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West coast rainfall and convective instability

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ABSTRACT
The paper examines the role of convective instability in producing spells of heavy rain at the coastal location on the windward side of the Western Ghats (WG) over the Indian west coast. For this purpose, cases have been selected, which include, days when isolated heavy rains occurred at Mumbai (Santacruz), with preceding and succeeding days reporting comparatively less rainfall. For each of these epochs, consisting of three consecutive days, we have computed Convective available potential energy (CAPE), Convective inhibition energy (CINE), Moist static energy (MSE), a non-dimensional kinetic parameter ('K'), vertically weighted average value of MSE (σ) etc., to obtain their daily variations. A simple convective precipitation model has been developed to obtain rainfall on these days. The model has two parts. In the first part convective updraft (w) at each level, solely due to the buoyancy effect, has been computed. In the second, this 'w' is used to compute convective precipitation rate using the physical model (Sarker 1966) for precipitation intensity. From the study it appears that a rise in rainfall from the first day to second day is generally associated with a corresponding rise in the positive value of CAPE and the parameter 'K', a fall in 'σ' and a fall in the negative value of CINE. While a fall in rainfall on the third day is associated with a corresponding fall in the positive value of CAPE. The study also shows that the proposed convective precipitation model, in general, is capable, at least qualitatively, of capturing the observed fluctuation in daily rainfall.

INTRODUCTION

It has been long known that mountains play an important role in the development of deep convection. First, mountains act as an elevated heat source that makes the air at mountain summits unstable, favouring the formation of convective cells. Moreover, in response to sustained heating at high elevations, horizontal pressure gradients are developed with lower pressure over the heated mountains than air at the same level, some distance away from the peak. This induces air to flow up the heated slopes and converge at the tops in a heated chimney, which provides moisture convergence as well.

Bonacina (1945) surveyed the regions of intense orographic rain around the world and the possible mechanisms. He concluded that orographic rain does not occur every time an air stream impinges on a mountain, but rather when air stream must has been conditioned by the prevailing synoptic situation. In particular he emphasized the importance of convective instability for the generation of intense orographic rain.

Mukherjee & Ghosh (1965) suggested a possible role played by the Khasi-Jayantia (KJ) hills in triggering pre-monsoon convective instability over Brahmaputra valley at night. They argued that during night only, when the KJ hills cool down more rapidly than the plane, strong katabatic wind blows down the slope towards the valley, which is being obstructed at least up to 1.5km at daytime. This in turn brings down southerly/south westerly moist air on the valley, where already easterly/east northeasterly dry air originating from Tibetan plateau prevails. So, a front like structure develops over the valley at night, which causes convective activity to take place at night.

Studies by Sarker (1966,1967) and De (1973), have indicated that in addition to forced orographic ascent, synoptic scale convergence, convective instability are also very important factors for heavy to very heavy precipitation rate on the windward slope.

From the study of Browning, Hill & Pardoe (1974) it appears that convective cells and meso-scale precipitation areas may be accentuated by orography. They are persistent over six hours or more and account for much of heavy precipitation.

Mukherjee & Ramanamurthy (1978) studied the contrasting rainfall features and associated thermodynamic behavior over Mumbai, lying on the windward side of the coastal mountain Western Ghats

(WG) along west coast of India. They found that higher rainfall days are associated with higher static stability, less convective instability and higher precipiTable water content (PWC). They also found that differences in static stability and PWC between consecutive days of contrasting rainfall were not significant.

Smith (1979) suggested that convection triggered by smooth orographic ascent brings the air to saturation and after some delay, raindrops form and fall to the ground. He also suggested that heating of the mountain slopes by insolation causes upslope winds leading to thermals above the mountain peak.

Grossman & Duran (1984) studied the influence of the Western Ghats, in India, in the formation of deep convective cells away from the mountain crest on the windward side. In this study the discrepancies between the observed coastal rainfall and the coastal rainfall computed using 2-D dynamical model (Sarker1966, 1967) was attributed to the frequent shallow and occasional deep convection near the coast. Watnabe & Ogura (1987) made a case study of heavy precipitation in the western parts of Japan and showed that during the heavy rainfall period convective cells form in succession over the sea about 50 km off the coast. As they moved eastward and approached the coastline, they developed rapidly and organized into a band structure. Further Watnabe & Ogura (1987) show that a mountain range of modest height could trigger the convective instability to enhance rainfall significantly along windward slope. Recent studies by Dutta & Rase (2004) have examined the nature of precipitation in the different regions of India including the Western Ghats.

Cotton (1988) studied the effect of the Colorado Rocky on the formation of mesoscale convective systems (MCS) to its lee and showed that heating of mountain slopes induces mountain-plane solenoidal circulations responsible for organizing convection on a broad range of scales. These shallow solenoidal circulations are strengthened and deepened by the systematic latent heating by the clusters of Cumulonimbus.

Dutta & De (1999) studied the convective instability during contrasting rainfall epochs over Mumbai. It was shown that in most cases the days with higher precipitation were associated with higher values of CAPE as compared to that on the days with lower rainfall.

Quite often it is observed that the coastal station Santacruz (SCz) in Mumbai on the windward side of the WG receives heavy rainfall on an isolated day, with preceding and succeeding days of comparatively low rainfall or dry weather. Dynamical models developed by Sarker (1966,1967), De (1973), SinhaRay (1988) for orographic rainfall, did not explain a significant part of such observed heavy rainfall on isolated days. So, in such cases forced orographic lifting alone is not the only mechanism responsible for such observed heavy rainfall but convective instability may play a crucial role in addition with synoptic scale convergence. But for convective instability to be triggered, a certain lifting of air parcel up to level of free convection (LFC) is required, which may be attained by orographic lifting, So it appears from the above that an attempt to study the role of convective instability in enhancing orographic rainfall rate is important.

The objective of the study in this paper is to show; at least qualitatively, that the isolated heavy rainfall on a day, preceded and succeeded by days with comparatively very less rainfall, is mainly due to the release of convective instability. For that, first we want to show that convective instability significantly varies in association with the fluctuation of daily rainfall in a series of three consecutive days and then to develop a dynamical model for convective precipitation rate.

