INITIAL STAGES AND SOME CHARACTERISTICS OF WEST AFRICAN LINE SQUALLS

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by

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DEDICATION

To

God for being, as ever, faithful to His promises, my father and late mother without whose prayers and support, this work would have been impossible.

ABSTRACT

A review of West African synoptic weather pattern reveals that the sub-region experiences a special kind of atmospheric disturbance - the line squall - whenever the south-westerlies cover, approximately, the whole of Nigeria. Various methods that have been used to study the squalls (i.e. observational investigations, satellite investigations and modelling) have not been very successful in isolating the trigger mechanism of the phenomenon.

It is been proposed that line squalls are initiated through the amplification (with time) of any wave-like perturbation along the surface of discontinuity between the south-westerlies and the north-easterlies. The amplifying perturbations could block the 650 mb. mid-tropospheric jet which further distorts the 'bump' formed by the undulating perturbation. This distortion forces the south-westerlies further up and they could condense. The precipitates fall into the underlying, dry jet and some of them evaporate; the latent heat of evaporation being supplied by the jet. The jet, now cooler, sinks. On sinking, the jet could hit the surface of the earth on which it forms the squall front and crawls; thereby lifting the south-westerlies ahead of it. The cycle of condensation, evaporation and sinking then continunes.

A gravity-wave model of this mechanism is presented through the solution of a frequency equation with the aid of a two-layer atmospheric model. The solution is an eigenvalue problem from which

many modes of different growth rates and phase velocities could be obtained. Some of these phase velocities will be complex - the real part representing the phase velocity while the imaginary part represents the amplification. Waves with the largest amplifications (i.e. the largest imaginary part) are those that could possibly block the 650 mb. mid-tropospheric jet and trigger off line squalls.

Among others, this proposal on the trigger mechanism of line squalls is able to explain:

(i) the preference of highlands as places of origin of line squalls,

(ii) the close association between the speeds of propagation of line squalls and the mid-tropospheric jet and

(iii) the observed overturning of the atmosphere after the passage of line squalls.

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Finally, I most sincerely thank my wife, Bola, for her patience, loyalty and co-operation throughout the period of this work.

CERTIFICATION

This is to certify that the work described in

this thesis was carried out under

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NOTATIONS AND SYMBOLS

u	x - component of velocity
v	y - component of velocity
w	vertical velocity
p	pressure
ω	$\frac{Dp}{Dt}$
ρ	density of air
g	acceleration due to gravity
ν	coefficient of viscosity
f	Corioli's parameter
s	twice the angular velocity of the earth
L	latent heat of vaporisation
H	scale height
r	condensed water vapour per unit mass of air parcel
q	specific humidity
C_p	specific heat at constant pressure
C_v	specific heat at constant volume
θ	potential temperature

κ	$\dfrac{C_p - C_v}{C_p}$
T	temperature
z	vertical height
σ	frequency of waves
α	wave number along x-direction
β	wave number along y-direction
R	gas constant

CHAPTER 1

INTRODUCTION

The weather pattern in West Africa could be classified into two main seasons: the wet and the dry. One of the distinguishing factors between the seasons is the amount of rainfall. Almost all the rainfall within the sub-region are recorded during the wet season which is approximately from April to September.

About half of the total annual rainfall recorded in any particular location within West Africa is due to the precipitation which usually accompanies isolated thermal convection in the atmosphere (Obasi, 1976). Such pockets of convective processes are referred to as LOCAL CONVECTIVE STORMS. As expected, the frequency and intensity of local convective storms vary from place to place; hence, total annual rainfall is not uniform in the sub-region.

Apart from local convective storms, there exists a special kind of storm that travels and thus covers a large area of land whenever it occurs. These 'travelling' storms are also accompanied by heavy rainfall. More than half of the total annual rainfall in West Africa is attributable to these storms which are known as LINE SQUALLS. The extent of 'travel' of line squalls (about two to three nights of 'travelling' at an average rate of 15 m/s (Fortune, 1977)) puts them in the class of mesoscale systems.

Line squalls can be defined in various ways depending on the mode of identification and place of occurrence. However, the definition given by Zipser (1977) seems all-encompassing and therefore adequate. According to him, line squalls are:

> cumulonimbus clouds, organised in linear fashion, associated with a pseudo-cold front (squall front) at the surface, propagating with considerable speed with respect to the ambient low-level air, in the general direction of the squall wind in the cold air behind the squall front.

Line squalls have been referred to by various names at different times and places. Hamilton and Archbold (1945) called them DISTURBANCE LINES (D.L.'s) because the squalls occur along a fictitious bow-shaped line that has N-W to S-E orientation (Fig. 1). Other names are: chubascos (in Central America), haboobs (in Sudan) and sumatras (in Malaysia).

These various names indicate the geographical spread of the occurrence of line squalls. Thus, they are not entirely a local West African (or tropical) sub-synoptic phenomenon. Marriot (1892), Prohaska (1907) and Browning and Ludlam (1962) reported the existence of this kind of storms in the mid-latitudes. Although mid-latitude and tropical line squalls propagate at speeds close to the speed of the mid-tropospheric winds in their respective regions (i.e about 15 m/s in West Africa), they differ in structure because the anvil in the latter extends behind rather than in front of the convective elements

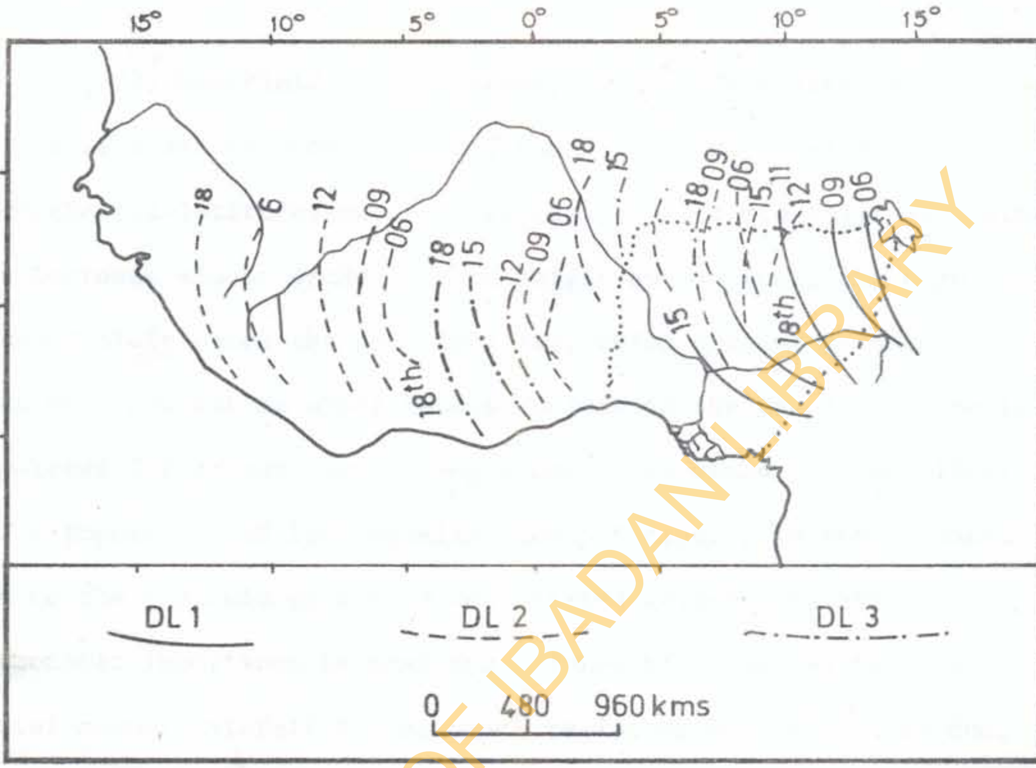

Fig. 1: Three disturbance lines (DL 1, DL 2, DL 3) in West Africa (the fictitious bow-shaped lines propagate westwards across West Africa)
(from Garnier, 1967)

(Fortune, 1977; Mansfield, 1977; Zipser, 1977). This difference in structure is a direct consequence of the variations in wind profile between the mid-latitudes and the tropics. In the mid-latitudes, wind speeds increase almost steadily with height and as such, the high wind speed aloft makes the updraught air, which originates ahead of the storms, flow out at upper levels whereas in the tropics, there is a low-level jet rather than strong upper level winds (Bolton, 1981).

The importance of line squalls lies principally in their contribution to the economic growth of West Africa (vis-a-vis agriculture). This economic importance is most appreciated if it is realised that the total annual rainfall in the Sahel region of West Africa is due, mainly, to these storms. This implies that years of little squall activities are usually dry in the Sahel. The consequent drought leads to loss of life (both human and animal) and poor agricultural output.

Apart from their economic and social implications, hazards in the field of aviation is another reason why studies on the initiation, maturity and dissipation of West African line squalls have engaged the attention of researchers for almost four decades now (Hamilton and Archbold, 1945; Dhonneur, 1970; LeRoux, 1976; Fortune, 1977; Okulaja, 1978; Bolton, 1981). On the global scale, the contribution of the phenomenon to the overall transfer of mass, energy and momentum is of interest to meteorologists. In chapters 2 and 3, reviews of some of the methods that have been employed in the study of line squalls are presented. Broadly, these methods could be classified into three: observational investigations, satellite investigations and modelling.

Whichever method is used to study line squalls however, complete theories on them must account for their growth, the fully-developed stage and dissipation. Furthermore, such theories must account for their orientation, direction of movement and the almost-uniform speed of propagation that is common to West African line squalls. It is also desirable to explain the limitation of line squalls to a particular season of the year and the similarities between a local convective storm and line squall.

Almost all features of the fully-developed (or mature) line squall have been documented (Hamilton and Archbold, 1945; Mansfield, 1977; Bolton, 1981). Outside West Africa, but within the tropics, the following works on line squalls are notable: Zipser (1969, 1977), Belts, Grover and Moncrieff (1976) and Miller and Betts (1977) while the description of the Wokingham storm by Browning and Ludlam (1962) gave a good picture of mid-latitude 'travelling' storms. Data from satellite photographs have also added to our knowledge of the matured line squall.

In spite of these works, however, there is still some doubt as to the meteorological factors on which the origin of line squalls are dependent. More precisely, the trigger mechanism which at verious times have been linked with insolation, orography and synoptic convergence, is still not fully understood. This study has, therefore, been directed towards the initial stages in the development of line squalls; paying special attention to the trigger mechanism. In chapter 4, the mechanism of initiation of line squalls, along with a description of the atmospheric features during the matured stages, are described.

The process of initiation of line squalls is mathematically modelled in chapters 5 and 6. A frequency equation is derived in chapter 5 from the basic set of hydrodynamical equations of motion. This frequency equation is solved in chapter 6. Since line squall is not a single mode but rather an amplifying patch of convection that consists of many modes of different growth rates and phase velocities (Bolton, 1981), solutions to the frequency equation are expected to reveal modes with the greatest amplifications. It is proposed that such modes play vital roles in the initiation of line squalls.

CHAPTER 2

A REVIEW OF LINE SQUALL OBSERVATIONS

2-1 West African synoptic weather pattern

As a background to understanding the various stages in the development and study of West African line squalls, a brief review of West African synoptic weather pattern is essential.

The air masses and the prevailing winds in West Africa are shown in figure 2. One of these air masses is labelled the MONSOON. According to Hamilton and Archbold (1945), the monsoon winds originate off the coast of South Africa and blow south-easterly for some time before they veer to their direction of south-west in which they arrive at the coast of West Africa. The long track of these winds over the Atlantic ocean accounts for their considerable moisture content by the time they arrive at the southern part of the sub-region. On the other hand, HARMATTAN WINDS blow across the northern part of West Africa. In contrast to the monsoon south-westerlies, this air mass is very dry because of its long track over the Sahera desert.

The north-easterlies and the south-westerlies meet at a zone of strong convergence called the Inter-Tropical Convergence Zone (ITCZ) (Fig. 2). Other names for this zone are: Inter-Tropical Front (ITF) and Inter-Tropical Discontinuity (ITD). The ITCZ moves North or South in sympathy (but with a phase lag of about one and a half months (Bolton, 1981)) with the motion of the sun between the tropics of Cancer and

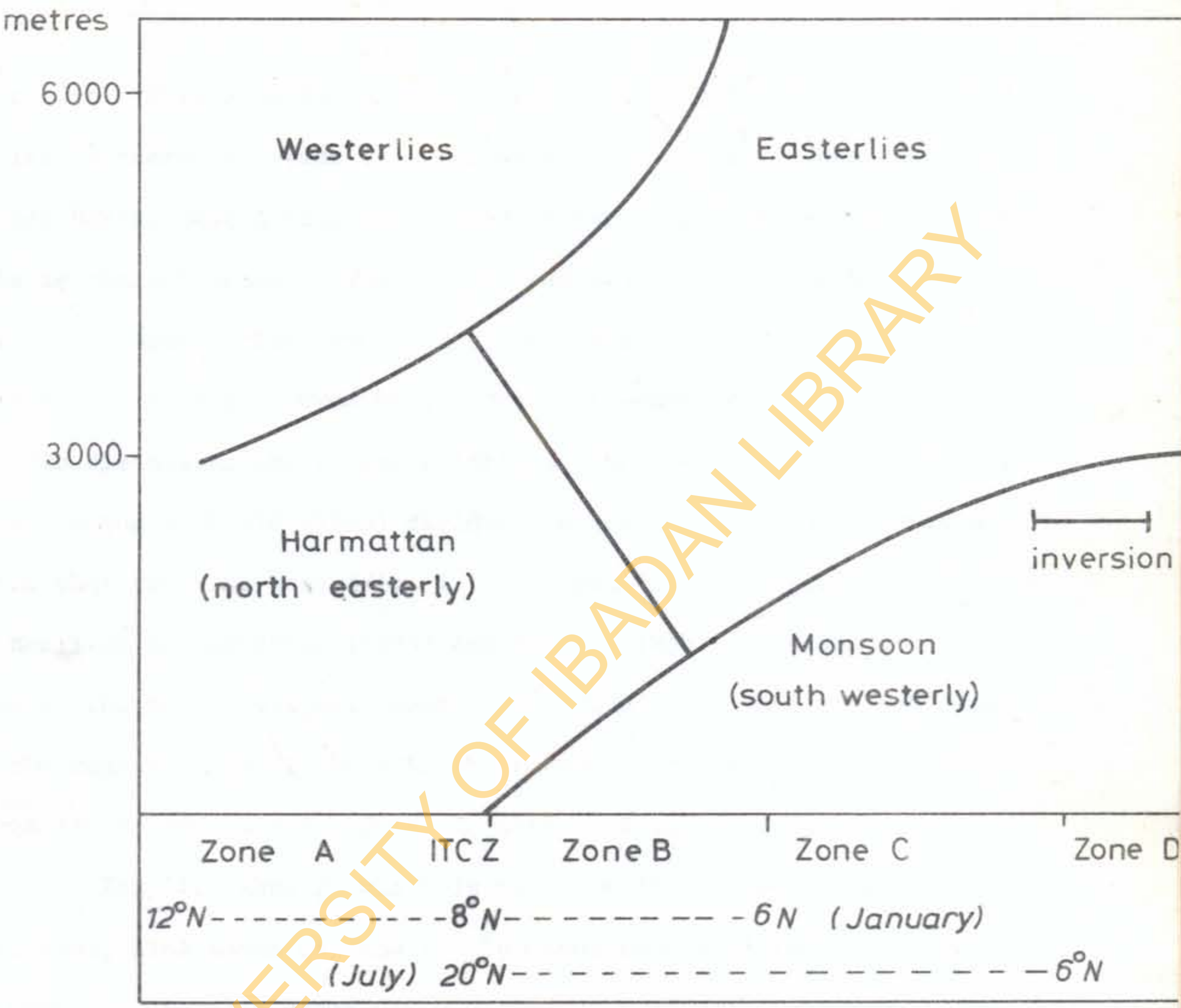

Fig. 2: North - South section of the prevailing winds over West Africa

(from Hamilton and Archbold, 1945)

Capricorn. This zone is farthest North (about 20°N) around July and at its southermost extent (about 8°N) around January. When the zone is far North, West Africa is covered by the monsoon south-westerlies; this is the wet season. Towards the end of the year when the zone is far South, West Africa experiences the dry season because the sub-region is almost entirely covered by the dry north-easterly harmattan winds.

