COMMAND AND STAFF TRAINING INSTITUTE

BANGLADESH AIR FORCE

Individual Staff Studies Programme (ISSP)

PROFESSIONAL SUBJECT -2 : METEOROLOGY
PHASE-16 : PART-II

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PHASE – 16 : PART-II

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CONTENTS

Ser No	Topic		Page No
1.	**Conduct to the Phase**		iv
	Meteorological Organizations.		01-09
	Task-1	World Meteorological Organization.	01
	Task-2	Bangladesh Meteorological Organization.	04
	Task-3	Bangladesh Air Force Meteorological Branch.	07
Sec-2	**Dynamic Meteorology.**		10-62
	Task-4	Hydrostatic Equilibrium.	10
	Task-5	Atmospheric Dynamics.	13
	Task-6	Horizontal Frictional Flow.	28
	Task-7	Variation of Horizontal Wind in Vertical.	42
	Task-8	Circulation and Vorticity.	49
Sec-3.	**Weather Systems**		63-140
	Task-9	Thunderstorm and Tornado.	63
	Task-10	Western Disturbance.	78
	Task-11	Tropical Cyclone.	83
	Task-12	Jet Stream.	101
	Task-13	Tropical Phenomena.	108
	Task-14	Atmospheric Circulation.	117
Sec-4.	**Analysis, Prognosis & Forecasting**		141-191
	Task-15	Stream Line Analysis.	141
	Task-16	Forecasting Mintra Level.	159
	Task-17	Forecasting Ice Accretion and Super Cooled Water.	161
	Task-18	Forecasting Thunderstorm.	167
	Task-19	Forecasting Fog and Low Stratus.	175
	Task-20	Forecasting Temperature.	185
Sec-5.	**Satellite Meteorology**		192-215
	Task-21	Satellite Imagery Interpretation.	192
	Task-22	Interpretation of Earth Surface.	199
	Task-23	Interpretation of Synoptic Scale Cloud.	205
	Task-24	Jet Stream as seen in Satellite Picture.	214
Sec-6.	**Monsoon Circulation**		216-239
	Task-25	Morphology of Monsoon Circulation.	216
	Task-26	Characteristic Features of South West Monsoon.	222
	Task-27	Onset & Withdrawal of South West Monsoon.	227
	Task-28	Monsoon Depression.	234
	Task-29	Break Monsoon.	238
Sec-7.	**Intensity and Movement of Tropical Cyclone**		240-276
	Task-30	Intensification of Tropical Cyclone	240
	Task31	Movement of Tropical Cyclone	270
Sec-8.	**Climatology**		277-348
	Task32	Climatology and World Weather	277
	Task33	Climatology of Bangladesh	312
	Task34	Climatology of Tibet & Himalayas	320
	Task35	Climatology of India	330

CONDUCT OF THE PHASE

PART-II, PHASE-16
SUBJ : PROFESSIONAL-2 (METEOROLOGY)

Ser No		Topic	Period Distr	Total Period
1.	**Meteorological Organizations.**			
	a.	World Meteorological Organization.	1	3
	b.	Bangladesh Meteorological Organization.	1	
	c.	Bangladesh Air Force Meteorological Branch.	1	
2.	**Dynamic Meteorology.**			
	a.	Hydrostatic Equilibrium.	2	14
	d.	Atmospheric Dynamics.	3	
	e.	Horizontal Frictional Flow.	3	
	f.	Variation of Horizontal Wind in Vertical.	3	
	g.	Circulation and Vorticity.	3	
3.	**Weather Systems**			
	a.	Thunderstorm and Tornado.	2	13
	b.	Western Disturbance.	2	
	c.	Tropical Cyclone.	3	
	d.	Jet Stream.	1	
	e.	Tropical Phenomena.	2	
	f.	Atmospheric Circulation.	3	
4.	**Analysis, Prognosis & Forecasting**			
	a.	Stream Line Analysis.	3	10
	b.	Forecasting Mintra Level.	1	
	c.	Forecasting Ice Accretion and Super Cooled Water.	1	
	d.	Forecasting Thunderstorm.	2	
	e.	Forecasting Fog and Low Stratus.	2	
	f.	Forecasting Temperature.	1	
5.	**Satellite Meteorology**			
	a.	Satellite Imagery Interpretation.	3	10
	b.	Interpretation of Earth Surface.	2	
	c.	Interpretation of Synoptic Scale Cloud.	3	
	d.	Jet Stream as seen in Satellite Picture.	2	
6.	**Monsoon Circulation**			
	a.	Morphology of Monsoon Circulation.	3	11
	b.	Characteristic Features of South West Monsoon.	3	
	c.	Onset & Withdrawal of South West Monsoon.	1	
	d.	Monsoon Depression.	2	
	e.	Break Monsoon.	2	
7.	**Intensity and Movement of Tropical Cyclone**			
	a.	Intensification of Tropical Cyclone	3	4
	b.	Movement of Tropical Cyclone	1	
8.	**Climatology**			
	a.	Climatology and World Weather	2	5
	b.	Climatology of Bangladesh	1	
	c.	Climatology of Tibet & Himalayas	1	
	d.	Climatology of India	1	
9.	**Preparation and Revision for TAE**			10
		Total		**80**

INTRODUCTION TO THE PHASE

Scope of the Phase

1. This phase deals with professional study materials on the subject of Meteorology which has been divided into two parts Part-I and Part-II. The materials contained in this volume constitute Part-II of the phase. The individual topic has been so presented as to help you gain theoretical background leading to its practical utilization in forecasting.

Guidance on Methods of Study.

2. In addition to this phase note you should study different manuals and Publications pertaining to this subject to improve your knowledge.

3. Self assessed exercises of different tasks are given in the phase note. Your study during the period is vital and your exercise results will reflect how much benefit you have derived forms it. You should try to solve some problems those come up during your study especially with Dynamic and Synoptic Meteorology.

SECTION – 1 : METEOROLOGICAL ORGANISATION

TASK-1 : WORLD METEOROLOGICAL ORGANISATION

Introduction

1. The earth's atmosphere is a whole, which cannot be divided to conform to political frontiers. Moreover, weather and climate in one place are normally influenced by conditions in distant areas. For these reasons, the nations of the world recognized as early as the last century the necessity of international co-operation to study the vagaries of the weather and to apply the knowledge gained for the benefit of mankind.

2. International collaboration in meteorology began in 1853 when a meeting of sea-going nations drew up a programme of weather observations over the oceans to contribute to the safety of life at sea.

3. To organize the necessary international collaboration in making and collecting weather observations, not only at sea but also on land, the International Meteorological Organization (IMO) was established on 23rd March, 1873 when 32 prominent meteorologists of 20 countries assembled in Viena for the purpose. A system for exchanging weather information between different countries was also set up by IMO.

4. During the present century, the development of aviation created a new need for detailed weather information. This gave a tremendous impetus to the study of weather. At the same time revolutionary advances in other branches of science and technology, such as radio communications greatly helped meteorologists in meeting this new challenge.

5. The creation of the United Nations provided a new framework for international collaboration in various areas, including technical and scientific fields. Consequently, in Washington in 1947 the World Meteorological Convention was adopted, establishing a new organization founded on a formal agreement between governments.

6. This convention was ratified by a large number of States and in 1951 the new World Meteorological Organization (WMO) replaced IMO, taking over its activities and traditions. In December, 1951, the General Assembly of the United Nations approved an agreement between the United Nations and the World Meteorological Organization by which the latter was recognized as a specialized agency of the United Nations.

World Meteorological Organization (WMO)

7. **Aims and Objectives.** The WMO is a specialized agency of the United Nations, created to co-ordinate, standardize and improve the services rendered by meteorology throughout the world to various human activities. It consists of more than 130 Member countries and territories, each maintaining its own Meteorological Service. The World Meteorological Organization has the following aims and objectives:

a. To facilitate international co-operation in the establishment of network of stations and centers to provide meteorological and hydrological services and observations.

b. To promote the establishment and maintenance of systems for the rapid exchange of meteorological and related information.

c. The promote standardization of meteorological and related observations and ensure the uniform publication of observations and statistics.

d. To further the application of meteorology to aviation, shipping, water problems, agriculture and other human activities.

e. To promote activities in operational hydrology and to further close co-operation between meteorological and hydrological services.

f. To encourage research and training in meteorology and as appropriate, in related fields.

8. **Structure of WMO.** The World Meteorological Organization comprises of the following important body, committee, association and commission:

a. The WORLD METEOROLOGICAL CONGRESS is the supreme organ of the organization. Representing all Members, it meets once every four years and fixes the policy, programme and budget of WMO.

b. The EXECUTIVE COMMITTEE, meeting at least once a year, prepares studies and recommendations for the Congress, supervises the implementation of Congress resolutions and informs and offers advice to Members on technical matters.

c. WMO Members are grouped into six REGIONAL ASSOCIATIONS, whose task is to co-ordinate meteorological activity within their respective regions.

d. Eight TECHNICAL COMMISSIONS, composed of experts, study the applications of meteorology, and problem and developments in specialized fields.

9. Completing the structure is the SECRETARIAT in Geneva, Switzerland. Under the Secretary-General, it carries out the technical and administrative work of WMO, is responsible for the various technical co-operation programmes, undertakes technical studies, prepares specialized publication and provides the secretariat for various WMO meetings.

Self-assessed Exercise on Task-1

Q1. What is IMO and when it was formed ?

Q2. Does the IMO still exist ?

Q3. What is the replacement of IMO ?

Q4. What is WMO and when it was established ?

Q5. Where is located the Secretariat of WMO ?

Q6. Write down the aims and objectives of WMO.

Answers to Self-assessed Exercise on Task-1

A1. This is International Meteorological Organisation. It was formed on 23rd March, 1873.

A2. No.

A3. WMO.

A4. This is World Meteorological Organisation, a specialized agency of United Nations. It was established in December 1951.

A5. In Geneva Switzerland.

A6. See Para 7.

TASK-2 : BANGLADESH METEOROLOGICAL DEPARTMENT

1. Like other agencies to the aid of human activities in the field of various development, Bangladesh Meteorological Department has been set up under the administrative control of the Ministry of Defence, Government of the People's Republic of Bangladesh to render services in respect of safety of aviation and marine activities, help growth of agricultural productions and extend aid to the development of Hydrology and Oceanography. Bangladesh is an active member of WMO since 1973.

2. Bangladesh Meteorological Department has been able to establish a good position in the World Meteorological Organization and is participating very actively in the World Weather Programme. After the independence of Bangladesh, the set up of the then meteorological not works available at that time was not sufficient.

3. The Department with the assistance of the World Meteorological Organization, has recently established a Meteorological Training Institute to train the personnel of this Department and other meteorologists in their professional fields. A modern IBM S-34 Computer has recently been installed with 48 K memory for achieving the meteorological observations of all the observing networks for research and planning necessities, in the newly formed Climatological Division of the Department. Bangladesh Meteorological Department has also established a modern Mechanical Workshop and Laboratory which is unique of its kind in this sub-continent. The Department has also taken up an elaborate scheme to develop the Meteorological Service by raising it upto the standard of a modern Meteo9orological Service in the next Five Year Plan, in which, priorities have been given on the following schemes:

a. Improvement of Telecommunication Service.

b. Improvement of Aviation Meteorology.

c. Addition of an improved Meteorological Radar for Hydrological uses.

d. Establishment of Agrimeteorological Services.

e. Storm Surge Research.

f. Establishment of a Marine Meteorological Study Centre leading to special studies on surface hydrology, oceanography and air-sea Inter-action.

4. Its existing various divisions like – Synoptic, Forecasting, Climatology, Electronics, Communication etc., functions under the control of the direction. By virtue of the national set up and position Bangladesh Meteorological Department also represents its country in the International Civil Aviation Organisation (ICAO) in the field of aviation meteorology.

5. As a matter of guideline the Bangladesh Meteorological Department extends the following assistances/aids to Air Force Meteorology :

a. Technical guideline as and when sought for.

b. Routine supply of international changes/modifications in codes/procedures

etc. in different parts of Meteorology.

c. Training facilities to officers and airmen in the Institute of meteorology at Sher-e-Bangla Nagar, Dhaka as and when a formal request is made by Air Headquarters.

d. Supply of all meteorological instruments, forms charts and consumable stores required for functioning of Departmental observatories manned by Air Force.

Self-assessed Exercise on Task-2

Q1. Under which Ministry Bangladesh Meteorological Department is set up ?

Q2. What are the assistances/aids Bangladesh Meteorological Department iextends to Air Force Meteorology ?

Answers to Self-assessed Exercise on Task-2

A1. Ministry of Defence.

A2. See Para 5.

TASK-3 : BANGLADESH AIR FORCE METEOROLOGICAL BRANCH

1. To provide meteorological facilities required for Air Force operations a meteorological branch has been organized in Bangladesh Air Force. In the national net-work, though technically assisted by Bangladesh Meteorological Department, the branch functions under the entire control of the Chief of Air Staff, Bangladesh Air Force.

2. A directorate of meteorology has been established under the Assistant Chief of Air Staff (Operations and Training) at Air Headquarters to formulate and implement plans and policies regarding meteorological facilities in Bangladesh Air Force. This unit remains responsible to the Assistant Chief of Air Staff (Operations and Training) for the following commitments:

a. Formulation of policy and planning of meteorological facilities in BAF.

b. Standardization of procedures and overall supervision of meteorological services in BAF formation in respect of observation, plotting, analysis of weather charts and forecasting.

c. Liaison with the Director of Bangladesh Meteorological Department in all technical matters on meteorology relating to practices and procedures in analyzing and forecasting. Implementation of technical procedures and practices in respect of observation, analysis and forecast, issued by WMO and subsequently communicated by the Director of Bangladesh Meteorological Department.

d. Liaison with Army and Naval Headquarters on matters relating to meteorological policy.

e. Planning and assessing the requirements of meteorological consumable stores, instruments, forms, charts etc. and advise Director of Supply for their procurement including printing of meteorological forms.

f. Advising Air Secretary and Director of Personnel on personnel matters concerning meteorological officers and airmen.

g. To advise Director of Training for arranging professional training of meteorological officers and airmen both at home and abroad and also to plan the training standard.

h. Policy, control and direction of Military Climatological Division and research activities.

j. To advise Director of Education and Director of Trg for procurement of publications, periodicals, journals and text books concerning meteorology.

k. To advise Director of Communications and Electronics on all matters related to Met Tele-communication and broadcast facilities.

l. Compiling and preparing airfield climatological summaries and tables of

surface and upper air meteorological elements.

m. Preparing climatography of aerodromes and regions of strategic importance.

n. Conducting investigations and developing long range forecasting methods for operational requirements.

p. Carrying out study research and developing objective forecast techniques on critical weather factors.

q. Keeping close co-ordination with Equipment Depot for timely supply of meteorological equipment, form and charts to meteorological squadrons.

r. Scrutinizing of met, forecast and warnings.

s. Check of meteorological instruments hold by meteorological squadrons for proper use and maintenance.

3. In the International Field of Meteorology, Air Force Meteorological Branch, is to maintain liaison as and when necessary, with the Director, Bangladesh Meteorological Department, who represents Bangladesh in the World Meteorological Organization (WMO) and International Civil Aviation Organization (ICAO).

Self-assessed Exercise on Task-3

Q1. To whom directorate of Meteorology remains responsible for its commitments ?

Q2. What are the functions of the Directorate of Meteorology in Air Headquarters ?

Answers to Self-assessed Exercise on Task-3

A1. Assistant Chief of Air Staff (Operations & Training).

A2. See Para 2.

SECTION-2 : DYNAMIC METEOROLOGY

TASK – 4 : HYDROSTATIC EQUILIBRIUM

Introduction

1. Hydrostatics is concerned with the mechanics of stationary fluids. In the case of a parcel of air in the atmosphere there will be no vertical motion if there is a complete balance between force of gravity and an upward directed force (pressure gradient force) arising from the normal decrease of pressure with altitude. This is known as hydrostatic equilibrium. In this task we shall first deal with force of gravity and the geopotential of a parcel of air. After that hydrostatic equation will be developed which is applicable to hydrostatic equilibrium. The hydrostatic equation is also used to determine the thickness of an atmospheric layer which lies between two pressure levels.

2. **The Force of Gravity**. By combining the effects of earths force of gravitation and the centrigugal force, we have force of gravity from equation

$$W = mg \quad \text{---------- (1)}$$

where W = force of gravity. It is seen from the above equation that the force of gravity on a parcel of unit mass is equal to the acceleration (g) due to gravity.

3. **Geopotential**. A body possesses energy due to its position in the earths gravitational field. The value of this energy depends on the magnitude of the force of gravity acting in the body and the vertical distance of the object above the standard reference level. Mean sea level is considered as the standard reference level. Mean sea level is considered as the standard reference level for measuring the gravitational potential energy. Let us take a sameple of air of mass one Kg at an altitude of z metre and located at a point where the acceleration of gravity is g metres S^{-2}. The gravitational potential energy is equal to mgz joule. So, the gravitational potential energy per unit mass is called the geopotential (Φ) of the sample of air i.e.

$$\Phi = gz \quad \text{---------- (2)}$$

The SI unit of specific energy (i.e. energy per unit mass) is the Joule per kg. WMO has adopted a unit of geopotential (i.e. gravitational potential energy per unit mass) which is 9.8 times greater than the SI unit. It is called geopotential metre (gpm). Hence,

1 geopotential metre = 9.8 Juule/Kg----------(3)

A geopotential meter would be the gravitational potential energy possessed by a sample of air of mass one Kg, if it were located one metre above MSL at a location where g = 9.8 m S^{-2}. From Eqn (2) & (3) it is clear that the geopotential (Φ)

$$\Phi = \frac{g}{9.8} z \text{ gpm} \quad \text{---------- (4)}$$

In reality, g does not differ from 9.8 m S^{-2}. Hence the magnitude of the geopotential in geopotential metre is approximately equal to the altitude (i.e. distance form MSL) of the

sample of air expressed in metres. For e.g. if the geopotential is 850 gpm then the jparcel of air is located approximately 850 m above MSL.

4. **<u>The Hydrostatic Equation.</u>** Let us consider a small cylindrical element of unit cross section as shown below :

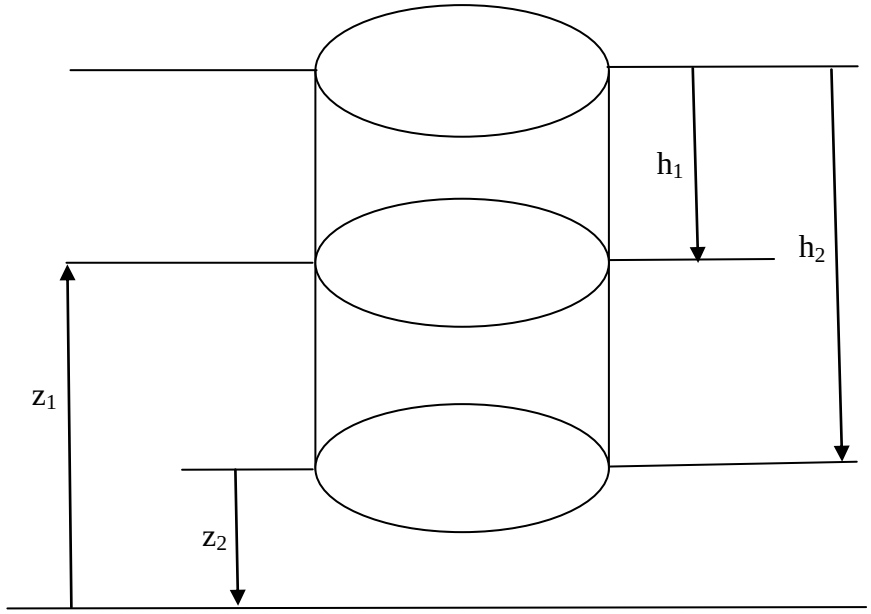


Fig-1

Now $P_1 = \rho g h_1$ pressure at p_1

$P_2 = \rho g h_2$ pressure at P_2

$P_2 = P_1 = \rho g (h_2 - h_1)$

But $h_2 - h_1 = z_1 - z_2$

Hence, $P_2 - P_1 = \rho g (z_1 - z_2)$

This equation is usually written as

$$P_1 - P_2 = -\rho g(z_1 - z_2) \qquad \text{----------(5)}$$

The minus sign indicates that as z increases in the atmosphere, the pressure decreases. Equation (5) is known as hydrostatic equation. It may be written as

$P = -\rho g \Delta z$ or in differential form

$$dp = -\rho g dz \qquad \text{---------- (6)}$$

Self Assessed Exercise on Task – 4

Q1. What do you understand by geopotential ?

Q2. Deduce Hydrostatic Equation.

Answers to Self Assessed Exercise Task – 4

A1. See Para 3.

A2. See Para 4.

TASK-5: ATMOSPHERIC DYNAMICS

Fundamental and Apparent Forces

1. **Fundamental Forces.** Forces which obey Newton's 2nd law in a coordinate system fixed in space:

   a. Pressure Gradient Force,

   b. Gravitational Force

   c. Friction

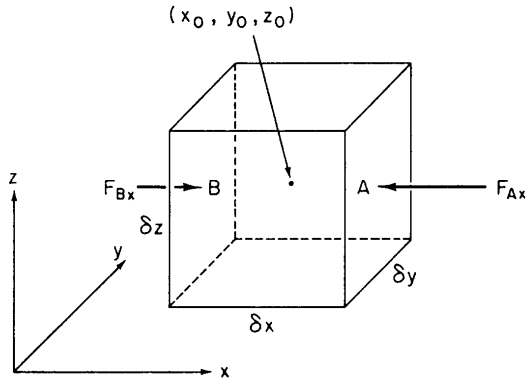
2. **Apparent Forces**. Forces which are present in an non-inertial coordinate system:

   a. Centrifugal Force

   b. Coriolis Force

3. **Pressure Gradient Force.** Consider a differential volume element of air

□***V*** **=** □***x***□***y***□***z***, centred at a point (x_o, y_o, z_o) where the pressure is p_o.



The x component of the pressure gradient force acting on a fluid element.

Pressure at wall *A* is given by

$$p_0 + \frac{\partial p}{\partial x}\frac{\delta x}{2} + \text{higher- order terms}$$

Neglecting the higher-order terms in this expansion, the pressure force acting on the volume element at wall *A* is

$$F_{Ax} = -\left(p_o + \frac{\partial p}{\partial x}\frac{\delta x}{2}\right)\delta y \delta z$$

where □*y*□*z* is the area of wall *A*. Similarly, the force acting on the volume element at wall *B* is just

$$F_{Bx} = +\left(p_o - \frac{\partial p}{\partial x}\frac{\delta x}{2}\right)\delta y \delta z$$

Therefore, the net *x*-component of the pressure force acting on the volume is

$$F_x = F_{Ax} + F_{Bx} = -\frac{\partial p}{\partial x}\delta x \delta y \delta z$$

The mass *m* of the differential volume element with density □ is

$$m = \square\ \square x \square y \square z$$

Thus, the *x*-component of the pressure gradient force per unit mass is

$$\frac{F_x}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$

Similarly, it can easily be shown that the *y* and *z* components of the pressure gradient force per unit mass are

$$\frac{F_y}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial y} \quad \text{and} \quad \frac{F_z}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial z}$$

so that the total **pressure gradient force** per unit mass is

$$\boxed{\frac{\mathbf{F}}{m} = -\frac{1}{\rho}\nabla p}$$

It is important to note that this force is proportional to the *gradient* of the pressure field, not to the pressure itself.

4. **<u>Gravitational Force.</u>**

$$\mathbf{F}_g = -\frac{GMm}{r^2}\hat{\mathbf{r}}$$

where G is the gravitational constant, M the mass of the earth, m the mass of object at a distance r from the centre of the earth.
In atmospheric science, we use z as the vertical coordinate which is defined as

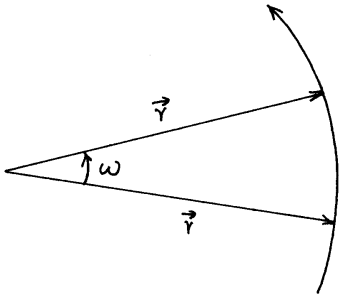
$$r = a + z$$

where *a* is the radius of the earth. Since *z* is a variable, $\mathbf{F}_g$ is also a variable. But $z << a$, so that the gravitational force per unit mass is

$$\mathbf{g}^* = \frac{\mathbf{F}_g}{m} = -\frac{GM}{a^2}\hat{\mathbf{r}} \cong \text{constant}$$

Friction or Viscous Force- to be discussed later.

5. <u>Centrifugal Force.</u>



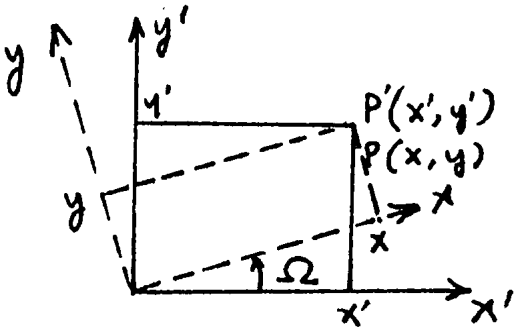
An object in a rotating frame of reference will have a centrifugal force which is the inertial reaction of the object on the centripetal force given by

$$\frac{\mathbf{F}}{m} = \omega^2 \mathbf{r}$$

where □ is the angular velocity.

<u>Dynamical Equations in a Rotating Coordinate System</u>

6. Consider a coordinate system rotating at a constant angular velocity □.



The equations describing the coordinates relative to the fixed system [denoted by (')] are:

$$\begin{pmatrix} x' \\ y' \end{pmatrix} = \begin{pmatrix} \cos\Omega t & -\sin\Omega t \\ \sin\Omega t & \cos\Omega t \end{pmatrix} \begin{pmatrix} x \\ y \end{pmatrix}$$

In the vertical (i.e. along axis of rotation), $z' = z$.

The forces along the three axes can be written in a similar way. Then, if the above equations are differentiated twice and the accelerations equated with the forces, we find

$$\ddot{x} - \frac{F_x}{m} = x\Omega^2 + 2\dot{y}\Omega$$

$$\ddot{y} - \frac{F_y}{m} = y\Omega^2 - 2\dot{x}\Omega$$

$$\ddot{z} - \frac{F_z}{m} = 0$$

where F_x, F_y and F_z are the component forces in the rotated system and (¨) means $d^2(\ )/dt^2$.

7. In the 1st two equations, 1st term on RHS - centrifugal force (per unit mass) which combines to give $\square^2 r$, where $r = \square\ (x^2 + y^2)$.

2nd term on RHS -accelerations due to the combined effects of rotation of the coordinates and motion of a particle relative to the rotating system- **Coriolis force** per unit mass

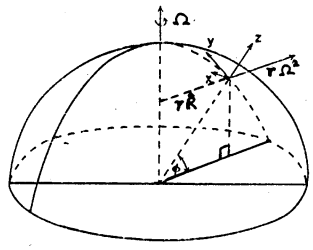
8. **<u>Properties of the Coriolis force</u>**

a. Only those components of motion (measured relative to a rotating frame) on a plane $\square$ to axis of rotation are affected;

b. For a counterclockwise rotation, this force is directed to the *right* of the component of motion on the same plane;

c. Because the force due to a motion in the *x*-direction (i.e. $\dot{x}$) is along the *y*-direction, this force only changes the direction of motion (i.e. a deflection) but *not* the speed.

9. Apply the equations to the earth:



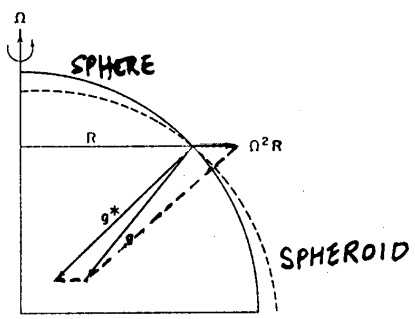
a. **<u>Centrifugal Force</u>**:

$$r\,\Omega^2\ \hat{\mathbf{R}} = -a\,\Omega^2 \cos\phi \sin\phi\,\hat{\mathbf{y}} + a\,\Omega^2 \cos^2\phi\,\hat{\mathbf{z}}$$

b. **<u>Coriolis Force</u>**:

DIRECTION OF MOTION	COMPONENT FORCE(S)
eastward ($u\hat{\mathbf{x}}$):	$-2\Omega u \sin\phi\,\hat{\mathbf{y}} + 2\Omega u \cos\phi\,\hat{\mathbf{z}}$
northward ($v\hat{\mathbf{y}}$):	$2\Omega v \sin\phi\,\hat{\mathbf{x}}$
upward ($w\hat{\mathbf{z}}$):	$-2\Omega w \cos\phi\,\hat{\mathbf{x}}$

10. **<u>Gravitation vs. Gravity</u>**



Because of the presence of the centrifugal force, a rotating earth that is spherical cannot be an equilibrium configuration. For equilibrium to exist, the earth must be an *oblate spheroid* the surface of which is ⊥ to the vector sum of the two forces. This vector is known as gravity. Since particles on these surfaces are "balanced", the surfaces are also levels of constant potential energy and therefore known as equipotential surfaces.

Combining all of the above gives the following set of **equations of motion**:

$$\boxed{\begin{aligned} \frac{d^2x}{dt^2} &= -\frac{1}{\rho}\frac{\partial p}{\partial x} + 2\Omega v \sin\phi - 2\Omega w \cos\phi + F_x \\ \frac{d^2y}{dt^2} &= -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\Omega u \sin\phi + F_y \\ \frac{d^2z}{dt^2} &= -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + 2\Omega u \cos\phi + F_z \end{aligned}}$$

which can be combined to give

$$\boxed{\frac{d\mathbf{V}}{dt} = -\frac{1}{\rho}\nabla p - 2\Omega \times \mathbf{V} + \mathbf{g} + \mathbf{F}}$$

where **V** is the vector wind.

Scale Analysis of the Equations of Motion

11. Objective*:* To simplify the equations by considering the magnitude (**scale**) of each of the terms and dropping those that are much smaller for a particular type of air motion under consideration.

Type of motion	Horizontal scale (m)
Molecular mean free path	10^{-7}
Minute turbulent eddies	10^{-2}-10^{-1}
Small eddies	10^{-1}-1
Dust devils	1 – 10
Gusts	10 - 10^{2}
Tornadoes	10^{2}
Cumulonimbus clouds	10^{3}
Fronts, squall lines	10^{4} - 10^{5}
Hurricanes	10^{5}
Synoptic cyclones	10^{6}
Planetary waves	10^{7}

Geostrophic Wind

12. For mid-latitude **synoptic-scale** weather systems, the following are the characteristic scales:

PARAMETER	SCALE
horizontal velocity *U*	10 m s^{-1}
vertical velocity *W*	1 cm s^{-1}
length *L*	10^6 m
depth *D*	10^4 m
time *L/U*	10^5 s
horizontal pressure fluctuation $\Delta P/r$	10^3 m^2 s^{-2}

13. Introduce another parameter - the **Coriolis parameter** *f* defined by

$$f = 2\Omega\sin\phi \cong 10^{-4}\ s^{-1} = f_o \qquad \text{for } \phi = 45°$$

Then, the two horizontal equations of motion have scales as follows:

		INDIVIDUAL TERMS		
x-component equation	$\frac{du}{dt}$	$-2\Omega v \sin\phi$	$2\Omega w \cos\phi$	$= -\frac{1}{\rho}\frac{\partial p}{\partial x}$
y-component equation	$\frac{dv}{dt}$	$+2\Omega u \sin\phi$		$= -\frac{1}{\rho}\frac{\partial p}{\partial y}$
Scale of terms	U^2/L	f_oU	f_oW	$\Delta P/\rho L$
Magnitude (m s^{-2})	10^{-4}	10^{-3}	10^{-6}	10^{-3}

Retaining only the largest terms,

$$\boxed{-fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} \quad \text{and} \quad fu = -\frac{1}{\rho}\frac{\partial p}{\partial y}}$$

This is known as the **geostrophic** relationship. It is a *diagnostic* relationship because no *d/dt* is involved, i.e. it cannot be used for prediction. Combining the two component equations gives

$$\boxed{\mathbf{V_g} = \mathbf{k} \times \frac{1}{f\rho}\nabla p}$$

where $\mathbf{V_g} = u_g\,\mathbf{i} + v_g\mathbf{j}$ is known as the **geostrophic wind**.

14. An atmosphere in which this equation is satisfied is said to be in **geostrophic balance**. Synoptic-scale weather systems in the mid-latitudes are in approximate geostrophic balance to within 10-15%. *But not so* in the tropics because $f \sim 10^{-5}$ so that the terms *d/dt* are of the same order as the Coriolis term. If the 10^{-4} terms are included, we have a set of *prognostic* equations:

$$\frac{du}{dt} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$

$$\frac{dv}{dt} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y}$$

15. To estimate the extent to which the atmosphere is in geostrophic balance, the ratio between the acceleration and the Coriolis term $(U^2/L)/(f_oU)$ can be calculated. This ratio is non-dimensional and called the **Rossby number**:

$$R_o = \frac{U}{f_o L}$$

The smaller the Rossby no., the closer is the atmosphere to geostrophic balance.

Hydrostatic Balance

16. Approximation for the vertical equation of motion

The scales of individual terms are:

	INDIVIDUAL TERMS			
z-component equation	$\frac{dw}{dt}$	$-2\Omega u \cos\phi$	$= -\frac{1}{\rho}\frac{\partial p}{\partial z}$	$-g$
Scale of terms	UW/L	f_oU	$P_o/\rho H$	g
Magnitude (m s^{-2})	10^{-7}	10^{-3}	10	10

P_o = surface pressure ~ 10^5 Pa, □ ~ 1 kg m^{-3} and H ~ 10 km

If only the largest terms are retained, then

$$-\frac{1}{\rho}\frac{\partial p}{\partial z} = g$$

which is simply the hydrostatic equation. Therefore, for these weather systems, the atmosphere is very close to hydrostatic balance.

A Mathematical Aside - Total Differentiation

17. Consider an instrument that measures the temperature at (x,y,z) to be T. If the instrument now moves with the wind which has a velocity $\mathbf{V} = u\mathbf{i} + v\mathbf{j} + w\mathbf{k}$, then after time □$t$ [during which the instrument is displaced (□x,□y,□z)], the temperature it measures will be T+□T.

Expanding the temperature change in a Taylor series,

$$\delta T = \left(\frac{\partial T}{\partial t}\right)\delta t + \left(\frac{\partial T}{\partial x}\right)\delta x + \left(\frac{\partial T}{\partial y}\right)\delta y + \left(\frac{\partial T}{\partial z}\right)\delta z + \left\{\begin{matrix}\text{higher}\\ \text{order}\\ \text{terms}\end{matrix}\right\}$$

Dividing through by □t and taking the limit □t □ 0,

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \frac{\partial T}{\partial x}\frac{dx}{dt} + \frac{\partial T}{\partial y}\frac{dy}{dt} + \frac{\partial T}{\partial z}\frac{dz}{dt}$$

where

$$\frac{dT}{dt} \equiv \lim_{\delta t \to 0} \frac{\delta T}{\delta t}$$

is the rate of change of *T* following the motion. Since

$$\frac{dx}{dt} \equiv u, \quad \frac{dy}{dt} \equiv v, \quad \frac{dz}{dt} \equiv w$$

Therefore,

$$\boxed{\frac{dT}{dt} = \frac{\partial T}{\partial t} + \left(u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} \right)}$$

or

$$\boxed{\frac{\partial T}{\partial t} = \frac{dT}{dt} - \mathbf{V} \cdot \nabla T}$$

The terms in this equation have the following interpretations:
dT/dt - **total** or **substantive** derivative, i.e. time rate of change following the motion
□*T*/□*t* - **local** derivative, i.e. time rate of change at a fixed point

-**V**·□T - temperature **advection**, the contribution to local temperature change due to air motion
+ve - *warm advection* (wind blowing from warm to cold)
-ve - *cold advection* (wind blowing from cold to warm)

Therefore, the local change of temperature is contributed by the rate of change of temperature following the motion (heating or cooling of the individual air parcels) and the advective rate of change of temperature.

18. Points to note:

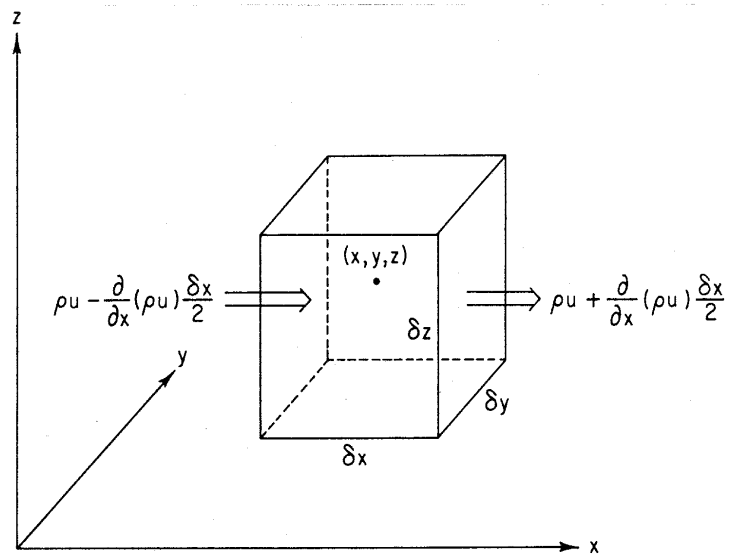
a. This relationship between the local and total changes applies to all variables.

b. If □/□t of a quantity is zero, the quantity is said to be in **steady state**.

c. *d/dt* does not have to follow the wind field. It can follow the motion of the object as well if the latter does not move with the wind.

<u>Continuity Equation</u>

19. Consider a volume element □*x*□*y*□*z* through which air passes.



Mass inflow into a fixed (eulerian) control volume, due to motion parallel to the *x* axis.

Rate of inflow of mass through the left-hand face (i.e. the *y-z* plane) per unit area is

$$\left[\rho u - \frac{\partial}{\partial x}(\rho u)\frac{\delta x}{2}\right]$$

while the rate of outflow per unit area through the right-hand face is

$$\left[\rho u + \frac{\partial}{\partial x}(\rho u)\frac{\delta x}{2}\right]$$

20. Since the area of each of these faces is □*y*□*z*, the net rate of flow into the volume due to the *x*-velocity component *u* is

$$\left[\rho u - \frac{\partial}{\partial x}(\rho u)\frac{\delta x}{2}\right]\delta y\,\delta z - \left[\rho u + \frac{\partial}{\partial x}(\rho u)\frac{\delta x}{2}\right]\delta y\,\delta z$$

$$= -\frac{\partial}{\partial x}(\rho u)\delta x\,\delta y\,\delta z$$

21. Similar expressions for the inflow in the *y* and *z* directions can be derived so that the net rate of mass inflow into the volume is

$$-\left[\frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v) + \frac{\partial}{\partial z}(\rho w)\right]\delta x\,\delta y\,\delta z$$

The rate of increase of mass per unit volume = □□/□*t*. By conservation of mass, these two terms must equal so that

$$\boxed{\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \mathbf{V}) = 0}$$

This is the *mass divergence* form of the **continuity equation** which relates the local rate of change of density with the divergence of the mass. Since

$$\nabla \cdot \left(\rho \mathbf{V}\right) = \rho \nabla \cdot \mathbf{V} + \mathbf{V} \cdot \nabla \rho$$

and

$$\frac{d}{dt} = \frac{\partial}{\partial t} + \mathbf{V} \cdot \nabla$$

an alternate form of the continuity equation is

$$\boxed{\frac{1}{\rho}\frac{d\rho}{dt} + \nabla \cdot \mathbf{V} = 0}$$

This is the *velocity divergence* form of the continuity equation which relates the fractional rate of change of density *following the motion of an air parcel* with the velocity divergence.

<u>Scale Analysis of the Continuity Equation</u>

22. The continuity equation can be written as

$$\underbrace{\frac{1}{\rho_o}\left(\frac{\partial \rho'}{\partial t} + \mathbf{V}_H \cdot \nabla \rho'\right)}_{\mathrm{A}} + \underbrace{\frac{w}{\rho_o}\frac{d\rho_o}{dz}}_{\mathrm{B}} + \underbrace{\nabla \cdot \mathbf{V}}_{\mathrm{C}} \cong 0$$

where □'= local deviation of density from its horizontally averaged value, $□_0$ (z).

a. **<u>Term A:</u>**
For synoptic scale motions, $□'/□_0 \sim 10^{-2}$ so that

$$\frac{1}{\rho_o}\left(\frac{\partial \rho'}{\partial t} + \mathbf{V}_H \cdot \nabla \rho'\right) \sim \frac{\rho'}{\rho_o}\frac{U}{L} \sim 10^{-7}\mathrm{s}^{-1}$$

b. **<u>Term B</u>**:

For motions in which the depth scale *D* is comparable to the density scale height *H*, $d(\ln □_0)/dz \sim D^{-1}$ so that

$$\frac{w}{\rho_o}\frac{d\rho_o}{dz} \sim \frac{W}{D} \sim 10^{-6}\ \mathrm{s}^{-1}$$

c. **<u>Term C</u>**:

$$\nabla \cdot \mathbf{V} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}$$

23. For synoptic-scale motions, the terms □*u*/□*x* and □□/□*y* tend to be of equal magnitude but opposite sign so that

$$\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \sim 10^{-1}\,\frac{U}{L} \sim 10^{-6}\ \mathrm{s}^{-1}$$

and

$$\frac{\partial w}{\partial z} \sim \frac{W}{D} \sim 10^{-6}\ \mathrm{s}^{-1}$$

Thus, terms *B* and *C* are each an order of magnitude greater than term *A* so that to a first approximation, terms B and C balance:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} + w\,\frac{d\,\ln \rho_0}{dz} \cong 0$$

which can be written in vector form as

$$\nabla \cdot \left(\rho_o \mathbf{V}\right) = 0$$

This equation says that for synoptic-scale motion, the *mass flux computed using the basic state density* ρ_o *is nondivergent.* In particular, if ρ_o constant (as in pure horizontal motion), the equation reduces to

$$\nabla \cdot \mathbf{V} = 0 \qquad \text{or} \qquad \frac{d\rho}{dt} = 0$$

which means that the fluid is **incompressible**.

<u>Thermodynamic Energy Equation</u>

24. 1st law: $dh = c_p\ dT - \alpha\ dp$

Divide through by *dt*,

$$\boxed{c_p \frac{dT}{dt} - \alpha \frac{dp}{dt} = \dot{Q}}$$

where $\dot{Q}$ is the heating rate or **diabatic forcing**.

25. **Summary of the Set of Equations Governing the Motion of Air**

NAME	EQUATION
Momentum	$\frac{d\mathbf{V}}{dt} = -\frac{1}{\rho}\nabla p - 2\Omega \times \mathbf{V} + \mathbf{g} + \mathbf{F}$
Continuity	$\frac{1}{\rho}\frac{d\rho}{dt} + \nabla \cdot \mathbf{V} = 0$
State	$p\alpha = RT^*$
Thermodynamic Energy	$c_p \frac{dT}{dt} - \alpha \frac{dp}{dt} = \dot{Q}$

Independent variables: *x, y, z, t*

Dependent variables: **V**(*u, v, w*), *p*, T^*, ρ

Pressure as Vertical Coordinate

26. In meteorology, pressure is often used as a vertical coordinate because the equations can be simplified. To do this, we have to transform

$$(\square/\square x)_z \text{ and } (\square/\square y)_z \;\square\; (\square/\square x)_p \text{ and } (\square/\square y)_p$$

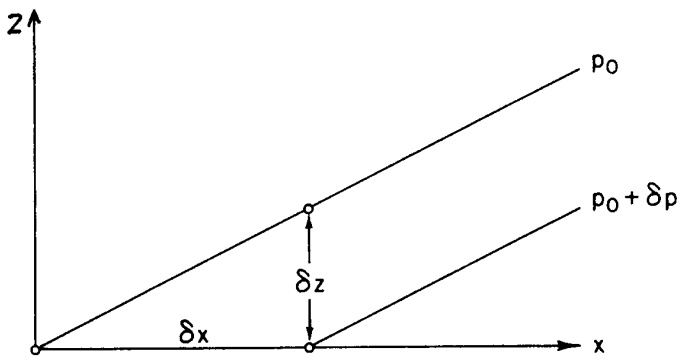


Fig. 1.11 Slope of pressure surfaces in the *x*,*z* plane.

$$\left[\frac{(p_o + \delta p) - p_o}{\delta x}\right]_z = \left[\frac{(p_o + \delta p) - p_o}{\delta z}\right]_x \left(\frac{\delta z}{\delta x}\right)_p$$

where the subscripts indicate variables which remain constant in evaluating the differentials. Then, in the limit $\square z \;\square\; 0$,

$$\left[\frac{(p_o + \delta p) - p_o}{\delta z}\right]_x \rightarrow \left(-\frac{\partial p}{\partial z}\right)_x$$

where the minus sign is included because □z < 0 for □p > 0. Taking the limits □x, □z □ 0,

$$\left(\frac{\partial p}{\partial x}\right)_z = -\left(\frac{\partial p}{\partial z}\right)_x \left(\frac{\partial z}{\partial x}\right)_p$$

Substitution from the hydrostatic equation then gives

$$-\frac{1}{\rho}\left(\frac{\partial p}{\partial x}\right)_z = -g\left(\frac{\partial z}{\partial x}\right)_p = -\left(\frac{\partial \Phi}{\partial x}\right)_p$$

Therefore, in the *isobaric* coordinate system (*x, y, p*), the density becomes implicit in the equations and the horizontal pressure gradient force is measured by the gradient of geopotential on isobaric surfaces.

<u>Equations in Isobaric Coordinates</u>

27. The horizontal momentum equations recall the approximate prognostic equations

$$\frac{du}{dt} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$

$$\frac{dv}{dt} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y}$$

Transformation from *z* to *p* gives, in vector form

$$\boxed{\frac{d\mathbf{V}_H}{dt} + f\,\mathbf{k} \times \mathbf{V}_H = -\nabla_p \Phi}$$

where $\mathbf{V}_H = u\mathbf{i} + v\mathbf{j}$ now is the *horizontal* wind vector and □$_p$ is the horizontal gradient operator holding *p* constant.

28. Note that *d/dt* is now defined as

$$\frac{d}{dt} \equiv \frac{\partial}{\partial t} + \frac{dx}{dt}\frac{\partial}{\partial x} + \frac{dy}{dt}\frac{\partial}{\partial y} + \frac{dp}{dt}\frac{\partial}{\partial p}$$

$$= \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y} + \omega\frac{\partial}{\partial p}$$

where □ = *dp/dt* is the "vertical motion" in isobaric coordinates, i.e. pressure change following the motion.

The geostrophic wind $\mathbf{V}_g$ now becomes

$$f\mathbf{V}_g = \mathbf{k} \times \nabla_p \Phi$$

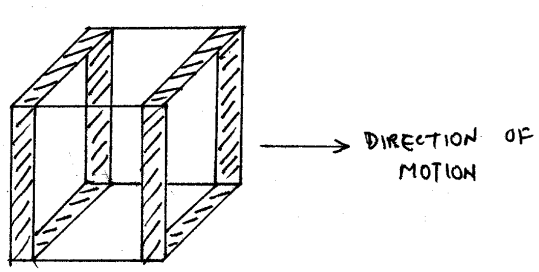
Taking the divergence of the equation,

$$\nabla_p \cdot \left(f\mathbf{V}_g \right) = \nabla_p \cdot \left(\mathbf{k} \times \nabla_p \Phi \right) = 0$$

Therefore, if f = constant, the divergence of the geostrophic wind at constant p is zero.

<u>Continuity Equation</u>

29. Consider a volume element of fixed mass □M moving with the fluid.



Volume of the fluid = □x□y□z = □x□y(-□p/□g)

$$\therefore \quad \delta M = -\frac{\delta x \delta y \delta p}{g}$$

30. Since the mass of the fluid is conserved following the motion,

$$\frac{1}{\delta M}\frac{d}{dt}\delta M = \frac{g}{\delta x \delta y \delta p}\frac{d}{dt}\left(\frac{\delta x \delta y \delta p}{g}\right) = 0$$

After differentiating using the chain rule, and changing the order of the differential operators, we obtain

$$\frac{\delta u}{\delta x} + \frac{\delta v}{\delta y} + \frac{\delta \omega}{\delta p} = 0$$

31. Taking the limit □x, □y, □p □ 0 gives the continuity equation in the isobaric system:

$$\boxed{\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)_p + \frac{\partial \omega}{\partial p} = 0}$$

This form of the continuity equation contains no reference to the density field, and does not involve time derivatives. It is therefore a *diagnostic* rather than a prognostic equation.

32. The thermodynamic energy equation Recall

$$c_p \frac{dT}{dt} - \alpha \frac{dp}{dt} = \dot{Q}$$

Expanding *dT/dt*,

$$c_p \left(\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + \omega \frac{\partial T}{\partial p} \right) - \alpha\omega = \dot{Q}$$

From the Poisson Equation, the dry-adiabatic lapse rate in isobaric coordinates is

$$\Gamma_d = \left(\frac{\partial T}{\partial p} \right)_\theta = \frac{\alpha}{c_p}$$

so that

$$\boxed{\frac{\partial T}{\partial t} + \mathbf{V}_H \cdot \nabla T + \left(\Gamma - \Gamma_d \right) \omega = \frac{\dot{Q}}{c_p}}$$

<u>Summary</u>

NAME	EQUATION
Horizontal momentum	$\frac{d\mathbf{V}_H}{dt} + f\,\mathbf{k} \times \mathbf{V}_H = -\nabla_p \Phi$
Continuity	$\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)_p + \frac{\partial \omega}{\partial p} = 0$
Hydrostatic	$\frac{\partial \Phi}{\partial p} = -\alpha$
State	$p\alpha = RT$
Thermodynamic energy	$\frac{\partial T}{\partial t} + \mathbf{V}_H \cdot \nabla T + \left(\Gamma - \Gamma_d \right) \omega = \frac{\dot{Q}}{c_p}$

Independent variables: *x, y, p, t* & Dependent variables: $\mathbf{V}_H(u, v)$, □, □, *T*, □

TASK-6 : HORIZONTAL FRICTIONLESS BALANCED FLOW

Objectives

After reading this chapter, you should be able to:-

- Understand the Equation of motion in natural co-ordinate system.
- Understand the classification of flow patterns and different solutions of gradient wind equation.
- Appreciate the relation between gradient and geostrophic winds and its application.
- Understand the concept of ageostrophic wind.
- Understand the representation of horizontal flow in spherical polar coordinate system.

Introduction

1. Observations reveal that every large-scale flow in the atmosphere is nearly horizontal. Appreciable vertical motion is usually confined to local convective currents, to boundary regions separating different air masses, or to mountainous regions where flow is influenced by topography. Regions of ascending air flow are often marked by the formation of clouds and precipitation due to lifting. On the other hand, regions of large-scale descending motion are frequently associated to widespread fog.

2. Although hydrostatic equilibrium is very nearly fulfilled in the atmosphere, vertical velocities do exist and play important role in resulting weather phenomena. Many important properties of winds can be deduced by considering horizontal flow alone. This chapter deals with the horizontal flow only, which is balanced, i.e., the sum total of all the forces acting on a particle is zero and there is no net acceleration. The assumption of frictionless flow essentially implies that the motion under study is at some higher level, at which, the effect of friction due to the uneven surface of the earth is negligible.

3. Despite the apparent complexity of atmospheric motion systems, the pressure (or geopotential height) and winds are closely related by rather simple approximate force balances. In order to gain more qualitative understanding of horizontal balanced atmospheric flow, it is useful to describe the wind flow by expanding the horizontal momentum equations into its components in Natural Coordinate system.

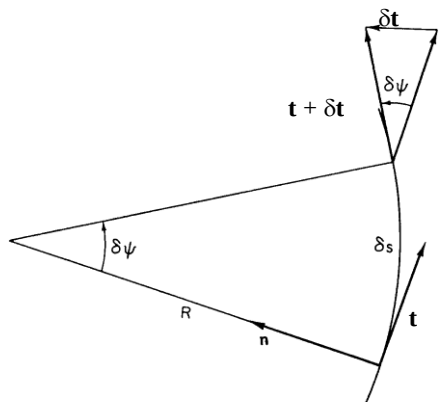
Equations of Motion in Natural Coordinate System

4. The natural coordinate system is defined by the orthogonal set of unit vectors $\hat{t}, \hat{n}, \hat{k}$, in which , $\hat{t}$ is oriented parallel to horizontal velocity at each point, $\hat{n}$ is normal to horizontal velocity in such a way that it is positive to the left perpendicular of the flow and $\hat{k}$ is directed

vertically upward (i.e., towards the local zenith). These unit vectors are perpendicular to each other and have the same sense as $\hat{i}, \hat{j}, \hat{k}$ in the rectangular (Cartesian) coordinate system. The advantage of using this coordinate system is that the flow is always considered in the direction of $\hat{t}$ and hence can never by negative (i.e., the flow will always be real and positive).

5. In such a system, the horizontal velocity would have only one component, i.e., tangential component. So, $\vec{V} = V\hat{t}$, where ***V*** is the horizontal wind speed and is equal to ds/dt , where ***s(x,y,t)*** is the curve followed by a parcel moving in horizontal plane. Thus, the acceleration can be found as:-

$$\vec{V} = V\hat{t} \Rightarrow \frac{d\vec{V}}{dt} = \hat{t}\frac{dV}{dt} + V\frac{d\hat{t}}{dt}$$



6. The rate of change of unit vector $\hat{t}$ following the motion (i.e., total rate of change) can be derived from the figure given above, as:-

$$\frac{d\hat{t}}{dt} = \frac{d\hat{t}}{ds}\frac{ds}{dt} = V\frac{d\hat{t}}{ds}$$

and
$$\frac{d\hat{t}}{ds} = \left|\frac{d\hat{t}}{ds}\right|\hat{n} = \frac{1}{R}\hat{n}$$

Where R is the radius of curvature following the motion (R=1/K, where K is the Curvature). By convention, R is taken to be positive, when the centre of curvature is in the +ve $\hat{n}$ direction. Thus, R is +ve for cyclonic (anti-clockwise) flow and –ve for anticyclonic (clockwise) flow. So, the acceleration would be given as:

$$\frac{d\vec{V}}{dt} = \frac{dV}{dt}\hat{t} + \frac{V^2}{R}\hat{n} \qquad \text{…….. (1.1)}$$

7. Therefore, the acceleration following the motion is the sum of two terms on the right hand side of this equation, which are called as the ‘Tangential Acceleration’ (in the tangential direction) and the ‘Centripetal Acceleration’ due to the curvature (in the normal direction).

8. If the flow is assumed to be horizontal and frictionless, then Pressure Gradient Force and the Coriolis Force will be the only forces acting on the parcel. Hence, the equation of motion in natural coordinate system would be as follows:

$$\frac{d\vec{V}}{dt} = -\vec{\nabla}_p \phi - f\hat{k} \times \vec{V}$$

and in component form as:

$$\frac{dV}{dt}\hat{t} + \frac{V^2}{R}\hat{n} = -\left(\frac{\partial \phi}{\partial s}\hat{t} + \frac{\partial \phi}{\partial n}\hat{n}\right) - f\hat{k} \times \left(V\hat{t}\right)$$

or

$$\frac{dV}{dt}\hat{t} + \frac{V^2}{R}\hat{n} = -\left(\frac{\partial \phi}{\partial s}\hat{t} + \frac{\partial \phi}{\partial n}\hat{n}\right) - fV\hat{n}$$

9. By comparing terms of both the sides in same directions, the equations of motion in component form can be obtained as:

$$\frac{dV}{dt} = -\frac{\partial \phi}{\partial s} \Rightarrow \frac{dV}{dt} + \frac{\partial \phi}{\partial s} = 0 \qquad \text{..... (1.2a)}$$

and

$$\frac{V^2}{R} = -\frac{\partial \phi}{\partial n} - fV \Rightarrow \frac{V^2}{R} + fV + \frac{\partial \phi}{\partial n} = 0 \qquad \text{..... (1.2b)}$$

10. From the Eqn 1.2(a), it is stated that the tangential acceleration is equal to the pressure (or contour) gradient in the tangential direction. If the flow is considered as balanced (i.e., non-accelerated), then the tangential acceleration would also be zero. In such case:-

$$\frac{dV}{dt} = -\frac{\partial \phi}{\partial s} = 0 \text{, which means:- } \frac{\partial \phi}{\partial s} = 0 \text{, or } \phi = Const$$

Which states that the horizontal frictionless balanced flow would be directed along the contour lines (or isobars). In other words, such a motion would flow parallel to the isobars or contour lines.

11. From the Eqn 1.2(b), it is evident that the centripetal acceleration is equal to the sum of the normal component of Pressure Gradient Force and Coriolis force, which always acts to the right perpendicular to the motion in northern hemisphere. In case of perfectly balanced flow, the centripetal acceleration would also be zero. So:

$$\frac{V^2}{R} = 0 = fV + \frac{\partial \phi}{\partial n} \Rightarrow V = -\frac{1}{f}\frac{\partial \phi}{\partial n} \qquad \text{..... (1.3)}$$

which shows that the horizontal, frictionless, non-accelerated and straight line flow would be such that the motion will be parallel to the contours (or isobars), keeping low values of

geopotential height (or pressure) to its left (refer Geostrophic Winds in the previous chapter also).

Classification of Flow Patterns

12. **Gradient Wind Equation**. The Eqn 1.2(b) is also known as the Gradient Wind Equation, which consists of three terms as:-

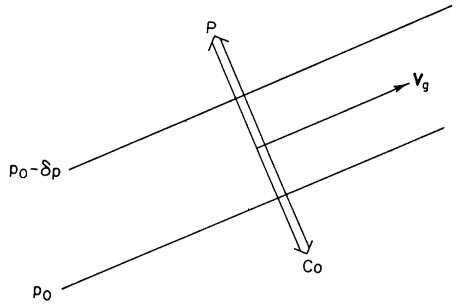
$$\frac{V^2}{R} + fV + \frac{\partial \phi}{\partial n} = 0 \quad \text{or} \quad V^2 + fRV + R\frac{\partial \phi}{\partial n} = 0$$

The first term is the Centripetal Acceleration, second one is the Coriolis Acceleration and the third one is the Pressure Gradient Force. Thus, this equation represents such a flow, which is frictionless and horizontal. Further, the tangential acceleration would be zero, as the contour value does not change in tangential direction (i.e., in the direction of the flow) (Eqn 1.2a). Therefore, the flow would be completely non-accelerated also. Such flow is known as '**Gradient Wind Flow**', which is balanced by the Pressure Gradient Force, Coriolis Force and Centrifugal (or Centripetal) Force.

13. **Geostrophic Flow**. From the definition of the Geostrophic wind flow, it is the horizontal, frictionless, non-accelerated and straight line flow. In case of a straight line flow, R becomes infinity (as K=0). So the centripetal acceleration vanishes and the flow is balanced by only the Pressure Gradient Force and Coriolis Force. The expression of such flow can be obtained from Eqn 1.2(b) as:

$$\frac{V_g^{\,2}}{R} = 0 = fV_g + \frac{\partial \phi}{\partial n} \Rightarrow V_g = -\frac{1}{f}\frac{\partial \phi}{\partial n} \qquad \text{..... (1.4)}$$

Which states that the Geostrophic winds flow parallel to the contours (or isobars), keeping low values to its left (as shown in the figure below).



14. **Gradient Flow.** As discussed earlier, the Gradient Wind is defined as the horizontal, frictionless, non-accelerated and curved flow, which is balanced by the Pressure Gradient Force, Coriolis Force and Centrifugal (or Centripetal) Force. The expression of such flow is given by the Gradient Wind Equation (Eqn 1.2b). The solution of this equation would be able to given the Gradient Wind, as this equation is the second order differential equation. The roots of the Gradient Wind can be obtained as:-

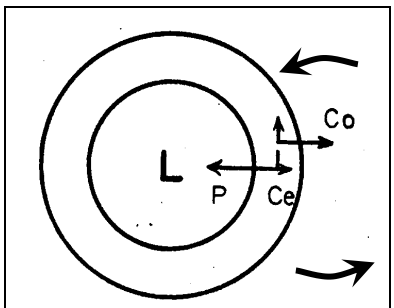
$$V = -\frac{fR}{2} \pm \sqrt{\left(\frac{fR}{2}\right)^2 - R\frac{\partial\phi}{\partial n}} = -\frac{fR}{2} \pm \sqrt{\frac{f^2R^2}{4} - R\frac{\partial\phi}{\partial n}} \quad \text{..... (1.5)}$$

This can be solved to find out various possible solutions. The most important thing to remember is that the wind speed has to remain +ve and real. From the close inspection, it is evident that total 08 combinations are possible, out of which, only 04 cases would emerge when wind speed would be +ve and real. Various solutions are given below:-

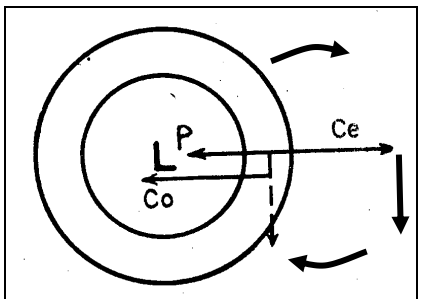
a. **<u>Cyclonic Flow (R > 0)</u>.**

(1) $\partial\phi/\partial n > 0$. Both the +ve and –ve roots would be unphysical, as it would produce –ve and imaginary solutions.

(2) $\partial\phi/\partial n < 0$. The –ve root would be unphysical, as it would produce –ve solution. However, the +ve root would produce +ve and real solution. Such a flow is known as '**Regular Low**' (refer the left figure given below). Examples of such flow patterns are seen in Monsoon Lows / Depressions, Heat Low etc.



Cyclonic flow around a low.

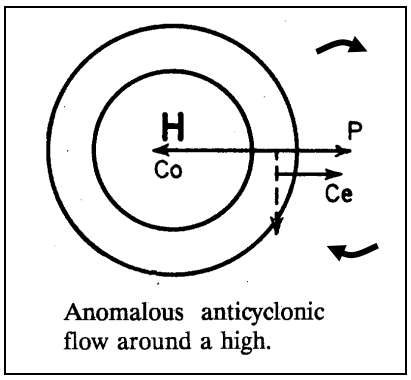


Anomalous anticyclonic flow around a low.

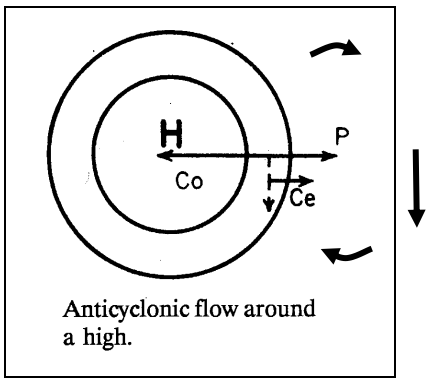
b. **<u>Anticyclonic Flow (R < 0)</u>.**

(1) $\partial\phi/\partial n > 0$. The –ve root would be unphysical, as it would produce –ve solution. However, the +ve root would be possible, but it would be anti-baric flow (i.e., anticyclonic flow around a central low pressure, which does not obey the basic law of keeping low pressure values to the left). Such a flow is known as '**Anomalous Low**' (refer the right figure given above). Examples of such flow patterns are seen in Tornadoes, top of Tropical Cyclones etc.

(2) $\partial\phi/\partial n < 0$. In this case, both the –ve and +ve roots would be possible and would produce +ve and real solutions. The description is as follows:



Anomalous anticyclonic flow around a high.



Anticyclonic flow around a high.

(a) <u>+ve Root</u>. Such flow is known as 'Anomalous High' (refer the left figure given above). In such a flow, the wind speed is not limited or restricted.

(b) <u>-ve Root</u>. Such flow is known as 'Regular High' (refer the right figure given above). In such a flow, the wind speed is limited or restricted by a certain condition (discussed in next paragraph).

15. **<u>Gradient Wind Solution for Anticyclonic Flow</u>.** In the northern hemisphere, R is always –ve for anticyclonic flow. As discussed earlier, the anticyclonic flow around a regular high pressure area requires the condition as $\partial\phi/\partial n < 0$, i.e., pressure is increasing towards the centre of the curvature. Now, from the gradient wind equation:

$$V = -\frac{fR}{2} \pm \sqrt{\left(\frac{fR}{2}\right)^2 - R\frac{\partial\phi}{\partial n}}$$

Where the first term is +ve (as R<0) and the term under the square-root will be +ve only if the following condition is satisfied:-

$$\left[\left(\frac{fR}{2}\right)^2 - R\frac{\partial\phi}{\partial n}\right] > 0 \Rightarrow \left(\frac{fR}{2}\right)^2 > \left|R\frac{\partial\phi}{\partial n}\right| \Rightarrow \frac{f^2R}{4} > \left|\frac{\partial\phi}{\partial n}\right|$$

and it would yield +ve and real wind speed (i.e., the solution) only for the –ve root, as the +ve root would yield the Anomalous High. If the above mentioned condition is not met (i.e., $\frac{f^2R}{4} < \left|\frac{\partial\phi}{\partial n}\right|$), then the wind speed would be imaginary. Therefore, this condition is the limiting condition for a regular anticyclonic flow to exist. The following cases can be discussed:-

a. If $\left|\partial\phi/\partial n\right| = 0$, then the Gradient wind would be zero.

b. If $\left|\partial\phi/\partial n\right| > f^2R/4$, then the Gradient wind would be imaginary.

c. If $\left|\partial\phi/\partial n\right| < f^2R/4$, then the Gradient wind would be real and +ve.

d. If $\left|\partial\phi/\partial n\right| = f^2R/4$, then the Gradient wind would be equal to $\left(-fR/2\right)$, which will be the maximum value of the wind speed possible in association with the anticyclonic flow.

16. **<u>Conclusions for Gradient Winds in case of Anticyclonic Flow</u>**. Base on the discussion on the solutions for Gradient winds in case of a regular anticyclonic flow, it can be concluded that the wind speed is liable to become imaginary, if Pressure Gradient becomes too large, Radius of Curvature becomes too small or Coriolis parameter becomes too small. Thus, if any one or any combination of these three conditions occur, gradient wind flow cannot exist. Physically, it means that:

a. It is not possible to have regular anticyclones at or near the equator, where the Coriolis parameter becomes too small.

b. It is not possible to have intense centres of anticyclones or high pressure areas with very small radius of curvature, except perhaps for short period of time.

c. Anticyclonic flow cannot sustain with very large pressure gradient. This means that strong pressure gradient and stronger winds are not possible close to the centre of the anticyclone. In practice, calm winds are the characteristics of the centre of an anticyclone.

d. In cases, when the above mentioned conditions are not satisfied for an existing anticyclone, it would dissipate rapidly.