DATA

Daily rainfall data of Santacruz (19°7′N, 72°51′E)(here after SCZ) on three consecutive days for the cases under study (shown in Table 1), were collected from National data centre, India Meteorological Department, Pune, India. Relevant radio-sonde data of SCz were also collected from same source.

Table-1. Daily rainfall data of Santacruz for five different cases

S.No.	Date	24 hour Rainfall (mm)
1	21.6.82	0.8
	22.6.82	135.8
	23.6.82	0.7
2	1.7.84	38.2
	2.7.84	544.3
	3.7.84	47.6
3	16.6.85	38.3
	17.6.85	345.5
	18.6.85	4.6
4	30.6.87	15.7
	1.7.87	153.9
	2.7.87	59.0
5	16.6.89	15.0
	17.6.89	111.9
	18.6.89	4.3

METHODOLOGY

Thermodynamic parameters related to rainfall:

Convective available potential energy (CAPE): CAPE has been computed using the formula

$$CAPE = \int_{Z_{LFC}}^{Z_{LNB}} \frac{g(T_{vp} - T_{ve})}{T_{ve}} dz = \int_{P_{LFC}}^{P_{LNB}} \frac{-R(T_{vp} - T_{ve})}{P} dp \quad \text{.... (1),}$$

where,

P_{LFC} = Pressure at the level of free convection (LFC).
P_{LNB} = Pressure at the level of neutral buoyancy (LNB).
Z_{LFC} = Height of the LFC above the mean sea level.
Z_{LNB} = Height of the LNB above the mean sea level.
T_{vp} = Virtual temperature of an air parcel at a pressure level 'p' following the pseudo-adiabat through the surface wet bulb temperature.
Tve = Virtual temperature of the environment at a pressure level 'p'.

Physically CAPE may be interpreted, as the maximum amount of potential energy, possessed by air parcel, solely due to convection, convertible to vertical kinetic energy.

Convective inhibition energy (CINE): The CINE has been computed using the formula

$$CINE = \int_{Z_{SFC}}^{Z_{LFC}} \frac{g(T_{vp} - T_{ve})}{T_{ve}} dz = \int_{P_{SFC}}^{P_{LFC}} \frac{-R(T_{vp} - T_{ve})}{P} dp \quad \text{.... (2),}$$

where, Z_{SFC} is the height of the surface above mean sea level and P_{SFC} is the surface pressure.

Physically, CINE may be interpreted as the amount of energy that must be supplied to an air parcel up to LFC following pseudo adiabat through the surface wet bulb temperature, to overcome the resistance inhibiting convection, owing to the low level stability of the atmosphere.

By using the value of CAPE and CINE, a non-dimensional kinetic parameter 'K' has been computed by using the formula

$$K = \frac{|CAPE| - |CINE|}{|CINE|} \quad \text{....(3).}$$

Clearly K is large, when CAPE is large and CINE is small. Atmospheric state is favourable for convective activity, when the value of K is large and moisture is available up to a considerable depth in the atmosphere.

Moist static energy (MSE): The MSE at any level is given by, $MSE = c_p T + gz + Lq$, where the symbols carry their usual meaning. Now, assume MSE_i be the moist static energy at level $z=z_i$. Then the stability parameter 'σ', following Dutta and De (1999), is defined by

$$\sigma = \frac{\sum_i z_i MSE_i}{\sum_i z_i} = \frac{\sum_i z_i (c_p \overline{T}_i + gz_i + Lq_i)}{\sum_i z_i} \quad \text{.... (4),}$$

where T_i, q_i are respectively the values of T and q at $z=z_i$.

Development of a model for convective precipitation intensity:

Development of a dynamical model for convective updraft:

Since the orographic influence is not significant in enhancing rainfall at the coastal location over the Western Ghats (Sarker1966, 1967), we assume that vertical motion is almost entirely due to convective instability or synoptic scale convergence.

To develop the present convective updraft model, we assume for simplicity that updraft is solely due to the buoyancy effect. In such case the effect of in-situ change as well as that of advectional change of vertical motion and also the effect of friction have been neglected. Also the effects of entrainment and compensating downward motion have not been considered in the present study i.e., vertical acceleration is solely due to the effect of buoyancy force. So, at any level the vertical momentum equation of the air parcel may be written as:

$$w\frac{\partial w}{\partial z} = g\frac{T_{vp}(z) - T_{ve}(z)}{T_{ve}(z)} \quad \text{.... (5).}$$

Hence,

$$w\frac{\partial w}{\partial p} = B(p) \quad \text{.... (6),}$$

where, $B(p) = -\frac{R}{p}\left[T_{vp}(p) - T_{ve}(p)\right]$ is the buoyancy force at pressure level 'p'. The above equation is solved, numerically, for 'w' at every pressure level using the condition that w=0 at LFC.

The partial derivative in equation (5) is approximated by backward difference scheme for levels above LFC and is approximated by forward difference scheme for levels below LFC.

Numerical solution

The atmosphere in the vertical is divided into discrete layers by inserting discrete pressure levels indexed as 1,2,3,....,LFC-1,LFC,LFC+1,....,N. Here index 1 corresponds to 1000hpa level and N corresponds to LNB, if found, otherwise 200hpa level.

Now, $w_{LFC} = 0$ .... (7)

For,

$$i < LFC, \quad \left(\frac{\partial w}{\partial p}\right)_i \approx \frac{w_{i+1} - w_i}{p_{i+1} - p_i} \qquad ....(8)$$

And for

$$i > LFC, \left(\frac{\partial w}{\partial p}\right)_i \approx \frac{w_i - w_{i-1}}{p_i - p_{i-1}} \qquad .... (9)$$

Hence, for levels above the level of LFC, we have, $w_i(w_i - w_{i-1}) = B_i$,

where,

$$B_i = -R\frac{(T_{vp} - T_{ve})_i}{p_i}\Delta p, \; \Delta p = p_{i+1} - p_i = p_i - p_{i-1}$$

Hence, $w_i = \dfrac{w_{i-1} \pm \sqrt{w_{i-1}^2 + 4B_i}}{2}$ ....(10)

Now for the levels above LFC, $T_{vp} > T_{ve}$ and also Dp < 0. Hence for those levels $B_i > 0$. Thus for those levels the quantity inside the square root in the above expression is greater than w_{i-1}^2. Hence if negative sign is retained in equation (10), then we get, $w_i < 0$, which is not realistic, because due to higher temperature the parcel is positively buoyant with respect to the environment. So, negative sign should be dropped in the above equation (10).