In discussing the characteristic weather pattern of West Africa, Hamilton and Archbold (1945) divided the sub-region into four weather zones that run East-West (Fig. 2). Balogun (1974) adopted the suggestions of Hamilton and Archbold (1945) and Walker (1959) and added a fifth zone to the four previously mentioned. Zone A of Hamilton and Archbold is the region north of the ITCZ while Zone B extends from the ITCZ to about 150 km southwards. Zone C, with a width of about 500 km lies south of Zone B. Zone D, which is over the land between July and September, lies south of Zone C. In other months of the year, Zone D is over the Atlantic ocean.

Within Zone D, there is an isothermal (or inversion) layer between 800 mb. and 850 mb. pressure levels. As we shall discuss later, the existence of this inversion might have something to do with the observed dissipation of line squalls around the coasts of West Africa. Zone C is distinguished from other zones because it is periodically traversed from East to West by D.L.'s (Hamilton and Archbold, 1945).

According to Garnier (1967), the movement of the ITCZ controls both the number and duration of the weather types experienced in different parts of West Africa. Broadly, the resulting pattern is a latitudinal one. For instance, Ibadan (7^{o} 26' N 3^{o} 54' E) experiences weather conditions in the following way:

Zone A - late December to part of January

Zone B - February and part, or all of March

Zone C - April (or part of March) to about mid or late July

Zone D - late July and part, or all of August

Zone C - late August to end of October or early November

Zone B - briefly in November and early December

Zone A - late December to part of January

Since Zone C is, as said earlier, traversed by D.L.'s, a station like Ibadan experiences intense D.L. activities twice in a year as the break-down of weather conditions given above shows. This explains the two peaks observed on the chart of monthly frequency of line squalls at some Nigerian stations (Fig. 3). The peaks coincide with the periods when the stations experience Zone C weather conditions. In section 4-1, it is shown that all the essential atmospheric features that favour the development of persistent storms are present at such stations during

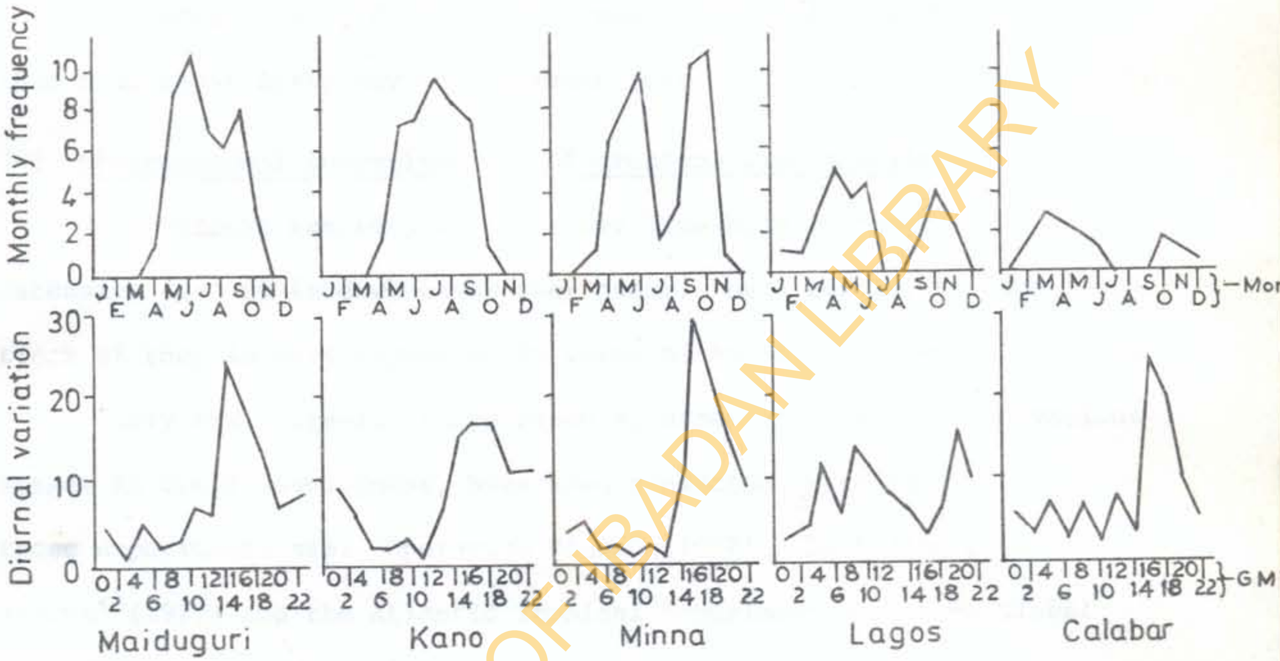

Fig.3: Monthly frequency and diurnal variation of squalls over 15 m/s at various Nigerian stations

(from Hamilton and Archbold, 1945)

the periods indicated by the peaks. Hence, it is not a coincidence that line squalls are, for instance, most-frequently observed in Ibadan at about April/May and September/October.

2-2 Observational investigations of tropical line squalls

As mentioned earlier, West African line squalls cover an extensive area of land whenever they occur. As a result, keeping track of them is very expensive in terms of human and material resources. Hence, very few large-scale experiments, aimed at analysing the various stages in their development, have been conducted up to date. Some of these experiments are: 'Operation Niger' (1972), 'Operation pre-GATE ASECNA*[1] (1973) and the Atlantic Tropical Experiment under the Global Atmospheric Research Programme GATE) (1974). Outside West Africa, but within the tropics, some experiments that are equivalent in scale are: Line Islands Experiment (1967), Barbados Experiment (1968) and the Venezuelan International Meteorological and Hydrological Experiment (VIMHEX) (1972). Before we review these experiments, the worthy contributions of a few individuals who have studies various aspects in the evolution of line squalls would be mentioned.

(a) Hamilton and Archbold (1945)

Hamilton and Archbold studied the meteorology of Nigeria and adjacent territories. In their report, they mentioned that D.L.'s

*ASECNA - Agence pour la Securite de la Navigation Aerienne could be identified on weather charts as species of fronts which move in a west-south-westerly (WSW) direction with a speed of 10-15 m/s. The authors describe how a D.L. manifests itself to a surface observer thus:

> Beforehand we have normal south-westerlies with normal or almost normal cloud development appropriate to the time of day. Then in the east we see a dark heavy band of Cb. approaching. This will be heralded at night by an impressive display of lightning which can be seen while the storm is as much as 100 miles (160 km) away... In the case of an active D.L. the Cb. will certainly reach to 20,000 feet (6000m) or more probably to 30,000 feet (9000m) and may even be much higher ...
>
> At the forward edge of the cloud base there is a well-marked roll of low cloud, roughly between the levels 600 and 3000 feet (200 - 1000m) though the top is not always plainly seen. Just before the arrival of the roll cloud the surface wind usually falls light or calm. When this cloud passes overhead, there is a sudden squall from between S.E and N.E. At Lagos the anemometer commonly records a squall of 30-40 m.p.h. (12 - 20 m/s); stronger gusts are experienced in the north but more than 60 m.p.h. (25 m/s) is uncommon though possible.
>
> The squall is the last warning to take cover. Two or three minutes later, a heavy downpour of rain sets in, reducing visibility to 1000 - 3000 yards (about 1000 - 3000m) or less and effectively to nil for pilots inside aircraft. At the same time the cloud base becomes ragged and very low ...
>
> About 2 to 3 hours after the beginning of the storm the rain ceases entirely and the surface wind falls calm or blows lightly once more from S.E. to S.W. After another hour the medium cloud lifts further and begins to break.

Hamilton and Archbold did not find any general method of forecasting the mode, place and time of origin of D.L.'s. We know, however, from their work that the passage of line squalls, whether over land or the ocean, overturns, and therefore stabilises, the atmosphere. After the passage, the surface dry-bulb temperature usually approximate the 700 mb. (or higher) wet-bulb potential temperature before the squall.

Analyses of the D.L. development by these authors, though commendable, was limited in scope because they relied mostly on ground meteorological stations and these were sparse at the time of their study. The situation had not markedly improve at the time Bolton (1981) carried out his investigations despite a time interval of over thirty years.

(b) Bolton (1981)

Bolton (1981) studied line squalls that affected Minna (9^{o} 37' N 6^{o} 34' E) over a period of about two years using mainly surface data. Because it is difficult for a single observer to distinguish between local convective storms and line squalls, Bolton set out the following criteria for identifying the latter:

(i) maximum squall speed at least 12 m/s,

(ii) rainfall at least 5mm,

(iii) storm duration at least 3h and

(iv) pressure jump at least 0.5 mb.

Using these conditions, sixty-six line squalls were identified during the period of study. The speeds of propagation of these storms agree well with the results of Hamilton and Archbold. Plates 1(a)-1(d) show the onset of a typical West African line squall as recorded by Bolton.

The author associated the trigger mechanism of line squalls with solar heating and orography; specifically mentioning that a combination of the two might be necessary before these storms develop. However, as he correctly pointed out, the fact that line squalls are generated over the sea indicates that these two mechanisms might not be very essential. There must then be other mode(s) of generation of line squalls. Large-scale experiments, which we shall now consider, did not fair better than individual contributions as far as isolating the mode(s) of generation of line squalls is concerned.

(c) 'Operation Niger' (1972)

This experiment, conducted on the bend of river Niger (0-5^{o}W) in mid-July 1972, was limited in affair and throughout the period, no long-lived D.L. was observed. Analysis of the only D.L. that passed through the experimental array revealed a weakening of the mid-level jet is the viscinity of the disturbance. From other works (Zipser, 1977; Bolton, 1981) and our own studies, the weakening of this jet is likely due to the fact that it descends to the surface of the earth in the course of initiation of line squalls. More will be said on this in chapter 4.

Plate 1(a): Line squall at Minna on 12 September 1977. View to the E at 1555 GMT showing advancing line of cumulonimbus (from Bolton, 1981).

Plate 1(b): Line squall at Minna on 12 September 1977. View to the NE at 1611 GMT showing roll cloud (from Bolton, 1981).

Plate 1(c): Line squall at Minna on 12 September 1977. View to SSE at 1613 GMT showing front of cumulonimbus belt and edge of rain area. (from Bolton, 1981)

Plate 1(d): Line squall at Minna on 12 September 1977. View to W at 1648 GMT showing individual cumulonimbus (from Bolton, 1981)

(d) 'Operation pre-GATE ASECNA' (1973)

This operation, centred in Bamako (Mali), was an intense analysis of ten days of observation from the sparse network of meteorological stations in West Africa. Five major disturbances (three of them line squalls) intercepted the array during the experiment. Some of the findings of the experiment are:

(i) line squalls are formed often to the east of latitude 5^{o}E and they cover about 1200 km. with an average speed of close to 15 m/s,

(ii) precipitation distribution is highly irregular,

(iii) there is a noticeable jump in the surface pressure upon the arrival of line squalls and

(iv) the atmosphere is homogenised after the passage of line squalls from a pre-squall stage of sharp contrast between low and mid-level air masses.

(e) GATE (1974)

This is, perhaps, the most intensive investigation conducted up to date on West Africa disturbances. The experiment made use of radar, radiosonde ascents, aircraft traverses and satellite data to study the storms over the eastern Atlantic (i.e. off the coast of Senegal). Data collected during the experiment are still being analysed by various scientists. We shall only mention the works of

Aspliden, Tourre and Sabine (1976) and Mansfield (1977) because this study has been able to justify some of their observations.

(i) <u>Aspliden et.al. (1976)</u>

These authors identified a total of 176 active D.L.'s during the three phases of GATE (i.e. 28 June - 16 July, 28 July - 15 August and 30 August - 19 September). The criteria for identification are:

- the D.L. should consist of large cumulonimbus clouds which tops to about 15 km (this implies that only the brightest cloud images on the geo-stationary Synchronous Meteorological Satellite - I (SMS - I) were considered; although not all cloud clusters qualify to be D.L.'s)
- such bright clouds should attain a size of at least $2^{\circ} \times 2^{\circ}$ during their life time
- the D.L. should be active for a minimum of 6h.

Very few (about 8%) of all the line squalls observed during GATE, as recorded by Aspliden et.al., formed over the ocean. Majority were generated and decayed over land. Figure 4 shows the number of disturbances which were generated as well as these that decayed in each 5° square over the continent and ocean during GATE.

Although these authors did not explain the process of formation of line squalls, we know from their work (Fig. 4) that the preferential

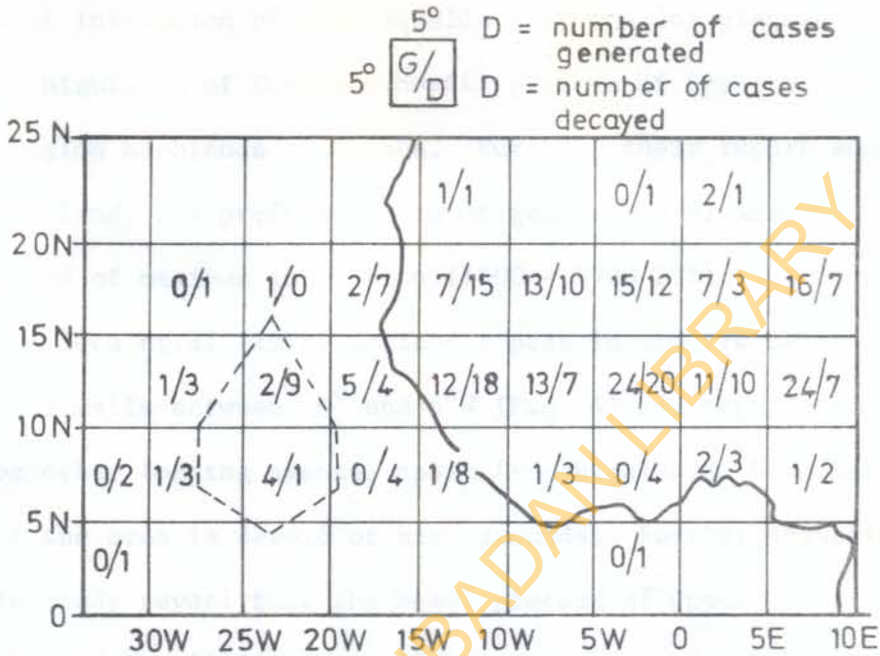

Fig. 4: Number of disturbance lines which generated and decayed in each 5° square over the continent and ocean during GATE

(from Aspliden et al.,1976)

places of initiation of line squalls are: the Jos plateau of Nigeria/ Adamawa highlands of Cameroun,Mossi: plateau of Upper Volta and the Fouta Djalon highlands of Guinea. Further, their report shows that, over the land, the preferred time of generation of line squalls is the period of maximum insolation (1400 - 1700 LST).

Aspliden et.al (1976) noticed a peak in the frequency of generation of line squalls between 0° and 5°W (Fig. 4) and postulated that there must be other forcing agents, apart from highlands, because they thought the area is devoid of high grounds. Further investigations in this study reveal that the Mossi plateau of Upper Volta is within the square 0-5W, 10-15N (Fig. 4) and this probably explains the observed peak in the frequency of generation of line squalls.

(ii) Mansfield (1977)

This author studied four occurrences of line squalls during phase phase III of GATE. As his time-height section of W-E wind component during the period shows (Fig. 5), the speed of the 650 mb. mid-tropospheric jet is usually a maximum just before the onset of line squalls. He then proposed that the attainment of the maximum speed within the troposphere of this jet, prior to the occurrence of line squalls, might have something to do with the initiation processes of these storms. From this study, such a link is still obscure.