17. **<u>Gradient Wind Solution for Cyclonic Flow</u>.** In the northern hemisphere, R is always +ve for cyclonic flow. Also, the cyclonic flow around a regular low pressure area requires the condition as $\partial\phi/\partial n < 0$, i.e., pressure is decreasing towards the centre of the curvature. Now, from the gradient wind equation:

$$V = -\frac{fR}{2} \pm \sqrt{\left(\frac{fR}{2}\right)^2 - R\frac{\partial\phi}{\partial n}}$$

Where the first term is –ve (as R>0) and the term under the square-root will always be +ve, as $R\,\partial\phi/\partial n < 0$. In that case, the +ve root would always yield a real and +ve wind speed and there is no limiting conditions for the maximum sustained wind speed in association with a regular cyclonic flow. Therefore, in theory, there is no upper limit for the wind speed and no lower limit for pressure gradient in case of a regular cyclone. However, the lower limit with it is zero, when $\partial\phi/\partial n = 0$. It is an observational fact also that more intense cyclonic circulations (e.g., associated with Tropical Storms etc.) are having vary tight radius of curvature. Further, in the real atmosphere, Cyclonic systems are generally much smaller in dimension than the

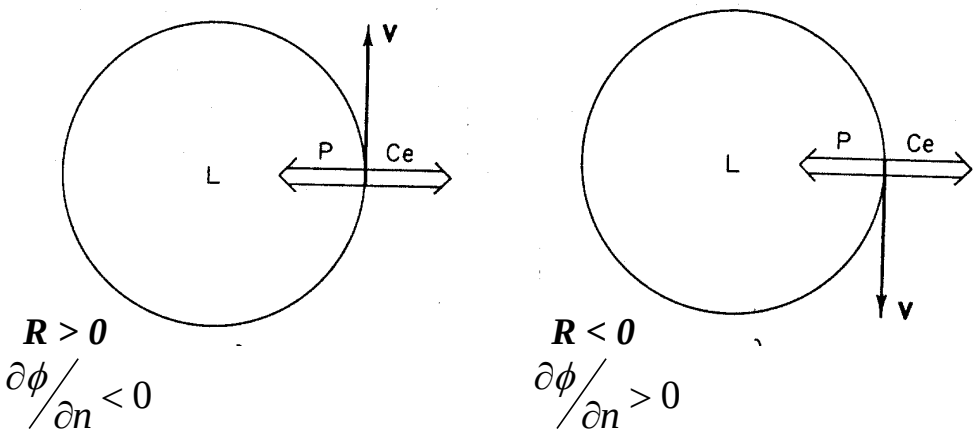
Anticyclonic systems. For example, Tropical Cyclones are the biggest cases of cyclonic flow in dimension, which are still much smaller in dimension compared to Sub-tropical Anticyclones.

18. **Cyclostrophic Flow**. If the horizontal scale of a disturbance is small enough, the variation of Coriolis force may be neglected. Further, if the Coriolis force is small compared with the Pressure Gradient force and the Centrifugal force, then also it can be neglected. Such a flow will be balanced by the Pressure Gradient force and Centrifugal force only. Such a flow is called as '**Cyclostrophic Flow**'. In such case, from the Gradient wind equation, we will get:-

$$\frac{V_c^2}{R} + \frac{\partial \phi}{\partial n} = 0 \Rightarrow V_c = \sqrt{-R\frac{\partial \phi}{\partial n}} \qquad \text{..... (1.6)}$$

In order to obtain +ve and real value of $\boldsymbol{V_c}$, it is necessary that $\left(-R\,{\partial\phi}/{\partial n}\right) > 0$, i.e., +ve. It can be either cyclonic or anticyclonic. Hence, the following types of Cyclostrophic flow are possible:-

a. **Cyclonic Case (R>0).** If R>0, then $\left({\partial\phi}/{\partial n}\right) < 0$, i.e., it is the case of a regular Low pressure area (as shown in the left figure given below).



b. **Anticyclonic Case (R<0).** If R<0, then $\left({\partial\phi}/{\partial n}\right) > 0$, i.e., it is the case of an Anomalous Low pressure area (as shown in the right figure given above). From the above discussion, it is clear that Cyclostrophic flow is not possible around a regular or Anomalous High pressure area. The Cyclostrophic balance is approximately valid if the ratio of the Centrifugal force to the Coriolis force is large. Such a ratio is just the Rossby number (***R_o = V/f R***). This happens if ***V*** is large (strong wind) or ***R*** is small (scale of disturbance small).

19. **Inertial Flow**. If the pressure field is horizontally uniform, or geopotential field is uniform on isobaric surface, so the horizontal pressure gradient vanishes. In such a case, the flow will be balanced by the Coriolis force and Centrifugal force only. Because these forces are all due to inertia of the fluid, such a flow is called as '**Inertial Flow**'. In such case, from the Gradient wind equation, we will get:-

$$\frac{V_i^2}{R} + fV_i = 0 \Rightarrow V_i = -fR \qquad \text{..... (1.7)}$$

In order to obtain +ve and real value of V_i, it is necessary that ***R*** is always –ve, which means that Inertial flow cannot be cyclonic in northern hemisphere. The radius of curvature ***R*** for such a flow is $R = -\frac{V_i}{f}$. Further, a uniform horizontal pressure field means that ***V*** is constant. Therefore, if ***f*** = constant, ***R*** = constant, which means that ***R*** is inversely proportional to ***f***. Thus, if changes in ***f*** are ignored, ***R*** will also be constant. But, ***f*** decreases equator-wards and increases pole-wards. So, ***R*** increases (decreases), as the particle moves equator-wards (pole-wards). Thus, air parcels follow circular *clockwise* paths (**anticyclonic** for Northern Hemisphere and **cyclonic** for the Southern Hemisphere). This type of motion is also referred to as an **'Inertial Oscillation'**. When the variation of ***f*** is slight and inertial path is nearly circular, the period of Inertial Oscillation will be given as:-

$$T_i = \frac{2\pi R_i}{V} = \frac{2\pi}{f} = \frac{2\pi}{2\Omega Sin\phi} = \frac{\frac{1}{2}Day}{Sin\phi} \qquad \text{..... (1.8)}$$

Such situations are rare in atmosphere, however, in oceans, currents are often generated by transient winds. Inertial motion is important mostly in oceanography where pressure gradients are weak. In the atmosphere, such an oscillation is important in some cases in which the pressure gradient is small. Cross Equatorial Flow during SW Monsoon is one example of the Inertial Flow.

20. **<u>Relation between Gradient and Geostrophic Winds</u>.** The expressions for Gradient and Geostrophic winds are given by Eqn 1.2(b) and Eqn 1.4 respectively. From Eqn 1.4:

$$V_g = -\frac{1}{f}\frac{\partial\phi}{\partial n} \Rightarrow \frac{\partial\phi}{\partial n} = -fV_g$$

By substituting this value in Eqn 1.2(b), we will get the required relation as:-

$$V^2 + fRV + R\frac{\partial\phi}{\partial n} = 0 \Rightarrow V^2 + fRV - fRV_g = 0$$

or

$$\frac{V_g}{V} = \frac{V}{fR} + 1 \Rightarrow V_g = V\left(1 + \frac{V}{fR}\right) \qquad \text{..... (1.9)}$$

21. **<u>Applications of Relation between Gradient and Geostrophic Winds</u>**. From Eqn 2.9, following applications can be deduced:-

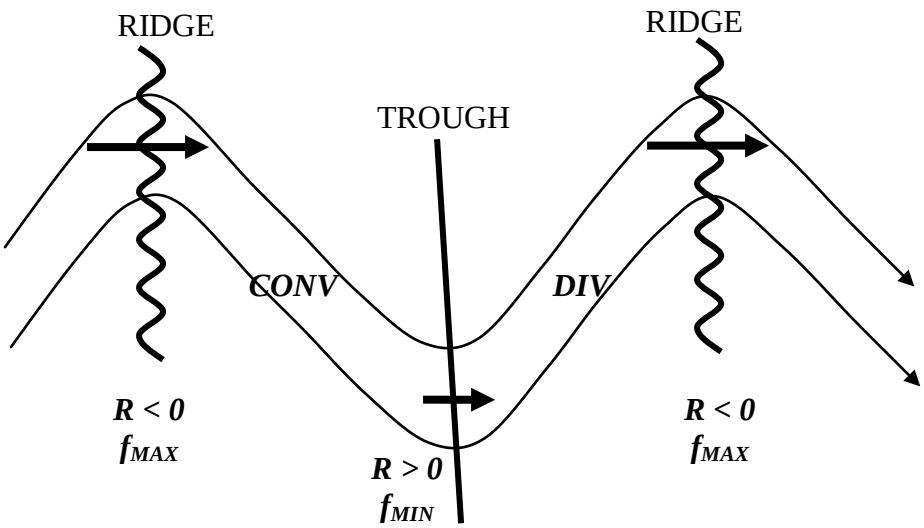
a. **<u>Cyclonic Flow</u>**. For a Cyclonic flow, R>0; so the Geostrophic wind will be greater than the Gradient wind.

b. **<u>Anticyclonic Flow</u>.** For an Anticyclonic flow, R<0; so the Geostrophic wind will be smaller than the Gradient wind.

c. Thus, it can be said that the Geostrophic Wind is an overestimate of Gradient Wind in a region of cyclonic curvature and an underestimate in a region of anticyclonic curvature. In other words, the Gradient winds will be Sub-Geostrophic for cyclonic curvature and Super-Geostrophic for anticyclonic curvature.

d. The relation also shows that the anti-baric Anomalous flow can exist only when $\left(V/fR\right) < (-1)$. Thus, such a flow is associated with small-scale intense vortices, e.g., Tornadoes etc.

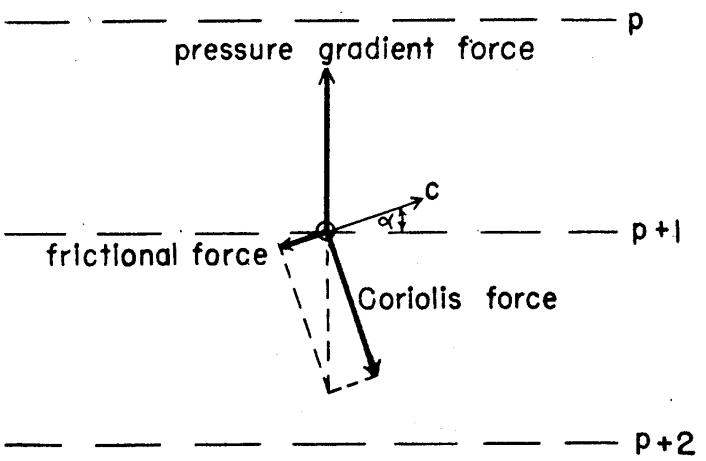
e. **Distribution of Divergence in a W'ly Wave**. In case of a W'ly wave, Gradient wind will be less than Geostrophic wind (i.e., sub-geostrophic) at Trough (as R>0) and more than Geostrophic wind (i.e., super-geostrophic) at Ridge (as R<0). Hence, as a result of variation of Gradient wind speed along the flow, Divergence will occur in the region from Trough to down-stream Ridge and Convergence from Ridge to down-stream Trough (as shown in the figure below). But, the Geostrophic wind speed is inversely proportional to the Coriolis parameter (***f***), hence it will be increasing from Ridge to down-stream Trough (as ***f*** decreases) and decreasing from Trough to down-stream Ridge (as ***f*** increases). Thus, both the effects will oppose each other. Therefore, it can be concluded that the Curvature effect would be dominant in case of Short waves and Latitude effect in case of Long waves.



Deviation from Geostrophy

22. In the real atmosphere, the winds do not always follow the Geostrophic Balance, especially in the lower latitudes. Friction also plays an important role in altering the winds from theoretical concepts of Geostrophy. The variation of Pressure Gradient can also alter the balance of forces. The deviation of Actual wind from the Geostrophic wind is known as '**Ageostrophic Wind**' or '**Geostrophic Departure**', which would be covered in details subsequently. Here, only the effect of Friction and variation of Pressure Gradient would be considered to study the deviation from Geostrophy.

23. **Effect of Friction**. If friction is present and acts opposite to the direction of motion, then the pressure gradient force must balance the sum of the frictional and the Coriolis forces (as shown in the figure given below).



Balance among pressure-gradient, Coriolis, and frictional forces

Therefore, "***The air moves at an angle to the isobars towards low pressure***". Assuming that the frictional force is proportional to the speed (with ***k*** being the proportionality constant) and the pressure gradient only exists in the *y*-direction, the equations of motion will become:

$$0 = fv - ku$$

$$0 = -fu - kv - \frac{\partial \phi}{\partial y}$$

By elimination we obtain:-

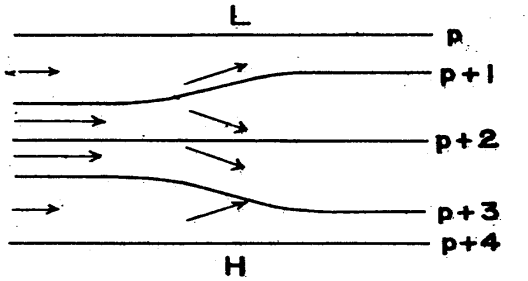
$$u = -\frac{f}{f^2 + k^2}\left(\frac{\partial \phi}{\partial y}\right) \quad and \quad v = -\frac{k}{f^2 + k^2}\left(\frac{\partial \phi}{\partial y}\right)$$

When ***k = 0*** (no friction), the flow is east-west. For ***k > 0*** (presence of friction), a north-south flow towards low pressure is introduced. By combining the two component equations, we will get:

$$c = \sqrt{\left(u^2 + v^2\right)} = \frac{1}{\sqrt{f^2 + k^2}}\left(\frac{\partial \phi}{\partial y}\right)$$

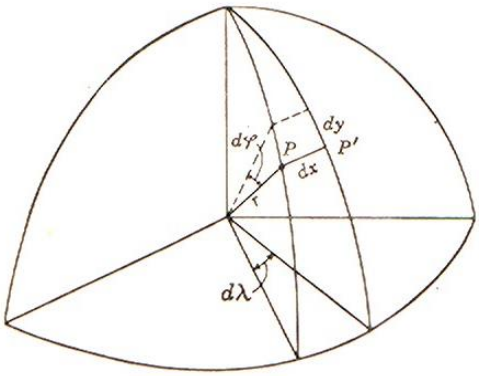
Where ***c*** is the resultant speed. Therefore, friction reduces the wind to below the geostrophic value.

24. **Effect of varying Pressure Gradient**. A change in the pressure gradient downstream will cause a cross-isobaric flow. For the pressure pattern, the actual wind will be cross-isobaric (as shown in the figure given below). An air parcel with initially high momentum moving into a region of more widely separated isobars will find the Coriolis force overbalancing the Pressure Gradient force. The wind of this nature will be Super-gradient wind. To restore balance, it will turn towards higher pressure (up the gradient). In the reverse case, an air parcel of low initial momentum moving into a region of higher pressure gradient will experience an imbalance such that the Pressure Gradient force exceeds the Coriolis force and this will result into Sub-gradient wind. In this case, the wind will turn towards lower pressure (down the gradient) to restore balance.



Equations of Motion in Spherical Polar Coordinate System

25. The shape of the earth is nearly spherical. Since the large-scale motions of the atmosphere essentially follow the shape of the earth, it is always easier and geometrically logical to express the equations of motion in spherical polar coordinates. The case of 'Horizontal' flow is then easily represented in such a frame.



26. In a spherical polar coordinate system, the position of a particle is given in terms of the Longitude (***λ***), Latitude (***Φ***) and Radial distance from the centre of the earth (***r***), where the earth is considered as the perfect sphere. Here, ***r*** may also be expressed as ***r=z+a***, where ***a*** is the radius of the earth and ***z*** is the altitude above mean sea level. Further, if ***x*** and ***y*** are the curvilinear distances towards east and north respectively along the sphere, then the curvilinear velocity components ***u***, ***v*** and ***w*** can be expressed as (refer the given figure):-

$$u = aCos\phi \frac{d\lambda}{dt}, \quad v = a\frac{d\phi}{dt} \quad and \quad w = \frac{dz}{dt}$$

and the velocity in vector form can be expressed as:-

$$\vec{V} = u\hat{i} + v\hat{j} + w\hat{k}$$

27. The most important point here is to remember that all the unit vectors (i.e., $\hat{i}, \hat{j}, \hat{k}$) will change in direction, as the particle moves. From a mathematical point of view, it implies that these unit vectors are functions of time, unlike those in case of rectangular Cartesian coordinate system. So, the acceleration can be found out as:

$$\frac{d\vec{V}}{dt} = \frac{du}{dt}\hat{i} + \frac{dv}{dt}\hat{j} + \frac{dw}{dt}\hat{k} + u\frac{d\hat{i}}{dt} + v\frac{d\hat{j}}{dt} + w\frac{d\hat{k}}{dt} \quad \text{..... (1.10)}$$

Where the total derivatives of unit vectors can be evaluated in terms of local changes with respect to space variables as:-

a. $\underline{d\hat{i}/dt}$. The total derivative can be written as:-

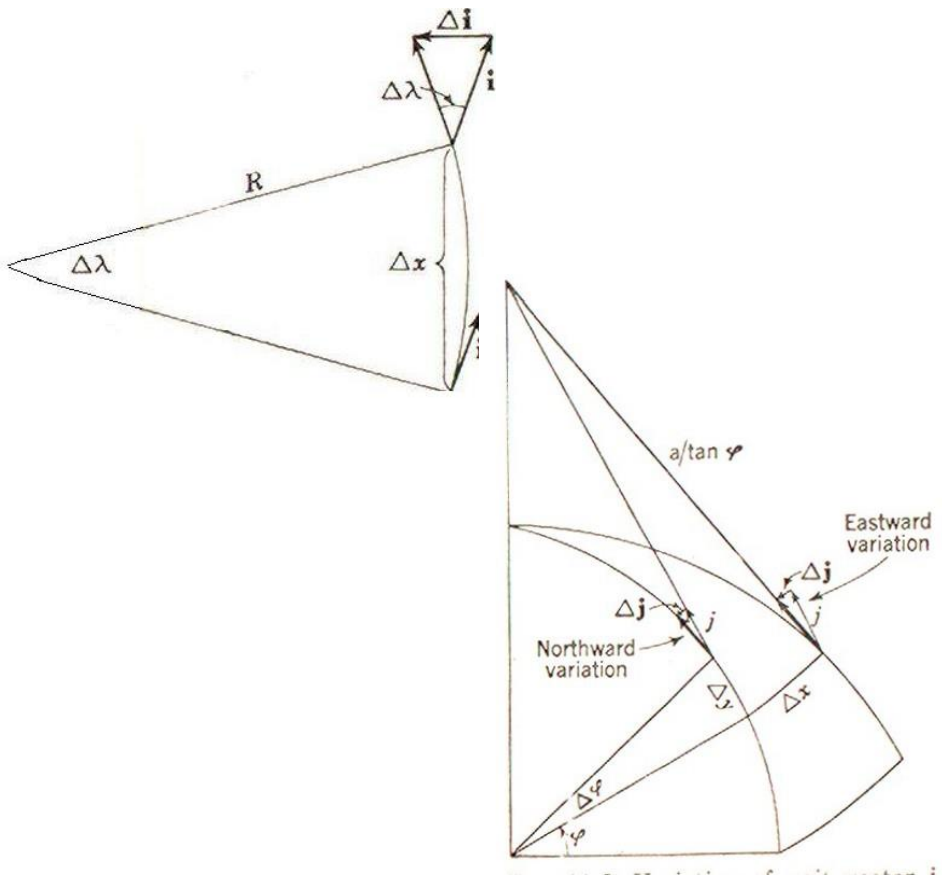
$$\frac{d\hat{i}}{dt} = \frac{\partial \hat{i}}{\partial t} + u\frac{\partial \hat{i}}{\partial x} + v\frac{\partial \hat{i}}{\partial y} + w\frac{\partial \hat{i}}{\partial z}$$

Where $\hat{i}$ would not change locally with respect to time (t), latitude (y) and height (z). So, it will be a function of longitude (x) only. From the figure given, it is clear that:-

$$\left|\frac{d\hat{i}}{dt}\right| = u\left|\frac{\partial \hat{i}}{\partial x}\right| = \frac{\Delta\lambda}{R\Delta\lambda} = \frac{1}{R} = \frac{1}{aCos\phi}$$

and its direction can be decomposed as $\hat{j}Sin\phi - \hat{k}Cos\phi$. So,

$$\frac{d\hat{i}}{dt} = \frac{u\tan\phi}{a}\hat{j} - \frac{u}{a}\hat{k}$$



b. $\underline{d\hat{j}/dt}$. The total derivative can be written as:-

$$\frac{d\hat{j}}{dt} = \frac{\partial \hat{j}}{\partial t} + u\frac{\partial \hat{j}}{\partial x} + v\frac{\partial \hat{j}}{\partial y} + w\frac{\partial \hat{j}}{\partial z}$$

Where $\hat{j}$ would not change locally with respect to time (t) and height (z). So, it will be a function of longitude (x) and latitude (y) only. From the figure given, it is clear that:-

$$\frac{\partial \hat{j}}{\partial x} = -\frac{\tan\phi}{a}\hat{i}$$

and $\dfrac{\partial \hat{j}}{\partial y} = -\dfrac{1}{a}\hat{k}$

So, $\dfrac{d\hat{j}}{dt} = -\dfrac{u\tan\phi}{a}\hat{i} - \dfrac{v}{a}\hat{k}$

c. $\underline{d\hat{k}/dt}$. The total derivative can be written as:-

$$\frac{d\hat{k}}{dt} = \frac{\partial \hat{k}}{\partial t} + u\frac{\partial \hat{k}}{\partial x} + v\frac{\partial \hat{k}}{\partial y} + w\frac{\partial \hat{k}}{\partial z}$$

Where $\hat{\boldsymbol{k}}$ would not change locally with respect to time (t) and height (z). So, it will be a function of longitude (x) and latitude (y) only. Similarl to earlier discussion, it is clear that:

$\dfrac{\partial \hat{k}}{\partial x} = \dfrac{1}{a}\hat{i}$ and $\dfrac{\partial \hat{k}}{\partial y} = \dfrac{1}{a}\hat{j}$ So, $\dfrac{d\hat{k}}{dt} = \dfrac{u}{a}\hat{i} + \dfrac{v}{a}\hat{j}$

28. Now, by substituting the values of total derivatives of all unit vectors in Eqn 1.10, we will get:

$$\frac{d\vec{V}}{dt} = \left(\frac{du}{dt}\hat{i} + \frac{dv}{dt}\hat{j} + \frac{dw}{dt}\hat{k}\right) + \left(\frac{u^2\tan\phi}{a}\hat{j} - \frac{u^2}{a}\hat{k}\right) + \left(-\frac{uv\tan\phi}{a}\hat{i} - \frac{v^2}{a}\hat{k}\right) + \left(\frac{uw}{a}\hat{i} + \frac{vw}{a}\hat{j}\right) \quad \text{or}$$

$$\frac{d\vec{V}}{dt} = \left(\frac{du}{dt} - \frac{uv\tan\phi}{a} + \frac{uw}{a}\right)\hat{i} + \left(\frac{dv}{dt} + \frac{u^2\tan\phi}{a} + \frac{vw}{a}\right)\hat{j} + \left(\frac{dw}{dt} - \frac{u^2}{a} - \frac{v^2}{a}\right)\hat{k}$$

If this equation is used for all the forces working on the particle, then:-

$$\left.\begin{aligned}
\frac{du}{dt} - \frac{uv\tan\phi}{a} + \frac{uw}{a} &= -\frac{1}{\rho}\frac{\partial p}{\partial x} + (2\Omega Sin\phi)v - (2\Omega Cos\phi)w + F_x \\
\frac{dv}{dt} + \frac{u^2\tan\phi}{a} + \frac{vw}{a} &= -\frac{1}{\rho}\frac{\partial p}{\partial y} - (2\Omega Sin\phi)u + F_y \\
\frac{dw}{dt} - \frac{u^2}{a} - \frac{v^2}{a} &= -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + (2\Omega Cos\phi)u + F_z
\end{aligned}\right\} \text{.... (1.11)}$$

These are the equations of motion in component form in spherical polar coordinate system.

Conclusion

29. The direction of motion and the normal to it are the basic orientations. So the natural coordinate system which has an s-axis tangent to the flow, along which s increases in the direction of flow, and an n-axis perpendicular to the flow ,along which n increases to the left of the direction of motion can explain appropriately different type of flow patterns, considering the balance between pressure gradient force, coriolis force and centrifugal force in the ideal frictionless situation. The theoretical concepts of geostrophy get disturbed in the presence of friction. The deviation of actual wind from the geostrophic wind due to the presence of frictional force is known as ageostrophic wind.

TASK-7 : VARIATION OF HORIZONTAL WIND IN VERTICAL

Objectives

After reading this chapter, you should be able to:-

- Understand the concept of Thermal Wind.
- Understand the different properties of Thermal Wind.
- Appreciate the applications of Thermal Wind.

Introduction.

1. Horizontal flow in the free atmosphere is approximately geostrophic for a broad horizontal flow, which is not often strongly curved. In order to examine the variation of geostrophic wind in vertical, the concept of Thermal Wind has been evolved.

The Thermal Wind

2. The Thermal Wind is defined as the vectorial difference of geostrophic winds at two levels, i.e., higher & lower levels. It is a fictitious wind and not found in the atmosphere. While the geostrophic wind is a level property, thermal wind is a layer property.

3. If geostrophic winds at higher and lower levels (separated by δz vertical distance, i.e., height difference) are $\vec{V}_{g_H}$ and $\vec{V}_{g_L}$, then:-

$$\vec{V}_{g_H} = \frac{1}{\rho f}\hat{k}\times\vec{\nabla}_H p_H \quad \text{and} \quad \vec{V}_{g_L} = \frac{1}{\rho f}\hat{k}\times\vec{\nabla}_H p_L$$

4. If the levels are chosen as isobaric surfaces (separated by δp), then these would be expressed as:-

$$\vec{V}_{g_H} = \frac{1}{f}\hat{k}\times\vec{\nabla}_p \phi_2 \quad \text{and} \quad \vec{V}_{g_L} = \frac{1}{f}\hat{k}\times\vec{\nabla}_p \phi_1$$

Where Φ_2 and Φ_1 represent the geopotential heights of both the isobaric levels. Then the thermal wind can be expressed as:-

$$\vec{V}_T = \vec{V}_{g_H} - \vec{V}_{g_L} = \frac{1}{f}\hat{k}\times\vec{\nabla}_p(\phi_2 - \phi_1)$$

or $$\vec{V}_T = \frac{1}{f}\hat{k}\times\vec{\nabla}_p(\Delta\phi) \qquad \text{..... (2.1)}$$

Where ΔΦ is the thickness of the layer. Thus, it is clear that "**the thermal wind blows parallel to the lines of equal thickness lines, keeping low values to the left, whereas the geostrophic wind blows parallel to the lines of equal geopotential values (i.e., contours), keeping low values to the left**". Hence, the thermal wind depends on the thickness gradient, while the geostrophic wind depends on the contour gradient.

5. The geostrophic wind components in isobaric coordinates are:

$$u_g = -\frac{1}{f}\frac{\partial \Phi}{\partial y} \text{ and } \qquad v_g = \frac{1}{f}\frac{\partial \Phi}{\partial x}$$

In order to find the variation of these with respect to height (or pressure), these components have to be differentiated with respect to *p*, as:-

$$\frac{\partial u_g}{\partial p} = -\frac{1}{f}\frac{\partial}{\partial y}\left(\frac{\partial \phi}{\partial p}\right) \text{ and } \qquad \frac{\partial v_g}{\partial p} = \frac{1}{f}\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial p}\right)$$

From the ideal gas law and the hydrostatic equation:-

$$\frac{\partial \Phi}{\partial p} = -\alpha = -\frac{RT}{p}$$

so, $$\frac{\partial u_g}{\partial p} = -\frac{1}{f}\frac{\partial}{\partial y}\left(-\frac{RT}{p}\right) \qquad \text{and} \qquad \frac{\partial v_g}{\partial p} = \frac{1}{f}\frac{\partial}{\partial x}\left(-\frac{RT}{p}\right)$$

so that,

$$p\frac{\partial u_g}{\partial p} = \frac{\partial u_g}{\partial \ln p} = \frac{R}{f}\left(\frac{\partial T}{\partial y}\right)_p \qquad \text{and} \qquad p\frac{\partial v_g}{\partial p} = \frac{\partial v_g}{\partial \ln p} = -\frac{R}{f}\left(\frac{\partial T}{\partial x}\right)_p$$

or in vectorial form:-

$$\frac{\partial \vec{V}_g}{\partial \ln p} = -\frac{R}{f}\hat{k} \times \vec{\nabla}_p T \qquad \text{..... (2.2)}$$

This equation is known as the Thermal Wind Equation, which relates the variation of geostrophic wind with pressure (height). Since pressure is the vertical coordinate, this is also referred to as the vertical wind shear of geostrophic wind. From this equation, it can be inferred that “the thermal wind blows parallel to the lines of equal mean temperature (i.e., mean isotherms) of the layer, keeping colder temperatures (i.e., low values) to the left in Northern Hemisphere”.

6. Integrating eqn (2.2) from p_0 to p_1, the thermal wind will be:

$$\vec{V}_T = \vec{V}_g(p_1) - \vec{V}_g(p_0) = -\frac{R}{f}\int_{p_0}^{p_1}\left(\hat{k} \times \vec{\nabla}_p T\right) d(ln\ p)$$

Now by letting $\overline{T}$ denote the mean temperature in the layer between pressure p_0 and p_1, the *x*- and *y*- components of the thermal wind are given as:-

$$u_T = -\frac{R}{f}\left(\frac{\partial \overline{T}}{\partial y}\right)_p \ln\left(\frac{p_0}{p_1}\right)$$

$$v_T = \frac{R}{f}\left(\frac{\partial \overline{T}}{\partial x}\right)_p \ln\left(\frac{p_0}{p_1}\right) \qquad \text{..... (2.3)}$$

Again, by integrating the hydrostatic equation between pressure p_0 and p_1, the hypsometric equation would be obtained as:-

$$\Phi_1 - \Phi_0 = R\overline{T}\ln\left(\frac{p_0}{p_1}\right)$$

So the thermal wind components can also be written as:-

$$u_T = u_g(p_1) - u_g(p_0) = -\frac{1}{f}\frac{\partial}{\partial y}(\Phi_1 - \Phi_0) \quad \text{and} \quad v_T = v_g(p_1) - v_g(p_0) = \frac{1}{f}\frac{\partial}{\partial x}(\Phi_1 - \Phi_0)$$

7. In the similar manner, the variation of geostrophic wind with height can also be found by differentiating it with respect to height *z*, as:-

$$\frac{\partial \vec{V}_g}{\partial z} = \frac{\partial \vec{V}_g}{\partial p}\frac{\partial p}{\partial z} = (-g\rho)\frac{\partial \vec{V}_g}{\partial p}$$

But $$\vec{V}_g = \frac{g}{f}\hat{k} \times \vec{\nabla} z$$

so, $$\frac{\partial \vec{V}_g}{\partial z} = (-g\rho)\frac{\partial}{\partial p}\left(\frac{g}{f}\hat{k} \times \vec{\nabla} z\right) = (-g\rho)\frac{g}{f}\left\{\hat{k} \times \vec{\nabla}\left(\frac{\partial z}{\partial p}\right)\right\}$$

or $$\frac{\partial \vec{V}_g}{\partial z} = (-g\rho)\frac{g}{f}\left\{\hat{k} \times \vec{\nabla}_H\left(-\frac{1}{g\rho}\right)\right\} = \frac{g}{f}\frac{p}{RT}\left\{\hat{k} \times \vec{\nabla}_H\left(\frac{RT}{p}\right)\right\}$$

or $$\frac{\partial \vec{V}_g}{\partial z} = \frac{g}{f}\left\{\hat{k} \times \frac{\vec{\nabla}_H T}{T}\right\} \qquad \text{..... (2.4)}$$

Which means again that vertical variation of geostrophic wind depends on horizontal thermal gradient and would blow parallel to isotherms, keeping low temperatures to the left in Northern Hemisphere.

8. By suitable substitution of Potential Temperature (θ) and Specific Volume (α), other relations can be obtained for vertical variation of geostrophic wind with height as:-

$$\frac{\partial \vec{V}_g}{\partial z} = \frac{g}{f}\left\{\hat{k} \times \frac{\vec{\nabla}_H \theta}{\theta}\right\} = \frac{g}{f}\left\{\hat{k} \times \frac{\vec{\nabla}_H \alpha}{\alpha}\right\} \qquad \text{..... (2.5)}$$

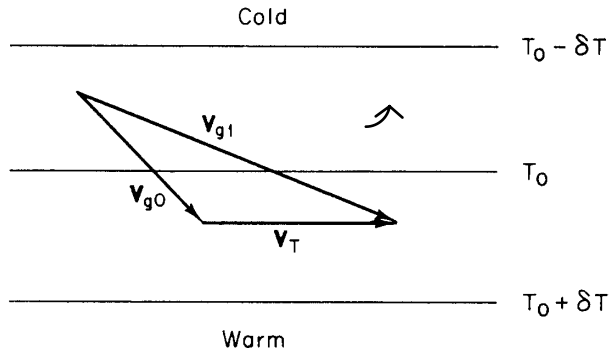
9. The vector quantity thus obtained (i.e., $\partial \vec{V}_g / \partial z$) can also be called as 'Thermal Wind Shear vector'.

<u>Thermal Wind and Thermal Advection</u>

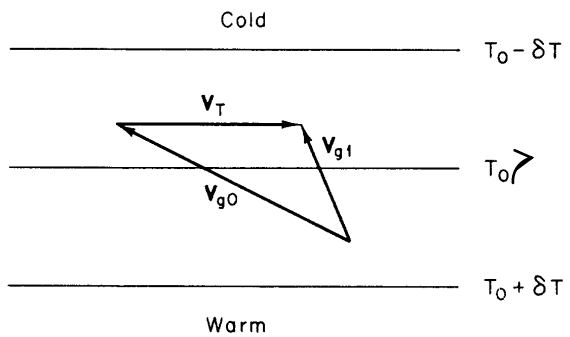
10. From Eqn (3.2) & (3.3), it is evident that the thermal wind blows parallel to mean isotherms, keeping colder temperatures to the left. This can be explained by the figures given below that thermal advection depends on how winds are changing with height, as:-

a. **<u>Winds Backing with Height</u>.** In the given example below, winds are backing (i.e., turning anticlockwise) with height. Now, since the thermal wind would blow parallel to isotherms, keeping colder temperature to left, winds at both the levels would

bring colder temperature towards warmer temperature. Hence, it would cause 'Cold Air Advection'.



b. **<u>Winds Veering with Height</u>**. In the given example below, winds are veering (i.e., turning clockwise) with height. Now, since the thermal wind would blow parallel to isotherms, keeping colder temperature to left, winds at both the levels would bring warmer temperature towards colder temperature. Hence, it would cause 'Warm Air Advection'.



11. Therefore, a geostrophic wind turning anticlockwise (i.e., backing) with height would be associated with Cold Air Advection, while the geostrophic wind turning clockwise (i.e., veering) with height would be associated with Warm Air Advection.

12. The thermal wind equation also suggests that if the geostrophic wind at one level and the mean temperature between two levels are known, the geostrophic wind at the second level can be calculated.

<u>Thermal Wind and Relative Stability</u>

13. One of the important applications of thermal wind is to find out the zones of relative stability/instability from the observed wind profile over a place at any given time. The concept used is the classical one that the warmer air under colder air would be unstable, while colder air under warmer air would be stable. This state can be qualitatively found out from the observed wind profile.

14. Let us consider three different levels over the same place, with such a wind profile that wind directions at each level change with height. In the figure below (Fig. 1), a typical example has been given, in which, winds are veering from the lowest level (level 1) to middle level (level 2) and then backing from middle level to higher level (level 3). Thus, dividing the entire column into two layers, thermal winds of both the layers (L & H, i.e., lower & higher) can be obtained, which would represent the colder and warmer regions in respective layers. For better appreciation, both the thermal winds have been separately drawn in larger proportions in Fig. 2.

2 $\vec{V}_{T_H}$

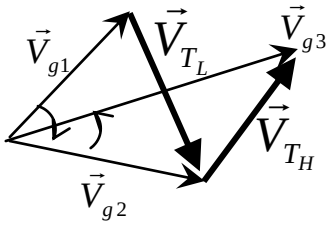


Fig. 1

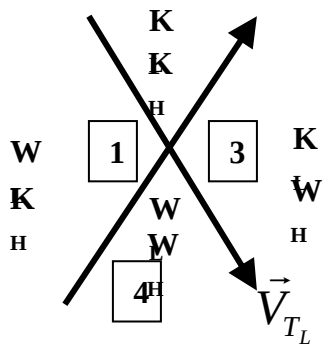
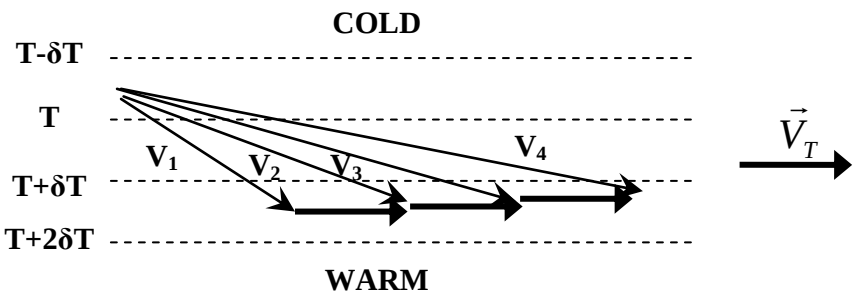


Fig. 2

15. As depicted in Fig. 2, thermal winds of both the layers would be segregating colder and warmer areas. Thus, dividing the entire area in four sectors, it can be seen that sector – 1 would have warmer air in lower layer with colder air in higher layer; on the other hand, sector – 3 would have colder air in lower and warmer air in higher layer. Hence, sector – 1 would be unstable, while sector – 3 would be stable. To examine the relative stability in sectors – 2 & 4, qualitative assessment would not be sufficient and quantitative assessment would have to be carried out.

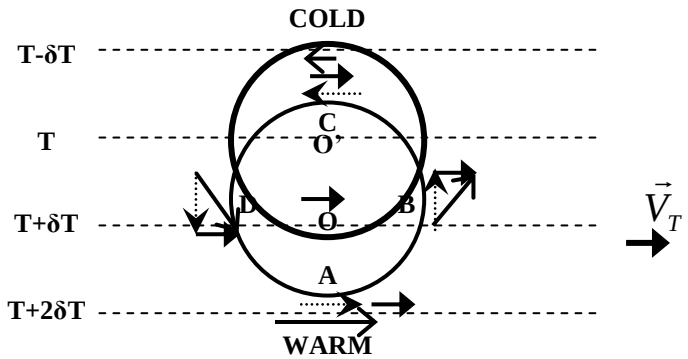
Applications of Thermal Wind

16. **Formation of Sub-tropical Jet Stream**. It is an observed fact that W'ly / NW'ly wind flow prevails in mid-latitudes right from the lowest levels, which gradually increase in strength with height. In the higher levels, very strong W'ly winds are found, which result into formation of 'Sub-tropical Jet Stream' (STJ). At a certain level, wind speeds reach to a maximum value, which is known as 'Level of Maximum Winds', above which wind speeds again start decreasing with height. This phenomenon of formation of STJ can be explained with the help of thermal wind concept. As mentioned earlier, thermal wind blows parallel to isotherms keeping colder temperatures to left in the NH and the upper level winds can be found if lower level winds are known with the horizontal temperature distribution within the layer. For mid-latitudes, colder temperatures in lower levels remain to the north, so thermal wind would remain W'ly, which would add to the existing W'ly / NW'ly flow of lower levels to give stronger upper level flow (as shown in the figure below). In this way, winds keep on strengthening with height till a particular level, where horizontal temperature gradient vanishes and thermal wind becomes zero. This happens at the level of tropopause of mid-latitudes, above which temperatures remain almost constant with height, while to the south (i.e., equator-wards) temperatures keep decreasing as tropical tropopause remains still higher. Hence, the level of maximum winds (i.e., Jet core) is found near the mid-latitude tropopause, above which horizontal thermal contrast gets reversed and north becomes warmer that south. As the result, thermal wind becomes E'ly and further higher level winds weaken in strength.

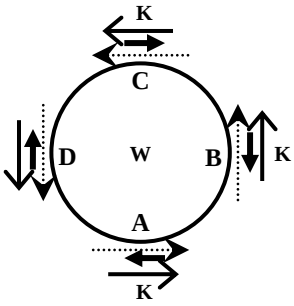


17. **SW-Monsoon : Formation of TEJ**. The formation of Tropical Easterly Jet Stream (TEJ) during SW-Monsoon season over Indian sub-continent can also be explained with the help of thermal wind concept in the same way as explained above. During SW-Monsoon season, the Heat low pressure area remains over Pakistan and neighbourhood, causing warmer temperatures over north and colder temperature to the south. As a result of this, thermal wind remains E'ly, which reduces the strength of SW'ly flow of lower levels over Arabian Sea and West Coast. Winds at higher levels keep decreasing and finally change into E'ly at one particular level. Further higher, E'ly winds keep strengthening and finally reach to the level of maximum wind at the level, where the existing N-S thermal gradient becomes almost zero. At this level, core of TEJ is found.

18. **Low Pressure Area in Unidirectional Thermal Pattern**. With the help of thermal wind concept, the slope (with height) of a Low Pressure Area can also be found out. To see this, let us consider the horizontal thermal pattern in such a manner that isotherms are nearly E-W oriented and north remains colder than south (i.e., horizontal thermal gradient is N-S oriented), as shown in the figure below. Due to this thermal gradient, thermal wind would be W'ly throughout. This W'ly thermal wind, when added to lower level winds (shown as thin arrows at points A, B, C & D), would give slightly different winds at next higher level (shown as thick arrows at various points). The noteworthy aspect is that the wind at centre (Point O), which was calm at the lower level, would become W'ly at next higher level and the point of calm wind (i.e., the centre) would shift to the north (i.e., Point O'). Further, at northern most point of the circulation (Point C), E'ly winds at lower level would weaken at next higher level. Thus, it can be inferred that the centre of the circulation shifts towards north with height. Therefore, it can be concluded that **the Low Pressure Area slopes towards the area of colder temperatures**.



19. **Intensity variation of Cold / Warm core Systems with Height**. Various pressure / winds systems in the atmosphere quite often reveal different thermal core characteristics. With the help of thermal wind concept, their intensity variations with respect to height can be ascertained. To show this effect, let us consider a simple case of a warm core Low (or Cyclonic Circulation). In such a case, centre of the Low would be warmer than the surroundings. So, thermal wind at each point would be opposite to the basic flow pattern at lower level. In the figure given below, lower level winds at various points (A, B, C & D) have been shown as thin arrows and thermal wind as thick arrows. As a result, winds at next higher level would be weaker (shown by large arrow-heads). So, it can be concluded that ***a warm core low weakens with height***. Winds at various points would keep weakening with height and flow pattern would change into an anticyclonic manner at a certain level



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20. Thus, it can be concluded that:

a. Warm core cyclones / Low pressure areas and Cold core anticyclones / High pressure areas weaken with height.

b. Cold core cyclones / Low pressure areas and Warm core anticyclones / High pressure areas intensify with height.

Conclusion.

21. It is very clear that the geostrophic wind must have vertical shear in the presence of a horizontal temperature gradient. The equation for the rate of the geostrophic wind components with height are most easily derived using the isobaric coordinate system. The combination of these components in vectorial form is referred as the thermal wind equation. The concept of thermal wind explains the relationship between turning of the geostrophic wind and temperature advection. The formation of subtropical jet stream, tropical jet stream and intensity variation of cold/warm core systems with height could also be explained with the help of thermal wind concept.

TASK-8 : CIRCULATION, VORTICITY AND DIVERGENCE

Introduction

1. Linear motion: measured by (linear) velocity. Circular motion: measured by angular velocity. There are alternative ways to measure circular motion than the angular velocity. Two primary measures of rotation in a fluid are:

   a. Circulation – macroscopic measure for a fluid area, which is a scalar integral quantity.
   b. Vorticity – microscopic measure at a point of the fluid, which is a vector.

Definition of the circulation

2. The circulation, C, about a closed contour in a fluid is defined as the line integral evaluated counterclockwise along the contour of the component of the velocity vector that is locally tangent to the contour. The **circulation** *C* around a given closed curve in a fluid is given by

$$C = \oint \mathbf{V} \cdot d\mathbf{s} = \oint V \cos \alpha \, ds$$

where **V** is the velocity vector, *d***s** is an element of the closed curve and □ the angle between **V** and *d***s***.*
Convention: *C* > 0 for anticlockwise integration. If *dx, dy, dz* are components of *d***s**, then

$$C = \oint \left(udx + vdy + wdz \right)$$

3. **Relationship Between Circulation and Angular Velocity**. For a circular disk of radius *r* and in solid-body rotation with angular velocity □ about the *z* axis,

$$\mathbf{V} = \square \times \mathbf{r}$$

where **r** is the distance to the axis of rotation. Then, the circulation *C* about the edge of the disk is

$$C = \oint \mathbf{V} \cdot d\mathbf{s} = \int_0^{2\pi} \Omega r^2 \, d\lambda = 2\Omega \pi r^2 \qquad (5.1)$$

Therefore, the circulation is just 2π times the angular momentum of the fluid ring about the axis of rotation. Note also from (5.1) that

$$\frac{C}{\pi r^2} = 2\Omega$$

That is, the circulation per unit area is twice the angular speed of the fluid ring. Since circulation can be calculated with other means, the angular momentum or angular speed can be deduced.

4. **Time Rate of Change of Circulation**. The time rate of change of circulation is important because circulation is related to the rotation of the fluid. Knowing *dC/dt* will provide an idea of the time change of rotation. Consider a closed loop of fluid particles moving with a 3-D velocity vector $\mathbf{U_a}$ in an *absolute* coordinate system:

$$\oint \frac{d_a \mathbf{U}_a}{dt} \cdot d\mathbf{l} = -\oint \frac{\nabla p \cdot d\mathbf{l}}{\rho} - \oint \nabla\Phi \cdot d\mathbf{l}$$

where the gravitational force **g** is represented as the gradient of the geopotential □□ **= g** = - *g***k**. The integrand on the left-hand side can be rewritten as

$$\frac{d_a \mathbf{U}_a}{dt} \cdot d\mathbf{l} = \frac{d}{dt}\left(\mathbf{U}_a \cdot d\mathbf{l}\right) - \mathbf{U}_a \cdot \frac{d_a}{dt}\left(d\mathbf{l}\right)$$

Since **l** is a position vector, $d_a\mathbf{l}/dt$ □ $\mathbf{U}_a$ so that

$$\frac{d_a \mathbf{U}_a}{dt} \cdot d\mathbf{l} = \frac{d}{dt}\left(\mathbf{U}_a \cdot d\mathbf{l}\right) - \mathbf{U}_a \cdot d\mathbf{U}_a$$

Since the line integral of a perfect differential is zero,

$$\oint \nabla\Phi \cdot d\mathbf{l} = \oint d\Phi = 0$$

and

$$\oint \mathbf{U}_a \cdot d\mathbf{U}_a = \frac{1}{2}\oint d\left(\mathbf{U}_a \cdot \mathbf{U}_a\right) = 0$$

Therefore,

$$\boxed{\frac{dC_a}{dt} = \frac{d}{dt}\oint \mathbf{U}_a \cdot d\mathbf{l} = -\oint \frac{dp}{\rho}} \qquad (5.2)$$

This is called the circulation theorem. The term $-\oint dp/\rho$ is called the **solenoidal** term. Note that if the density is constant, the line integral again becomes zero. In this case, the absolute circulation is a constant. This is known as **Kelvin's Circulation Theorem**.

Since part of the absolute circulation consists of the circulation due to the rotation of the earth C_e, it is necessary to calculate C_e in order to obtain the time rate of change of the relative circulation *C.*
If $\mathbf{U}_e$ = □ **x r** represents the velocity of the earth at **r**, the relative circulation C_e is

$$C_e = \oint \mathbf{U}_e \cdot d\mathbf{l} = \iint_A \left(\nabla \times \mathbf{U}_e\right) \cdot \mathbf{n} dA$$

where *A* is the area enclosed by the contour, and the unit normal vector **n** is defined by the counterclockwise sense of the line integration. For the area *A* to be a horizontal plane on the earth, **n** is parallel to the local vertical so that

$$\left(\nabla \times \mathbf{U}_e\right) \cdot \mathbf{n} = 2\Omega \sin\phi = f$$

so that

$$C_e = 2\Omega \sin\phi A = 2\Omega A_e$$

where A_e is the projection of area A onto the equatorial plane. The relative circulation C is therefore given by

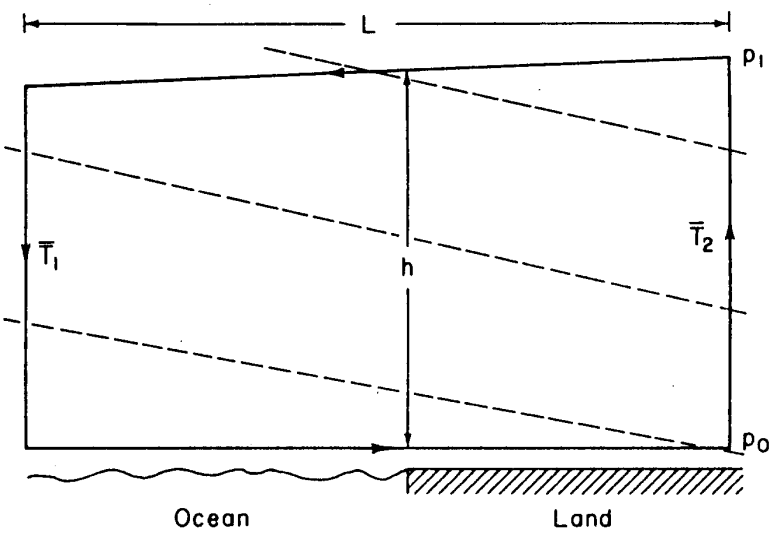
$$C = C_a - C_e = C_a - 2\Omega A_e$$

Differentiating the equation and substitituting from (8.2) gives

$$\boxed{\frac{dC}{dt} = -\oint \frac{dp}{\rho} - 2\Omega \frac{dA_e}{dt}}$$

This is known as the **Bjerknes Circulation Theorem**.

5. **<u>The Solenoidal Term</u>** : **<u>The Land-Sea Breeze Circulation.</u>** Consider the following distribution of pressure and density:



Substituting the ideal-gas law into (5.2),

$$\frac{dC_a}{dt} = -\oint RT\, d\ln p$$

The only contribution to the integral comes from the vertical segments so that

$$\frac{dC_a}{dt} = R\ln\left(\frac{p_o}{p_1}\right)\left(\overline{T}_2 - \overline{T}_1\right) > 0$$

This suggests that the circulation will increase and in a direction s.t. denser fluid sinks and lighter fluid rises. This will convert potential energy of the mass distribution of the fluid to kinetic energy of the circulation. Such a circulation (one in which the circulation moving in the same sense as the direction of the solenoidal term would impose) is called a **direct solenoidal circulation**.

In an atmosphere in which the pressure and density (or temperature) surfaces are at an angle (which is called a **baroclinic** atmosphere), air would always try to set up a direct solenoidal circulation so as to reduce this angle, i.e. to decrease the **baroclinity** of the atmosphere.

An **indirect solenoidal circulation** is one in which the circulation is moving in the *opposite* sense as the direction of the solenoidal term would impose. Such a circulation converts kinetic energy to potential energy and weakens with time.

6. **The Coriolis Term**. This term is important usually if the solenoidal term is small (*i.e.* pressure and density surfaces coincide, which represents the condition in a **barotropic** atmosphere). It will contribute to a change in the relative circulation if the projected area onto the equatorial plane changes through

a. A change in the actual area accomplished by changing the convergence or divergence -(*divergence effect*), or

b. A change in the latitude accomplished by moving the area northwards or southwards -(*latitude effect*)

A horizontal gradient of vertical velocity across the circuit could lead to individual fluid particles moving in/out of the horizontal plane and *A* will change with time - *tipping effect*

Vorticity

7. Vorticity is a *microscopic* measure of the rotation of a fluid and is a vector defined as the curl of the velocity vector.

absolute vorticity $\omega_a \equiv \nabla \times \mathbf{U}_a$
relative vorticity $\omega \equiv \nabla \times \mathbf{U}$

In meteorology, we are mainly concerned with the vertical components of vorticity:

absolute vorticity $$\eta = \mathbf{k} \cdot \left(\nabla \times \mathbf{U}_a\right)$$

relative vorticity $$\zeta = \mathbf{k} \cdot \left(\nabla \times \mathbf{U}\right)$$

planetary vorticity $$f = \mathbf{k} \cdot \left(\nabla \times \mathbf{U}_e\right) = 2\Omega \sin \phi$$

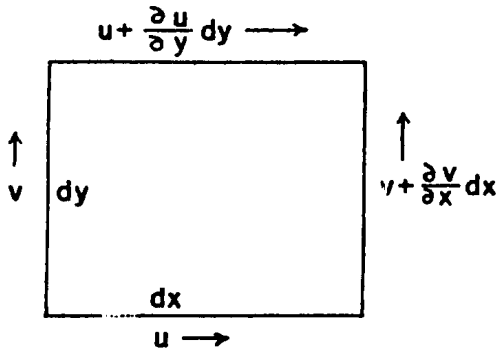
Therefore,

$$\square = \square + f$$

In Cartesian coordinates

$$\boxed{\begin{aligned}\zeta &= \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\\ \eta &= \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} + f\end{aligned}}$$

8. **<u>Relationship Between Circulation and Vorticity</u>**. Consider the circuit formed by an infinitesimal rectangle:



The contribution to circulation around this rectangle is

$$dC_{dxdy} = udx + \left(v + \frac{\partial v}{\partial x}dx\right)dy - \left(u + \frac{\partial u}{\partial y}dy\right)dx - vdy$$

or

$$dC_{dxdy} = \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)dxdy$$

Therefore, for a small element of area, the vorticity of the fluid in the area is equal to the circulation around the perimeter per unit area.

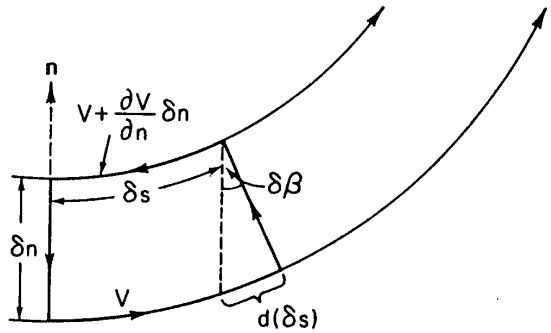
This is actually a special case of Stokes' Theorem:

$$\oint \mathbf{V}\cdot d\mathbf{s} = \int_A \left(\nabla \times \mathbf{V}\right)\cdot d\mathbf{A}$$

or in scalar form,

$$\oint (udx + vdy) = \iint_A \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)dxdy$$

9. **<u>Vorticity in Natural Coordinates</u>**. Consider the flow in the following situation:



The circulation around the contour is

$$\delta C = V[\delta s + d(\delta s)] - \left(V + \frac{\partial V}{\partial n}\delta n\right)\delta s$$

Since

$$d(\square s) = \square\square\ \square n$$

$\square C\square$ is equal to

$$\delta C = \left(-\frac{\partial V}{\partial n} + V\frac{\delta\beta}{\delta s}\right)\delta n\,\delta s$$

so that as $\square n$, $\square s \square 0$,

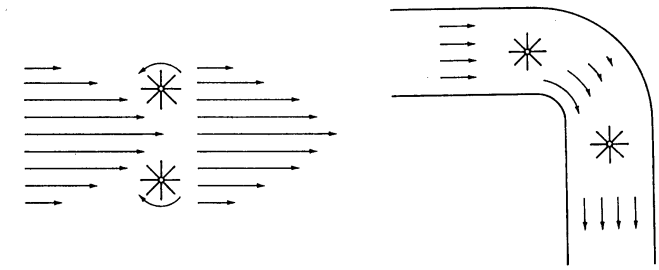
$$\boxed{\zeta = \lim_{\delta n,\delta s\to 0}\frac{\delta C}{\delta n\,\delta s} = -\frac{\partial V}{\partial n} + \frac{V}{R_s}}$$

$\square\ \square$ consists of two parts:

a. $-\square V/\square n$: the rate of change of wind speed $\square$ to the direction of flow - **shear vorticity**

b. V/R_s turning of the wind along a streamline - **curvature vorticity**:

Therefore, straight flow can contain vorticity due to shear of the wind while curved flow may have zero relative vorticity if the effect of shear cancels that of curvature.



10. **Divergence.** The divergence of a fluid is defined as

$$\nabla \cdot \mathbf{V} = \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right)$$

A flow directed outward is called pure **divergence** and a flow inward is pure **convergence** or negative divergence.

11**.** **The Vorticity Equation.** Similar to calculating *dC/dt*, we wish to derive an equation describing the change in vorticity with time. To do this, we use the equations of motion and differentiate the *x*-component equation w.r.t. *y* and the *y*-component equation w.r.t. *x*:

$$\frac{\partial}{\partial y}\left\{ \frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} \right\}$$

$$\frac{\partial}{\partial x}\left\{ \frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y} \right\}$$

Subtracting the two equations gives

$$\frac{\partial \zeta}{\partial t} + u\frac{\partial \zeta}{\partial x} + v\frac{\partial \zeta}{\partial y} + w\frac{\partial \zeta}{\partial z} + (\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)$$

$$+ \left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z} \right) + v\frac{df}{dy} = \frac{1}{\rho^2}\left[\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x} \right]$$

Since *f* only depends on *y*,

$$\frac{df}{dt} = v\frac{df}{dy}$$

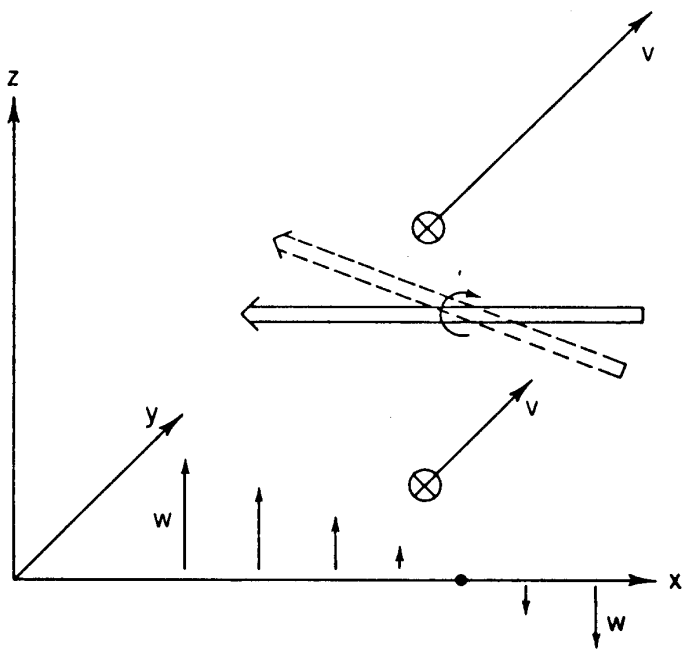
so that

$$\frac{d}{dt}(\zeta + f) = \underbrace{-(\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)}_{A} \underbrace{-\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right)}_{B}$$

$$+ \underbrace{\frac{1}{\rho^2}\left[\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right]}_{C}$$

Term A - **divergence term**
The absolute vorticity, (+ *f*) is almost always +ve. Therefore, convergence increase in . Convergence means increase in mass increase in thickness of column -*stretching of vortex tube*.

Term B - **tilting term**
A horizontal shear of the vertical velocity will tilt a vorticity vector in the horizontal to the vertical:



Vorticity generation by the tilting of a horizontal vorticity vector (double arrow).

Term C - **solenoid term**

A crossing between the density and pressure surfaces leads to change in. Therefore, it is essentially the same term as the solenoidal term in the circulation theorem. This term disappears if the vorticity equation is derived in isobaric coordinates.- (EXERCISE)

12. **<u>Scale Analysis of the Vorticity Equation</u>**. Using the same scales as before for mid-latitude synoptic-scale weather systems and noting that

$$\frac{df}{dy} = \beta \sim \frac{f_o}{L} \sim 10^{-11}\ \mathrm{m^{-1}\ s^{-1}}$$

we can scale the various terms in the vorticity equation:

$$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \leq \frac{U}{L} \sim 10^{-5}\ \mathrm{s^{-1}}$$

Since f_o is ~ 10^{-4} s^{-1}, we see that the relative vorticity for synoptic-scale weather systems is small compared with the planetary vorticity so that the divergence term

$$(\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \cong f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)$$

The magnitudes of the other terms are scaled as follows:

$$\frac{\partial \zeta}{\partial t}, u\frac{\partial \zeta}{\partial x}, v\frac{\partial \zeta}{\partial y} \sim \frac{U^2}{L^2} \sim 10^{-10}\ \mathrm{s^{-2}}$$

$$w\frac{\partial \zeta}{\partial z} \sim \frac{UW}{LH} \sim 10^{-11}\ \mathrm{s^{-2}}$$

$$v\frac{df}{dy} \sim U\beta \sim 10^{-10}\ \mathrm{s^{-2}}$$

$$f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) \leq \frac{f_o U}{L} \sim 10^{-9}\ \mathrm{s^{-2}}$$

$$\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) \leq \frac{WU}{LH} \sim 10^{-11}\ \mathrm{s^{-2}}$$

$$\frac{1}{\rho^2}\left[\frac{\partial \rho}{\partial x}\frac{\partial p}{\partial y} - \frac{\partial \rho}{\partial y}\frac{\partial p}{\partial x}\right] \sim \frac{\delta\rho\delta p}{\rho^2 L^2} \sim 10^{-11}\ \mathrm{s^{-2}}$$

Keeping only terms with magnitudes ~ 10^{-10} s^{-2},

$$\frac{d_{\mathrm{h}}(\zeta + f)}{dt} = -f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)$$

Where

$$\frac{d_h}{dt} = \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y}$$

Therefore, for mid-latitude synoptic-scale weather systems, the generation of vorticity is almost entirely due to the effect of convergence.

For an incompressible fluid,

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -\frac{\partial w}{\partial z}$$

so that

$$\frac{d_{\mathrm{h}}(\zeta + f)}{dt} = (\zeta + f)\frac{\partial w}{\partial z}$$

If the wind is geostrophic, the relative vorticity is independent of height (*Why?*). Then, integrate the equation from $z = D_1$ to $z = D_2$,

$$(D_2 - D_1)\frac{d_{\mathrm{h}}(\zeta_{\mathrm{g}} + f)}{dt} = (\zeta_{\mathrm{g}} + f)[w(D_2) - w(D_1)]$$

Since $w = dz/dt$, the vertical velocities at D_1 and D_2 can be written as

$$w(D_2) = \frac{dD_2}{dt} \text{ and } w(D_1) = \frac{dD_1}{dt}$$

Define the depth of the fluid $H = D_2 - D_1$ and substituting into the above equation,

$$\frac{1}{(\zeta_{\mathrm{g}} + f)}\frac{d_{\mathrm{h}}(\zeta_{\mathrm{g}} + f)}{dt} = \frac{1}{H}\frac{d_{\mathrm{h}}H}{dt}$$

or

$$\frac{d_{\mathrm{h}}}{dt}[\ln(\zeta_{\mathrm{g}} + f)] = \frac{d_{\mathrm{h}}}{dt}[\ln H]$$

which reduces to

$$\boxed{\frac{d_{\mathrm{h}}}{dt}\left(\frac{\zeta_{\mathrm{g}} + f}{H}\right) = 0}$$

The term $(_{\mathrm{g}} + f)/H$ is called the **potential vorticity** for a barotropic (or incompressible) fluid. This equation says that *potential vorticity of a barotropic fluid is conserved following the motion of a parcel.*

If $w = 0$, the conservation equation further reduces to

$$\frac{d_{\mathrm{h}}}{dt}\left(\zeta + f\right) = 0$$

which states that *the absolute vorticity is conserved following the motion if it is purely horizontal.* Note that in this case, the flow does not have to be geostrophic.

13. **<u>Potential Vorticity</u>**. Since potential temperature is conserved following the motion in unsaturated adiabatic flow, it is of interest to study the variation of vorticity on adiabatic surfaces. From the Poisson Equation, it can be shown (*you should do it*) that

$$\rho \propto p^{c_v/c_p}$$

so that density is only a function of pressure on surfaces. This means that the solenoidal term in the circulation theorem will be zero, *i.e.*

$$\oint \frac{dp}{\rho} \propto \oint dp^{(1-c_v/c_p)} = 0$$

Therefore, the circulation theorem on a surface is the same as that for a barotropic fluid, *i.e*.

$$\frac{d}{dt}(C + 2\Omega A \sin\phi) = 0$$

where the circulation C is evaluated for a closed loop which encompasses an area A on a surface. If the surface is approximately horizontal, C can be related to the vertical component of relative vorticity as

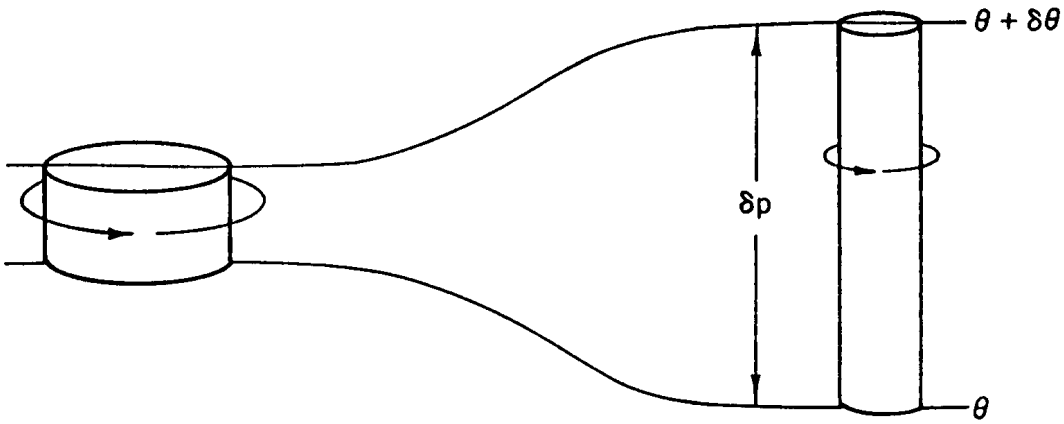
$$\zeta = \lim_{A\to 0} \frac{C}{A}$$

so that the circulation theorem can be written as

$$A(_{\mathrm{o}} + f) = \text{constant}$$

since $f = 2$ sin.