Hence, for the levels above LFC, we have,

$$w_i = \frac{w_{i-1} + \sqrt{w_{i-1}^2 + 4B_i}}{2} \qquad .... (11).$$

Now, for levels below the level of LFC, we have, $w_i(w_{i+1} - w_i) = B_i$, where,

$$B_i = -R\frac{(T_{vp} - T_{ve})_i}{p_i}\Delta p, \; \Delta p = p_{i+1} - p_i = p_i - p_{i-1}$$

Hence,

$$w_i = \frac{w_{i+1} \pm \sqrt{w_{i+1}^2 - 4B_i}}{2} \qquad .... (12)$$

Now for the levels below LFC, $T_{vp} < T_{ve}$ and also Dp < 0. Hence for those levels $B_i < 0$. Thus for those levels the quantity inside the square root in the above expression is greater than w_{i+1}^2. Hence if positive sign is retained in equation (12), then we get, $w_i > 0$, which is not realistic, because due to lower temperature the parcel is negatively buoyant with respect to the environment. So, positive sign should be dropped in the above equation (12).

Hence, for the levels below LFC, we have,

$$w_i = \frac{w_{i+1} - \sqrt{w_{i+1}^2 - 4B_i}}{2} \qquad .... (13).$$

Using the radio-sonde data of the concerned place and using the algorithm developed by Krishnamurti (1986), we obtain the virtual temperature of environment and that of air parcel and then at each discrete pressure level 'i' , the value of B_i is computed. And then using these values of B_i, convective updraft at each level is computed using equations (7), (11) and (13).

Physical model for computation of rainfall intensity For this model the inputs are vertical velocity (w) as obtained from the section 3.2, the density (r) and saturation-mixing ratio (q) of the air parcel. Following Sarker (1966,1967), rainfall intensity due to the net convergence in the layer bounded by the ith pressure level and $(i+1)$th level is given by

$$R_i = \{w_i\rho_i(q_i - q_m) + w_{i+1}\rho_{i+1}(q_m - q_{i+1})\} \times 0.036 \text{ mm/hr},$$

where, w_i, q_i, ρ_i are respectively the vertical velocity, saturation mixing ratio and density at ith pressure level and $w_{i+1}, q_{i+1}, \rho_{i+1}$ are those at $(i+1)$th level, $q_m = \dfrac{q_i + q_{i+1}}{2}$ and R_i is the rainfall intensity due to the net convergence in the layer bounded by the ith pressure level and $(i+1)$th level. Hence the rainfall intensity obtained from the column up to LNB is $\sum_{1}^{N-1} R_i$ mm/hr. In the above summation we have taken $R_i = 0$ if, $R_i \leq 0$.

RESULTS AND DISCUSSION

Now we discuss each of the cases presented in the study. In each case Mumbai (Santacruz), a station on the west coast reported isolated heavy rainfall on a day followed and preceded by days with comparatively lesser amount of rainfall. The thermodynamic structure of the atmosphere on these three days of the spell has been studied using radio sonde data of relevant station for these days. We have computed the rainfall intensity from the convective precipitation model, as discussed in section 3. Now we shall discuss the synoptic features and thermodynamic structure of the atmosphere for the following cases:

Synoptic features

Synoptic features for all the five cases have been given in table 1. From this table it is clear that in most of the cases dominant synoptic features were (i) west coast trough extending northward upto north Maharashtra or even up to south Gujarat (ii) MTC over Gujarat /Maharashtra-Goa coast/East Arabian sea (iii) Cycir over Gujarat/ East central Arabian sea, which might be attributed for rainfall over SCz. But

Table-2. Synoptic features for all the five cases

Case No.	Period	Synoptic features
1.	20 June'82 to 22 June '82	1. Throughout the period there was a trough on the sea level pressure field off the west coast of India extending from north Gujarat coast to Lakshadweep. 2. A mid-tropospheric cyclonic circulation (MTC) over south Gujarat on 21st June, which became less marked on 22 June '82.
2.	30 June'84 to 2 July'84	1. Throughout the period there was a trough on sea level pressure field off the west coast of India extending from Gujarat/south Maharashtra coast to Lakshadweep. 2. Also there was an east-west trough extending from Saurashtra and Kutch to Gangetic West Bengal, with an embedded cyclonic circulation over Gujarat.
3.	15 June'85 to 17June'85	1. Throughout the period there was west coast trough in sea level pressure field off the Maharashtra coast. 2. There was MTC off Maharashtra-Goa coast on 15 and 16 June, which lay over east central Arabian sea and adjoining Maharashtra on 17 June.
4.	29 June'87 to 1July '87	1. On 29 June there was a cyclonic circulation (cycir) over east central Arabian Sea off Karnataka coast between 3.1km and 4.5km above mean sea level, which became less marked on 30 June. 2. On 29 there was a cycir over Gujarat region between 3.1km and 3.6km, which on the next day moved northward to west Rajasthan and lying between 2.1km and 3.6km. 3. On 30 June there was a trough in the lower levels over east Arabian Sea extending from South Gujarat coast to Karnataka coast. This trough was present on 1 July also. 4. On 1 July one MTC formed over Gujarat region.
5.	15 June'89 to 17June'89	1. Throughout the period there was trough off the west coast of India extending from Gujarat coast to Karnataka/Kerala coast, extending up to 2.1km above sea level on 15 and up to 0.9km on the next two days.

interestingly, in most of the cases it is found that synoptic features, listed in Table-2, did not vary significantly on all the three days of a spell to account for significant fluctuation in the daily rainfall of that spell.