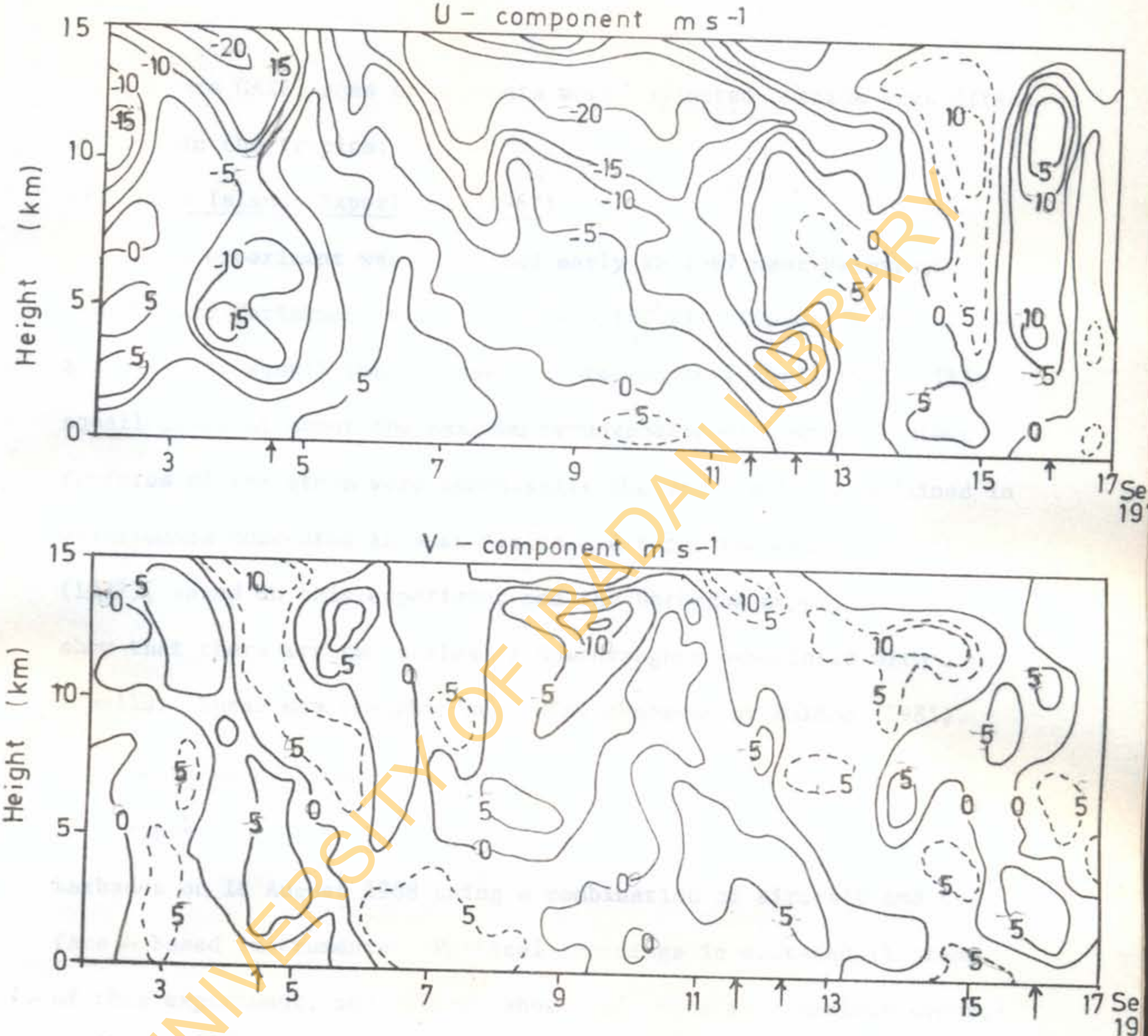

Fig. 5: Time-height section of wind component over a GATE ship 2 - 17 September 1974.
(from Mansfield, 1977)

Before GATE, some experiments were conducted outside West Africa but within the tropics.

(f) Line Islands Experiment (1967)

This experiment was conducted early in 1967 near Palmyra, Fanning and Christmas Islands (150^{o}W - 170^{o}W). Zipser (1969) analysed a particular squall that crossed the region on 1 April 1967. This squall moved at about the maximum tropospheric wind speed. Other features of the storm were essentially the same as those obtained in experiments conducted in West Africa. A separate analysis by Zipser (1977), based on this experiment and the Barbados experiment of 1968, show that there are two scales of downdraughts associated with line squalls. These downdraughts were also observed by Bolton (1981).

(g) Barbados Experiment (1968)

Zipser (1977) intensively studied a line squall that passed Barbados on 18 August 1968 using a combination of aircraft and surface - based instruments. Vertical soundings in post-squall areas of this experiment, and others, show that there is a maximum separation between temperature, T, and dew point, Td, at about 900 mb. When the soundings approached the base of the anvil, which is slightly above 600 mb., this separation narrows and thereby create an 'onion' shape for characteristic soundings in post-squall areas (Fig. 6).

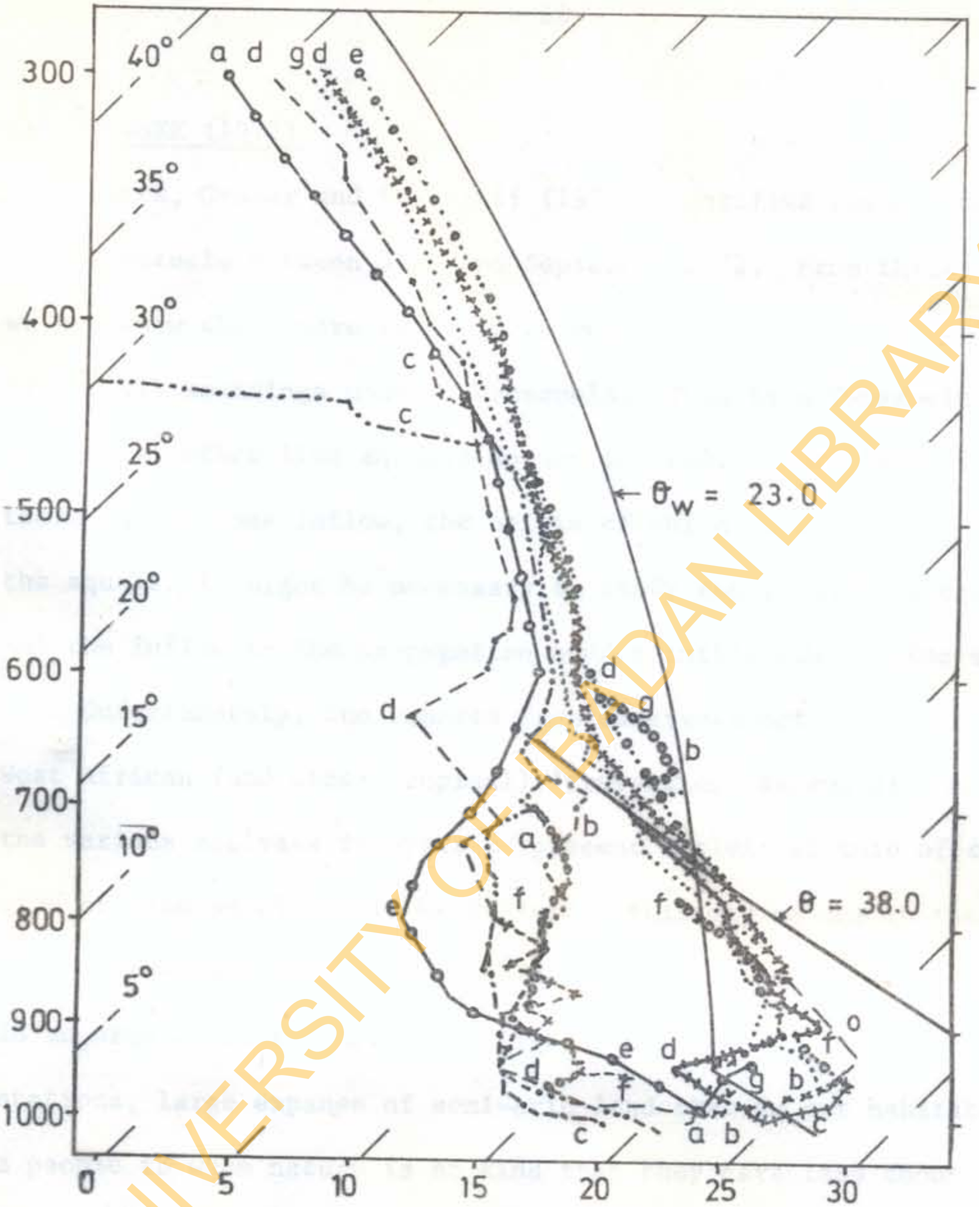

Fig. 6: Tephigram showing T, T_d for various post-squall soundings

(a) GATE Queen air sounding, mostly in rain, 5 Sept. 1974.
(b)(c)(d) Various GATE oceanic soundings, 12 Sept. 1974
(e)(f) Venezuelan soundings, 28 Aug. 1969.
(g) Barbados sounding (no T_d available), 18 Aug. 1968

(from Zipser, 1977)

(h) VIMHEX (1972)

Betts, Grover and Moncrieff (1976) identified seven line squalls over Venezuela between June and September 1972. From their report, we now know that there is a single reversal of the shear at 700 mb. in pre-squall soundings made in Venezuela. This is a Venezuela feat which most other line squalls do not possess. Apart from the reversal, there exists some inflow, the origin of which is not clear, behind the squall. It might be necessary to study the role(s) of the reversal and the inflow in the propagation and/or initiation of line squalls.

Unfortunately, the reports from observational investigations of West African (and other tropical) 'travelling' storms did not synthesise the various analyses in order to present a clear picture of the evolution of line squalls. In West Africa, this may be due to the enormity of the problems involved in keeping track of a sub-synoptic phenomenon in an area of scarce aircraft flights, sparse network of meteorological stations, large expanse of semi-arid land that is not habitable and a people to whom nature is so kind that they care less about weather. Satellite investigations of tropical storms provide better opportunities for synthesising the various events that occur during the evolution of line squalls.

2-3 Satellite investigations of line squalls

Investigation of West African severe weather systems by satellite is relatively new when compared with other forms of studies. Although satellite investigation of atmospheric disturbances is aesthetically appealing, it is not as popular as would be expected because access to the satellite data is very limited. Furthermore, among those who have access to satellite records, the criteria for identifying different atmospheric phenomena on satellite photographs vary. However, there have been few cases wherein satellite images have been most wonderfully utilised to study specific atmospheric disturbances; the work of Fortune (1977) is a case in point.

Fortune (1977) analysed 48 hours of time-lapse satellite imagery of a family of line squalls in West Africa. These images were received from SMS - I which provided visible and infra-red images of West Africa every 30 min. He (Fortune) examined particular images, taken during phase III of GATE, on the University of Wisconsin (U.S.A.) Man-Computer Interactive Data-Access System (McIDAS) which displays single images or time-lapse sequences on a television terminal. Sequential thirty-minute SMS - I images have the advantage of displaying instantaneous atmospheric features over a large area (e.g. West Africa). Thus, in terms of cost-benefit ratio, satellite investigations seems to be the least expensive method of studying tropical disturbances.

Line squalls can be identified in both visible and infra-red images as distinct cloud masses characterised by high brightness and, generally, a convex leading edge (Fortune, 1977). On satellite, images, the roll cloud (Plate 1(b)) can sometimes be seen, detached from the main cloud mass, as a low-level feature on the visible images only. Behind the main cloud mass, points of brilliance could be seen while the rear of line squalls are usually indistinct and fibrous. Plates 2(a) - 2(e) show different positions (with time) of the line squall of 5 September 1977 which Fortune analysed.

Latrasse (1972) classified African squalls into two types according to the pattern of the satellite pictures. Type I images have a single well-defined arc-shaped front (Fig. 7) that faces the direction of propagation. At the rear of this arc is a compact and very bright cloud mass. The rear edge of the cloud mass is not as sharp as the rear of the arc. Type II images (Fig. 7) appear like indented arcs. This type of squalls usually arise from the merging of Type I cloud masses. Type II squalls often split into sister squalls during the course of propagation. Apart from these two types , other forms of gust front exist. Some of these have been reported by Gurka (1976); although his observations were over the temperate region.

It is possible for SMS - I images to miss the process of initiation of line squalls because of the relatively large time interval between

Pate 2(a): The 'Fortune' squall of 5 September 1974. Visible SMS-I image of West Africa at 0800

(Arrows indicate position of squall line)

Plate 2(b): The 'Fortune' squall of 5 September 1974. Visible SMS-I image of West Africa at 0930 (Arrows indicate position of squall line)

Plate 2(c): The 'Fortune' squall of 5 September 1974. Visible SMS-I image of West Africa at 1100 (Arrows indicate position of squall line)

Plate 2(d): The 'Fortune' squall of 5 September 1974. Visible SMS-I image of West Africa at 1230

(Arrows indicate position of squall line)

Plate 2(e): The 'Fortune' squall of 5 September 1974. Visible SMS-I image of West Africa at 1400 (Arrows indicate position of squall line).

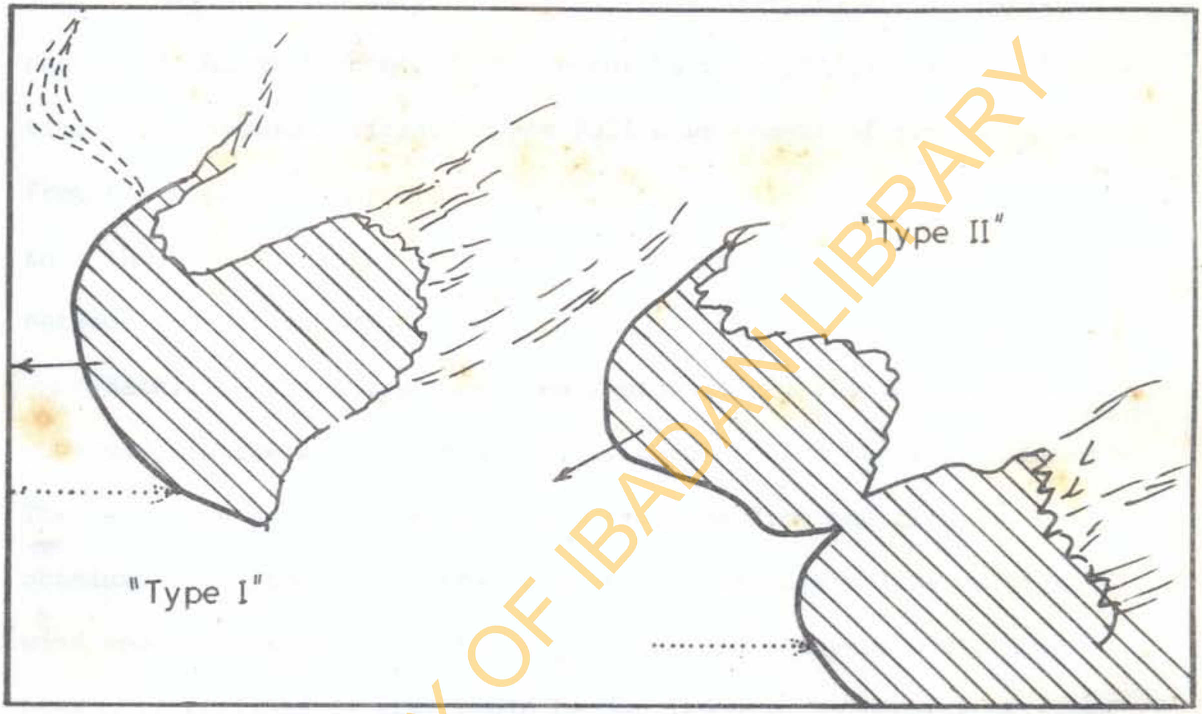

Fig. 7 : Two types of arc lines in satellite imagery
(from Latrasse, 1972)

successive pictures. It is therefore, not surprising that satellite imagery has not been very successful in identifying the formative causes of the disturbances. We do not have empirical evidence for the time lapse between initiation and full development of line squalls but from the results of numerical experiments available so far, we suppose that this time would be less than thirty minutes - the time interval between successive SMS - I images.

Satellite investigations cannot provide all the information necessary in the study of the origin, maturity and decay of line squalls. They are supposed to complement observations from the ground. Data obtained from ground stations on in-cloud parameters like vertical wind speed, moisture content, temperature and pressure would have been most reliable if they could be supplemented by reports obtained from direct flights into storm centres. Such direct flights are, however, hazardous. Consequently, studies on tropical disturbances have concentrated on modelling. The prohibitive cost of the option of observational investigations as a method of studying line squalls also favours the alternative of modelling.

CHAPTER 3

A REVIEW OF LINE SQUALL MODELS

Models of line squall that start from the complete set of hydrodynamical equations fall into two main categories: mathematical and numerical. There are other models that do not start from these equations but from some assumptions on the nature of motions of severe storms. Into this class falls models that have attempted to formulate the trigger mechanism of line squalls. Firstly, mathematical models shall be considered because of their relevance to this study.

3-1 Mathematical models

These models can be classified into two main divisions: the simple linear gravity wave models which are mainly appropriate to initial development of the storms and the non-linear steady state models that are applicable to the mature stage.