Consider the following distribution of surfaces so that a parcel is confined between two surfaces $_{\mathrm{o}}$ and $_{\mathrm{o}}$ + which are separated by a distance p:



A cylindrical column of air moving adiabatically, conserving potential vorticity.

The mass of the parcel $M = A\,p/g$ must be conserved following the motion so that

$$A = \frac{Mg}{\delta p} = \left(\frac{\delta\theta}{\delta p}\right)\left(\frac{Mg}{\delta\theta}\right) = \text{const} \times \left(\frac{\delta\theta}{\delta p}\right)$$

since = constant. Substituting and let p 0,

$$\boxed{(\zeta + f)\frac{\partial\theta}{\partial p} = \text{constant}}$$

The quantity (+ f)(/p) is called the **potential vorticity**. This is the same as the definition before for a barotropic fluid because in this case, density is a constant so that

$$A = \frac{M}{\rho\delta z} = \frac{\text{constant}}{\delta z}$$

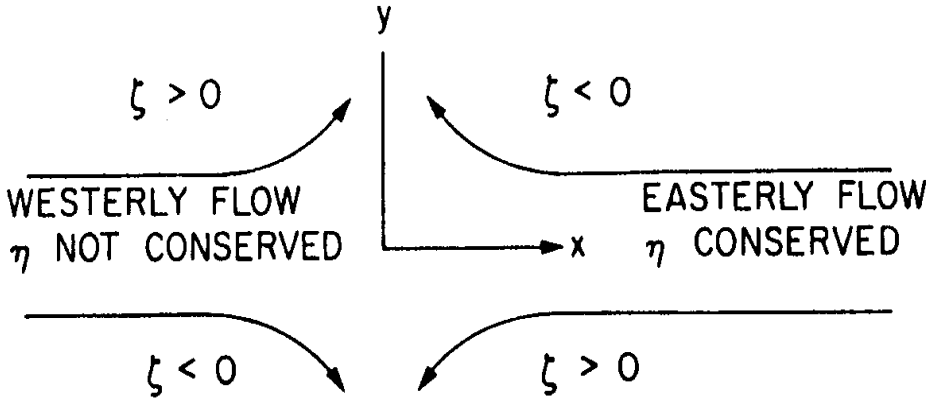
where z is the depth of the fluid. Eliminating A, we obtain

$$\frac{\zeta + f}{\delta z} = \text{constant}$$ The potential vorticity is therfore a measure of the ratio of the absolute vorticity to the effective depth of the column.

Either equation suggests that *potential vorticity is conserved in adiabatic, frictionless motion*. Any change in potential vorticity must therefore be due to diabatic forcing (latent heat release, solar radiation, other heat sources) or friction.

Because the conservation property depends on two factors: the absolute vorticity (ζ+f) and the stability of the atmosphere ($\partial\theta/\partial p$), two situations can be considered.

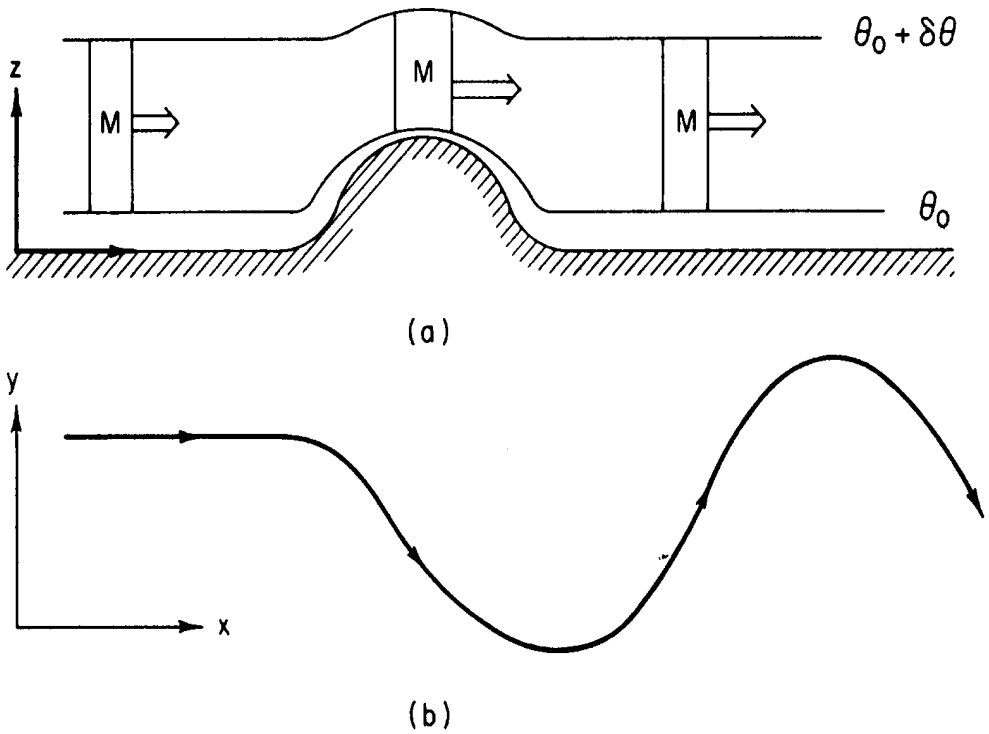
CASE 1 - /p = constant. Then, (+ f) is conserved following a parcel.



Absolute vorticity conservation for curved flow trajectories.

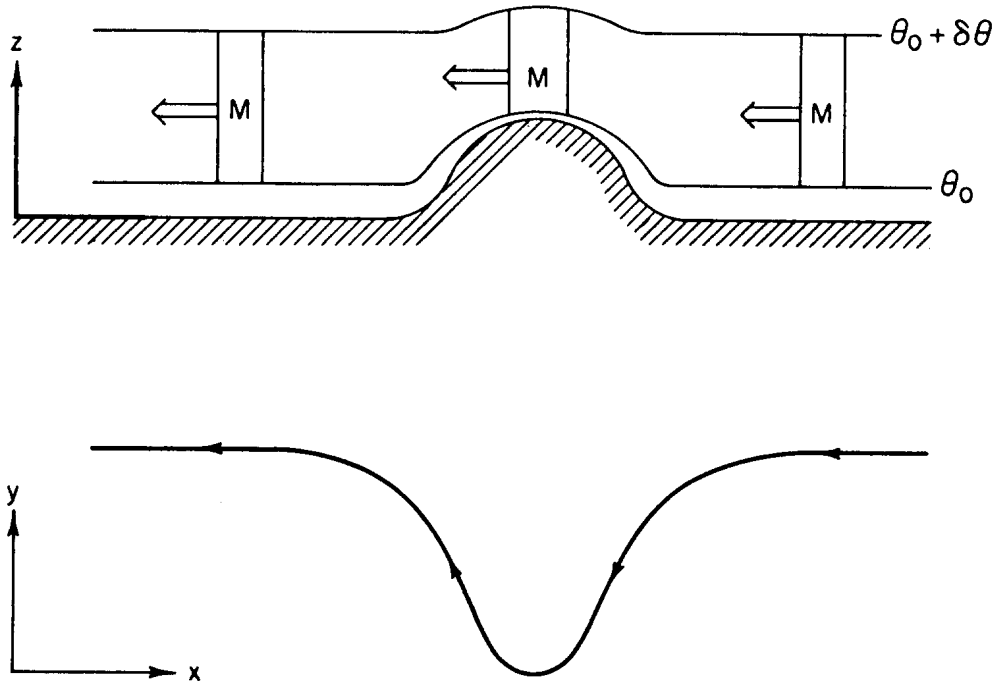
CASE 2 - Flow over a mountain (/p constant)

14. **Westerly flow**



Westerly flow over a topographic barrier: (a) the depth of a column as a function of x; (b) the trajectory of a parcel in the x,y plane.

15. **Easterly Flow.**



Easterly flow over a topographic barrier.

SECTION – 3 : WEATHER SYSTEMS

TASK – 9 : THUNDERSTORM AND TORNADO

Objectives

After reading the chapter you should be able to :

- Assimilate the conditions favourable for the formation of CB cloud.
- Understand Life Cycle of a CB Cell.
- Understand structure of Thunderstorm.
- Assimilate the Flying Hazards of a Thunderstorm and Avoidance and flying techniques in a Thunderstorm.

Introduction

1. Cumulonimbus is a cumulus cloud, which develops upto great heights due to instability and a high degree of humidity in a deep layer of air. The release of energy due to the over-turning of air in the unstable layers gives rise to a storm. The electrical charges developed in the cloud give rise to lightning and thunder. Thunderstorms are one or more convective cells in which electrical discharges are seen as lightning or heard as thunder.

2. The thunderstorm is by far the most dangerous weather phenomenon from the point of view of aviation.

Conditions Favourable for Cumulonimbus Cloud Formation

3. The necessary conditions for the formation of Cumulonimbus (Cb) clouds are as follows:

a. Lapse rate steeper than the SALR throughout a layer at least 5 to 6 kilometers in depth, permitting development of clouds to heights at which the temperature is below 0ºC.

b. Adequate supply of moisture from below.

c. A process which produces saturation in the region of the steep lapse rate or a triggering mechanism.

4. As the instability increases cloud grows upwards, some of the surrounding unsaturated air is entrained into the cloud mass. Consequently, some of the cloud droplets evaporate. If the humidity of the surrounding air is very low, the evaporation becomes dominant and arrests further growth of the cloud. Well-developed Cbs are thus possible only when the humidity aloft is sufficiently high.

5. The triggering mechanism, which sets off a thunderstorm, may be:

a. Local convection (insolation).

b. Orographic lifting.

c. Convergence.

d. Frontal lifting.

e. Radiational or katabatic cooling.

Cellular Structure

6. The fundamental structural element of the thunderstorm is the unit of convective circulation known as the convective cell. A mature thunderstorm contains several of these cells, which vary in diameter from 2 to 10 kms. By radar analysis and measurement of drafts, it has been determined that, generally, each cell is independent of surrounding cells of the same storm. Each cell progresses through a cycle which lasts from 1 to 3 hours. In the initial stage (cumulus development), the cloud consists of a single cell, but as the development progresses, new cells form and older cells dissipate.

Life Cycle of a Cell

7. The life cycle of the thunderstorm cell consists of three distinct stages; they are the cumulus stage, the mature stage, and the dissipating or anvil stage. (See fig. 5.1)

a. **Cumulus Stage**. Although most cumulus clouds do not become thunderstorms, the initial stage of a thunderstorm is always a cumulus cloud. The chief distinguishing feature of this cumulus or building stage is an updraft which prevails throughout the entire cell. Such updrafts vary from a few feet per second in the early cells to as much as 100 feet per second in mature cells.

b. **Mature Stage.** The beginning of surface rain, with adjacent updrafts and downdrafts, initiates the mature stage. By this time the top of the average cell has attained a height of 25,000 feet or more. As the raindrops begin to fall, the frictional drag between the raindrops and the surrounding air causes the air to begin a downward motion. Since the lapse rate within a thunderstorm cell is greater than the moist adiabatic rate, the descending saturated air soon reaches a level where it is colder than its environment; consequently, its rate of downward motion is accelerated, resulting in a downdraft. (See fig. 5. 2) A short time after the rain starts its initial fall, the updraft reaches its maximum speed. The speed of the updraft increases with altitude. Downdrafts are usually strongest at the middle and lower levels, although the variation in speed from one altitude to another is less than in the case of up-drafts. Downdrafts are not as strong as updrafts; downdraft speeds range from a few feet per second to 40 feet per second or more. Significant downdrafts seldom extend to the top of the cell because, in most cases, only ice crystals and snowflakes are present, and their rate of fall is insufficient to cause appreciable downdrafts. The mature cell, then, generally extends far above 25,000 feet, and the lower levels consist of sharp updrafts and downdrafts adjacent to each other. Large water droplets are encountered suspended in the updrafts and descending with the downdrafts as rain.

c. **Dissipating (Anvil) Stage**. Throughout the life span of the mature cell, more and more air aloft is being dragged down by the falling raindrops. Consequently, the downdraft spreads out to take the place of the dissipating updraft. As this process progresses, the entire lower portion of the cell becomes an area of downdraft. Since this is an unbalanced situation and since the descending motion in the downdraft effects a drying process, the entire structure begins to dissipate. The high winds aloft have now carried the upper section of the cloud into the anvil form, indicating that gradual dissipation is overtaking the storm cell.

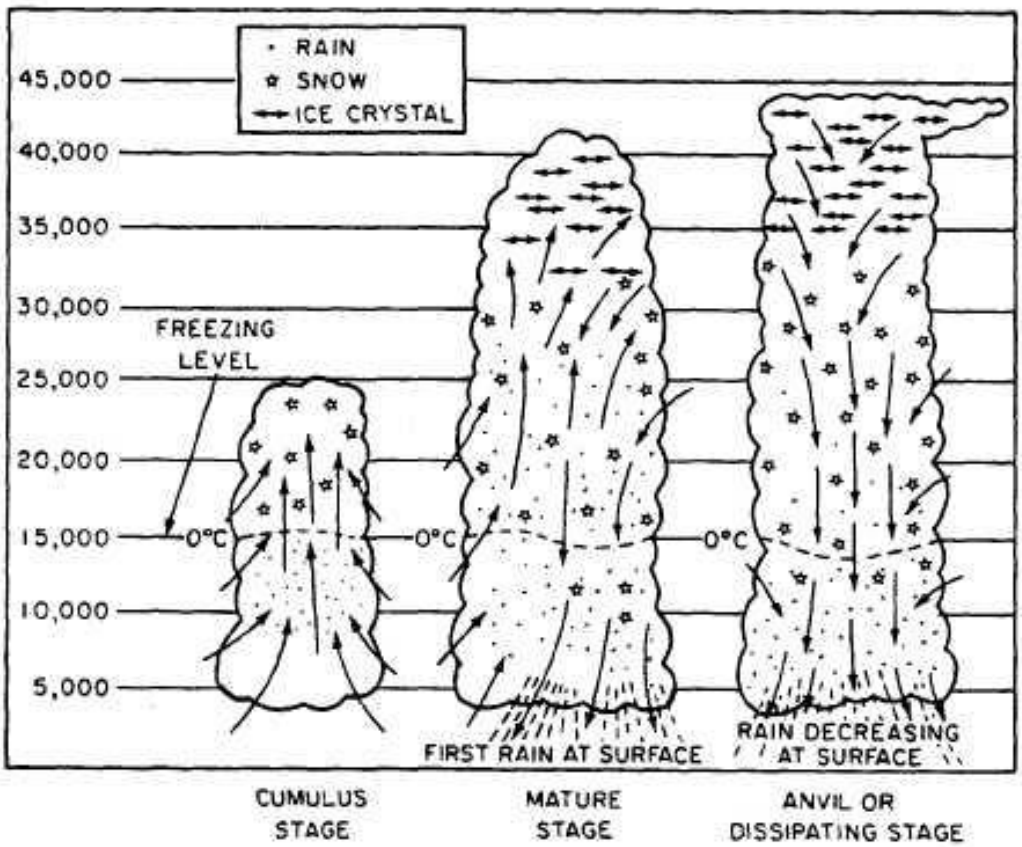


Fig 5.1. Life cycle of a thunderstorm cell

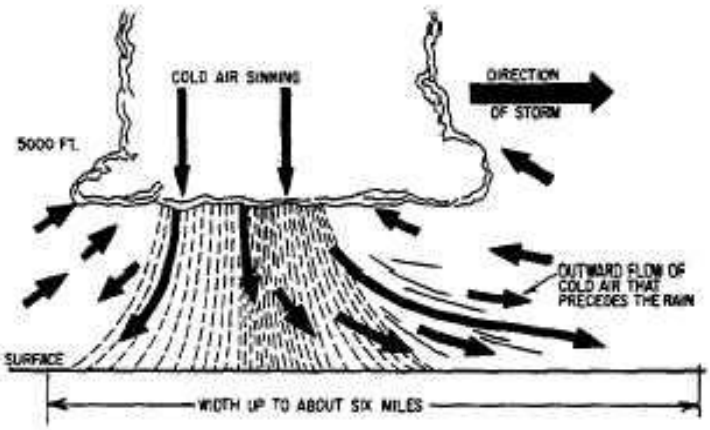


Fig 5.2 Downdraft beneath a thunderstorm cell in the mature stage. Arrows represent wind flow. Dashed lines indicate rainfall

Vertical Development

8. Thunderstorms have been accurately measured as high as 67,000 feet and some severe thunderstorms attain an even greater height. More often the maximum height is from 40,000 to 45,000 feet. In general, air-mass thunderstorms extend to greater heights than do frontal storms. Rising and descending drafts of air are, in effect, the structural bases of the thunderstorm cell. A draft is a large-scale vertical current of air that is continuous over many thousands of feet of altitude. Downdraft speeds are either relatively constant or gradually varying from one altitude to the next. Gusts, on the other hand, are smaller scaled discontinuities associated with the draft proper. A draft maybe compared to a great river flowing at a fairly constant rate, whereas a gust is comparable to an eddy or any other random motion of water within the main current.

Thunderstorm Weather

9. The hydrometers and turbulence of a thunderstorm that we observe and record at the surface are easily recognised. The weather within the thundercloud itself is another story. Visual observations from aircraft are difficult because of the speed with which they pass through the thunderclouds, and man has yet to devise an instrument that will measure all hydrometers in the cloud. Let us look at those forms of precipitation turbulence and icing occurring with and within thunderclouds, as we know them today.

10. **Rain**. Liquid water in a storm may be ascending if encountered in a strong updraft; it may be suspended, seemingly without motion, yet in an extremely heavy concentration; or it may be falling to the ground. Rain, as normally measured by surface instruments, is associated with the downdraft. This does not preclude the possibility of a pilot entering a cloud and being swamped, so to speak, even though rain has not been observed from surface positions. Rain is found in almost every case below the freezing level. In instances in which no rain is encountered, the storm probably has not developed into the mature stage. Statistics show that although heavy rain is generally reported at all levels of a mature storm, the greatest incidence of heavy rain occurs in the middle and lower levels of a storm.

11. **Hail**. Hail, if present, is most often found in the mature stage. Very seldom is it found at more than one or two levels within the same storm. When it is observed, its duration is short. The maximum occurrence is at middle levels for all intensities of hail.

12. **Snow**. The maximum frequency of moderate and heavy snow occurs several thousand feet above the freezing level. Snow, mixed, in many cases, with supercooled rain, may be encountered in updraft areas at all altitudes above the freezing level. This presents a unique icing problem: wet snow packed on the leading edge of the wing of the aircraft resulting in the formation of rime ice.

13. **Turbulence**. There is a definite correlation between turbulence and precipitation. The intensity of associated turbulence, in most cases, varies directly with the intensity of the precipitation.

14. **Icing**. Icing may be encountered at any level where the temperature is below freezing. Both rime and clear ice occur, with rime predominating in the regions of snow and mixed rain and snow. Since the freezing level is also the zone of greatest frequency of heavy turbulence and generally heavy rainfall, this particular altitude appears to be the most hazardous for aircraft.

15. **Surface Wind**. A significant hazard associated with thunderstorm activity is the rapid change in surface wind direction and speed immediately before storm passage. The strong winds at the surface accompanying thunderstorm passage are the result of the horizontal spreading out of downdraft current from within the storm as they approach the surface of Earth. The total wind speed is a result of the downdraft divergence plus the forward velocity of the storm cell. Thus, the speeds at the leading edge, as the storm approaches, are greater than those at the trailing edge. The initial wind surge, as observed at the surface, is known as the First Gust.

16. The speed of the first gust is normally the highest recorded during storm passage, and the direction may vary as much as 180° from the previously prevailing surface wind. First-gust speeds increase to an average of about 16 knots over prevailing speeds, although gusts of over 78 knots (90 mph) have been recorded. The average change of wind direction associated with the first gust is about 40°.

17. In addition to the first gust, other strong, violent, and extremely dangerous downdraft winds are associated with the thunderstorm. These winds are referred to as Downbursts. Downbursts are subdivided into Macrobursts and Microbursts.

a. **Macrobursts.** Macrobursts are larger scale downbursts. Macrobursts can cause widespread damage similar to tornadoes. These damaging winds can last 5 to 20 minutes and reach speeds of 130 knots (150 mph) or more.

b. **Microbursts.** Microbursts are smaller scale downbursts. A microburst can last 2 to 5 minutes and can also reach wind speeds in excess of 130 knots. Microbursts produce dangerous tailwinds or crosswinds and windshear for aircraft and are difficult to observe or forecast.

18. Downbursts are not the same as first gusts. First gusts occur in all convective cells containing showers and are predictable and expected. Downbursts, however, do not occur in all convective cells and thunderstorms.

Electric Field

19. In clear air of normal density a critical electrical field of about 3,000,000 volts per metre is required for breaking down the insulation and for a spark to pass. In cloud it is reduced to about 1,000,000 volts per metre. Such intense electrical fields are built up locally within a Cb cloud. The upper portions of a Cb cloud acquire a positive charge and the lower portions a negative charge. The exact cause of the generation of these charges is not yet fully understood. However, the electrical field is sufficiently strong to permit a lightning discharge. Discharges may take place within the same cloud, from one cloud to another and less frequently from cloud to the earth. The lightning discharge is in the form of stepped strokes, 1 to 4 km long with a channel diameter of 1 to 10 m.

20. The air through which the discharge has passed is rendered white hot and expands suddenly, giving the clap of thunder. If the lightning stroke is long, the thunder may be in the form of peals as the sound from different parts of it takes longer time for travel and reach the ground.

Movement of Thunderstorms

21. Although large thunderstorms modify the airflow in their immediate neighbourhood, most thunderstorms drift slowly with the wind prevailing in the layers in which they are embedded. As a first approximation, the direction of movement may be deduced from the prevailing winds at 3 to 5 kilometers above sea level. The speed of movement is less than the speed of the wind at these levels. When the winds aloft are weak or variable, thunderstorms show little movement. In such cases the shower from the cloud is confined to a limited area, resulting at times in exceptionally heavy falls of short duration, popularly known as cloudburst. If the thunderstorm has appreciable movement in the rain stage, the shower gets distributed over a larger belt.

Regeneration of Thunderstorms

22. The cold downdraught from one thunderstorm can activate a nearby cumulus cloud and cause the latter to build-up rapidly into a cumulonimbus. This process is known as regeneration. Regeneration can happen successively, giving the impression of movement of the same Cb or thunderstorm cloud. In reality it is a chain reaction by means of which thunderstorms occur in quick succession over an extended belt covering some hundreds of kilometers. The violent summer thunderstorms known as Norwesters or Kalbaisakhis which affect Bengal and Bihar owe their origin to regeneration from the primary thunderstorms over the Chotanagpur hills which build-up in the afternoon.

Flying Hazards of Thunderstorms

23. Thunderstorms pose a variety of hazards to aircraft in flight. The more important ones are listed below :

a. **Squall, Windshear and Microbursts**. Squall means sudden increase in wind speed & possibly change of direction. Squalls from thunderstorm wrecked mariners for centuries before aviation commenced. Flying has led to a continuing study of such airflow and yet its pattern cannot be precisely forecast or measured despite an intense study setup by ICAO in 1977. Major airports issue windshear alerting messages and pilots are expected to report any windshear experienced. Sometimes the normal variation of wind speed with height becomes greatly accentuated, perhaps decreasing from 40 knots or more at 1000-2000 ft to less than 10 knots near the surface. A varying cross wind component complicates the matter. The gust front of self-propagating storms may be 15 to 20 km ahead of the storm, which generated it. It may be marked by newly-forming cumuliform clouds or by a line of duststorms in desert countries, but equally it may be quite invisible. Microbursts are particularly intense and localised – probably not more than half a km across – and especially hazardous when they are overhead or near the runway in use. Again, if on the approach in thuderstorm conditions it is found that abnormal levels of power are necessary to maintain the airspeed, attitude and glideslope, then wind shear go-around should be initiated. The US Federal Aviation Administration (FAA) initiated the ‘low level windshear alert system (LLWAS)’ at major US airports where windshear problems are a frequent hazard. LLWAS detects not only mocroburst but also any hazardous windshear.

b. **Heavy Showers**. These may reduce ground visibility to very low values for short durations.

c. **Fractostratus Clouds**. These clouds are actually ragged fragments of the base of the Cb; the fragmentation occurs due to turbulence caused by the squall in the region between the ground and the cloud base. At times the base of these fragments may be as low as 500 ft even over comparatively plain ground.

d. **Poor In-flight Visibility**. In the interior of a Cb cloud the in-flight visibility is practically nil. Cb clouds have a larger amount of water content per unit volume as compared to other clouds, both the concentration as well as the size of drops being higher.

e. **Draughts.** Within the cloud the strong up-draughts and downdraughts may cause sudden and large variations in the altitude of an aircraft. In general the updraughts are stronger than the downdraughts, both being predominant in the middle and upper parts of the cloud. In the Thunderstorm project in USA the traverses were made at cruising speed of about 160 kt. The highest vertical displacement encountered was 6,000 ft in the upward direction and 1,400 ft in the downward direction, both at a flight level of about 25,000 ft. The storms traversed were of the sub-tropical variety. It is possible that in summer thunderstorms in the tropics the displacements may be of a somewhat higher order.

f. **Gusts**. Near the borders of the updraughts and downdraughts the large difference in the velocities creates friction. This friction is responsible for vigorous eddies which, once formed, travel horizontally with the wind and also in the vertical direction with the draughts. They are mainly responsible for the extreme bumpiness within the cloud, although the effect of the draughts is also mixed up. The severity of bumpiness increases with height upto the middle part of the cloud and remains constant upto about 10,000 ft below the top. Further aloft it decreases rapidly. Bumpiness is much less in the (cloud-filled) lanes between two adjacent cells of the same storm.

g. **Ice Accretion**. The formation of ice on the parts of an aircraft in flight is known as “ice accretion” or merely as ‘icing”. Icing can affect the aerodynamics of the aircraft even upto the stage of loss of control. It is thus a serious hazard requiring careful study of the different types of icing, the meteorological conditions under which they occur and the techniques of flying which avoid or minimise the risk of icing. Although many types of aircraft are fitted with de-icing equipment, this provides only partial protection and its successful operation is facilitated by knowledge of the type of icing and the possible rate of accumulation.

h. **Hail.** Aircraft which encounter hail in flight, can suffer serious damage. There have been instances wherein airframes have been badly dented, windscreens holed. Perspex astrodomes shattered, de-icing boots ripped off and radiator found badly bent. Other things being equal the damage is more in the case of faster aircraft.

j. **Lightning.** Apart from distracting aircrew and temporarily dazzling their eyes lightning may interfere with radio communication and may seriously affect the magnetic compass performance. The aircraft itself does not, however, get damaged due to lightning strike.

Avoidance and Flying Techniques.

24. The take-off of an aircraft should be postponed while there is any risk of flying into the area of an active thunderstorm cell before completing the initial climb. Similarly on arriving in such conditions the approach and landing should, if possible, be delayed or a diversion carried out under advice from ATC.

25. The flight planning should aim at selecting a route and flight level, which will have the least possible traverse in the dangerous parts of thunderstorms.

26. Flying through thunderstorms should be avoided as far as possible. Sometimes, however, it may be necessary to penetrate them. This is always dangerous, but by employing certain techniques it is possible to minimise the risks involved.

27. The following procedures are recommended:-

a. Select a level below the freezing level or above 25,000 ft if the aircraft is capable of this. The height band between the freezing level and 25,000 ft is the worst in regard to bumpiness, icing and hail formation.

b. Fasten safety belts and secure any loose articles.

c. Select safe speed for penetration of turbulent zone so that stalling does not occur due to pronounced gusts. For most types of aircraft speeds of penetration of turbulent zones have been laid down.

d. Check all instruments.

e. When within the storm maintain constant heading.

f. Disregard the indications of the ASI. This gives low readings as the Pitot tube gets partially blocked due to heavy rain. Radar is one of the best instrumental aids to thunderstorm flying. The area delineated by the radar echo indicates the area of maximum water content which also coincides with maximum turbulence. In the case of line squalls, the best possible course for penetration, or the least active areas can be seen in the radar echo picture. Weather radar pictures on the ground is a good aid not only to the forecaster, but to the aircrew also. When an aircraft is not fitted with an airborne weather radar, aircrew in flight can get while in flight, the latest weather radar report from the nearest Weather Radar Centre.

Tornado

28. Tornadoes occur with severe thunderstorms. They are violent circular whirlpool of air shaped like on inverted funnel of tube hanging from a cumulonimbus cloud. The air motion is characterized by a cyclonic upward spiral, causing rapid expansion, cooling, and condensation, which forms the dark cloud of the tornado funnel. The width of the storm at the ground level average about 300 yards although it may vary from 20 to 30 yards to a mile. Their paths over the ground are often only a few miles long, and they move at a speed of 30 to 55 miles per hour. It often travels in an erratic Skipping path, so that the tapering end of the funnel may hop over one area and descent to wreak further havoc over another area. Although highest tornado wind speed have never been measured precisely, property damage and other effects indicate that they may exceed 300 miles per hour. Their very low pressure gives them great suction having estimated vertical velocity upto 200 miles per hour. The great destructiveness of tornadoes is caused by these two types of high winds. They sometimes lift very heavy objects off the ground. In the Northern hemisphere, tornadoes generally occur in the warm air in the southern section of a cyclone or extern and southeastern portion of a well developed trough and more often with lines of thunderstorms. They are more common in spring and early summer.

Briefing and Thunderstorms

29. Thunderstorms, at times, create acute flight safety hazards for pilots during local flying or route flying including landing and take off. The accuracy of for forecast and briefing by meteorological forecasters, will have definite contribution towards flight safety and will usher confidence in the mind of the aircrews to face the worst situation if need be, without a failure. Remember this; “so far as flying is concerned, no thunderstorm should be regarded as light. It is better to get to one’s destination late than not at all; 180° turns have saved many lives”. Both the meteorological forecasters who is carrying out the briefing and the pilots who is going to undertake the flight, should keep in mind the following recommended hints when thunderstorms are expected or in existence :

a. Pilots, particularly those flying light aircraft, should avoid thunderstorms.

b. Pilots, even those flying large aircraft, should avoid thunderstorms because the safety of the aircraft is not necessarily related to its size.

c. Pilot when flying around thunderstorms should not venture closer than 5 miles to any mature visible storm cloud with over hanging anvils because of the possibility of encountering hail.

d. Pilots should not attempt to fly under orographic thunderstorms even if the area on the other side of the mountains or hills can be seen. Wind flow sufficiently strong to give necessary lifting action to produce the thunderstorms will probably also create dangerous turbulence between the peaks. This turbulence, together with downdrafts, could result in disaster.

e. Pilots should avoid flying under any thunderstorms if possible, because of updrafts under growing storms and downdrafts which can be quite strong in the rain core.

f. For a decision whether or not to risk engulfment by thunderstorms in an area where thunderstorms are known to exist should be based upon the fol :

(1) Experience in flying through thunderstorms.

(2) The type of aircraft and the amount of stress it can stand.

(3) The vertical extent and the physical appearance of the thunderstorm clouds.

(4) Whether or not the aircraft is radar equipped.

g. Any thunderstorm indicating echo tops of 35, 000 feet or higher on ground-based radar should be avoided, regardless of other considerations, because of the extreme severity of hazards associated with them.

h. Radar, whether airborne or ground-based, does not eliminate the hazards of the thunderstorms. It merely helps to locate the most severe conditions at the instant. Since the radar scope indicates only precipitation areas within thunderstorms, hazards can be encountered even in “Soft spots”. Thunderstorms having frequent, vivid lightning discharges usually are dangerous. The airborne weather radar is to be used as a thunderstorm avoidance rather than as a thunderstorm penetration tool.

j. A thunderstorm cloud which is “boiling up” or growing rapidly generally contains more hazards than one which is showing little change. The first 15 minutes after the towering cumulus cloud in its upward growth has penetrated the freezing level is the time in the thunderstorm’s life span when it normally offers its greatest hazards.

k. For a decision to undertake a flight through thunderstorms the following practices will prove beneficial :

(1) Cockpit lights are put to highest intensity. Dark glasses are often used to avoid temporary blindness from lightning flashes.

(2) Penetration altitudes are to be established. Altitudes from the freezing level upward to – 10^0C (roughly 5,000 feet above the freezing level) are avoided. The “softest” altitude in a thunderstorm cloud is usually between lower 4,000 and 6,000 feet. Four thousand feet above the highest terrain in the area should be the minimum penetration altitude.

(3) Airspeed is changed to the manufacturer’s recommenced one for turbulent air penetration. This reduces the structural stresses resulting from turbulence and is best for maintaining control.

(4) Power setting is changed to establish the reduced airspeed before entering the storm.

(5) Power setting is kept constant in the storm because airspeed indicator given false readings in vertical drafts or in heavy rain.

(6) Altitude is kept straight and level. Unnecessary manoeuvering is avoided.

(7) If the auto-pilot is used the “altitude hold mode” is kept in the off position so that the auto-pilot does not place the aircraft in a “Pitch down” altitude to compensate for an updraft and in a “Pitch up” for “nose up” altitude to compensate for a downdraft.

(8) The quickest way through is often straight ahead; and turns should be restricted to small alterations designed to achieve the smoothest path through the storm.

Thunderstorm Forecast

30. Forecasting of occurrence of thunderstorm, like all other meteorological elements affecting aircraft operations, requires the forecaster to be well conversant with the meteorological conditions which contribute for its formation and development. Forecasting needs a continuity in analysis of changes taking place in the meteorological conditions which are ultimately responsible for occurrence of different types of weather. Once the forecasters are able to correlate these changes with the generation of a particular weather, it is not difficult for him to forecast the same, whether it is for local air field or for the route. He is to know fully the causes first before he makes an attempt to forecast a thunderstorm. Certain conditions, which are creating particular situations for the occurrence of thunderstorm of different intensities have been put together for the forecaster’s easy references.

31. **General Situation & Thunderstorm.** The forecasters must be conversant with the following :

a. Position and movement of fronts and troughs.

b. Source, trajectory and modification of air masses.

c. The general wind flow at all levels.

d. Terrain and peculiarities of the area.

e. Prior local weather conditions and tendencies.

32. **Radar Observation and Thunderstorm.** The forecasters must utilize the radar observations in the following manner :

a. Comparison with radar and other reports from surrounding station.

b. Detection of approaching squall lines which give lines of echoes and are rarely smooth.

c. Determination of the position of maximum intensity echoes with respect to the station – if echoes are upwind at middle levels, the station is probably in the path of the most severe storms.

d. Realization that cells which move more rapidly than other are affected by the maximum down rush velocities. When two cells merge, intensification results and severe weather phenomena may occur.

e. Careful interpretation of radar echoes, coupled with micro analysis of surface observations, permits identification of the intersections of lines of activity which are essentially micro cyclones. When a thunderstorm cell moves into them, serve as triggering mechanisms for tornadic activity.

f. Detection of dark spots within the echoes of severe storms Altering the antenna angle will show the dark spot to extend to great heights, and tilt into upper level winds. These dark spots occur with the most severe storms and tornadic winds occur at the surface near the base of the spot in many instances.

33. **Local Observations Thunderstorm.** There are close correlations between the behaviour of clouds, pressure, temperature/dew point, winds and the occurrence of thunderstorms. The forecasters must analyse the behaviour of these elements and use them as tools for forecasting.

a. **Behaviour of Clouds.**

(1) In some instances, lower clouds will decrease in amount with the approach of a severe storm, even an hour or more in advance.

(2) Any advance of middle clouds should be observed suspiciously for evidences of mamatus. The path of severe storms is generally immediately south of the lowest axis of mamatus formation. If the mamatus is below 14,000 feet, the barometer may be observed to “pump”, indicating turbulent conditions aloft in the immediate area. This is a sign that a rapidly developing severe storm is nearby.

(3) Alto-cumulus castellatus resembling a row of turrets with well defined flat base, and Alto-cumulusfloccus --- broken irregular tufts having cumuliform tops but ragged, eroding bases (Figure 3 a and b) are cumyuliform clouds at medium levels. These clouds indicate instability in the middle troposphere. Their appearance is often and indicator of probable development of CB and thunderstorms in the vicinity of the observing station. When first forming they resemble patches of Altocumulus lenticular. However, within minutes cumuliform tops sprout upwards. These clouds may be found at several levels simultaneously and also coexisting with stratiform clouds both Ac and As; the whole sky then takes on a chaotic appearance.

b. **Behaviour of Pressure**. Increasing instability of the air structure is generally reflected in erratic behaviour of the barometer. Without contradicting the

concept of falling pressure with approach of a storm, it is sometimes noted that, with the advance of a fast-moving squall line or bubble, the rate of pressure fall decelerate, or even rise lightly, prior to a more rapid fall. If the barometer is falling very rapidly it may be concluded that strong vertical currents exist.

c. **Behaviour of Temperature and Dew – Point**. An increased in temperature and without a decrease in dew-point over the normal diurnal change shows that moisture is being added to the lower levels of the local air column. Severe storms are likely to develop in the area having maximum moisture. If a falling barometer accelerates its fall simultaneously with an increase in temperature and dew-point, the storm warning is more certain.

d. **Behaviour of Winds**. The surface winds usually decrease in speed and back with the approach or a severe storm; however, in most cases, this indication occurs too late to be of much help. If the surface wind increases in speed and shifts from southerly to westerly, the disturbance is probably developing north or northeast of the station. If the surface wind increases in speed and shifts from easterly to southerly, the storm is approaching. If the winds above 8,000 feet increase rapidly in comparison to previous observations, the site is probably in the area of destructive thunderstorm winds. Also, if the winds above 8,000 feet show a sharp decreases in speed, the probability of tornadoes is increased, horizontal wind shear and/ or strong vertical currents developing.

34. **Instability & Thunderstorm.** One of the contributory situations is the presence of instability in the atmosphere which is indicated as “Stability Indix”. Most widely used is, perhaps, the instability index proposed by Showalter. This index is the difference between the observed temperature at 500 mb and the temperature that a parcel of air at 850 mb would acquire if it were lifted dry-adiabatically to its lifting condensation level and thence wet-adiabatically to 500 mb. Given below is the Showalter Index and the expected weather

INDEX			**WEATHER**
(+3, + and higher)	…	…	No thunderstorm
(+ 2, +1, and 0)	…	…	Showers, Light Thunder may occur.
(-1, -2 and –3)	…	…	Thunderstorm of Moderate intensity.
(-4, -5 and –6)	…	…	Thunderstorm of Severe Intensity
(-7, -8 and lower)	…	…	Tornadoes occur.

It should be noted that the stability index is unrepresentative when a frontal surface or an inversion is present between 850 and 500 mb.

Conclusion

37. The hazardous weather, which is most detrimental to flight safety, is the thunderstorm having unusual characteristics, all of which may effect aircraft operations at the same instant. To impart some knowledge in respect of both theoretical and practical aspect of a thunderstorm forecasting and keeping in view the safe aircraft operations, an effort has been made to accumulate materials for this paper from different meteorological text books, manuals and research works.

38. Assuming that the general air structure is critical for thunderstorm and tornado development, correlation of the general synoptic situation and radar observations with a microanalysis of the local area and fluctuations of one's own instruments will provide indications of the activity to be expected locally at least an hour in advance. While it is often difficult, the problem of local storm warnings is by no means hopeless for the persistent and conscientious forecasters. Forecasters' knowledge, experience, sincere and devoted approach to the problem of forecasting a thunderstorm will definitely produce accurate results and hence arouse confidence in the mind of the users. Calculations of peak gust and hail size can be studied from part I phase 9 also.

Self-Assessed Exercise on Task-9

Q1. What are the requirements of thunderstorm formation ?

Q2. What is the level of free convection ?

Q3. What agents are responsible for lifting the air ?

Q4. What is the life cycle of convective cell ?

Q5. Do you get any precipitation during cumulus stage of a thunder cloud ? If the answer is 'No', explain why not ?

Q6. What are the characteristics of a dissipating thunder cloud ?

Q7. How the existence of a thunderstorm cloud are indicated ?

Q8. How different type thunderstorms are named ?

Q9. When a thunderstorm cloud starts giving lightning ?

Q10. How the frequency of lighting and the rainfall are related ?

Q11. How the name "Tornado" is characterized ?

Q12. How occurrences of thunderstorms and the behaviour of surface wind is correlated?

Q13. What are the showaller stability indices for thunderstorms of moderate intensity and of tornadoes ?

Q14. Discuss the Meteorological conditions which are required to be considered while forecasting the formation and development of thunderstorm.

Answers to Self Assessed Exercise

A1. Unstable air, initial lifting or trigger action and high moisture.

A2. This is a level at which the conditional unstable air, when lifted, becomes warmer than its surroundings.

A3. Strong surface heating, cold or cold occluded front, mountainous terrain and the vertical motions resulting from air coming together from two or more directions (convergence).

A4. 20 minutes to 1½ hours.

A5. Usually no pricipitiation falls during this stage because the drops and ice particles are carried upward and are more or less suspended in the ascending air current.

A6. a. Top of the cloud generally takes the sharp of an anvil.

b. Lower portion of the cloud gives straliform look.

c. Down drops are generally predominant.

A7. Para 15 and 16.

A8. Thunderstorms are classified according to the lifting agent involved in the process of their formation and development.

A9. Lifting shorts when its top reaches the level of –20°C.

A10. The period of the greatest frequency of lighting is often followed immediately by the period heaviest rainfall.

A11. This is a violent circular wheilpool of air shaped like an invected funnel of tuble hanging from a cumulonimbus cloud. The air motion is characterized by a cyclonic upward spiral, causing rapid expansion, cooling, and condensation, which forms the dark cloud of tornado funnel. The width of the storm at the ground level may vary from 20 to 30 yards to a mile. It moves at 30 to 50 miles per hour. Horizontal wind speed may exceed 300 miles per hour, where as the vertical wind speed may be as high as 200 miles per hours.

A12. Para 33d.

A13. S.I. (-1,-2,-3) for moderate thunderstorms.
S.I.(-7,-8 and lower) for tornadoes.

TASK – 10 : WESTERN DISTURBANCE

Objectives

After reading the chapter you should be able to :

- Assimilate Characteristics of Western Disturbance.
- Understand mechanism of formation of Western Disturbance.
- Understand structure, sequence of clouding and Aviation Weather Hazards associated with Western Disturbance.

Introduction

1. On the global charts, one can see eastward moving troughs in the zonal westerlies. These upper air troughs in the sub-tropical westerlies often extend down to the lower troposphere of the northern latitudes of the Indian sub-continent, particularly during the winter season. These often give rise to closed cyclonic circulations on the sea level charts. In Indian meteorological terminology, these have come to be known as western disturbances. Bangladesh and Northern India normally experience fine to fair weather during the winter months. But this phenomenon is disturbed by disturbance from the west associated with overcast to showery weather. These disturbances form due to the influence of extra-tropical cyclones moving from west to east north of 35^0N. The track of these disturbances shifts southwards with the passage of winter season. On average 3 to 5 disturbances affect the area in each month accompanied by cloudy to rainy/showery weather and with cold wave condition. The most affected areas are Jammu and Kashmir, Punjab, Hariana, Himachal Pradesh etc, which get most of their rainfall from such disturbance.

Definition

2. Western Disturbance (WD) is an eastward moving system seen as an upper air trough in sub-tropical westerly after extending to lower troposphere of North Indian latitude during winter. It is defined as a low or trough at the surface or the upper air in the region of westerly wind regime (North of latitude 20^0N) usually giving rise to changes in the pressure, wind and temperature fields and accompanied by clouding with or without precipitation. The features at 500 mb level and above are described as trough in the easterlies. The low that develops further south of WD over Rajasthan or Madhya Padesh is described as induced low. When two or more isobars at a 2 mb interval can be drawn on the sea level chart, the disturbance is described as Western Depression. When WD moves across India and neighbouring countries, the following features are observed:

a. There is general strengthening of the westerly flow.

b. Westerly maximum over North India may strengthen and often reach values of 130-150 kts.

3. Sometimes, Western Disturbance is not found at surface chart but can easily be detected as convergence at 0.9 km or 1.5 km.

Types of WDs

4. WDs are classified into the following two types:

a. **Active WD.** They originate mostly from cold fronts or as induced low from strong upper air trough at 300 mb. These active WDs are responsible for giving rain/thunderstorm and even sometimes hails and snowfall over Jammu and Kashmir and Himachal Pradesh.

b. **Inactive WD**. These type of WDs originate from warm front or they appear as remnants of high latitude cyclone or seen as trailing end of extra-tropical cyclone that happen to move southward without fresh moisture feeding from the Arabian sea. This type gives cloudy to overcast weather and die out gradually. Features associated with active and inactive WDs are shown in the Table 1.

Table 8.1: Features associated with active and inactive WD.

Parameter	Active WD	Inactive WD
Westerly Trough at 300 mb	Not seen in the vicinity, accentuation takes place if trough comes within 5-7^0 of longitude in the rear of WD. The WD moves rapidly if the trough overtakes it. WD normally gives precipitation over Indian region.	Superimposed over the longitude low level position of WD or slightly (1-2^0 long) ahead of it. WD moves away rapidly with the trough without causing any precipitation over the northwest Indian regions.
Position of Jet Maxima.	Shifts southwards and also increase in speed is observed.	No appreciable change in speed or position.
Temperature difference between Jodhpur and Delhi at 300mb.	Shows an increase in the day of advent of a WD.	Shows a decrease over the day of advent of a WD.
Anticyclone at 700mb level over Gujrat or along West Coast.	Exhibits gradual southward shift with the approach of a WD.	No appreciable shift is observed.

Origin and Movement

5. The westerly troughs or waves and the associated low level cyclonic systems often originate in the West Atlantic or Mediterranean regions. These are of the extra-tropical frontal type of cyclones. Secondaries develop over the Persian Gulf. The mountainous terrain of Iran, Afghanistan and Pakistan modifies and often obliterates the frontal characteristics of the systems at the lower levels. The movement is east or northeastwards.

6. The WDs originated from eastern Mediterranean Sea move eastward and become weak disturbance over land areas with warm core centre. They meet fresh cold surge from the southern part of Russia. This cold surge carries moisture from Caspian Sea as shown in Figure 2. Thus cold and moist air combines with warm core lows to intensify further. These surface lows get the influence of strong westerly jet and keep on moving west. Some of the WDs move further south and draws in fresh moisture from Arabian Sea. Such warm moisture feeding makes the WDs strong enough to move to Sub-Himalyan West

Bengal and adjacent Bihar. Again the westerly trough at 300 mb helps the surface lows to intensify. Moisture from Bay of Bengal meet such lows to enable them to give weather like overcast sky, light to moderate rain/thundershowers.

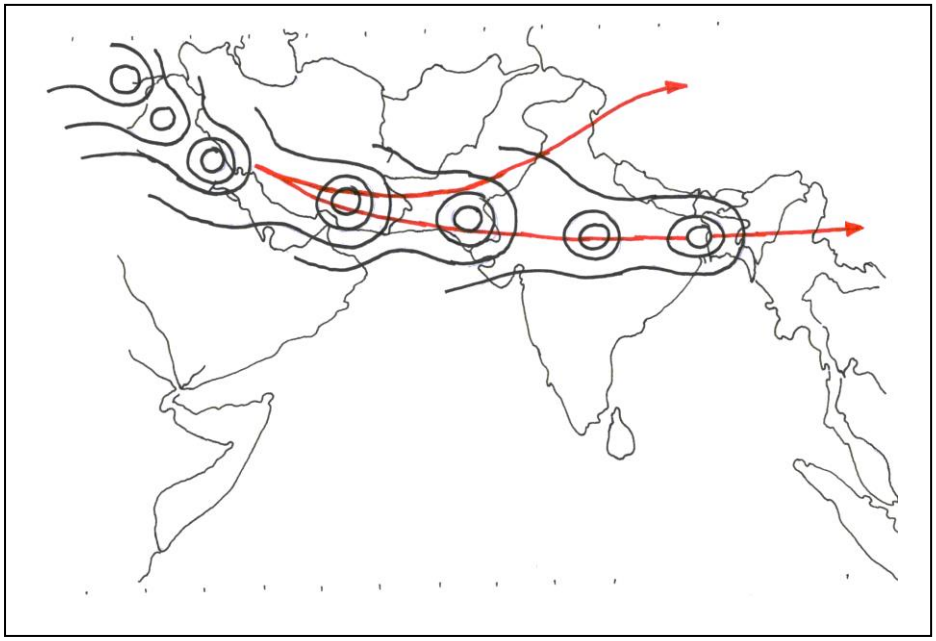


Fig 8.1: Movement of WD.

Identification of WDs

7. WDs are generally identified over the surface charts as closed isobars provided sufficient surface observations are available. It is generally difficult to trace the WD with scanty data. However, satellite cloud picture helps to identify the presence of WDs by patterns of cloud bands. The cloud patterns are often amorphous (undefined) in structure and cover a wide area. The active WDs show longitudinal or north-south oriented cloud bands. Some of them show clear-cut vortices in the cloud pictures. Convergence of wind in the pilot charts also indicates the presence of WD.

Features of WDs

8. The following features are observed either alone or in combination with other:

a. Appearance of low with one or two closed isobars on the surface or appearance of convergence in lower troposphere.

b. Change of surface winds from north/northwestly over Rajasthan and along the belt of 22^0N to 25^0N.

c. Appearance of strong winter Jet or Sub-tropical Jet with core reaching upto 130 -150 kts over northern part of India at or above 24,000 feet.

d. Appearance of strong westerly trough at 300 mb due to passage of extra-tropical cyclone.

Influence of Mountain Ranges

9. In the Western Himalayas, the characteristics of the extra-tropical depressions undergo considerable modification due to the mountain ranges. The speed of the fronts, the slope of the frontal surfaces and the associated weather phenomena, all get modified. The factors involved in the extent of modification are the height and horizontal extent of the ranges, the angle of the front in respect to the range and the type of front. The main effects are:

a. Widening of the area of precipitation on the windward side.

b. Prolongation of the duration and intensification of the preci-pitation on the windward side.

c. Narrowing of the area of precipitation and dissolution of the front on the leeward side.

d. Retardation of the fronts and occasionally formation of a fresh low pressure system behind the cold front.

e. Speeding up of occlusion process.

Characteristics

10. In the Indian region, these disturbances are mostly in an occluded stage, though sometimes well-marked lows or depressions with the characteristic warm front, warm sector, cold front and the associated weather moving from west to east can be clearly discerned. The movement can be clearly traced on the charts from West Asia onwards. Sometimes, no surface low can be drawn on the charts but the weather system can be seen progressing eastward. However, in such cases the progression of well-defined upper air waves can be seen.

11. Orography plays a major part in the breaking up, stagnation and intensification of these disturbances. Some of the disturbances split into two in the hilly regions of Pakistan, one moving north and the other eastwards. In the Western Himalayas over J & K and Himachal Pradesh areas, some of the disturbances break up and dissipate while some take a more easterly course and travel as far east as Assam and North Burma.

12. A depression located in the coastal region, west of the Nile may move east-north-east and affect North Iraq Iran, Afghanistan, Northern Pakistan and Kashmir. Many of the depressions moving over Iraq separate into two near the Iraq-Iran area due to the obstruction of the high terrain there. The northern cell moves northeast over Iran and the southern cell travels east along the Persian Gulf. The southern cells intensify over the Persian Gulf area though they weaken further east since the circulation in the lower levels is cut off over the highland regions of Iran, Baluchistan and Afghanistan. Hence, by the time they cross over to India, they are mostly weak and irregular disturbances.

13. Over northwest India they revive and intensify again with moisture fed from the Arabian Sea. Depressions which take a more southerly course often enter India along the Mekran coast, sometime developing warm and cold fronts, the contrasting air masses being the moist warm Arabian Sea current and the cold, dry north/northwest winds from the temperate regions.

14. The nearly closed-in character of the Punjab with high hills to the north, northeast and west and the open terrain to the east and south explains the tendency of western disturbances to slow down and intensify in this area.

Sequence of Weather

15. WDs generally move about 5^0 to 6^0 per day, but some of the WDs of active nature move very fast. The steadily moving WDs show a fine sequence of weather. These are as follows:

a. Synoptic chart analysis shows a rise of temperature and fall of pressure.

b. Visual observation or satellite picture observation shows the appearance of high clouds followed by medium and low clouds.

c. Light to moderate precipitation with warm front oriented disturbances followed by quick clearance of clouds.

d. Moderate thundershowers or snowfall from cold front oriented WDs.

e. Rise of pressure and fall of temperature after the clearance of WD.

f. Formation of radiation fog in the morning hours one or two days after the clearance of rain bearing WDs.

Conclusion

16. WDs originated from clod fronts or their occlusions; give rain/thundershowers associated with snowfall over Kashmir, Punjab and Himachal Pradesh areas. During such periods, flying becomes impossible there. Again WDs from warm fronts or their occlusions are associated with cloudy/overcast weather and sometimes with light to moderate rain/thunderstorm. This causes poor visibility and hence flying remains suspended. Over Bangladesh and adjoining areas, the WDs give light to moderate shower for 1 to 2 days. This does not allow military flying to continue over the area but civil flying can continue with certain restrictions. After the passage of WDs, radiation fog forms in the morning periods, causing flying suspended till the foggy weather is ove

TASK – 11 : TROPICAL CYCLONE

Objectives

After reading the chapter you should be able to :

- Assimilate seasonal genesis parameters required for the formation of Tropical Cyclone.
- Understand parameters necessary for the formation of a tropical Cyclone.
- Understand physical process and Life Cycle of a tropical Cyclone.

Seasonal Genesis Parameters

1. The seasonal tropical cyclone frequency can be directly related to a combination of six physical parameters known as primary genesis parameters.

2. **Primary Genesis Parameters**.

a. Low-Level Relative Vorticity (Zr)

b. Coriolis Parameter (f)

c. The Inverse of the Vertical Shear (Sz) of The Horizontal Wind between the lower & upper Troposphere, or 1/Sz.

d. A Sea Surface Factor (STF) or the Sea Temperature Above 79°F between the Surface and 200 feet ocean depth.

e. Vertical Gradient of θe between the surface and 500 hPa level ($\partial\theta e/\partial p$).

f. Mid-troposphere relative humidity (RH).

Seasonal Genesis Parameter (SGP) $\propto$ (VP) x (CP) x (VSP) x (STF) x (MSP) x (RHP)

where,

VP = Vorticity Parameter (Zr)

CP = Coriolis Parameters (f)

VSP = Vertical Shear Parameter (1/Sz)

STF = Sea Temperature Factor (STF)

MSP = Moist Stability Parameter ($\partial\theta e/\partial p$)

RHP = Relative Humidity Parameter (RHP)

SGP $\propto$ (Zr) x (f) x (1/Sz) x (STF) x ($\partial\theta e/\partial p$) x (RH)

3. **Units of Parameters**.

a. **Vorticity Parameter**.

VP = (Zr + 5), [950 hPa]
Where, Zr is determined in units oF 10^{-6} Sec $^{-1}$

b. **Coriolis Parameters**.

CP = f or $2\Omega \sin \phi$
Where, Ω is the rotation rate of the earth and ϕ denotes latitude

c. **Vertical Shear Parameter**.

VSP = 1/(Sz+ 5), [between 950 and 200 hPa]
Where, Sz = $|\partial Vh/\partial p|$ is determined in units of m/sec per 750 hPa.

d. **Sea Temperature Factor**.

STF = $(T_{SST} – 79°F)/\{1+(T_{SST} – T_{200FT})\}$

e. **Moist Stability Parameter**.

MSP = $(\partial\theta e/\partial p) + 5$
Where, $\partial\theta e/\partial p$ is in the units of °K per 500 hPa.

f. **Relative Humidity Parameter**.

RHP = RH [between 700 and 500 hPa] which varies from 0.0 – 0.5

RHP = 0.5 for mean 700 – 500 hPa humidity ≥70%,
0.4 For mean humidity of 55 – 60%,
0.2 For humidity of 46 – 54%,
0.05 For humidity of 40 – 46% and
0.0 For values < 40%.

4. **Dynamic & Thermal Potential**

SGP = Dynamical Potential X Thermal Potential
Units 1.5 x 10^{-10} SEC^{-2} °K per m/sec

Dynamical Potential = (Zr) x (f) x (1/Sz)
Units 10^{-12} sec^{-2} per m/sec per 750 hPa

Thermal Potential = (STF) x $(\partial\theta e/\partial p)$ x (RHP)
Units °K per 500 hPa

Parameters Necessary for Tropical Cyclone Formation

5. It is not yet known what are the necessary and sufficient conditions for the formation of Tropical Cyclone. However, it has been possible to qualitatively define a number of necessary conditions (Gray).

6. **Low level Relative Vorticity** (ζ). It has been found that TCs form only in regions of positive low level vorticity. The larger the low level vorticity the greater the potential of Tropical Cyclogenesis. It has also been found that the low level vorticity in developing cloud clusters is twice that in non-developing disturbances. The variations of low level vorticity was one of the few parameters showing a strong correlation with intensification.

7. **Coriolis Parameter**. Cyclones do not form within 4-5° of the equator. The influence of the earth's rotation would thus appear to be of primary importance. Fig 4 shows the frequency of location of initial formation of TC by latitude. Only 13% of the TCs form poleward of 20° latitude and nearly all those occur in the Northern Hemisphere (NH). Genesis does not occur poleward of 22° latitude in the Southern Hemisphere. In the NH genesis occurs at latitude as high as 35°N. Globally 65% of the TCs has genesis in the latitude belt 10° and 20°.

8. **Vertical Wind Shear**. Tropical cyclones form under conditions of minimum vertical shear of the horizontal wind between lower and upper troposphere. The individual level cluster winds *(V)* may be quite different than that of the cluster or disturbance velocity *(Vd).* The magnitude of *(V – Vd)* determine the extent to which sensible heat and water vapour are advected, or ventilated, through the developing tropical disturbance. Ventilation in the upper level tends to remove sensible heat, while low level ventilation primarily drives the cluster increasing cooling due to re-evaporation. Both these processes inhibit the enthalpy building up required for genesis. If ventilation is small, an upper level enthalpy increase can occur. This will lead to gradual surface pressure decrease and cyclone formation. Once the cyclone has formed, the environmental motion relative to the system becomes very small and further enthalpy and moisture gain can be readily accomplished.

9. **A Warm Sea Surface Temperature (SST)**. It has been found that Seas temperature above 26°C to a depth of 60m is favourable for the development of Tropical Cyclone. The influence of hurricane on the ocean underneath is usually found to extend down to 60m. Ocean thermal energy potential per unit area (E) defined as ocean thermal energy above 26°C down to a depth of 60 m is given by

$$E = \int C\,(T-26)\,dz$$

Where the integral is from the surface to 60 m or to where T-26°C is the ocean density. C the Specific heat of water and T the oceanic temperature in °C. It has been found that Typhoons weaken when they cross the track of a previous typhoon, which has produced a lower SST due to up welling.

10. **Vertical Gradient of θ_e between the surface and 500 hPa ($\partial\theta_e / \partial\rho$)**. This represents the degree of conditional instability. Deep cumulus convection is the primary mechanism for TC. For this θ_e difference between the boundary layer and mid troposphere should be large so that air parcels are acted on by large buoyancy force. Gray took

($\partial\theta_e / \partial\rho$) between the surface and 500 hPa levels. This parameter typically ranges between 15° and 20°K.

11. **High Relative Humidity (RH) in the Middle Troposphere**. It has been observed that TCs form only in regions where seasonally averaged value of middle level humidity are high. Regions with low RH in the middle troposphere are unfavourable for Tropical Cyclogenesis for two reasons. First, convective cloud which originates in the boundary layer are eroded by entrainment of dry air as they rise through the middle troposphere. Second, since much of the mass convergence in TCs occurs above the boundary layer dry middle level air means less total moisture convergence in a column and less latent heat release. Although large mid troposphere RH are necessary for Tropical Cyclogenesis this parameter does not differ significantly in convective systems which intensify to TCs and those which do not.

12. **Physical Processes**.

a. At the low levels, friction causes air to flow into the low-pressure region. As air flows in, it will increase in tangential speed due to conservation of angular momentum and thus v_θ increases towards the centre.

b. At the point where the increase in angular momentum is balanced by the increase in frictional dissipation, air can no longer flow in and will have to rise. This radius is the RMW.

c. As air rises, moisture condenses, forming clouds and rain. Latent heat is also released. At the same time, cyclonic angular momentum will be transported upward. However, some of this will be lost due to turbulent mixing so that the tangential winds decrease with height.

d. When the rising air hits the tropopause or the stable lower stratosphere, it will have to spread out, which leads to outflow. At this point, v_θ is still +ve so that in the upper troposphere the outflow is *cyclonic*. However, as the air moves outward, conservation of angular momentum will eventually cause v_θ to become -ve, i.e. anticyclonic.

e. The outflow will also advect the warmer air (due to latent heat release) outward, leading to warming at the upper levels at large distances.

f. Another branch of the rising air when hitting the tropopause will meet and converge near the centre, forcing the air to sink which leads to warming due to adiabatic compression. This warm core is also consistent with the decrease of tangential winds with height.

g. A warm core system must have a high height region at the upper troposphere. Because the stratosphere is stable, this leads to a lower thickness in the lower stratosphere and therefore cold.

The Dynamics of Tropical Cyclones

13. Dynamically, storms (existing and potential) are subject to influences from the surrounding areas. In the Tropics, we usually encounter two layers in the troposphere, and these levels have very different characteristics. Most of the time, a steady trade wind blows in the lower levels (surface to 500 hPa), while a succession of large cyclonic and anticyclonic vortices are present in the upper troposphere in the 400- to 150-hPa strata.

14. Aside from the surface chart, the 700-hPa level chart is very helpful in determining low-level flow patterns. The 200-hPa level is representative of the upper layer, whereas the 500-hPa level, often located in the transition zone, is of much less use in tropical than in extratropical forecasting. However, forecast considerations should not be limited to the two layers in the Tropics. Middle latitude weather also influences and shares in the control of weather changes in low latitudes. The position and movement of troughs and ridges in the westerlies affect both the formation and motion of tropical storms. Therefore, middle latitude analysis is important. Since forecasts generally run from 1 to 3 days, there should also be a hemispheric analysis.

15. The climatological approach to the forecast problem should also be taken into consideration. It is obvious that a forecaster must be familiar with regional and seasonal changes; for instance, areas of frequent tropical cyclone formation mean storm tracks, the scattering of individual tracks about the mean, and the month-to-month variations of all of the above. Mean tracks and other data on Pacific storms are available in several publications. These maps and charts reveal that the "climatological" approach gives some useful information, but it cannot be relied on in anyone specific area, or in the case of any particular storm to the exclusion of synoptic indications for the forecast. These probability considerations are used mainly for long-range planning and when data are unavailable. Therefore, we reach the conclusion that climatological information should be treated as a weighing factor to be included after evaluating synoptic data.

Synoptic Conditions Favorable for Development

16. The following list contains conditions favorable for development of tropical cyclones. In this list, no attempt is made to separate the surface from the upper air indications, as the two are most often occurring simultaneously and are interrelated.

a. Marked cyclonic turning in the wind field.

b. A shift in the low-level wind direction when easterlies are normally present. Wind-speeds generally 10 knots or more.

c. Greater than normal cloudiness, rainfall, and pressure falls. Cloudiness that increases in vertical as well as horizontal extent.

d. Moderate to strong outflow aloft.

e. Sea level pressures lower than normal. The value is dependent upon the region analysed; however, a deviation of greater than 3 hPa is the normal criterion.

f. Easterly wind speeds 25 percent or more above normal in a limited area, especially when the flow is Cyclonic.

g. Temperatures above normal at sea level, generally 26°C (79°F) or more in the lower layers.

h. Moisture above normal at all levels.

j. Westerlies greater than average, north of the latitude of the seasonal maximum.

k. Easterlies weaker than average in a wide zone.

l. Easterlies decreasing with height.

m. The latitude of the subtropical ridge is higher than normal above the 500-hPa level.

n. Evidence of a fracture in a trough aloft (at 200 hPa).

p. Long waves those are slowly progressive.

q. A zone of heavy convection is present, indicating the absence of the trade inversion.

r. An increasing surface pressure gradient north and west of the suspect area.

s. The disturbance has relative motion toward the upper ridge at or above 400 hPa.

Life Cycle

17. Tropical cyclones are sustained by the energy released through the latent heat of condensation. This energy is furnished to tropical cyclones by the warm waters over which they develop and move. The warm moist air is lifted by a combination of convergence and instability, and upon condensing, the latent heat is liberated. When these systems move over land, the energy source is cut off, and they eventually dissipate. Tropical cyclones also weaken on moving over the colder waters of higher latitudes. Any decrease in the moisture and heat supplied by warm ocean waters weakens them.

18. The average life-span of tropical cyclones is about 6 days from the time they form until they either move over land or recurve to higher latitudes. Some last only a few hours, while others last as long as 2 weeks. The life cycle of the average tropical cyclone is divided into four stages.

19. **Formative or Incipient Stage**. This stage starts with the birth of the cyclonic circulation and ends at the time hurricane intensity is reached. This stage can be slow, requiring days for a weak circulation to begin, or in the case of an easterly wave, it can be relatively explosive, producing a well-formed eye in as little as 12 hours. In this stage, the minimum pressure reached is about 975 hPa. A good indication that a system of this type has formed or is forming is the appearance of westerly winds (usually 10 knots or more) in those tropical latitudes where easterly winds normally prevail.

20. **Immature or Intensification Stage**. This stage lasts from the time the system reaches hurricane or typhoon intensity until the time it reaches its maximum strength and lowest central pressure. The lowest central pressure often drops well below 975 hPa, and the winds spiral in a tight ring around the eye with a fair degree of symmetry. Usually, the radius of strongest winds are no more than 60 miles around the centre. This development may take place gradually or occur in less than a day. The cloud and precipitation fields develop into narrow, inward-spiraling bands.

