v_\alpha^{\;2})\right\}.$$

It may be having a wide variation depending on the type of obstructing ground structures and the direction of squalls relative to the orientation of the structures. Assuming that the obstructions are either practically absent or that their contribution is negligible, we get from the above formula, $\triangle p_\circ = +2{\cdot}8$ mb which is a fairly reasonable value, as far as the order of magnitudes is concerned.

3.2. *Case of severe hailstorms*

When the falling hydrometeors are in the form of raindrops or only small hailstones, the time taken by them to reach the ground is favourable for appreciable heat flow from the air to the hydrometeors. This heat flow enables the hail to melt and also contributes towards the evaporation of water under the cloud. If, however, the falling hydrometeors be in the form of large hailstones as in the case of severe hailstorms, their terminal velocity would be large and hence the time of contact between the hail and the surrounding air would be relatively very short. Thus the fall of a large hailstone may be accompanied by little exchange of heat between the hydrometeor and the environment and hence by little change in the temperature of air inside and outside the cloud. In other words, for the first few minutes of large hailstones reaching the ground, the quasi-static perturbation pressure may be controlled by the term $-1{\cdot}8\,b$. Then, for every 1 gm of hailstones per sq. cm, the barograph would register a fall of 1·8 mb. Obviously, this condition cannot last for long. Soon after the release of large hailstones from the air column, smaller hailstones and the water drops would cool the air and resume the normal control of quasi-static pressure as shown in Section 3·1 and would make positive contributions to the value of $\triangle p_\circ$. If such a process takes place in the atmosphere, as it possibly can, we should expect the fall of large hailstones from a thundercloud to be accompanied first by a fall and then by a rise of pressure, *i.e.*, a sort of "dip" in the barograph trace.

It may be emphasized that this dip in the barograph trace is a possibility and not a necessity in the case of severe hailstorms. For, the nearly idealised conditions required for the control of $\triangle p_0$ by the term $-1{\cdot}8\,b$ would not always occur in the atmosphere. Quite often, the other terms in the formula for $\triangle p_0$ seem to be having overwhelming influence at the time when large hailstones are being released from the air column. The negative contribution then made by the term $-1{\cdot}8\,b$ gets obscured on a barograph trace.

This theoretical possibility of hailstorm being accompanied by a pressure dip is supported by some of the actual barographic records analysed by the author. For the sake of illustration and completeness, one of the three such published (Asnani 1958) records is shown in Fig. 2. Reference may also be made to another barographic 'dip' described in the *Indian J. Met. Geophys.*, 1957, **8**, pp. 333-334.

4. Concluding remarks

The author is quite aware of the limiting assumptions under which the quantitative estimate of thunderstorm pressure "high" has been made in this paper. In spite of these limitations, two encouraging results follow—

(*i*) Order of magnitude of the perturbation pressure computed in this paper compares well with the actually observed magnitudes.

(*ii*) There is a theoretical possibility, though not a necessity, that severe hailstorms may sometimes be accompanied by pressure dips. This is supported by some actual barographic records analysed by the author.

5. Acknowledgements

The author is grateful to Dr. P. R. Pisharoty for his guidance in this work. Sincere thanks are also due to the Director General of Observatories, New Delhi who gave all facilities for research, including study leave, for completing the Ph.D. thesis.

REFERENCES

Asnani, G. C.	1958	*Indian J. Met. Geophys.*, **9**, 2, pp. 117-132.
Byers, H. R. and Braham, R. R.	1949	*The Thunderstorm*, Thunderstorm Project, Washington
Ludlam, F. H. and Scorer, R. S.	1953	*Quart. J. R. met. Soc.*, **79**, pp. 317-341.
Mull, S. and Rao Y. P.	1950	*Indian J. Met. Geophys.*, **1**, 2, pp. 116-136.
Schaffer, W.	1947	*Bull. Amer. met. Soc.*, **28**, pp. 351-355.
Scorer, R. S. and Ludlam, F. H.	1953	*Quart. J. R. met. Soc.*, **79**, pp. 94-103.