Chapter 2
Tropical Disturbances (Quasi-stationary Waves, Easterly/Westerly Waves, Lows and Depressions, Cyclonic Storms, and Meso-Scale Disturbances)

2.1 Introduction

It is widely believed that weather in the tropics is largely seasonal and monotonously of one type; hot and humid during summer and cold and dry during winter. In reality, it is not always quite so, for, the monotony is often broken by different types of tropical disturbances ranging from synoptic-scale wave disturbances that form in the tradewinds that converge into the circulations around the equatorial trough of low pressure, and by severe local storms originating in meso-scale and small-scale disturbances. The synoptic-scale wave disturbances which have a mean horizontal scale of about 1000–3000 km and a Lagrangian time scale of 3–7 days usually move westward with a speed of about 5–8 m s\(^{-1}\). Some of these disturbances, under favorable conditions, develop into depressions and cyclones. They are mostly thermally driven and powered by the latent heat of condensation of water vapour carried by the tradewinds. For this reason, the most intense of these disturbances form and develop over the warmer parts of the oceans where there is almost unlimited supply of heat and moisture from the underlying surface for their development. They are called cyclones in the Indian Ocean, hurricanes in the Atlantic and typhoons in the Pacific Oceans, though they are fundamentally the same. Out of the large number of lows and depressions that form in the tropics, a limited number only develop into cyclones or severe cyclones and, out of these, only a small percentage develop an ‘eye’ at the center. The severe cyclones are associated with extremely high winds and torrential rain. Some of them cause storm surges which inundate many coastal belts and cause heavy loss of life and property. Many of them breed deadly tornadoes which knock down many standing structures and cause heavy loss of life and property. Tornadoes have drawn the attention of meteorologists from early times and we have an extensive literature on the subject, dealing with their formation, structure, development and movement.

Table 2.1 gives a list of the synoptic-scale disturbances classified on the basis of their scale and tangential wind speed, adopted by most meteorological services.

Most of the above-mentioned disturbances originate in quasi-stationary waves when they interact with traveling disturbances. Embedded within the above-mentioned large- or synoptic-scale disturbances, there are several types of meso- or small-scale disturbances of shorter duration which are very violent in nature. Just...
as in midlatitudes, these include thunderstorms, hailstorms, tornadoes, squalls, etc., but seldom fronts. In the present chapter, we first look at the structure and properties of quasi-stationary wave disturbances that form in zonal currents due to differential heating between land and sea. We then look at the structure and properties of wave disturbances that form and move in the tradewind easterlies and are called Easterly waves. The problem of development of a quasi-stationary trough of low pressure into a depression or a cyclone is addressed next. The role of condensation heating in development is emphasized. It is hypothesized that with enhanced condensation heating mainly on one side of the trough axis and without any appreciable change of structure, a quasi-stationary trough cannot develop further. A transition to a depression or cyclone requires, as a first step, docking of a traveling wave with a stationary wave; a phenomenon which brings about a change of structure with an additional zone of condensation heating. A developing low pressure system requires condensation heating to extend tangentially around the center, gradually turning the quasi-stationary wave disturbance into an axi-symmetric circulation system. Thus, a docking of a traveling wave with a quasi-stationary wave appears to be a prerequisite or necessary first step in development. The role of tropical easterly waves and subtropical/midlatitude westerly waves in development of tropical wave disturbances is discussed in some detail. Some aspects of subsynoptic or meso-scale disturbances and severe local storms are also discussed.

### 2.2 Quasi-stationary Waves

A quasi-stationary wave forms in meteorological fields over several parts of the tropics wherever land–sea thermal contrast exists and is characterized by higher temperature and lower pressure (‘heat lows’) over the warm land and lower temperature and higher pressure (‘cold highs’) over the neighboring cold ocean during summer. The fields are reversed during winter. At low levels, the circulation around the ‘heat low’ has westerlies on the equatorward side of the trough of the heat

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**Table 2.1 Classification of synoptic-scale tropical disturbances**

<table>
<thead>
<tr>
<th>Classification</th>
<th>Horizontal scale (km)</th>
<th>Range of wind speed (m s⁻¹)</th>
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</thead>
<tbody>
<tr>
<td>Low/trough</td>
<td>2000–2500</td>
<td>&lt;8.5</td>
</tr>
<tr>
<td>Depression</td>
<td>1500–2500</td>
<td>8.5–13.5</td>
</tr>
<tr>
<td>Deep depression</td>
<td>1500–2000</td>
<td>14.0–16.5</td>
</tr>
<tr>
<td>Cyclonic storm/tropical storm</td>
<td>1000–1500</td>
<td>17.0–31.5</td>
</tr>
<tr>
<td>Tropical cyclone (severe cyclonic</td>
<td>&lt;1000</td>
<td>&gt;32.0</td>
</tr>
<tr>
<td>storm/hurricane/typhoon)</td>
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low and easterlies on the poleward side. However, in the presence of an easterly thermal wind, the low-level westerlies give way to upper-level easterlies at some height. Both low-level westerlies and upper-level easterlies get involved in wave development process.

The existence of a quasi-stationary wave over a region is reflected in several meteorological fields, as shown by a detailed study by Saha and Saha (1996) for the monsoon region of southern Asia. Here we produce evidence from their study in respect of wind and temperature fields.

### 2.2.1 Quasi-stationary Wave in Wind Field

The wave structure in the wind field is revealed best by the zonal anomaly of the meridional component of the wind (v-component) at 850 and 700 hPa along 20°N during July, presented in Fig. 2.1.

Between the northerly and the southerly flows at both the pressure surfaces in Fig. 2.1, the location of three monsoon troughs, one each over the eastern part of the Arabian sea, the western part of the Head Bay of Bengal and the western part of South China sea can be identified. The mean zonal wavelength and amplitude of the wave in the v-field would appear to be about 2000–2500 km and 4 m s$^{-1}$ respectively. The wave structure is revealed not only in the low-level westerlies but also in the upper level easterlies (not shown), and other parameters.

![Fig. 2.1](image.png) Zonal anomaly (deviation from zonal mean) of the meridional component (v) of the wind (m s$^{-1}$) at 850 hPa (dashed) and 700 hPa (dotted) along 20°N during July. Positive southerly, negative northerly (from Saha and Saha, 1996)
2.2.2 Quasi-stationary Wave in Temperature Fields

Figure 2.2 shows the distribution of zonal anomaly (deviation from zonal mean) of a 10 year (1976–1985) mean July temperature over southern Asia along latitude

![Diagram showing the distribution of zonal anomaly over southern Asia along latitude 20°N.](image)

**Fig. 2.2** Quasi-stationary wave in a 10 year (1976–1985) mean July temperature anomaly (deviation from the zonal mean) field along 20°N over Southern Asia. Upper panel (a) shows the distribution of the anomaly at MSL, 850 and 700 hPa; Lower panel (b) shows the distribution of the anomaly in a vertical section up to 50 hPa (from Saha and Saha, 1996)
2.2 Quasi-stationary Waves

20°N, at MSL, 850 and 700 hPa (upper panel) and at different pressure surfaces in the troposphere (lower panel).