Thermodynamic structure

Case-1. 21 June '82-23 June '82

We have used the radio sonde (RS) data of SCz, averaged between 0000UTC and 1200UTC observations, for 20June'82-22June'82 and also for comparison we have used the observed rainfall data of SCz for the dates from 21June'82-23June'82. Values of CAPE, CINE, MSE and s for the dates from 20June'82-22June'82 and also convective rainfall intensity for the dates from 21June'82-23June'82, were computed for the purpose of study.

The daily variations of CAPE and CINE on these three days are shown in Fig.1(a). It shows a sharp increase in rainfall from 0.8mm on 21June'82 to 135.2mm on 22June'82 associated with a sharp rise in the positive value of CAPE from 1865.32 J/Kg on 20June'82 to 3330.96J/kg on 21June'82 and also associated with a fall in the negative value of CINE from –8.09J/kg on 20June'82 to –2.78J/kg on 21June'82. This is also reflected in most of the cases.

It is also seen that the sharp fall in rainfall to 0.7mm on 23June'82 is associated with a sharp fall in CAPE to 1897.92 J/kg and rise in CINE to –180.48J/kg on 22June'82.

The daily variation of 'σ' for the dates from 20June'82 to 22June'82 is shown in Fig.1(b). It shows that a rise in rainfall on the 2[nd] day is associated with a fall in s. Fig.1(c) shows the daily variation of the parameter 'K' for the dates from 20June'82 - 22June'82. It shows that the rise in rainfall on the 2[nd] day is associated with a significant rise (from 229.45 to 1197.20) in the value of 'K' and the fall in rainfall is associated with a fall (from 1197.20 to 9.52) in the value of 'K'.

The daily variation of actual and the computed rainfall intensities are shown in Fig.1(d). From the fig it is clear that the fluctuation in the daily-observed rainfall and that computed are in the same sense, but the computed rainfall intensities are largely overestimated. This may be attributed to the other factors (may be more dominant in this case) like advectional, in-situ change in w, the entrainment effect, the effect of compensating downward motion and the effect of friction, which have not been incorporated in this simplified model.

However, the present convective rainfall model is, at least qualitatively able to capture the fluctuation in the daily-observed rainfall.

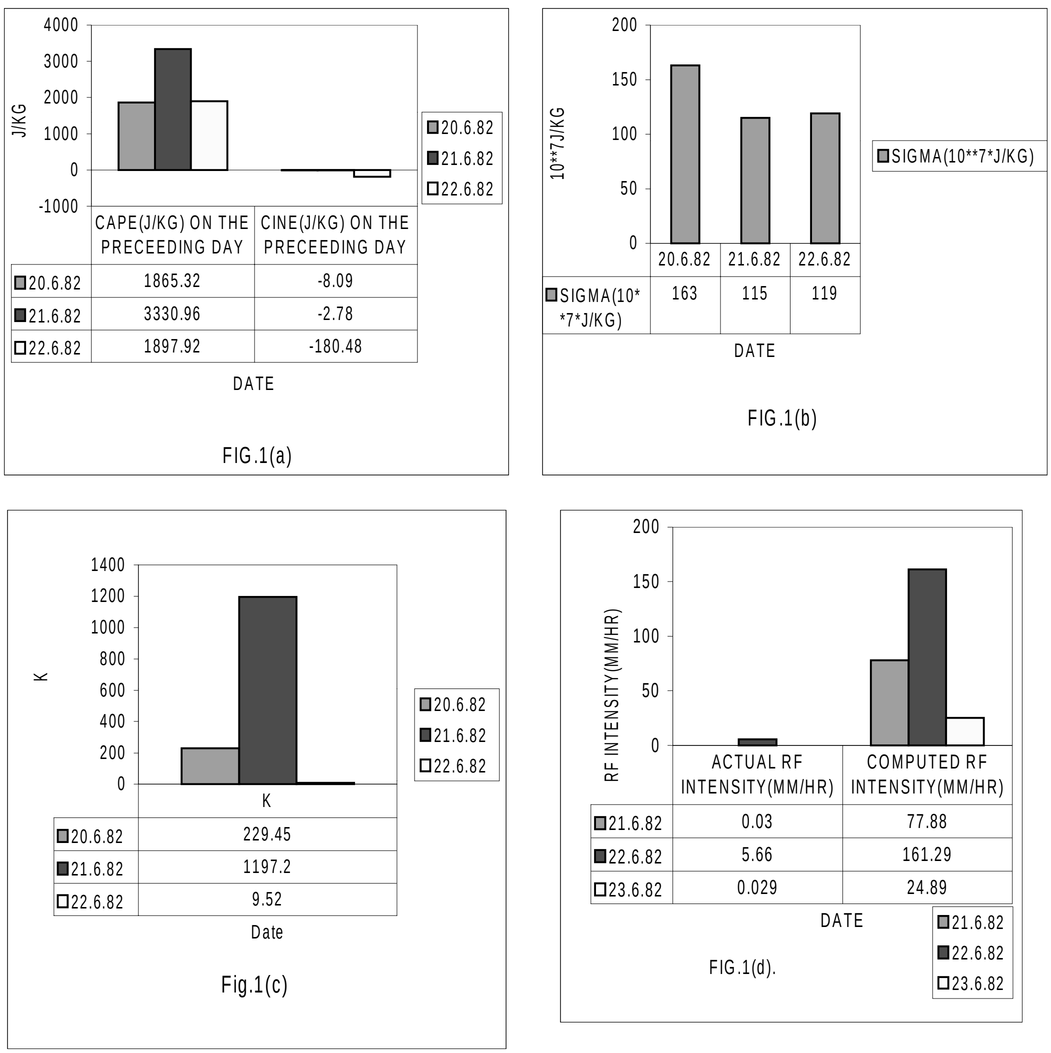

Figure 1 (a): Daily variation of cape/cine during 20.6.82-22.6.82. (b): Daily variation of sigma during 20.6.82-22.6.82. (c): Daily variation of 'k' during 20.6.82-22.6.82. (d): Daily variation of rainfall intensity during 21.6.82-23.6.82.