21. **Mature Stage**. This stage lasts from the time a typhoon or hurricane attains its maximum intensity until it weakens to below this intensity or transforms, and is reclassified, as an extra-tropical cyclone. In this stage, the typhoon or hurricane may exist for several days at nearly the same intensity level, or decrease slowly. Strong winds extend farther and farther from the centre, affecting a much greater area. The winds and weather usually extend out farther in the right semicircle. By the time this stage is reached, the tropical cyclone is usually well advanced toward the north and west, or it has recurved under the influence of the mid-latitude westerlies. The typhoons of the Pacific usually last longer in the mature stage and grow to larger sizes than the hurricanes of the Atlantic.

22. **Decaying Stage**. This stage is characterised by rapid decay, as in the case of many tropical cyclones that move inland, or as associated with recurvature and transformation into an extratropical cyclone. In the case of a tropical cyclone moving over land, it steadily loses its strength and character. In regard to transformation, regeneration frequently occurs and results in maintenance or redevelopment of strong winds and other hurricane or typhoon characteristics. There is no set duration for the time a tropical cyclone may be in any one stage. It is entirely possible that a stage may be skipped or gone through in such a short time that it may be missed without synoptic data. Satellites permit almost constant viewing of tropical systems and have led to a classification system based on the appearance of the cloud system associated with tropical cyclones.

Intensification of Tropical Cyclone

23. Only when easterlies extend vertically to 25,000 feet or more at the latitude of the vortex is intensification possible. This most frequently occurs when the subtropical ridge lies poleward of its normal position for the season. The following characteristics are indications of intensification:

a. Movement is less than 13 knots.

b. The disturbance is decelerating or moving at a constant speed.

c. The system has a northward component of motion.

d. A migratory anticyclone passes to the north of the storm centre.

e. A strong net outflow (divergence) is manifested by anticyclonic flow in the upper levels (200 hPa).

f. Long waves are slowly progressive. (Applicable to genesis also)

g. The trade inversion is absent; convection deep.

24. Other considerations to bear in mind:

a. Intensification occurs when the cyclone passes under an upper-level trough or cyclone, provided there is relative motion between the two. There is some indication that intensification does not take place when the two remain superimposed.

b. Poleward movement of the cyclone is favourable for intensification; equatorward motion is not favorable.

c. Intensification occurs only in areas where the sea surface temperature is 79°F or greater, with a high moisture content at all levels.

d. Other factors being equal, deepening will occur more rapidly in higher than in lower latitudes. (Coriolis force is stronger.)

e. When a storm moves overland, the intensity will immediately diminish. The expected amount of decrease in wind speeds can be 30 to 50 percent for storms with winds of 65 knots or more; and 15 to 30 percent for storms with less than 65 knots. If the terrain is rough, there is more decrease in each case than if the terrain is flat.

25. Once an intense tropical cyclone has formed, there will be further changes in its intensity and in the course of its motion within the Tropics and during recurvature. The forecaster should consider these changes in connection with the predicted path of movement.

Movement of Tropical cyclones

26. Tropical cyclones usually move with a direction and speed that closely approximates the tropospheric current that surrounds them. Logically, therefore, charts of the mean flow of the troposphere should be used as a basis for predicting the movement of tropical cyclones, but lack of observations generally precludes this approach. Generally there is a tendency for tropical cyclones to follow a curved path away from the Equator; however, departures from this type of track are frequent and of great variety,

27. Tropical cyclones move toward the greatest surface pressure falls and toward the area where the surface pressure falls increase fastest with time. Calculation is necessary for this rule to be used.

28. Numerous theories have been advanced to explain the cyclone tracks of the past and to predict those of the future. Observational data have never been sufficient to prove or disprove most of them.

29. Tropical storms move under the influence of both external and internal forces. The external forces are a result of air currents that surround a storm and carry it along. The internal forces appear to produce a tendency for a northward displacement of the storm (which probably is proportional to the intensity of the storm), a westward displacement that decreases as latitude increases, and aperiodic oscillation about a mean track.

30. **Initial Movement of Intense Cyclones.** The movement of cyclones that are undergoing or have just completed initial intensification (about 24 hours) will be as follows:

a. Storms developing in westward moving wave troughs in the easterlies move toward the west and also with a pole ward component given by an angle of approximately 20° to the right of the axis of the trough looking downstream. (See fig. 2.1.)

b. The motion of storms developing from preexisting vortices can be extrapolated from the track of these vortices.

c. Storms developing on the eastern edge of polar troughs initially move up the trough.

d. Pacific storms forming on equatorial shearlines usually are most active in the southern and southwestern portions at the start and they move slowly northeast. After 1 to 2 days, the influence of the trades becomes dominant and the storms turn back toward a direction ranging from west to north. (This appears to be true mainly for cyclones deepening near or west of 130°E.)

e. If a storm is discovered without information as to its past movement and the upper air current, it is best to start it along the climatological mean track and then secure the data necessary to determine the steering current.

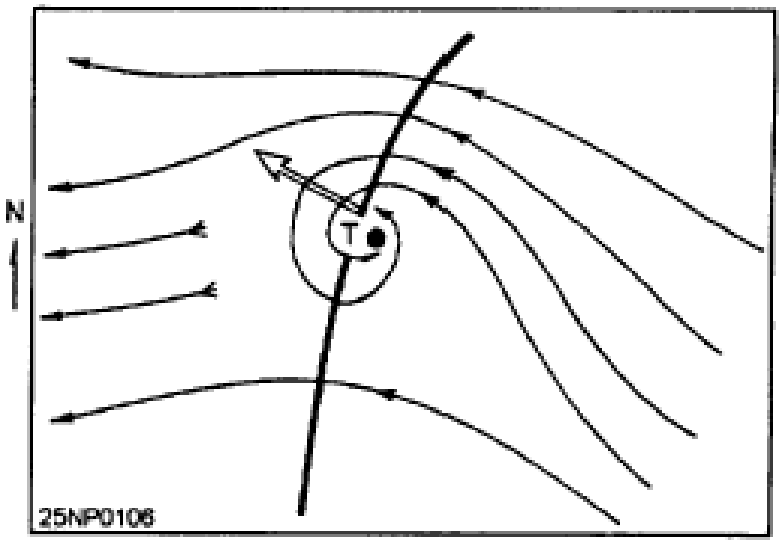


Fig 2.1. Initial movement of tropical cyclones forming in a wave trough In the easterlies

Characteristics of Tropical Cyclone

31. To make the most efficient analysis of available data in the vicinity of cyclone, you must be familiar with the normal wind, pressure, temperature, clouds, and weather patterns associated with these storms. No two hurricanes or typhoons are exactly alike. On the contrary, there are very great variations between storms. However, certain general features appear with sufficient frequency to permit mean pattern classifications. These features serve as a valuable guide in reconstructing the picture of an individual cyclone from sparse data.

32. Since meteorological elements are not distributed uniformly throughout all sections of these storms, it is customary to describe the storms in terms of right and left semicircles or four quadrants. The division into semicircles is along a line extending through the centre of the cyclone and in the direction toward which the storm is moving. The general features of hurricanes or typhoons given in the following section apply mainly to the mature stage.

33. **Surface Winds**. The surface winds blow inward in a counterclockwise direction toward the centre. The winds in the left-rear quadrant have the greatest angle of inflow (in the Northern Hemisphere). The diameter of the area affected by hurricane or typhoon force winds may be in excess of 100 miles in large storms or as small as 25 to 35 miles. Gale-force winds sometimes cover an area 500 to 800 miles or more. The maximum extent of strong winds is usually in the direction of the major subtropical high-pressure centre, which is most frequently found to the right of the storm's path in the Northern Hemisphere. Surface wind speeds of 140 knots have been successfully recorded, but accurate measurements of peak wind speeds in large mature storms have not been possible with any reliable degree of accuracy.

34. **Surface Pressure**. The sea-level isobars are an excellent tool with which to analyse these storms. The isobars take on a nearly symmetrical or elliptical shape, although deformations in the isobaric pattern are not uncommon. For instance, the tightest isobaric spacing (strongest pressure gradient) is found to the right of a storm's line of movement, and a trough often extends southward from these storms. The central pressures of mature storms are well below average. Central pressures of 890 to 930 hPa are not uncommon.

35. **Surface Temperature**. In contrast to extratropical cyclones, the tropical cyclone may show no cooling, or very little, toward the storm centre. This indicates that the horizontal adiabatic cooling caused by lower pressures is largely offset by the heat added through the condensation process. Upper-air temperatures have been found warmer by 5 'C or more.

36. **Clouds**. The cloud patterns of tropical cyclones also differ from those of extratropical cyclones. In mature tropical cyclones, almost all the cloud forms are present, but by and large the most significant clouds are the heavy cumulus and cumulonimbus which spiral inward toward the outer edge of the eye. These spiral bands, especially the leading ones, are also referred to as BARS. Cirrus and Cirrostratus occupy the largest portion of the sky over these storms. In fact, cirrus, becoming more dense, then changing to cirrostratus and lowering somewhat is more often than not a mariner's first indication of an approaching distant storm or the development of one in the near vicinity. The appearance of the sky is very similar to that of an approaching warm front. A typical cloud distribution chart for a tropical cyclone is found in fig 2.1.

37. **The Eye**. The eye of a storm is one of the oddest phenomena known in meteorology. Precipitation ceases abruptly at the boundary of a well-developed eye; the sky partly clears; the sun or stars become visible; the wind subsides to less than 15 knots, and at times there is a dead calm. In mature storms, the eye's diameter averages about 15 miles, but it may attain 40 miles in large typhoons. The eye is not always circular; sometimes it becomes elongated and even diffuse with a double structure appearance. The eye is constantly undergoing transformation and does not stay in a steady state.

38. **Precipitation**. Very heavy rainfall is *normal* amount and distribution of precipitation cannot be said to exist. Precipitation is generally concentrated in the inner core, where the slope of the barograph trace is the highest. Amounts of 20 inches are not uncommon. Over the open sea, rainfall is considered of operational interest primarily from the standpoint of its effect on ceiling and visibility. Over land, the threat of flooding is of paramount importance and must always be taken into consideration. Orographic effects produce concentrations of rainfall that often result in costly floods. Hurricane- or typhoon-force winds pushing moisture-laden tropical air up a steep mountain slope often results in phenomenal rainfall. A fall of 88 inches was recorded during one storm in the Philippines. At the other extreme, as little as a trace was recorded at a station in Florida that had winds up to 120 knots during the passage of a hurricane.

39. **State of the Sea**. The winds of a tropical cyclone produce wind waves which move outward from its centre. The size and speed of the waves depends upon the wind velocity and the length of the over-water trajectory of the winds. As the wind waves move away from the centre, their height diminishes, their length is reduced, and they become low undulating waves known as swell. The wind waves and swell move outward from the storm centre at a rate that nearly always exceeds the forward speed of the storm. Since these waves precede the storm, they are one of the first signs of its approach.

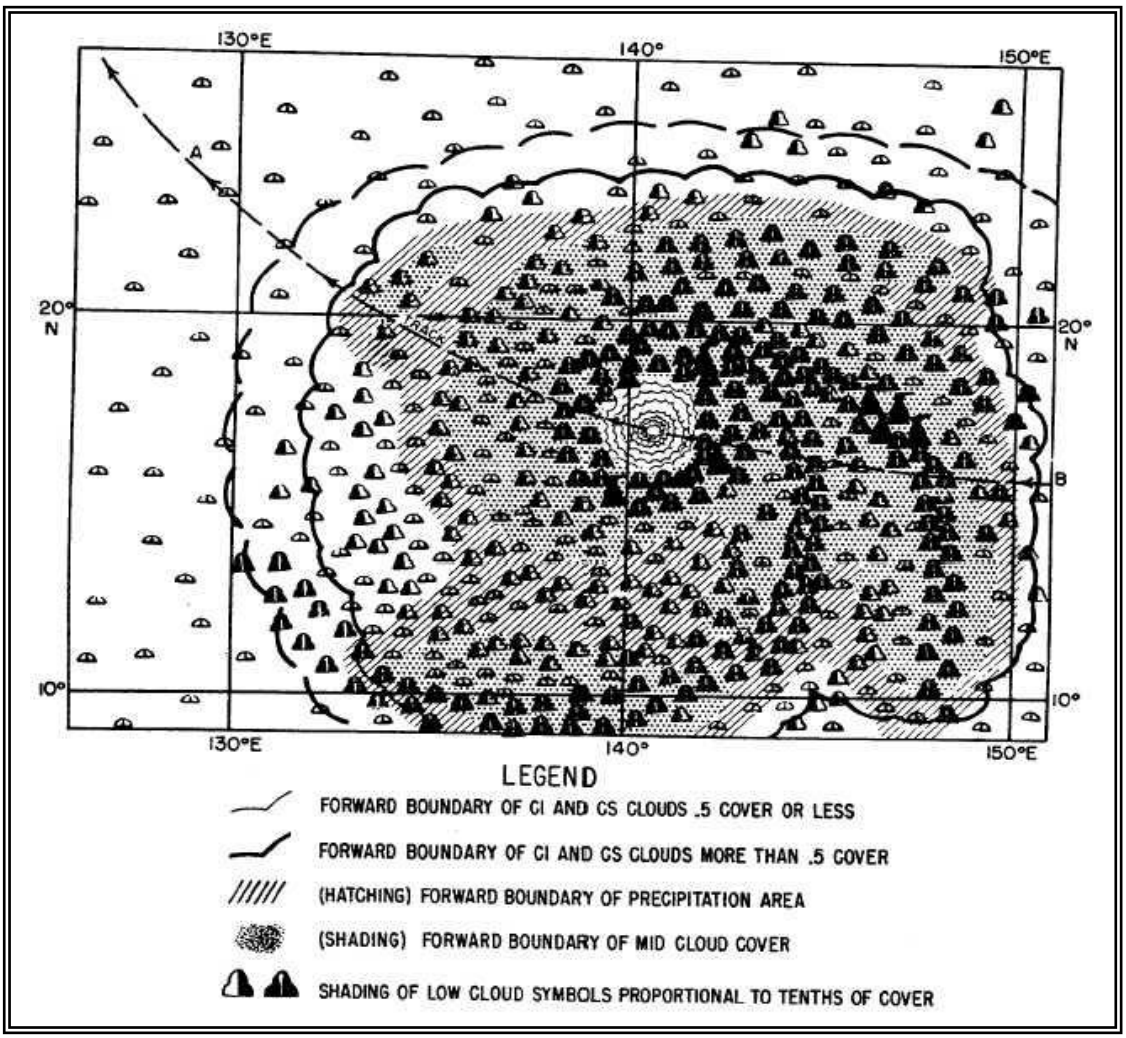


Fig 2.1 Typical cloud distribution associated with a tropical cyclone

40. In the open sea, the combination of wind and swell direction gives some indication of a storm's location, while the period of the swell (that is, the time, in seconds, between the passage of successive wave crests) is helpful in determining a storm's intensity. The wave period of the swells associated with a tropical cyclone is considerably longer than other swell waves normally observed in the tropics. In fact, in the Caribbean Sea and the Gulf of Mexico, long-period swell waves are not common except in connection with tropical cyclones. In these waters during hurricane season, the appearance of heavy swell waves with a period of 9 to 15 seconds is an indication of the existence of a tropical cyclone of storm intensity in the direction from which the swells are coming. Longer swell wave periods, 12 to 15 seconds, are almost a certain sign of a cyclone. Fig 2.1 shows the direction of wind and swell around a hurricane centre. One of the most severe effects of hurricanes and typhoons is the damage to coastal areas caused by large ocean waves. The most severe waves occur where land partially surrounds bodies of water, such as the Bay of Bengal and the Gulf of Mexico. Strong sustained winds in the right semicircle push water into coastal areas, thereby causing tides to rise by as much as 10 to 15 feet above normal. This effect is referred to as a storm tide. An added and even greater threat is the so-called hurricane wave. This term applies to the marked rise in the level of the sea near the centre of intense tropical cyclones. This rise can reach 20 feet or more and seriously affects small islands and continental shorelines. These waves have produced many of the major hurricane disasters of history. The hurricane wave may occur as a series of waves, but it is usually one huge wave. In partially enclosed seas, the hurricane wave may superimpose itself on the storm tide. There is usually little warning of its approach; however, it should be anticipated near, and to the right of intense tropical cyclone centres.

41. **Vertical Characteristics**. The vertical structure of a mature tropical cyclone also differs considerably from the extratropical cyclone. The first difference is the tropical cyclone's warm core. Between roughly 850 hPa and 100 hPa, the air temperature is generally 5°C to 15°C warmer at levels within the core than at corresponding levels in the surrounding atmosphere. The slope is assumed to be nearly vertical, and cloud distribution and the fact there is no air mass difference tend to support this assumption. Strong convergence occurs in the layer from the surface to roughly 10,000 feet, while strong divergence occurs above 30,000 feet. A net outflow of air is usually very pronounced in the vicinity of the 200-hPa level, and it is for this reason that the 200-hPa chart is one of the primary analysis tools in the tropics. The outflow at the 200-hPa level is manifested by anticyclonic circulation unless the storm is unusually severe. In such cases, cyclonic flow then extends to even the 200-hPa level.

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d. Pacific storms forming on equatorial shearlines usually are most active in the southern and southwestern portions at the start and they move slowly northeast. After 1 to 2 days, the influence of the trades becomes dominant and the storms turn back toward a direction ranging from west to north. (This appears to be true mainly for cyclones deepening near or west of 130°E.)

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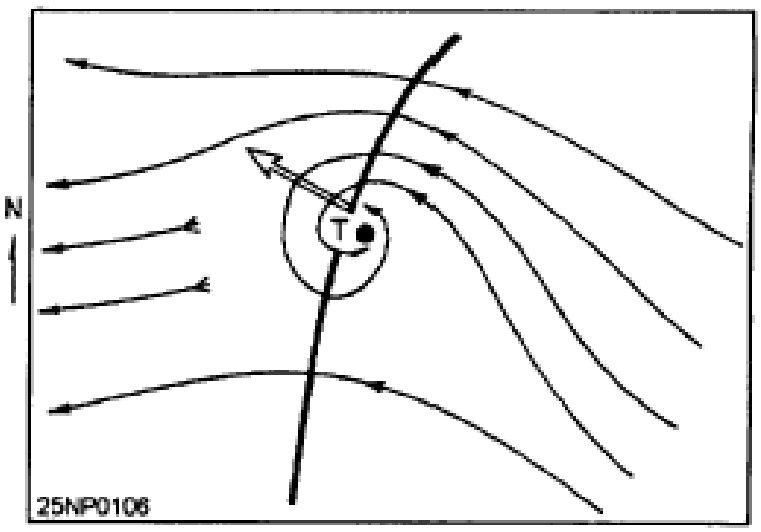


Fig 2.2. Initial movement of tropical cyclones forming in a wave trough In the easterlies

47. **Steering.** The movement of a tropical cyclone is determined to a large extent by the direction and speed of the basic current in which it is embedded. This concept appears to work well as long as the cyclone remains small and remains in a deep broad current. By the time a tropical cyclone has reached hurricane intensity, these conditions seldom exist. It then becomes necessary to integrate the winds at all levels through which the cyclone extends and in all quadrants of the storm to determine the effective steering current. Although the principle is not fully understood and has not yet received universal acceptance as a valid rule, it does have practical applications. For example, changes in winds atone or more levels in the area surrounding the cyclone can sometimes be anticipated, and in such cases, a qualitative estimate of the resulting change in the movement of the cyclone can be made. Conversely, when it appears likely that none of the winds in the vicinity of the cyclone will change appreciably during the forecast period, no change in the direction and speed of movement should be anticipated.

48. Further, this rule applies to situations of nonrecurvature and only some cases of recurvature. It is difficult to determine the steering current, since the observed winds represent the combined effects of the basic current and disturbances. As most storms extend into the high troposphere, it is better to calculate a "steering layer" than a steering level, since presumably the wind throughout most of the troposphere influences the storm movement. One writer recommends an integration of the mean flow between the surface and 300 hPa, over a band 8° in latitude centered over the storm. Another writer indicates that for moderate and intense storms the best hurricane steering winds would be found in the layer between 500 and 200 hPa and averaged over a ring extending from 2° to 6° latitude from the storm centre.

49. **Streamline Analysis and Vector Averages**. Other practical applications of the steering concept to short-range tropical cyclone motion use differing approaches in attempting to measure the basic current. One is by streamline analysis of successive levels to find a height at which the vertical circulation diminishes to a point such that the winds are supposedly representative of the undisturbed flow. Another method is to take vector averages of reconnaissance winds near the zone of strongest winds in the storm.

50. **Use of Observed Winds Aloft**. When sufficient data are available, it has been found that the use of streamline analysis of successive levels usually gives valuable indications of tropical storm movement for as much as 24 hours in advance. However, since wind observations are usually scarce in the vicinity of a hurricane, their analysis is necessarily rather subjective. Some forecasters claim dependable results in using this concept when data were available to high levels near the storm. The technique is not based on the assumption that wind at any single level is responsible for steering the storm, since the forces controlling movement are active through a deep layer of the atmosphere.

51. However, as successive levels are analysed, a level is found at which the closed cyclonic circulation of the storm virtually disappears. This *steering* level coincides with the top of the warm vortex and varies in height with different stages and intensities of the storm. It maybe located as low as 20,000 feet or, in the case of a large mature storm, as high as 50,000 feet. It has been found in analysis that most weight should be given to the winds in advance of the storm within a radius of 200 to 300 miles in preference to those in the rear quadrants. The hurricane generally moves with a speed of 60 to 80 percent of the current at the steering level.

52. **The Direction of Movement.** The direction of movement is not always exactly parallel to the steering current, but has a component toward high pressure that varies inversely with the speed of the current, ranging from almost 0° with rapid movement to as much as 20° with speeds under 20 knots. In westward moving storms, a component of motion toward high pressure could result from the poleward acceleration arising from the variation of the Coriolis parameter across the width of the storm. This would indicate that, to the extent that this effect accounts for the component of motion toward high pressure, northward moving storms would fit the direction of the steering current more closely than westward moving ones. The tendency for poleward drift would be added to the speed of forward motion in the case of a northward moving storm so that it would approach more closely the speed of the steering current. Empirical evidence supports this hypothesis.

53. Corrections for both direction and rate of movement should be made when this is indicated by the windflow downstream in the region into which the storm will be moving. For prediction beyond several hours, changes in the flow pattern for a considerable distance from the storm must be anticipated. It should also be remembered that intensification or decay of a storm may call for use of a higher or lower level, respectively, to estimate the future steering current.

54. **Recurvature.** One of the fundamental problems of forecasting the movement of tropical cyclones is that of recurvature. Will the cyclone move along a relatively straight line until it dissipates, or will it follow a track that curves poleward and eastward? When recurvature is expected, the forecaster must next decide where and when it will take place. Then, he or she is faced with the problem of forecasting the radius of the curved track. Even after the cyclone has begun to recurve, there are a great variety of paths that it may take. At any point, it may change course sharply.

55. The most common recurvature situation arises when an extratropical trough approaches a storm from the west or when the storm moves west to northwest toward a stationary or slowly moving trough.

56. **Indicators of Possible Recurvature.** Some of the indicators of possible recurvature are as follows:-

a. If the base of the polar westerlies lowers to 15,000 to 20,000 feet west of the storm's latitude, and remains in this position, recurvature may then be expected to occur.

b. However, if there is the building of a dynamic high or an eastward movement of this high to the rear of the advancing trough, and the westerlies dissipate in the low latitudes, the storm will move past the trough to its south and continue its westward path.

c. The above rule also holds true in cases where the polar trough moves from the west against a blocking high. The higher latitude portion of the trough continues to move eastward while the southern segment of the trough is retarded and is no longer connected with the upper portion of the trough.

d. Recurvature may be expected when an anchor trough is about 500 miles west of the storm and when the forward edge of the westerlies is from 500 to 700 miles to the west. Correlated with this parameter, RJ Shafer found that a spot value of the thickness between 850 to 500 hPa 7.5 degrees of latitude to the northwest of the storm was one of the most valuable of all parameters on recurvature or as an indication of future movement of Atlantic hurricanes. This thickness reflects the relative strength of cold troughs to the west essential for recurvature. Low values of thickness, 14,000 feet (approximately 4,270 meters) or less, almost always indicate recurvature; and high values, 14,200 feet (4,330 meters) or more, generally indicate continued westward motion.

e. The major trough west of the storm (in the westerlies) is slowly progressive.

f. Long waves are stationary or slowly progressive.

g. There is a rapid succession of minor troughs aloft.

h. The climatological mean track indicates recurvature (use with caution).

j. When the neutral point at the southern extremity of the trough in the westerlies at the 500-hPa level lies at or equatorward of the latitude of the cyclone, recurvature into the trough will usually occur. In this situation, the cyclone would normally be under the influence of southerly winds from the upper limits to a level well below the 500-hPa level while approaching the trough.

k. When the subtropical ridge at the 500-hPa level is broad and consists of large anticyclones, recurvature usually occurs. This case represents a low index situation in which the cyclone remains under the influence of a single, large, slow-moving anticyclone for a relatively long time.

l. Weak troughs exist between two separate subtropical high cells. Sometimes tropical storms move northward through these very weak breaks in subtropical highs.

57. **Nonrecurvature.** The following flow patterns are associated with nonrecurvature:

a. Strong subtropical anticyclone or ridge to the north of the storm with the mean trough in the westerlies located far to the west of the longitude of the storm. If this pattern develops strongly over the western oceans or continents, a storm will generally be driven inland and dissipate before it recurves into the western end of the ridge.

b. Flat (small-amplitude waves) in the westerlies at latitudes near or north of the normal position. A narrow subtropical ridge separates the westerlies from the tropical trough. In many cases the mean trough in the westerlies may be located at the same longitude of the storm.

58. **Motion during Recurvature.** The following rules apply during recurvature only:

a. When the radius of recurvature of a storm is greater than 300 miles, it will not decelerate and may even accelerate. The storm will slowdown if the radius of recurvature is less than 300 miles. In general, when the radius of recurvature is large, it is usually very uniform. A small radius will occur along a brief portion of the track. (See fig. 2.3)

b. A large radius of recurvature is to be expected if the high northeast of the tropical cyclone has a vertical axis (fig. 2.4 (A)). This will occur when long waves are stationary.

c. When the high slopes south to southeast with height (fig. 2.4 (B)), the cyclone is transferred rapidly from the influence of upper easterlies to that of the upper westerlies and the track has a short bend. This occurs when long waves are progressive.

59. The following rules refer to short-term (24 hours or less) forecasting:

a. Tropical cyclones move toward the area of greatest surface pressure falls (12- to 24-hour pressure change).

b. Tropical cyclones move toward the area where the surface pressure falls increase most rapidly with time. For calculation, 3-hourly reports are necessary. Take, for example, the 24-hour pressure change at 1800 and subtract it from the 24-hour pressure change at 1500. This gives the acceleration of the pressure fall. If this quantity is computed for a network of stations and isolines are drawn for a suitable interval, the negative centre helps to pinpoint the expected cyclone position. This rule is used at intervals of 12 hours or less, and is especially helpful in determining the precise place of landfall for a storm that is just offshore.

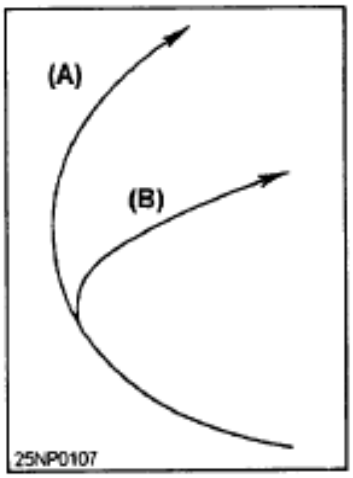


Fig 2.3. (A) Recurvature at a constant speed;
(B) First decelerating and then accelerating

60. **Changes in Intensity during Movement out of the Tropics.** During movement out of the Tropics the storm comes under influences different from those in its origin and tropical path. The following rules and observations can be helpful in forecasting the storm's changes in intensity:

a. A storm drifting slowly northward with a slight east or west component will preserve its tropical characteristics.

b. Storms moving northward into a frontal area or area of strong temperature gradients usually become extratropical storms. In this case the concentrated centre dies out rapidly while the area of gale winds and precipitation expands. The closed circulation aloft gives way to a wave pattern and the storm accelerates to the northeast.

c. If the tropical cyclone recurves into a trough containing a deep, slowly moving surface low, it normally will overtake this low and combine with it. Temporary intensification may result.

d. If the tropical cyclone recurves into a trough containing an upper cold core low, it appears to move into the upper low in most cases.

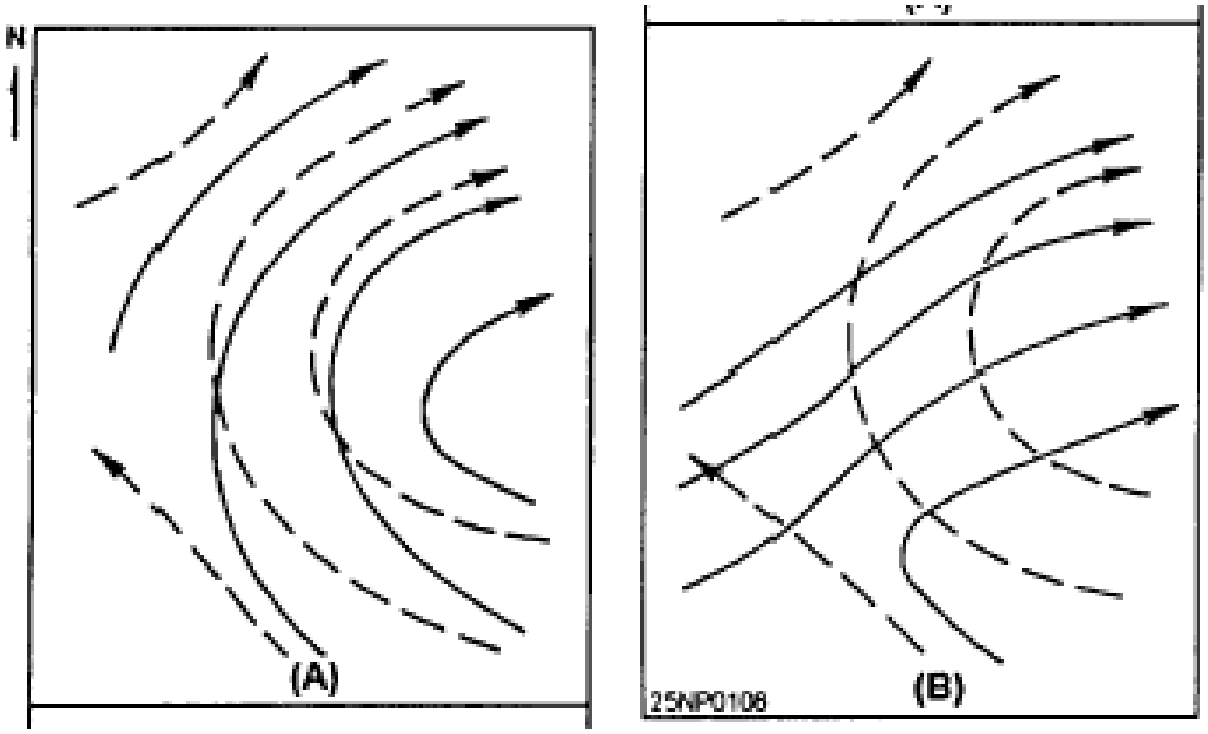


Fig 2.4. Surface flow pattern (dashed line) and 300- to 200-hPa flow pattern (solid lines).

TASK – 12 : JET STREAMS

Introduction

1. The first indication of a strong current encircling the earth was provided by Bjerknes and collaborators, around 1930, who computed meridional profiles of the thermal wind, through the Eurasian region. Since about 1942 direct wind observations have shown that there is highly concentrated current that encircles the earth. Because of its high concentration and great speed, it has been referred to as the Jet Stream.

2. The existence of strong winds at high level has been known in a general way for a long time through occasional high level pilot balloon observations or Ci movements. During the World War II the study of the subject gained prominence when the subject was suddenly pushed into the forefront by the requirements of aviation as well as the introduction of radio and rawinsondes as routine observational tools. American B-29 first began to bomb Japan when the pilots reported encountering such high velocity head winds at 20,000 to 30,000 feet that their ground speed was reduced nearly to zero. Similarly when they turned downwind, then ground speed became nearly double their air speed. There reports were hard for meteorologist to believe, but subsequent high-altitude weather studies disclosed the presence of narrow bands of high velocity winds of 200 to 300 miles per hour in other parts of earth.

3. Since 1950 high-level wind observation with the help of radio or rawinsonde, aircraft and rockets have given us large amount of information on jet stream.

Definition

4. The commission for Aerology of WMO, recommended the following definition for a jet stream, which has been provisionally adopted by WMO: ‘A jet stream is a strong narrow current, concentrated along a quasi-horizontal axis in the upper atmosphere or in the stratosphere, characterised by strong vertical and lateral wind shears and featuring one or more velocity maxima’. For operational purposes the commission also recommended the criteria, ‘Normally a jet stream is thousands of kilometres in length, hundreds of kilometres in width and some kilometres in depth. The vertical shear of wind is of the order 5 to 10 m/sec per km and the lateral shear is of the order 5 m/sec per 100 km. An arbitrary lower limit of 30 m/sec (≅ 60 knot) is assigned to the speed of the wind along the axis of a jet stream’. A typical figure of Jet Stream is shown in Figure 1.

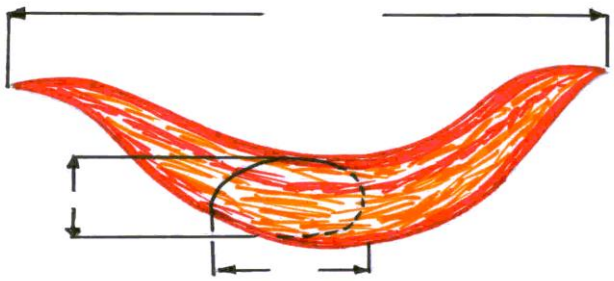


Figure 1: A typical figure of Jet Stream

Causes of Jet stream

5. The following physical condition may be regarded as the causes of Jet Stream:

a. Concentration of the large fractions of kinetic energy of the atmosphere in the Jet Stream belt.

b. Concentration of the bulk vertical temperature change in a narrow zone.

Existence of Jet stream

6. Although the origin of the jet stream is still obscure, recent observations have provided a fairly complete description of this spectacular phenomenon. The jet is essentially circumpolar in character and completely girds the earth in a meandering course in both the Northern and Southern Hemisphere. In places it appears to consist of several individual filaments of high velocity air and in other parts of filaments combine to form a single great stream of rushing air moving west to east. These winds attain their highest velocity at elevations between 30,000 and 40,000 feet. Although jet stream is sinuous (turning while moving) in character its average location is about 30° to 35°.

7. In the upper troposphere the westerly winds increase in the velocity from the poles equator ward until they reach a maximum speed at about 30°, after which they decline rapidly before reaching the region of weak easterly winds. Thus these high velocity winds aloft line immediately about the subtropical highs and horse latitude at the surface which are region of variable and weak winds.

8. In the stratosphere the horizontal temperature gradient are usually opposite in direction to those of the tropospheric systems. This is consistent with the tropopause in a warm air mass being higher than that in a cold air stream. In the stratosphere the air over the equator is colder than that over the poles and similarly the stratosphere over a warm sector is colder than that on the cold side of the lower level fronts. The thermal wind therefore, increases with height to a maximum at or near the tropopause and then, as a result of the reversal of the horizontal temp gradient, decreases with height in the stratosphere. Hence, winds are particularly strong in a channel near tropopause. This type of wind distribution as a tube rather than a channel of strong winds are jet streams.

Classification and Location of Jet Stream

9. Atmospheric jet streams occur at all levels but the classification is based upon the frequently observed jet streams in the synoptic or climatological charts of the troposphere and the lower stratosphere. The types and location of Jet Stream are shown in Table 1.

Table 1: Classification and Location of Jet Stream.

Ser no	**Types of Jet Stream**	**Location**	**Level**
1	Polar Front Westerly Jet	Between 30°N - 70^{0}N	300mb - 200mb
2	Sub Tropical Westerly Jet	25°N - 35°N	250mb - 150mb
3	Tropical Easterly Jet	Equator to 15°N	200mb - 100mb
4	Low Level Jet	Upto 15°N	Below 6 km
5	Arctic Stratospheric Jet	Arctic & Antarctic	20 km
6	Arctic Jet	Above 70^{0}	Below 300mb

Polar Front Westerly Jet (PFJ)

10. PFJ is found in the high-level middle latitude westerlies between 30^0N-70^0N at a mean height of about 9.0 km close to the middle latitude tropopause. The PFJ migrates in winter to as far as latitude 30^0N and in summer to as far north as latitude 70^0N. The core speeds are stronger in winter than in summer. The main direction is westerly, although on many occasions the Jet Stream may be North Westerly or South Westerly. Normally speeds are of 80 kts to 100 kts in winter. It lies at a level between 300mb to 200mb. As per Singh (1971) PFJ is located to the north of main STJ at latitude 43^0N at 11.0 km (240 mb) with an average speed of 100 kts. It is located over the 500 mb level temperature of –29^0C ($\pm 3^0$C) and the 300 mb temperature of –53^0C ($\pm 5^0$C). PFJ is produced by a temperature difference and is closely related to the polar front. At high-level wind blow along the isotherms and closely packed isotherms indicate strong winds.

Sub Tropical Westerly Jet (STJ) over Indian Sub-continent

11. The main characteristics of the sub-tropical jet stream over Indian Sub-continent in winter are as follows:

a. The mean jet stream over the Indian sub-continent (in winter) lies near latitude 27°N at a height of about 12 km (200 hPa). The mean wind speed at core is about 100 knots. There is a slight shift in the axis of the jet (by about 2-3 degrees of latitude) southwards as the season advances and towards end of winter or early spring, the jet stream reaches the southern most position.

b. Various rawin stations prove that the mean winds of jet stream flow across India from Jodhpur to Guwahati. The jet stream continues further northeastwards (to the east of India) across South China, to Japan and the core speed progressively increases downstream, reaching a maximum over South Japan. At 200 hPa, the mean wind at Shionomisaki in South Japan is nearly 150 knots, the highest over the area.

c. The STJ lies entirely within the tropical troposphere about 3-4 km below the tropical tropopause. The jet is in the region of a break in the tropopause between the tropical tropopause (near 200-250 hPa).

d. The jet is caused by a concentration of the horizontal temperature gradient below the jet level and a reversal of the gradient above the jet level. At and near the jet stream level, the temperature gradient is very small or nearly zero.

e. The STJ is a very steady phenomenon and is seen practically daily on the charts during the winter, in more or less the same geographical location. Palmen postulated the generation of the STJ, as due to the pole ward drift of air in the upper branch of the Hadley cell of the general circulation, with partially conserved absolute angular momentum. Consistent with this concept, the STJ is located near the pole ward boundary of this cell. The near constancy in the latitudinal position of the circumhemispherical STJ is thus related to the constancy of the Hadley cell. Over Indian Sub-continent, the presence of the Himalayas is also believed to contribute to the stability of the geographical location of the STJ. In comparison the Polar Front Jet (PFJ), though quite pronounced on individual day's charts, varies considerably in space and time. As a result, it is not clearly delineated on the mean charts. It is the STJ alone

that is clearly pictured on the mean charts. Subtropical and Polar Jets are shown in Figure 2.

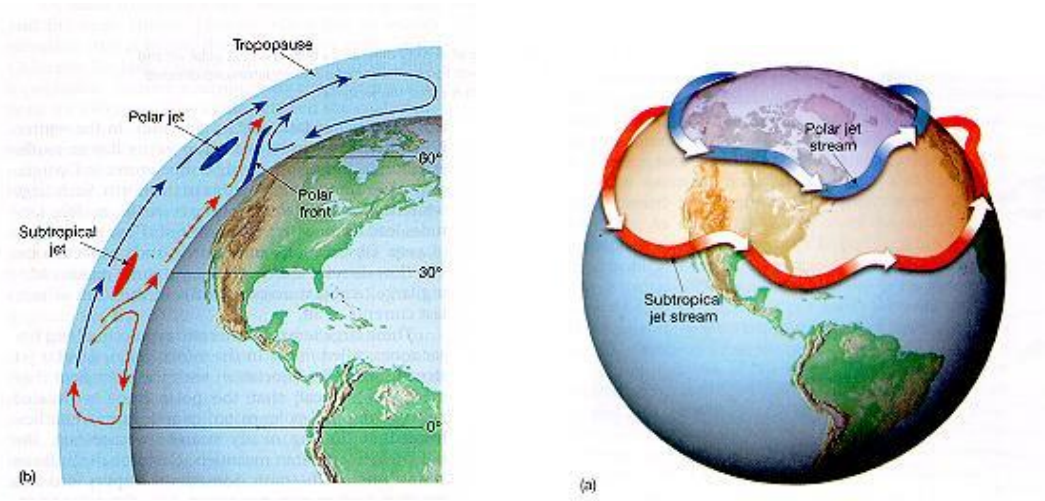


Figure 2 : Subtropical and Polar Jets.

<u>Tropical Easterly Jet Stream (TEJ) over Indian Sub-continent</u>

12. The main characteristics of the TEJ over Indian regions are as follows:

a. It occurs in the summer tropical region above 300 hPa level; strong easterly winds develop between the equator and subtropical ridges. The speed may be 60 knots.

b. It becomes relatively stronger during summer season rather than the winter season of the respective hemispheres.

c. Easterlies well developed during northern summer over Southeast Asia, particularly over South India.

d. The wind speed has large variation of the order of 20 knots, while the wind direction has very little variation.

e. Maximum intensity of easterly wind appears around latitudes 9°N, ie around 77 knots.

f. Vertical wind shear is of the order of 10 knots/km. The shear above the level of the maximum wind is slightly higher than below the level of maximum wind.

g. Horizontal wind shears are very weak being of the order of 10 kts/500 km. Anticyclonic shear is to the north of the wind maximum.

<u>Low Level Jet Stream</u>

13. These do not satisfy the WMO recommended definition of jet stream speed $\geq$ 30 m/sec. This vertical and horizontal extends are also not comparable with other types of jet stream. The characteristics of LLJ are discussed as:

a. These low level jet streams are generally located in the lower levels, within 1 to 2 km of the troposphere.

b. These are strongly influenced by orography, friction, diurnal cycle or heating and corresponding, variation of pressure gradient and static stability.

c. Favourable locations of occurrence of low-level jet streams or strong air current are as follows:

(1) Slopes of mountain parallel to the anticyclone flow around the subtropical anticyclone, eg over west central USA, along the Peru Coast in South America.

(2) North-South oriented continental coasts near cross equatorial flow, eg East America low-level jet.

(3) Marsabit (north Kenya) jet stream.

Arctic Stratospheric jet Stream (SJ)

14. Stratospheric jet stream is found at levels about 20 km in the Arctic and Antarctic regions. These are Westerlies in winter and Easterlies in summer.

Arctic Jet Stream

15. Arctic Jet Stream is found in the high level westerlies of the arctic regions at mean height of about 7 km to 8 km, above 70^0 close to the polar tropopause.

Jet Stream as Seen in Satellite Picture

16. Satellite picture shown extensive cloud formation associated with Jet Stream which are probably one of the longest cloud systems seen in satellite pictures. There are following four types of jet associated cloud organisations:

a. Long shadow lines (shadow is black line between clouds).

b. Sharp edged large Ci sheet.

c. Longitudinal Ci bands or streams. (longitudinal bands indicates large vertical wind shear, speed of jet 80 kts and small lapse rate).

d. Transverse bands in Ci formations.

17. Sub-paragraph a & b are associated with PFJ while sub-paragraph c & d are associated with STJ. The jet axis could be placed parallel to the shadow line or the northern boundary of the Ci sheet, slightly to the north of it. The jet cirrus is most often observed when the winds follow a broad anticyclone turning path. Where jet stream turns cyclonically, the Ci as rarely detectable. Jet cirrus usually forms in association with large amplitude troughs. Clouds form only ahead of the trough line.

Jet Stream and Aviation Hazards

18. **Missile Launch.** Conditions for the missile flight are most adverse when the momentum increases with height. This is most likely to happen in situations when the bottom of jet stream is situated in the layer between 10,000 and 20,000 feet pressure altitude, much lower than usual.

19. **Navigational Problems.** Jet stream shows vertical wind shear which influences the optimum flight level, the horizontal structure of the wind filed is of much greater importance in the choice of optimum flight route. Those aircraft whose speed is greater when wind speed in strong jet streams are hardly affected by jet stream. Fuel consumption also increases when aircraft fly against jet stream wind direction.

Clear Air Turbulence (CAT)

20. The most disastrous effect of jet stream is clear air turbulence though each and every jet stream is not associated with Clear Air Turbulence (CAT). CAT leads to bumpiness in flight causing injuries to passenger and stresses in the material, which lead to metal fatigue shortening life of plane. In the jet stream region there are following 3 main areas with excessive turbulence frequency and with unstable flow conditions:

a. On the anti-cyclonic side of the jet stream approximately at the level of jet axis.

b. On the cyclonic side of the jet axis near the tropopause.

c. Below the jet axis in the region of the baroclinic and thermally stable layer of the jet stream front.

21. CAT usually extends about 200 km to each side of the jet stream core and may extend from 80 to 200 km downstream from the mountain ridge. If a known mountain wave zone exists in 250 hPa level wind, the CAT forecast area to mountain wave zone rather than centred on jet stream core. The areas of turbulence in jet stream is shown in Figure 3.

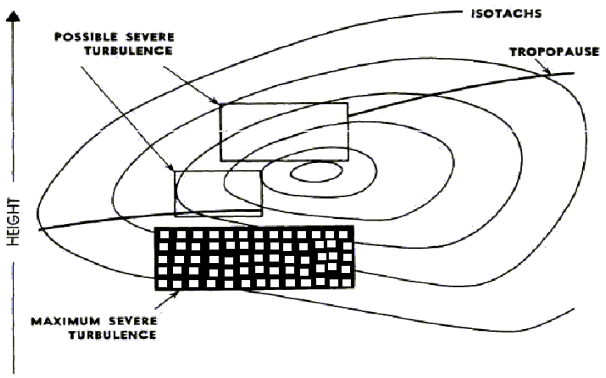


Figure 3: Areas of turbulence in Jet stream.

Weather

22. **Development of Weather near Jet Stream.** Jet streams form between two great blocks of dissimilar air, cold air from Polar Regions meets warm air from the tropics along the

polar front. The warmer and colder the air masses in the first 5 miles of the troposphere the stronger the jet for pressure higher aloft in the warm air mass and lower in cold air mass. Jet

strength is a measure of the sharpness of approaching fronts and the strength of surface winds. Jet streams send warning of the approach of the surface winds and weather by forming characteristic Ci clouds in the warm air on its tropical side in either hemisphere. The *right entrance* and *left exit* of the jet maxima are favourable areas of development of enhanced weather activity, Figure 4.

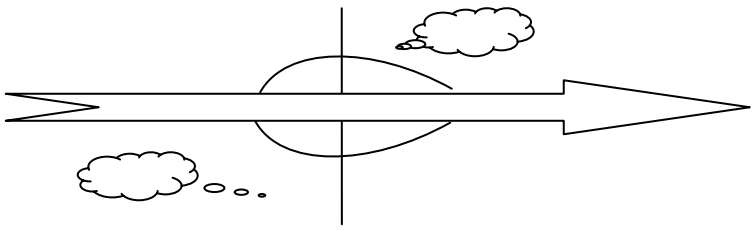


Figure 4: Right entrance and left exit of jet maxima are favourable areas for development of weather activity.

23. **Precipitation and Moisture Distributions.** Anticyclonically curved jet stream maxima show an increasing tendency for the formation of precipitation in the right entrance quadrant whereas cyclonically curved maxima show this tendency in the left exit quadrant. Jet streams with approximately straight axes show a distinct shift of maximum precipitation toward the right side in the entrance region. In the delta region the jet maximum shows a significant excess of precipitation amount in the left sector.

Conclusion.

24. Jet stream is a strong current with a minimum speed of 60 knots encircling the earth. It has got higher vertical wind shear is more in the jet stream than that of the horizontal wind shear. Though the exact cause of jet stream is not yet known; however, vertical temperature change and concentration of large kinetic energy in the jet stream belt are considered the main causes of jet stream. Jet streams are classified according to their locations. Main jet streams observed over Bangladesh and India sub-continent are Subtropical Westerly Jet Stream (STJ) and Tropical Easterly Jet Stream (TEJ). The mean location of SJT is 27^0N at a height of 12 km (40,000 feet) with average speed of 100-150 knots.

25. Jet stream not only can be identified by rawin-sonde observations but also by cloud imageries in satellite picture. Jet stream affects the missile launch and creates navigational problems. The most hazardous CAT is found near the jet stream.

26. Jet stream also favours the development of weather. Right entrance and left exit of jet maxima are favourable areas of development of weather activity. Anticyclonically curved jet stream maxima show an increasing tendency for the formation of precipitation in the front left quadrant whereas cyclonically curved maxima show this tendency in the right rear quadrant. Jet streams with approximately straight axes show a distinct shift of maximum precipitation toward the right side in the entrance region.

TASK – 13 : TROPICAL PHENOMENA

Introduction

1. The tropical phenomena (features) we are about to discuss do not, for the most part, correspond to any mid-latitude phenomena. Shear lines, easterly waves, tropical cyclones, the tropical easterly jet stream, and the intertropical convergence zone are all a part of tropical analysis, and you must have an understanding of these phenomena in order to relate them to weather occurring within the tropics. Some of the features discussed will be in more detail than others; therefore, for a more in-depth look at all the above features,

Shear Line

2. A shear line is a narrow zone across which there is an abrupt change in the horizontal wind component. As was stated earlier, the primary characteristic of these lines is the low-level horizontal wind discontinuity. The discontinuity results from strong subsidence (downward vertical motion) associated with an intensifying high-pressure system poleward of the polar front. As the tail end of the polar front advances into the tropics, the subsidence causes frontolysis. In lower levels of the troposphere, strong easterly winds develop poleward of the line and weaker easterly winds equatorward of the line. Figure 9.19 illustrates shear line formation. In streamline analysis, the area where the polar front becomes a shear line is denoted by a neutral point. See figure 9.20.

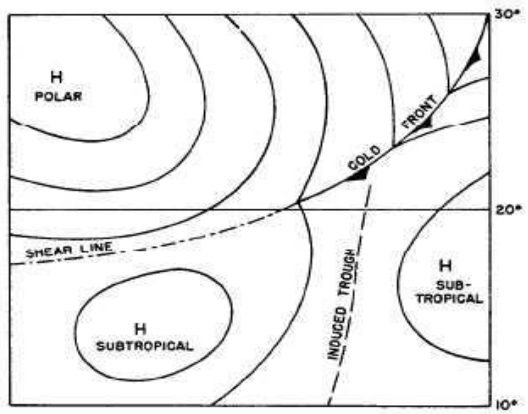


Figure 9.19. Shear line resulting when the polar high invades the tropics and reinforces the subtropical high (also note induced trough)

3. Because of the subsidence on the poleward side of shear lines, the clouds along these lines exhibit lower tops than those of the frontal zones. Isolated convective buildups occur along the lines, but the predominant clouds are of the lower varieties, with tops 10,000 to 15,000 feet or less. The three most common clouds found along shear lines are altostratus, high stratocumulus, and nimbostratus. Rainy periods are common along with scattered in-stability showers, reduced visibilities, and low ceilings. The seas poleward of a shear line can build rapidly, because of the increased wind speed and relatively small change in wind direction. With the passage of a shear line, the sudden increase in wind speed is known as the "surge of the trades." Over North Africa in winter, polar outbreaks

are not accompanied by great amounts of precipitation (until reaching the equatorial zone), but strong surface winds poleward of the front and/or shear line raise widespread dust, called “Harmattan haze,” in West Africa.

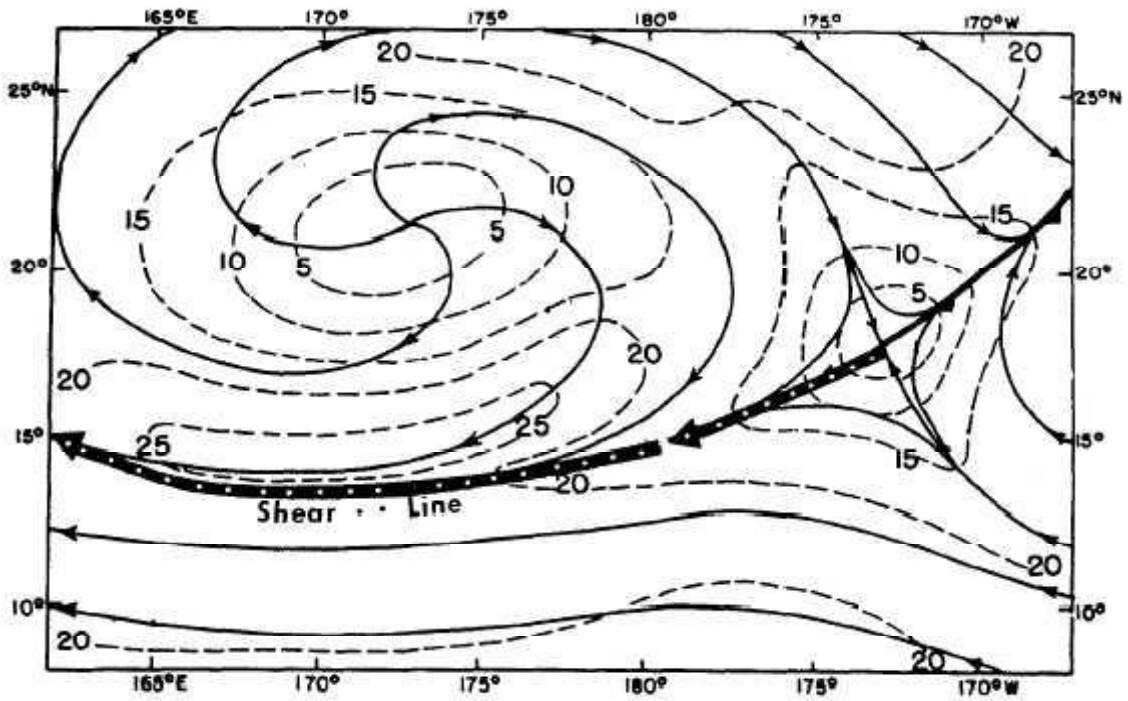


Figure 9.20. A shear line in the wind field

Easterly Waves

4. An easterly wave is defined as a migratory wave-like disturbance of the subtropical easterlies or put more simply, a trough in the easterlies that moves from east to west. See figure 9.21. The wave is usually oriented from south-southwest to north-northeast on surface charts and lower-level upper-air charts in the Northern Hemisphere and from north-northwest to south-southeast in the Southern Hemisphere. The deformation of the isobars may be more pronounced at the 700-hPa level than at the surface. Easterly waves lie within the boundaries of the tropical easterlies-between 25°N and 20°S. Easterly waves are classified as stable, neutral, and unstable.

Stable Easterly Waves

5. Stable waves slope east with height and have divergence ahead, and convergence to the rear, of their troughs. The height of the moist layer is at a minimum approximately 20 miles ahead of the trough line, but it rises rapidly near the trough line and attains its maximum height in the region of strongest convergence behind the wave. Further to the east of this zone of maximum convergence, the top of the moist layer lowers again to its normal height. As was discussed earlier, the height of the moist layer plays a major role in the type of clouds and weather that occurs within the tropics.

6. To the west, ahead of a stable easterly wave, the characteristic cloud is cumulus, and the visibility is poor in haze. To the east, behind the wave, cumulus congestus and cumulonimbus with showers and thunderstorms are characteristic. The cumulonimbus are usually arranged in parallel bands. Layers of altocumulus, altostratus, and cirrus connect the cumulonimbus. Visibility is good except in the precipitation.

7. Near and to the rear of the trough line, temperatures may lower from the evaporation of falling rain and the vertical stretching of the atmosphere created by the low-level convergence. However, this cooling does not amount to more than 4°F to 6°F in 24 hours and does not extend above 15,000 feet. Ahead of and to the rear of the convective area, the temperature remains normal or may even be slightly higher than normal.

8. Surface pressure falls of 0.5 hPa to 3.0 hPa in 24 hours occur ahead of the trough, and rises of a similar magnitude occur behind. The greatest relative pressure changes usually occur between 10,000 and 15,000 feet. Between 25,000 and 35,000 feet, there is normally no pressure change.

9. Easterly waves usually do not move as fast as the current in which they are embedded. A stable wave usually moves at a moderate speed of approximately 10 knots.

Neutral and Unstable Easterly Waves

10. When an easterly wave becomes vertical or slopes westward with height, the wave is deepening (intensifying). When vertical, the wave is said to be neutral. The lines of cumulonimbus and areas of precipitation, which are found east of the axis of a stable wave, move west to a position along the trough axis.

11. Neutral waves usually have a short life (often less than 24 hours) and they often revert to stable, but sometimes become unstable waves. When a wave slopes westward with height, the wave is said to be unstable. This type of wave has the most violent weather, and is often associated with the development of tropical cyclones that reach hurricane intensity. The showers and thunderstorms are found west, or ahead, of the trough line.

12. Satellite imagery and increased amounts of meteorological data have shown that the classical easterly wave does not occur as frequently as originally believed.

Intertropical Convergence Zone (ITCZ)

13. As satellite pictures have greatly increased the amount of information provided meteorologists, the pictures have also led to some changes in theories concerning various phenomena. One such phenomenon so affected is the Intertropical Convergence Zone, or as it is frequently referred to now, the zone of Inter-tropical Confluence (ITC).

14. The ITCZ, or ITC, is an area where the winds of the Northern and Southern hemispheres converge. The Regional Tropical Analysis Centre denotes this zone as “A nearly continuous fluent line (usually confluent) representing the principal asymptote of the Equatorial Trough.” In general, it is an area where horizontal convergence of airflow is occurring.

15. Figure 9.22 shows a typical cloud band associated with the ITCZ. It is within this cloud band that disturbances frequently occur. The cloud band, at times, is narrow (2 to 3 degrees latitude) and continuous for thousands of miles. At other times, it is discontinuous and is characterised by a number of large cloud areas 5 to 10 degrees in latitude across. On occasion, vertical cloud patterns are observed within the ITCZ cloud band.

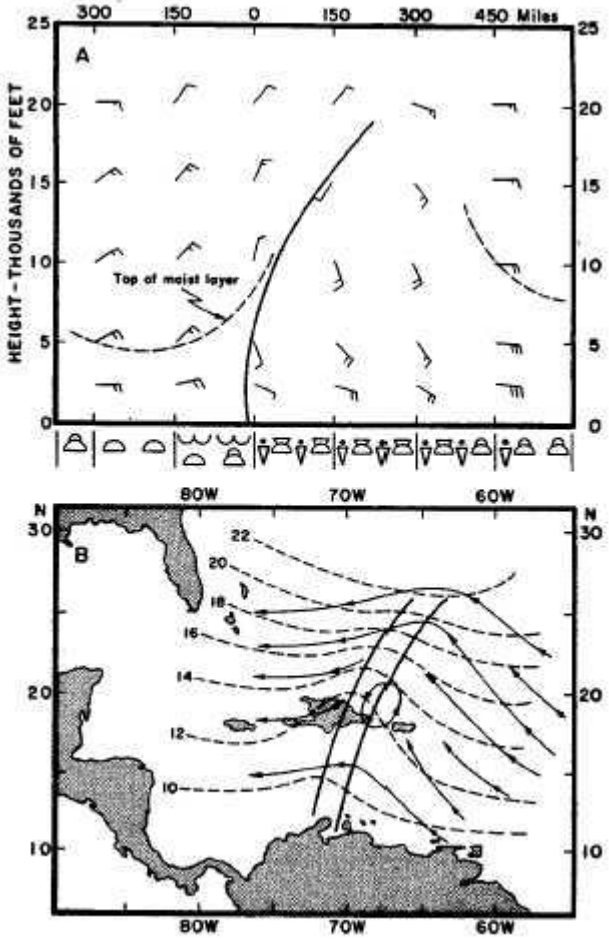


Figure 9.21. Model of an easterly wave, both vertical and horizontal

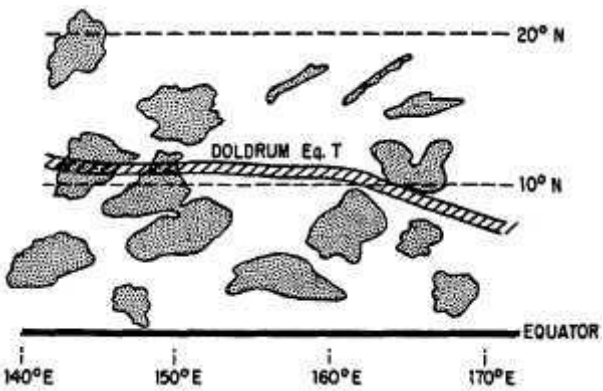


Fig 9.22. Typical satellite-observed cloud cluster patterns relative to a doldrum equatorial trough

16. Generally, disturbances along the ITCZ move from east to west and can move poleward and develop into tropical storms. These disturbances are referred to as "equatorial waves" and are most frequent in the doldrum portions of the equatorial trough. Here, low-level cyclonic wind shear is present over large areas. This, together with friction, produces the forced convergence necessary for the development of the individual cloud systems that form the ITCZ cloud band. Figure 9.23 shows the cloud pattern typical of an active doldrum trough in the Western Pacific. It has been determined that the presence of

surges of airflow from one hemisphere to the other is one of the controlling factors of ITCZ clouds. As air moves across the equator, anti-cyclogenesis takes place. This results in a reduction or clearing of clouds along one portion of the ITCZ cloud band and an intensification of the cloud band in advance of the burst of cross-equatorial flow.

Weather along the ITCZ

17. The degree and severity of the weather along the ITCZ varies, based on the degree of convergence. The zone of disturbed weather may be as little as 20 to 30 miles in width or as much as 300 miles. Under typical conditions, frequent rainstorms, cumulus and cumulonimbus-type clouds, and local thunderstorms occur. Violent turbulence may be associated with these storms, and cloud bases may lower to below 1,000 feet, or even be indistinguishable, in heavy showers. Their tops frequently exceed 40,000 feet. An extensive layer of altocumulus and altostratus usually occurs, due to the spreading out of the upper parts of the convective clouds. These layers vary in height and thickness with the currents of the air masses. At higher levels, a broad deck of cirrostratus spreads out on both sides of the zone. Icing in the heavy cloud masses is likely to reach serious proportions at altitudes above 15,000 feet, as this is roughly the average freezing level in equatorial regions.

18. Surface winds are generally squally in the areas of heavy showers. Usually, the winds in the squalls do not exceed 15 to 25 knots, but winds of 40 to 50 knots or higher have been reported. Visibility is usually good except when reduced by the heavy shower activity.

19. The intensity of the ITCZ varies inter-diurnally, from day to day, and to a lesser degree annually. Over ocean areas, precipitation reaches its maximum just before dawn, with a minimum occurring in the late morning or early afternoon. With the exception of coastal areas under the influence of onshore winds, the reverse is true over land—the daily maximum occurs in the late morning.

Seasonal Variation of ITCZ

20. Through the use of satellite data, the seasonal meridional displacement of the ITCZ band has been determined. See figure 9.24. Imagery has shown the cloud band to have somewhat different characteristics in different parts of the world. In the Atlantic, it is centered about 3°N in the winter season and moves to about 8°N by late summer. In the Pacific, east of 150°W, there is no apparent seasonal fluctuation. Seasonal pressure changes over the North American continent may be responsible for the seasonal shift of the cloud band in the eastern part of the Pacific. There is some evidence of a second cloud band associated with the ITCZ in the Eastern Pacific. Besides the main band north of the equator, a weak band appears at 5°S in the January through March time frame. The strength of this second band varies from year to year; in some instances it fails to develop at all. Over the Indian Ocean, during the Southern Hemisphere summer, the ITCZ cloud band is much broader than it is over either the Atlantic or eastern Pacific Oceans.

Summary

21. Tropical meteorology differs considerably from the meteorology of higher latitudes. Seasonal weather changes (especially temperature) are minimal; the tropics are hot throughout the year. Geographically, this region of Earth is predominantly oceanic, and weather observations are therefore sparse. You must make the most of every observation

and have an understanding of each element's representativeness. Temperature and pressure fields are not nearly as important in the tropics as in higher latitudes, because gradients are so weak. The most significant element is the wind. It reflects the changes in pressure, and as long as you take orographic effects into consideration at island and coastal stations, wind reports take on significant importance with regard to the overall analysis. Wind regimes play a major role in tropical meteorology. The trades are the predominant wind of the tropics. The trade wind belt is found between the horse latitudes and the doldrum belt, and fair weather is the norm in this region. The doldrums is an area of convergent wind flow; the trades of each hemisphere come together in this belt and create widespread convective activity and precipitation. The winds in the doldrums are, for the most part, light and variable, much like the winds of the horse latitudes. The horse latitudes are those regions of Earth that correspond to the location of the subtropical high-pressure belts of each hemisphere. The weather in the horse latitudes is generally fair.

22. Streamline analyses are used in conjunction with surface isobaric and upper-level constant-pressure analyses to adequately portray the tropical atmosphere. Other charts, such as the 24-hour pressure change, weather distribution, and time sections, also play a key role in analysing the weather in this region. Weather features common in tropical analysis are easterly waves, vortices, asymptotes, shear lines, the Intertropical Convergence Zone, tropical cyclones, and the tropical easterly jet stream.

23. Of all tropical weather phenomena, the most destructive and feared is the tropical cyclone. There are three classifications assigned to these low-pressure systems. The first is the tropical depression. It is so designated when at least one closed isobar appears at the surface and winds are less than 34 knots. Advisories on the location and expected development of tropical depressions are numbered. When a depression intensifies and produces sustained winds of 34 to 63 knots, it is upgraded to a tropical storm and named. When sustained winds exceed 63 knots, the cyclone is reclassified as a hurricane or typhoon. Since high winds and seas, heavy rain, severe thunderstorms, and tornadoes accompany these storms, we track their progress in order to provide as much advance warning as possible to ships and shore activities.