The upper panel brings out the land-sea thermal contrasts with the warm sectors generally appearing over the land and the cold sectors over the sea. The contrasts appear to stand out more clearly in the lower panel which suggests that the stationary wave is basically a phenomenon of the lower troposphere, its upper boundary being at about 500 hPa. It appears to have a wavelength of about 2500 km and amplitude of about 1°C in the mean temperature field. Above 500 hPa, the structure of the wave appears to change considerably, as it comes under the influence of the upper-tropospheric warm high over Southern Asia which forms part of the planetary-scale monsoon wave at 200 hPa (Kanamitsu and Krishnamurti, 1978).

A characteristic feature of the thermal structure above 500 hPa is a pronounced secondary temperature maximum (warm anomaly) situated directly over the low-level temperature maximum over the land and a minimum over the low-level temperature minimum over the ocean.

Features similar to those presented in Fig. 2.2 were also found along latitudes 15 and 25°N, but not along 10, or 30°N. The meridional domain of the stationary wave would, therefore, appear to be between 15 and 25°N with its strongest intensity along 20°N.

2.2.3 Structure of the Quasi-stationary Wave in Circulations

Figure 2.3 shows schematically the (a) horizontal (LL-lower level, UL-upper level), and (b) zonal-vertical circulations through a quasi-stationary wave over India (along 20°N) (after Saha and Saha, 1996).

2.2.3.1 Horizontal Circulation

The plan view in Fig. 2.3(a) shows the horizontal circulation in the stationary wave at the lower level (LL) and the upper level (UL). Note that in (a), there is flow convergence at the lower level and divergence at the upper level to the west of the low pressure center and divergence at the lower level and convergence at the upper level to the east.

2.2.3.2 Zonal-Vertical Circulation

Figure 2.3(b) shows the vertical circulation along the zonal section p, q, r, s, t in (a).

While interpreting the vertical circulation in Fig. 2.3(b), it must be borne in mind that the prevailing zonal wind over the region is westerly at low levels which gives way to easterlies above about 700 hPa. This variation of the wind with height is due to the thermal wind being easterly over the region. For this reason, the vertical tilt of the troughs (Tr) and ridges (R) is steeply eastward in the lower troposphere, and slightly westward in the upper troposphere. Further, it is to be borne in mind
that due to adiabatic changes in the atmosphere, the thermally indirect circulation with warmer air sinking and colder air rising generates available potential energy, while the direct circulation with warmer air rising and colder air sinking increases kinetic energy. Thus, the direct and indirect circulations are intimately coupled to each other and co-operatively drive the observed stationary wave circulation.

Note also that deep convection with heavy rainfall (RR) occurs to the west of the low-level trough of low pressure over the region between p and r where the low-level convergence is capped by upper-level divergence. Shallow convection with little or no rain occurs to the east between r and t where upper-level convergence is superimposed on low-level divergence. The difference can, perhaps, be understood
2.3 Traveling Easterly (E’ly) Waves

Several studies have addressed the problem of easterly waves over the North Atlantic and North Pacific Oceans that travel westward and under favorable conditions give rise to development of hurricanes and typhoons. We shall refer to some of these studies in the pages that follow. There is converging evidence that the E’ly waves originate in a zone of strong cyclonic wind shear in the belt of the tradewind easterlies that converge into the equatorial trough of low pressure over these oceans.

2.3.1 Easterly Waves in Tropical North Atlantic

Following early work of Dunn (1940) who noted that the periodic westward movement of disturbed weather across the Caribbean islands in the latitudes of Puerto Rico was associated with the movement of isallobaric lows and highs at mean sea level in regular succession, Riehl (1945, 1954) made detailed synoptic case studies and found that the observed movement of disturbed weather and isallobaric centers at surface was due to the westward movement of waves in the overlying easterly tradewinds. His detailed analysis threw light on the structure and properties of these
waves. On the basis of his studies in the Caribbean, Riehl (1945) proposed a model of an easterly wave, the characteristic features of which were the following:

1. The westward passage of the wave over a station is marked by a sharp clockwise turning of the wind direction from NNE to SSE;
2. A marked jump in the depth of the sub-cloud moist layer; and
3. A change in cloud and weather pattern from fair-weather small-scale cumulus clouds to large-scale towering cumulus and cumulonimbus clouds and rain and thunderstorms.

Riehl (loc. cit.) showed that the depth of the moist layer, which is related to development of clouds, is directly controlled by the field of motion in the vertical.

- Where an upper-tropospheric divergent flow is superimposed on a lower-tropospheric convergent flow, vertical motion is strongly upward and it leads to a deepening of the moist layer and development of large clouds.
- A superposition of upper-tropospheric convergence on lower-tropospheric divergence suppresses upward motion and vertical growth of the moist layer.

The vertical variation of the basic current can be easily understood in terms of a change of thermal wind. The profile (a) will prevail in regions where the thermal wind is westerly, and (b) where the thermal wind is easterly. A possible implication of a change of thermal wind across a borderline between the two regions in the development of a tropical disturbance will be discussed in a later part of this chapter.

Fujita (1971) used cloud motion vectors at low and high levels from satellite cloud imagery around 15 GCT on 14 July 1969 to construct streamlines to demonstrate the presence of easterly waves and their associated cloud systems over the tropical north Atlantic.

### 2.3.2 Easterly Waves in Tropical North Pacific

As in the Atlantic, easterly waves form in the tradewinds over the north Pacific and travel westward. Palmer (1952), Yanai (1961a,b, 1963, 1968a), and several others studied the structure and properties of these waves using different methods. Yanai et al. (1968), Wallace and Chang (1969), Nitta (1970), Wallace (1971) and others used the technique of power-spectrum and cross-spectrum analysis for the purpose. A compositing technique used by Williams (1970) and Reed and Recker (1971) yielded valuable information regarding the average properties of these waves. Chang (1970) used time composite satellite cloud pictures to identify and track these wave disturbances over long distances in the tropical Pacific. A number of workers (e.g., Krishnamurti and Baumhefner, 1966; Yanai and Nitta, 1967; Holton, 1971)
carried out diagnostic studies, using hydrodynamical equations, to infer different characteristics and obtain dynamically consistent pictures of these waves.