Case-2. 1 July '84-3 July '84

We have used the radio sonde (RS) data of SCz, averaged between 0000 UTC and 1200 UTC observations, for 30June'84-2July'84 and also for comparison we have used the observed rainfall data of SCz for the dates from 1July'84-3July'84. Values of CAPE, CINE, MSE, σ and K for the dates from 30June'84-2July'84 and also convective rainfall intensity for the dates from 1July'84-3July'84, were computed for the purpose of study.

The daily variations of CAPE and CINE on these three days are shown in Fig.2(a). There is a sharp increase in rainfall from 38.2mm on 1July'84 to 544.3mm on 2July'84. It is associated with a fall in the negative value of CINE from –360.68J/kg on 30June'84 to –165.54J/kg on 1July'84 and with a sharp rise in CAPE from 117.66J/kg on 30June'84 to 1365.27J/kg on 1July'84. It is also seen that the sharp fall in rainfall to 47.6mm on 3July'84 is again associated with a sharp fall in CAPE to 36.45 J/kg and with a sharp increase in CINE to –226.94 J/kg on 2July'84.

The daily variation of 'σ' on for the dates from 30June'84-2July'84 is shown in Fig.2(b). The Fig. shows that s decreases from the 1st to 2nd day in association with the increase in rainfall from the 1st to 2nd day and s increases from the 2nd to 3rd day in association with the decrease in rainfall from the 2nd to 3rd day.

Fig.2(c) shows the daily variation of the parameter 'K' for the dates from 30June'84-2July'84. It shows that the rise in rainfall on the 2nd day is associated with a rise (from –0.67 to 7.25) in the value of 'K' and the fall in rainfall on the 3rd day is associated with a fall (from 7.25 to –0.84) in the value of 'K'.

The daily variation of actual rainfall and the computed rainfall intensity has been shown in Fig.2(d). From the Fig. it is seen that in this case, similar to case1, the fluctuations in the daily-observed rainfall and that computed are matching. But unlike to case1, in this case computed rainfall intensities are closer to observed rainfall intensities.

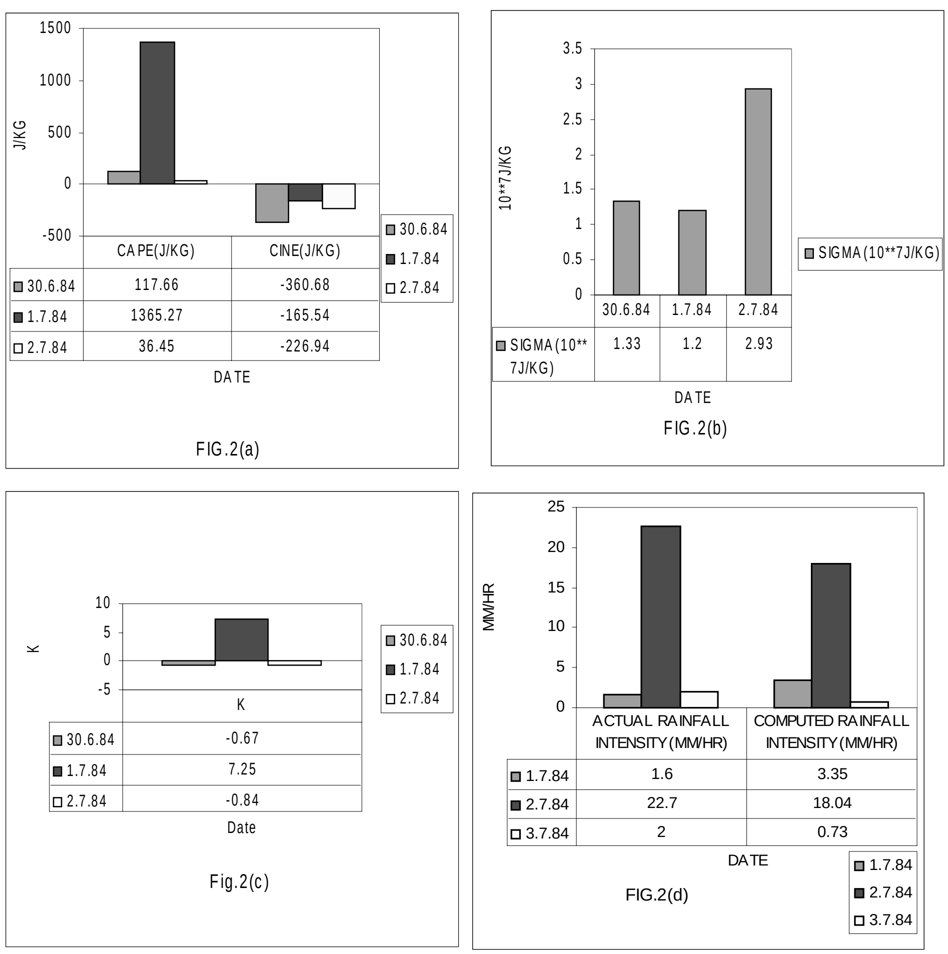

Figure 2 (a): Daily variation of cape/cine during 30.6.84-2.7.84. (b): Daily variation of sigma during 30.6.84-2.7.84. (c): Daily variation of 'k' during 30.6.84-2.7.84. (d): Daily variation of rainfall intensity during 1.7.84-3.7.84.

Case-3. 16June'85-18June'85

We have used the radio sonde (RS) data of SCz, averaged between 0000 UTC and 1200 UTC observations, for 15June'85-17June'85 and also for comparison we have used the observed rainfall data of SCz for the dates from 16June'85-18June'85. Values of CAPE, CINE, MSE, and s for the dates from 15June'85-17June'85 and also convective rainfall intensity for the dates from 16June'85-18June'85, were computed for the purpose of study.