Southern Oscillation Index

24. Circulation indices are introduced as simplified measurements of complex states of the atmospheric motion. The Southern Oscillation Index (SOI) is frequently used. Walker & Bliss (1932) described the southern oscillation (SO) as when the surface pressure in the South Pacific Ocean is high, the pressure in the Indian Ocean from Africa to Australia tends to be low. The SO describes conditions associated with low temperatures in the Pacific and Indian Oceans and rainfall amounts that vary inversely with pressure. Walker & Bliss defined SO as a function of the surface pressures at Santiago, Honolulu, Manila and Darwin, and the rainfall in India and Chile, and the Nile Flood. Walker & Bliss found that the SOI is proportional to the Indian rainfall, Nile flood, and the meridional pressure gradient between the subtropical and the sub-equatorial regions, and the zonal pressure gradient between the South Pacific and the Australia regions. Walker & Bliss applied the index to prepare seasonal forecasts of rainfall and pressure in India; however, the verifications are not available. A simplified version of the SOI can be determined from the difference between the mean monthly sea level pressure at Easter Island, Chile and Darwin, Australia. The US National Weather Service (NWS) defined the SOI as the difference between the mean monthly sea level pressure at Tahiti, Archipel de Tahiti and Darwin.

25. Troup (1965) expanded Walker & Bliss' SOI and found that when there is a westerly anomaly over the Indian Ocean with high SOI (i.e., greater than +1 standard deviation) during the period from June to August, the Indian Ocean southeast trades weaken and the Arabian Sea branch of the low-level westerlies strengthen. The South Pacific trades and westerlies are stronger than normal. Troup concluded that the SO can be interpreted as an large-scale air exchange, between eastern and western hemisphere, driven by the temperature difference between the relatively cool Southeast Pacific and the relatively warm West Pacific and Indian Oceans. Troup also found that there are various displacements of eddy-like pressure systems associated with the SO. Troup also stated that the trade wind intensification can produce upwelling in the South American coast, thus the SO is related to the El Nino phenomena.

The El Nino Phenomena

26. The Paita sailors name the counter-current flowing from north to south the El Nino current (El Nino means the Child Jesus in Spanish) because this warmer than normal current has been observed to appear around Christmas, immediately after Christmas, in January, February, or March. The El Nino phenomena is a warm ocean current setting south along the Peruvian coast. The period and magnitude of El Nino phenomena are irregular.

27. The El Nino has been interpretated by conceptual models linking a meteorological control, a oceanic feedback, and a teleconnection. The meteorological control is that an ocean wide weakness of the northern trade winds over a sequence of years would permit a large-than-normal volume of warm water to assemble in the north Eastern Tropical Pacific, increasing the chances for a warm water invasion over ocean areas normally occupied by Peru Current water. The oceanic feedback is that large-scale SST effects on the climate and weather not only locally but also almost globally. The teleconnection is that the regional weather and climate changes have to be viewed as large-scale connections between anomalous in widely separate locations of the global (Bjerknes, 1959, 1961; Philander, 1990; Wuethrich, 1995).

28. An El Nino phenomena has the following characteristics:-

a. During normal years, the coastal Peru is the driest littoral in the world. The Peru Current moves northward from the very cold waters of the Antarctic Circle. The Coriolis effects on the current combined with offshore winds tend to deflect the Peru Current away from the shore, resulting in upwelling. The local SST is about 2°C cooler than the main current. In the north of Peru, north of 6°S, the equatorial counter current set southward from the Gulf of Guayaquil, the current is about 6°C warmer than Peru current and low in salinity. This regular phenomenon in the north of Peru occurs from late December till March. The confluence region between the cold and warm currents is a rapid transition zone climatologically and ecologically.

b. During abnormal years, the warm equatorial countercurrent becomes strong, reaches far south about 12-15°S, and causes displacement of the cold Peru Current away from the coast. During such period the SST in Callao (12°S) may rise to 80°F from the normal, 60°F. At the same time, the ITCZ can move to the desert of northern Peru. For example, in March 1925, the rainfall amount at Trujillo (8°S) was 395 mm, where the average monthly rainfall of the eight preceding years was only 4.4 mm. In addition to the increase of rain in the deserts of Chile and Ecuador, the El

Nino associated with drought at Australia. Such changes in ocean currents can cause catastrophic change in ecology, such as large fluctuations of plankton, fish, bird and human population.

c. The opposite mode of El Nino with an occasional cold tropical ocean current is called La Nina (Philander, 1990). Both El Nino and La Nina may be regarded as the two extremes of the SO as far as the ocean is concerned.

d. Rasmussen & Carpenter (1982) pointed out that the term El Nino has been used to encompass the large-scale features of warming episodes associated with upwelling along the equator and the South American coast. Rasmusson (1984) compiled a list of the 76 onset years of El Nino event during the period from 1726 to 1983. 1987 and 1992 are also El Nino years. Rasmusson found that 67 out of 71 cases since 1803 have a repeating interval with a range of between 1 and 5 years.

El Nino and Southern Oscillation

29. Meteorologically speaking, the El Nino is not a local phenomena. In some cases, El Nino may be related to a change in trade winds in terms of the weakening of the South Pacific anticyclone. In other cases, the warmer equatorial counter-current and the ITCZ are pushed far into the Southern Hemisphere by the southward displacement of the subtropical anticyclones in the Northern Hemisphere. Thus, the El Nino and Southern Oscillation (ENSO) have been treated as one system. Rasmussen & Carpenter (1982) developed a composite picture of the five phases of ENSO; they are pre-onset, onset, peak, transition, and mature phases.

a. **Pre-onset Phase.** This occurs during August-October of the year preceding El Nino. The SSTs in the central and eastern equatorial Pacific (about 150°E-80°W, 15°S-15°N) are about 0.2 to 0.4 °C colder than normal. The surface wind direction anomaly is westerly in the vicinity of the equator, west of the dateline.

b. **Onset Phase.** This occurs during November-January prior to the maximum SST on the Ecuador-Peru coast. The SST around the dateline (170°E-170°W) becomes 0.2 to 0.4 °C warmer than normal. In the same area, the easterly anomaly of surface winds are replaced by westerly anomaly. Most of the eastern equatorial Pacific (east of 170°W) is slightly colder (about 0.2 °C) than in the pre-onset phase.

c. **Peak Phase.** This occurs during March-May of El Nino year. Almost the entire tropical Pacific Ocean is anomaly warm, with the maximum (greater than 1°C) located off the coast of Peru and Chile. The ITCZ in the eastern North Pacific is displaced south of its normal position, but still north of the equator. The westerly anomaly of the western, central and eastern Pacific have increased from those during the onset phase. The trades over the entire Pacific become very weak.

d. **Transition Phase.** This occurs during August-October of the El Nino year. The warm SST anomalies prevail in the entire equatorial Pacific from 160°E to the South American coast, with a zone of maximum temperature along the equator. The ITCZ is located north of the maximum SST zone at about 5-10 degrees latitude. In the central Pacific (160°E-130°W), the maximum SST zone and the low- level convergence zone are collocated. The westerlies in the western equatorial Pacific (130°E-150°W) are well established.

e. **Mature Phase.** This occurs during December-February following the El Nino. The teleconnections between the tropics and the Northern Hemisphere subtropics are strongest during this phase. The temperatures around the zone of maximum sea-surface temperature have been increased another 0.2-0.4 degree C. The positive SST anomalies reach 1.5-2.0 degree C over the central and eastern Pacific. The SSTs along the South American coast have returned to near normal, the easterlies in the eastern equatorial Pacific are stronger than that during the pre-onset phase. The magnitudes of low-level equatorial convergence over the central Pacific are increased over that of the transition phase by about 20 percent.

30. Fraedrich (1988) used a 140-year (1844-1983) record of ENSO observations to show the predictability of ENSO is about one to two years, provided that the initial conditions are reliable. This means that the long-term ENSO predictions are still difficult. Gray (1984) found that the yearly numbers of Atlantic hurricanes are correlated with the El Nino phenomena. Chan (1985) did a similar study, but for the Pacific typhoons. However, the ultimate causes of the El Nino or ENSO effects are not yet satisfactorily established.

Cumulus Convection

31. Cumulus convection is a common phenomenon in the conditionally unstable regions where there are steep lapse rate and deep low- level moist and convergent layer. Convection is one of the effective mechanisms for the vertical transports of heat, moisture and momentum in the troposphere. These transports provide physical links in maintaining the large-scale circulation and the cumulus- scale convection. The term "scale interaction" has been used to describe physical links between motions with different spatial and temporal scales.

32. One example of thermodynamics interactions between cumulus and large-scale circulations is the notion of conditional instability of the second kind; Charney & Eliassen, 1964). This notion was postulated to explain nonlinear intensification of a pre-existed weak low-level cyclonic circulation into a mature tropical cyclone, as a cooperative process between large-scale diabatic heating and cumulus-scale entrainment and detrainment. This notion also offers the basis for many developments of cumulus parameterization schemes for the numerical weather prediction models with much coarser spatial resolution (about 200-400 km) than the mean size of deep cumulus (i.e. 1-10 km). The coarse-resolution models can therefore compute the heat, moisture and momentum transports due to cumulus in terms of model resolvable parameters such as temperature, specific humidity and velocity. Without going into the cumulus-scale spatial resolution, the time step of integration can be made large as long as the numerical model is computational stable. This feature can save computational time of numerical models and the time saved can be used for other computations. So cumulus parameterization scheme is highly desirable for numerical weather prediction model development and application.

TASK-14 : ATMOSPHERIC CIRCULATION

Introduction

1. To understand large-scale motions of the atmosphere, it is essential that the meteorologists study the primary or general circulation of the atmosphere as a whole.

2. The sun's radiation is the energy that sets the atmosphere in motion, both horizontally and vertically. The vertical motion is caused by the rising and expanding of the air when it is warmed, or the descending and contracting of the air when it is cooled. The horizontal motion is caused by differences of atmospheric pressure; air moves from areas of high pressure toward areas of low pressure. Differences of temperature, the cause of the pressure differences, are due to the unequal absorption of the Sun's radiation by Earth's surface. The differences in the type of surface; the differential heating; the unequal distribution of land and water; the relative position of oceans to land, forests to mountains, lakes to surrounding land, and the like, cause different types of circulations of the air. Due to the relative position of Earth with respect to the Sun, much more radiation is absorbed near the equator than at other areas, with the least radiation being absorbed at or near the poles. Consequently, the principal factor affecting the atmosphere is incoming solar radiation, and its distribution depends on the latitude and the season. Chapter discusses general circulation, secondary circulations and tertiary circulations.

General Circulation

3. The general circulation theory attempts to explain the global circulation of the atmosphere with some minor exceptions. Since Earth heats unequally, the heat is carried away from the hot area to a cooler one as a result of the operation of physical laws. This global movement of air which restores a balance of heat on Earth is the general circulation.

World Temperature Gradient

4. Temperature gradient is the rate of change of temperature with distance in any given direction at any point. World temperature gradient refers to the change in temperature that exists in the atmosphere from the equator to the poles.

5. The change in temperature or temperature differential, which causes atmospheric circulation can be compared to the temperature differences produced in a pan of water placed over a gas burner. As the water is heated, it expands and its density is lowered. This reduction in density causes the warmer, less dense water to rise to the top of the pan. As it rises, it cools and is forced to the edges of the pan. Here it cools further and then sinks to the bottom, eventually working its way back to the centre of the pan where it started. This process sets up a simple circulation pattern due to successive heating and cooling.

6. Ideally, the air within the troposphere may be compared to the water in the pan. The most direct rays of the Sun hit Earth near the equator and cause a net gain of heat. The air at the equator heats, rises, and flows in the upper atmosphere toward both poles. Upon reaching the poles, it cools sufficiently and sinks back toward Earth, where it tends to flow along the surface of Earth back to the equator. (See fig. 4.1) Simple circulation of the atmosphere would occur as described above if it were not for the following factors:-

a. Earth rotates, resulting in an apparent force known as the Coriolis force (a deflecting force). This rotation results in a constant change to the area being heated.

b. Earth is covered by irregular land and water surfaces that heat at different rates. Regions under the direct rays of the Sun absorb more heat per unit time than those areas receiving oblique rays. The heat produced by the slanting rays of the Sun during early morning may be compared with the heat that is produced by the slanting rays of the Sun during winter. The heat produced by the more direct rays at midday can be compared with the heat resulting from the more direct rays of summer.

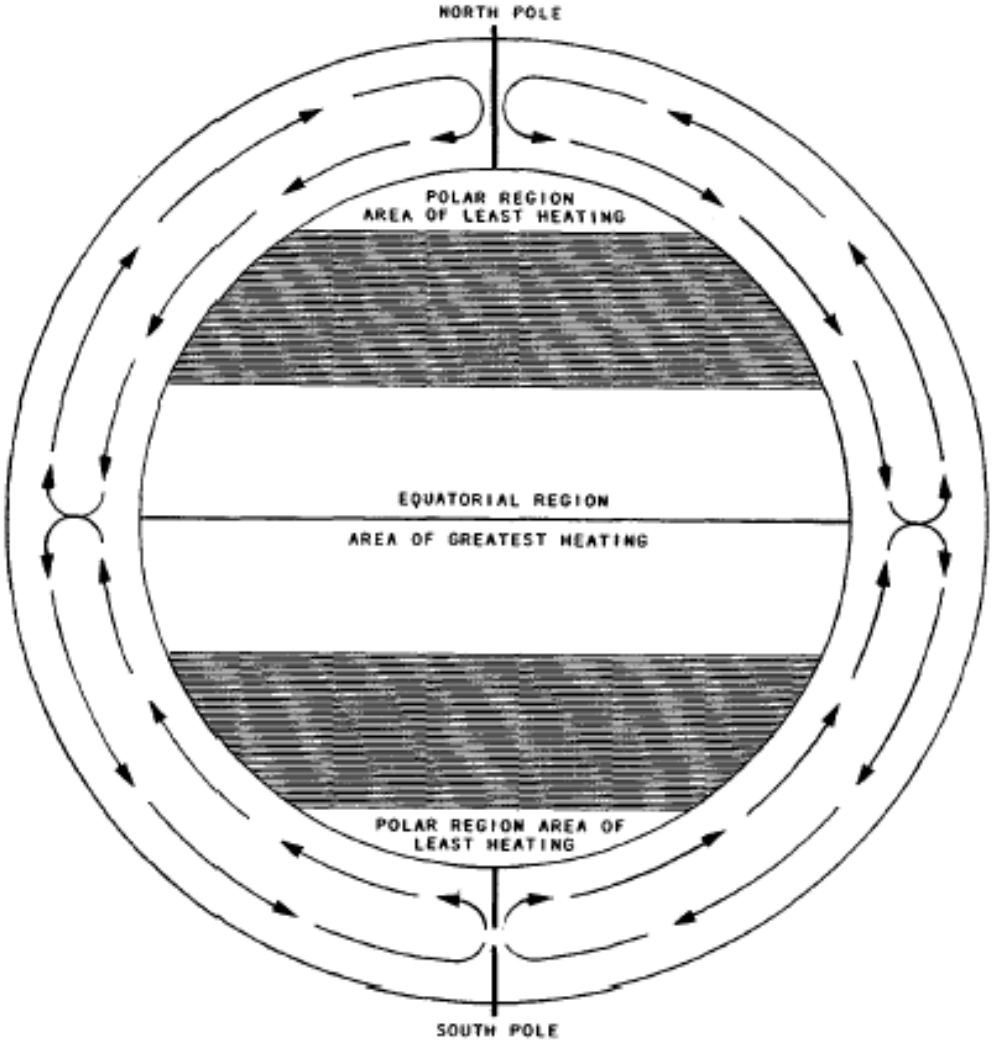


Fig 4.1. Simple Circulation

7. The length of day, like the angle of the Sun's rays, influences the temperature. The length of day varies with the latitude and the season. Near the equator there are about 12 hours of daylight with the Sun's rays striking the surface more directly. Consequently, equatorial regions normally do not have pronounced seasonal temperature variations.

8. During the summer in the Northern Hemi-sphere, all areas north of the equator have more than 12 hours of daylight. During the winter the situation is reversed; latitudes north of the equator have less than 12 hours of daylight. Large seasonal variation in the length of the day and the seasonal difference in the angle at which the Sun's rays reach Earth's surface cause seasonal temperature differences in middle and high latitudes.

9. The weak temperature gradient in the sub-tropical areas and the steeper gradient poleward can be seen in figures 4.2 and 4.3. Note also how much steeper the gradient is poleward in the winter season of each hemisphere as compared to the summer season.

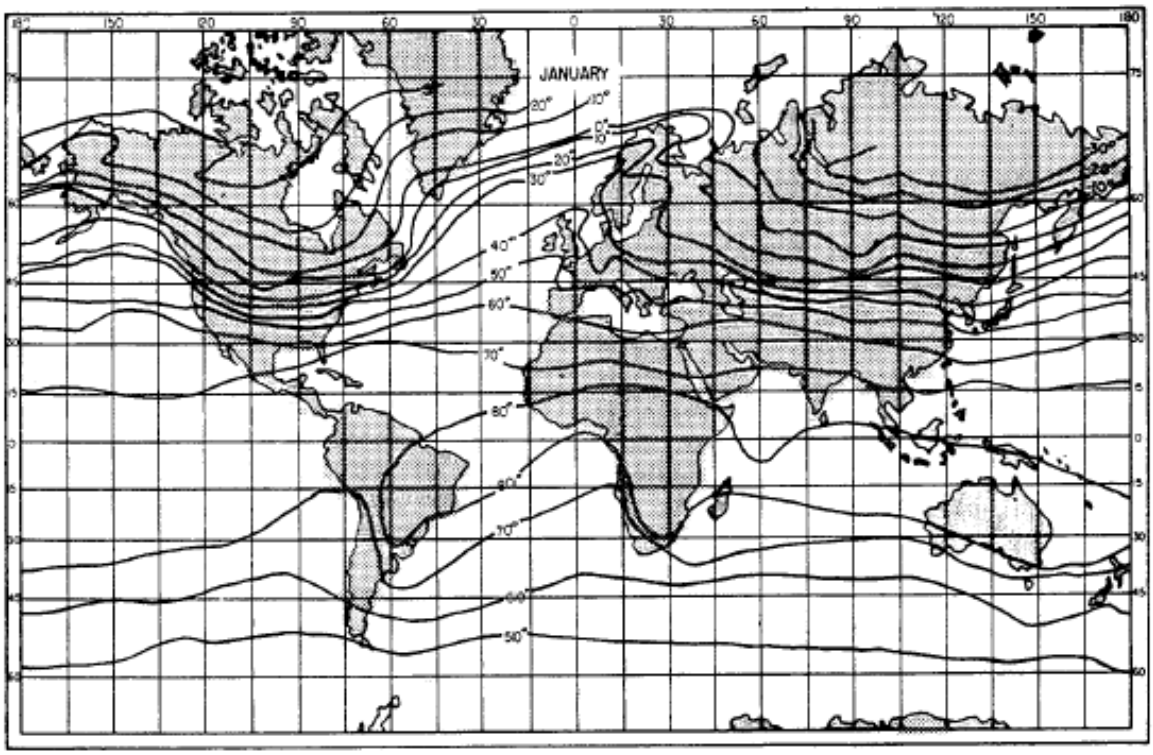


Fig 4.2 Mean World Temperature for January

Pressure over the Globe

10. The unequal heating of Earth's surface due to its tilt, rotation, and differential insolation, results in the wide distribution of pressure over Earth's surface. Study fig 4.2 and 4.3. Note that a low-pressure area lies along the intertropical convergence zone (ITCZ) in the equatorial region. This is due to the higher temperatures maintained throughout the year in this region. At the poles, permanent high-pressure areas remain near the surface because of the low temperatures in this area throughout the entire year. The subtropical high-pressure areas at 30°N and S latitudes are caused mainly by the "piling up" of air in these regions. Other areas are also dominated by relatively high or low pressures during certain seasons of the year.

Elements of Circulation

11. Temperature differences cause pressure differences which in turn cause air movements.

12. The following sections show how air movements work and how they evolve into the various circulations—primary, secondary, and tertiary.

13. To explain the observed wind circulation over Earth, three basic steps are used. The first step is to assume Earth does not rotate and is of uniform surface; that is, all land or all water. The second step is to rotate Earth, but still assume a uniform surface. The third step is to rotate Earth and assume a nonuniform surface. For now, we deal with the first two steps, a nonrotating Earth of uniform surface and a rotating Earth of uniform surface.

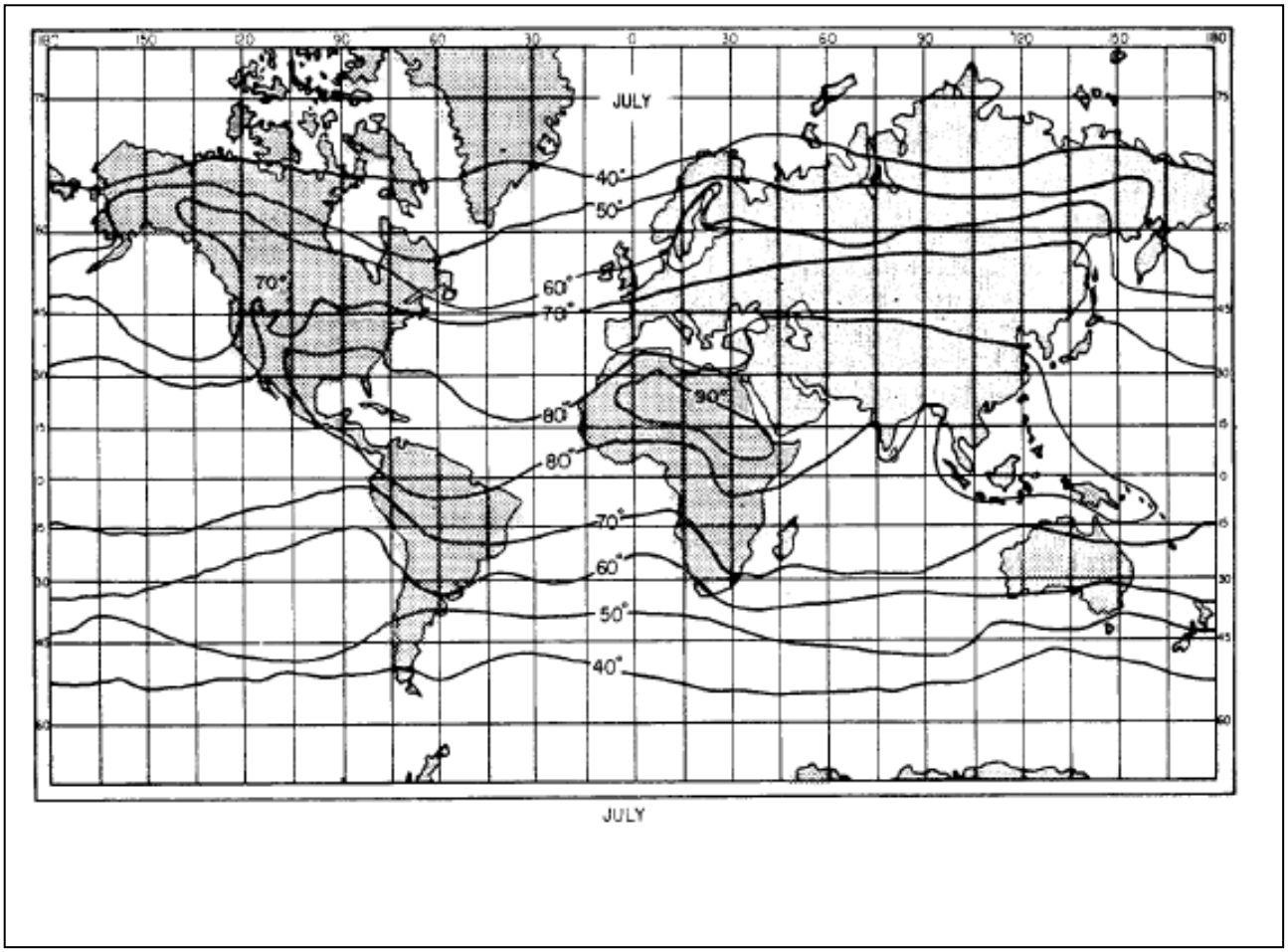


Fig 4.3 Mean World Temperature for July

Static Earth

14. The circulation on a non-rotating Earth is referred to as the thermal circulation because it is caused by the difference in heating. The air over the equator is heated and rises (low pressure), while over the poles the air is cooled and sinks (high pressure). This simple circulation was shown in figure 4.1.

Rotating Earth and Coriolis Force

15. In thermal circulation, the assumption was made that the earth did not rotate, but of course this is not true. The rotation of Earth causes a force that affects thermal circulation, causing it to be deflected to the Right of the direction of movement in the Northern Hemisphere and to the Left of the direction of movement in the Southern Hemisphere. This force is called the Coriolis force.

16. The Coriolis force is not a true force. It is an apparent force resulting from the west-to-east rotation of Earth. The effects, however, are real.

17. Arctic rivers cut faster into their right banks than their left ones. On railroads carrying only one-way traffic, the right hand rails wear out faster than the left-hand rails. Artillery projectiles must be aimed to the left of target because they deflect to the right. Pendulum clocks run faster in high latitudes than in lower latitudes. All these phenomena are the result of the Coriolis force, which is only an apparent force. The most important phenomena is that this force also deflects winds to the right in the Northern Hemisphere. Therefore, it is important to understand how this force is produced. As Earth rotates, points on the surface are moving eastward (from west to east) past a fixed point in space at a given speed. Points on the equator are moving at approximately 1,000 miles per hour, points on the poles are not moving at all, but are merely pivoting, the points somewhere between are moving at speeds between 1,000 and zero miles per hour depending upon their relative position. Refer to view in fig 4.1.

18. Assume that a missile located at the North Pole is launched at a target on the equator. The missile does not have any eastward lateral velocity, but the target has an eastward velocity of 1,000 miles per hour. The result is that the missile appears to be deflected to the right as the target moves away from its initial position. Refer to view fig 4.1.

19. A similar condition assumes that a missile located on the equator is launched at a target at the North Pole. The missile has an eastward lateral velocity of 1,000 miles per hour; while the target on the pole has no lateral velocity at all. Once again the missile appears to be deflected to the right as a result of its initial eastward lateral velocity. Refer to view in fig 4.6.

20. Due to Earth's rotation and the Coriolis effect, the simple circulation now becomes more complex as shown in fig 4.6 & 4.7. The complex circulation resulting from the interplay of the Coriolis effect with the flow of air is known as the 3-cell theory. (See fig. 4.7.)

The 3-Cell Theory

21. According to the 3-cell theory, Earth is divided into six circulation belts—three in the Northern Hemisphere and three in the Southern Hemisphere. The dividing lines are the equator, 30°N and S latitude, and 60°N and S latitude. The three cells of general circulation of the Northern Hemisphere are similar to those of the Southern Hemisphere. (Refer to fig. 4.7 during the following discussion)

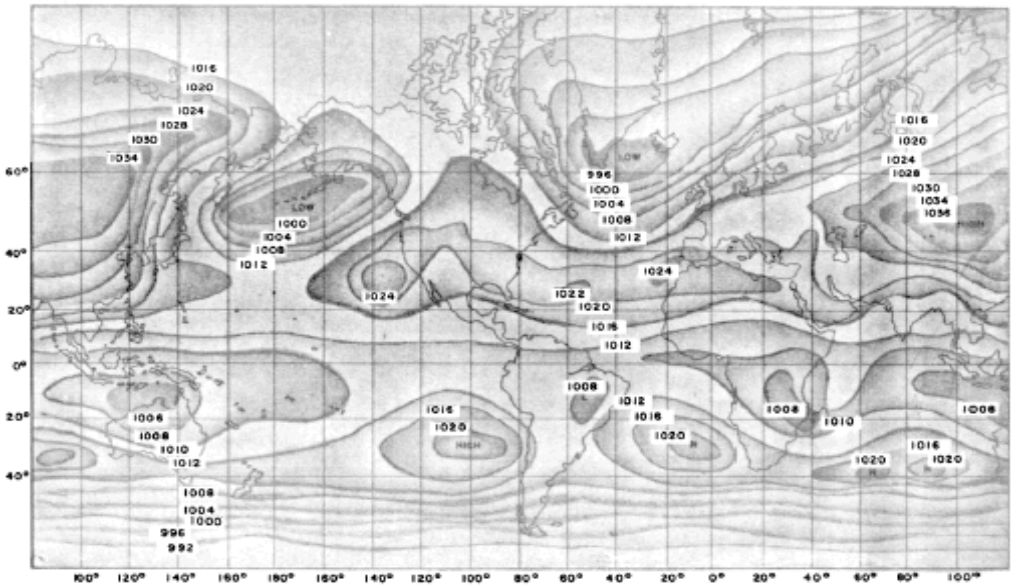


Fig. 4.4 Mean World Pressure for January

22. First, note the tropical cell of the Northern Hemisphere which lies between the equator and 30°N latitude. The air at the equator heats and rises due to convection. When it reaches the upper portions of the troposphere, it tends to flow toward the North Pole. By the time the air has reached 30°N latitude, the Coriolis effect has deflected it so much that it is moving eastward instead of northward. This results in a piling up of air (convergence) near 30°N latitude and a descending current of air (subsidence) toward the surface which forms a belt of high pressure. When the descending air reaches the surface where it flows outward (divergence), part of it flows poleward to become part of the mid-latitude cell; the other part flows toward the equator, where it is deflected by the Coriolis ffect and forms the northeast trades.

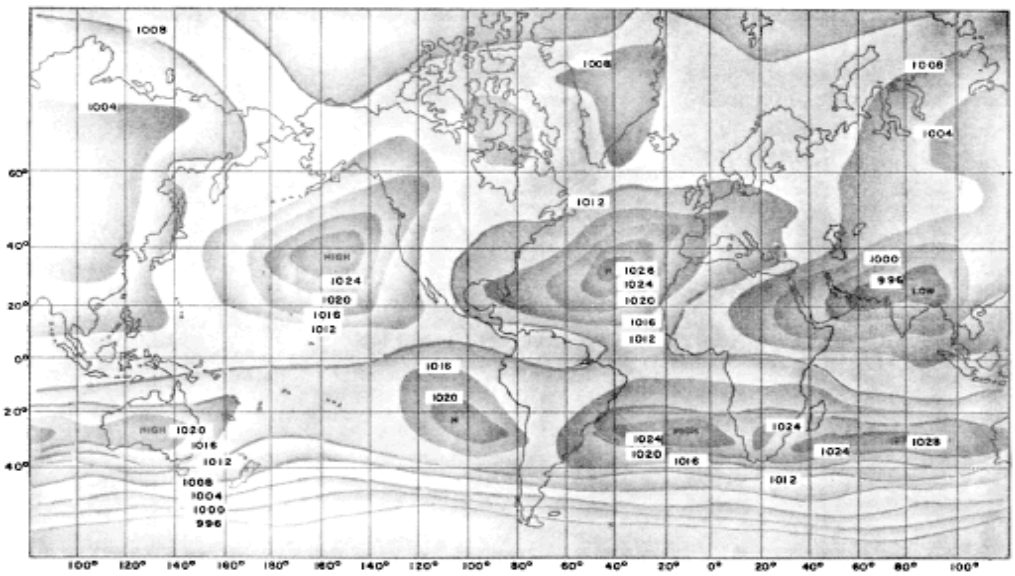


Fig. 4.5 Mean World Pressure for July

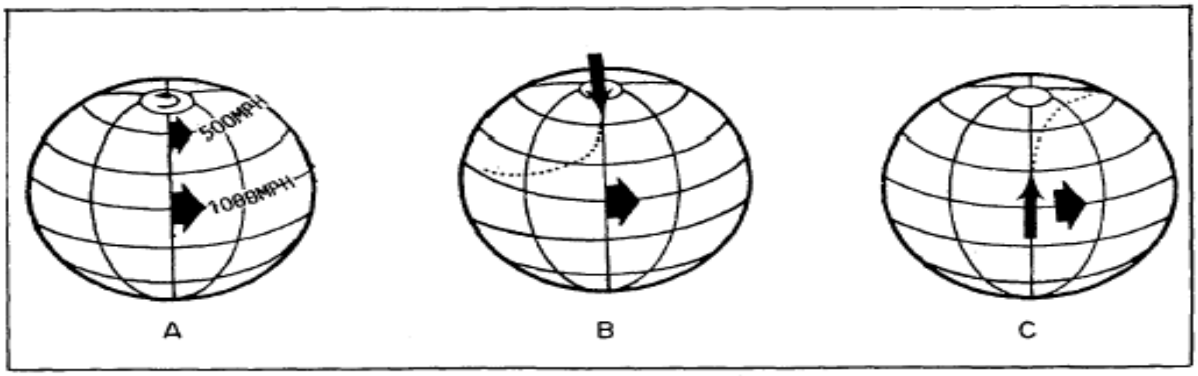


Fig 4.6 Coriolis Force

23. The mid-latitude cell is located between 30° and 60°N latitude. The air which comprises this cell circulates poleward at the surface and equatorward aloft with rising currents at 60° (polar front) and descending currents at 30° (high-pressure belt). However, in general, winds both at the surface and aloft blow from the west. This is easily explained for the surface wind by the Coriolis effect on the poleward-moving surface air. The west wind aloft is not as easily explained. Most meteorologists agree that this wind is frictionally driven by the west winds in the two adjacent cells.

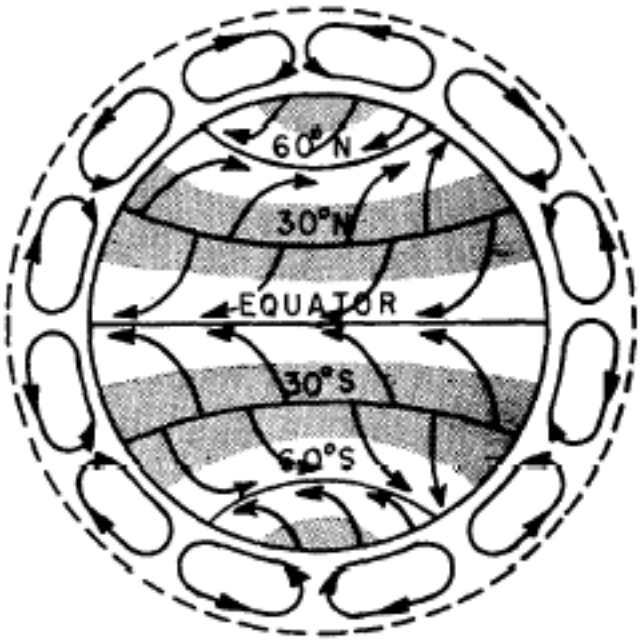


Fig 4.7 Coriolis Effect on Windflow

24. The polar cell lies between 60°N latitude and the North Pole. The circulation in this cell begins with a flow of air at a high altitude toward the pole. This flow cools and descends at the North Pole and forms a high-pressure area in the polar regions. After reaching the surface of Earth, this air tends to flow equatorward and is deflected by the Coriolis effect so that it moves from the northeast. This air converges with the poleward flow from the mid-latitude cell and is deflected upward with a portion circulating poleward again and the remainder equatorward. The outflow of air aloft between the polar and mid-latitude cells causes a semipermanent low-pressure area at approximately 60°N latitude.

25. To complete the picture of the world's general atmospheric circulation, we must associate these prevailing wind and pressure belts with some basic characteristics.

World Winds

26. In the vicinity of the equator is a belt of light and variable winds known as the doldrums. On the poleward side of the doldrums are the Trade Winds; the predominant wind system of the tropics. These easterly winds are the most consistent on Earth, especially over the oceans. Near 30°N and 30°S latitudes lie the sub-tropic high-pressure belts. Winds are light and variable. These areas are referred to as the Horse Latitudes.

27. The prevailing westerlies, which are on the poleward side of the subtropical high-pressure belt, are persistent throughout the mid-latitudes. In the Northern Hemisphere, the direction of the westerlies at the surface is from the southwest. In the Southern Hemisphere, westerlies are from the northwest. This is due to the deflection resulting from the Coriolis effect as the air moves poleward.

28. Poleward of the prevailing westerlies, near 60°N and 60°S latitudes, lies the belt of low pressure known as the polar front zone. Here, converging winds result in ascending air currents and consequent poor weather.

Wind Theory

29. Newton's first two laws of motion indicate that motion tends to be in straight lines and only deviates from such lines when acted upon by another force or by a combination of forces. Air tends to move in a straight line from a high-pressure area to a low-pressure area. However, there are forces that prevent the air from moving in a straight line.

Wind Forces

30. There are four basic forces that affect the directional movement of air in our atmosphere: pressure gradient force (PGF), the Coriolis effect, centrifugal force, and frictional force. These forces, working together, affect air movement. The direction that the air moves is determined by the forces that are affecting it at that particular time. Also, the different names given to the movement of the air (geostrophic wind, gradient wind, etc.) depends on what forces are affecting it.

Pressure Gradient

31. The rate of change in pressure in a direction perpendicular to the isobars is called pressure gradient. Pressure applied to a fluid is exerted equally in all directions throughout the fluid; e.g., if a pressure of 1013.2 millibars is exerted downward by the atmosphere at the surface, this same pressure is also exerted horizontally outward at the surface. Therefore, a pressure gradient exists in the horizontal (along the surface) as well as the vertical plane (with altitude) in the atmosphere.

32. **Horizontal Pressure Gradient**. The horizontal pressure gradient is *steep* or *flat* or when the isobars are far apart. (See fig. 4.4 & 4.5)

33. **Vertical Pressure Gradient**. If isobars are considered as depicting atmospheric topography, a high pressure system represents a hill of air, and a low pressure system represents a depression or valley of air. The vertical pressure gradient always indicates a decrease in pressure with altitude, but the rate of pressure decrease (gradient) varies directly with changes in air density with altitude. Below 10,000 feet altitude, pressure decreases approximately 1 inch of mercury per 1,000 feet in the standard atmosphere. The vertical cross section through a high and low (view in fig. 4.4) depicts the vertical pressure gradient. A surface weather map view of the horizontal pressure gradient in the same high and low is illustrated in view the fig 4.5.

Pressure Gradient Force

34. The variation of heating (and consequently the variations of pressure) from one locality to another is the initial factor that produces movement of air or wind. The most direct path from high to low pressure is the path along which the pressure is changing most rapidly. The rate of change is called the pressure gradient. Pressure gradient force is the force that moves air from an area of high pressure to an area of low pressure. The velocity of the wind depends upon the pressure gradient. If the pressure gradient is strong, the wind speed is high. If the pressure gradient is weak, the wind speed is light. (See fig. 4.8.)

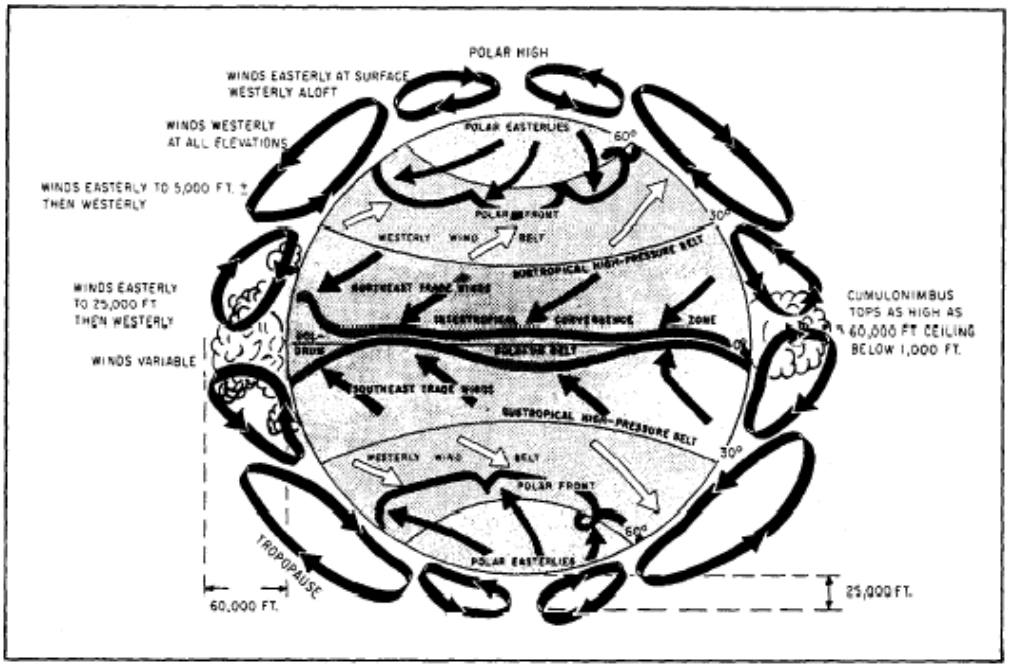


Fig 4.8 Idealised pattern of the general circulation (The 3-cell theory)

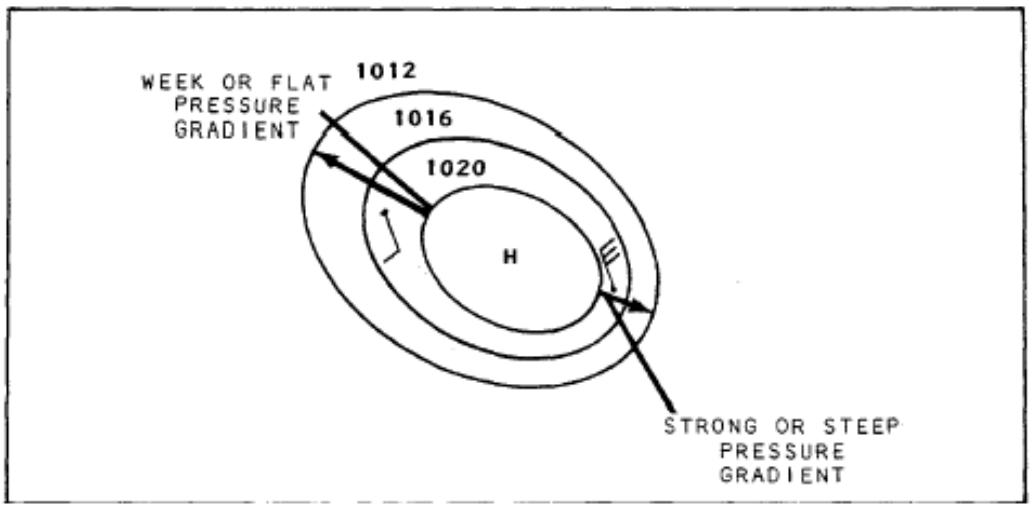


Fig 4.9. Horizontal Pressure Gradient

35. Figure 4.9 shows that the flow of air is from the area of, high pressure to the area of low pressure, but it does not flow straight across the isobars. Instead the flow is circular around the pressure systems. Pressure gradient force (PGF) causes the air to begin moving from the high-pressure to the low-pressure system. Coriolis (deflective) force and centrifugal force then begin acting on the flow in varying degrees. In this example, frictional force is not a factor.

Coriolis Effect

36. If pressure gradient force were the only force affecting windflow, the wind would blow at right angles across isobars (lines connecting points of equal barometric pressure) from high to low pressure. The wind actually blows parallel to isobars above any frictional level. Therefore, other factors must be affecting the windflow; one of these factors is the rotation of Earth. A particle at rest on Earth's surface is in equilibrium. If the particle starts to move because of a pressure gradient force, its relative motion is affected by the rotation of Earth. If a mass of air from the equator moves northward, it is deflected to the right, so that a south wind tends to become a southwesterly wind.

37. In the Northern Hemisphere, the result of the Coriolis effect is that moving air is deflected to the right of its path of motion. This deflection to the right is directly proportional to the speed of the wind; the faster the wind speed, the greater the deflection to the right, and conversely, the slower the wind speed, the less the deflection to the right. Finally, this apparent deflective force is stronger at the polar regions than at the equator.

Centrifugal Force

38. According to Newton's first law of motion, a body in motion continues in the same direction in a straight line and with the same speed unless acted upon by some external force. Therefore, for a body to move in a curved path, some force must be continually applied. The force restraining bodies that move in a curved path is called the *centripetal force;* it is always directed *toward* the centre of rotation. When a rock is whirled around on a string, the centripetal force is afforded by the tension of the string.

39. Newton's third law states that for every action there is an equal and opposite

reaction. Centrifugal force is the reacting force that is equal *to* and opposite in direction to the centripetal force. Centrifugal force, then, is a force directed outward from the centre of rotation.

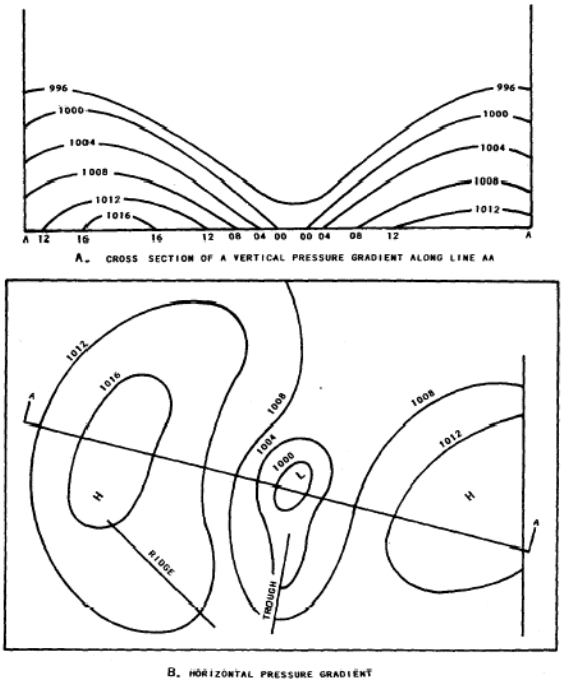


Fig 4.10 Vertical Pressure Gradient

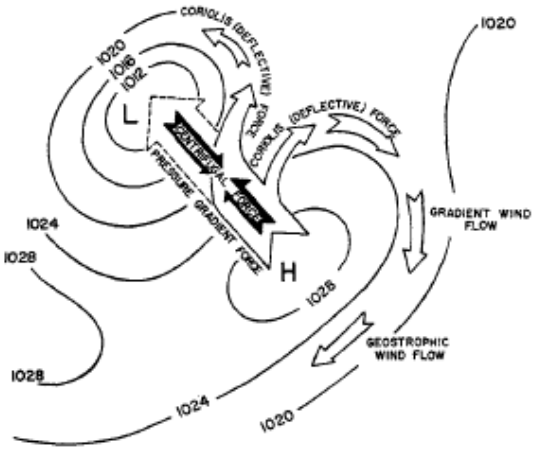


Fig 4.11 Examples of Circulation about High and Low Pressure Systems

40. As you know, a bucket of water can be swung over your head at a rate of speed that allows the water to remain in the bucket. This is an example of both centrifugal and centripetal force. The water is held in the bucket by centrifugal force tending to pull it outward. The centripetal force, the force holding the bucket and water to the centre, is your arm swinging the bucket. As soon as you cease swinging the bucket, the forces cease and the water falls out of the bucket. Figure 3.10 is a simplified illustration of centripetal and centrifugal force.

41. High- and low-pressure systems can be compared to rotating discs. Centrifugal effect tends to fling air out from the centre of rotation of these systems. This force is directly proportional to the wind speed; the faster the wind, the stronger the outward force. Therefore, when winds tend to blow in a circular path, centrifugal effect (in addition to pressure gradient and Coriolis effects) influences these winds.

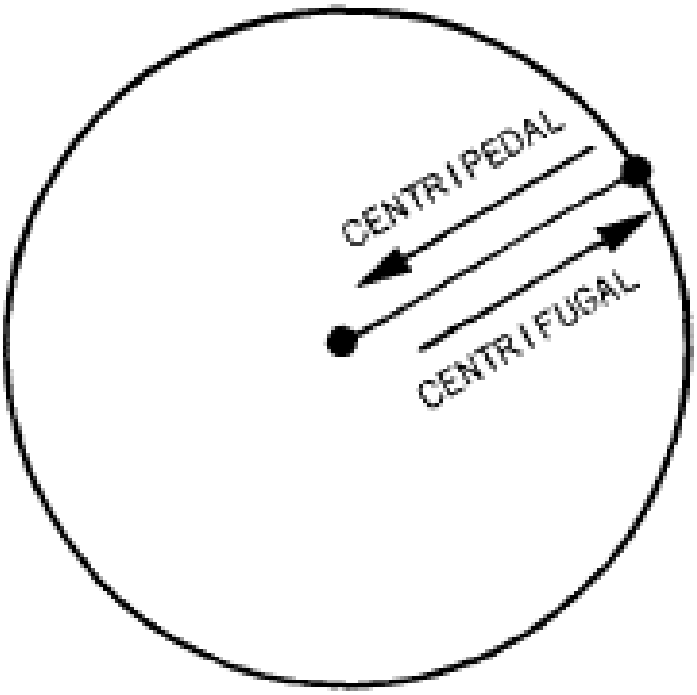


Fig 4.12 Simplified illustration of Centripetal and Centrifugal Force

Frictional Force

42. The actual drag or slowing of air particles in contact with a solid surface is called friction. Friction tends to retard air movement. Since Coriolis force varies with the speed of the wind, a reduction in the wind speed by friction means a reduction of the Coriolis force. This results in a momentary disruption of the balance. When the new balance (including friction) is reached, the air flows at an angle across the isobars from high pressure to low pressure. (Pressure gradient force is the dominant force at the surface.) This angle varies from 10 degrees over the ocean to more than 45 degrees over rugged terrain. Frictional effects on the air are greatest near the ground, but the effects are also carried aloft by turbulence. Surface friction is effective in slowing the wind to an average altitude of 2,000 feet (about 600 meters) above the ground. Above this level, called the gradient wind level or the second standard level, the effect of friction decreases rapidly and may be considered negligible. Air above 2,000 feet normally flows parallel to the isobars.

Wind Types

43. Since there is a direct relationship between pressure gradient and wind speed and direction, we have a variety of wind types to deal with. We discuss below the relationship of winds and circulations, the forces involved, and the effect of these factors on the general circulation.

Geostrophic and Gradient Wind

44. On analysed surface weather charts, points of equal pressure are connected by drawn lines referred to as isobars, while in upper air analysis, points of equal heights are connected and called isoheights.

45. The variation of these heights and pressures from one locality to another is the initial factor that produces movement of air, or wind. Assume that at three stations the pressure is lower at each successive point. This means that there is a horizontal pressure gradient (a decrease in pressure in this case) for each unit distance. With this situation, the air moves from the area of greater pressure to the area of lesser pressure. If the force of the pressure were the only factor acting on the wind, the wind would flow from high to low pressure, perpendicular to the isobars. Since experience shows the wind does not flow perpendicular to isobars, but at a slight angle across them and towards the lower pressure, it is evident that other factors are involved. These other factors are the Coriolis effect, frictional force, and centrifugal effect. When a unit of air moves with no frictional force involved, the movement of air is parallel to the isobars. This wind is called a gradient wind. When the isobars are straight, so only Coriolis and pressure gradient forces are involved, it is termed a geostrophic wind.

46. Let’s consider a parcel of air from the time it begins to move until it develops into a geostrophic wind.

47. As soon as a parcel of air starts to move due to the pressure gradient force, the Coriolis force begins to deflect it from the direction of the pressure gradient force. (See fig). The Coriolis force is the apparent force exerted upon the parcel of air due to the rotation of Earth. This force acts to the right of the path of motion of the parcel of air in the Northern Hemisphere (to the left in the Southern Hemisphere). It always acts at right angles to the direction of motion. In the absence of friction, the Coriolis force changes the direction of motion of the parcel until the Coriolis force and the pressure gradient force are in balance. When the two forces are equal and opposite, the wind blows parallel to the straight isobars (view fig). The Coriolis force only affects the direction, not the speed of the motion of the air. Normally, Coriolis force is not greater than the pressure gradient force. In the case of super-gradient winds, Coriolis force maybe greater than the pressure gradient force. This causes the wind to deflect *more* to the right in the Northern Hemisphere, or toward higher pressure.

48. Under actual conditions, air moves around high and low pressure centres toward lower pressure. Turn back to fig. Here, the flow of air is from the area of high pressure to the area of low pressure, but, as we mentioned previously, it does not flow straight across the isobars (or isoheights). Instead, the flow is circular around the pressure systems.

49. The Coriolis force commences deflecting the path of movement to the right (Northern Hemisphere) or left (Southern Hemisphere) until it reaches a point where a balance exists between the Coriolis and the pressure gradient force. At this point the air is no longer deflected and moves forward around the systems. Once circular motion around the systems is established, then centrifugal force must be considered.

50. Centrifugal force acts outward from the centre of both the highs and the lows with a force dependent upon the velocity of the wind and the degree of curvature of the isobars. However, the pressure gradient force is acting towards the low; therefore, the flow in that direction persists. When the flow is parallel to the curved portion of the analysis in fig, it is a Gradient Wind. When it is moving parallel to that portion of the analysis showing straight flow, it is a Geostrophic Wind.

51. We defined pressure gradient as being a change of pressure with distance. This means that if the isobars are closely spaced, then the pressure change is greater over a given distance; it is smaller if they are widely spaced. Therefore, the closer the isobars, the faster the flow.

52. Geostrophic and gradient winds are also dependent, to a certain extent, upon the density of the atmosphere and the latitude. If the density and the pressure gradient remain constant and the latitude increases, the wind speed decreases. On the other hand, if the latitude decreases, the wind speed increases. If the density and the latitude remain constant and the pressure gradient decreases, the wind speed decreases. If the pressure gradient and the latitude remain constant and the density decreases, the wind speed increases. If the density increases, the wind speed decreases.

53. True geostrophic wind is seldom observed in nature, but the conditions are closely approximated on upper-level charts.

Cyclostrophic Wind

54. In some atmospheric conditions, the radius of rotation becomes so small that the centrifugal force becomes quite strong in comparison with the Coriolis force. This is particularly true in low latitudes where the Coriolis force is quite small to begin with. In this case, the pressure gradient force is nearly balanced by the centrifugal force alone. When this occurs, the wind is said to be cyclostrophic. By definition, a cyclostrophic wind exists when the pressure gradient force is balanced by the centrifugal force alone.

55. This exact situation rarely exists, but is so nearly reached in some situations that the small Coriolis effect is neglected and the flow is said to be cyclostrophic. Winds in a hurricane or typhoon and the winds around a tornado are considered cyclostrophic.

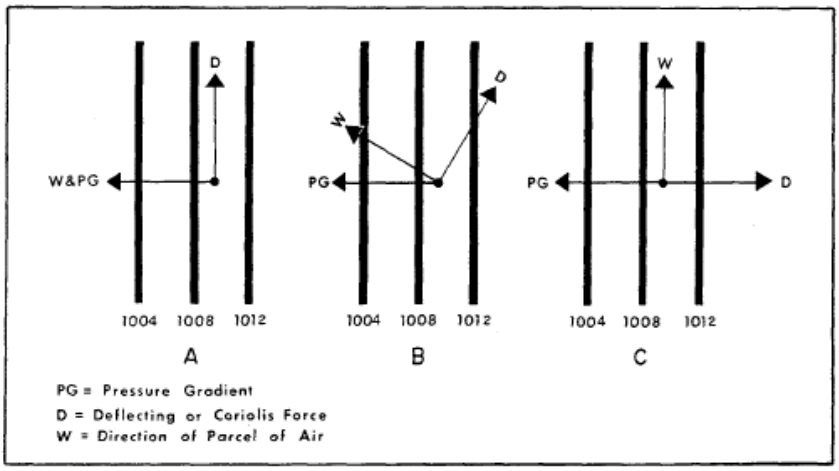


Fig 4.13 Development Cycle of a Geostrophic Wind

Movement of Wind around Anticyclones

56. The movement of gradient winds around anticyclones is affected in a certain manner by the pressure gradient force, the centrifugal force, and the Coriolis force. The pressure gradient force acts from high to low pressure, and the Coriolis force acts opposite to the pressure gradient force and at right angles to the direction of movement of the parcel of air. The centrifugal force acts at right angles to the path of motion and outward from the centre about which the parcel is moving. (See fig) In the case of a high pressure centre, the pressure gradient force and the centrifugal force balance the Coriolis force. This phenomenon may be expressed in the following manner:-

PG + **CF** = **D**

Movement of Wind around Cyclones

57. As in the case of anticyclones, gradient winds around cyclones are affected by the pressure gradient force, the centrifugal force, and the Coriolis force, but the balance of the forces is different. (See fig) In a cyclonic situation the pressure gradient force is balanced by the Coriolis force and the centrifugal force. This balance may be expressed in the following manner:-

PG = **D** + **CF**

58. Centrifugal force acts with the pressure gradient force when the circulation is anticyclonic and against the pressure gradient force when the circulation is cyclonic. Therefore, wind velocity is greater in an anticyclone than in a cyclone of the same isobaric spacing.

Variations

59. It has been determined that, given the same density, pressure gradient, and latitude, the wind is weaker around a low-pressure cell than a high-pressure cell. This is also true for gradient and geostrophic winds. The wind we observe on a synoptic chart is usually stronger around low cells than high cells because the pressure gradient is usually stronger around the low-pressure cell.

Geostrophic and Gradient Wind Scales

60. The geostrophic wind is stronger than the gradient wind around a low and is weaker than a gradient wind around a high. This is why the isobar spacing and contour spacing, for a curved flow, differs from that determined by a geostrophic wind scale. If the flow under consideration is around a high-pressure cell, the isobars are farther apart than indicated by the geostrophic wind scale. If the flow is around a low-pressure cell, the isobars are closer together than indicated by the geostrophic wind scale. Geostrophic and gradient wind scales are used to determine the magnitude of these winds (based on isobar or contour spacing) and to determine the isobar or contour spacing (based on observed wind speeds).

61. There are a number of scales available for measuring geostrophic and gradient flow of both surface and upper air charts.

62. Many weather plotting charts used by the Meteorological Offices have geostrophic wind scales printed on them for both isobaric and contour spacing.

63. The most common scales in general use can be used for both surface and upper air charts. The scales are in 4 mb and 60 mb intervals. An example of a geostrophic wind scale is shown in fig. Note that latitude accounts for the increases in gradients. In tropical regions, the geostrophic wind scales become less reliable because pressure gradients are generally rather weak.

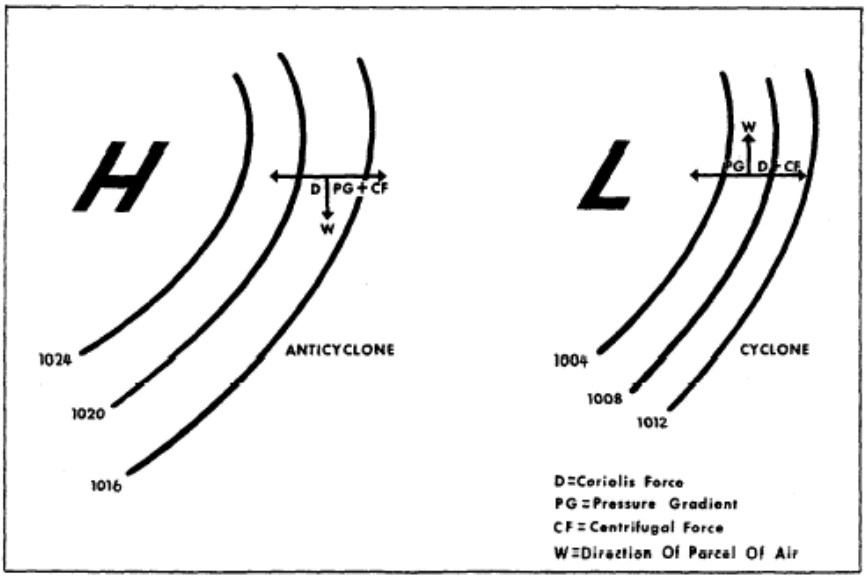


Fig 4.14 Forces acting on Pressure Systems

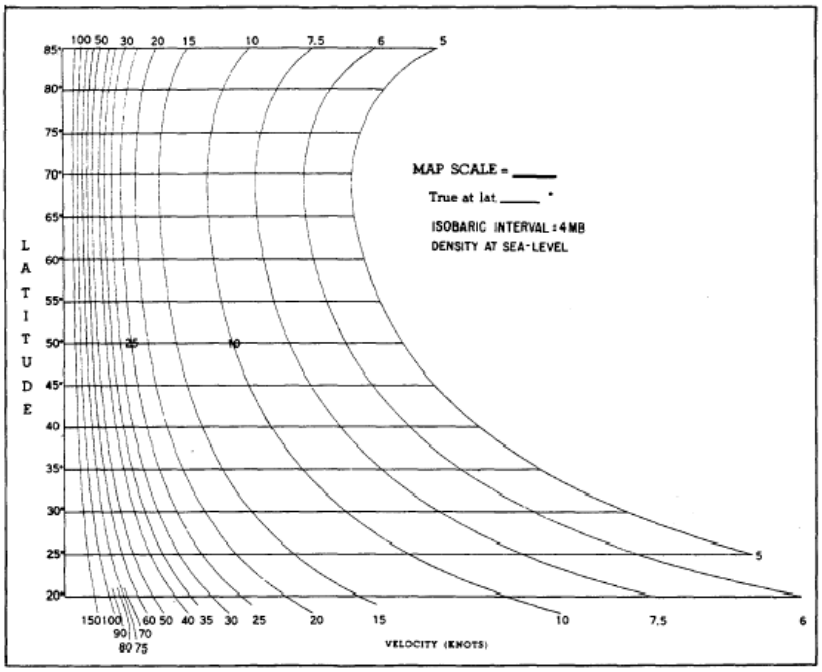


Fig 4.15 Geostrophic Wind Scale

Secondary Circulation

64. Now that you have a picture of the general circulation of the atmosphere over Earth, the next step is to see how land and water areas offset the general circulation. The circulations caused by the effect of Earth's surfaces, its composition and contour, are known as secondary circulations. These secondary circulations give rise to winds that often cancel out the normal effect of the great wind systems.

65. There are two factors that cause the pressure belts of the primary circulation to break up into closed circulations of the secondary circulations. They are first, the nonuniform surface of the earth and second, the difference in heating and cooling of land and water. The surface temperature of oceans changes very little during the year. However, land areas sometimes undergo extreme temperature changes with the seasons. In the winter, large high-pressure areas form over the cold land and low-pressure areas form over the relatively warm oceans. The reverse is true in summer when highs are over water and lows form over the warm land areas. The result of this difference in heating and cooling of land and water surfaces is known as the thermal effect.

66. Circulation systems are also created by the interaction of wind belts of pressure systems or the variation in wind in combination with certain distributions of temperature and/or moisture. This is known as the dynamic effect. This effect rarely, if ever, operates alone in creating secondary systems, as most of the systems are both created and maintained by a combination of the thermal and dynamic effects.

Centres of Action

67. The pressure belts of the general circulation are rarely continuous. They are broken up into detached areas of high and low pressure cells by the secondary circulation. The breaks correspond with regions showing differences in temperature from land-to water surfaces. (Turn back to fig. Compare the temperature distribution in views A and B of figures 3.4 to the pressure distribution in views fig) Note the gradient over the Asian Continent in January. Compare it to the warmer temperature over the ocean and coastal regions. Now look at view A of fig and note the strong region of high pressure corresponding to the area. Now look at the same area in July. Note the way the temperature gradient flattens out and warms. Look at view fig and see the low-pressure area which has replaced the high-pressure region of winter. These pressure cells tend to persist in a particular area and are called centres of action; that is, they are found at nearly the same location with somewhat similar intensity during the same month each year.

68. There is a permanent belt of relatively low pressure along the equator and another deeper belt of low pressure paralleling the coast of the Antarctic Continent. Permanent belts of high pressure largely encircle Earth, generally over the oceans in both the Northern and Southern Hemispheres. The number of centres of action are at a maximum at about 30 to 35 degrees from the equator.

69. There are also regions where the pressure is predominantly low or high at certain seasons, but not throughout the year.

70. In the vicinity of Iceland, pressure is low most of the time. The water surface is warmer (due to warm ocean currents) than the surface of Iceland or the icecaps of Greenland. The Icelandic low is most intense in winter, when the greatest temperature contrast occurs, but it persists with less intensity through the summer. Near Alaska, a similar situation exists with the Aleutian low. The Aleutian low is most pronounced when the neighboring areas of Alaska and Siberia are snow covered and colder than the adjacent ocean. These lows are not a continuation of one and the same cyclone. They are, however, regions of low pressure where lows frequently form or arrive from other regions. Here they remain stationary or move sluggishly for a time, then the lows move on or die out and are replaced by others. Occasionally these regions of low pressure are invaded by traveling high-pressure systems.

71. Two areas of semipermanent high pressure also exist. There is a semipermanent high-pressure centre over the Pacific westward of California and another over the Atlantic, near the Azores and off the coast of Africa. Pressure is also high, but less persistently so, west of the Azores to the vicinity of Bermuda. These subtropical highs are more intense and cover a greater area in summer than in winter. They also extend farther northward in summer. In winter, these systems move south toward the equator, following the solar equator.

72. The largest individual circulation cells in the Northern Hemisphere are the Asiatic high in winter and the Asiatic low in summer. In winter, the Asiatic continent is a region of strong cooling and therefore is dominated by a large high-pressure cell. In summer, strong heating is present and the high-pressure cell becomes a large low-pressure cell. (See fig) This seasonal change in pressure cells gives rise to the monsoonal flow over India and Southeast Asia. Another cell that is often considered to be a centre of action is the polar high. Both Arctic and Antarctic highs have considerable variations in pressure, and these regions have many traveling disturbances in summer. For example, the Greenland high (due to the Greenland icecap) is a persistent feature, but it is not a well-defined high during all seasons of the year. The Greenland high often appears to be an extension of the polar high or vice versa.

73. Other continental regions show seasonal variations, but are generally of small size and their location is variable. Therefore, they are not considered to be centres of action.

74. An annual average pressure distribution chart (fig. 4.16) reveals several important characteristics. First, along the equator there is a belt of relatively low pressure encircling the globe with barometric pressure of about 1,012 hPa. Second, on either side of this belt of low pressure is a belt of high pressure. This high pressure area in the Northern Hemisphere lies mostly between latitudes 30° and 40°N with three well-defined centres of maximum pressure. One is over the eastern Pacific, the second over the Azores, and the third over Siberia; all are about 1,020 hPa. The belt of high pressure in the Southern Hemisphere is roughly parallel to 30°S. It, too, has three centres of maximum pressure. One is in the eastern Pacific, the second in the eastern Atlantic, and the third in the Indian Ocean; again, all are about 1,020 hPa. A third characteristic to be noted from this chart is that, beyond the belt of high pressure in either hemisphere, the pressure diminishes toward the poles. In the Southern Hemisphere, the decrease in pressure toward the South Pole is regular and very marked. The pressure decreases from an average slightly above 1,016 hPa along latitude 35°S to an average of 992 hPa along latitude 60°S. In the Northern Hemisphere, however, the decrease in pressure toward the North Pole is less regular and not as great. This is largely due to the distribution of land and water: note the extensive landmasses in the Northern Hemisphere as compared to those of the Southern Hemisphere.