Reed and Recker (1971) summarized the average structure and properties of easterly waves in the equatorial western Pacific region as follows: the meridional wind maxima of nearly opposite phase occurred in the lower and upper troposphere. Negative temperature deviations were found in the vicinity of the wave trough at low and high levels; positive deviations were observed at intermediate levels. Highest relative humidities occurred in the trough region. This was also the region of strongest upward motion and maximum cloud amount and rainfall. The maximum upward velocity of 2.5 cm s\(^{-1}\) was found at 300 hPa. Convergence was strongest in the sub-cloud layer; divergence was concentrated near 175 hPa. The maximum anticyclonic vorticity was also observed at that height. Besides the above, there were several other properties of these waves that were revealed by their study. According to theoretical reasoning (Riehl, 1954) and numerical modeling (Holton, 1971), it was apparent that the wave structure would be sensitive to the vertical profile of the basic current in which the wave was embedded in addition to the horizontal profile. This was actually found to be so in the equatorial western Pacific where the vertical structure of these waves changed systematically as they travelled from the easternmost to the westernmost islands from near the dateline to a longitude of about 135E. The wave axis, which tilts eastward with height at the eastern end of the islands, becomes vertical in the central region and acquires an opposite tilt in the west. The change in the tilt is believed to be caused by the variation with longitude of the vertical shear of the basic current, as shown in Fig. 2.4. It may be seen that the vertical wind shear at Koror (7.5N, 135E) which lies at the western end of the network is very different from that at Majuro (7.0N, 171E) which lies near the eastern end and also from that of the mean of three island stations which lie in between.

Yanai (1964) suggested a model of the easterly wave based on his studies in the equatorial eastern Pacific, which is shown in Fig. 2.5. The upper panel shows a side view of the wave with isolines of vorticity (thin continuous lines) and the directions of vertical motion (double-shaft arrows) on the left and vertical distribution of the basic current on the right. The lower panel shows a horizontal view of the wave with streamlines (solid arrows) and isotachs (thin dotted lines) on the left, and the latitudinal distribution of the basic current at 850 hPa on the right.

### 2.3.3 Easterly Waves in the Indian Ocean Region

Easterly waves constitute a prominent feature of the atmospheric circulation over the Indian Ocean monsoon region. They are observed in quasi-stationary as well as traveling modes (For quasi-stationary wave mode, see Sect. 2.2).
Fig. 2.4  Mean zonal wind speed over western North Pacific for the period July–September, 1967. The profile labelled KEP is the mean for Kwajalein (9.0N, 167E), Eniwetok (12N, 162E) and Ponape (7.0N, 157.5E). The thin vertical line denotes the average wave speed (−9 m s$^{-1}$) observed during the period of study (after Reed and Recker, 1971)

The traveling E’ly waves usually originate over the Pacific Ocean and arrive over the Indian Ocean region after crossing the SE Asia peninsula. However, south of the equator, they originate in the extreme eastern parts of the Indian Ocean adjoining the northwestern coast of Australia. They usually move westward along the south equatorial trough of low pressure between the equator and 15°S.

Traveling E’ly waves that enter the Northern Indian Ocean region from the east appear to have more or less the same structure as the stationary E’ly wave over the region, shown in Fig. 2.3. However, their vertical structure varies considerably depending upon the prevailing wind structure.

An inclined quasi-stationary monsoon trough appears to offer the most favorable orientation for development of the trough into a depression, for, it provides, inter alia, an east–west component of the circulation associated with it to interact with an approaching E’ly wave. Where the trough is under influence of two opposing forces, it gets inclined in the direction of the more powerful heat low. In the Indian Ocean region, inclined monsoon troughs may be found on both sides of the equator. In the Arabian Sea sector, development is favored when the trough is inclined in a NE–SW
Fig. 2.5 Model of an easterly wave in tropical eastern Pacific (after Yanai, 1964)

direction, while an inclination in a NW–SE direction favors development in the Bay of Bengal sector.

2.4 Development of Waves

2.4.1 Meaning of Development

Before we discuss the problem of development of a trough of low pressure into deeper pressure systems, such as a depression or a cyclone, it may be desirable to clarify the meaning of the term ‘development’. The development of a low pressure system means a faster and greater fall of pressure at the center of the low pressure system relative to its surroundings, as measured by the Laplacian of pressure tendency, $\nabla^2 (\delta p/\delta t)$.

Alternatively, it can also be defined by an increase in the absolute vorticity of the circulation, i.e., by $d\eta/dt$, where $\eta$ is the absolute vorticity and the operator $d/dt$ denotes change following motion. If the system is stationary, $d/dt$ can be replaced by $\partial/\partial t$, the local tendency. It can be easily shown by integrating the hydrostatic equation that the pressure tendency at any given location depends upon the net flow divergence and/or mass advection in the overlying column of air, as shown below:
The pressure $p_s$ at the earth’s surface ($z=0$) is given by

$$p_s = \int_0^\infty \rho g \delta z,$$

where $\rho$ is density of air, and pressure is assumed to be zero at height infinity.

Differentiating $p_s$ with respect to $t$, and using the continuity equation,

$$\frac{\partial \rho}{\partial t} = -\nabla \cdot \rho \mathbf{V}$$

we get the pressure tendency equation in the form

$$\frac{\partial p_s}{\partial t} = -\int_0^\infty \nabla_H \cdot (\rho \mathbf{V}) g \delta z - \int_0^\infty \frac{\partial (\rho w)}{\partial z} g \delta z$$

where $\mathbf{V}$ is the horizontal velocity vector, $w$ is vertical velocity, $\nabla_H$ is horizontal Del operator and the first term on the right-hand side gives the contribution of horizontal mass convergence or divergence through the sides of the column, while the second term gives the contribution of vertical mass convergence or divergence through the base of the column. The second term may be simplified to $g(\rho w)_0$, since $\rho$ and $w$ are both zero at infinity.

We may, therefore, write the pressure tendency equation in the simplified form

$$\frac{\partial p_s}{\partial t} = -\int_0^\infty \nabla_H \cdot (\rho \mathbf{V}) g \delta z + g(\rho w)_0$$

Thus, the pressure tendency at a place is negative or positive, according as there is net mass divergence from, or convergence into, the overlying air column. Lines drawn through places of equal pressure tendency are called isallobars.