The daily variations of CAPE and CINE on these three days are shown in Fig.3(a) and that of s is shown in Fig.3(b). It shows a sharp rise in CAPE from 46.15J/kg on 15June'85 to 879.08J/kg on 16June'85 associated with a sharp increase in rainfall from 38.3mm on 16June'85 to 345.5mm on 17June'85, a slight fall in CAPE to 843.77 J/kg on 17June'85 associated with a sharp fall in rainfall to 4.6mm on 18June'85. We also see that, although there is a rise in the positive value of CAPE from the 1st to the 2nd day in response to the rise in rainfall on the 2nd day, there is also an increase in the negative value of CINE on the 2nd day and in contrast there is a significant fall in CINE along with CAPE on 3rd day.

Fig.3(b) shows that s has reduced considerably on the 2nd day in response to the increase in rainfall and remained same on the 3rd day.

Fig.3(c) shows the daily variation of the parameter 'K' for the dates from 15June'85-17June'85. It shows that the rise in rainfall on the 2nd day is associated with a rise (from –0.40 to 3.00) in the value of 'K', but the fall in rainfall on the 3rd day is not associated with a fall in the value of 'K', rather associated with a rise from 3.00 to 9.10.

The daily variation of actual rainfall and the computed rainfall intensity has been shown in Fig.3(d). It is seen that in this case also the fluctuations in the daily-observed rainfall and that computed rainfall are in the same sense. Computed rainfall intensity is closer to the observed rainfall intensity on the 1st day, while on the other two days it is overestimated.

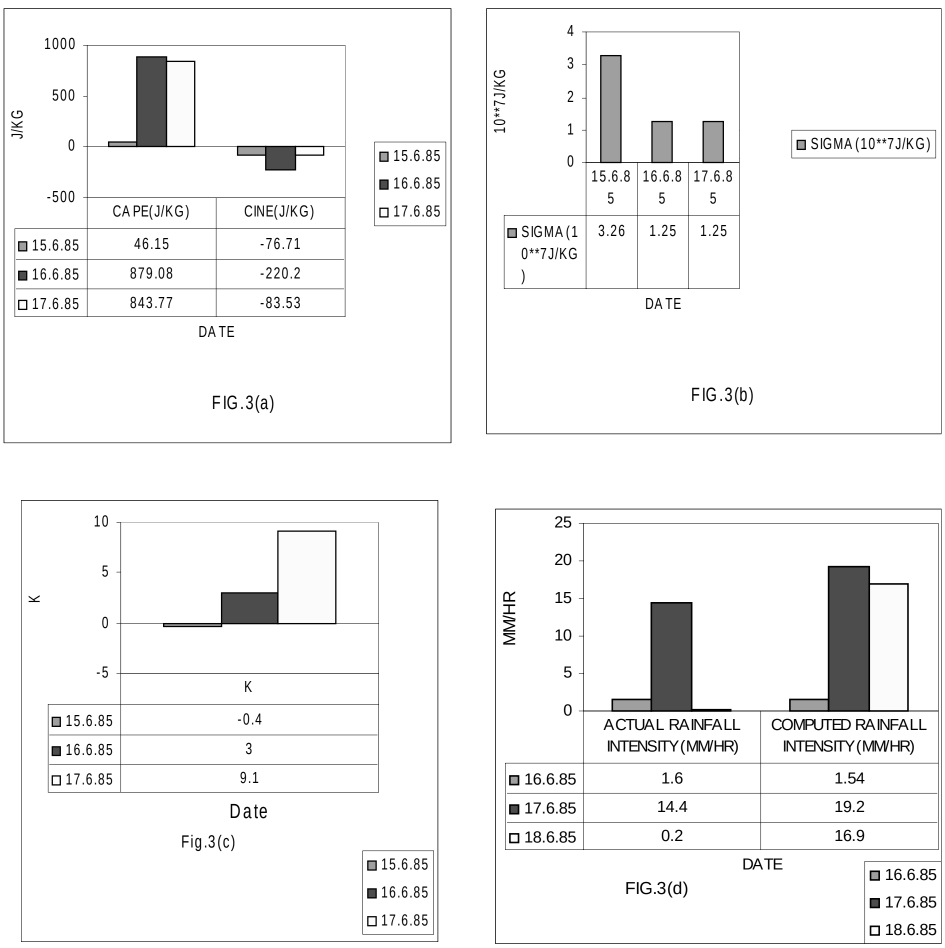

Figure 3 (a): Daily variation of cape/cine during 15.6.85-17.6.85. (b): Daily variation of sigma during 15.6.85-17.6.85. (c): Daily variation of 'k' during 15.6.85-17.6.85. (d): Daily variation of rainfall intensity during 16.6.85-18.6.85.

Case-4: 30June'87-2July'87

We have used the radio sonde (RS) data of SCz, averaged between 0000 UTC and 1200 UTC observations, for the dates from 29June'87-1July'87 and also for comparison we have used the observed rainfall data of SCz for the dates from 30June'87-2July'87. Values of CAPE, CINE, MSE, s for the dates from 29June'87-1July'87 and also convective rainfall intensity for the dates from 30June'87-2July'87, were computed.

The daily variations of CAPE and CINE on these three days are shown in Fig.4(a) and that of s are shown in Fig.4(b). It shows a sharp increase in CAPE from 1.08J/kg on 29June'87 to 1403.3J/kg on 30June'87 associated with a rise in rainfall from 15.7mm on 30June'87 to 153.9mm on 1July'87 and a fall in CAPE to 492.53 J/kg associated with a sharp fall in rainfall to 59.0mm on 2July'87. It is also seen that, although there is a rise in the positive value of CAPE from the 1st to the 2nd day of the spell in response to the rise in rainfall on the 2nd day, there is also an increase in the negative value of CINE on the 2nd day and there is a significant fall in CINE along with CAPE on 3rd day.

Fig.4(b) shows that s has an increasing trend during the said period.

Fig.4(c) shows the daily variation of the parameter 'K' for the dates from 29June'87-1July'87.The Fig. shows that the rise in rainfall on the 2nd day is associated with a rise (from –0.88 to 6.06) in the value of 'K', but the fall in rainfall on the 3rd day is not associated with a fall in the value of 'K', rather associated with a significant rise from 6.06 to 1301.99.