75. While the pressure belts that stand out on the average annual pressure distribution chart represent average pressure distribution for the year, these belts are rarely continuous on any given day. They are usually broken up into detached areas of high or low pressure by the secondary circulation of the atmosphere. In either hemisphere, the pressure over the land during the winter season is decidedly above the annual average. During the summer season, the pressure is decidedly below the average, with extreme variations occurring such as in the case of continental Asia. Here the mean monthly pressure ranges from about 1,033 hPa during January to about 999 hPa during July. Over the northern oceans, on the other hand, conditions are reversed; the summer pressure there is somewhat higher. Thus in January the Icelandic and Aleutian lows intensify to a depth of about 999 hPa, while in July these lows fill and are almost obliterated. The polar high in winter is not a cell centered directly over the North Pole, but appears to be an extension of the Asiatic high and often appears as a wedge extending from the Asiatic continent. The cell is displaced toward the area of coldest temperatures—the Asiatic continent. In summer, this high appears as an extension of the Pacific high and is again displaced toward the area of coolest temperature, which in this case is the extensive water area of the Pacific.

76. In winter over North America, the most significant feature is the domination by a high-pressure cell. This cell is also due to cooling but is not as intense as the Asiatic cell. In summer, the most significant feature is the so-called heat low over the southwestern part of the continent, which is caused by extreme heating in this region.

Migratory Systems

77. General circulation, based on an average of wind conditions, is a more or less quasi-stationary circulation. Likewise, much of the secondary circulation depends on more or less static conditions that, in turn, depend on permanent and semipermanent high- and low-pressure areas. Changes in the circulation patterns discussed so far have been largely seasonal. However, secondary circulation also includes wind systems that migrate constantly, producing rapidly changing weather conditions throughout all seasons, especially in the middle latitudes. The migratory circulation systems are associated with air masses, fronts, cyclones, and anticyclones.

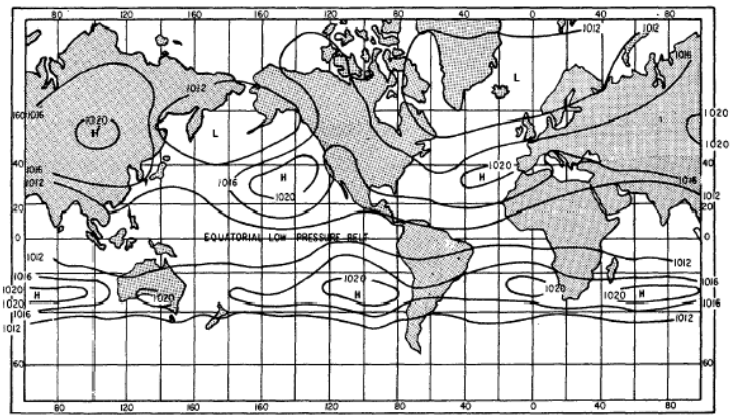


Fig 4.16 Average Annual Pressure Distribution Chart

Anticyclones

78. An anticyclone (high) is an area of relatively high pressure with a clockwise flow (wind circulation) in the Northern Hemisphere and counter-clockwise flow in the Southern Hemisphere. The windflow in an anticyclone is slightly across the isobars and away from the centre of the anticyclone. (See fig) Anticyclones are commonly called highs or high-pressure areas.

79. The formation of an anticyclone or the intensification of an existing one is called Anticyclogenesis. Anticyclogenesis refers to the development of anticyclonic circulation as well as the intensification of an existing anticyclonic flow. When a high-pressure centre is increasing in pressure, the high is building or intensifying. Although a high can build (or intensify) without an increase in anticyclonic flow, it is rare. Normally, building and anticyclogenesis occur simultaneously.

80. The weakening of anticyclonic circulation is anticyclolysis. When the pressure of a high is decreasing, we say the high is weakening. Anticyclolysis and weakening can occur separately, but usually occur together. The vertical extent of pressure greatly depends on the air temperature. Since density increases with a decrease in temperature, pressure decreases more rapidly vertically in colder air than in warmer air.

81. In a cold anticyclone (such as the Siberian high), the vertical extent is shallow; while in a warm anticyclone (such as the subtropical high), the vertical extent reaches high into the upper atmosphere due to the slow decrease in temperature with elevation.

Cyclones

82. A cyclone (low) is a circular or nearly circular area of low pressure with a counterclockwise flow. The flow is slightly across the isobars toward the centre in the Northern Hemisphere and clockwise in the Southern Hemisphere. (See fig. 3.16.) It is commonly called a low or a depression. This use of the word cyclone should be distinguished from the colloquial use of the word as applied to the tornado or tropical cyclone (hurricane). The formation of a new cyclone or the intensification of the cyclonic flow in an existing one is called cyclogenesis. When the pressure in the low is falling, we say the low is deepening. Cyclogenesis and deepening can also occur separately, but usually occur at the same time. The decrease or eventual dissipation of a cyclonic flow is called cyclolysis. When the pressure in a low is rising, we say the low is filling. Cyclolysis and filling usually occur simultaneously.

83. Cyclones in middle and high latitudes are referred to as extratropical cyclones. The term *cyclone* refers to hurricanes and typhoons.

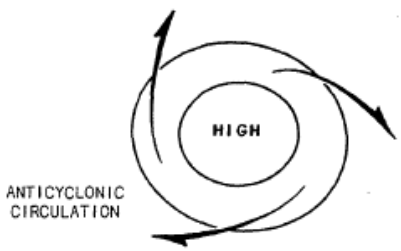


Fig 4.17 Anticyclone

Vertical Structure of Secondary Circulations (Pressure Centres)

84. To better understand the nature of the pressure centres of the secondary circulation, it is necessary to consider them from a three-dimensional standpoint. With the aid of surface and upper air charts, you will be able to see the three dimensions of these pressure systems as well as the circulation patterns of the secondary circulation as established at higher levels in the troposphere and lower stratosphere.

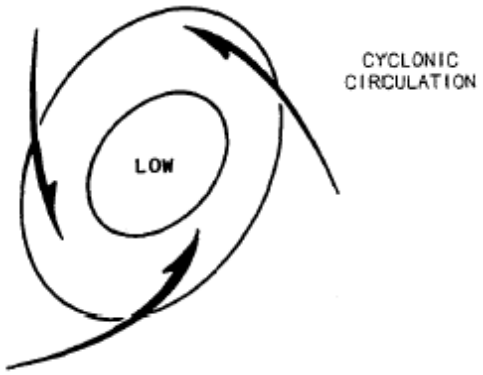


Fig 4.18 Cyclone

85. The study of gas laws showed that volume is directly proportional to temperature. Stated another way, we might say that the thickness of a layer between two isobaric surfaces is directly proportional to the mean virtual temperature of the layer. Because the atmosphere is always moist to some degree, virtual temperature is used. Mean virtual temperature is defined as the average temperature at which dry air would have the same pressure and density as moist air. Thus, lines representing thickness are also isotherms of mean virtual temperature. The higher the mean virtual temperature, the thicker the layer, or vice versa. The thickness between layers is expressed in geopotential meters. The shift in location, as well as the change of shape and intensity upward of atmospheric pressure systems, is dependent on the temperature distribution.

86. An example of the effects of virtual temperature can be demonstrated by two columns of air placed side by side—one cold and one warm. The constant pressure surfaces in the cold column are closer together than those in the warm column of air. Figure 3.17 shows an increase in thickness between two given pressure surfaces for an increase in mean virtual temperature. Note the increase in the distance between the constant pressure surfaces; P, P1, etc., from column A to column B.

87. The thickness between two pressure surfaces can be derived by using the hypsometric equation. Thickness may also be determined from tables, graphs, etc.

Vertical Structure of High-Pressure Systems

88. The topographic features that indicate the circulation patterns at 500 millibars in the atmosphere correspond in general to those at lower and higher levels. However, they may experience a shift in location as well as a change in intensity and shape. For example, a closed high on a surface synoptic chart may be reflected by a ridge aloft. In addition, upper air circulation patterns may take on a wavelike structure in contrast to the alternate closed lows, or closed high patterns at the surface level. The smoothing of the circulation pattern aloft is typical of atmospheric flow patterns.

Cold Core Highs

89. A cold core high is one in which the temperatures on a horizontal plane decrease toward the centre. Because the temperature in the centre of a cold core high is less than toward the outside of the system, it follows that the vertical spacing of isobars in the centre of this system is closer together than on the outside. Although the pressure at the centre of these systems on the surface may be high, the pressure decreases rapidly with height. (See fig. 3.18) Because these highs are often quite shallow, it is common for an upper level low to exist above a cold core high. Note: For the purpose of illustration, fig are exaggerated with respect to actual atmospheric conditions.

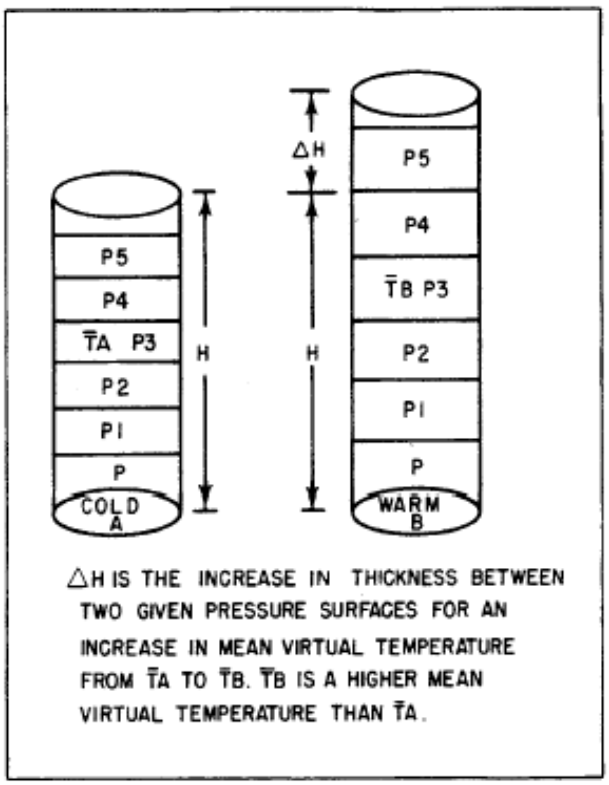


Fig 4.19 Thickness of two strata as a function of Mean Virtual Temperature

90. If the cold core high becomes subjected to warming from below and to subsidence from aloft, as it moves southward from its source and spreads out, it diminishes rapidly in intensity with time (unless some dynamic effect sets in aloft over the high to compensate for the warming). Since these highs decrease in intensity with height, thicknesses are relatively low. In the vertical, cold core highs slope toward colder air aloft. Examples of cold core highs are the North American High and the Siberian High and the migratory highs that originate from these anticyclones.

Warm Core Highs

91. A warm core high is one in which the temperatures on a horizontal level increase toward the centre. Because the temperatures in the centre of a warm core high are higher than on the outside of the system, it follows that the vertical spacing of isobars in the centre is farther apart than toward the outside of the high. For this reason, a warm core high increases in intensity with altitude and has an anticyclonic circulation at all levels (see fig). In the vertical, warm core highs slope toward warmer air aloft.

Examples of warm core highs are the Azores or Bermuda High and the Pacific High.

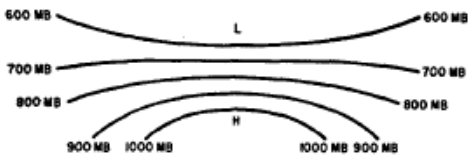


Fig 4.20 Cold Core High

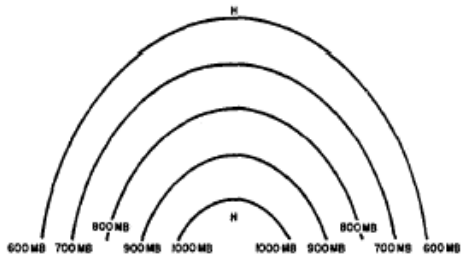


Fig 4.21 Warm Core High

Vertical Structure of Low-Pressure Systems

92. Low pressure systems, like high pressure systems, are generally a reflection of systems aloft. They, too, experience shifts in location and changes in intensity and shape with height. At times, a surface system may not be evident aloft and a well developed system aloft may not reflect on a surface analysis.

Cold Core Lows

93. The cold core low contains the coldest air at its centre throughout the troposphere; that is, going radially outward in any direction at any level in the troposphere, warmer air is encountered. The cold core lows (fig. 4.22) increase in intensity with height. Relative minimums in thickness values, called cold pools, are found in such cyclones. The temperature distribution is almost symmetrical, and the axis of the low is nearly vertical. When they do slope vertically, they slope toward the coldest temperatures aloft. In the cold low, the lowest temperatures coincide with the lowest pressures.

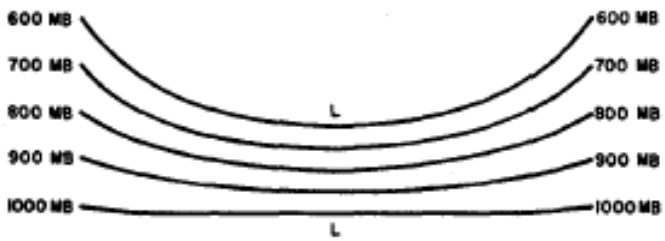


Fig 4.22 Cold Core Low

94. The cold low has a more intense circulation aloft from 850 to 400 millibars than at the surface. Some cold lows show only a slight evidence in the surface pressure field that an intense circulation exists aloft. The cyclonic circulation aloft is usually reflected on the surface in an abnormally low daily mean temperature and in precipitation and unstable hydrometeors.

95. At high latitudes the cold pools and their associated upper air lows show some tendency for location in the northern Pacific and Atlantic Oceans where, statistically, they contribute to the formation of the Aleutian and Islandic lows.

Warm Core Lows

96. A warm core low (fig. 4.23) decreases in intensity with height and the temperature increases toward the centre on a horizontal plane. The warm low is frequently stationary, such as the heat low over the Northwest India and southwestern United States in the summer; this is a result of strong heating in a region usually insulated from intrusions of cold air that tend to fill it or cause it to move. The warm low is also found in its moving form as a stable wave moving along a frontal surface. There is no warm low aloft in the troposphere. The tropical cyclone, however, is believed to be a warm low because its intensity diminishes with height. Because most warm lows are shallow, they have little slope. However, intense warm lows like the heat low over the southwest United States and hurricanes do slope toward warm air aloft. In general, the temperature field is quite asymmetrical around a warm core cyclone. Usually the southward moving air in the rear of the depression is not as warm as that moving northward in advance of it.

Dynamic Low

97. A dynamic low is a combination of a warm surface low and a cold upper low or trough, or a warm surface low in combination with a dynamic mechanism aloft for producing a cold upper low or trough. It has an axis that slopes toward the coldest tropospheric air. (See fig. 4.23) In the final stage, after occlusion of the surface warm low is complete, the dynamic low becomes a cold low with the axis of the low becoming practically vertical.

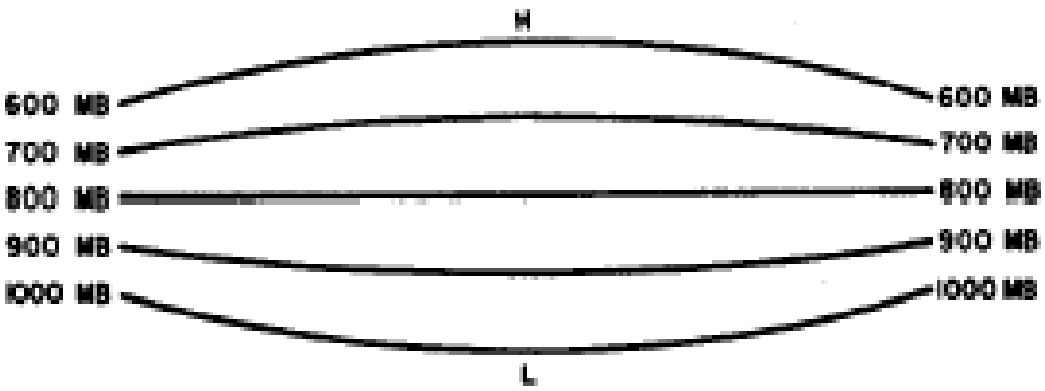


Fig 4.23 Warm Core Low

Dynamic High

98. The dynamic high is a combination of a surface cold high and an upper-level warm high or well-developed ridge, or a combination of a surface cold high with a dynamic mechanism aloft for producing high-level anticyclogenesis. Dynamic highs have axes that slope toward the warmest tropospheric air. (See fig. 4.21) In the final stages of warming the cold surface high, the dynamic high becomes a warm high with its axis practically vertical.

CHAPTER-4 : ANALYSIS PROGNOSIS AND FORECASTING

TASK – 15 : STREAMLINE ANALYSIS

Plotting

1. Wind arrows and barbs are plotted and the wind direction and speed written in parentheses alongside the barb. Streamline charts are usually plotted every 12 hours, but data from off-time periods are used to supplement the 12-hourly data.

Streamline Analysis

2. The weak horizontal pressure and temperature gradients of the tropics do not adequately explain much of the weather occurring there. Therefore, another element is used to better explain tropical weather patterns. That element is the wind. An analysis of the wind field is called a streamline analysis. A tropical analysis is a continuous representation of the wind field constructed from plotted wind observations and data gathered from satellite imagery.

3. **Wind Estimates from Satellite Pictures.** Clouds over the tropics, viewed from satellite altitude, reveal many features of the flow. The distribution of widespread cloud systems has definite relationships to major trough and ridge positions, and wind estimates for both upper and lower tropospheric levels can be obtained from analysis of cumulus and cirrus cloud formations as seen in singular satellite pictures or from cloud motion measurements taken from a series of pictures.

4. Two levels of the tropical atmosphere are usually plotted and analysed: in the lower troposphere, it is the gradient level, and in the upper troposphere, it is the 300-or 200-hPa level.

5. **Low-Level Streamlines**. Gradient level (2,000 to 3,000 feet) winds are most often chosen for analysis, because the wind frequency is fairly equal, and since cloud bases average 2,000 feet, balloons are often lost above this level. Ship's surface winds can also be used to supplement the winds of this field, and satellite pictures permit wind determination from low-level cloud fields.

6. Research has shown that many cumulus cloud lines observed between 10°N and 10°S over oceans are oriented in approximately the same direction as the surface wind. However, because the orientation of cloud lines relates to wind shear, these lines may depart considerably from the surface wind direction. Therefore, a great deal of caution must be exercised when using cloud lines to define low-level wind flow.

7. There are no hard-and-fast rules that permit positive identification of the particular cloud lines that approximately parallel the surface wind. It has been observed that cloud lines that are very long, very narrow, and either wavy, zig-zag, or have knots (wide places along a line) are the type that most often parallel the surface wind. Fig 4.1 shows cloud lines that approximately parallel the wind flow. Surface wind reports are entered on the picture as white arrows.

8. Two drawbacks to using the gradient level are ‘a’ the orographic influences and ‘b’ in the trade wind belt, tropical disturbances are not readily apparent at this level; the wind fluctuations at 2,000 feet may be small and non-informative. Thus, the choice of level to be analysed must be based on local and synoptic peculiarities of the area.

9. **Upper-Level Streamlines**. All the rules applicable to low-level streamlines also apply to the streamlines of the upper troposphere. Upper-level streamline analyses are primarily used to locate areas of cyclonic and anticyclonic flow. Anticyclonic outflow areas are the most important areas to look for, because they are your first clue to tropical cyclone development aloft. You will find, more often than not, that upper-level wind data (200-hPa level, or 40,000 ft) diminishes to less than half that available for low-level analyses. In view of this, you will find qualitative streamline analysis more suitable for this level in most cases, and that satellite pictures play an, even more important part in the analysis.

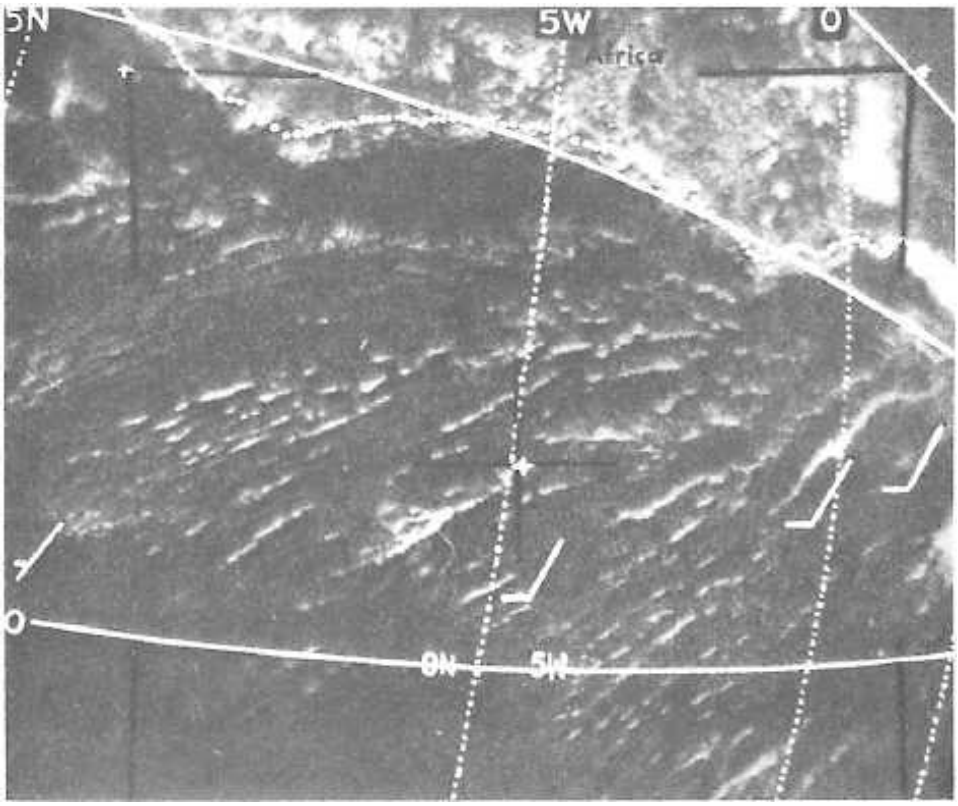


Fig 4.1 Low-level clouds off the coast of Africa

10. Just as low-level winds are estimated from cloud patterns seen in satellite imagery, so too are upper-level winds. Two of the most useful cloud patterns used in determining upper-level flow are plumes from the tops of cumulonimbus (Cb) clouds and cirrus cloud shields associated with subtropical jet streams.

11. **Upper-level Winds from Cumulonimbus Cloud Tops**. Vertical wind shear through an atmospheric layer containing clouds is one of the prime factors governing the shape of cloud formations. Most of the cloud lines and bands seen in satellite pictures parallel the thermal wind through the cloud layer (base to top). Under some circumstances, the direction of the shear is the same, or nearly the same, as the wind direction in the layer containing the cloud. This is the case if ‘a’ the wind changes speed with height but not

direction or ‘b’ the wind changes direction with height, but the winds at the top of the Cb are much stronger than those near its base. When either of these conditions occurs, it is possible to estimate the wind direction directly from the orientation of the Cb plumes. These conditions are common in tropical regions, where the speed of the wind at the level of cumulonimbus anvils is several times greater than the mean wind speed of the wind through the lower portion of the Cb. In these cases, wind estimates based on orientation and dimensions of cirrus blow-off are quite accurate and closely approximate the winds near the cloud top. On the other hand, there are two conditions that can cause considerable error in wind direction estimates derived from Cb anvils: ‘a’ light winds at the cirrus level and ‘b’ a large difference between the direction of the shear through the convective layer and the direction of the wind at the top of the layer.

12. When the speed is light (less than 20 knots), the wind direction is variable causing the correlation between plume orientation and wind direction to be less reliable. If there is a large difference between the direction of the shear through the convective layer and the direction of the wind at the top of the layer, the plumes will parallel the shear but not the wind at the top of the layer. This condition is common in frontal areas where temperature advection is strong, but occurs less frequently in the tropics.

13. **Upper-level Flow from Cirrus Cloud Shields**. It is common to observe huge masses of cirriform clouds extending poleward from the tropics. These clouds occur in advance of 200-hPa troughs and represent a poleward transport of momentum and high-level moisture from the tropical region. A wind maximum is usually located on the poleward edge of these cloud formations. As with strong jet streams, the direction of the upper-level wind parallels the cloud edge. Because the air is moving away from the equator and accelerating, it is crossing the contour pattern toward lower pressure. For this reason, the general orientation of the cloud shield and the striations within it can differ as much as 15 degrees from the orientation of the upper-tropospheric contours. When this is taken into account, cirrus formations of this type provide good estimates of wind direction but less precise information as to wind speed. Fig 4.2 shows a typical cirrus cloud shield. The double shafted, white arrows (A to C) represent the 200-hPa wind maximum suggested by the cirrus cloud edge. The black, single-shafted arrows represent the 200-hPa flow based on individual cirrus elements that lie within the larger scale formation. The vertical cloud pattern (D) is associated with a cut-off low.

14. A cirriform cloud shield extends northeast from A in advance of an upper-level trough associated with the vortex. There are numerous small-scale lines of clouds (E to F) that are oriented approximately perpendicular to the wind direction. These lines are believed to be caused by horizontal shear and should not be confused with cumulonimbus plumes which parallel the vertical shear. There are two basic methods of streamline analysis in use; the discontinuous or qualitative method and the streamline-isotach method. If reports are sparse, the discontinuous or qualitative method is generally used. The most complete analysis is made with the streamline-isotach method, and this is the more recommended procedure.

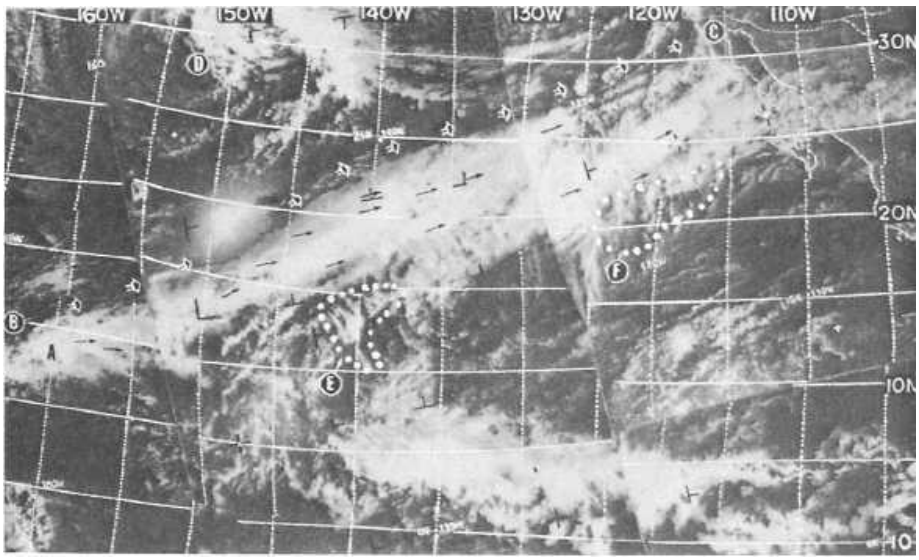


Fig 4.2 A typical cirrus cloud shield

Discontinuous (Qualitative) Streamline Analysis

15. This method involves a single set of lines drawn tangential to the wind direction and spaced in proportion to the wind speed. This means that in discontinuous streamline analysis the spacing between the stream lines is inversely proportional to the wind speed, whereas wind direction remains along them. See the streamline analysis in figure 4.3.

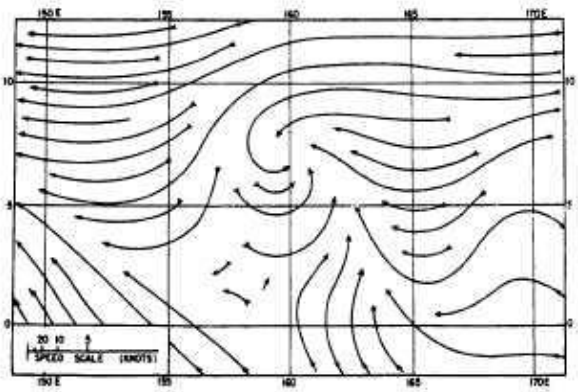


Fig 4.3 Discontinuous or qualitative streamline analysis

Streamline - Isotach Analysis

16. This method consists of two sets of lines: streamlines, representing the wind direction, and isotachs (labeled in knots), representing the wind speed. The two sets of lines give a continuous representation of the wind field, from which wind direction and wind speed can be determined at any point on the chart. Familiarity with circulation patterns is a necessity when using this method of analysis.

17. **Asymptotes**. These are streamlines in the wind field away from which neighboring streamlines diverge (positive asymptotes) or toward which they converge (negative asymptotes). We are primarily interested in the negative asymptotes lying between converging currents since, in the lower levels, they are most frequently associated with bad weather. Significant negative asymptotes are frequently found in the convergent flow around well-developed cyclonic indrafts and in the broad zone of convergence between the trade wind currents of the two hemispheres (ITCZ of the doldrums). Asymptotes may or may not represent lines of true horizontal mass divergence and convergence. Typical examples of asymptotes are shown in figure 4.4 Diffluent asymptotes are drawn in blue and confluent asymptotes in red. They begin and end one-inch before and after the first and last converging or diverging streamline. See figure 4.5.

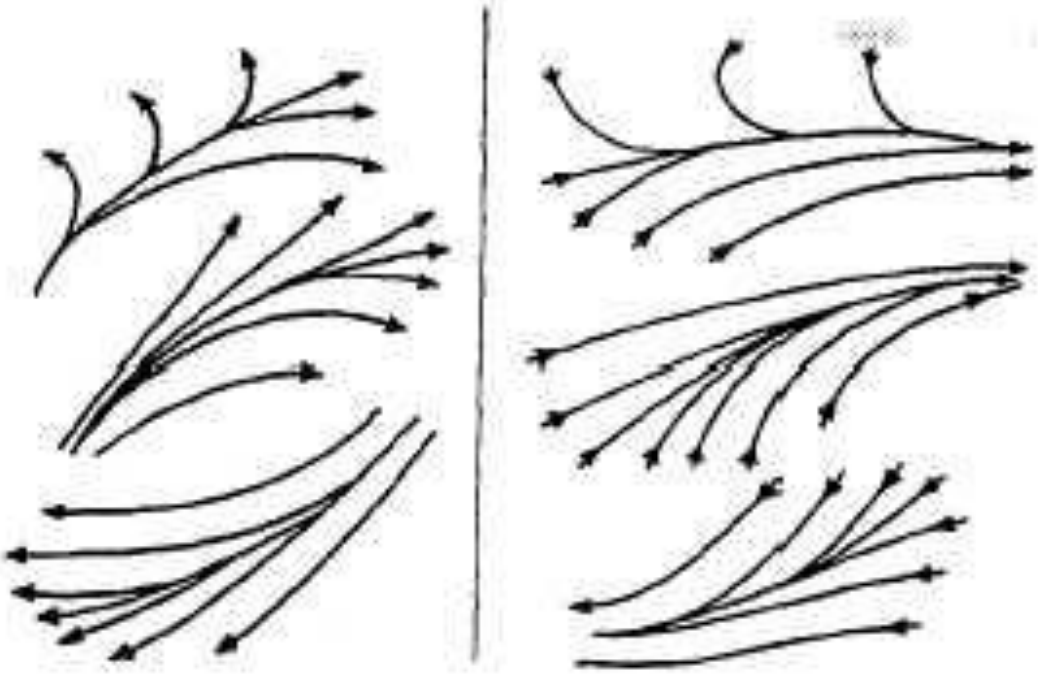


Fig 4.4 Streamline asymptotes of difluence (divergence) left; of confluence right

18. **Waves**. These are perturbations in the streamlines similar to the wavelike arrangements of troughs and ridges in isobaric patterns. Waves that do not extend across the entire width of the current in which they are embedded are called damped waves. In this case, the streamlines on one or both sides of the current have smaller amplitude than those in which the wave is more pronounced. Fig 4.6 illustrates a damped wave in the streamlines.

19. **Singular Points**. Singular points are defined as the point in the wind field where wind speed is zero. These points act as ‘source’ or ‘sink’. Centres of pure rotation, divergence and deformation or different combinations of these are the examples of singular points. Cyclones, anticyclones and col occur at these points. At these points more than one stream lines can be drawn. We also get vortices from streamlines spiraling outward or inward. These are singular points in isogon field also and are meeting point of isogon.

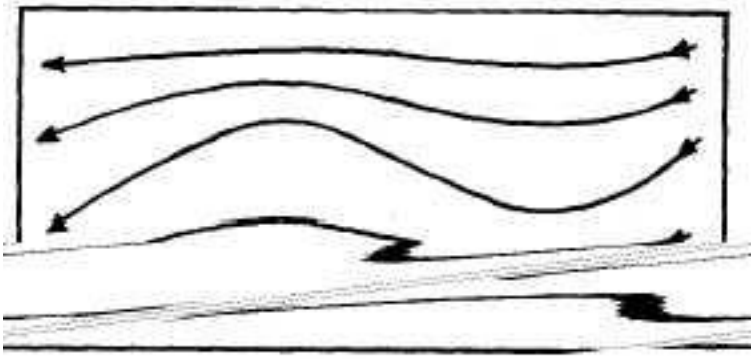


Fig 4.5 A damped wave in the streamlines

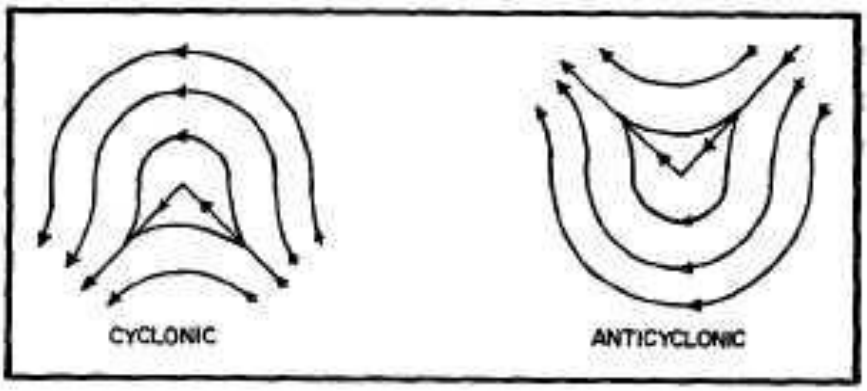


Fig 4.6 Cusps

20. **Cusps.** Cusps represent an intermediate pattern in the transition between a wave and a vortex. They are relatively unimportant in synoptic wind analysis since they are short-lived, and there is normally insufficient data to determine their presence. Figure 4.7 illustrates two variations of this class of singular points.

21. **Vortices**. These are centres, outdrafts, and indrafts of anticyclonic or cyclonic circulation. Fig 4.8 shows six of the most commonly seen vortices. Anticyclonic outdrafts and cyclonic indrafts are frequently found at low levels in the atmosphere. Data is usually too sparse at the upper levels to determine outdraft or indraft characteristics of vortices. Therefore, many upper level vortices are drawn as pure cyclones or anti-cyclones, for lack of more detailed information. Satellite pictures reveal even the weaker, smaller-scale vortices, but conventional data is required to determine the level of most intense circulation.

22. **Neutral Points**. These are the points at which two asymptotes, one of directional confluence (convergence) and one of directional difluence (divergence), come together. They are similar to cols of isobaric analysis in that they represent a saddle between two areas of anticyclonic flow and two areas of cyclonic flow. Neutral points in the streamlines are shown in fig 4.9.

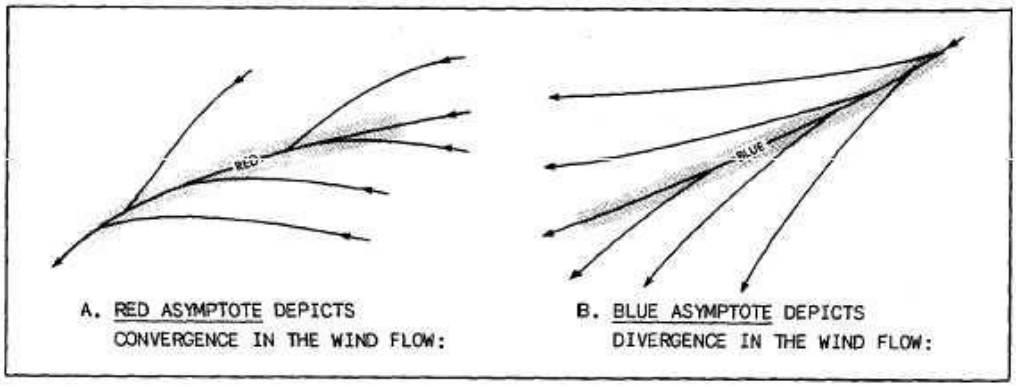


Fig 4.7 Colored asymptotes

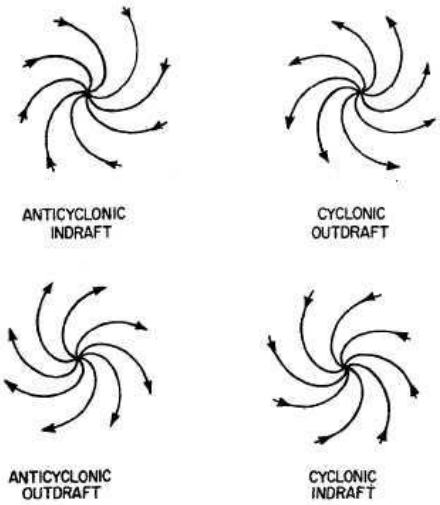


Fig 4.8 Vortices in the streamlines (Northern Hemisphere)

Analysis Procedures

23. As with any weather analysis, you must look to the past to better understand the present. Make sure past history is transferred onto the current streamline chart. In streamline analysis the past locations of cyclones, anticyclones, waves, neutral points, and asymptotes of convergence and divergence are marked using a yellow pencil. Your next step is to evaluate the plotted data for obvious errors. The last preliminary step is dependent on your access to real-time satellite pictures. Pictures coinciding with your chart time should be examined.

24. Before drawing your first streamline, you should tentatively locate and mark (in pencil) the centre positions of anticyclonic and cyclonic centres and their associated neutral points. Also, mark the centre positions of tropical depressions, storms, and hurricanes/typhoons according to information taken from appropriate advisories or warnings, and label the centres with standard symbology. Note that each of the above features is a singular point. Draw the streamlines in the same manner as any other cyclonic circulation. However, do not forget that on upper-level charts (300- or 200-hPa), the wind flow changes from cyclonic to anticyclonic. As hurricanes track northward and enter mid-latitudes, the streamline analysis is dropped and a contour (pressure) analysis is begun. It is up to the command to determine at what point the switch is made.

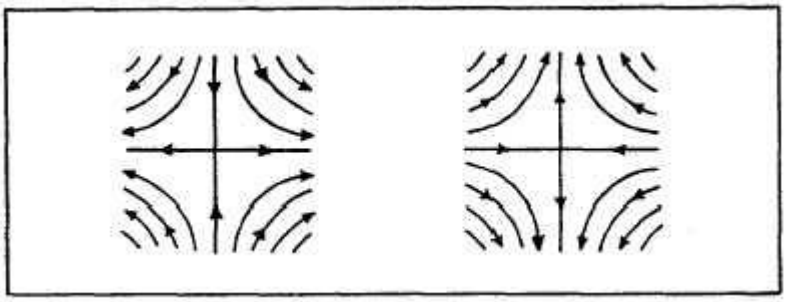


Fig 4.9 Neutral points in the streamlines

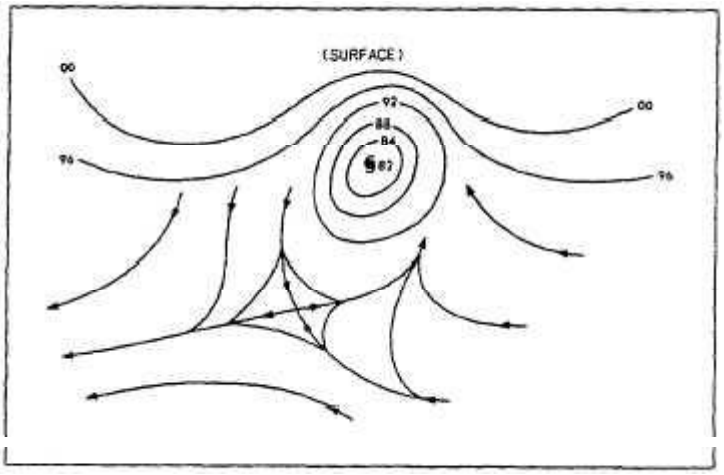


Fig 4.10 Low-level streamline-surface contour analysis combination

25. Fig 4.10 illustrates the low-level streamlines associated with a hurricane and the switch from streamline to contour analysis.

26. A recommended area in which to start drawing streamlines is at the base of the subtropical ridge line. The wind flow in this area is fairly straight (undisturbed). These winds are the trades, and they are the dominant feature of most tropical streamline analyses. Draw a few streamlines through this area first. Then, analyse the major features in the subtropical ridge—the anticyclones, and neutral points. Upon completing this, locate and draw streamlines within any other extensive area of undisturbed wind flow. Your next step is to look for cyclones and their associated neutral points. Asymptotes are then drawn to these singular points and also to areas of confluent and diffluent winds. After all major features and asymptotes are drawn, sketch additional streamlines until a smooth and continuous pattern of wind direction is established. You should remember that some interpolation is almost always required in drawing streamlines. Streamlines must parallel the wind arrows, although some leeway is given in areas of wind speeds that are 5 knots or less.

27. The analysis of the wind field is incomplete without the isotachs. They are dashed green lines which, like streamlines, are first sketched in pencil. They represent areas of equal wind speed. Isotachs are drawn at 5-knot intervals up to 20 knots, then every 10 knots thereafter. Figure 4.11 is an example of an isotach pattern typically found on a streamline analysis. The major axes of isotach patterns tend to parallel streamlines. The higher wind-speed areas show elongated patterns, while areas of low wind speed show a much broader isotach pattern. Colour coding is used to denote certain wind speed areas. Areas with speeds 30 knots or greater are shaded in purple, while areas with wind speeds of 10 knots or less are shaded yellow. In addition to this shading, the terms *MAX* and *MIN* are placed inside the highest and lowest isotachs respectively.

28. Isotach maxima and minima have preferred areas of occurrence within the streamline pattern. The maxima are usually found upstream from centres of confluent wind flow on either side of asymptotes. Maxima also tend to occur down-stream from the centres of streamline divergence on either side of asymptotes. See fig 4.12 In the trades, or any other major east-west current, elongated maxima are usually found near the centre of the current. Fig 4.13 shows that in very broad currents, two or more maxima may exist side by side. Isotach spacing reflects the strength of the wind; the higher the wind speed, the tighter the spacing. Isotach minima occur at all singular points and also along asymptotes. Around neutral points, they take on a saddle-like appearance away from the area of minimum winds. See fig 4.14.

29. From the illustrations presented, you can see that isotachs kink when they cross streamlines. With cyclonic circulations, they kink with the flow (in the same direction). With anticyclonic circulations, they kink against the flow. Fig 4.15 illustrates these two patterns.

30. Before you finalise your analysis, look over your chart objectively and ask yourself the following questions:-

a. Do the streamlines conform to the general flow pattern?

b. Have I justified throwing out every wind report that I could not draw to?

c. If I cannot justify discarding a wind report, is there any way I can draw to it?

d. Have I drawn the streamlines parallel to the plotted winds, rather than at an angle?

e. Is the chart consistent with other levels?

f. Is the chart consistent with history?

g. Are the streamlines and isotachs consistent?

h. Have I drawn any unnecessary lines?

j. Have I given more weight than necessary to light and variable winds?

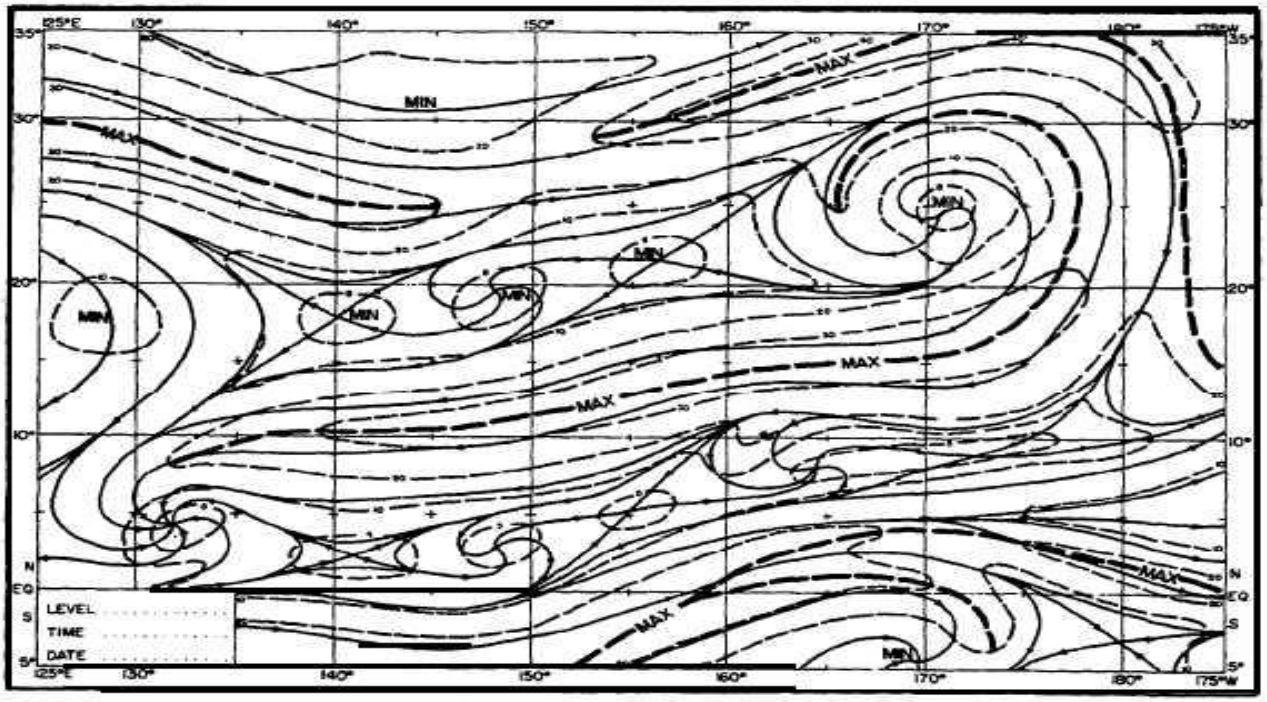


Fig 4.11 Streamline-isotach analysis over a large ocean area

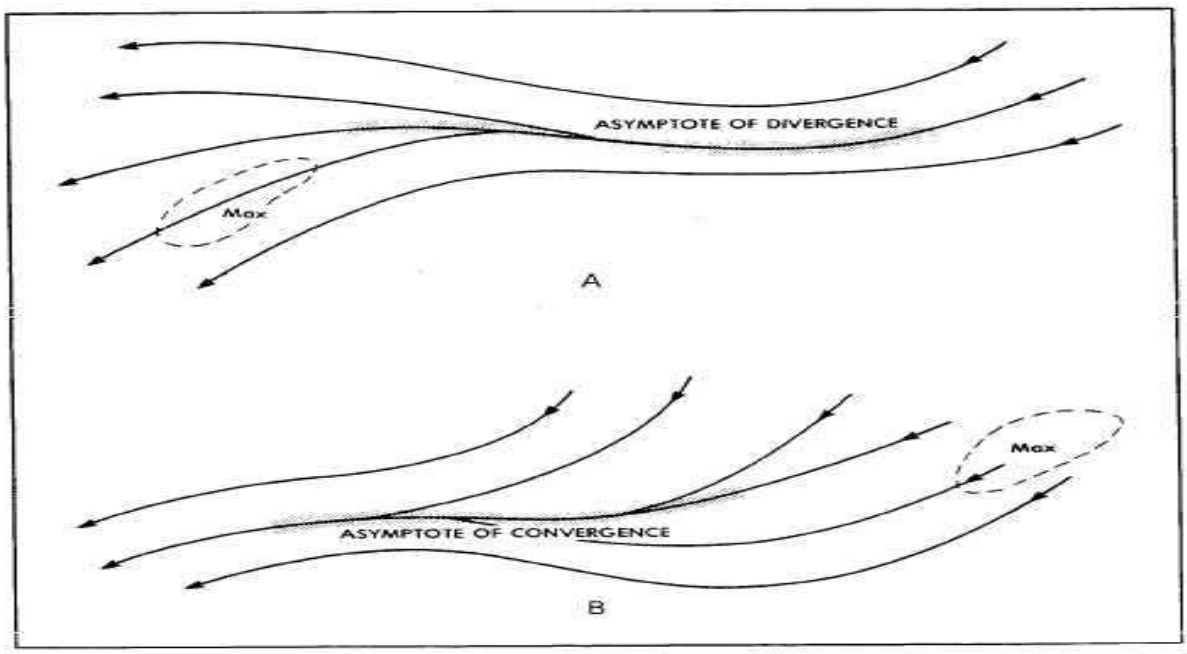


Fig 4.12 Isotach maxima relationships to asymptotes of divergence and convergence

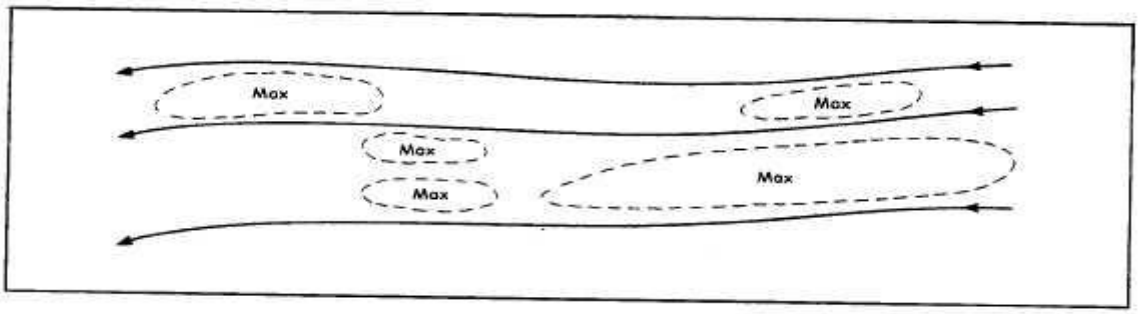


Fig 4.13 Isotach maxima in broad easterly flow; maxima usually parallel streamlines and are sometimes found side by side

Correlation of Wind and Weather

31. In as much as the vast majority of clouds are formed by upward motion of air, and vertical convergence of air can be deduced from the horizontal components of direction and speed of the wind field taken over large areas, many weather patterns in general, and specifically tropical weather patterns, have a direct relation-ship to streamline patterns at all levels.

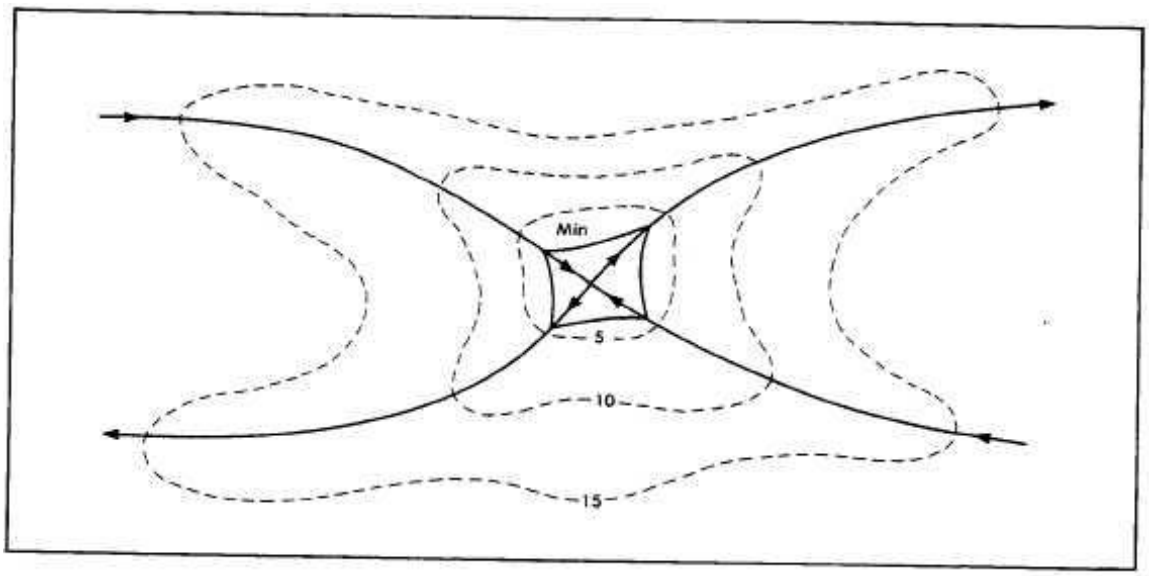


Fig 4.14 Isotach pattern around neutral points

32. Complete correlation of wind and weather requires analysis of the wind field at several levels; however, data and time limitations usually permit correlation of only two levels: the low-level streamlines between 2,000 and 5,000 feet and the streamlines of the 200-hPa level. The correlation rules are as follows:

a. In regions of moderate to strong divergence in the low-level streamlines, the predominant cloud is usually cumulus humilis with less than one-half coverage. When great vertical shear accompanies the moderate divergence, the cumulus is drawn out and sheared-off and may be accompanied by dependent stratocumulus.

b. North of the trade wind maximum, in the Northern Hemisphere, the vertical shear in the lower layers is usually great and the trade wind inversion low. The predominant cloud here is stratocumulus. The clouds exist in broken patches if the low-level-streamline pattern is divergent. Even with weak to moderate convergence at this level, very little cumulus forms; this usually results in an increase in the amount of stratocumulus west and south of the subtropical high-pressure centres.

c. When there is little vertical shear in the lower layers, and the low-level streamlines show weak to moderate convergence south of the trade wind maximum, expect deeper cumulus. The amount of cumulus will also increase over that expected in divergent flow, but not to a marked extent. Even with very strong convergence, the cumuliform clouds show many breaks, the general effect of increased convergence being to increase the height of the convective clouds.

d. An asymptote of convergence in the low-level streamlines is accompanied by a line of cumulus congestus or cumulonimbus if the asymptote coincides with an isotach minimum in the wind field.

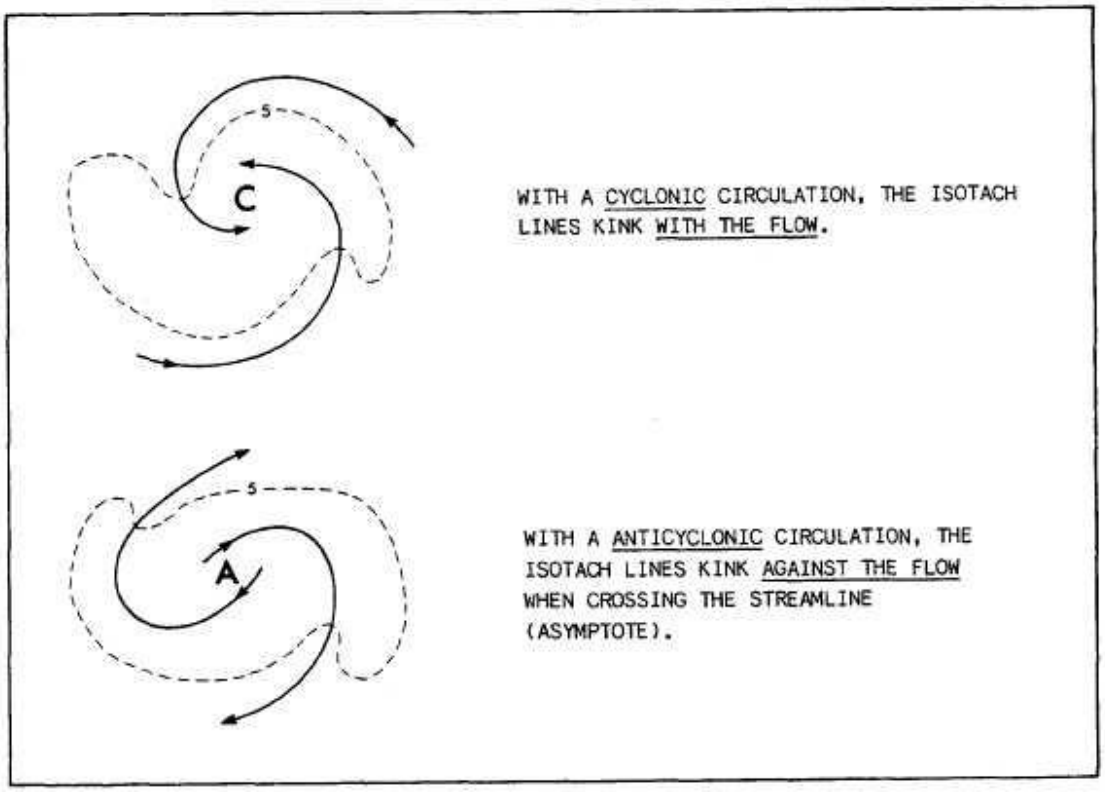


Fig 4.15 The typical isotach patterns associated with cyclones and anticyclones

e. Polar fronts that penetrate into the tropics rarely reach 15°N or S latitude over oceanic regions. Such fronts are often detectable as a line of towering cumulus. The highly modified front may, in such cases, appear on the low-level streamline chart as an asymptote of convergence running east to west or northeast to southwest from the neutral point between two middle-latitude anticyclones (high-pressure centres). In most cases, temperature contrast across such fronts is negligible, even though the cloud line may persist for some time after air mass contrasts have disappeared.

f. An asymptote of convergence in the lower streamline field is not necessarily accompanied by a cloud line. If strong speed divergence occurs along part of the line, there may be very little heavy clouds or precipitation accompanying it; furthermore, bad weather may be found some distance from the line where speed convergence predominates.

Weather Distribution Charts and Analysis

33. The weather distribution chart is a graphic representation of the clouds and weather over an area. Outside of the tropics most meteorologists are reluctant to spend the time necessary in analysing such a chart, because the surface synoptic chart, with clouds and weather plotted at each station, and satellite pictures are adequate for most purposes. For tropical operations, however, the weather distribution chart is almost indispensable in the areas of flight briefing and forecasting. In instances where precise and detailed analysis of weather over an area or terminal is required, it is often necessary to follow the continuity of the weather and cloud patterns in the same fashion as we follow the continuity of low-pressure systems and fronts. As a briefing tool, the weather distribution chart is perhaps the most practical means of presenting the weather to non-meteorologists.

34. All methods of drawing weather distribution charts involve the use of standard synoptic cloud symbols. Any deviation in the methods is mostly confined to the representation and analysis of cloud amounts. The choice of method is determined locally.

35. The data included on the plotted charts should be as complete and as extensive as possible. In addition to the regular land and ship synoptic reports, other reports entered on the map include reconnaissance, off-hour ship, PIREP, and aircraft in-flight reports. These additional reports are designated as to type and time of the report. Past weather should also be entered, because it aids in correlating the movement of weather systems, and it may fill blank spots on the charts on occasion.

Weather Distribution Analysis

36. If weather distribution maps form a part of standard daily analysis in the weather office, the task of drawing the map is greatly simplified by the fact that the major weather systems of the tropics have continuity from map to map. From day to day, large-scale systems moving through an opera may show the same characteristic intensity, as judged by the amount, depth, and arrangement of clouds and the precipitation pattern. On the other hand, these systems may increase in intensity or die away. However, even during the explosive deepening of some tropical cyclones, the changes are sufficiently slow to be followed on the map sequences. In some regions, systems may form aloft and remain stationary for many days, slowly increasing in intensity from day to day. Under these circumstances, the whole process of deterioration in the weather can be followed in detail.

37. The following discussion of weather distribution analysis is limited to times when weather distribution charts are not part of the normal weather office routine.

a. If you are not already familiar with the climatology of the area, consult the best available information on the seasonal cloud and weather characteristics of the area. If the equatorial trough lies near or across your opera, be on the alert for "equatorial fronts''—lines of cumulus congestus and/or cumulonimbus that appear similar to cold fronts of high latitudes. If operating in the trade wind area, find out the characteristic distribution and heights of cumulus and stratocumulus. In other words, climatological knowledge should help you form a mental image of the normal weather distribution map of the region. If you find radical departures from the climatological pattern, pay particular attention to the anomalous features, making every effort to delineate them on the map.

b. **Examine Latest Satellite Imagery**. Without a doubt, satellite imagery provides the most complete picture of the cloud cover over the tropics. When available, imagery shows the extent and orientation of cloud systems. Also, when used in conjunction with standard surface and upper-air data, the weather associated with cloud systems is monitored with a far greater degree of success than was possible before the advent of environmental satellites.

c. **Examine Aircraft Reports**. First priority should be given to reconnaissance reports.

d. **Outline Areas of Middle Clouds**. The first step in analysing the plotted weather distribution chart is to outline the middle clouds, following observed reports and satellite pictures, if available. If no satellite picture is available, you will have to make a reasonable interpolation as to the extent of mid clouds. If cumulonimbus are absent, any mid cloud is probably independent. Independent systems are usually band-like or sickle-shaped with frayed edges of patchy altocumulus. All middle clouds are not independent. If orographic cumulus or cumulonimbus are widespread, fairly large patches of mid clouds maybe reported in this area. However, these patches will be detached from one another and will rarely form a very extensive alto-system with a definite shape. If reports of cumuliform clouds indicate that the mid cloud in any part of the region is dependent, you should keep the mid cloud outline to the minimum extent that it is compatible with the reports and the topography; at the same time, show clearly the connection between the mid cloud and the convective pillars. In passing, it should be emphasised that over the open sea, lines of cumulus congestus or cumulonimbus are often associated with alto-systems, but this does not necessarily mean the mid cloud is dependent; on the contrary, the most likely situation, during the late stages of deepening of an upper-level cyclone, is for one or more asymptotes of convergence, with accompanying cumulus or cumulonimbus lines, to develop under a preexisting deep alto-system and merge with it aloft.

e. **Delineate the High Clouds**. Follow the same principles as used in drawing the mid clouds. Usually, there is little difficulty in distinguishing between dependent and independent cirrus. Almost all independent alto-systems are accompanied by cirrus or cirrostratus sheets, either separate and at a much higher level, or fused with it in the areas of precipitation. Usually cirrus sheets cover a wider area than the alto-system, and often cirrus in broad bands cover a great area on either side of the alto-system, the bands being oriented parallel to the main axis of the alto-system. The delineation of cirrostratus sheets of independent formation can be of great assistance in tracing the genesis and development of an upper-level cyclonic system. In the early stages of development of such systems, only a small amount of mid cloud may be present, and the circulation may be evident only in the layers above 30,000 feet (9,144 metres). On the weather distribution chart, however, this early stage of upper-cyclone development is often accompanied by extensive sheets of cirrostratus. In the absence of satellite data, a series of maps showing the development of cirrostratus sheets is often the best way to follow the gradual intensification of such systems.

f. **Outline Areas of Low Clouds**. By the time you have analysed the distribution of the mid and high-clouds, you should have studied most of the lower cloud reports and have formed definite ideas about their distribution. Low clouds of orographic origin should be obvious by this time. The more front-like lines of cumulus and cumulonimbus should be clearly marked. Often, very extensive areas may be covered by more less uniform distributions of cumulonimbus. Delineate these areas as sharply as possible.

g. Next, note all precipitation and special phenomena. The analysis is then complete. Although various combinations of cloud types and amounts occur, the following patterns tend to be the most prominent:-

(1) A large amount of low clouds, cumulus congestus and cumulonimbus, with mid and high clouds and reports of current or past showers or rain. This is a typical pattern associated with tropical disturbances.

(2) Few or no mid or high clouds either with few low clouds or with large amounts of cumulus humilis (fair-weather cumulus) or stratocumulus. This combination is indicative of suppressed convection.

(3) Near-average amounts of low clouds, occasional cumulus congestus, and mid and high clouds in varying amounts. In such a transition pattern, it is most important to study the upper clouds, since these are independently formed in such cases rather than derived from towering cumuliform clouds. Thickening cloud cover aloft, especially altostratus overcast denotes an approaching, developing disturbance.

Surface Chart — Isobaric Analysis

38. It is essential to draw for 2-hPa intervals in the trades and for 1-hPa intervals close to the equator. In addition, the 1-hPa interval should be used whenever the isobars are so widely spaced that even 2-hPa isobars do not reveal the important features of the pressure field. It is thus readily seen how important it is to screen pressure reports carefully. A few bad pressures can distort the isobaric pattern in a grotesque way. All pressures from normally reliable stations should be drawn quite strictly and disregarded or modified only when excellent reasons exist. The 2-hPa spacing is required because of the weak pressure gradient in the tropics.

39. The low-level streamlines, especially when at 200 feet, determine the character of the surface isobars to within about 5° of the equator. In the centre of the equatorial zone, you must rely on the pressure reports alone. In fixing the course of the isobars, use the low-level streamlines. The streamlines are an invaluable aid in sorting out good and bad pressures, especially in the trades. However, streamline analysis is concerned exclusively with the wind field and does not convey information about the field of mass and its changes, which is essential for many prediction problems.

24-Hour Pressure-Change Charts

40. As an added plotting step on the surface chart, the 24-hour pressure changes that have occurred at all reporting stations should be computed and written just above their plotted sea level pressure values on the surface map. It will help if the plotter circles these changes so that they stand out among the mass of information on the surface chart. Over the ocean area, the procedure is to tabulate the pressure at all 5° latitude-longitude intersections and to compute and plot the pressure differences. Isallobars are then drawn at intervals of 0.5 to 1.0 hPa, depending on the magnitude of the changes.

41. In an alternative method, the changes maybe obtained by using a light table and graphically subtracting the pressures of the current chart from the previous chart. This method, however, often proves less satisfactory because the isobars tend to run fairly parallel and intersect at very small angles, making interpretation difficult.