According to Table 2.1, a low pressure system can develop into many different stages, ranging from a depression to a most intense hurricane. It is estimated (Frank, 1971) that only about 10% of all tropical disturbances reach storm strength and about one in three depressions makes the transition to storm or hurricane. Further, the formation of an ‘eye’ in a hurricane marks a most crucial final stage in the transformation of a tropical disturbance. Most tropical disturbances (90%) fail to form the eye and never reach the highest intensity. Further, it is observed worldwide that most disturbances develop from a pre-existing trough of low pressure, such as that found in association with quasi-stationary monsoon troughs discussed in the preceding section and that after formation and development they usually move along the axes of the monsoon troughs. As a quasi-stationary feature of the tropical circulation, the quasi-stationary wave is a dynamically stable system but may get destabilized whenever a traveling Easterly wave interacts with it.
2.4.2 Development of a Quasi-stationary Monsoon Trough into a Depression

2.4.2.1 Role of Tropical E’ly Waves

In the past, mechanisms such as barotropic, baroclinic, or combined CISK-barotropic-baroclinic instability have been suggested to explain the formation of monsoon depressions. However, observations show that in spite of the presence of these types of instabilities in the mean monsoon atmosphere, it is only occasionally (i.e., once or twice a month) that a depression forms and only a small percentage of the depressions develop further to storm strength. So, what actually triggers the formation of a monsoon depression in the first place?

Some past studies (e.g., Koteswaram and George, 1958; Saha et al., 1981) suggested the involvement of easterly waves in the process. Koteswaram and George (1958) suggested that when upper-air divergence ahead of an approaching easterly wave trough was superimposed upon the convergent area of the low-level monsoon trough, the trough would develop into a depression. Saha et al. (1981) found that in a 10-year (1969–1978) period, out of the 52 depressions that formed over the Bay of Bengal during July–August, 45 (87%) could be traced to an interaction with westward-propagating easterly waves which moved in from areas to the east.

When a traveling easterly wave approaches the quasi-stationary monsoon wave and docks with it, the docking strengthens a pre-existing zone of subsidence to the east of the trough axis by added subsidence of the indirect circulation of a new zone of condensation heating created to the east of the subsidence zone. Enhanced subsidence leads to increased adiabatic warming and a fall of surface pressure immediately to the east of the trough axis. Thus, the initiation of condensation heating on either side of trough axis with a zone of enhanced subsidence warming in between produces a complete depression wave. The depression circulation is centered immediately to the east of the monsoon trough axis.

The development of the quasi-stationary trough into a depression is accelerated if the trough is oriented in a NW–SE direction and a cross-equatorial flow from the winter hemisphere converges into the TCZ. The involvement of cross-equatorial flow in development of a monsoon depression was suggested by several studies in the past (e.g., Sikka, 1980b, Sikka and Gray, 1981; Saha and Saha, 2004a,b) in the case of the monsoon trough over the Bay of Bengal, though without mentioning the branch of the cross-equatorial flow likely to be involved. However, it is important to note here that with a monsoon trough oriented in a NW–SE direction, it is only the Bay of Bengal branch of the cross-equatorial flow which can lead to cyclogenesis of the trough; not the Arabian Sea branch which activates the ITCZ. A simultaneous activation of both ITCZ and TCZ will, of course, accelerate the development. We may visualize the development process as illustrated in Fig. 2.6.

There is enough observational evidence from different parts of the tropics, which appears to support the conceptual model of the formation of a monsoon depression, suggested above and shown schematically in Fig. 2.6. Rao (1976) cites numerous
Fig. 2.6 Schematic illustrating the formation of a monsoon depression in a quasi-stationary monsoon wave trough by interaction with a traveling E’ly wave

examples of typical cloud distributions in the field of a monsoon depression over the Indian monsoon region, which strongly support the structure suggested in Fig. 2.6.

Several theoretical and numerical experiments (e.g., Charney and Eliassen, 1964; Krishnamurti et al., 1976; Shukla, 1978) on the formation of a monsoon depression have concluded that the eddy available potential energy required for development of a pre-existing perturbation in the monsoon current is, in all probability, supplied by condensation heating. The conceptual model presented here suggests a mechanism which can bring about the additional zone of condensation heating required for the development.

2.4.2.2 Role of Subtropical W’ly Waves

Like an easterly wave, a wave in subtropical/midlatitude westerlies can initiate the development of the quasi-stationary monsoon trough. It is well-known that W’ly waves in the course of their eastward travel often develop large amplitudes when they reach the longitudes of the quasi-stationary monsoon wave and interact with it across the subtropical belt in the upper troposphere (see Fig. 2.7).

The interaction produces extended troughs and ridges when the waves get into the same phase. During interaction, an exchange of energy occurs between the waves in which cold subsiding and adiabatically-warming air from higher latitudes move
Fig. 2.7 Interaction of the quasi-stationary monsoon trough with a W’ly wave trough in the upper troposphere in the northern hemisphere (trough locations are shown by thick dashed lines)

equatorward to intensify the subsidence warming to the west of the monsoon trough axis and warm rising and adiabatically-cooling air from lower latitudes move poleward to intensify the rising currents to the east of the midlatitude W’ly wave trough axis. So, when, after coupling, the waves separate, and the westerly trough moves away eastward, a fracture occurs in the extended trough axis and a cut-off low pressure forms around the easterly wave trough, which may subsequently develop into a monsoon depression.

It should, however, be noted that in this interaction, it is the TCZ which gets involved and activated and becomes a zone of penetrative convection, clouding and precipitation; not the ITCZ on the southwestward side of the trough axis.

Thus, as a result of this interaction, we have two zones of condensation heating with a zone of enhanced subsidence warming between them, which are favorable for the formation of a monsoon depression. Here also, as in the case of interaction with an easterly wave alone, the depression is likely to be centered immediately to the east of the low-level monsoon trough axis. The arrival of a cross-equatorial flow from the winter hemisphere to the trough zone at low levels at this time may lead to explosive cyclogenesis.

There is evidence to suggest that eastward-propagating large-amplitude W’ly waves interact with quasi-stationary monsoon troughs over all continents, leading to formation of monsoon depressions. Over the Indian subcontinent, the W’ly waves which move across the Himalayas and the Tibetan plateau influence the formation, development and movement of monsoon depressions over the region, especially the Bay of Bengal (e.g., Saha and Chang, 1983).

Summarizing the above paragraphs, we may state that both E’ly and W’ly waves can contribute to the development of a quasi-stationary monsoon trough into a depression. Explosive development may result if a cross-equatorial current from the winter hemisphere is also drawn into the circulation around the monsoon trough at low levels to activate the ITCZ and the TCZ during the interaction. The low frequency of such a favorable combination may, perhaps, explain why a monsoon
Depression forms so infrequently, usually one or two a month, though both easterly and westerly waves interact with the quasi-stationary monsoon trough more frequently.