The daily variation of actual rainfall and the computed rainfall intensities are shown in Fig.4(d). It can be seen that in this case, similar to case1, the fluctuation in the daily observed rainfall and that computed agree qualitatively, but computed rainfall is an overestimate of rainfall on all the days except the 1st day. So, in this case also the model can qualitatively capture the fluctuation in the daily-observed rainfall over SCz in a series of three consecutive days.

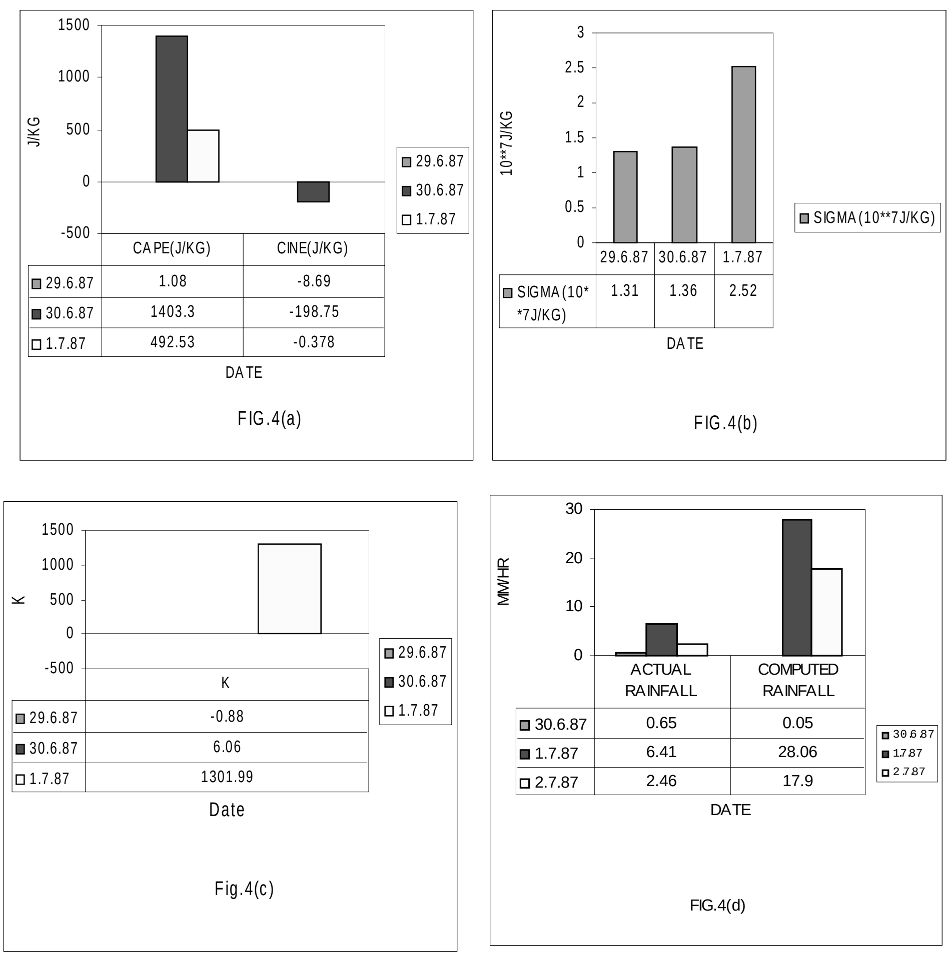

Figure 4 (a): Daily variation of cape/cine during 29.6.87-1.7.87. (b): Daily variation of sigma during 29.6.87-1.7.87. (c): Daily variation of 'k' during 29.6.87-1.7.87. (d): Daily variation of rainfall intensity during 30.6.87-2.7.87.

Case-5. 16June'89-18June'89

We have used the radio sonde (RS) data of SCz, averaged between 0000 UTC and 1200 UTC observations, for the dates from 15June'89-17June'89 and also for comparison we have used the observed rainfall data of SCz for the dates from 16June'89-18June'89. Values of CAPE, CINE, MSE, s for the dates from 15June'89-17June'89 and also convective rainfall intensity for the dates from 16June'89-18June'89, were computed for the purpose of study. The daily variations of CAPE and CINE on these three days are shown in Fig.5(a) and that of s are shown in Fig.5(b). It shows a sharp rise in CAPE from 1460.2J/kg on 15June'89 to 2280.5J/kg on 16June'89 and a drop in the negative value of CINE from –199.19J/kg on 15.6.89 to –1.36J/kg on 16June'89 associated with a sharp increase in rainfall from 15mm on 16June'89 to 111.9mm on 17June'89. The Fig. also shows a sharp fall in CAPE to 32.59 J/kg and a rise in CINE to –104.83J/kg on 17June'89 associated with a sharp fall in rainfall to 4.3mm on 18June'89.

Fig.5(b) shows that as the rainfall increases from first to the 2nd day, s decreases and as rainfall decreases on the 3rd day, s increases.

Fig. 5(c) shows the daily variation of the parameter 'K' for the dates from 15June'89-17June'89.The Fig. shows that the rise in rainfall on the 2nd day is associated with a significant rise (from 6.33 to 1675.47) in the value of 'K' and the fall in rainfall on the 3rd day is associated with a significant fall in the value of 'K' to –0.69.

The daily variations of actual and the computed rainfall intensities are shown in Fig.5(d). It is seen that, in this case, similar to case1, the pattern in the fluctuation of daily-observed rainfall intensity is similar to that of computed daily rainfall intensity. But the computed rainfall intensities are an overestimate. However, this over estimation is less on the 3rd day. So, in this case also the model can qualitatively capture the fluctuations in the observed rainfall over SCz for the epoch.