42. The main objective of 24-hour pressure-change charts is to track the isallobaric centres and note their changes in intensity. In addition, these charts furnish a check on the ocean isobaric analysis; if large regions of intense falls suddenly appear where none existed previously, you will do well to reexamine such areas. As was stated previously, the tropics is a region that experiences little change in its daily weather; therefore, any sudden intense pressure changes would signal a breakdown in its continuous nature. In general, isallobaric analysis yields good continuity. Usually, you will look for concentrated fall or rise centres that cover limited areas only. These are the areas of intensely disturbed weather. Extensive areas of fairly uniform pressure fails are considered harmless.

43. Occasionally falls of 1 to 2 hPa over most, if not all, of the region of analysis will appear suddenly on one day, only to be cancelled out by corresponding rises within 24 to 48 hours. Such atmospheric pressure waves on a huge scale are not understood at present, but they do not appear to be of any significance. When such widespread falls occur, an area of isallobaric falls often appears as a centre of still larger falls. On the other hand, an area of isallobaric rises may take the form of an area of small falls embedded in an area of larger falls.

44. Subsequent to the analysis, the isallobaric centres should be related to the low-level streamlines and the weather pattern; later they should be compared with the distribution of wind and temperature aloft. Fall centres containing extensive bad weather and associated with warming in the troposphere and anticyclonic circulation at 200 hPa must be regarded as potentially dangerous. Rise centres associated with bad weather generally do not indicate immediate deepening, even though the intensity of the convection may be severe.

Upper-Air Charts and Analysis

45. The two most important constant-pressure charts used in tropical analysis are the 200- and 500-hPa charts. Upper-level troughs and ridges over lower latitudes usually reach their greatest intensity near 200 hPa, which makes this level suitable for high tropospheric analysis. Also, pressure pattern flying is as feasible over tropical oceans, at least 10° to 15°N and S, as it is at higher latitudes. Thus, for high-altitude jet aircraft operations, a 200-hPa analysis is indispensable.

46. The 500-hPa level is not an ideal level for tropical upper-air analysis, the reason being that many disturbances and their associated temperature fields occupy only the lower and upper troposphere. Nevertheless, this level is important because of its use in differential (thickness) analysis. It is impractical to analyse the entire tropical layer between 1,000 and 200 hPa, so we break it into two layers, 1,000-500 hPa and 500-200 hPa. Constant-pressure and differential analyses are carried out in the same manner as in mid-latitudes, although there are some general considerations that must be taken into account.

General Upper-Air Analysis Considerations

47. Several rules guide the upper-air analysis:-

a. Draw upper contours parallel to upper winds when the wind is 10 knots or greater.

b. Upper heights and thickness between levels vary in a limited range in relation to location and season. Frequency distribution charts provide us with the normal and extreme values. Heights and thicknesses outside observed ranges should be questioned, while higher and lower values within the permissible range are significant because of their association with definite types of flow patterns.

c. Vertical wind shear vectors should indicate the orientation of the mean temperature field though not necessarily the magnitude of the mean temperature gradient. This applies especially to the 500-200 hPa layer.

d. Since the mean temperature of a layer several hundred hPa thick is nearly conservative from day to day, the thickness fields on successive days should be readily comparable.

48. Regions of relatively warm or cold air should be traceable from day to day without much difficulty, and the intensity of the centres should not change drastically with no good reason. In particular, large areas of cold or warm air should not suddenly appear where none existed previously.

Analysis Procedure

49. The following is a recommended set of steps in conducting your analysis of the tropical atmosphere:-

a. Analyse constant-pressure charts; especially the 1,000-, 500-, and 200-hPa levels.

b. Compute height differences between levels and check the results for pattern and change from the preceding day.

c. Analyse thickness charts, either graphically or from charts of thickness values and wind shears plotted at stations and at each 5° latitude-longitude intersection over the ocean. The latter is the preferable method, but time limitations will normally prevent its preparation. Depending on the area and season, contours are drawn at 15- or 30-metre intervals.

d. Correlate the cloud and weather chart with upper-level streamline and contour patterns and thickness changes. Some correlations are as follows:

(1) In a zone of cloud reports indicating definite subsidence, cooling from the previous day won't be observed unless a cold pocket is being advected into the area.

(2) A deepening upper low must have either cloudiness on the inflow side into the low or colder air advected into the low from outside the system. There is no other physical process except radiation that could achieve the central cooling necessary to account for 200-hPa height falls.

(3) If a 200-hPa high intensifies and extensive cloudiness exists beneath it, this can only mean that latent heat of condensation is being liberated in such a way as to produce warming aloft. Since this heat can be partly converted to kinetic energy, the possibility of a tropical storm forming must be considered.

TASK 16 : FORECASTING OF MINTRA LEVEL

Definition

1. The level below which condensation trails do not form is called the Mintra Level. It is a function of temperature and pressure and therefore, fluctuates up and down in the atmosphere under varying conditions of relative humidity.

2. To meet the operational requirements of jet aircraft, information regarding Mintra Level is to be included as a routine in local and flight forecasts.

Forecasting

3. For forecasting Mintra Level a table showing the critical values of pressure and temperature is given in Table I. A graph of the same values is given in Figure 5.

Pressure (MB)	Relative Humidity (%)			
	0	60	90	100
40	-69.1	-66.9	-63.9	-61.3
50	-67.4	-65.0	-62.0	-59.2
60	-65.9	-63.4	-60.3	-57.7
70	-64.6	-62.0	-59.0	-56.2
80	-63.3	-60.9	-58.0	-55.0
90	-62.4	-59.8	-56.9	-53.9
100	-61.5	58.9	-56.0	-53.0
150	-58.0	-55.1	-52.3	-49.0
200	-55.4	-52.6	-49.6	-46.2
250	-53.2	-50.3	-47.4	-44.0
300	-51.5	-48.6	-45.6	-42.0
400	-48.6	-45.6	-42.7	-39.0
500	-46.5	-43.3	-40.3	-36.7
600	-44.7	-41.4	-38.4	-34.6
700	-43.0	-39.9	-36.8	-32.9
800	-41.7	-38.3	-35.2	-31.3
900	-40.3	-37.0	-34.0	-30.0
1000	-39.2	-36.0	-32.9	-28.9

Figure 5

4. The minimum relative humidity required in the environment to given rise to jet exhaust trail formation at the given values of pressure and temperature.

5. Since the atmospheric humidity must always be between 0 and 100% the graph has been divided into relative humidity zones of 0,60,90 and 100%. If a pressure temperature value lies to the left of the 0% line, the contrails should always form, even in absolutely dry air. This is indicated as the “Yes” area. If the value falls to the right of the 100% line, the contrails should never form even in saturated air. This is marked as “No” area. If the point falls between the 0 and 100% lines, the contrail is to form provided the actual relative humidity is equal to or greater than the value indicated at the point on the graph. This is the uncertain or “Possible” area.

6. For day-to-day forecasting a graph is to be constructed on a Tephigram as in Figure 16 using the data in Table 1 and kept protected under Perspex for permanent use. The point of intersection between dry temperature curve and the 100% humidity line gives the Mantra Level.

TASK 17 : FORECASTING OF ICE ACCRETION AND SUPER-COOLED WATER

Introduction

1. Ice accretion is a recognized hazard to aviation. Although modern aircraft are fitted with various de-icing devices and effective methods are available to remove ice deposits, none of those methods provide complete freedom from this hazard. Recently, in a few aircraft accidents it has been suspected that even super-cooled moisture laden air could in itself be a potential hazard owing to its ability to adversely affect the engine performance of certain types of jet aircraft.

The Effect of Ice Accretion on the Performance of Aircraft.

2. **Airframe Icing:** Ice formation near the leading edges of the airframe modifies the airflow round the affected part. It normally increases drag and reduces lift. Ice may as well form on other parts of the wing and fuselage and cause considerable increase in drag. Efficiency of radio and radar equipment may be adversely affected by ice on masts, aerials and radomos.

3. **Engine Icing.** Ice formation in the air-intake and induction system of a reciprocating piston engine results in loss of engine power due to obstruction of the air passage and to disturbance of the fuel motoring. It may even cause a flame out. Moveable parts also may become inoperative. Engine icing is not only a low temperature phenomenon but may occur at air temperatures will above freezing point and in clear air as well as in clouds and precipitation. For some types of jet aircraft, even the super-cooled moisture alone can cause adverse effects on the performance of the compressors.

Meteorological Factors of Ice Accretion

4. The Meteorological factors are:

   a. Super-cooled water content of the air.
   b. Ice-crystal content of the air.
   c. Temperature and humidity.
   d. Water droplet and ice-crystal size distribution.

5. Of the four meteorological factors, the most important on is super-cooled water-content of the air. The principal factors governing the quantity of super-cooled water :

   a. The temperature.
   b. The rate of ascent of the air.
   c. The initial water vapour content of the ascending air

6. Any successful forecast of ice-accretion, therefore, would involve the ability to forecast the above parameters.

Super-Cooling of Water-Droplets

7. Water, in finally divided form as it occurs in the atmosphere, consists of minute drops, varying in diameter form a few microns to a few millimeters. These drops, however, never freeze at 0°C in the absence of ice-crystals, and can exist in the atmosphere in the liquid state at temperatures well below 0°C. Laboratory experiments have shown that the temperature of spontaneous freezing varies with dropsize, but at temperature of about 0°C, even the smallest size droplest freeze spontaneously. This suggests the temperature limits for existence of super-cooled water in clouds is 0°C to –40°C. However, actual flight reports on few occasions have brought evident of super-cooled water existing even at lower temperature but these were on rate occasions and night have been under special physical circumstances.

Super-Cooled Water and cloud Types

8. In our latitudes the cloud types which extend into sub-freezing temperatures are mainly the following:

a.	Large Cumulus	Usually super-cooled water upto –23°C.
b.	Cumulonimbus	Usually super-cooled water drops upto 14°C to 30°C mixture of super-cooled water drops and ice-crystals. Some water drops may exist upto –40°C beyond which only ice-crystals are expected.
c.	Altocumulus	Mainly super-cooled water upto-10°C and at times below this temperature.
d.	Altostratus	Usually ice-particles only but occasionally super-cooled water.
e.	Nimbostratus	Mainly super cooled water.

Measurements of Super-Cooled Water

9. The actual measurements of super-cooled water content throughout the range of possible values, and at all possible temperatures is extremely difficult and can only be performed by instrumented aircraft meant for meteorological research.

10. The maximum liquid water content in clouds can be calculated from thermodynamic considerations. Such calculations have been performed by American workers and by Best UK.

11. For the simple model of a convective type of cloud consisting of vertical current of air in which water condenses as a result o cooling by ascent, the temperature falls of at saturated adiabatic lapse-rate and the cloud is isolated from the surrounding, the values of liquid water content for three different cloud base.

Temperature at 950 mb are given in table-II

Pressure	Height Above 950 mb	Temperature	W
Mb	Ft	oC	Gm. m^{-3}
Cloud-bade temperature 38°F at 950 mb.			
400	21.068	-45.6	3.01
500	16.036	-31.8	3.27
600	11.704	-20.8	3.18
700	7.896	-12.0	2.71
800	4.501	-5.1	1.89
900	1.433	+0.7	0.72
950	0	+3.3	0.00
Cloud-bade temperature 70°F at 950 mb.			
250	34.212	-41.1	5.99
300	30.042	-29.9	6.58
400	23.082	-14.5	7.18
500	17.406	-3.9	6.91
600	12.618	+3.7	6.14
700	8.473	+9.7	4.91
800	4.811	+14.8	3.19
900	1.526	+19.2	1.08
950	0	+21.1	0.00
Cloud-bade temperature 80°F at 950 mb.			
200	40.861	-34.8	6.87
250	35.611	-22.8	7.66
300	31.141	-14.1	8.17
400	23.811	-2.4	8.50
500	17.911	+6.2	7.91
600	12.951	+12.7	6.76
700	8.683	+16.3	5.43
800	4.923	+22.0	3.70
900	1.561	+25.9	1.44
950	0	27.8	0.00

12. For layer type clouds, which are formed by turbulence (such as stratocumulus), and upper-limit to the liquid water content can be estimated by assuming that the air is lifted by eddies along the wet-adiabatic curve from the base of the cloud to the top. Table III below shows the liquid water-content of air which has arisen along the wet-adiabatic through a height of 3.000 feet above the cloud base to reach the temperature shown in the table at pressure level of 9.00 mb. Liquid water content in layer cloud 3.000 feet deep.

Table-III

Temperature at cloud top (900 mb)				
0°C	-10°C	-20°C	-30°C	-40°C
Grams per cubic meter				
Liquid water content				
1.59	1.05	0.65	0.39	0.18

13. The above tables should form a guide rather than a rule in estimating the liquid water-content in any cloud situation, since there may be variation between calculated theoretical values and actual values.

Orographic Effect

14. Three orographic effects would be considered for forecasting icing conditions:

a. **Modification of Free Air Temperature:** Force ascent on hills raises or lowers temperatures according to whether the air is unstable or stable with respect to appropriate adiabatic quantitative assessments can be made from the general rules.

b. **Modification to Liquid Water-Content:** The liquid water content tends to be greater when air rises over hills.

c. **Effect of Lee-Waves:** In ice waves vertical action will modify temperature distribution and liquid water-content in clouds in a manner similar to that caused by high grounds.

Severity and Chance of Icing in Convective Clouds

15. For convective clouds, the following rules are suggested :

a. At temperature below-40^{0}C, the chance of icing is very little.

b. At heights where the temperature is between – 23^0 and –40^oC the chance of moderate or severe icing is little except in newly developed convective cells. However, light icing is possible.

c. At heights where temperature is between 0^0C and –23^0C, the rate of icing may be severe over a substantial depth of cloud for a wide range of cloud base temperatures. Figure 6 will be found helpful in assessing the severity of ice accretion, along with the estimate of probable liquid-water content. Best has shown that super-cooled water contents of 1 and 0.5 grams per cubic meter are approximately the limiting values separating severe, moderate and light icing.

General Remarks

16. It should be borne in mind that most of the icing statistics that have been gathered and are available in literature are from observations made in USA, USSR, UK, Canada and Australia. Very little or no work has been done in tropics and particularly in our country in this field. Therefore, the suggested rules must be used with discretion, and the best performance on the part of a forecaster would be to judge every situation very carefully from the knowledge of the physical processes that he expects to be occurring in the atmosphere at any particular space and time.

Self Assessed Exercise on 16 & 17

Q1. What is Mintra level ?

Q2. Does Mintra level remain fixed ?

Q3. Under what condition the condensation trail must form ?

Q4. How airframe icing affect the aircraft performances ?

Q5. What is the most important requirement for ice formation ?

Q6. What is super – cooled water drops ?

Q7. What is the range of temperature within which the super-cooled water drops are found ?

Q8. What Orographic effects are to be considered for force casting icing ?

Q9. Write short notes on super – cooled water.

Q10. “Aircraft icing is one of the major weather hazards to “aviation”, Explain. What are the effects of aircraft icing ?

Q11. What is super cooled water ? In our latitudes what are the cloud types which extend into sub-freezing temperatures, Describe elaborately.

Answer to the Self Assessed Exercise on 16 & 17

A1. It is the level below which condensation trails do not form.

A2. It is not a fixed level, rather it fluctuates up and down in the atmosphere under varying conditions of relative humidity.

A3. If a pressure temperature value lies to the left of the 0% RH line.

A4. Ice formation near the leading edges changes the airflow pattern. It increases the drag and reduces lift. Ice formation of masts, aerials and redoes affect the radio and Radar efficiency.

A5. Presence of super-cooled water droplets in the air.

A6. Water drops those exists at temperatures below freezing points.

A7. The temperature range is O°C to – 40°C.

A8. Para 14a, b and c. of Task 17.

TASK 18 : THUNDERSTORM FORECASTING

Chapter objectives

After reading this chapter, you should be able to:-

- Assimilate the different method of Forecasting thunderstorm.
- Understand clouds formation under different conditions.
- Assimilate indications of instability in atmosphere by different method.

Introduction

1. The standard method of forecasting air mass thunderstorms has long consisted primarily of an analysis of rawinsonde data with particular emphasis on the so-called "positive areas."

2. Many times conditions are favourable for thunderstorm development, with a large positive energy area showing up on the sounding, with no ensuing thunderstorm activity. At other times, thunderstorms may occur when they are not forecasted. Clearly, factors other than instability are important, and, at times, of overriding importance.

3. A number of thunderstorm forecasting methods have been developed, but many of these are beyond the scope of this book. The forecasting of convective clouds by using variations of the parcel method are covered in this section. Further, a method for the prediction of these storms that enables the forecaster to arrive at a fairly accurate, reasonably objective forecast will be discussed.

The Parcel Method

4. The temperature of a minute parcel of air is assumed to change adiabatically as the parcel is displaced vertically from its original position. If, after vertical displacement, the parcel has a higher virtual temperature than the surrounding atmosphere, the parcel is subjected to a positive buoyancy force and will be further accelerated upwards; conversely, if its virtual temperature has become lower than that of the surrounding air, the parcel will be denser, and eventually return to its initial or equilibrium position.

Formation of Clouds by Heating From Below

5. The first step is to determine the convection temperature or the surface temperature that must be reached to start the formation of convection clouds by solar heating of the surface air layer. The procedure is to first determine the CCL on the plotted sounding and, from the CCL point on the T curve, proceed downward along the dry adiabat to the surface pressure isobar. The temperature read at this intersection is the convection temperature.

6. Fig 5.1 shows an illustration of forecasting afternoon convective cloudiness from a plotted sounding. The dewpoint curve was not plotted to avoid confusion. The dashed line with arrowheads indicates the path the parcel of air would follow under these conditions. You can see that the sounding was modified at various times during the day.

7. To determine the possibility of thunderstorms by the use of this method and from an analysis of the sounding, the following conditions must exist:

a. Sufficient heating must occur.

b. The positive area must exceed the negative area.

c. The greater the excess, the greater the possibility of thunderstorms.

d. The parcel must rise to the ice crystal level.

e. Generally, this level should be –10°C and below.

f. There must be sufficient moisture in the lower troposphere. This is the most important single factor in thunderstorm formation.

g. Climatic and seasonal conditions should be favourable.

h. A weak inversion (or none at all) should be present in the lower levels.

h. An approximate height of the cloud top may be determined by assuming that the top of the cloud will extend beyond the top of the positive area by a distance equal to one-third of the height of the positive area.

Formation of Clouds by Mechanical Lifting

8. When using this method, it is assumed that the type lifting will be either orographic or frontal, Here we will be concerned with the Lifting Condensation Level (LCL) and the Level of Free Convection (LFC). The LCL is the height at which a parcel of air becomes saturated when it is lifted dry adiabatically. The LCL for a surface parcel is always found at, or below, the CCL. The LFC is the height at which a parcel of air is lifted dry adiabatically until saturated, and thereafter would first become warmer than the surrounding air. The parcel will then continue to rise freely above this level until it becomes colder than the surrounding air.

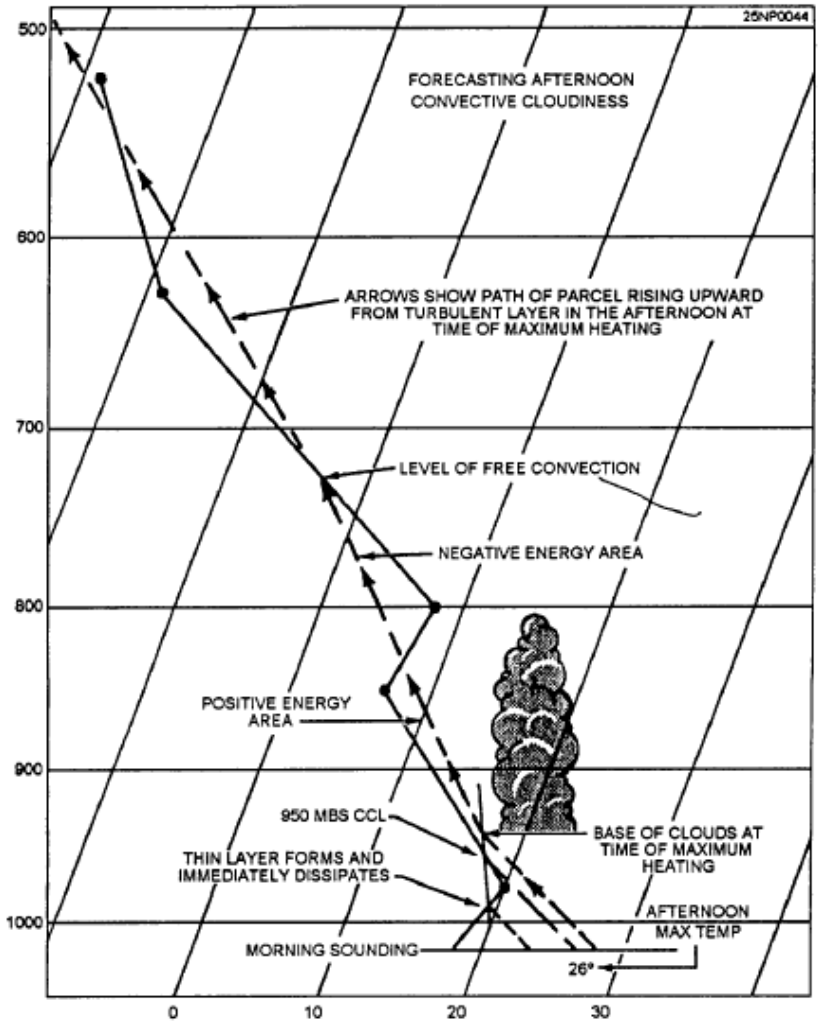


Fig 5.1 Forecasting afternoon convective cloudiness

9. Fig 5.2 shows the formation of clouds due to mechanical lifting. This figure shows the formation of a stratified layer of clouds above the LCL, and to the LFC. At the LFC and above, the clouds would be turbulent. The tops of the clouds extend beyond the top of the positive area due to overshooting, just as clouds formed due to heating. To determine the possibility of thunderstorms from this method, the following conditions should be met:-

a. The positive area must exceed the negative area; the greater the excess, the greater the possibility of thunderstorms,

b. There must be sufficient lifting for the parcel to reach the LFC, The frontal slope or the orographic barriers can be used to determine how much lifting can be expected,

c. The parcel must reach the ice crystal level (–10°C and below).

d. Even though the positive area does not exceed the negative area, cloudiness occurs after the parcel passes the LCL, and precipitation may occur after the parcel passes the ice crystal level.

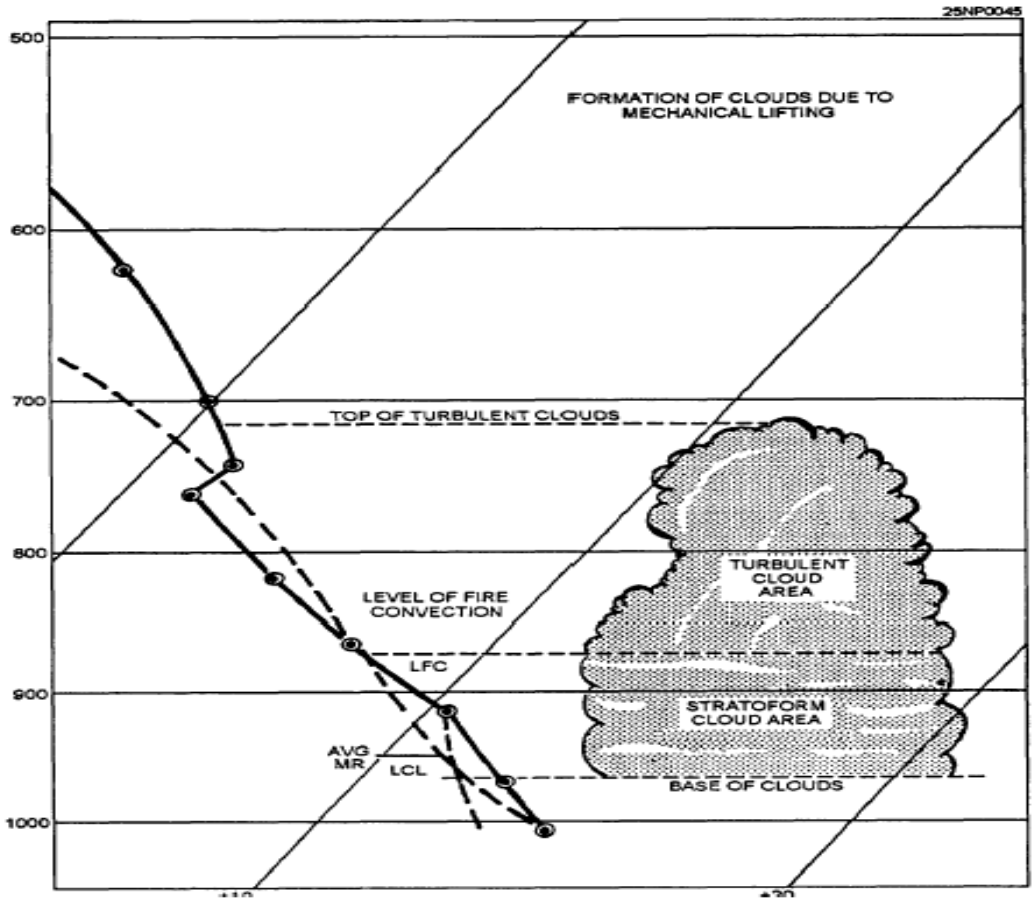


Fig 5.2 Formation of clouds due to mechanical lifting

10. One main advantage of this method is that it can be done quickly and with relative ease. A major disadvantage is that it assumes that the parcel does not change its environment, and that it overestimates or underestimates the stability and instability conditions.

Instability Indications from the Wet-Bulb Curve

11. A layer of the atmosphere is potentially unstable if the potential wet-bulb temperature decreases with altitude. Potential instability refers to a layer that is lifted as a whole. The wet-bulb temperature may be found by lifting each individual point on the sounding dry adiabatically to saturation, and then back to its original level moist adiabatically, By connecting the points on a sounding, a wet-bulb curve can be constructed.

12. If the wet-bulb curve slopes to the right with increasing altitude, the potential wet-bulb temperature increases with height, and the layer is potentially stable.

13. If it slopes to the left with increasing height, more than the saturation adiabats, the layer is potentially unstable. If none of the potential curves intersect the sounding, thunderstorms are not likely to occur.

Air-Mass Thunderstorms

14. This method is a reasonably simple method for forecasting air-mass thunderstorms. It does require prediction of short-range changes in the vertical distribution of temperature and moisture.

a. The first consideration is to eliminate those areas whose soundings disclosed one or more of the following moisture inadequacies.

b. The dewpoint depression is 13°C or more at any level from 850 through 700 hPa.

c. The dewpoint depression sum is 28°C or more at 700- and 600-hPa levels.

d. There is dry or cool advection at low levels.

e. The surface dewpoint is 60°F or less at 0730 local with no substantial increase expected before early afternoon.

f. There is a lapse rate of 21°C or less from 850 to 500 hPa.

g. There is a freezing level below 12,000 feet in an unstable cyclonic flow, producing only light showers.

h. After eliminating all soundings that meet one or more of the previous six conditions, you should use the following parameters to make the forecast.

j. The lapse rate between 850 and 500 hPa.

k. The sum of the dewpoint depressions at 700 and 600hPa in degrees C.

Note: The lapse rate is the difference in temperature between these two pressure levels. For example, if the temperature at 850 hPa was 15°C and at 500hPa it was –10°C, the difference would equal 25°. These two computations are used as arguments for the graph in fig 5.3.

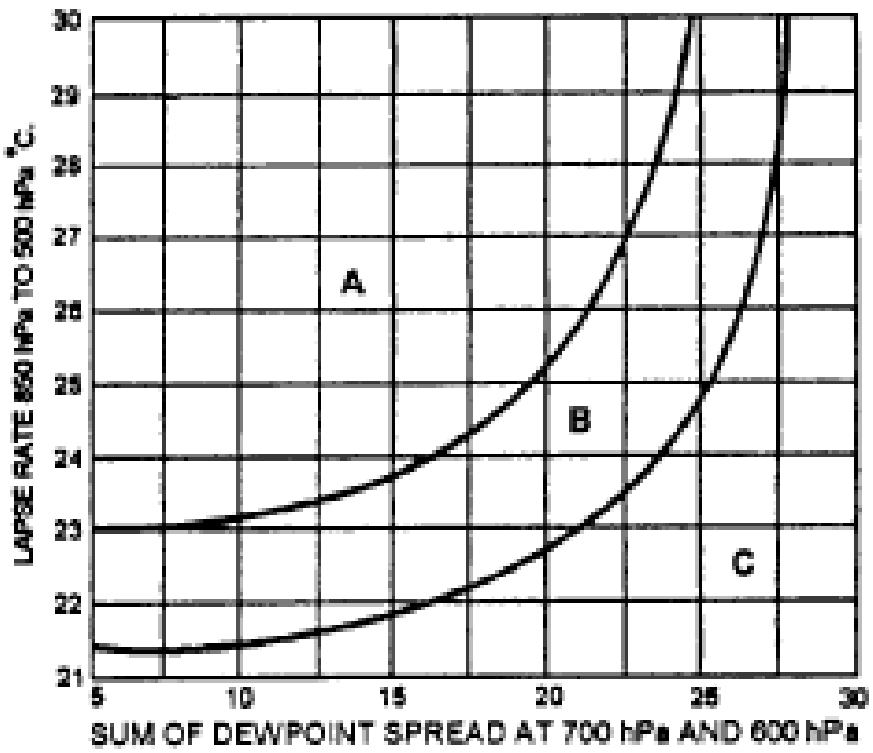


Fig 5.3 Local area thunderstorm graph. Area "A" is isolated thunderstorms, with a 12 to 1 chance of at least one rain gauge in dense network receiving rain. Area "B" is scattered thunderstorms, with a 4 to 1 chance of reported rain. Area "C" is no rain

15. One further condition for the development of thunderstorms is the absence of large anticyclonic wind shear, which is measured at 850 hPa. No horizontal shear at this level may exceed 20 knots in 250 miles measured toward the low-pressure area from the sounding station. Fig 5.4 illustrates how this measurement is made.

Stability Indexes as an Indication of Instability

16. The overall stability or instability of a rawinsonde sounding is sometimes conveniently expressed in the form of a single numerical value called the stability index. Such indexes have been introduced mainly as aids in connection with particular forecasting techniques. Most of the indexes take the form of a difference in temperature, dewpoint, wet-bulb temperature, or potential temperature in height or pressure between two arbitrarily chosen surfaces. These indexes are generally useful only when combined, either objectively or subjectively, with other data and synoptic considerations. Used alone, they are less valuable than when plotted on stability index charts and analysed for large areas. In this respect they have the value of alerting the forecaster to those soundings, routes, or areas that should be more closely examined by other procedures.

17. There are a number of methods that maybe used for determining stability or instability. Among these methods are the Showalter Stability Index (SSI), the Lifted Index (LI), the Fawbush-Miller Stability Index (FMI), and the Martin Index (MI).

18. Fig 5.5 shows the computation of the SSI. For forecasting purposes, the significance of the index values to forecasting is as follows:

a. When the index is +3°C or less, showers are probable, and some thunderstorms may be expected in the area.

b. The chance of thunderstorms increases rapidly for index values in the range of +1°C to -2°C.

c. Index values of -3°C or less are associated with severe thunderstorms.

d. When the value of the index is below –6°C, the forecaster should consider the possibility of tornado occurrence. However, the forecasting value of all index categories must, in each case, be evaluated in the light of the moisture content of the air and of other synoptic conditions.

Forecasting Movement of Thunderstorms

19. Radar can be an invaluable aid in determining the speed and the direction of movement of the thunderstorm. Sometimes it is desirable and necessary to estimate the movement from winds aloft. There is no completely reliable relationship between speed or direction of the winds aloft in forecasting thunderstorm movement. However, one study reveals that there is a marked tendency for large convective rainstorms to move to the right of the wind direction in the mean cloud layer from 850 to 500 hPa (or the 700-hPa wind) with a deviation of about 25 degrees. There appears to be little correlation between wind speed and speed of movement at any level, although the same study mentioned above revealed that 82 percent of the storms move within plus or minus 10 knots of a mean speed of 32 knots.

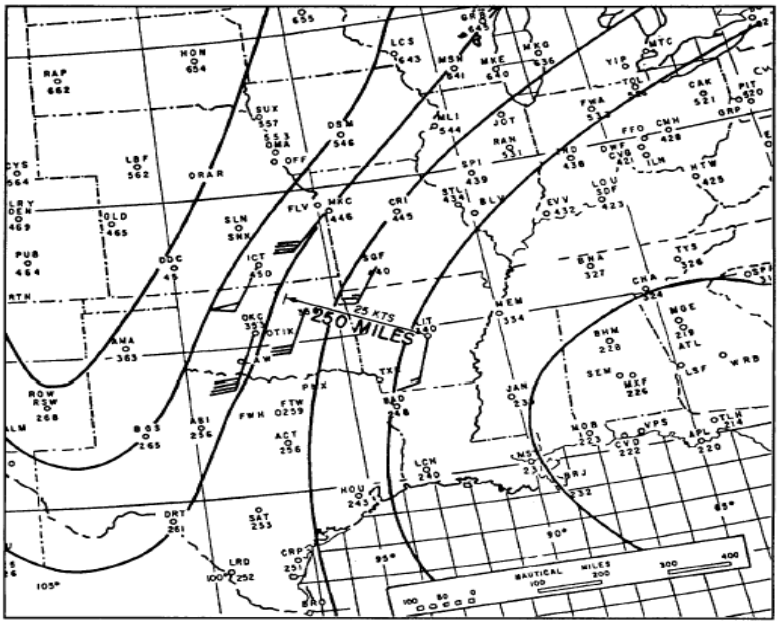


Fig 5.4 Example of anticyclonic shear and curvature at 850 hPa preventing thunderstorms, with otherwise favourable air mass conditions present in the Little Rock area

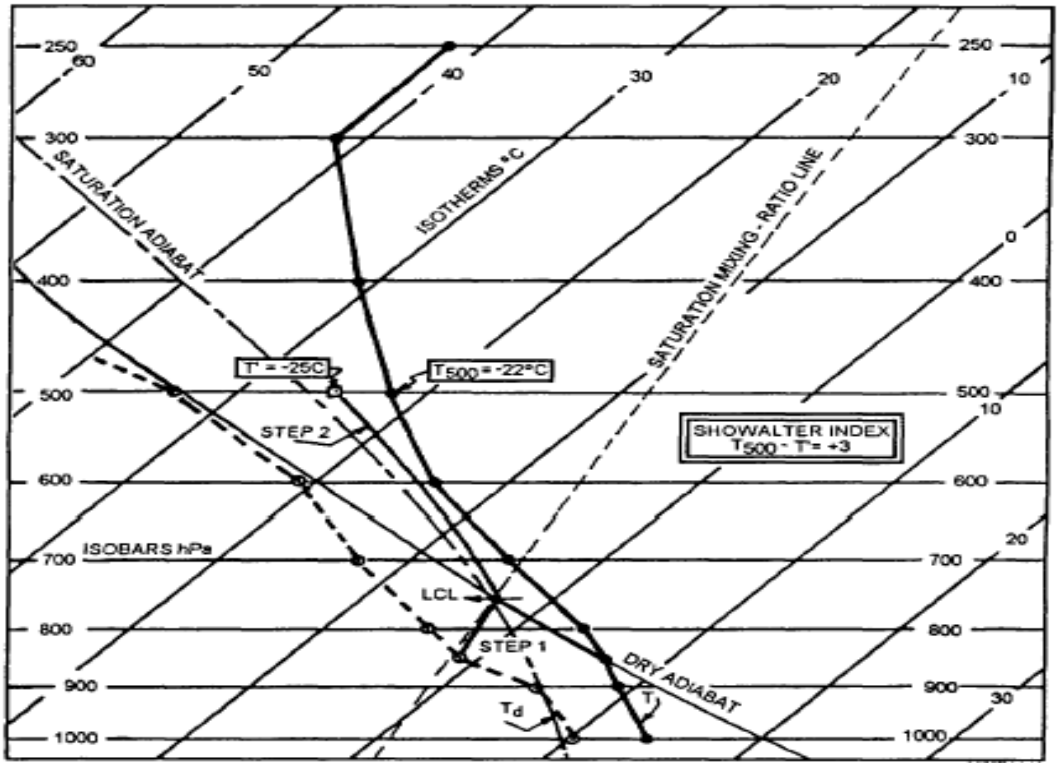


Fig 5.5 Computation of the Showalter Stability Index (SSI)

Forecasting Maximum Gusts with Non-frontal Thunderstorms

20. Maximum gusts associated with thunderstorms occur over a very small portion of the area in which the thunderstorm exists, and usually occur immediately before the storm's passage. Nevertheless, the possibility of damage to aircraft and installations on the surface is so great that every available means should be used to make the best and most accurate forecast possible to forewarn the agencies concerned.

Estimation of Gusts From Climatology and Storm Intensity

21. The forecaster is aware that the season of the year and the station location have a great bearing upon the maximum winds to be expected. Certain areas of the world, have a history of severe thunderstorm occurrence with associated strong winds during the most favourable seasons of the year. For this reason you should have the thunderstorm climatology for your station, as well as for the general area, available as to time of occurrence, season of occurrence, and the associated conditions.

22. The storm's intensity and reports from neighboring stations can give you a good indication of conditions to expect at your station.

TASK 19 : FORECASTING FOG AND STRATUS

Chapter objectives

After reading this chapter, you should be able to:-

- Assimilate the different method of Forecasting Fog and Stratus.
- Understand clouds formation under different conditions.
- Understand Factors to be considered in Fog and stratus formation.

Introduction

1. Fog and stratus clouds are hazardous conditions for both aircraft and ship operations. You will frequently be called upon to forecast formation, lifting, or dissipation of these phenomena. To provide the best information available, we will discuss the various factors that influence the formation and dissipation of fog and stratus.

General Procedure for Forecasting Fog

2. The synoptic situation, time of year, climatology of the station, air-mass stability, amount of cooling expected, strength of the wind, dewpoint-temperature spread, and trajectory of the air over favourable types of underlying surfaces are basic considerations you should take into account when forecasting fog.

Consideration of Geography and Climatology

3. Certain areas are more favourable climatologically for fog formation during certain periods of the year than others. All available information pertaining to climatology should be compiled for your station or operating area to determine the times and periods most favourable for fog formation.

4. You should also determine the location of the station with respect to air drainage or upslope conditions. Next, determine the type of fog to which your location would be exposed. For example, inland stations would be more likely to have radiation fog and shore coastal stations advection fog. A determination should then be made of air trajectories favourable for fog formation at your station.

Frontal Fog

5. Frontal fogs are associated with the forecasting of the movement of fronts and their attendant precipitation areas. For example, fogs can form in advance of a warm front, in the warm air section behind the warm front (when the warm air dewpoint is higher than the cold air temperature), or behind a slow moving cold front when the air becomes saturated.

Airmass Fog

6. The first step in the forecasting of air-mass fogs is to determine the trajectory of the air mass and estimate the changes that are expected to occur during the night. If the air mass has been heated during the day and there was no fog the preceding morning, no marked cloud cover during the day, and no over water trajectory, fog will not form during the night. However, during fall and winter, nights are long and days are short, and conditions are generally stable. When a fog situation has been in existence, the same conditions tend to remain night after night, and the heating during the day is insufficient to effectively raise the temperature above saturation. Also, determine if the air has had a path over extensive bodies of water, and whether this path was sufficient to raise the humidity or lower the temperature sufficiently to form fog. Then construct nomograms, tables, etc., by using dewpoint depression against time of fog formation for various seasons and winds; modify these in the light of each particular synoptic situation.

Factors to be Considered in Fog and Stratus Formation

7. Wind, saturation of the air mass, nocturnal cooling, and airmass trajectories have a role in the formation of fog or stratus clouds.

8. **Wind.** Wind velocity is an important consideration in the formation of fog and/or low ceiling clouds. When the temperature and dewpoint are near one another at the surface and eddy currents are 100 feet or more in vertical thickness, adiabatic cooling in the upward portion of the eddy could give the additional cooling needed to bring about saturation. Any additional cooling would place the air in a temporary supersaturated state. The extra moisture will then condense out of the air, producing a low ceiling cloud. Adiabatic heating in the downward portion of the eddy will usually evaporate the cloud particles. If all cloud particles evaporate before reaching the ground, the horizontal visibility should be good. However, if many particles reach the ground before evaporation, the horizontal visibility will be restricted by moderate fog. Clouds that form in eddy areas may at first be patchy and then become identified as *ragged stratus.* If the cloud forms into a solid layer, it will be a layer of stratus. When conditionally unstable air is present in the eddy, or if the frictional eddy currents are severe enough, stratocumulus clouds will form in the area. See fig 6.1.

9. **Saturation of the Air Mass.** The saturation curve in fig 6.2 shows the amount of moisture in grams per kilogram the air will hold at various temperatures. The air along the curve is saturated and is at its dewpoint. Any further cooling will yield water as a result of condensation; hence, fog or low ceiling clouds (depending upon the wind velocity) will form.

10. **Nocturnal Cooling.** Nocturnal cooling begins after the temperature reaches its maximum during the day. Cooling will continue until sunrise, or shortly thereafter. This cooling affects only the lower limits of the atmosphere. If nocturnal cooling reduces the temperature to a value near the dewpoint, fog or low clouds will develop. The wind velocity and terrain roughness will control the depth of the cooled air. Calm winds will allow a patchy type of ground fog or a shallow, continuous ground fog to form. Winds of 5 to 10 knots will usually allow the fog to thicken vertically. Winds greater than 10 knots will usually cause low stratus or stratocumulus to form. See fig 6.3 and I for examples of fog and stratus formation. The amount of cooling at night is dependent on soil composition, vegetation, cloud cover, ceiling, and other factors. Cloud cover based below 10,000 feet has a greenhouse effect on surface temperatures, absorbing some terrestrial radiation and

reradiating a portion of this heat energy back to be absorbed by the land. This causes a reduction in nocturnal cooling. Nocturnal cooling between 1530 local and sunrise will vary from as little as 5° to 10° (with an overcast sky condition based around 1,000 feet), to 25° or 30°F with a clear sky or a cloud layer above 10,000 feet. Other factors and exceptions must also be considered. If a front is expected to pass the station during the night, or onshore winds are expected to occur during the night, the amount of cooling expected would have to be modified in light of these developments.

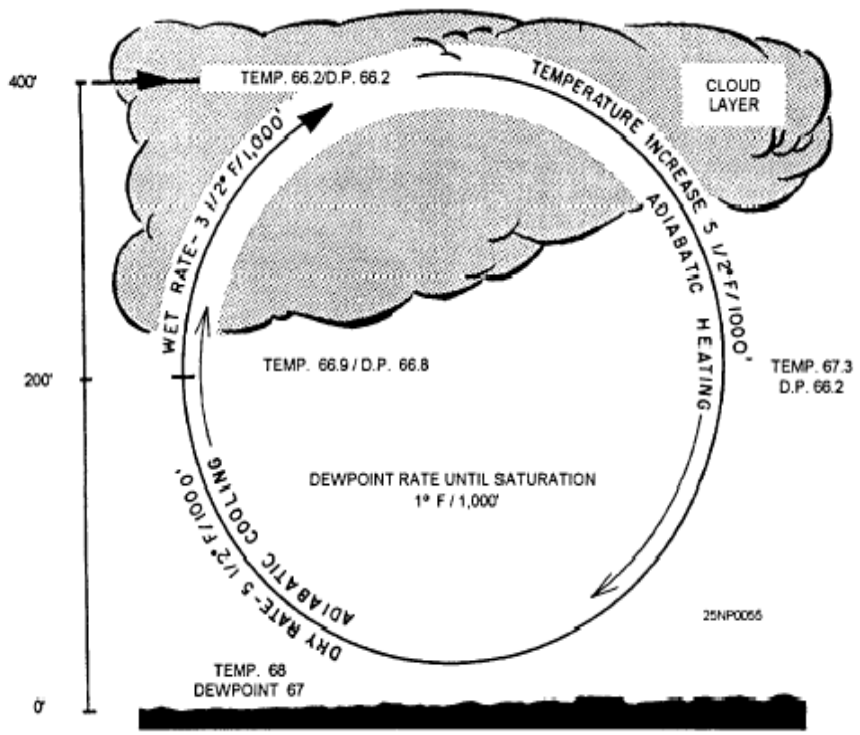


Fig 6.1 How wind velocity can cause a low cloud layer

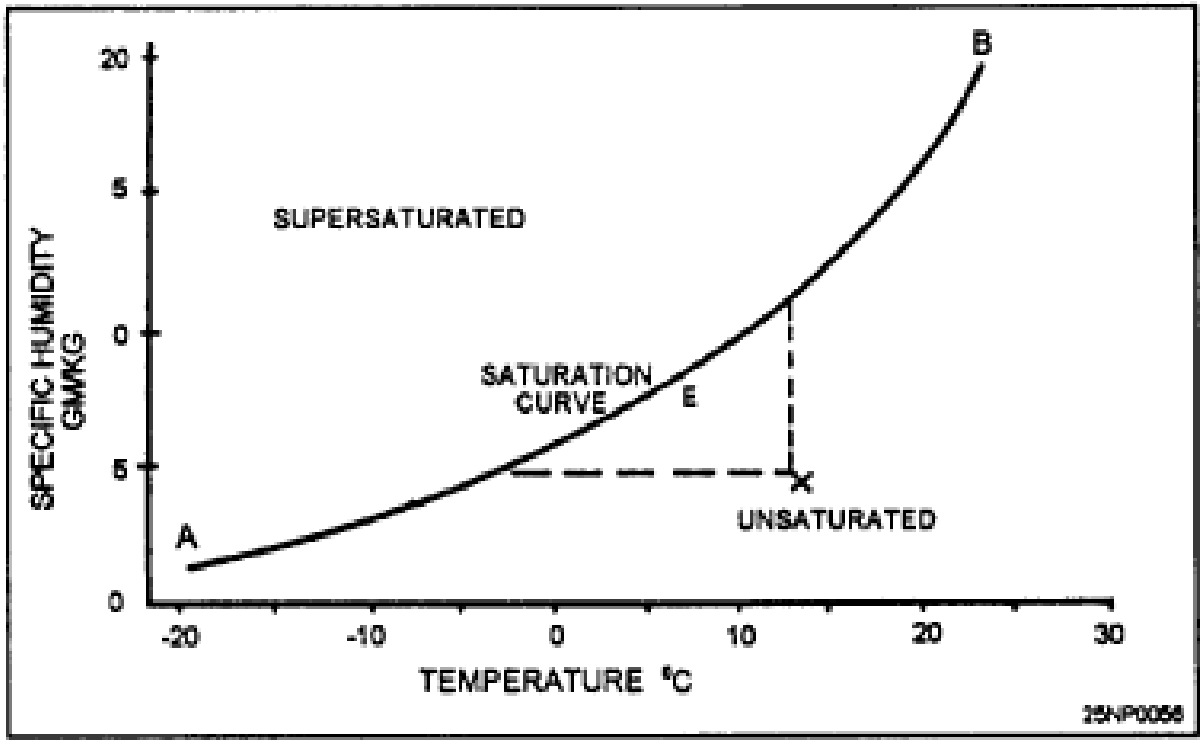


Fig 6.2 Saturation curve

11. **Airmass Trajectories.** The trajectory of an air mass during the forecast period can be another important factor in fog formation. Warm air moving over a colder surface is a primary fog producer. This can happen when a station is in the warm sector, following a warm frontal passage. Cooling of the air mass takes place, allowing condensation and widespread fog or low stratus to form. To determine the probabilities of condensation behind a warm front, compare the temperature ahead of the front with the dewpoint behind the front. If the temperature ahead of the warm front is lower than the dewpoint behind the front, the air mass behind the front will cool to a temperature ahead of the front, causing condensation and the formation of fog or low temperature near the stratus. Over water areas, warm air passing over cold water may cause enough cooling to allow condensation and the production of low clouds. Another instance in which trajectory is important is when cold air moves over a warmer water surface, marsh land, or swamp, producing *steam fog.* In addition, air passing over a wet surface will evaporate a portion of the surface moisture, causing an increase in the dewpoint. Whenever there is a moisture source present, air will evaporate a portion of this moisture, unless the vapour pressure of the air is as great, or greater, than the vapour pressure of the water. A dewpoint increase may be enough to allow large eddy currents, nocturnal cooling, or terrain lifting to complete the saturation process and allow condensation to occur.

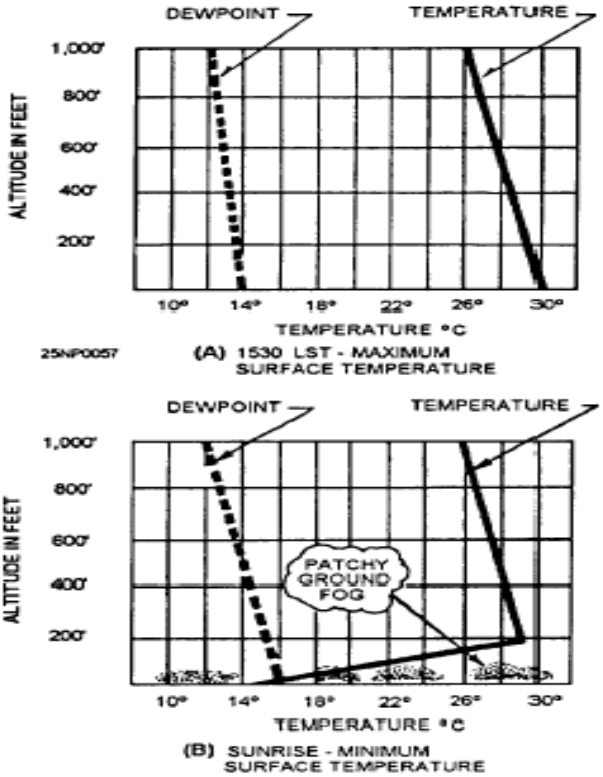


Fig 6.3 Nocturnal cooling over a land area producing patchy ground fog (calm winds) (A) 1530 Local max surface temperature; (B) sunrise-min surface temperature

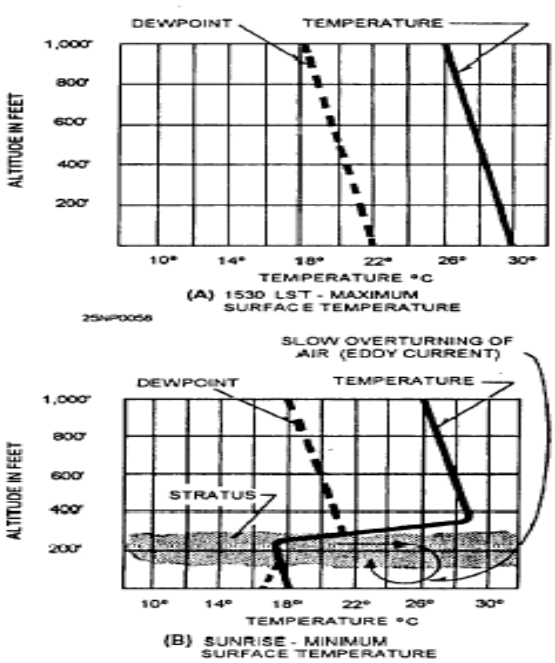


Fig 6.4 Nocturnal cooling over a land area producing stratus (10- to 15-knot wind) (A) 1530 local maximum surface temperature; (B) sunrise-minimum surface temperature

Use of the Skew T Log P Diagram in Forecasting the Formation & Dissipation of Fog

12. One of the most accepted methods for forecasting the formation and dissipation of fog makes use of an upper air sounding plotted on the Skew T Log P Diagram. The plotting of an upper air sounding is useful in forecasting both the formation and dissipation of fog, but it can be used more objectively in forecasting fog dissipation.

13. The use of an upper air sounding to determine the possibility of fog formation must be subjective. A study of the existing lapse rate should be made to determine the stability or instability of the lower layers. The surface layer must be stable before fog can form. If it is not found to be stable, the cooling expected during the forecast period must be considered, and this modification should be applied to the sounding to determine if the layer will be stable with the additional cooling.

14. The difference between the temperature and the dewpoint must be considered. If the air temperature and the dewpoint are expected to coincide during the period covered by the forecast, a formation of fog is very likely. The expected wind speed must be considered. If the wind speed is expected to be strong, the cooling will not result in a surface inversion favourable for the formation of fog, but may result in an inversion above the surface, which is favourable for the formation of stratus clouds.

Determination of Fog Height

15. An upper air sounding taken during the time fog is present will show a surface inversion. The fog will not necessarily extend to the top of the inversion. If the temperature and dewpoint have the same value at the top of the inversion, you can assume that the fog extends to the top of the inversion. However, if they do not have the same value, you can determine the depth of the fog by averaging the mixing ratio at the surface and the mixing ratio at the top of the inversion. The intersection of this average mixing ratio with the temperature curve is the top of the fog layer.

16. Two methods that may be used to find the height of the top of the fog layer, in feet, are reading the height directly from the pressure-height curve on the Skew T - Log P Diagram or by using the dry adiabatic method.

a. In using the pressure-height curve method, locate the point where the temperature curve and the average mixing ratio line intersect on the Skew T - Log P

Diagram. Move this point horizontally until the pressure-height curve is intersected. Determine the height of the fog layer from the value of the pressure-height curve at this intersection.

b. The dry adiabatic method is based on the fact that the dry adiabatic lapse rate is 1°C per 100 m, or 1°C per 328 ft. Using this method, follow the dry adiabat from the intersection of the average mixing ratio line with the temperature curve to the surface level. Find the temperature difference between the point where the dry adiabat reaches the surface and the point of intersection of the dry adiabat and the average mixing ratio. For example, in figure 15.18, the dry adiabat at the surface is 25°C. The temperature at the intersection of the dry adiabat and the average mixing ratio is 20°C. By applying the dry adiabatic method with a lapse rate of 1°C per 328 ft, we find the height of the top of the fog layer as follows:

$$1/328 = 5/x; \quad x = 1{,}640 \text{ ft}$$

Dissipation

17. To determine the surface temperature necessary for the dissipation of fog by using the Skew T - Log P Diagram, trace dry adiabatically from the intersection of the average mixing ratio line and the temperature curve to the surface level. The temperature of the dry adiabat at the surface level is the temperature necessary for dissipation. This temperature is known as the critical temperature. This temperature is an approximation, since it assumes no changes will take place in the stratum from the time of observation to the time of dissipation. This temperature should be modified on the basis of local conditions. See fig 6.5

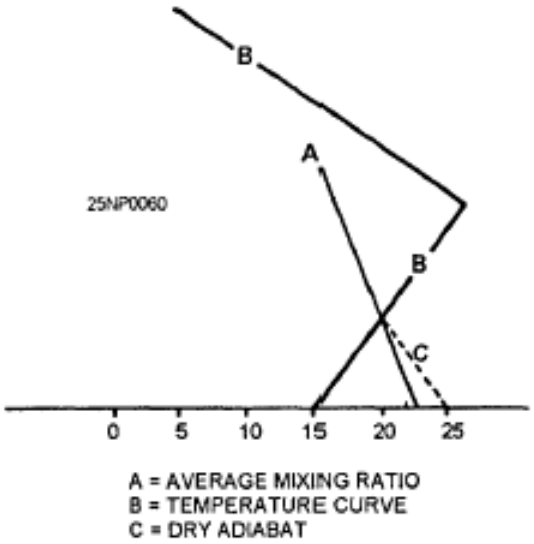


Fig 6.5 Dry adiabatic method of determining fog height

18. In considering the dissipation of fog and low clouds, you should consider the rate at which the surface temperature will increase after sunrise. Vertically thick fog, or multiple cloud layers, will slow up the morning heating at the surface. If advection fog is present, the fog may be lifted off the ground to a height where it is classified as stratus. If ground fog is present, the increase in surface air temperature will cause the fog particles to evaporate, thus dissipating the fog. Further heating may evaporate advection fog and low clouds.

Forecasting Advection Fog over the Oceans

19. In the absence of actual temperature and dewpoint data and with a stationary high (a southerly flow is assumed), use the following method to forecast advection fog over the ocean:-

a. Pick out the point on an isobar at which the highest sea temperature is present (either from the surface chart or a mean monthly sea temperature chart). Assume that at this point, the air temperature is equal to that of the water and has a dewpoint two degrees lower.

b. Find the point on the isobar northward where the water is two degrees colder. From this point on, patchy light fog should occur.

c. From a saturation curve chart (fig. 6.2), find how much further cooling would have to occur to give an excess over saturation of 0.4 gm per kg, and also 2.0 gm per kg. The first represents the beginning of moderate fog and the second represents drizzle.

d. As the air continues around the northern ridge of the high, it will reach its lowest temperature, and from then on will be subject to warming. The pattern will then be drizzle until the excess is reduced to 2.0 gm per kg, and moderate fog until 0.4 gm per kg is reached.

20. If actual water and temperature data are available, use these in preference to climatic mean data. If the high is moving, trajectories will have to be calculated. The fog is usually less widespread than calculated, and drizzle is less extensive. Also, clearing and lifting on the east side of the high is slightly faster. This method appears to work well in the summer over the Aleutian areas where such fog is frequent.

Forecasting Upslope Fog

21. Orographic lifting of the air will cause adiabatic cooling at the dry adiabatic rate of 5.5°F per 1,000 feet. If an adequate amount of lifting occurs, fog or low clouds will form. This process can create challenges for the forecaster.

22. The procedures for determining the probability of fog or low clouds during nighttime hours at stations having upslope winds are as follows:

a. Forecast the amount of nocturnal cooling,

b. Determine the expected amount of upslope cooling by using the following steps:

(1) Determine the approximate number of hours between sunset and sunrise.

(2) Estimate the expected wind velocity during the nighttime hours.

(3) Multiply (i) by (ii). This will give the distance the upslope wind will move during the period of the day when daylight heating cannot counteract upslope cooling.

(4) Determine the approximate terrain elevation difference between the station and the distance computed in c. Elevation difference should be in feet.

(5) Multiply the elevation difference by the dry adiabatic rate of cooling. (Example, 2.5 times 5.5 = 13.75°F of upslope cooling.)

(6) Add the expected amount of upslope cooling to the expected nocturnal cooling to arrive at the total amount of cooling.

(7) Determine the late afternoon temperature dewpoint spread at the station under consideration. If the expected cooling is greater than the late afternoon spread, either fog or low clouds should be expected. Wind velocity will determine which of the two conditions will form.

Forecasting Stratus Formation and Dissipation

23. Fog and stratus forecasting are so closely tied together that many of the fog forecasting rules and conditions previously mentioned also apply to the forecasting of stratus clouds.

Determining the Base and Top of a Stratus Layer

24. One of the first steps in forecasting the dissipation of stratus is to determine the thickness of the stratus layer. The procedure is as follows:

a. Determine a representative mixing ratio between the surface and the base of the inversion.

b. Project this mixing ratio line upward through the sounding.

c. The intersection of the average mixing ratio line with the temperature curve gives the approximate base and maximum top of the stratus. Point A in fig 6.6 is the base of the stratus layer, and point B is the maximum top of the layer. Point A is the initial base of the layer; but as heating occurs during the morning, the base will lift. Point B represents the maximum top of the stratus layer; although in the very early morning, it might lie closer to the base of the inversion. However, as heating occurs during the day, the top of the stratus layer will also rise and will be approximated by point B.

25. If the temperature and the dewpoint are the same at the top of the inversion, the stratus will extend to this level. To determine the height of the base and the top of the stratus layer, use either the method previously outlined for fog, or the pressure altitude scale.

Determining Dissipation Temperatures

26 To determine the temperature necessary for the dissipation of a stratus layer, the following steps are provided:

a. From point A in fig 6.6, follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the temperature required to be reached for stratus dissipation to begin. This is point C.

b. From point B in fig 6.6, follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the surface temperature required for the dissipation of the stratus layer to be complete. This is point D.

Determining Time of Dissipation

27. After determination of the temperatures necessary for stratus dissipation to begin and to be completed, a forecast of the time these temperatures will be reached must be made. Estimate the length of time for the required amount of heating to take place; and on the basis of this estimate, the time of dissipation may be forecasted. Remember to take into consideration the absence or the presence of cloud layers above the stratus deck. In addition, consider the trajectory of the air over the station. If the trajectory is from a water surface, temperatures will beheld down for a longer than normal period of time.

28. One rule of thumb used widely in forecasting the dissipation of the stratus layer is to estimate the thickness of the layer; and if no significant cloud layers are present above and normal heating is expected, forecast the dissipation of the layer with an average of 360 feet per hour of heating. In this way an estimate can be made of the number of hours required to dissipate the layer.

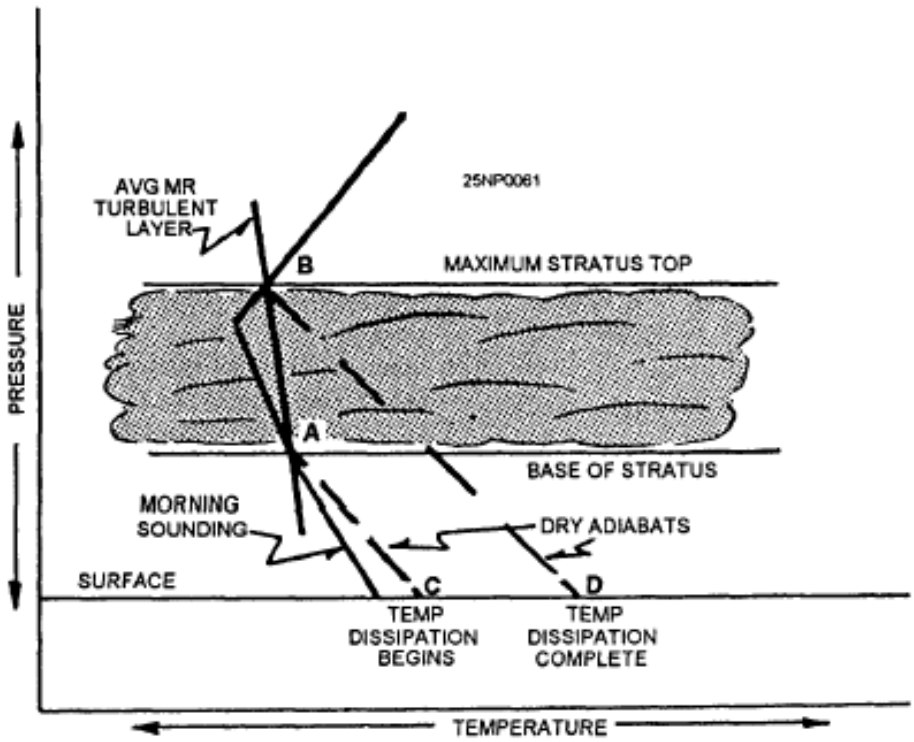


Fig 6.6 Sounding showing the base and the top of stratus layers. Also note temperature at which dissipation begins and temperature when dissipation is complete

TASK 20 : FORECASTING TEMPERATURE

Importance of Temperature.

1. The following are the important factors of temperature:

a. Minimum temperature is important for all types of aircraft because of formation of ice.

b. Runway temperature is important at the time of take-off and landing of aircraft specially jets.

c. Clothing requirements for troops.

d. For constructional purposes.

e. Chill factor.

f. Discomfort index : D.I. = 15+0.4 (T+Td).

g. Cold waves and heat waves.

h. In agriculture for forecasting frost.

j. Instrument shelter etc.

Factors affecting temperature

2. Horizontal advection:

a. **From or over water.** Modification in lower layers can effect the surface temperature.

b. **Over Land.** Air mass is modified when passing over land surface, because of temperature of the land and terrain.

3. **Meteorological Elements.**

a. **Winds**

(1) Direction.
(2) Speed.
(3) Duration.

b. **Cloud Cover**

(1) Amount.
(2) Thickness.
(3) Duration.
(4) Types.
(5) Height of base.

c. Dew Point temperature.

d. **Precipitation.**

(1) Height of formation.
(2) Types
(3) Evaporation, freezing or melting on the surface.

4. **Location**

a. Elevation.

b. Latitude.

c. **Environment:**

(1) Nearness of water bodies.
(2) Source region.
(3) Type of soil.
(4) Vegetations.

Factors Affecting Minimum Temperature

5. The following factors are affecting the minimum Temperature :

a. Air mass type during the coldest part of the day expected to pass over the station.

b. Trajectory over water keeps the minimum temperature up.

c. Trajectory over snow affects the minimum temperature.

d. Amount of cloud over. Increased cloud cover raise the minimum temperature.

e. Condensation and freezing raise the minimum temperature.

f. Evaporation lowers the minimum temperature.

g. Wind speed. Higher wind speed will raise the minimum temperature.

h. Frontal passages. Lowest minimum will occur during the second night after frontal passage.

j. Nearness of water bodies. Diurnal range of temperature,

(1) Over water 2°F
(2) Over the coast 8°F
(3) 300 miles inland 15°F.

Factors Affecting Maximum Temperature.

6. The following factors are affecting the maximum temperature :

a. Air mass passage during the hottest part of day.

b. Local winds (land breeze and sea breeze).

c. Cloudiness lowers the maximum temperature.

d. Precipitation lowers the maximum temperature.

e. Stability at the lower portion of the soundings will raise the maximum temperature.

f. Existence of snow cover has tendency to lower the maximum temperature.

g. Thermal convection or surface heating. Surface heating during day usually affects the lower 5000 feet of the atmosphere. Thus, the maximum will equal very nearly the potential temperature of the air at 850 mbs for stations near sea level. Mountain or plateau stations will have maximum temperatures very near that of the potential temperature at 700 mbs.

General Methods for Forecasting Temperature.

7. **Persistency.** If there is no day-to-day change of air mass, cloud cover, wind direction and speed, and insolation, there will be very little change in day-to-day temperatures.

8. **Advection.** On most days horizontal advection is the most important factor that produces surface temperature change at a given place. With rare exceptions, south winds bring warming, north winds bring cooling. Therefore, the surface chart determines the sign of the temperature change for the most part. Temperature raises ahead of a low and drops after its passage. If the pressure drops by 5 milibars in a 12 hour period an increase of temperature by 6-8°F may be expected. This rise in temperature follows rather than proceeds the fall in pressure. For a rather precise estimate of the advective temperature change, the forecaster can make trajectory calculations with use of the prognostic charts.