### 2.4.3 Development of a Depression into a Cyclonic Storm/Tropical Cyclone/Hurricane/Typhoon

While interaction with a traveling wave appears to be an essential first step in the formation of a depression, further development of a depression into a tropical storm or tropical cyclone appears to be a long drawn-out and complicated process requiring the following pre-requisites for further growth.

- A depression must be in an environment which would favor condensation heating by sustained moisture convergence at low level over a long period of time. The following environmental conditions have been found to be necessary for the formation of a cyclonic storm or tropical cyclone;
- Pre-existence of a warm land, or an ocean with SST exceeding 26.5°C, and a closed low or trough of low pressure which ensures cyclonic circulation, flow convergence and positive relative vorticity at low levels;
- A deep mixed layer of the ocean with continuous availability of water vapour either through surface evaporation or advection of moist air, which ensures strong moisture convergence at low levels;
- Existence of conditional instability and weak vertical wind shear in the overlying atmosphere;
- Strong upper air divergence above low level convergence, which ensures penetrative convection, rapid build-up of large cloud clusters and precipitation and release of latent heat of condensation; such conditions are usually ensured in monsoon troughs which are inclined to latitudes, as depicted in Fig. 1.5;
- Release of enough latent heat of condensation to enforce strong subsidence warming and rapid fall of surface pressure near the center of the depression.
- For a tropical cyclone with an eye to form, condensation heating must extend all around the center of the depression. In other words, the cyclone must be axi-, or quasi-symmetric.

Furthermore, the development of a depression involves a reduction in the horizontal scale of the disturbance to no more than 30% of the original scale (see Table 2.1). How is this reduction brought about and with what effect on the physical and dynamical characteristics of the depression in the process? These are important questions to which we can provide only tentative answers at present, since the whole process is nonlinear and there are many dissipative processes at work. In fact, the process to full axial symmetry is a long and complicated one and there is little guarantee that a given disturbance will mature to its fullest extent. However, the process...
may be visualized as follows: The introduction of two zones of condensation heating on either side of the center of a low or depression induces increased subsidence warming and fall of pressure near the center. In response to the pressure fall, air from the outer boundary region starts flowing inward. The inflowing air becomes increasingly convergent with strong upward motion and development of towering clouds and precipitation. Another consequence of the inward flow is a rapid strengthening of the tangential wind, because the inflowing air is required to conserve absolute angular momentum. The scale of the disturbance thus gets gradually reduced till it stabilizes at some minimum radial distance from the center, where the pressure gradient force is balanced by the centrifugal, Coriolis and the frictional forces. This radial distance marks the inner boundary of a broad zone of maximum tangential wind, strongest updraft, and torrential precipitation, which constitutes the eye-wall.

In the course of the development process, the cloud zone which was originally confined to two narrow zones on either side of the depression center gradually extends along the circumference till the disturbance attains its nearly full axial symmetry with potential to develop an eye at the center. However, the process of formation of an eye in a tropical cyclone appears to be a complicated one. It is discussed further in Chap. 3, which deals specifically with the formation, structure and properties of tropical cyclones.

It follows from the foregoing considerations that depressions which have limited sea travel or which move over cold oceans have little chance to develop further. Also, the requirement of a weak vertical shear of the basic wind appears to be a very crucial one, for, it is important that the cool, moist air that converges at low levels in a trough zone must be lifted by upper level divergence as close to the trough zone as possible and that this can only happen in a low-level strongly convergent zone, such as the TCZ, as discussed in Chap. 1.

Gray (1968) in a worldwide survey of tropical cyclones for a 20-year period identified the initial genesis points of cyclones (Fig. 2.8). According to this survey, the areas most susceptible to formation of tropical cyclones are: (1) a wide
belt of ocean extending from the eastern Arabian Sea to the Western Pacific Ocean up to the dateline north of the equator; and (2) a more extensive belt extending from the coast of South Africa to about 160°W south of the equator. Two other major cyclone-breeding areas lie in the Western Atlantic Ocean and Eastern Pacific Ocean in the northern hemisphere. It is observed that in all the areas, development occurs around inclined troughs of low pressure. Undoubtedly, one may expect to see many more genesis points than found by Gray’s survey after introduction of earth-orbiting satellites. For example, a tropical cyclone formed over the southwesterly part of the Atlantic Ocean off the coast of Brazil in March, 2004, which was visible in a satellite cloud imagery, but not reported by conventional observational systems.

2.5 Meso-Scale Disturbances and Severe Local Storms in the Tropics

2.5.1 General Considerations – Source of Energy of the Storm

Besides large and synoptic-scale wave disturbances discussed in the foregoing sections, several types of subsynoptic- or meso-scale disturbances with diameters ranging from a few hundred meters to about 100 km, such as thunderstorms, hailstorms, tornadoes and squall-lines, also form in the tropical atmosphere. Most of them form in a conditionally unstable atmosphere within the synoptic-scale disturbances and develop by mutual co-operation between the large and the small scales. A small-scale disturbance with shallow convection may form in a conditionally unstable environment when it is heated from below or cooled from above, but its further growth in most cases depends upon the support of the synoptic- or large-scale disturbance which provides the necessary mechanism for its development. Thus, the basic requirements for development of a meso-scale tropical storm in any locality appear to be largely the same as enunciated in the case of development of a tropical cyclone in Sect. 2.4.3, viz,

(i) The existence of a conditionally unstable atmosphere;
(ii) Moisture convergence at low levels; and
(iii) An effective mechanism to raise the low-level moisture to higher levels and release the latent heat of condensation for further growth of the cloud through penetrative convection, the net effect of which is an adiabatic warming of the environment by subsidence.

Normally, depending upon the extent of mutual co-operation between the cloud and its environment, a small-scale cloud may grow to different stages and display different physical and dynamical characteristics. For example, it may grow to a stage where it gives only thunder and lightning, but no high wind, rainfall or hail at the ground. On the other hand, it may develop to a stage where it displays all these characteristics and, perhaps, additionally also a highly developed tornado vortex.
When fully developed, most of them make a significant contribution to the total precipitation and the circulation and weather phenomena in the tropics, apart from its impact upon society. In this section, we discuss some aspects of their formation, internal structure, electrical properties, and other physical and dynamical properties to the extent revealed by observations from several field projects, such as the well-known thunderstorm project (Byers and Braham, 1949) in USA. In fact, these experiments have largely confirmed some of the basic characteristics of atmospheric convection, such as strong updrafts inside the cloud cell, entrainment and detrainment of air, and net warming of the environment by subsidence.