3-1-1 Gravity wave models

Gravity wave models omit non-linear effects in order that the equations of motion could be tractable. This linearity permits separation of variables; the variables being small deviations from steady state values. In the stable gravity wave model, the static stability of the atmosphere is taken to be greater than zero. Such positive values for static stability are normally assumed while simulating convective processes in dry air. Results show that in dry air, wave-like perturbations propagate without any amplification. As will be shown later, release of latent heat in moist thermal convection modifies the value of static stability thereby making it negative and in such environments, wave-like perturbations amplify without any propagation. Since this study is interested in the initial stages in the development of line squalls, gravity wave models will be developed later on.

3-1-2 Non-linear steady state model

An example of non-linear steady state models is the two-dimensional convection in shear by Moncrieff and Green (1972). They did not investigate the internal structure of convective storms; rather, their Cartesian set of axes moved with the speed of the storm such that at any instant, the motion is stationary relative to these axes. The flow was assumed to be steady and adiabatic and the entropy of the system written as a function of the streamfunction and height. From this expression, a conservative quantity was derived which allowed the flow field to be discussed without any integration of advection derivatives.

In discussing the flow field, Moncrieff and Green derived a non-dimensional number R which lacks any direct physical interpretation but is numerically related to the Richardson number R_i. This relationship is $$R_i = - \frac{RH}{Z_*}$$

where H = upper boundary of model (or cloud tops) and

Z_* = steering level i.e. the height at which there is no motion relative to the storm.

With the aid of this relation, the authors tried to predict the speed of propagation of some 'travelling' storms. Results show that the predictions are reliable if R_i is of the order of unity. The reliability however waves for storms that have considerable motion along the third dimension. This, as the authors pointed out, might be due to the two-dimensional nature of the model.

3-2 Numerical models

The basic equations used in numerical modelling (as a matter of fact, in mathematical modelling) are: the first law of thermodynamics, the equation of state for ideal gases and the equations for the conservation of mass, momentum and water substances. These equations have been applied in simulating convective storms in two-dimensions (Ogura, 1963; Takeda, 1971; Hane, 1973; Schlesinger, 1973; Mitchell and Hovermale, 1977; Moncrieff, 1978) and three dimensions (Ogura and Charney, 1962; Pastushkov, 1973; Steiner, 1973; Miller and Pearce,1974; Moncrieff and Miller, 1976; Schlesinger 1978,1980).

3-2-1 Two-dimensional models

These are simplifications of the dynamical description of the flow within severe storms. Results from these simulations are qualitatively representative of the structure and motion within storms. Quantitatively, however, some results (e.g. Hane, 1973) do not agree entirely with observations because the atmospheric phenomena being simulated have considerable motions along the third dimension. The numerical model of deep, moist convection in two dimensions by Schlesinger (1973) is fairly representative of other two-dimensional models.

The reliability of numerical simulations of atmospheric disturbances as far as predicting and understanding atmospheric phenomena is concerned depends largely on the amount of physics specified in the equations governing atmospheric motions and the boundary conditions applied in solving the equations. Since nature is complex, we cannot but have some idealisation and assumptions in setting out the equations. Some basic assumptions made in two-dimensional numerical models are:

(i) all y-derivatives are zero (i.e. parameters are invariant along the horizontal direction perpendicular to the direction of motion of storms)

(ii) the Coriolis force is neglected (most numerical experiments make this assumption because the square of the time-scale of the disturbances simulated is usually less than f^{-2}),

(iii) motions are anelastic i.e. local time changes of air density are neglected in the continuity equation (Ogura and Charney (1962) have shown that this assumption filters out waves of little meteorological significance e.g. acoustic waves. Further, Ogura and Phillips (1962) showed that this assumption along with that of Batchelor (1953) that the distribution of pressure and density are always close to the distribution of pressure and density in an adiabatically-stratified atmosphere are necessary in deriving the anelastic equations),

(iv) non-inclusion of the solid phase of water,

(v) no friction on the earth's surface (i.e. no-drag condition along the surface),

(vi) supersaturated water vapour condenses and water drops in an unsaturated region evaporate instantenously until saturation is realised in the region and

(vii) both ascent and descent are moist-adiabatic in saturated air.

Justification of some of these assumptions have been examined in detail by Schlesinger (1972).

Generally, in two-dimensional numerical models, the equation of continuity is written as:

$$\frac{\partial}{\partial x}(\rho_o u) + \frac{\partial}{\partial z}(\rho_o w) = 0.$$

The equation, written in this form, is accurate to about 1 per cent (Moncrieff and Green, 1972). This equation inplies the existence of a streamfunction, ψ, such that

$$u = \frac{1}{\rho_o}\frac{\partial\psi}{\partial z}$$

and $$w = -\frac{1}{\rho_o}\frac{\partial\psi}{\partial x}$$

The streamfunction, the vorticity equation and the conservation equations form a comparatively simple modelling system. Using this system, a Poisson-type of equation is solved for ψ. This equation is then integrated rafher than the primitive equations.

In the two-dimensional numerical model of Takeda (1971), this integration was carried out to study the effect of vertical profile of ambient wind on a precipitating convective cloud. Some earlier attempts (Arnason, Greenfield and Newburg, 1968; Liu and Orville,1969) had included ambient wind field, vertical shear and precipitation in their models. Takeda improved on these works by taking cloud physical process into account more realistically; although he did not include the ice phase. He grouped precipitating convective clouds into three types according to the way the clouds develop:

(i) convective clouds formed on both sides of an initial cloud due to diverging downdraughts in an atmosphere of sufficient conditional instability and in the presence of weak shear,

(ii) short-lived clouds that form when the direction of the vertical wind shear is constant with height and the shear is strong and

(iii) 'long-lasting' clouds that are formed when the vertical wind shear changes direction at a certain level.

Takeda's work indicate that there is a critical range of height (about 2.5 km) within which the shear must change direction for the formation of 'long-lasting' (or 'travelling') clouds. An important deduction from this experiment is that two air masses, oppositely directed and of different stabilities (e.g. one moist and the other dry) are necessary atmospheric conditions before 'travelling' clouds could develop. In particular, models of line squalls should be formulated using two-layer atmospheric models.

Hane's (1973) two-dimensional model was on the structure and tendency for self-maintenance of line-squall thunderstorm. He showed that there is a tendency towards convergence upon a common solution when various initial perturbations are imposed on the undisturbed atmosphere and that the time-lapse before convergence is a function of the initial difference between perturbations. Hence, the problem of initial perturbation is, according to Hane, not very important. Some of the parameters he calculated (e.g. vertical velocity and cloud size), like those of most two-dimensional numerical models, are larger than the observed values. This is probably due to the non-inclusion of motions along the third dimension. For instance, the downward motion, as calculated by Hane, is more intense than it would have been in nature where some flow would have taken place out of the plane of his model. He concluded that the capacity of the dry middle-level air to evaporate rain and to be cooled, along with the capacity of the moist low-level air to condense and to be warmed, ultimately control the line squall thunderstorm circulations. This study agrees with his conclusion.

Schlesinger (1973) investigated the influence of ambient conditions on the behaviour of deep moist convection in the atmosphere. His results revealed that the intensity of convection increased with low-level moisture supply and decreased with increasing mid-tropospheric wind shear. In cumulus (or cumulonimbus) convection, the intensity of convection usually decreases with increasing wind shear because the shear inhibits the formation of new bubbles at the top of the cloud. In the study of line squalls, however, vertical wind shear is very vital to the development and sustenance of the squalls.

The inability of two-dimensional models to correctly predict the speed of propagation of severe storms that have considerable motion along the third dimension, the fact that in-cloud parameters simulated from these models are usually larger than observed values and the reality that line squalls are three-dimensional in nature are some of the reasons that, at various times, prompted the extension of numerical experiments into three dimensions. The initial problem in three-dimensional numerical simulations was how to extend the vorticity system to three dimensions. This problem was surmounted by modelling with the primitive equations.

3-2-2 Three-dimensional models

Ogura and Charney (1962) used the primitive equations in their three-dimensional simulation of thermal convection. The authors did not explicitly treat the individual clouds and their internal circulations; rather, they dealt with the general problem of thermal convection on a large scale. Their model was designed to study the dynamical behaviour of all forms of thermal convection. From a frequency equation, which was derived from the primitive equations, they showed that wave-like disturbances generated in dry air propagate without any noticeable amplification whereas disturbances in moist air amplify with time. In chapter 5, their work is extended to describe waves generated along the surface of discountinuity between dry and moist air masses.

The numerical simulation of cumulonimbus clouds by Miller and Pearce (1974) was also in three dimensions. Their formulations were in pressure coordinates. An advantage of the use of pressure coordinates is the availability of data at pressure levels. Also, according to Miller and Pearce, the use of pressure coordinates could be prompted by the neatness of handling the air density in the mementum and continuity equations. Although the continuity equation appears neat in the pressure coordinates, it should be mentioned that this is the exact form of the equation. Since Ogura and Charney proved that the elimination of the local time changes of air density in the continuity equation filters out acoustic waves, it implies that waves of no meteorological significance would be implicit in pressure-coordinate formulations. Miller (1974) showed how such waves could be suppressed.

The model of Miller and Pearce responded sensitively to cloud physical processes in spite of the fact that their parameterization of liquid water was not too refined. It is therefore essential for numerical experiments to take into account, in a more realistic manner, as done by Takeda (1971), the size distribution of water drops, rainfall intensity, evaporation, congulation and terminal fall velocity of water drops.

A major problem in numerical experimentation is the specification of boundary conditions (B.C.). Often, the vertical components of velocity are taken to be zero at the surface and the top of convective elements; the top being usually near the tropopause. The most satisfactory way of choosing lateral boundaries would have been the specification of a large distance from the storm centre, where the disturbance of the velocity field would be zero throughout the period of convective activity (Miller and Pearce, 1974). Current computer facilities would not, however, allow this. Thus, various B.C.'s have been assumed for the lateral boundaries. Details of some forms of these B.C.'s could be obtained from Takeda (1970,1971), Hane (1973) and Miller and Pearce (1974).

3-3 Other theoretical models

Apart from mathematical and numerical models of thermal convection, there are other theoretical models. Two of these are: the plume(or jet) model and the bubble-theory model.

3-3-1 Plume model

In this model, the cloud is assumed to behave like a steady turbulent plume which entrains the air of the environment (Ogura,1963). This model cannot be easily applied to natural convection because there is, usually, no well-defined source-region in nature. Another hindrance is that the theory disregards any development in time of the depth or width of the cloud. Mixing at the top of the cloud is also neglected in the plume model.

3-3-2 Bubble-theory model

This model assumes that convection consists of separated volumes of bouyant fluid rising from an ambient fluid (Scorer and Ludlam, 1953), The most primitive form of this model - the classical 'parcel model' - does not allow heat and mass transfer through the boundary of rising columns of air (i.e. no entrainment) and it ignores pressure disturbance (Ogura, 1963). The model's assumption of uniform properties on the horizontal plane is definitely an over-simplification in the light of more-recent numerical experiments.

3-4 Syntheses of line squalls

In the course of synthesising the varous stages in the evolution of line squalls, the trigger mechanism of line squalls have been modelled; albeit with some inadequacies. Generally, these models do not start from the complete set of the hydrodynamical equations of motion. Examples of these models are those of Tepper (1950) and LeRoux (1976).

Tepper (1950) proposed a 'pressure jump line' mechanism for line squalls. He hypothesised that the initial impetus to the formation of line squalls is the acceleration of the cold front in a border between warm and cold air masses wherein an inversion or an isothermal layer exists in the warm zone. As a consequence of the acceleration, a pressure jump is formed (Fig. 8) along the surface of the inversion or isothermal layer. The air above the inversion is vigorously forced up as the pressure jump passes and, depending on the stability of this air, condensation and precipitation of moisture could occur. The air below the inversion is also forced upwards and the precipitation falling into it may, depending also on the stability of this air mass, reach the ground or evaporate completely.

After inducing the pressure jump, the cold front decelerates and the jump moves ahead of it. The deceleration causes the formation of a rarefaction wave which, according to Tepper, moves faster than the jump and eventually overtakes and destroys it. The destruction of the jump would apparently mark the termination of line squall. The total life span of squalls initiated by the Pressure Jump Line mechanism is thus a function of: the intensity of the initial acceleration of the cold front, the duration of the acceleration and the nature of the deceleration which follows the acceleration (Tepper, 1950). Such squalls could also be terminated whenever the inversion along which the pressure jump is travelling either descends to the ground or disappears completely.

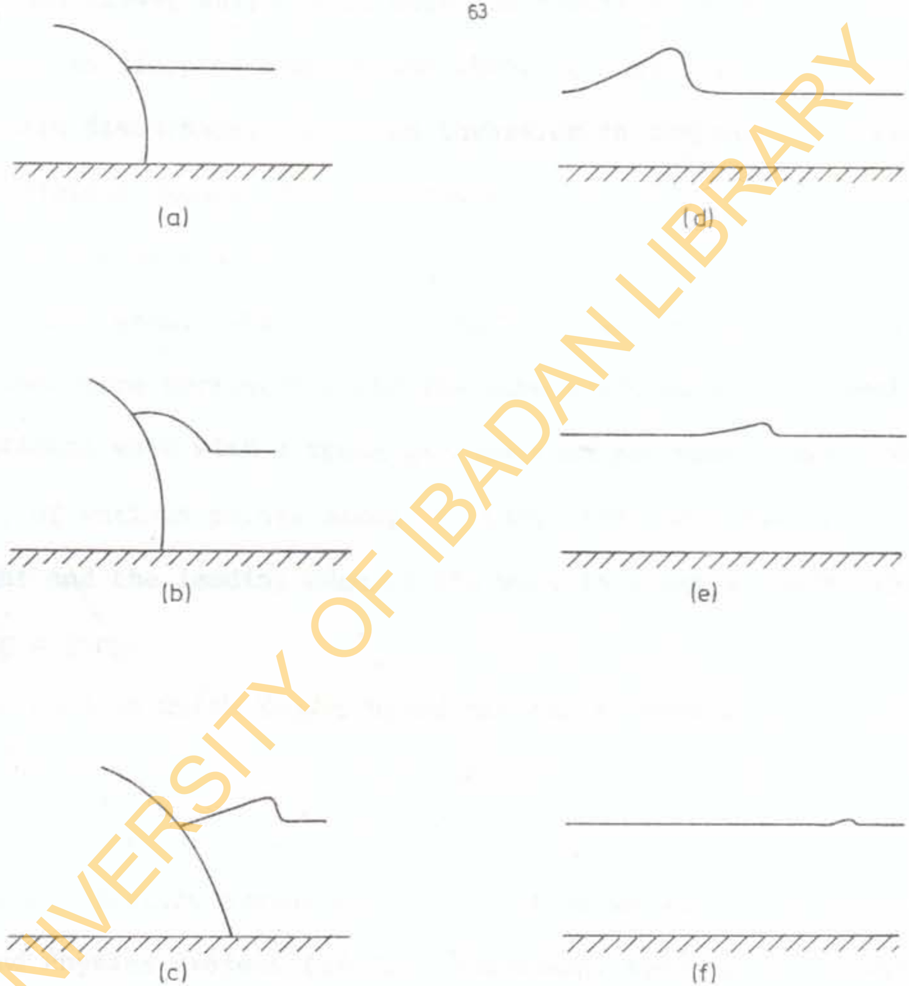

Fig. 8: Idealized history of a line squall

(a) Initial position of the cold front and the inversion in the warm sector

(b) The cold front accelerates inducing a pressure jump on the inversion

(c) The cold front decelarates while the jump moves out ahead with the rarefaction wave following

(d), (e) & (f) Successive positions of pressure jump being overtaken by the rarefaction wave

(from Tepper, 1950)

Tepper's hypothesis was based on the works of Prieswerk (1940) and Freeman (1948) which dealt with the theory of a gas flow in gas dynamics. He (Tepper) adapted the theory to describe the flow of atmospheric disturbances along an inversion in response to gravity. In the original theory, Freeman discussed the motion of a disturbance induced by the acceleration of a vertical-plane piston and propagated along an inversion. The level of the inversion would rise as a bump in its immediate surrounding and the wave produced would travel as a gravitational wave with a speed greater than the wind speed. The velocity of various points along the wave crest increases as a function of height and the leading edge of the wave is alsomost vertical; thus, creating a jump.

The data on which Tepper based his theory were read from the original traces of pressure, wind, temperature and precipitation taken at 55 automatic recording stations operated by the United States (of America) Weather Bureau during the cloud seeding experiment of the Cloud Physics Project (1948). One-minute synoptic maps were constructed from the data. Some important features of the maps are:

1. Pressure gradients are very intense (the order of magnitude is 1.15 mb km^{-1}).
2. The leading edge of the pressure gradient undulates violently and has an irregular motion.