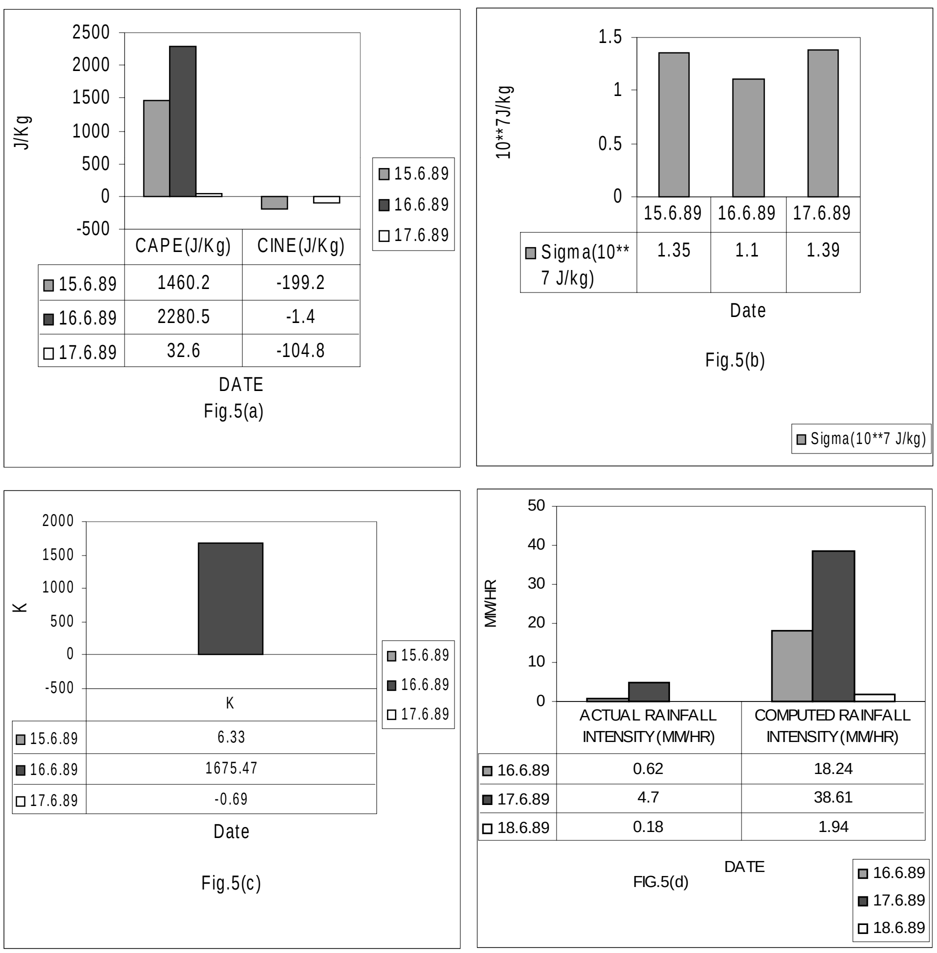

Figure 5 (a): Daily variation of cape/cine during 15.6.89-17.6.89. (b): Daily variation of sigma during 15.6.89-17.6.89. (c): Daily variation of 'k' during 15.6.89-17.6.89. (d): Daily variation of rainfall intensity during 16.6.89-18.6.89.

Table-3. Comparison of the vertical velocities obtained from different studies in India.

S.No	Study	System	Order of computed vertical velocity
1.	Rao and Rajamani (1975) using diagnostic omega equation.	Synoptic scale systems like Monsoon depression.	0.1-4cm/sec with maximum reaching to 5cm/sec between 850 and 700hpa.
2	Mukherjee et al (1977) using simple parcel method: Without entrainment effect (I) With entrainment effect (II)	Thunder storms over Mumbai during pre-monsoon season.	(I) 47.5m/sec (II) 28.6m/sec
3	Mukherjee and Choudhury (1979) using parcel method.	Excessive overshooting of cumulonimbus over Calcutta during pre-monsoon season.	88.6m/sec
4	Sinha ray et al (1982) using 2-D orographic rainfall model.	Orographic rainfall during southwest monsoon season.	20-30cm/sec reaching to maximum value 50-60cm/sec below 850hpa.
5	Mukherjee et al (1984) using continuity equation in p-co-ordinate.	Off shore vortex over east Arabian sea during southwest Monsoon season.	Maximum value 2.6cm/sec at 5km level.
6	Present study using simple parcel method.	Isolated heavy to very heavy rainfall (Daily).	40-60m/sec during heavy rain spells.

From the foregoing discussions it appears that in most of the cases synoptic features in a spell did not vary significantly to account for the significant fluctuation in daily rainfall in that spell.

Comparison of the convective updraft

The convective updraft computed using the technique explained in section 3.2.2 was compared with the typical vertical velocities computed by earlier during different seasons in India and are shown in Table 3.

The Table shows that the results of the present study are similar to those of Mukherjee, Kumar & Krishnamurthyl (1977) and Mukherjee & Choudhury (1979). The high values of vertical velocities computed using a simple buoyancy model might be due to non-inclusion of entrainment, advection and downward compensating flow and frictional drag, which are important. However, the model attempts to bring out that convection plays a significant role in release of precipitation in the western coast close to the Western Ghats. The orographic lifting may trigger off some of these heavy rain spell away from the mountain peaks by acting in a symbiotic manner with the convective activity.

CONCLUSIONS

Important conclusions of the study are:

In most of the cases it is found that synoptic features did not vary significantly on all the three days of a spell to account for significant fluctuation in the daily rainfall of that spell.

In most of the cases, a sharp change in rainfall from a comparatively low rainfall to a comparatively very high rainfall is associated with a marked rise in the positive value of CAPE and a fall in the negative value of CINE. Likewise a fall in observed rainfall is associated with a fall in CAPE. The changes range from 850J/kg to 1500J/kg in case of CAPE and from 150J/Kg to 200J/Kg in case of CINE.

A sudden rise in rainfall is associated with a fall in the weighted average value of MSE (σ). Change in s ranges from 11.0E5 J/Kg to 200.0E5 J/Kg.

In general, a sudden rise in rainfall on 2nd day is associated with a rise in the value of the non-dimensional kinetic parameter 'K'.

Although the computed values are much higher, the present model, in general, captures, at least qualitatively, the fluctuations in the daily-observed rainfall intensities in an epoch of three consecutive days.

The study reveals that, in the orographic region also sudden variation of rainfall away from the peak could be due to a favourable interaction between orographic forcing and convective instability.

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