2.5.2 Thunderstorms

A thunderstorm derives its name from its characteristic property of producing thunder and, inevitably, lightning in the sky besides stormy conditions including high, squally winds and sudden heavy downpour at the ground. Occasionally, it may turn into a hailstorm or even breed tornadoes. Because of its association with thunder and lightning, it is also sometimes called an ‘electrical storm’, though the term is normally used in the context of space weather. Generally, it appears in the sky as a giant cumulonimbus cloud with a large spread-out anvil at the top and occasional lightning flashes. However, when it is far away from a station, one may see lightning flashes only, but hear no thunder. But, as it draws near, lightning flashes become more frequent and one may also hear thunder from it. Since light waves travel a million times faster than sound waves through the air, the difference in their times of arrival gives an indication of the distance of the storm from a station at any time. Severe weather usually follows when the storm is directly over a station.

A thunderstorm derives its kinetic energy from the available potential energy stored in a conditionally unstable atmosphere in which the equivalent potential temperature decreases with height up to at least 3 km, and increases aloft. In such an atmosphere, cloud growth is limited to small-scale cumulus clouds only. Their further growth to higher levels is prevented by the strong static stability of the upper troposphere.

In the tropics, heat thunderstorms usually develop in late spring and early summer months. During that period, the temperature at the ground and the air layer immediately above it goes through a diurnal cycle, with a temperature inversion near the ground during late night and early morning hours and a steep temperature lapse rate developing during the daytime. So, it is during the late hours of the day that conditions become favorable for penetrative convection and rapid development of the cloud, resulting in the formation of a large cumulus or cumulonimbus cloud.

2.5.2.1 Cellular Structure of a Thundercloud and Vertical Currents

When viewed from outside, a thundercloud may appear as a single large cloud-mass with a diameter of a few kilometres (usually 10–12 km) and top rising to great heights, sometimes even beyond the tropopause at 16–18 km a.s.l., and lasting for a few hours. This may be compared with an individual fair weather cumulus
cloud which has an average diameter of about 1 km, a height of 1–2 km, and a life span of 15–20 min only. In reality, and as found by several aircraft probes during a thunderstorm project (Byers and Braham, 1949), there are several individual cloud cells inside the large cloud masses, all almost merged into one another in some kind of a conglomerate under a single canopy, each having a short lifespan of its own while being replaced by a new cell. Thus, the cloud cells inside a large developing cumulonimbus cloud are in violent agitation, somewhat like rising bubbles in a boiling cauldron. The rise and fall of the seething cloud cell tops can be clearly seen from an aircraft flying above the clouds. It is found that during the developing stage, updrafts dominate over downdrafts inside the cloud. But when precipitation starts falling through the cloud downdrafts increase and ultimately dominate over updrafts, thereby signalling the demise of the thunderstorm.

Direct observations or calculations of vertical air velocity inside a growing cumulus or cumulonimbus cloud are scanty. Fortunately, the Thunderstorm Project, referred to in the preceding para, studied the structure and circulation patterns of thunderclouds in various stages of development and found by direct aerial measurements that there is only updraft in the interior of these clouds till the precipitation stage is reached when updraft is replaced by downdraft in the portion of the cloud through which rain or snow falls. The general distribution of updraft velocity with height in the central portion of a growing cumulus cloud and a mature cumulonimbus cloud and the radial distribution of updraft velocity from the central region to the periphery of a cloud of horizontal extent of 8 km at the height of maximum updraft at about 9.0 km are shown in Fig. 2.9(a) and (b) respectively.

The vertical distribution Fig. 2.9(a) shows that there is a gradual increase of updraft velocity from a value of about 5 m s\(^{-1}\) at height 1.5 km to 20 m s\(^{-1}\) at 8–10 km and then a gradual decrease to low velocity aloft. The radial distribution (b) shows that the maximum updraft of about 18–20 m s\(^{-1}\) occurs near the center of

![Fig. 2.9](image)
the storm with a gradual decrease of the velocity to a value of 2–3 m s\(^{-1}\) at about 3.5 km from the center.

### 2.5.2.2 Hydrometeors Inside the Thundercloud

The updrafts in a thundercloud lift moisture to different levels of the atmosphere leading to formation of water substances in different phases of water. In the first stage, as moist air is lifted to levels above the condensation level at about 1.5 km, water vapour condenses on the existing cloud condensation nuclei to form cloud drops. Some of the drops increase in size as they are carried upward by collecting smaller drops and deposition of new water vapour round them but they remain in liquid form till they reach the freezing level at about 5 km. Above the freezing level, drops don’t freeze immediately, since they can remain in a supercooled state to a temperature as low as \(-40^\circ\)C. However, as they are lifted by strong updrafts to the cooler regions of the upper troposphere, an increasing number of them turn into snow crystals. Beyond a height of about 10 km, most of the snow crystals turn into ice crystals. In fact, the anvil part of a thundercloud near the tropopause consists mainly of ice crystals.

### 2.5.2.3 Precipitation, Downdraft and Squalls

The water substances inside the storm in the form of supercooled water drops, snow and ice crystals, after they grow to large sizes, start falling through the cloudmass under their own weight with a velocity which depends upon the updraft velocity and the aerodynamic properties and resistance of the environmental air through which they fall. As they fall through the freezing level, the ice and snow crystals begin melting by drawing heat from the warmer environment and thereby cool the air through which they fall by evaporation from their surfaces. The result is a downdraft of extremely cold air in the precipitating part of the cloudmass. On reaching the ground along with large drops of rain and sometimes pellets or hailstones, the downdraft spreads in all directions producing squally winds. In meteorology, a squall is defined as a strong wind characterized by a sudden rise of its velocity to 16 knots or more which is sustained for at least two minutes.

Squalls of abnormally high winds can do a lot of damage to lightly-built structures, uproot trees, and knock down living creatures out in the open.

### 2.5.2.4 Lightning and Thunder

Lightning and thunder are two most characteristic properties of a thunderstorm. Lightning is an electrical discharge that occurs between two oppositely charged regions inside a cumulonimbus cloud, or between the cloud and the earth’s surface. The basic requirement is a separation of electric charges between two parts of the cloud and build-up of a strong electrical field between them. By induction, the field produces a path of ion-molecules through which the discharge occurs.
More than half of all lightning discharges take place inside the cumulonimbus clouds and are known as intracloud discharges. But, lightning also occurs between the cloud and the ground which are known as streak or forked lightnings. Because of its intrinsic interest in the context of harmful effects upon living creatures by way of death and destruction, ignition of forest fires, and disturbances to power and electrical communications, etc., the cloud-to-ground lightning has been studied more extensively than any other form of lightning. Most of the electrical energy in a lightning discharge goes into producing heat. It is estimated that air along the path of a lightning discharge may be raised to a temperature exceeding 10,000°C. A part of this heat energy goes into producing radiation and light, while a part goes into producing sound waves through longitudinal compression and rarefaction which we hear as thunder.