3. The pressure jump is independent of the other parameters like wind shift, rain gush, wind speed maximum etc.

4. Other meteorological parameters, apart from pressure maximum, have an intimate relation one to another (e.g. these parameters are propagated with about the same speed - 20 to 21 m/s).

5. The average speed of all meteorological parameters exceeds the average maximum speed of the surface wind

6. The precipitation belt lies behind the pressure jump.

It has been observed (Mansfield, 1977), in agreement with one of the conclusions of Tepper (number 5 above), that the speed of the West African line squall and consequently that of all meteorological parameters associated with it, is the maximum within the troposphere. There are, however, other features of the West African line squall that cannot be easily explained by the Pressure Jump Line theory. Some of these features are:

1. There is no conclusive evidence that inversions exist prior to the onset of line squalls. Our experience in West Africa is that the layer of inversion observed during the period of frequent line squall activities occurs around the coast rather than the regions of initiation of line squalls (Fig. 2).

2. The speed of propagation of line squalls, according to the Pressure Jump theory, depends on the variations in intensity and duration of the initial acceleration of the cold front and hence, this speed must vary from one occurrence of line squall to the other. This is contrary to the almost-uniform speed of propagation of West African Line squalls which have been observed to be, approximately, the speed of mid-tropospheric winds.
3. The observed overturning of the atmosphere after the passage of line squalls.
4. The preference of highlands as places of origin of line squalls.

Another area where the Pressure Jump Line theory seems inappropriate as far as West African Line squalls are concerned is the identification procedure. Tepper identified line squalls through the showery precipitation that accompanies lines of storm which appear in the warm sector of cyclones, almost parallel to the cold front, along which there is intense convective activity. While this procedure was good at the time of his study because of the little amount of work done then on the dynamics and kinematics of line squalls, it is now known that precipitation that accompanies line squall is very erratic.

With the present-day knowledge (obtained from numerical experiments and observations) precipitation, being irregular, might not have any meaningful relationship with other meteorological parameters during line squall activities. Irregularities in the amount of rainfall accompanying line squalls make Tepper's identification procedure for these storms inadequate as far as West Africa is concerned.

Some other attempts were made, after the Pressure Jump Line theory, at explaining the trigger mechanism of line squalls. In one of such attempts, LeRoux (1976) explained that the key element is a surface of discontinuity, between two air masses, upon which a descending cold-core wind jet impinges; although he did not advance a viable mechanism for the descent of the jet. A series of events would then follow the impingement and these will lead to the formation of line squalls. Danielsen (1974) had earlier analysed cases in which a descending jet core triggers dust storms.

LeRoux observed no relationship between the intensity and speed of propagation of line squalls and thereby concluded that line squalls are energised solely by the easterly jet. In effect, the line squalls proposed by his mechanism will not be self-perpetuating (Fortune, 1977). This is contrary to the observations of Hamilton and Archbold (1945), Browning and Ludlam (1962), Fortune (1977), Mansfield (1977), Zipser (1977) and Bolton (1981). It is now known that fully-developed tropical line squall propagates in a way that suggests an in-built regenerative mechanism that sustains motion.

Our explanation, in the next chapter, of the trigger mechanism of line squalls will be a modified form of the proposal of LeRoux (1976). In our discussion, we hope to shed some light on those areas where the Pressure Jump Line mechanism and the explanation of LeRoux on the trigger mechanism of line squalls seem to be inadequate in formulating the initial stages in the evolution of West African line squalls.

CHAPTER 4

EVOLUTION OF LINE SQUALL

4-1 Necessary atmospheric features in the development of line squalls

As stated earlier, two air masses of different stabilities are essential before line squalls could develop. One of these air masses - the lower one - must be convectively unstable (Hamilton and Archbold, 1945; Browning and Ludlam, 1962; Hane, 1973). Some other features that are essential in the organisation of line squalls are: cold front zones (i.e. borders between cold and warm air masses) (Prohaska, 1905; Tepper, 1950; Takeda, 1971), a moisture profile that drops betwean 600 mb. and 500 mb. (Hane, 1973), a mid-level tropospheric jet maximum (Mansfield, 1977) and strong vertical shear (Browning and Ludlam, 1962).

Strong vertical shear in the ambient winds aids severe storms because it separates the downdraught from the updraught and allows the former to thrust under the bouyant columns (Browning, 1964). The development of new cells on the downshear side of a storm is also enhanced by a strong vertical shear (Fortune, 1977). Figure 9 shows the vertical profile of wind over West Africa between the months of April and September. From this figure, it is observed that strong vertical shear is created in the lower troposphere of West Africa by the 650 mb. mid-tropospheric jet. Thus, throughout the period referred to as the wet season, strong vertical shear, an essential

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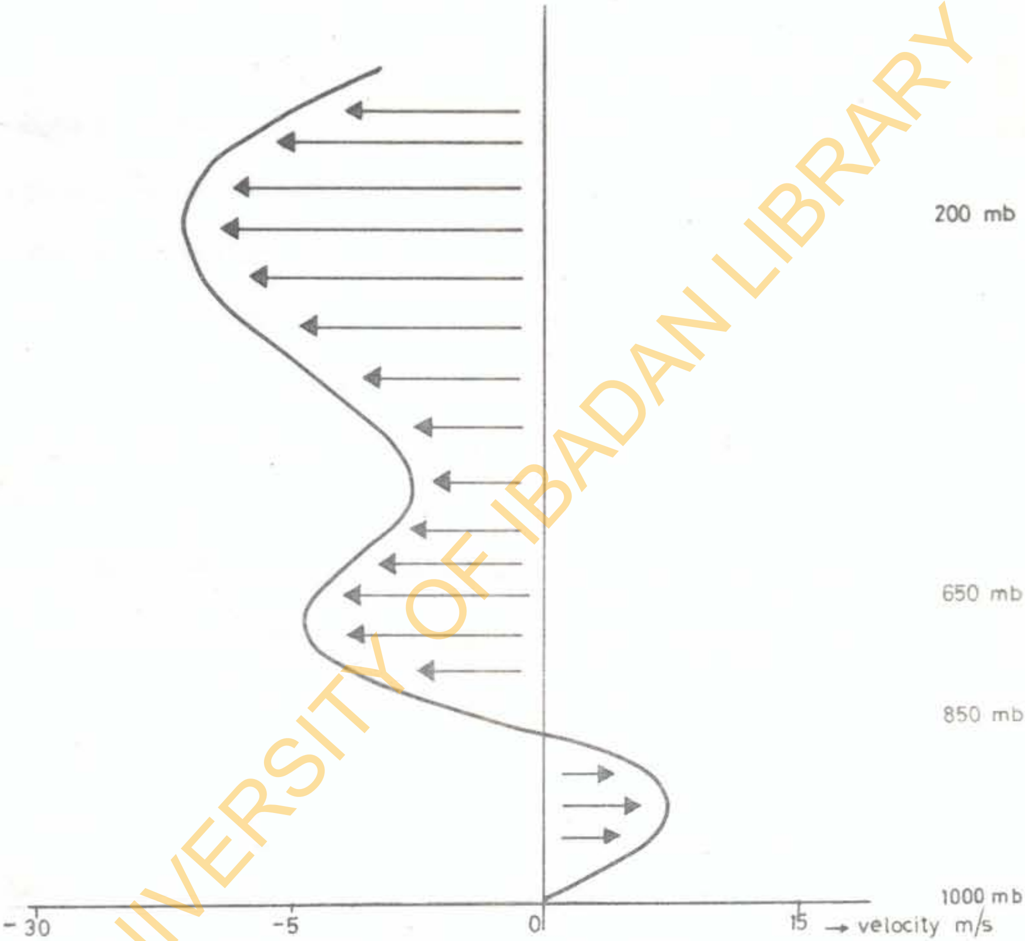

Fig.9 : A sketch of the vertical profile of wind over West Africa between April and September

(Deduced from soundings made at the meteorological station attached to the airport at Ibadan – 7° 26′ N 3° 54′ E) atmospheric feature in the development of line squalls, exists in West Africa.

Another feature that enhances the development of line squalls is a drop in moisture profile between 600 mb. and 500 mb. (Hane,1973). This is likely to be so in West Africa because some mixing might occur, below the 650 mb. jet (Fig. 9), between the dry north-easterlies and the humid mansoon winds since the surface of discontinuity is not a sharp demarcation. This mixing tends to make the layer of air above the 650 mb. jet to be drier than the layer below it and consequently, there is a drop in moisture profile between 700 mb. and 500 mb.

In West Africa, an example of a convectively-unstable air mass is the monsoon wind. If a parcel of the monsoon south-westerlies is sufficiently lifted, condensation occurs and the latent heat released makes the parcel warmer than its environment. This results in the bouyancy of such parcels of air. The south-westerlies are therefore convectively unstable. Convective instability is, as mentioned earlier, important in the development of line squalls.

The conclusion from the foregoing is that the basic atmospheric features that aid the development of line squalls are present in West Africa during the period of frequent squall activities. However, the existence of these features alone cannot give rise to line squalls.

Other factors play vital roles. A probable factor is the African easterly waves.

Burpee (1972) traced the origin of the easterly waves to an area between Khartoum (32^{o}E) in Sudan and Ft. Lamy (now N'Djamena) (15^{o}E) in Chad. The waves propagate westwards across West Africa from June to early October (Burpee, 1972) with a period of about 3.5 days and wavelength of about 2000 km (Carlson, 1969 a,b). These waves are sometimes associated with the development of line squalls. The probable link, from our studies, is the evidence of Fortune (1980) that a necessary condition for the development of line squalls is a synoptic pattern that generates convergence and the observation of Reed, Norquist and Recker (1977) that there is a convergence ahead of the African easterly wave trough. The deduction from these is that line squalls could be initiated in the viscinity of the easterly wave trough. This idea is still not a consensus because of some clear distinctions between the line squall phenomenon and easterly waves.

One major distinction between line squalls and easterly waves is the speed of propagation of the two phenomena. Line squalls propagate at about 15-20 m/s (Hamilton and Archbold, 1945 ; Aspliden et.al., 1976; Fortune, 1977; Mansfield, 1977; Zipser,1977; Bolton, 1981) while easterly waves propagate at about half this value

(Carlson, 1969 a,b; Burpee, 1972). As a further distinction between line squalls and African easterly waves, Carlson (1971) found that the waves have a warm core at high levels and a cold core at middle and low levels with no clear pattern of cloudiness or vertical motion whereas line squalls are cold-core in low levels (due to the downdraught) and they possess distinctive cloud and vertical motion patterns.

In the remaining sections of this chapter, we shall endeavour to show the important role(s) played by the observed atmospheric features that have been enumerated above, in the development and maturity of line squalls.

4-2 Initial stages and trigger mechanism of line squall

Wave-like disturbances are known to be generated along surfaces of discontinuity between air masses; the discontinuity between the south-westerlies and the north-easterlies in West Africa is not an exception. It will be shown in the next chapter that, depending on the stabilities of the air masses on either side of the discontinuity, such wave-like disturbances could intensify in vertical velocity, pressure, temperature etc. as well as propagate through the system in which they are generated. This 'conflict' between the air masses was long ago recognised as a cause of violent weather (Hubert, 1926).

In this study, the initiation of line squalls is associated with these perturbations.

The five sections of figure 10 show the East-West section through the lower troposphere in West Africa. Figure 10 a represents the undisturbed atmospheric structure of the sub-region wherein the north-easterlies are on top of, but oppositely directed to, the south-westerlies. The boundary between the air masses is at a pressure level of about 850 mb. (Fig. 9). Figure 10b shows the onset of a wave-like disturbance along the surface of discontinuity. The disturbance could intensify as shown in figure 10c and this intensification could be large enough to cause a blockage of the 650 mb. mid-tropospheric jet.

On encountering the surface of discontinuity, the jet further distorts the surface and in the process, forces the monsoon south-westerlies to ascend more (Fig. 10d). Due to the forced convection, a rising parcel of the convectively-unstable south-westerlies could condense. Precipitations from the rising parcel of air could fall into the jet that now underlies the monsoon winds. Some of these precipitates evaporate in the dry jet; the latent heat of evaporation being supplied by the jet. This evaporative loss of heat cools the jet and it sinks.

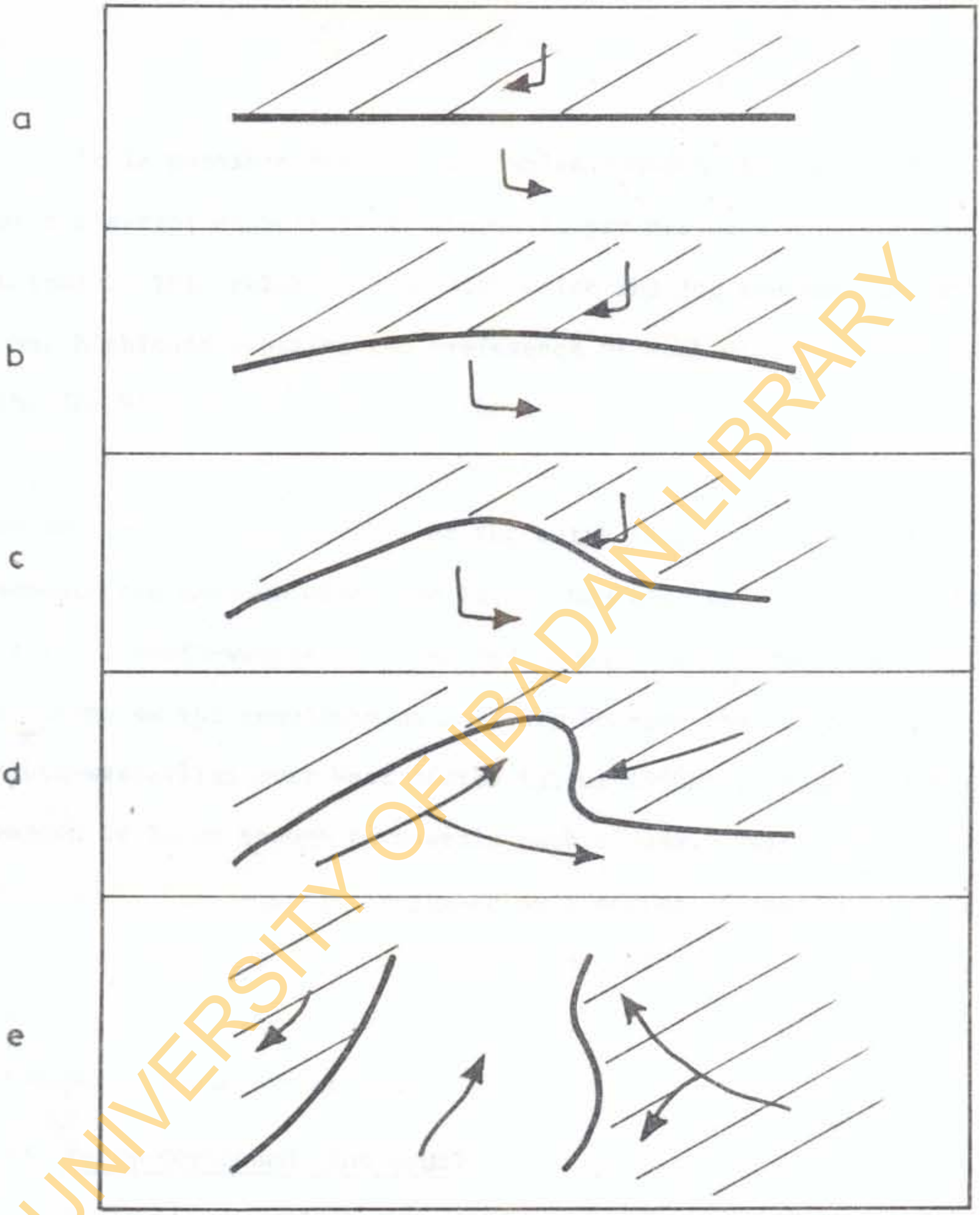

Fig.10: The process of formation of line squalls
(from Le Roux, 1976)

It is possible for the jet, while sinking, to hit the surface of the earth; especially if the whole process occurs over a high terrain. This relative ease with which the jet touches the ground over highlands explains the preference of such elevated areas like the Jos Plateau of Nigeria as the source region of line squalls.