Lightning activity over the globe is believed to contribute importantly to the maintenance of the earth’s electric field. It is well-known that the ionosphere which is positively charged loses its charge continuously to the ground which is negatively charged. This implies that unless replenished regularly, the ground will soon lose its negative charge and the atmosphere will lose its electric field. It is believed that this requirement is fulfilled by lightning activity over different parts of the globe, which supplies the much-needed negative charge from the cloud to the ground. The cloud also sends positive charge from its upper part to the ionosphere. The harmful effects of lightning strikes can be minimized by keeping indoors during a thunderstorm.

Fig. 2.10  Schematic showing the structure of a thunderstorm (after Newton, 1966). Arrow shows the direction of air motion. Symbols: A rectangle represents an ice pellet, a star a snowflake, and a circle a raindrop. F.L. stands for freezing level. Double-shaft arrow shows the direction of the wind believed to be steering the storm
2.5 Meso-Scale Disturbances and Severe Local Storms in the Tropics

is a common practice in most parts of the globe to erect lightning rods in buildings in order to protect them from possible damage from lightning strikes.

2.5.2.5 Structure of a Thunderstorm

Figure 2.10, after Newton (1966), is a sketch showing some of the many above-mentioned features of a mature thunderstorm, the top of which shoots up to a level well above the tropopause. It is likely that the structure will differ if the top remains below the tropopause level, but it is believed that the main features shown will remain similar.

2.5.2.6 Squall Line

A squall line is a line of active thunderstorms which may be continuous or with breaks over a length of 100 km or more. It is accompanied by all the characteristics usually associated with a thunderstorm, viz., strong updrafts and downdrafts, precipitation, lightning, thunder and squalls. It is a type of meso-scale convective severe local storm system which is usually associated with moving winter cold fronts and summertime troughs of airmass discontinuity. It is distinguished from other types of severe local storms by a smaller width-length ratio.

2.5.3 Hailstorms

A hailstorm is a mature thunderstorm from which significant hailstones fall to the ground. The size of hailstones can vary from 5 mm to 15 cm or more in diameter. Bilham and Relf (1937) quote a report of a severe hailstorm at Porter, Western Nebraska in USA, in 1928, in which the largest of the round hailstones was as big as a large grapefruit. It measured 43 cm in circumference indicating a diameter of about 13.7 cm, had a density of about half of that of water and weighed 681 g. A cumulonimbus is the only type of cloud that is known to yield hailstones.

Hailstorms are of intrinsic interest because of their destructive power. They are known to inflict damage to aircraft in flight and standing crops on the ground. They can also destroy plant and animal life. In tropical countries, such as India and Nigeria, hailstones can be of even larger sizes. There was once a report of a hailstorm in Rajasthan in India which wiped out a big herd of cattle comprising of giant-size buffaloes grazing in the field. On 27 May, 1959, an Indian Airlines aircraft flying at 5700 m a.s.l. near Delhi in India was caught up in a hailstorm and the hailstones encountered caused holes in the aircraft, the largest of which had a diameter of about 15 cm (Saha, 1962). The clean-cut of the holes which were practically spherical in shape would give an impression that the hailstones that caused these holes were of comparable size, although the impression may not be quite justified in the absence of data based upon relationship between the size of the hailstone and that of the hole in different ranges of size and energy of the hailstone. Wegener (1911) recorded that the largest known hailstone had a weight of 1 kg and the size
of a skittle ball (Kegelkugel), indicating a diameter of about 14 cm if a mean density of 0.7 g/c.c. were assumed.

Because of the potential dangers attached to hailstones, they have drawn the attention of meteorologists from early times. However, little definite is known as to how hailstones form and grow inside a thundercloud. Several theories and hypotheses have been advanced in this regard. For a summary of some of these studies, the reader is referred to an article by the author (Saha, 1962).

A plausible theory of the formation of large hailstones must take into account the following facts of observation, reiterated below:

1. A hailstone consists of concentric shells of opaque and clear ice.
2. It forms in convective clouds the tops of which reach great heights in the atmosphere. In the tropics, the tops of cumulonimbus clouds have been known to extend to 12–5 km a.s.l. or even greater heights.
3. The mean distribution of updrafts in towering cumulus and cumulonimbus clouds is such that there is a gradual increase of updraft velocity from a value of about 2.5 m s\(^{-1}\) at height 1.5 km to about 10 m s\(^{-1}\) at 8–10 km and then a gradual decrease to lower velocity aloft (Byers and Braham, 1949).
4. All thunderstorms do not precipitate hailstones, although it is generally agreed that the evolution of an ice phase is a common feature of all thunderclouds. Observations show that only a very small proportion of even the severest thunderstorms yield large hailstones.
5. There is a pronounced geographical and seasonal variation in the distribution of hailstones. In general, thunderstorms occurring in particular areas and seasons only yield hailstones.
6. In the size-spectrum, hailstones of very large diameter are found to occur in low concentration (1/m\(^3\) or less).
7. Most often, hailstones are observed to fall earlier than raindrops. Sometimes they fall simultaneously.
8. Hail starts rather suddenly and usually lasts from a few minutes to less than half an hour.
9. There may be several ups and downs inside and outside the cloud before a hailstone comes to the ground.
10. The presence of a large thickness of cloudmass between the base of the cloud and the level of freezing of large clouddrops prevents the growth of hailstones by prematurely precipitating out the larger drops in the form of heavy rain.

There have been several theories regarding the formation of large hailstones in a hailstorm, but one that takes care of most of the above-mentioned facts of observation and rather promising is one that was advanced by Schumann (1938). He gave the following equation for the ultimate diameter of a hailstone:

\[
L = (\sigma/m)(2D - 6) - (4\mu/m)(3k\rho\sigma/\pi g)^{1/2} \times \{(2D)^{1/2} - 2.45\}
\]  
\[ (2.5.1) \]
where

\[ L \] is the depth of fall of the hailstone above the freezing level,
\[ \sigma \] the average hailstone density,
\[ m \] the concentration of condensed water,
\[ u \] the updraft velocity,
\[ \kappa \] the Reynolds number,
\[ \rho \] the air density,
\[ g \] the acceleration due to gravity, and
\[ D \] the ultimate diameter of the hailstone, the initial diameter being assumed 3.0 cm at height 9.0 km a.s.l. which is the mean height of maximum updraft.

Using the above equation and the following values (in c.g.s. system) of the above-mentioned parameters,
\[ L = 4.5 \times 10^5 \text{ cm}, \quad \sigma = 0.7 \text{ gm (c.c)}^{-1}, \quad m = 20 \times 10^{-6} \text{ gm (c.c)}^{-1}, \]
\[ u = 2000 \text{ cm s}^{-1}, \quad \kappa = 0.1, \quad \rho = 5 \times 10^{-4} \text{ gm (cm)}^{-3}, \quad \text{and} \quad g = 981 \text{ cm s}^{-2}, \]
Schumann obtained a value of 12.1 cm for \( D \).

The present position regarding these theories is that several questions relating to formation of large hailstones remain unanswered. In the absence of direct observations on many of the parameters involved, it is rather difficult to discuss the relative merits and demerits of the various theories that have been advanced from time to time. It is, however, generally agreed that strong updraft promotes growth of large hailstones.

But it needs to be emphasized that the size of a hailstone that is balanced by updraft is the maximum size with which it leaves the cloud or reaches the ground. The growth of a hailstone beyond the size which can be supported by updraft occurs while it descends through a large mass of supercooled drops and snowflakes in the supercooled region of the cloud. This appears to be an all-important point in any plausible theory of formation of large hailstones.

Regarding the observed structure of hailstones consisting of alternate shells of white ice and clear ice, there have been two main schools of thought. One led by Humphreys (1940) holds the view that the formation of alternate shells of opaque and clear ice is due to the alternating movements of a frozen cloud or rain drop between the realms of rain and snow across the freezing level by vertical aircurrents. The other theory, due to Gaviola and Fuertes (1947), seeks to explain the observed structure by assuming that while the surface is wet, a transparent layer forms, and when it dries, an opaque white layer forms.

### 2.5.4 Tornadoes

A tornado is a mesoscale violently-rotating atmospheric vortex protruding downward from the base of a large cumulonimbus cloud in the shape of a funnel which often reaches the ground with disastrous effects on life and property. It occurs mostly during passage of a heavy thunderstorm or squall line over a locality. Because of its
funnel shape, it is also sometimes called a funnel cloud. When appearing over water, it is called a water spout.

The typical characteristics of a full-grown tornado vortex as revealed by observational and theoretical studies are the following:

(a) A diameter of 100–200 m, a variable depth below the cloud base with intermittent touch-downs, a life-span varying from an hour to several hours, and an erratic path varying in length from 1 to 100 km;
(b) Revolving winds with high positive relative vorticity and updraft locally inside the cloud, as found in a helical vortex;
(c) A maximum tangential velocity, 100–150 m s\(^{-1}\);
(d) A maximum updraft velocity, 100–150 m s\(^{-1}\);
(e) Extremely low pressure and high temperature inside the vertical column;
(f) Entrainment of environmental air at explosive rate;
(g) Frequent lightning flashes and increasingly thunderous and roaring sound at the approach of a tornado;
(h) Often heavy precipitation following the passage of a tornado.

It should, however, be emphasized that there have been few direct observations made inside a live tornado, because of the obvious dangers to human lives and inability of any meteorological instruments to withstand the impact of the high revolving winds. This means that most of our current information about tornadoes is derived indirectly from detailed post-mortem examination of death and destruction it leaves behind.

[The author once investigated a tornado which occurred in the district of Cooch Bihar in eastern India in 1963. The indirectly estimated values of some of the above-mentioned characteristics are quoted mostly from the findings of that study.]

2.5.4.1 Tornado Circulation and Intensity

Circulations in a tornado, both horizontal and vertical, are sometimes so strong that they can uproot large trees, bend electric poles, knock down buildings and houses, wipe out crop fields, pick up heavy objects from the ground and throw them around like deadly projectiles, derail running trains, sink boats and ships plying in rivers, and occasionally lift boats from river banks and throw them into water, and so on. There was a report of a tornado which while moving over a house in eastern India picked up a young girl of about 10 years from the compound of the house and later deposited her safely a few hundred metres down the path. Further down, it sucked up all the muddy water from a wide canal 1.2 m deep and deposited it all along its route. The direction of the circulation and probable speed of the horizontal and vertical components of the wind are usually estimated from the extent and intensity of these damages and destructions. Casualties to human and animal life are usually caused by collapse of poorly-built houses and by their exposure to heavy metallic projectiles when exposed in the open. In USA, windspeeds are sometimes estimated on the basis of observed damages using a scale known as the Fujita scale (Fujita,
1981). The Fujita scale (also known as F-scale) is a six-point scale which, according to the Meteorological Glossary (second edition) of the American Meteorological Society (2000), corresponds to the following wind-speed estimates:

Further clarification of the Fujita-scale is furnished by the following descriptions:

- **F-0** Slight damage to chimneys; branches broken; shallow-rooted trees knocked over.
- **F-1** Surface of roofs peeled off; mobile homes pushed off foundations or overturned; moving autos pushed off roads.
- **F-2** Roofs torn off frame houses; mobile homes demolished; boxcars pushed over; large trees snapped or uprooted.
- **F-3** Roofs and some walls torn off well-constructed houses; trains overturned; most trees in forests uprooted; heavy cars lifted off ground and thrown.
- **F-4** Well-constructed houses leveled; structures with weak foundations blown off; large missiles generated.
- **F-5** Strong frame houses lifted off foundations and carried considerable distances; automobile-sized missiles flying through the air for distances in excess of 100 m; trees debarked.

### 2.5.4.2 Geographical and Seasonal Distribution of Tornadoes

Although tornadoes occur in most parts of the globe, they seem to have a strong geographical and seasonal bias. In USA where their frequency of occurrence per unit area appears to be the highest in the world with more than 1000 per year, majority of the tornadoes form over the Great Plains of the midwest comprising the States of Nebraska, Kansas, Oklahoma, and north Texas, and the southeastern States of Mississippi, Alabama, Louisiana and Florida. This continuous zone of the States is often referred to as the tornado alley. They can occur throughout the year at any time of the day but the maximum frequency is during spring and early summer. Multiple tornadoes have been observed in association with certain types of synoptic-scale disturbances. They usually move from west to east with prevailing upper winds. In West Africa where they occur mostly during summer, they move from east to west. In eastern India and Bangladesh, they occur mostly during spring and early summer months and are known as ‘Kal-Baisakhis’ or deadly storms of the month of Baisakh (April–May) and usually move from a NW-ly direction.
Observations appear to suggest that tornadoes tend to form in thunderclouds in which a large number of ‘mammatos’ appear at the cloudbase suggesting the presence of strong vertical currents inside them. The exact mechanism of how the vertical currents organize themselves to generate a tornado, however, is not known and open to speculations. One school of thought views a tornado as a meso-scale revolving storm with a horizontal and vertical structure very similar to that of a mature miniature tropical cyclone.
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