The active phase in the evolution of line squalls starts as soon as the jet hits the surfaee of the earth (Fig. 10e). The jet goes beneath the monsoon winds and lifts them up. As a result of this lifting, south-westerlies rise and develop into large clouds. Clouds as large as the cumulonimbus may form because the depth of the south-westerlies over West Africa (about 1500m) throughout the wet season is large enough to sustain such clouds. This explains the reason why line squalls propagate as a series of cumulonimbus cells. The horizontal extent of propagation depends on the extent of travel of the jet over the surface as well as the condition within the atmosphere ahead of the storm.

4-3 Fully-developed line squall

The mechanism of propagation of fully-developed line squall is self-sustaining. Moist low-level air is lifted by the mid-tropospheric jet as the latter moves over the surface of the earth. The rising, moist air condenses and the precipitates fall into the underlying jet.

Since the jet is dry, some of the rain evaporate in it. The evaporative loss of heat coools the jet and it sinks; thereby continuing its motion over the surface and consequently, lifting the moist air ahead. Hane (1973) presented a similar mechanism for propagating severe storms. The sustenance of the storm, however, requires an atmosphere ahead of the storm that is convectively-unstable, sheared in vertical wind profile and lacking other storms (Dhonneur, 1970).

In the case of an environment that had experienced local convective storms before the arrival of the squall, precipitation is usually very low by the time the squall arrives. Thus, it is not uncommon for squalls to dissipate completely in areas where severe convective storms had previously occurred.

Vertical wind shear, another vital factor in the continuous sustenance of line squalls, is responsible for the structure of inflow ahead and outflow to the rear that is usually observed in West African line squalls. This structure is an overturning of the atmosphere because high winds aloft are brought to the surface and the lower winds transported upwards. This overturning stabilises the atmosphere after the passage of squalls (Hamilton and Archbold, 1945; Zipser, 1977; Bolton, 1981). The high wind is brought to the surface through the descent of the 650 mb. mid-tropospheric jet.

Also, sustenance of line squalls depend on the continuous descent of the 650 mb. jet. As mentioned earlier, this descent is enhanced by the evaporative loss of heat by the dry jet as a result of the precipitates falling, and evaporating, in it. Precipitation is an integral part of convective instability. Hence, convective instability in the atmosphere ahead of the storm is essential for the sustenance of line squalls.

The descent of the jet creates a region of strong convergence which is often called the squall front. According to Mitchell and Hovermale (1977), this front is the gusty wind surge that leads the surface outflow from an intense convective storm downdraught. Strong convergence and the fact that the descending jet is colder than the monsoon winds aid the ascent of the latter.

The rate at which convective elements develop ahead of 'travelling' storms depends on how fast the squall front travels on the surface. The squall front is, however, the leading edge of the mid-tropospheric jet as the jet crawls over the surface. Hence, the close association between the speeds of propagation of line squalls and that of the mid-tropospheric jet. As observed (Aspliden et al., 1976; Mansfield, 1977), line squalls propagate at speeds of about 1 m/s over and above the speed of the mid-tropospheric jet. This slight increase in speed might be due to the faster rate at which the jet is sinking as a result of a combination of its own weight and negative bouyancy produced by evaporation.

On the average, West African line squalls propagate in a direction WSW. This is so because the monsoon winds that give rise to the development of cumulonimbus clouds come from a direction that is more WSW than SW and the north-easterlies are from a direction that is more ENE than NE. That is, the jet, whose leading edge is the squall front, comes from a direction that is approximately ENE.

Immediately following the squall front in a 'travelling' storm is the core of precipitation (Fig. 11). The intensity of rainfall decreases as one moves further away from the front because the rising columns of monsoon winds would have lost most of their moisture content and thereafter, merge with the anvil. However, Zipser (1977) observed heavy rain in the region 30-100 km behind the squall front (Fig. 11). This rain is from the anvil base.

The anvil is usually thick and extensive. This thickness is of the order of 8 to 10 km (Zipser, 1977) in the region 30 - 100 km behind the squall. Whereas the anvil could extend to about 300 km behind the storm, the highest could tops is about 15 km (Fig. 11). This large horizontal extent of the squall system with respect to the vertical extent justifies the use of the hydrostatic relation in mathematical modelling of line squalls.

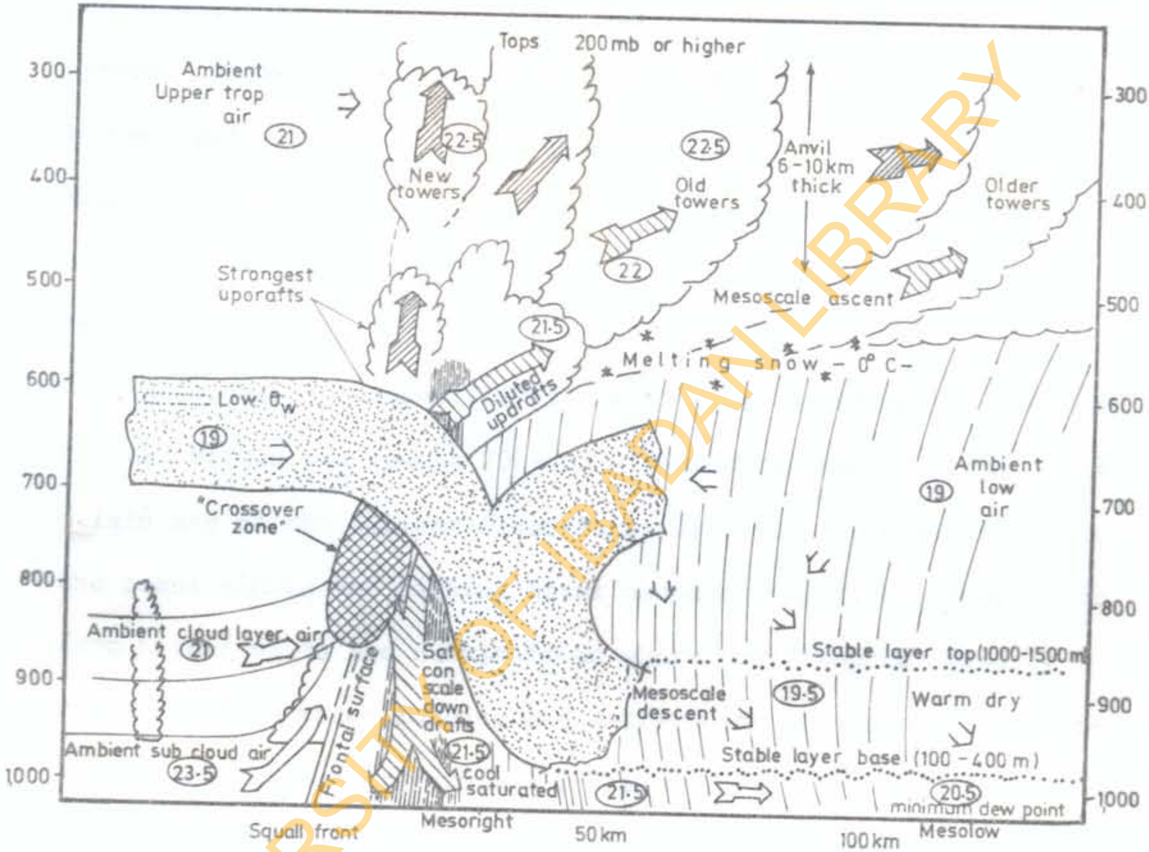

Fig. 11: Schematic cross section through a class of squall system. All flow is relative to the squall line which is moving from right to left. Circled numbers are typical values of θ_w in °C

(from Zipser, 1977)

The extent and persistence of line squall systems suggest the possibility of organised ascents in the upper half of the troposphere. These ascents would be very vital to the sustenance of the storm. But ascents would normally give rise to downdraughts. Mansfield (1977), Miller and Betts (1977), Zipser (1977) and Bolton (1981) have all observed two scales of down draughts in line squall systems.

According to Zipser (1977) one of these downdraughts could be located within the lowest 200-400 m behind the squall front. Its origin may be the ambient cloud-layer air (900-800 mb).lifted into the cumulonimbus and sinking after becoming negatively bouyant or from higher up in the cumulonimbus where downdraughts normally originate from water loading and entrainment of air with low wet-bulb potential temperature, θ_{ω}, (Zipser, 1977).

The other scale of downdraughts has lower θ_{ω}. Zipser (1977) postulated that their origin must therefore be from areas of low θ_{ω} in the ambient winds. This source must be typically above 750 mb. These downdraughts approach the line squall from the front and the rear (Fig. 11). While overtaking from the rear is more easily understood because of descending 650 mb. jet, the entry from the front is more difficult to visualise. However, the picture becomes clearer if it is realised that along the squall front, cumulobimbus clouds are not continuous in distance ar in time.

Fujita (1963) noticed mesoscale high pressure regions immediately following line squall passage as well as the occasional presence of a mesoscale low-pressure area some distance behind line squalls. Although this observations was over the temperate zone, Zipser (1977) has also observed mesohighs and mesolows within the tropics. A large depth of cold saturated air combine with large liquid water content (Sanders and Emanuel, 1977) to cause a rise in pressure in regions immediately following line squall passage. Mesolows exist at a distance of about 100 km behind the squall front.

Data compiled by Aspliden et al. (1976)(Fig. 4) show that line squalls scarcely propagate beyond the coasts of West Africa. Hence, it would be right to conclude that line squalls decay around the coastal areas. This decay might be as a result of extensive convective activities around the coasts such that squalls propagating into these areas die out. Another reason for the decay might be the fact that the 650 mb. jet now impinges on the ocean surface rather than the solid earth on which it could spread out and form squall fronts which force the south-westerlies to rise. The layer of inversion (Fig. 2) observed around the coasts could also contribute to the decay of line squalls by inhibiting convective processes. This layer acts like a 'lid' that prevents upward growth of clouds(Balogun, 1974).

This 'lid' is never taken off unless the force of convection is enought to destabilise the inversion layer. In the absence of precipitation from convective elements, line squalls would dissipate completely.

Attention can now be focused on a gravity wave model for the initial stages in the development of line squalls.

CHAPTER 5

FUNDAMENTAL EQUATIONS

5-1 Basic and perturbation equations

The basic equations in the theory of atmospheric circulations are given below in notational forms. All symbols have their usual meaning. However, for the sake of clarity, the meanings of these notations and symbols have been given.

$$\frac{\partial u}{\partial t} + \nabla\left(\frac{u^2}{2}\right) + f\Lambda u + \frac{\nabla P}{\rho} + g = \nu\nabla^2 u \qquad \ldots\ldots\ 5.1$$

$$\frac{D}{Dt}\left(\frac{Lr}{C_p} + \theta\right) = 0 \qquad \ldots\ldots\ 5.2$$

$$\frac{D\rho}{Dt} + \rho\nabla u = 0 \qquad \ldots\ldots\ 5.3$$

$$\frac{D}{Dt}(r+q) = 0 \qquad \ldots\ldots\ 5.4$$

$$\frac{T}{\theta} = \left(\frac{P}{P_o}\right)^{\kappa} \qquad \ldots\ldots\ 5.5$$

5.1 is the momentum equation with viscosity term,.

5.2 is the energy equation and 5.3 is the equation for continuity of mass. 5.4 represents the conservation of water substance and 5.5 is the definition of potential temperature.

The Coriolis parameter, f, in the momentum equation could be neglected because the square of the time scale of line squalls is much less than f^{-2}. All molecular effects (e.g. viscosity) are also neglected because the scale of motion is much more than molecular dimensions.

If $\frac{Dr}{Dt} = Q$, 5.1 - 5.5,

after simplification, become:

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} + \frac{1}{\rho}\frac{\partial P}{\partial x} = 0 \qquad \text{..... 5.6}$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + \frac{1}{\rho}\frac{\partial P}{\partial y} = 0 \qquad \text{..... 5.7}$$

$$\frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z} + \frac{1}{\rho}\frac{\partial P}{\partial z} + g = 0 \qquad \text{......5.8}$$

$$\frac{\partial \theta}{\partial t} + u\frac{\partial \theta}{\partial x} + v\frac{\partial \theta}{\partial y} + w\frac{\partial \theta}{\partial z} + \frac{LQ}{c_p} = 0 \qquad \text{..... 5.9}$$

$$\frac{1}{\rho}\left(\frac{\partial \rho}{\partial t} + u\frac{\partial \rho}{\partial x} + v\frac{\partial \rho}{\partial y} + w\frac{\partial \rho}{\partial z}\right) + \frac{\partial u}{\partial y} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \qquad \text{..... 5.10}$$

$$\frac{\partial q}{\partial t} + u\frac{\partial q}{\partial x} + v\frac{\partial q}{\partial y} + w\frac{\partial q}{\partial z} + Q = 0 \qquad \text{..... 5.11}$$

$$\frac{T}{\theta} = \left(\frac{P}{P_o}\right)^{\kappa} \qquad \text{..... 5.12}$$

The set of hydrodynamical equations 5.6 - 5.12 consists of non-linear equations and, hence, they are difficult to solve. The perturbation method reduces them to a linear form.

The basic assumption of the perturbation method is that atmospheric motions are made up of small perturbations superimposed on a basic atmospheric state. Fundamentally, the basic motion as well as the perturbations must separately and jointly satisfy the hydrodynamical equations. Thus, the atmospheric variables in 5.6 - 5.12 could be expressed as linear combinations of their basic and perturbation values as:

$$\left.\begin{aligned} u &= u_s + u' \\ v &= v_s + v' \\ w &= w \\ \theta &= \theta_s + \theta' \\ q &= q_s + q' \\ P &= P_s + P' \end{aligned}\right\} \qquad \ldots\ldots 5.13$$

where subscript 's' denotes the basic values and the primed variables are the perturbation quantities. The set of equations 5.6 to 5.12 could be converted into their equivalents in the pressure co-ordinates. As stated earlier, an advantage of the pressure co-ordinates is the availability of more data (on atmospheric parameters) along isobaric surfaces. Further, the frequency equation being sought appears neater in the pressure co-ordinates. (The equivalence of the frequency equation in z-coordinates have been documented by many authors e.g. Giwa (1965)). The set of values in 5.13 could be incorporated, along with the co-ordinate conversion, into the hydrodynamical equations to give, after linearisation:

$$\left.\begin{aligned}
&\frac{\partial u'}{\partial t} + u_s \frac{\partial u'}{\partial x} + \omega \frac{\partial u_s}{\partial P} + \frac{1}{\rho_s} \frac{P'}{\partial x} = 0 \\
&\frac{\partial v'}{\partial t} + u_s \frac{\partial v'}{\partial x} + \frac{1}{\rho_s} \frac{\partial P'}{\partial y} = 0 \\
&-\frac{1}{\rho_s g}\left(\frac{\partial \omega}{\partial} + u_s \frac{\partial \omega}{\partial x}\right) + \frac{1}{\rho_s} \frac{\partial P'}{\partial P} + \frac{RT'}{P_s} = 0 \\
&\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial \omega}{\partial P} = 0 \\
&\frac{\partial \theta'}{\partial x} + u_s \frac{\partial \theta'}{\partial x} + \omega \frac{\partial \theta_s}{\partial P} + \frac{LQ}{C_p} = 0 \\
&T' = \theta' \left(\frac{P_s}{P_o}\right)^{\kappa}
\end{aligned}\right\} \quad \ldots\ 5.14$$

5.14 is a set of perturbation equations which could be used to derive a frequency equations.

5-2 <u>Frequency equation</u>

The perturbations superimposed on the hydrodynamical equations constitute the disturbances in the atmosphere. Since atmospheric disturbances are wave-like in nature, these perturbations could be expressed in the form:

$$u' = u(p)\, e^{i(\sigma_o t + \alpha x + \beta y)}$$

$$v' = v(p) e^{i(\sigma_o t + \alpha x + \beta y)}$$

etc.

such that

$$\frac{\partial u'}{\partial t} = i\sigma_o u'$$

$$\frac{\partial u'}{\partial x} = i\alpha u'$$

and

$$\frac{\partial u'}{\partial y} = i\beta u'$$

i.e.

$$\left.\begin{aligned} \frac{\partial}{\partial t} &= i\sigma_o \\ \frac{\partial}{\partial x} &= i\alpha \\ \frac{\partial}{\partial y} &= i\beta \end{aligned}\right\} \qquad \ldots\ldots\ 5.15$$

If it is assumed that density does not vary along the horizontal (i.e. x direction) in the mean atmosphere,

$$\frac{\partial}{\partial x}\left(\frac{P'}{\rho_s}\right) = \frac{1}{\rho_s}\frac{\partial P'}{\partial x}$$

and so, if $G' = \frac{P'}{\rho_s}$

$$\frac{1}{\rho_s}\frac{\partial P'}{\partial x} = \frac{\partial G'}{\partial x}$$

and

$$\frac{1}{\rho_s}\frac{\partial P'}{\partial y} = \frac{\partial G'}{\partial y}$$

Since, as mentioned in chapter 4, the horizontal extent of the line squall phenomenon is much more than its vertical extent, the hydrostatic relation could be used in this mathematical formulation.

In effect, the vertical acceleration terms will be neglected in the set of equations in 5.14. If 5.15 is incorporated into 5.14 and tracer element, μ, placed on the latent heat term, the following set of equations is obtained:

$$i\sigma^{*}u' + \omega\frac{\partial u_s}{\partial P} + i\alpha G' = 0 \qquad .... \; 5.16$$

$$i\sigma v' + i\beta G' = 0 \qquad .... \; 5.17$$

$$\frac{\partial G'}{\partial P} + \frac{RT'}{P_s} = 0 \qquad .....5.18$$

$$i\alpha u' + i\beta v' + \frac{\partial \omega}{\partial P} = 0 \qquad .... \; 5.19$$

$$i\sigma\theta' + \omega\frac{\partial \theta_s}{\partial P} + \frac{LQ}{C_p}\mu = 0 \qquad .... \; 5.20$$

$$T' = \theta'\left(\frac{P_s}{P_o}\right)^{\kappa} \qquad .... \; 5.21$$

Eliminating u', v', θ', T' and G' from the set of equations 5.16 - 5.21, the result is:

$$\sigma^2 \frac{\partial^2 \omega}{\partial p^2} + \frac{R^2 T_s}{P_s^2 g}\left[\frac{\mu L g P_s}{R\theta_s C_p}\frac{\partial r}{\partial P} - \frac{g}{C_p} - \frac{\partial T_s}{\partial z}\right]\nabla^2\omega = 0 \qquad .... \; 5.22$$

If $$\Gamma = \frac{R^2 T_s}{g^2}\left[\frac{g}{C_p}\left(1 - \frac{\mu L P_s}{R\theta_s}\frac{\partial r}{\partial P}\right) + \frac{\partial T_s}{\partial z}\right], \qquad ..... \; 5.23$$

5.22 becomes

$$\frac{\partial^2 \omega}{\partial p^2} + \frac{g\Gamma}{P_s^2}\left(\frac{\alpha^2 + \beta^2}{\sigma^2}\right)\omega = 0 \qquad .... \; 5.24$$

---

* $\sigma = \sigma_o + u_s\alpha$

5.24 is the frequency equation being sought and its solution will be discussed in the next chapter.

5-3 Effects of atmospheric stability on the frequency equation

The measure of atmospheric stability in 5.24 is Γ and as defined in 5.23

$$\Gamma = \frac{R^2 T_s}{g^2}\left[\frac{g}{C_p}\left(1 - \frac{\mu L P_s}{R\theta_s}\frac{\partial r}{\partial p}\right) + \frac{\partial T_s}{\partial z}\right]$$

If γ represents the environmental lapse rate and Γ_d denotes the dry adiabatic lapse rate,

$$\Gamma = \frac{R^2 T_s}{g^2}\left[\Gamma_d\left(1 - \frac{\mu L P_s}{R\theta_s}\frac{\partial r}{\partial p}\right) - \gamma\right] \qquad \ldots\ldots 5.25$$

In a dry atmosphere, the latent heat term in 5.25 will be non-existent because there would not be any condensation

$$\text{i.e.} \quad \mu = 0$$

Thus,

$$\Gamma_d = \frac{R^2 T_s}{g^2}\left[\Gamma_d - \gamma\right]$$

where the subscript '2' is used to distinguish the atmospheric stability within such a dry environment. In the context of West Africa, 5.26 would represent the stability condition within the north-easterlies.

In a moist environment, $\mu = 1$ because of the possibility of condensation during a convective process. This possibility would then justify the retention of the latent heat term. The term

$$\Gamma_d[1 - \frac{LP_s}{R\theta_s}\frac{\partial r}{\partial p}]$$

in 5.25 would then represent a modification of the dry adiabatic lapse rate due to precipitation. If a layer of saturated air is assumed, this term will be the wet adiabatic lapse rate, Γ_ω.

$$\text{i.e.} \quad \Gamma_\omega = \Gamma_d[1 - \frac{LP_s}{R\theta_s}\frac{\partial r}{\partial p}] .$$

Since the south-westerlies are moist, the stability of this air mass could be written as:

$$\Gamma_1 = \frac{R^2 T_s}{g^2}[\Gamma_\omega - \gamma] \qquad \text{........ 5.27}$$

The south-westerlies have been shown in section 4-1 to be convectively unstable and as such, the environmental lapse rate, γ, within this air mass lies between Γ_ω and Γ_d.

$$\text{i.e.} \quad \Gamma_\omega < \gamma < \Gamma_d .$$

This inequality implies that Γ_1 (5.27) will be negative. If such a negative value of Γ is used in the solution of 5.24, the condition for obtaining a sinusoidal function of time for any perturbation will be

$$\sigma^2 < 0.$$

This implies that the frequency of the waves generated is purely imaginary and such waves are known to amplify, without any propagation, with time.

In practice, an environmental lapse rate that is greater than the dry adiabatic lapse rate is a sign of instability. Such an environment cannot be maintained for along period because mechanical mixing would occur and thereby stabilise the atmosphere. This means that the environmental lapse rate, even in dry air, is scarcely larger than Γ_d and if it does, it is for a short period. Therefore, Γ_2 (5.26), which is a measure of the stability of the north-easterlies, will be positive. If such a positive value of Γ is used in the solution of 5.24, the condition for obtaining a sinusoidal function of time would be

$$\sigma^2 > 0.$$

The implication of this is that the frequency of the waves generated will be real. Such waves propagate, without any amplification, with time.

Given a situation, therefore, where there are two air masses, whose separate stability conditions give complex and real values for wave frequencies respectively, one on top of the other (e.g. the north-easterlies on the south-westerlies) the waves generated along the surface of discontinuity would both propagate as well as a amplify with time. The velocities of such waves would be complex; the real part represents the phase velocity while the imaginary part is a measure of the amplification of the wave.

CHAPTER 6

SOLUTION TO THE FREQUENCY EQUATION

Equation 5.24 could be solved by assuming a single-layer model for the atmosphere. This kind of solution had been discussed by Ogura and Charney (1962), Miller (1974) and Bolton (1981). However, since line squalls require two layers of air with different static stabilities for their initiation and sustenance, it would be more realistic to solve equation 5.24 with the aid of a two-layer model of the atmosphere.

6-1 Boundary conditions

Figure 12 shows a section of the interface between the air masses in a two-layer atmospheric model. If w and u are the vertical and horizontal components of velocity respectively, the velocity component perpendicular to the surface is

$$w\cos\theta - u\sin\theta .$$

This value must be continuous if no air parcels accumulate at the interface. For small ε (fig. 12),

$$\frac{\partial \varepsilon}{\partial x} = \tan\theta \approx \sin\theta.$$

Also, $\cos\theta \approx 1$

Therefore, $w\cos\theta - u\sin\theta \approx w - u\frac{\partial \varepsilon}{\partial x}$

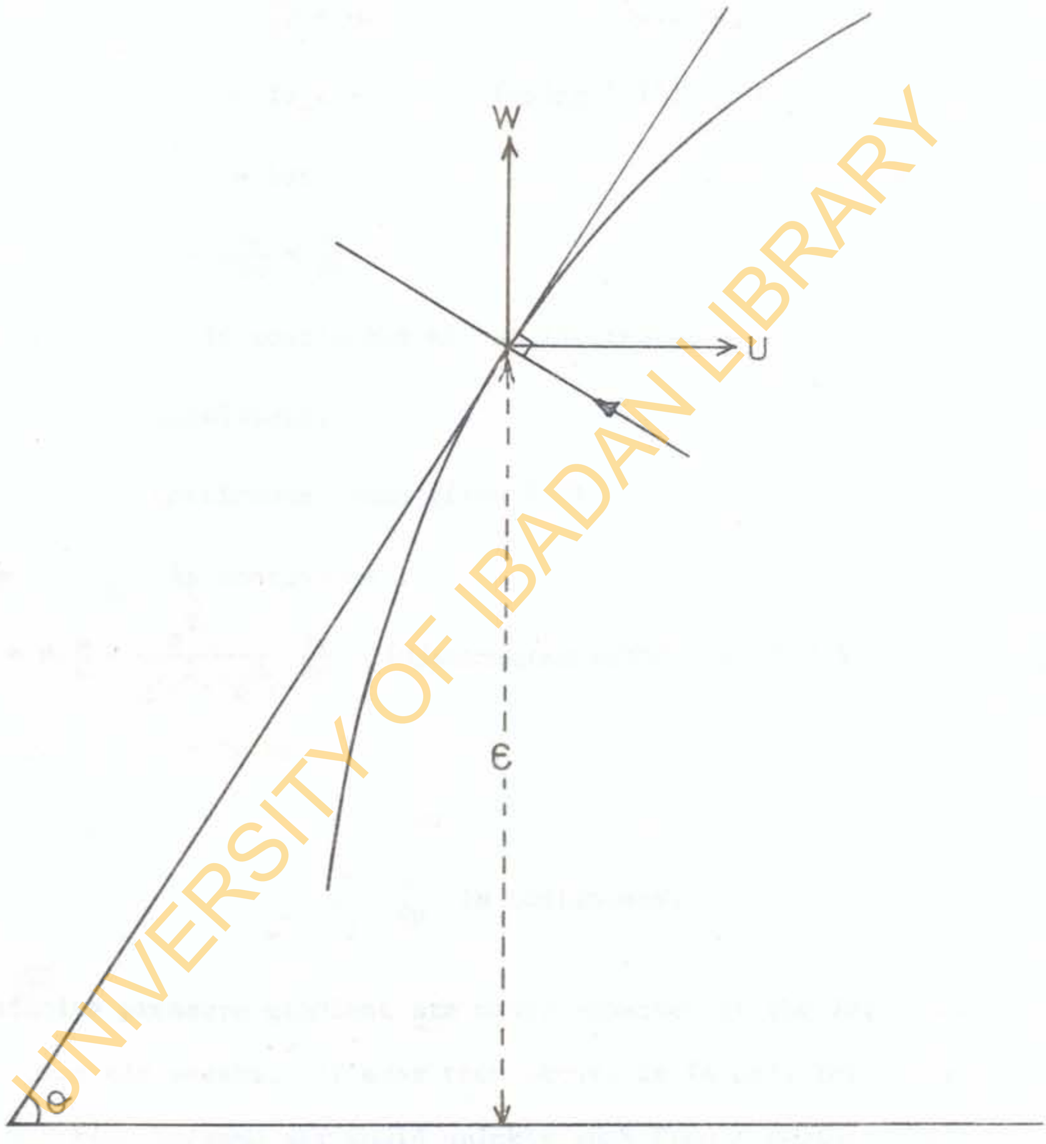

Fig.12: A section of the interface between the air masses in a two-layer atmospheric model

$$w = \frac{D\varepsilon}{Dt} \quad \text{(fig. 12)}$$

i.e. $$w = \frac{\partial\varepsilon}{\partial t} + u\frac{\partial\varepsilon}{\partial x} \qquad \ldots.\ 6.1$$

$$= i\sigma_o\varepsilon + \alpha u_s\varepsilon \quad \text{(using 5.15)}$$

i.e. $$w = i\sigma\varepsilon \qquad \ldots\ldots\ldots\ldots\ 6.2$$

From 6.1, $w - u\frac{\partial\varepsilon}{\partial x} = \frac{\partial\varepsilon}{\partial t} = i\sigma_o\varepsilon$

but $w - u\frac{\partial\varepsilon}{\partial x}$ is continuous at the interface, hence

ε is continuous.

If ε is continuous, then (from 6.2)

$\frac{w}{i\sigma}$ is continuous .

$$w = H\frac{\omega}{p} - \frac{\sigma^2}{g(\alpha^2 + \beta^2)}\frac{\partial\omega}{\partial p} \quad \text{(discussions with Giwa, 1981)}$$

where H is scale height.

Since $\frac{w}{\sigma}$ is continuous, then

$$\frac{H}{\sigma}\frac{\omega}{p} - \frac{\sigma}{g(\alpha^2+\beta^2)}\frac{\partial\omega}{\partial p} \text{ is continuous.}$$

Infinite pressure gradient are never observed at the interface between air masses. If ever they occur, it is only for a very short time because air would quickly rush from a region of high pressure to that of low pressure. So, pressure is necessarily continuous at the interface.

i.e. $p(z+\varepsilon)$ is continuous.

$$p(z+\varepsilon) = p(z) + \varepsilon\frac{\partial p}{\partial z}$$

$$= P_s(z) + p'(z) + \varepsilon\frac{\partial P_s(z)}{\partial z} + \underline{\varepsilon\frac{\partial p'}{\partial z}}$$

$P_s(z)$ is a constant and the underlined term is a product of small quantities which makes it negligible; therefore,

$$p'(z) + \varepsilon\frac{\partial P_s(z)}{\partial z} \text{ is continuous .}$$

By definition, $\omega = \frac{Dp}{Dt}$

$$= \frac{\partial p'}{\partial t} + w\frac{\partial p}{\partial z} + u\frac{\partial p'}{\partial z}$$

Using 5.15 and the hydrostatic relation,

$$\omega = i\sigma p' - \rho g w$$

$$\text{i.e. } \frac{\omega}{i\sigma} = p' - \frac{\rho g w}{i\sigma} \qquad \ldots\ldots 6.3$$

It has been proved that

$p'(z) + \varepsilon\frac{\partial P_s(z)}{\partial z}$ is continuous. So, using 6.2,

$p'(z) + \frac{w}{i\sigma}\frac{\partial P_s(z)}{\partial z} = p'(z) - \frac{\rho g w}{i\sigma}$ is continuous.

From 6.3, this means

$\frac{\omega}{\sigma}$ is continuous.

For the scale of motion considered in this study, the surface of the earth is usually assumed to be flat. Because of the flatness, the vertical component of velocity is taken to be zero at the surface i.e. $w = 0$.

To summarise, the following terms are continuous at the interface between the air masses:

$$\frac{\omega}{\sigma}$$

and
$$\frac{H}{\sigma}\,\frac{\omega}{p} - \frac{\sigma}{g(\alpha^2+\beta^2)}\,\frac{\partial\omega}{\partial p} .$$

At the surface, $w = 0$ i.e. $\frac{H}{\sigma}\,\frac{\omega}{p} - \frac{\sigma}{g(\alpha^2+\beta^2)}\,\frac{\partial\omega}{\partial p} = 0$ and at the top, $\omega = 0$ (this is a free surface condition).

6-2 Mathematical derivations

To simplify mathematical derivations in our two-layer atmospheric model, parameters like density, temperature and horizontal velocity that should normally vary with height are taken to be constant within each layer. Subscripts '0', 'B' and 'T' indicate the values of these parameters at the surface, boundary between the air masses and the top of the atmosphere respectively while subscripts '1' and '2' are used to identify atmospheric parameters within the lower and top layers respectively.

We recall equation 5.24 i.e.

$$\frac{\partial^2\omega}{\partial p^2} + \frac{g\Gamma}{P_s^2}\,\frac{(\alpha^2+\beta^2)}{\sigma^2}\,\omega = 0 .$$

The solution to this differential equation is of the form

$$\omega = AP^{m_1} + BP^{m_2} \qquad \ldots.\ 6.4$$

where A and B are constants and m_1, m_2 are the roots of

$$\left[m(m-1) + \frac{g\Gamma(\alpha^2+\beta^2)}{\sigma^2} \right] = 0 \qquad \ldots 6.5$$

From 6.5, let

$$m_1 = \frac{1}{2} + \sqrt{\frac{1}{4} - \frac{g\Gamma(\alpha^2+\beta^2)}{\sigma^2}}$$

and $$m_2 = \frac{1}{2} - \sqrt{\frac{1}{4} - \frac{g\Gamma(\alpha^2+\beta^2)}{\sigma^2}}.$$

If $\lambda = \frac{1}{4} - \frac{g\Gamma(\alpha^2+\beta^2)}{\sigma^2}$

$$\omega = AP^{1/2+\lambda} + BP^{1/2-\lambda} \text{ (from 6.4).}$$

At the boundary between the air masses, it was shown that

$$\frac{H}{\sigma}\frac{\omega}{p} - \frac{\sigma}{g(\alpha^2+\beta^2)}\frac{\partial\omega}{\partial p}$$

and $\frac{\omega}{\sigma}$ are continuous.

At the surface, $w = 0$ and at the top $\omega = 0$.

Applying these conditions to 5.24 with the following substitutions:

$$Q_1 = \frac{\sigma_1^2}{g(\alpha^2 + \beta^2)}$$

and $$Q_2 = \frac{\sigma_2^2}{g(\alpha^2+\beta^2)}$$, two equations are obtained:

$$\left\{[H_B - Q_1(\lambda_1 + \tfrac{1}{2})] - [H_B - Q_1(\tfrac{1}{2} - \lambda_1)]\left[\frac{H_S + Q_1(\lambda_1 + \frac{1}{2})}{H_S + Q_1(\frac{1}{2} - \lambda_1)}\right]\left(\frac{P_o}{P_B}\right)^{2\lambda_1}\right\}\left(\frac{P_B}{\sigma_1}\right)^{\lambda_1} \cdot A_1$$

$$-\left\{[H_B - Q_2(\lambda_2 + \tfrac{1}{2})] - [H_B - Q_2(\tfrac{1}{2} - \lambda_2)]\left(\frac{P_T}{P_B}\right)^{2\lambda_2}\right\}\left(\frac{P_B}{\sigma_2}\right)^{\lambda_2} \cdot A_2 = 0 \qquad \ldots\ 6.6$$

$$\left\{1 - \left[\frac{H_S + Q_1(\lambda_1 + \frac{1}{2})}{H_S + Q_1(\frac{1}{2} - \lambda_1)}\right]\left(\frac{P_o}{P_B}\right)^{2\lambda_1}\right\}\left(\frac{P_B}{\sigma_1}\right)^{\lambda_1} \cdot A_1 - \left\{1 - \left(\frac{P_T}{P_B}\right)^{2\lambda_2}\right\}\left(\frac{P_B}{\sigma_2}\right)^{\lambda_2} \cdot A_2 = 0 \qquad \ldots\ 6.7$$

6.6 and 6.7 constitute a set of simultaneous equations in A_1 and A_2. For non-trivial solutions, the determinant of the coefficients must vanish.

If $$R_1 = \left[\frac{H_o + Q_1(\lambda_1 + \frac{1}{2})}{H_o + Q_1(\frac{1}{2} - \lambda_1)}\right]\left(\frac{P_o}{P_B}\right)^{2\lambda_1}$$

and $R_2 = \left(\frac{P_T}{P_B}\right)^{2\lambda_2}$, the value of this determinant is

$$\sigma_1^2 \lambda_1\left(\frac{1 + R_1}{1 - R_1}\right) - \sigma_2^2 \lambda_2\left(\frac{1 + R_2}{1 - R_2}\right) + \frac{\sigma_1^2}{2} - \frac{\sigma_2^2}{2} = 0 \qquad \ldots.\ 6.8$$

Recalling from chapter 5,

$$\sigma_1 = \sigma_o + \alpha u_1$$

$$\sigma_2 = \sigma_o + \alpha u_2 .$$

Substituting for σ_1 and σ_2 in 6.8,

$$\sigma_o^2 \{[2\lambda_1(\frac{1 + R_1}{1 - R_1}) + 1] - [2\lambda_2(\frac{1 + R_2}{1 - R_2}) + 1]\} +$$

$$\sigma_o\{2\alpha u_1[2\lambda_1(\frac{1 + R_1}{1 - R_1}) + 1] - 2\alpha u_2[2\lambda_2(\frac{1 + R_2}{1 - R_2}) + 1]\}+$$

$$\{\alpha^2 u_1^2[2\lambda_1(\frac{1+R_1}{1-R_1}) + 1] - \alpha^2 u_2^2[2\lambda_2(\frac{1+R_2}{1-R_2}) + 1]\} = 0 \qquad \ldots .\ 6.9$$

At a glance, 6.9 looks like an ordinary quadratic equation in σ_o but a closer observation reveals that the coefficients are not constants. This means that no analytic solution exists for 6.9. Thus, the equation reduces to an eigenvalue problem to determine σ_o such that there are non-zero solutions.

Some of the eigenvalues obtained in the solution of 6.9 will be complex and as said earlier, the imaginary part of such complex values would represent the amplification of any wave-like perturbation. The modes with the largest amplifications (i.e. the largest imaginary parts) are those that are proposed as being responsible for the onset of line squalls. These modes are those that will likely amplify up to the extent of blocking the 650 mb. mid-tropospheric jet as explained in the proposal for the initiation of line squalls (section 4-2). A single-layer model could be considered as a special case of equation 6.9.

6-3 Special case

In a one-layer model, $u_1 = u_2 = u$ (say) and $\lambda_1 = \lambda_2 = \lambda$ (say). If these are incorporated into 6.9,

$$\lambda[(\frac{1+R_1}{1-R_1}) - (\frac{1+R_2}{1-R_2})][\sigma_o^2 + 2\alpha u\sigma_o + \alpha^2 u^2] = 0 \qquad \ldots 6.10$$

The conditions for 6.10 to be true are:

(i) $\sigma_o^2 + 2\alpha u\sigma_o + \alpha^2 u^2 = 0$

i.e. $\frac{\sigma_o}{\alpha} = -u$.

This represents the velocity of the ambient wind without any perturbation whatsoever.

(ii) $\lambda = 0$.

This imlies $\sigma^2 = 4g\Gamma\alpha^2$.

For any single layer of air, this condition represents a phase velocity of

$$\frac{\sigma}{\alpha} = u \pm 2\sqrt{g\Gamma}$$

where $2\sqrt{g\Gamma}$ 'represents' the perturbation on the ambient wind u. As considered in section 5-3, if Γ is positive, the perturbation propagates, without any amplification, through the ambient winds. On the other hand, if Γ is negative, the perturbation amplifies with time without any propagation. These conditions depict wave-like perturbations in dry and moist air respectively.

(iii) $\left(\frac{1+R_1}{1-R_1}\right) - \left(\frac{1+R_2}{1-R_2}\right) = 0.$

Substituting for R_1 and R_2,

$$\left(\frac{P_T}{P_S}\right)^{2\lambda} = \frac{1 + 2\lambda - \frac{2H_0}{Q}}{1 - 2\lambda - \frac{2H_0}{Q}} \qquad \ldots\ldots \; 6.11$$

Equation 6.11 had been obtained by Millier (1974) while solving a frequency equation for the speed of propagation of external gravity waves. These are 'free-surface' waves which occur in one-layer models through the use of particular boundary conditions. Simplified cases of these waves in hydrostatic, neutral and non-hydrostatic, neutral systems have been discussed by Miller (1974).

CHAPTER 7

SUMMARY AND CONCLUSION

In this study, the importance of line squalls and the reasons for studying the phenomenon have been stressed. In order to understand the various stages in the development of line squalls, the West African synoptic weather pattern was reviewed in chapter 2. In the review, it was noted that line squalls are most-frequently observed whenever an area experiences the weather Zone C of Hamilton and Archbold (1945). Various methods that had been used to study these squalls (i.e. observational investigations, satellite investigations and numerical simulations) were mentioned and the problems that still exist highlighted; the main problem being the isolation of the formative cause of line squalls.

Although observational investigations of West Africa disturbances were not very successful in isolating the trigger mechanism of line squalls, they offer good descriptions of line squalls such that the phenomenon can now be easily identified. For instance, Hamilton and Archbold (1945) and Bolton (1981) described how D.L manifests itself to a surface observer.

As recorded by observational investigations, a cross-section through a D.L. along its direction of propagation shows that the updraught slants over the downdraught in a way that permits the precipitation from the former to fall into and possibly evaporate in the latter. This mechanism permits the persistence of line squalls. The main shortcoming of observational investigations is that they lack any synthesis of the various stages in the development of line squalls.

Satellite investigations of tropical storms provided a better alternative to observational investigations as far as the synthesis of various experimental analyses are concerned. This is so because sequential analysis of successive satellite photographs could show the various stages in the development of line squalls. On such satellite photographs, the following criteria could be used to identify D.L.'s:

1. a D.L. should consist of large cumulonimbus clouds with tops to about 15 km;
2. such bright cloud clusters should attain a size of at least $2^o x\ 2^o$ during its life-time and
3. a D.L. should be active for a minimum of 6h.

Line squalls identified this way fall into two categories according to the pattern of satellite pictures: type I squalls have a well-defined arc-shaped front that faces the direction of propagation while type II squall fronts appear like indented arcs. The amount of information that could be extracted from a single satellite photograph is so much that this form of investigation is probably the cheapest means of studying tropical disturbances.

Limited access to satellite data along with the hazards, and cost, of observational investigations were mentioned as some of the reasons which favour the option of modelling in the study of tropical disturbances. There are two main categories of models: numerical and mathematical. Mathematical models could be further separated into gravity wave models and non-linear steady state models. Non-linear steady state models are applicable to the mature stage in the evolution of line squalls while the gravity wave models are appropriate in describing the initial development of the storms. Since this study is primarily interested in the initial stages in the development of line squalls, the gravity wave model was developed.

A review of numerical models showed how the basic hydrodynamical equations of motion could be utilised in two-dimensional numerical experiments; the discussion being based principally on the work of Schlesinger (1973). Some specific findings of two-dimensional numerical models are:

1. the conclusion drawn from Takeda (1971) that two air masses of different atmospheric stabilities and oppositely directed to one another are essential before 'long - lasting' clouds could develop;

2. the deduction from the work of Hane (1973) that initial perturbation is not very important in simulating tropical disturbances because there is always a tendency towards convergence upon a common solution irrespective of the initial perturbations assumed and

3. the observation of Schlesinger (1973) that the intensity of convection increases with moisture supply from the lower air mass and decreases with mid-tropospheric wind shear.

Three-dimensional numerical models were discussed as extensions of two-dimensional models; the extension being necessary because values of in-cloud parameters obtained from two-dimensional models are usually larger than observed values. Furthermore, such extensions are more realistic because the line squall phenomenon is three-dimensional in nature.

Notable among three-dimensional models is the work of Ogura and Charney (1962) who derived a frequency equation from which they showed that wave-like disturbances generated in a typical pre-storm 'tropical' environment amplifies with time. This study extended their work to describe waves generaged along the surface of discontinuity between dry and moist air masses.

From numerical experiments, basic atmospheric features that are essential before line squalls could develop were deduced. These features include:

1. a two-layer air mass wherein the lower layer is convectively unstable;
2. cold front zones (i.e. borders between warm and cold air masses);
3. a moisture profile that drops between 600 mb. and 500 mb pressure levels;
4. a mid-tropospheric wind maximum and
5. strong vertical shear.

These features were shown in section 4-1, to be present in West Africa during the wet season.

Another possible feature that was mentioned is the existence of an inversion or isothermal layer in the atmosphere. The Pressure Jump Line theory of Tepper (1950) on the trigger mechanism of line squalls was based on this feature. This theory links the formation of line squalls with the development of a pressure jump which is a consequence of the sudden acceleration of the cold front in a border between warm and cold air masses wherein an inversion or isothermal layer exists in the warm zone. Some observed features of the West African line squalls that cannot be easily explained by this theory are:

1. the almost-uniform speed of propagation of line squalls;

2. the observed overturning of the atmosphere after the passage of squalls and
3. the preference of highlands as source regions of line squalls.

LeRoux (1976) also made an attempt at formulating the trigger mechanism of line squalls. The defect in his explanation is that his model could not justify the self-sustaining nature of line squalls. This study's presentation of the trigger mechanism of line squalls was a modification of the proposal of LeRoux.

The trigger mechanism of line squalls proposed in chapter 4 is based on the amplification of any wave-like disturbance along the surface of discontinuity between south-westerlies and north-easterlies. The waves generated along the surface of discontinuity could amplify up to the extent of blocking the 650 mb. jet. The jet, by impinging on the surface, further distorts the 'bump' that has been formed and thereby forces the monsoon air to rise more. The rising monsoon air could precipitate and the precipitates fall into (with some of them evaporating in) the underlying jet. The jet, now cooler, sinks and while sinking, could hit the surface of the earth especially if the whole process occurred over a table-land. On the surface, a squall front is formed as a result of strong convergence.

As the squall front moves over the surface, it lifts the convectively-unstable south-westerlies ahead of it and since the depth of this air mass during the wet season is enough to sustain large clouds, cumulonimbus clouds could form. The extent of line squall activities over the surface depends on how far the jet could travel on the surface. This picture of the trigger mechanism of line squalls seems to be more adequate than previous presentations because it has been able to account for a number of observed features of the disturbance. Among others, this proposal has accounted for the following:

1. the limitation of West African line squalls to a particular season of the year (line squalls are limited to the wet season because this is the period when the weather Zone C of Hamilton and Archbold (1945) cover a substantial area of land and as shown, all essential atmospheric features that aid the development of line squalls are present in any area experiencing the weather Zone C),
2. the preference of highlands as places of origin of line squalls,
3. the close association between the speeds of propagation of line squalls and the mid-tropospheric jet,
4. the orientation and direction of movement (WSW) of D.L.'s,

5. the role of precipitation in the sustenance of line squalls and

6. the observed overturning of the atmosphere after the passage of line squalls.

While discussing the evolution of line squalls in chapter 4, the two scales of downdraughts in a fully-developed line squall system were mentioned. One of these downdraughts is within the lowest 200-400 m behind the squall front. The other scale of downdraughts has lower wet-bulb potential temperature, θ_{ω}, and probably originate from areas of low θ_{ω} (i.e. pressure level 750 mb. and above). In a fully-developed squall system, heavy rainfall that immediately follow the squall front cause a rise in pressure (mesohigh) while about 100 km behind the front, there exists a mesolow.

To justify the proposed trigger mechanism mathematically, chapter 5 and 6 were devoted to the development of a gravity wave model for initiation of line squalls. From the basic hydrodynamical equations governing atmospheric motions, perturbation equations were derived (in pressure co-ordinates). One advantage of the use of pressure co-ordinates is that more data on atmospheric parameters are available along isobaric surface. Further, the frequency equation, which was derived from the set of perturbation equations, appears neater in pressure co-ordinates.

In the derivation of the frequency equation in chapter 5, a new definition emerged for the static stability of the atmosphere. The effects of this stability term on the possible solutions to the frequency equation were discussed in section 5-3. It was shown that in a moist environment, the condition for wave-like solution to the frequency equation is that any wave-like perturbation must amplify with time without any propagation while in a dry environment, any wave-like perturbation must propagate, without any amplification, with time. In a situation wherein two air masses, one on top of the other, have different stabilities, waves generated along the surface of discontinuity would amplify as well as propagate with time. This is the situation along the surface of discontinuity between the north-easterlies and the south-westerlies in West Africa.

Such amplifying patch of convection consist of many modes of different growth rates and phase velocities. All possible phase velocities will be solutions to equation 6.9 which was obtained by solving the frequency equation with the aid of a two-layer model of the atmosphere. Some phase velocities which are possible solutions to equation 6.9 will be complex; the real part representing the phase velocity while the imaginary part is a measure of the amplification of the wave-like disturbance. Waves with the largest amplifications (i.e. largest imaginary parts) along the surface of discontinuity between the north-easterlies and the south-westerlies are those that might likely block the 650 mb. mid-tropospheric jet and thereby initiate line squalls.

This study is by no means exhaustive on West African line squalls. For instance, the role, if any, of the African easterly waves in the process of initiation of the disturbance needs further study. Dhonneur (1970) noted that the two phenomena - easterly waves and line squalls - could interact in spite of the fact that they are distinct. A possible interaction is, as mentioned in the thesis, the observed coincidence between the arrival of easterly wave troughs and the formation of line squalls. If this coincidence were true, there should be a periodicity in the occurrence of line squalls since the easterly wave trough would normally arrive in West Africa at a regular interval of about 3.5 days. This periodicity would, however, be subject to the existence of other atmospheric parameters that are essential to the development of line squalls.

Hamilton and Archbold (1945) and Aspliden et al. (1976) have observed that line squalls are generated during the period of maximum isolation. It will be necessary to investigate how this observation fits into the model proposed in this study on the trigger mechanism of line squalls.

The determination of the particular wavelength that gives rise to maximum amplification in the possible solutions to equation 6.9 is an avenue for further numerical work.

The dissipation of line squalls needs further investigation. Although few suggestions were made on the dissipative process of line squalls, there is need to fully justify the reason(s) why line squalls dissipate completely around the coasts of West Africa.

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