9. **Temperature Chart.** Temperature charts as proposed at numerous forecasting centers include the maximum temperature for the preceding day, the minimum temperature for the preceding night, and the 24 hour changes of both the temperatures. Care must be used in extrapolating areas of temperature change as depicted on those charts. For example, on area of rising temperature in a southerly flow may show little continuity as increasing moisture and extensive cloud develop and diminish the diurnal temperature rise. Usually, we can extrapolate lines of zero temperature change with relative success. We can also express the temperature chart in terms of deviation from average and track the motion of centers of positive and negative departure. The sign of these centers is preserved over appreciable periods during their motion.

10. **Climatological Data.** The forecaster should be familiar with the pertinent climatology of the region for which he makes forecasts. He should be aware, for example, the he is forecasting record-breaking temperatures when he does so. In recent years there has been considerable discussion concerning the development of so-called synoptic climatology as an aid in forecasting.

11. **Statistical Charts.** Those charts give frequency distribution of temperatures and temperature changes. This is useful for forecasting of temperatures for periods more than 48 hours.

Forecasting of Minimum Temperature

12. **Beyden's Method**

a. **Restrictions.**

(1) No air mass change between the time of maximum temperature and minimum temperature.

(2) No precipitation.

(3) T_w 0°C.

(4) Data of 1700 local time.

b. **Formula:**

T_{min} = T_w + T_d +, where

T_w = Wet-bulb temperature

T_d = Dew point temperature

K = The refinement factor found from a scatter diagram using the observed cloud cover at 1700 LT and the forecast wind speed at 1700 LT and the forecast wind speed at 23-2400 hours.

13. **Keysser's Method**

a. **Restrictions.**

(1) No air mass change.
(2) Little cloudiness.
T_w at 1700 LT is the T_{min} on the following night.

14. **Nichol's Method.** With same restrictions as in Keysser's, T_d at the time of maximum temperature will be the minimum temperature on the following night.

15. **Smith's Method**

a. **Restrictions.**

(1) Few Clouds.
(2) Wind less than 5 knots.
(3) No air mass change.

b. **Formula.**

$T = T_{MAX} - T_{MIN}$ where

T_{MAX} = Monthly average of maximum temperatures.

T_{MIN} = Monthly average of minimum temperatures.

Again,

$T_{MAX} - \Delta T = T_{MIN}$ where

T_{MAX} = Maximum temperature of the day.

Self Assessed Exercise on Task 20

Q1. What factors affect the temperature ?

Q2. What factors affect the minimum temperature ?

Q3. What factors affect the maximum temperature ?

Q4. What is indicated by "T" in Smiths" formula ?

Q5. How "T" in Energy Triangle method of forecasting maximum temperature expressed?

Answer Self Assessed Exercise on Task 20

A1. See Para 2, 3 and 4.

A2. See Para 5.

A3. See Para 6.

A4. See Para 15b.

A5. See Para 16, Step 4.

SECTION – 5 : SATELLITE METEOROLOGY

CHAPTER-21 : SATELLITE IMAGERY INTERPRETATION

Introduction

1. Satellites is a 20th century additions to our arsenal of meteorological data-gathering equipment. Radar came into existence during World War II, while meteorological satellites were first launched in 1960. Meteorologists have studied the imagery produced by satellite sensors and radar in order to correlate it to the weather occurring at the time. The chapter covers environmental satellite analysis, interpretation of weather echoes on radar and the lightning detection and tracking system presentations.

2. Satellite imagery is very important because it provides weather information for many areas of Earth that have no weather observing stations or the observing network is inadequate. We just finished discussing one such area- the tropics. The polar latitudes are also lacking in their number of reporting stations. Even in mid-latitudes, there are limited weather reports from over the vast oceanic areas. You will find satellite pictures extremely helpful in these areas, as well as in the more dense reporting areas. More often than not, imagery will provide the first indication of developing low-pressure centres, fronts, squall lines, and many other phenomena. These features can be located and tracked, thereby lending credence to your weather charts and briefings.

Environmental Satellite Imagery Analysis

3. Satellite pictures, whether received via real-time tracking equipment, facsimile circuits, or video monitors, are now commonplace in weather offices. They are tools that can make your overall job as an analyst easier, but first you must learn to use the tools. Your goals should be to correctly interpret the imagery and to incorporate it with other analyses and briefings. The following text covers some of the most important features found in satellite imagery. Realise that this material is not all inclusive.

Types of Imageries

4. Today's weather satellites are equipped with sensors capable of providing us with visible and infrared pictures. This dual sensing capability permits picture-taking both day and night. It also permits imagery switching from one picture to the next in order to compare the IR and visible pictures of the same region. You will note as you look at compared pictures that there can be quite a bit of contrast between visible and IR pictures of the same features. Therefore, to correctly analyse satellite pictures, you must have an understanding of the various types of imagery.

Visible Imagery

5. Visible satellite photos look very much like standard black-and-white photographs. Actually, they are constructed line by line from digital information gathered by the satellite sensor. The pictures are produced in black and white and various shades of gray. They are a measure of the Sun's reflection off clouds, land, and sea surfaces. The amount of reflectivity is dependent, on (a) the position of the satellite in relation to the Sun, (b) the reflective power of the surfaces being scanned by the camera, and (c) to some extent,

moisture. With regard to a satellite's relative position to the Sun, early morning and late evening visible pictures sometimes show up partially or totally black because the Sun is not high enough above the horizon in relation to the area of Earth being scanned by the camera.

Figure 5.1. Partial picture owing to darkness (sunset)

6. Figure 5.1 is an example of such an occurrence. There is simply not enough sunlight to permit a picture of this area. The reflective power of a surface or body (albedo) is a factor that makes features highly visible or practically invisible. Surfaces that are highly reflective (high albedos) show up bright white in pictures, while surfaces with low albedos appear dark. Table 5.1 lists the reflective power of some features. The last factor affecting reflectivity is moisture. Drier areas of Earth, such as deserts, reflect more sunlight than moist areas and in comparison appear brighter.

Table 5.1. Reflectivity of Various Surfaces

Surfaces		Percentage of Reflectivity
1	Large Thunderstorm	92 %
2	Fresh New Snow	88 %
3	Thick cirrostratus	74 %
4	Thick stratocumulus	68 %
5	White Sands	60 %
6	Snow, 3-7 days old	59 %
7	Thin stratus	42 %
8	Thin cirrostratus	32 %
9	Sand, no foliage	27 %
10	Sand and brushwood	17 %
11	Coniferous forests	12 %
12	Water surfaces	09 %

7. In addition to the required sunlight, there are other factors that control the appearance of features in visible imagery. The height of the satellite above Earth controls the area of coverage and the size of the features seen in imagery; the higher the satellite, the greater the coverage and the smaller the features. Camera resolution controls imagery detail; the smaller the resolution, the greater (finer) the detail. The focal point of the camera system defines the most useable area of the picture; features at the edges of pictures are not as easily interpreted as those near the picture's centre (the focal point).

Infrared (IR) Imagery

8. Infrared camera systems (scanning radiometers) permit the viewing of Earth's weather in darkness as well as light. Infrared imagery and enhanced infrared imagery are created by a computer using the radiometer's measurements of electromagnetic radiation emitted from Earth's surface features and clouds. The computer converts the radiation readings to temperatures and assigns shades of gray based on a temperature-gray scale. In unenhanced imagery, the lower radiation measurements are synonymous with colder temperatures, while the higher readings relate to warmer temperatures. The gray shades assigned range from pure white (coldest) to black (warmest). An unenhanced IR picture is shown in figure 5.2. In terms of Earth and its cloud cover, the following features will usually appear in unenhanced IR imagery:-

a. Hot deserts — black.

b. Lakes and oceans - near black. Clouds with low tops (Cu, Sc, St) - dark gray.

c. Snow, ice, and clouds with medium tops (As, Sc) - light gray I Clouds with high tops (Cb, TCu, Ci) – white.

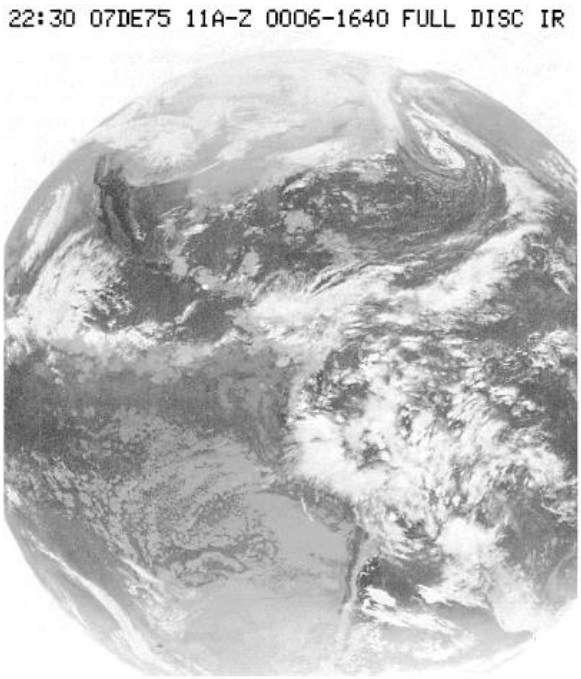
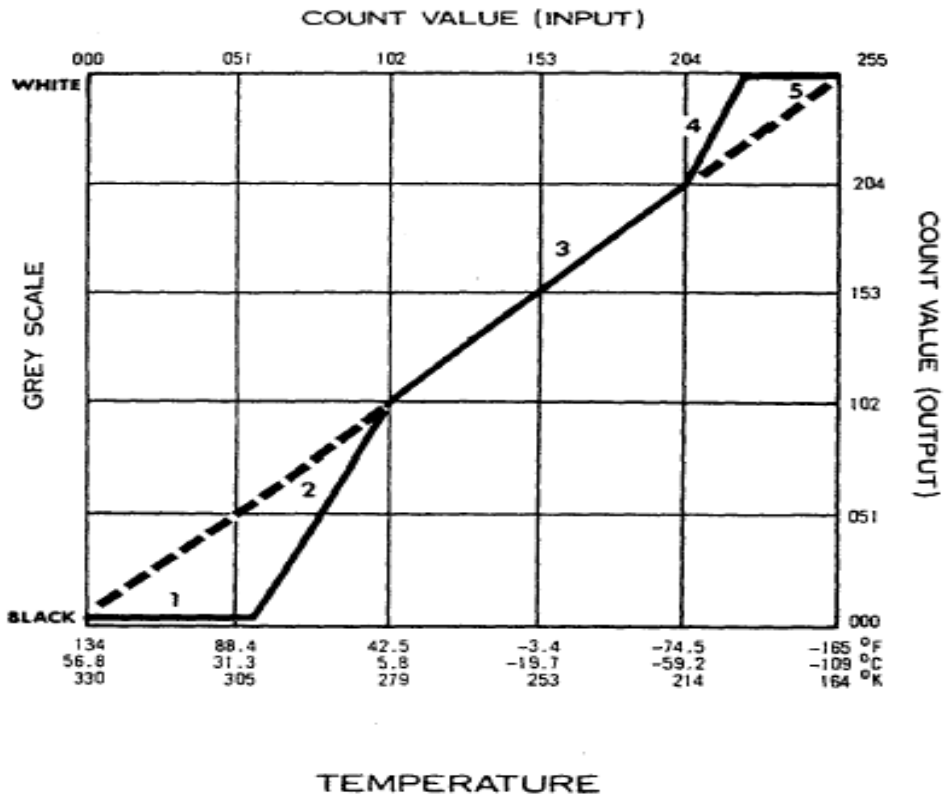


Fig 5.2. Unenhanced IR picture

Enhanced IR Imagery

9. An enhancement process is simply a modification of the temperature-gray scale in order to better differentiate features seen by the IR camera. The computers used with the Defence Meteorological Satellite (GOES) systems are programmed with many enhancement curves (temperature-gray scales). Figure 5.3 shows the difference between the unenhanced temperature curve (actually a straight line) and an enhanced curve known as the ZA curve. You will notice the ZA curve modifies the shading assigned at the top and bottom of the temperature spectrum. The very warm temperatures in *Segment 1* (56.8°C to 29.3°C) are all shaded black, while the very cold temperatures in *Segment 5* (– 75.2°C and colder) are all pure white. *Segments 2* and *4* are assigned a greater number of gray shades over smaller temperature ranges to increase the contrast of features seen within these ranges. The finer definition in shading in these areas makes for easier recognition of significant cloud and surface features, such as cirrus and stratus decks, fog, haze, ocean current boundaries, and terrain features. *Segment* 3 uses the same shading as that of the unenhanced curve, because the unenhanced curve does a good job over this range of temperatures. As you can see, the ZA curve only slightly modifies the unenhanced curve; therefore, IR pictures produced using this curve are interpreted in much the same manner as unenhanced IR pictures.



SEGMENT NUMBER	°C TEMPERATURE TO	COMMENTS REASON FOR SEGMENT ENHANCEMENT
1	56.8 to 29.3	Little or no Met. Data
2	28.8 to 6.8	Low level/Sea Surface Difference
3	6.3 to −55.2	Middle level—no enhancement
4	−56.2 to −75.2	Upper Level enhancement
5	−75.2 to −110.2	Little or no Met. Data

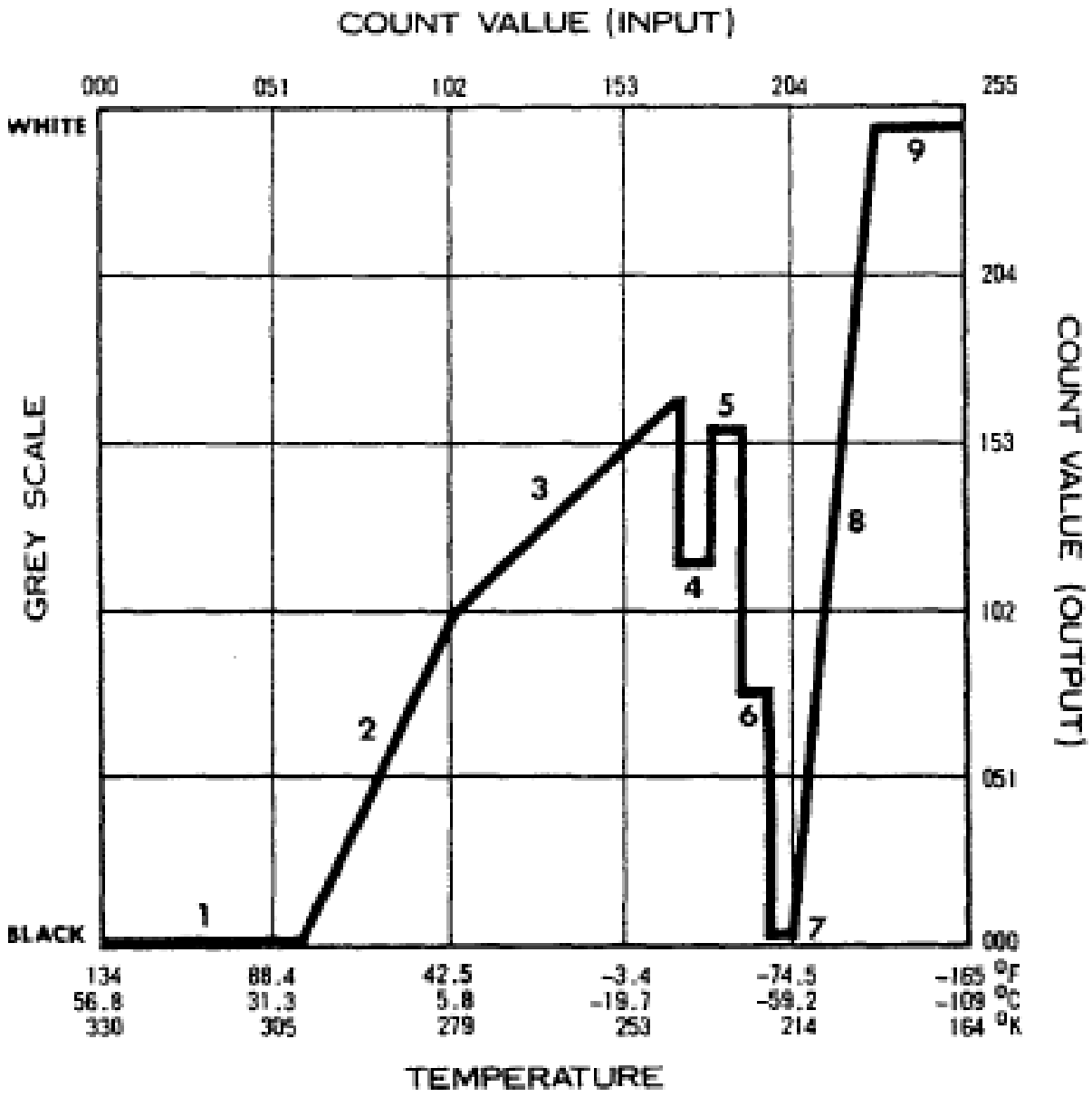
Fig 5.3. Enhanced IR temperature curve compared to IR unenhanced temperature curve

10. For a contrast to the slightly modified enhancement curve, you should look at an MB enhancement curve. See figure 5.4. Notice the degree of modification. This curve better defines areas of convective activity, and it is used to estimate rainfall amounts. Figure 5.5 is an example of the imagery produced by the MB curve. This picture shows how the shading assigned by *Segments* 4 through 7 of the curve accentuate an area of convective activity. There are many enhancement curves used to produce GOES and DMSP (Defence Meteorological Satellite Program) imagery.

11. Infrared imagery (enhanced or unenhanced) provides far more information to meteorologists and oceanographers than its visible counterpart. However, a far superior degree of interpretation is possible when infrared and visible imagery are viewed together, since both exhibit strengths and weaknesses in differing conditions. A product of such comparisons is now incorporated in the GOES system. The computer combines unenhanced visible imagery with enhanced IR imagery to produce what is known as a composite picture. Composite imagery combines the best qualities of both visible and IR imagery. These pictures are primarily used in convective storm analysis and as a forecasting aid.

Effects of Naturally Occurring Temperature Variations on IR Pictures

12. Large diurnal temperature variations over land surfaces produce quite a contrast in daytime and nighttime IR pictures. In the daytime, warm land appears darker in IR imagery and contrasts well with cool bodies of water, which appear gray, and with cold middle and high clouds, which appear white. As the land cools at night, its shading becomes much lighter in the IR image. The closer the land temperature to the temperature of water and/or clouds, the smaller the contrast between these features and the less obvious their boundaries. If land and water surfaces become the same temperature, the boundaries disappear. Seasonal and latitudinal temperature variations are another cause of variation in the appearance of IR imagery. In the winter, especially at night, clouds over cold land are not easily detectable because they are often in the same temperature range as the cool land. On the other hand, in the tropics, there is always a large temperature contrast between the surface and middle and high clouds, so these clouds are easily recognisable. The effects of diurnal, latitudinal, and seasonal temperature variations can be compensated for, to a degree, by the use of IR enhancement curves.



SEGMENT NUMBER	°C TEMPERATURE	COMMENTS
1	58.8 to 29.3	Little or no useful Met Data (Black)
2	28.8 to 6.8	Low Level/Sea Surface Difference
3	6.3 to -31.2	Middle Level - No Enhancement
4	-32.2 to -42.2	First Level Contour (Med Gray)
5	-43.2 to -53.2	(Light Gray)
6	-54.2 to -59.2	Thunderstorm (Dark Gray)
7	-60.2 to -63.2	Enhancement (Black)
8	-64.2 to -80.2	Overshooting Tops Enhancement
9	-81.2 to -110.2	(White)

Fig 5.4. MB enhancement curve

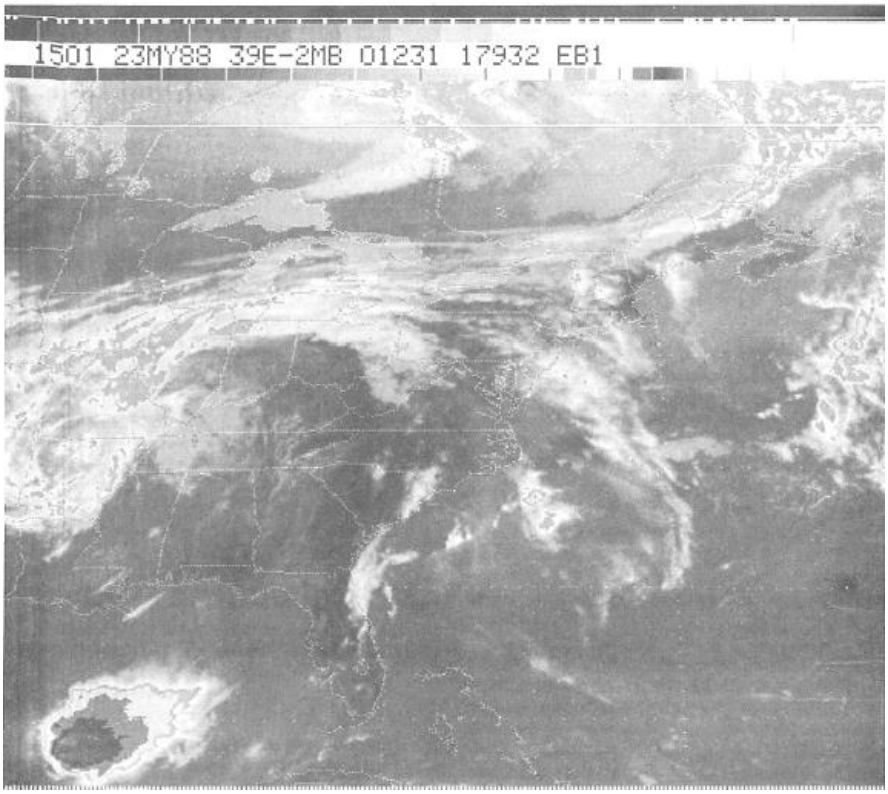


Fig 5.5. EIR picture using MB temperature curve

Water Vapour (WV) Imagery

13. Water vapour imagery is derived from radiation at wavelengths around 6-7µm. This is not an atmospheric window but is a part of the spectrum where water vapour is the dominant absorbing gas. It has absorption band centred on 6.7 µm. In wavelength regions of strong absorption, most of the radiation reaching the satellite originated high in the troposphere. The stronger the absorption, the higher is the originating level of the emission that ultimately reaches the satellite. As the relative humidity decreases, so the main contribution to the radiance received at the satellite comes from lower in the troposphere. Water vapour imagery is usually displayed with the emitted radiation converted to temperature, like normal IR imagery. Since temperature decreases with height, regions of high upper tropospheric humidity appear cold (light) and regions with low humidity appear warm(dark). In other words, when the upper troposphere is dry, the radiation reaching the satellite originated from farther down in the atmosphere, where it is warmer and appear darker on the image.

TASK-22 : INTERPRETATION OF EARTH SURFACE

Objectives

After reading this chapter, you should be able to:

- Interpret the land and sea surface features in satellite imageries
- Identify the snow over mountains and distinguish it from clouds
- Identify ship trails and sun glint in the images and estimate the sea surface conditions
- Understand the basic principles behind interpretation of dust haze and use infrared, watervapour and 3.7µm imageries

Structure

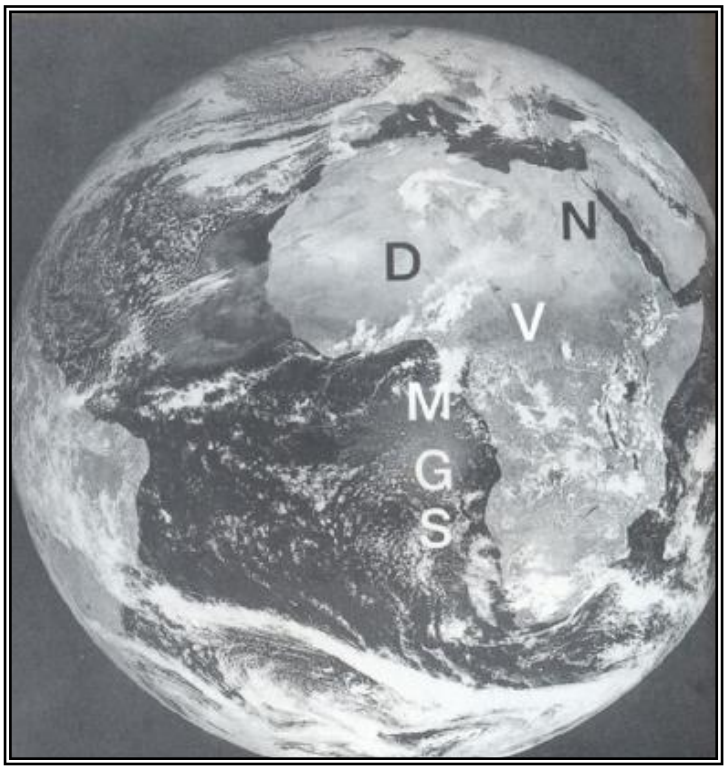
1. Introduction
2. Land Surface Features
3. Sea Surface Features
4. Cloud Features Modified by Atmospheric Pollutants

Introduction

1. The brightness of the earth's surface in visible imageries depends on the time of the day, season of the year, the geographical location etc. Clouds are brighter with high sun angle than when solar elevation is low. Decreasing solar elevation will result in reduction in the level of illumination and a corresponding reduction in the brightness of cloud response will occur across the image in the direction of decreasing sun angle. However at low sun angles the sides of deep convective clouds or thick cloud layers can still produce quite bright responses leading to highlights appearing within at the relatively darker portions of the image. It is easy to distinguish dry land from a water surface and to identify shorelines, rivers, lakes, islands etc. Dry land, depending on the type of relief, assumes various grey shades in visible imageries. The tone of the underlying surface in IR imageries depends on temperature and so its appearance is affected on the latitudinal, diurnal and seasonal variations. During day light hours and in summer dry land surfaces appear darker than the water surface but at night it has a lighter tone than a water surface. IR imageries reveal only large irregularities of relief of vegetation pattern, which are associated with marked temperature gradients whereas visible pictures can reveal even small terrain variations.

Land Surface Features

2. **Deserts and Vegetated Areas.** Dense vegetated areas appear relatively dark in visible imageries. Mountains with thick forests can be easily identified when it is surrounded by sparsely vegetated plains. In deserts with very sparse vegetation combined with red and yellow soils and rocks makes the earth's surface highly reflective. These areas appear brightest in satellite pictures as shown in Fig. 4.1. Surface temperature differences between deserts and vegetated regions are often modified diurnally by watervapour absorption in moist low levels of the atmosphere. Hence the radiation emitted from the surface reaches the satellite undiminished only when the atmosphere is dry. When it is moist the radiation from the high cold levels only reaches and hence the IR imagery is not much influenced by the surface characteristics.



D - Sahara Desert
V - Vegetation
M - Meteosat Sub Point
G - Glint
S - Meteosat Sub Point
N - Nile

Fig.4.1 : Meteosat Vis image at 1200UTC /09 Feb 88

3. **Snow**. A knowledge of surface geography together with an appreciation of the climatological variation of snow cover through the seasons is a basic prerequisite for the analysis of satellite imagery in respect of snow. The tone of snow on visible imagery varies from bright white to light grey depending on the nature of the relief, vegetation, the age of snow and the illumination. In areas of relatively flat terrain without trees snow surfaces appear uniformly white in tone. In areas covered by extensive forests a snow covered surface appears patchy, brighter patches correspond to areas where there are no trees and dark patches to forest areas. In IR imagery, snow covered terrain often appears whiter than its surroundings. The brightness of the snow area depends on the temperature of the surrounding ground, which is not snow covered, whether there is vegetation within the snow-covered area, the type of vegetation, and on how much of the vegetation is snow covered. At night the snow surface cools more rapidly than its surroundings and during the day it warms gradually. In visual pictures, differentiating between snow, ice, fog, clouds, and land is difficult at best. Differentiating between snow and ice is hard, and it becomes more difficult when the ice field is old. Snow cover on mountains is the easiest to detect in

visual pictures, as the valleys are dark (due to vegetation) while the mountain ridges are bright (due to the reflection from the snow). This combination of bright snow and dark valleys produces a dendritic (branching) pattern, as seen in Fig. 4.2. With regard to albedo, snow shows up very bright than clouds and ice fields. Ice has an albedo slightly less than clouds, and it shows up in gray shade when it is beside snow. Frozen lakes appear as bright spots in visual pictures. They are easily distinguished from clouds in sequential satellite pictures, because they are stationary and the clouds move. The same method of comparison is used to determine snow and ice fields from clouds.

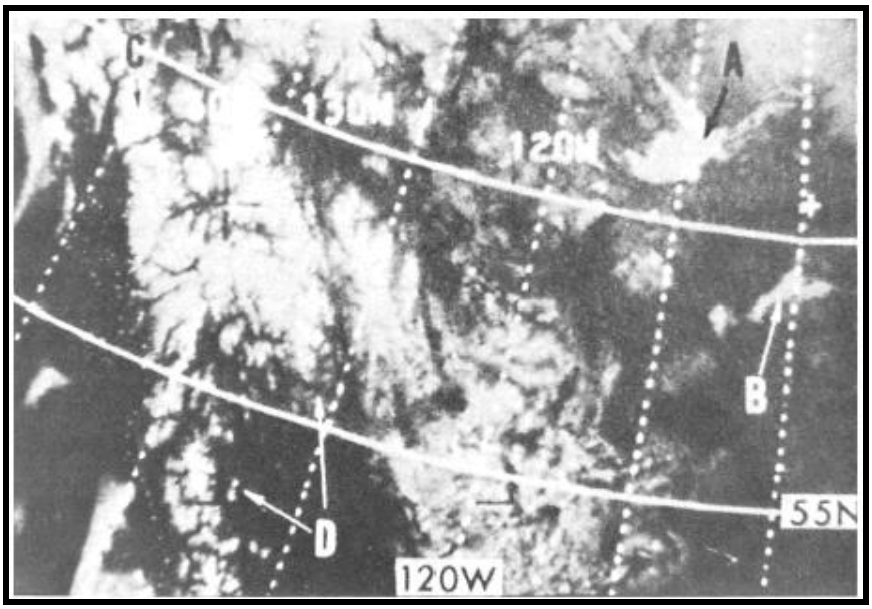


Fig. 4.2: Dendritic pattern created by snow on mountain tops

Sea Surface Features

4. **Sea Surface - Sun Glint**. The reflection of the sun on water surface can appear either as a large dull diffuse area or as a small bright spot. Variations in size and brightness of a sun glint are dependent on the roughness of water surface on which it appears. The brightest spot usually indicate a smooth undisturbed sea surface. Sunlight is reflected from the water surface such as the oceans and large lakes, and this reflecting phenomenon is called sun glint. On the visible image, it appears as a large bright region and its position is depending on season and time of day. The size and intensity of sun glint alters according to the water surface conditions, that is, the sun glint is small and bright for a sea surface not waves under calm winds. On the other hand, if the winds are intense and the sea surface waves are high, the sun glint is large and dark. That is, the sea surface conditions can be known through the sun glint.

5. Fig. 4.3 shows the mechanism of sun glint. A calm sea surface acts as a mirror. Sunlight reflected from the sea surface at the point B is directly incident upon the sensor of the satellite. On the other hand, sunlight reflected at the point A is not directly incident upon the sensor of the satellite. That is, the satellite sees directly the sunlight reflected at the point B, which is seen as the brightest region (sun glint) on the visible region. A sun glint moves over the image from east to west within a day. Across the seasons, its center swings between 11.75° N and 11.75° S.

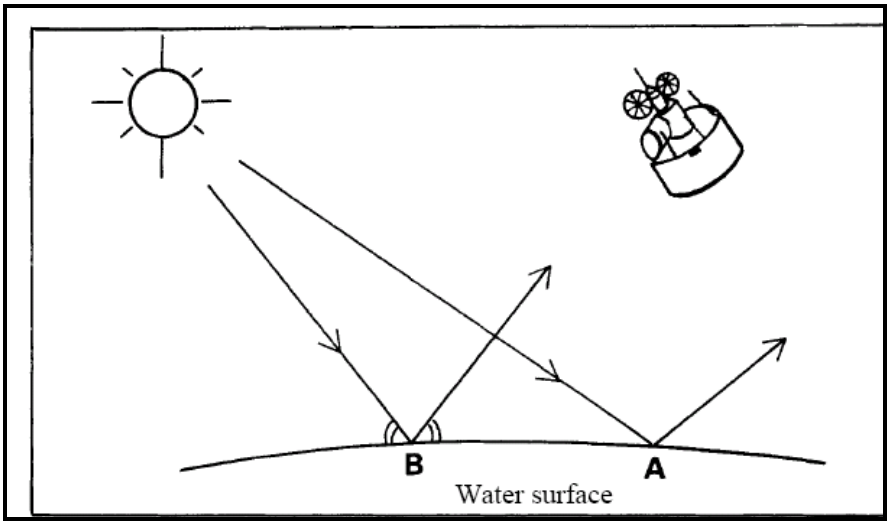


Fig. 4.3: Mechanism of sun glint

6. On the visible image in Fig. 4.4, the shining region around 10° N and 140° E (marked by a wedge in the figure) is sun glint. Since this sun glint is small and bright, the sea surface around here can be estimated to be calm. The position of sun glint lies in the northern hemisphere because it was July, and around 140° E because the image was taken at noon (Japan Standard Time). In Fig. 4.5, the sun glint lies around 10° S and 160° E (marked by a wedge in the figure). The position of the sun glint is on the southern hemisphere because it was December and around 160° E because the image was taken at 900 (Japan Standard Time).

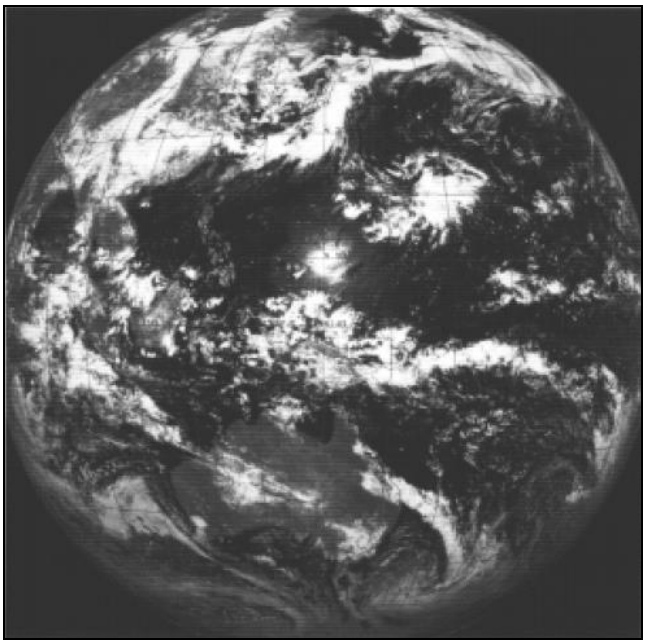


Fig. 4.4 : Visible image at 03UTC, 22 Jul 98 .

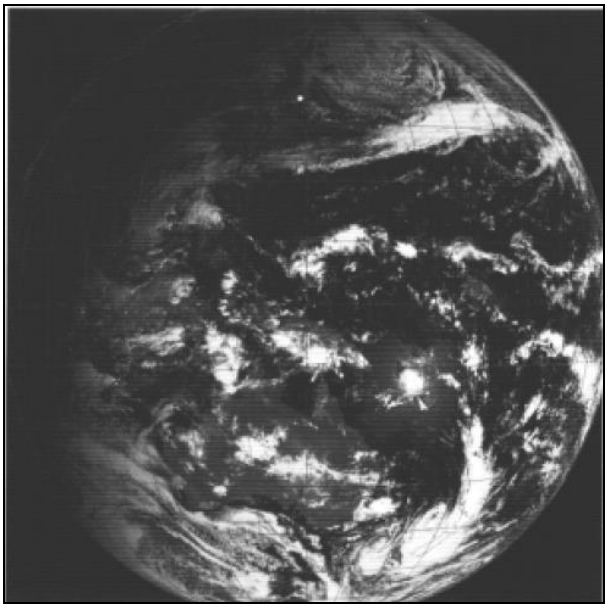


Fig. 4.5 : Visible image at 00UTC, 18 Dec 98

7. **Sea Surface Temperature Gradients**. In satellite images, changes in IR emissions between regions of the Earth's surface with different temperatures are most apparent at coastlines. Other temperature contrasts are often too diffuse for the human eye to pick out when images are presented in shades of grey representing broad bands of temperature. Enhancement of the image over the appropriate range of temperature coupled with the use of false colours, can improve discrimination.

Cloud Features Modified by Atmospheric Pollutants

8. **Dust and Haze**. Dust is characterized by a dull, hazy and filmy appearance similar to thin cirrostratus. Dust can extend to adjacent water bodies and sometimes obscure the coastline. The edges of dust areas are ill defined. Blowing dust and sand can be observed in satellite imageries when the reflectivity of the suspended particles differs greatly from the reflectivity of the underlying 'surface. Haze particles scatter quite effectively at the blue end of the visible wavelength and hence areas of haze show up better in images of this channel. Particles of dust or sand carried by the wind form a cloud, which causes land marks to appear blurred or to disappear in visible images. To be detected in IR imagery a dust cloud must be very deep. There are two reasons for the same. Firstly the dust must be composed of large enough particles (over 15 µm) to obscure radiation omitted from the ground at least partially. Secondly it must be deep enough to have low temperature otherwise it will not be able to distinguish it from the surface.

9. As we have seen from the previous paragraph, satellite imagery plays a crucial role in locating dust. This paragraph explores the capability of dust detection for different sensors aboard different satellites. It also looks at new dust enhancement products that help forecasters see blowing dust on images in the succeeding paragraphs. The AVHRR images shown in Fig. 4.6 and Fig. 4.7 show important properties of blowing dust. First, examine the visible and long wave IR window images over central Asia. There are two areas of blowing dust in this scene, one over water and the other over land. Can you identify them? You'll probably be able to spot the regions of thick dust, but there are surrounding regions of thinner dust that are harder to identify.

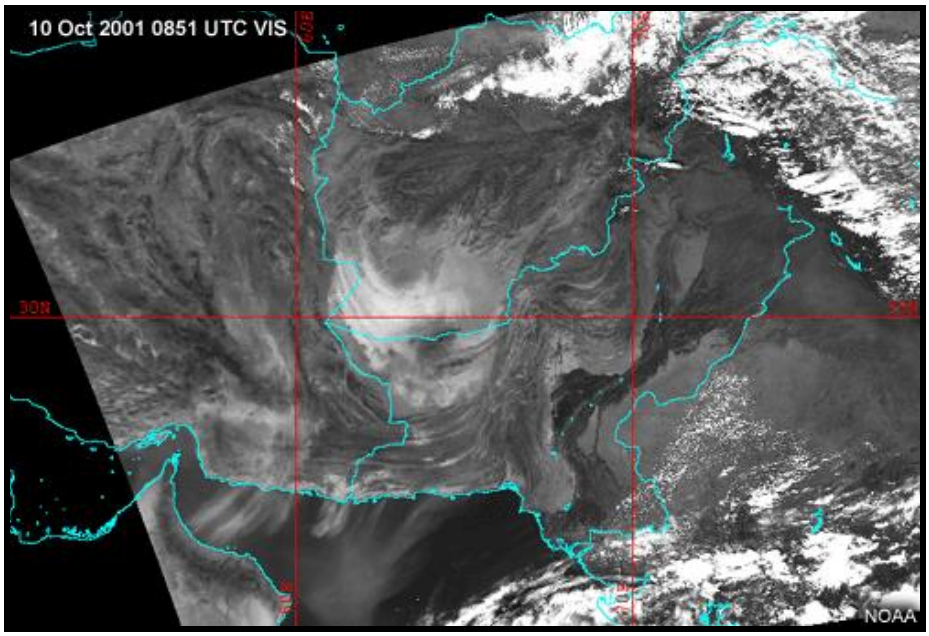


Fig. 4.6: Visible image at 0851UTC/10 Oct 01

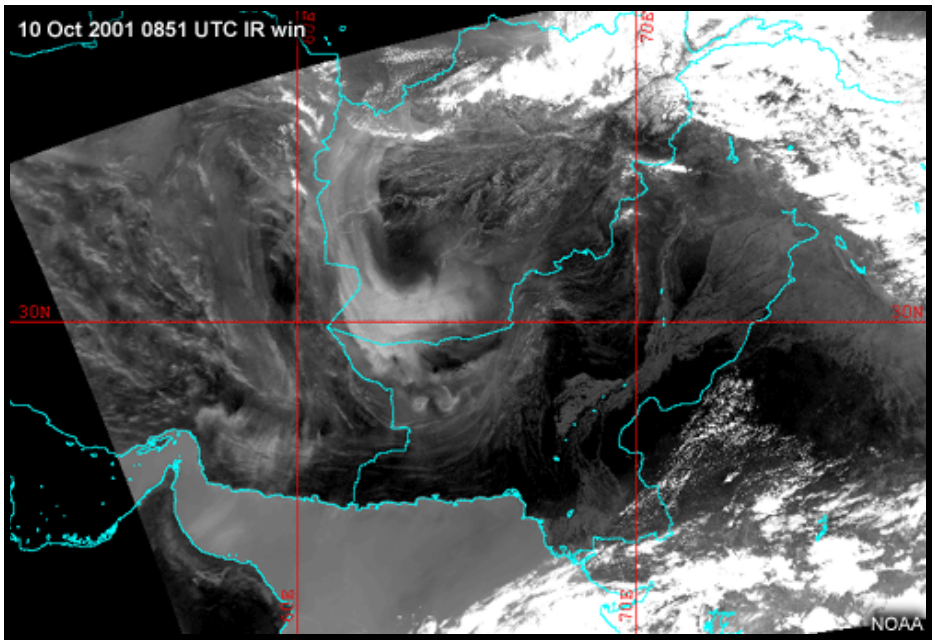


Fig. 4.7: IR image at 0851UTC/10 Oct 01

TASK-23 : INTERPRETING SUBSYNOPTIC AND SYNOPTIC SCALE CLOUDS

Cloud Features

1. Synoptic scale is defined as the scale of the migratory high- and low-pressure systems of the lower atmosphere, with wavelengths of 1,000 to 2,500 km. Cloud patterns that relate to the synoptic and sub synoptic scale are associated with fronts, high- and low-pressure systems, jet streams, etc.

Easterly (Tropical) Waves

2. In the tropics, spiraling (vertical) cloud patterns occur in the formative, mature, and dissipative stages of tropical cyclones, while weak tropical waves are often identifiable by curved cloud lines, spiraling cloud bands, and/or inverted-V- shaped cloud patterns that form east of these waves. Figure 6.1 shows the inverted-V-shaped pattern. This figure also shows another distinctive and characteristic feature of many weak, non-developing waves. The feature is known as the "screaming eagle" cloud pattern, and it can be seen above the second wave of the inverted-V-shaped cloud pattern. The broad wing-like pattern is also shown in figure 6.2

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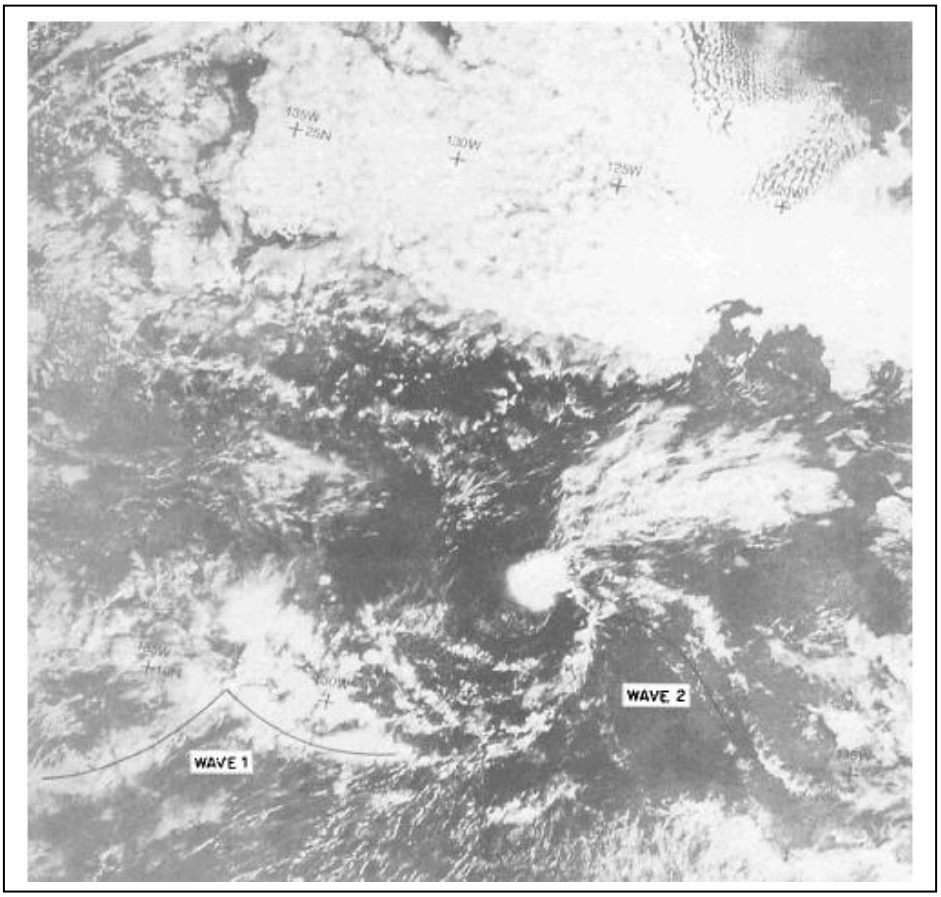


Fig 6.1 Inverted V-shaped pattern of tropical wave

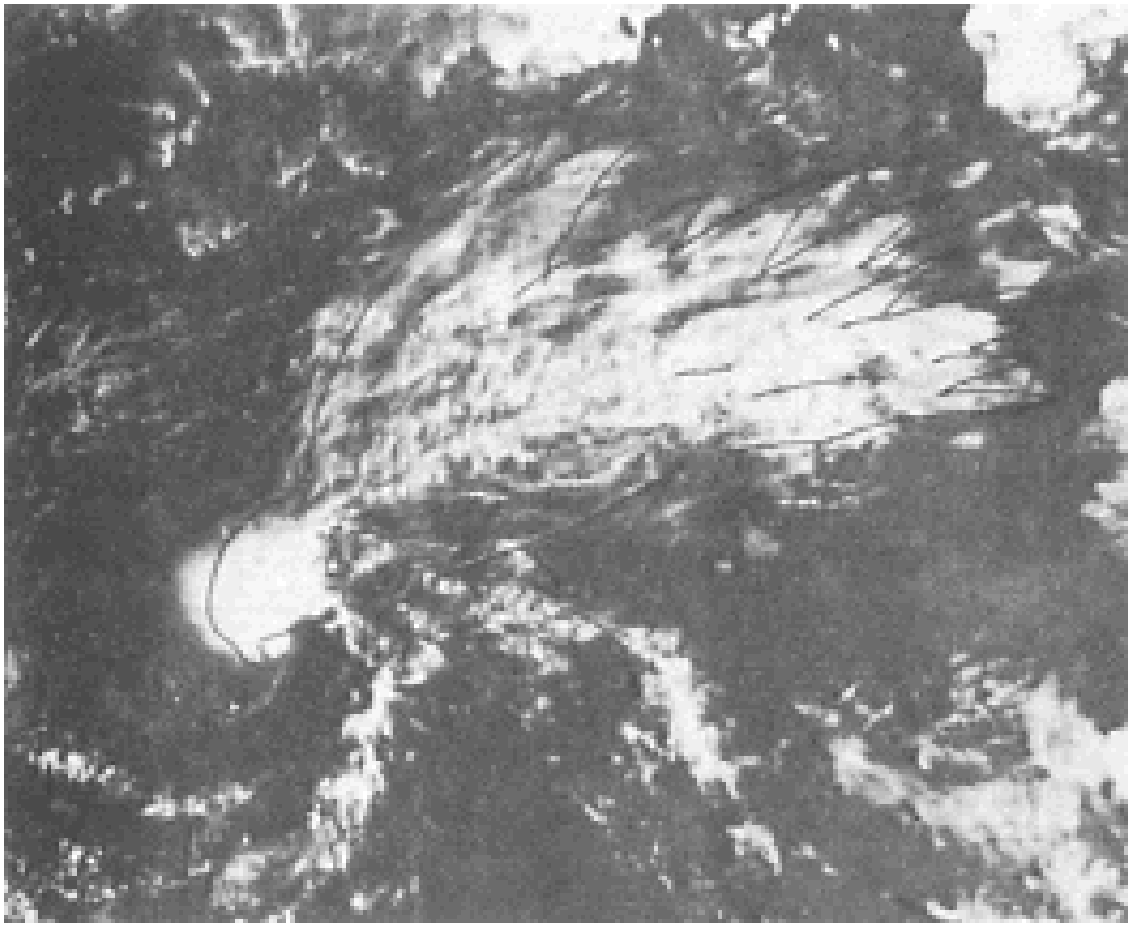


Figure 6.2. Detail of screaming eagle cloud pattern. 1603 GMT 03 October 1971

Table 6.1 Appearnce of Cloud in Satellite Picture

CLOUD TYPE	SIZE	SHAPE (ORGANIZATION)	SHADOW	TONE (BRIGHTNESS)	TEXTURE
CIRRUS CIRROSTRATUS	TYPICAL LENGTHS OF ORGANIZED BANDS HUNDREDS OF MILES, WIDTHS OF SINGLE BANDS MAY BE 25 - 50 MILES; EXTENSIVE LAYERS OF CIRROSTRATUS MAY ALSO COVER LARGE AREAS.	LONG BANDS, PARALLEL TO UPPER TROPOSPHERCIC WINDS, OFTEN HAVING SHARPLY DEFINED LEFT BOUNDARY, RELATIVE TO AN OBSERVER FACING DOWNWIND; RIGHT BOUNDARY SOMETIMES WELL-DEFINED, BUT IS MORE FREQUENTLY INDISTINGUISHABLE WHEN IT APPEARS OVER A MIDDLE CLOUD LAYER.	NORMALLY PRESENT AS A DARK LINE ALONG ONE EDGE; MOST NOTICEABLE WHEN SHADOW IS CAST ON A LOWER CLOUD LAYER OR A SMOOTH SURFACE WITH HIGH REFLECTIVITY.	TYPICALLY LIGHT GREY, BUT TONE IS DEPENDENT ON SUN ANGLE; TRANSLUCENT, LOWER CLOUDS AND GEOGRAPHICAL FEATURES ARE USUALLY ONLY PARTLY OBSCURED BY CIRRIFORM CLOUD.	CIRROSTRATUS NORMALLY HAS A UNIFORM TEXTURE, WHILE CIRRUS TENDS TO BE MORE FIBROUS. CLOUD BANDS PERPENDICULAR TO THE WIND INDICATE WAVE STRUCTURE.
ANVIL CIRRUS (DETACHED FROM CUMULONIMBUS	MAY BE QUITE EXTENSIVE, COVERING AREAS OF AROUND 500 HUNDRED MILES OR MORE IN LENGTH AND WIDTH.	CHAOTIC APPEARANCE WITH ALIGNMENT OF CLOUDS STREAKS PARALLEL TO UPPER TROPOSPHERIC WINDS, DIFFUSE OR POORLY - DEFINED EDGES.	ONLY DETECTABLE WHEN CLOUD LAYERS ARE SUFFICIENTLY THICK AND SHADOW FALLS ON AN ILLUMINATED LOWER CLOUD LAYER OR BRIGHTLY REFLECTIVE LAND OR WATER SURFACE.	LIGHT GREY OR WHITE, DEPENDING ON CLOUD THICKNESS AND SUN ANGLE.	FIBROUS WITH NUMEROUS STREAKS, OR MORE UNIFORM TEXTURE WHEN DENSE CIRRUS LAYERS ARE CONCENTRATED WITHIN A SMALL AREA.
ALTOSTRATUS ALTOCUMULUS	EXTENSIVE SHEETS OR BANDS COVERING AREAS TENS TO HUNDREDS OF THOUSANDS OF SQUARE MILES; BANDS MAY BE TWO OR THREE HUNDRED MILES ACROSS.	ORGANIZED INTO VORTICES, BANDS, LINES OR LARGE COMMA-SHAPED AREAS ASSOCIATED WITH CYCLONES AND FRONTS CHARACTERIZED BY PERSISTENCE IN FORM OVER PERIODS OF 12-24 HOURS OR MORE; SINCE CLOUD IS ASSOCIATED WITH SYNOPTIC SCALE MOTION SYSTEMS; USUALLY WELL-DEFINED BOUNDARIES.	OFTEN PRESENT ALONG ONE EDGE; SHADOW ENHANCED IF IT APPEARS ON A LAYER OF LOWER CLOUD.	VERY WHITE; ONE OF THE TWO BRIGHTEST CLOUD FORMS, THE OTHER BEING CUMULONIMBUS-DUE TO GREAT VERTICAL DEPTH OF CLOUD WHITEST CLOUD LAYERS ARE OFTEN ASSOCIATED WITH NIMBOSTRATUS AND PRECIPITATION AT THE GROUND.	STRATIFORM CLOUD WITH UNIFORM TOP SURFACE HAS UNIFORM TEXTURE. IF CONVECTIVE CLOUDS ARE PRESENT OR THE MIDDLE CLOUD IS NOT SOLID, VARIATIONS IN TEXTURE WILL APPEAR AS A RESULT OF SHADOWS, BREAKS, OR THICKNESS VARIATIONS.
WAVE CLOUDS CIRRUS ALTOCUMULUS STRATOCUMULUS	NARROW PARALLEL BANDS OF THE ORDER OF TEN TO A HUNDRED MILES IN LENGTH; UNIFORM SPACING OF CLOUD IS CHARACTERISTIC OF THESE CLOUD FORMS.	UNIFORMLY SPACED; PARALLEL BANDS, MORE OR LESS PERPENDICULAR TO THE WIND DIRECTION AT CLOUD LEVEL. MOST OFTEN FOUND LEE OF HILLS AND MOUNTAINS. NOTABLE EXAMPLES APPEAR OVER THE ROCKIES, APPALACHIANS, LABRADOR, AND OTHER RANGES.	NOT USUALLY DISCERNIBLE	GREY, OCCASIONALLY WHITE, DEPENDING ON SUN ANGLE AND VERTICAL THICKNESS OF CLOUD.	CONTINUOUS OR BROKEN PARALLEL BANDS; MAY BE VERMICULATED.
CUMULUS TOWERING CUMULUS	INDIVIDUAL CUMULUS CLOUD CELLS ARE NORMALLY TOO SMALL TO BE DISCERNIBLE AT 800 MILES. RATHER WHAT APPEARS SIMILAR TO INDIVIDUAL CUMULI AS SEEN FROM THE GROUND ARE GROUPS OF CLOUDS HAVING A REGULAR ORGANIZATION OR PATTERN NOT NORMALLY DETECTABLE FROM SURFACE OBSERVATIONS: DIMENSIONS OF CLOUD GROUPS 3 TO 10 MILES.	WITH LIGHT WINDS, CLOUD GROUPS PRESENT A UNIFORM CELLULAR PATTERN OR MAY BE ORGANIZED IN SINGLE OR PARALLEL STREETS, STRAIGHT OR GENTLY CURVED, GENERALLY PARALLEL TO THE WINDS. OCCASIONALLY, HOLLOW POLYGONAL CELLS, CRESCENTS OR SOLID CELLS WILL APPEAR IN THE OVERALL PATTERN; USUALLY LUMPY APPEARANCE.	USUALLY PRESENT WITH TOWERING CUMULUS; DISTINCTIVE SHADOWS ON DOWN SUN SIDE; SHADOWS NOT SO EVIDENT WITH SMALLER CLOUDS OR CLOUD GROUPS.	BROKEN DARK GREY, GREY OR WHITE, DEPENDING ON DIMENSIONS AND THICKNESS OF CLOUD GROUPS AS SEEN FROM SATELLITE ALTITUDES; SMALLER CLOUD GROUPS ARE DARKER IN TONE WHILE CUMULUS CELLS SMALLER THAN THE THRESHOLD RESOLUTION OF THE CAMERA (2 MILES) WILL NOT BE VISIBLE IF SEPARATION IS GREATER THAN 2 MILES; AREAS OF SMALL CUMULUS WILL APPEAR IN BROKEN GREY TONE.	NON-UNIFORM, ALTERNATING PATTERN OF WHITE, GREY AND DARK GREY OFTEN HAVING GREAT REGULARITY DUE TO CONTAINED SHADOWS; HOLLOW CENTERS MAY BE PRESENT IN A RING OF CELLS.
CUMULONIMBUS	INDIVIDUAL ISOLATED CUMULONIMBUS CLOUDS ARE THE ORDER OF TENS OF MILES IN DIAMETER; COMBINED CLUSTERS OF SUCH CLOUDS MAY PRESENT A PATTERN AS LARGE AS A HUNDRED MILES ACROSS DUE TO MERGING OF CIRRUS ANVILS.	ISOLATED CELLS HAVE SHARPLY DEFINED EDGES ON ONE SIDE WITH CIRRUS ANVIL SPREADING OUT ON THE OPPOSITE SIDE IN THE PRESENCE OF PRONOUNCED WIND SHEAR; OTHERWISE THEY APPEAR AS ISOLATED, WHITE, NEARLY CIRCULAR CELLS.	SHADOWS USUALLY PRESENT AND WELL-DEFINED WITH CUMULONIMBUS	VERY WHITE, PARTICULARLY TOPS HAVE CHARACTERISTIC BRIGHTNESS.	UNIFORM TEXTURE, SHARPLY DEFINED EDGES, ALTHOUGH CIRRUS PLUMES ARE OFTEN QUITE DIFFUSE BEYOND MAIN CELLS.
STRATOCUMULUS	APPARENT SIZE OF CELLS 2-10 MILES ALTHOUGH LAYERS WILL HAVE NO DISTINCTIVE SIZE.	STREETS OR BANDS ALIGNED WITH THE BOUNDARY LAYER WINDS OR EXTENSIVE AREAS WITH WELL-DEFINED BOUNDARIES.	SHADOWS MAY SHOW STRIATIONS ALONG THE WIND.	SMALL CLOUD GROUPS ARE MOSTLY GREY OVER LAND. THICK OVERCAST STRATOCUMULUS LAYERS OVER OCEANS OFTEN APPEAR WHITE DUE TO CONTRAST IN REFLECTIVITY.	OVERCAST STRATOCUMULUS CLOUD LAYERS OFTEN SHOW HOLLOWS WITH DIFFUSE CENTERS.
STRATUS	VARIABLE	VARIABLE, EXCEPT WHEN STRATUS CLOUD IS LOWER THAN SURROUNDING TERRAIN, IN WHICH CASE IT ASSUMES THE SHAPE OF A VALLEY, MOUNTAIN, OR COAST LINE, ETC BOUNDARY WELL-DEFINED BUT MAY HAVE A RAGGED EDGE.	NORMALLY NOT DISCERNIBLE, BUT PRESENCE IS DEPENDENT ON HEIGHT OF STRATUS LAYER ABOVE GROUND.	WHITE OR GREY, DEPENDING ON VERTICAL CLOUD THICKNESS AND SUN ANGLE.	UNIFORM
FOG	VARIABLE	VARIABLE, IRREGULAR, BUT IN THE CASE OF FOG OVER BODIES OF WATER, SHAPE CONFORMS TO THAT OF SURROUNDING LAND; BOUNDARIES SHARPLY-DEFINED AND MAY BE THE ONLY DISTINGUISHING CHARACTERISTIC FROM STRATUS.	NORMALLY NOT DISCERNIBLE.	WHITE OR GREY, DEPENDING ON THICKNESS OF FOG LAYER AND SUN ANGLE, NORMALLY IF DEPTH OF FOG LAYER EXCEEDS 1000 FEET IT APPEARS WHITE.	VERY UNIFORM

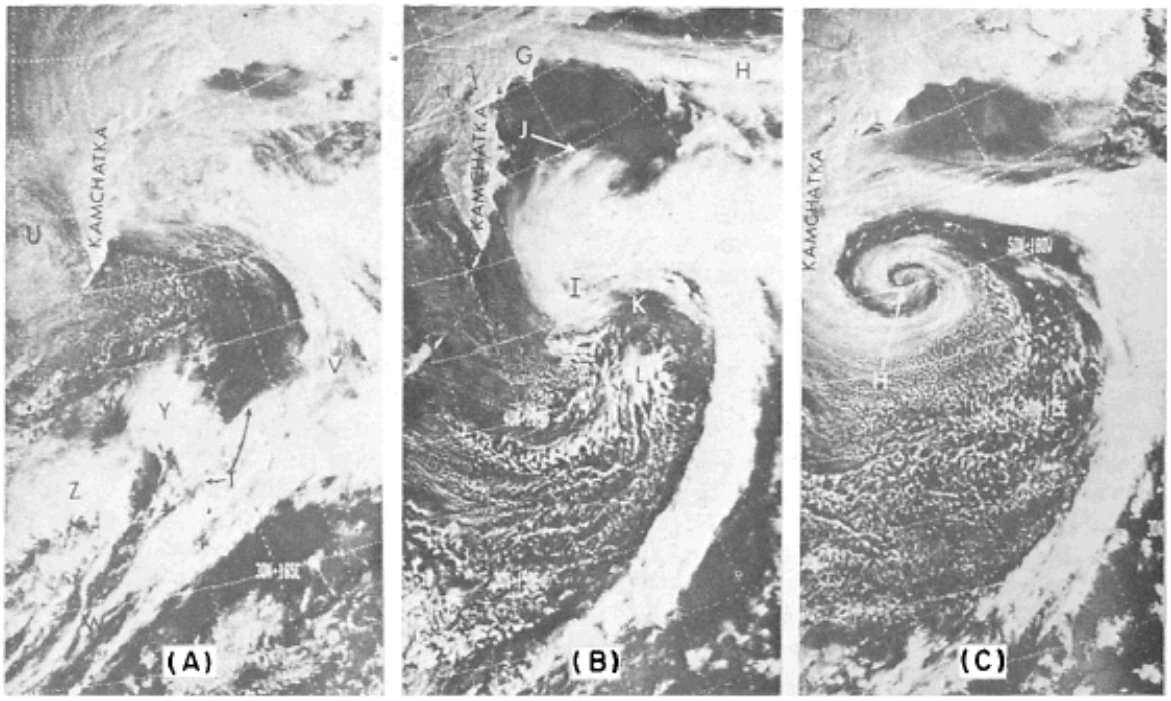


Fig 6.5. Development of a Pacific storm; (A) visible. NOAA-2 SR. orbit 6426.2214 GMT. March 12: (B) orbit 6439. 2307 GMT, March 13; (C) orbit 6451, 2209 GMT, March 14, 1974

Tropical Cyclones

3. Satellite pictures provide the earliest warning of tropical cyclone existence and permit a highly accurate tracking capability. Initially, a tropical cyclone is simply a dense mass of cumuliform, cirriform, layered middle clouds without a recognisable shape or pattern. It's the size and location of these cloud masses that make them suspect to tropical cyclone classification.

4. A suspect cloud mass has an average diameter of 3 degrees latitude or more except when the cloud, mass is within or adjacent to the ITCZ of the Atlantic, Pacific, or South Indian oceans. Here, the diameter must average at least 6 degrees latitude. In the Arabian Sea and Bay of Bengal, suspect cloud masses must attain a diameter of 8 degrees latitude or more. All satellite pictures and any surface and upper-air observations from an area of suspicion are studied to determine whether tropical cyclogenesis is taking place.

5. Recognisable cloud patterns evolve as tropical cyclones develop, intensify, and weaken. In the early stages, the clouds take on the shape of curved bands or lines. Cloud bands are usually seen 36 hours before a disturbance reaches tropical storm force. During this pre-storm period, the cloud pattern is highly variable. Frequent, short period, mainly diurnal changes occur, which can cause a disturbance to appear quite intense during the day and much weaker at night. As a tropical cyclone reaches tropical-storm force, the cloud bands become more organised. At the weak tropical-storm stage, the band(s) curve halfway around the centre of the system. As the storm intensifies, the banding completely surrounds the centre, forming an eye, at the weak hurricane/typhoon stage. Further intensification is indicated by increasing clouds around the eye, or by the eye becoming better defined.

6. **Tropical Cyclone Intensity Analysis**. For over 25 years meteorologists have studied satellite imagery of tropical cyclones. An outgrowth of studies correlating cloud patterns with tropical cyclone intensity levels was the development of an intensity analysis technique. Today, fairly accurate estimates are possible based on cloud patterns seen in imagery. The technique currently in use was developed by Mr. Vernon Dvorak. The Dvorak technique is based on cloud feature measurements, models, and cloud pattern changes. The technique is applicable to all types of imagery (visible, IR, and enhanced IR [EIR]). The procedures and rules for each type of imagery are included in their entirety in appendix at the end of the chapter. The following terminology is associated with satellite cloud pattern analysis and the Dvorak technique.

7. **Banding features (BF)**. The amount of coiling (banding) in the cloud system.

8. **Central Cold Cover (CCC)**. A circular, cold overcast near the storm centre that develops as the curved band cloud pattern dissipates. ‘This feature is only distinguished using enhanced IR imagery. It can occur at any stage of storm development and may last for several hours or several days. When it persists, it signifies the end of tropical cyclone development.

9. **Central Dense Overcast (CDO)**. The dense overcast mass of clouds covering the most tightly curved cloud bands. It may cover the lower, curved cloud features or surround the eye of a well-developed tropical cyclone. This pattern pertains to visible imagery only.

10. **Central Features (CF)**. The character of the cloud mass associated with the cloud system centre and the overcast around the centre. A central dense overcast and eye are examples of central features.

11. **Cloud Features**. Those aspects of the cloud system used in intensity analysis; for example, central features, central dense overcast, banding, etc.

12. **Cloud System Centre (CSC)**. The focal point of all curved cloud lines or bands of the tropical cloud system, or put another way, the point toward which all cloud lines or bands merge. T-number (tropical numbers). The classification system used to denote the intensity of tropical cyclones. A T1 is representative of minimal tropical cyclone intensity, while a T8 represents maximum possible intensity.

13. **Procedures and Rules**. The analysis procedure consists of two major parts. In part 1, intensity estimates are obtained by measuring cloud features seen in imagery. In part 2 the cloud patterns seen in imagery are compared to model cloud patterns. Part 2 is only necessary when cloud measurements aren’t possible and/or to limit the intensity level (T-number) obtained from the measurements made in Part 1. Analysis procedures are summarised in figure 6.6.

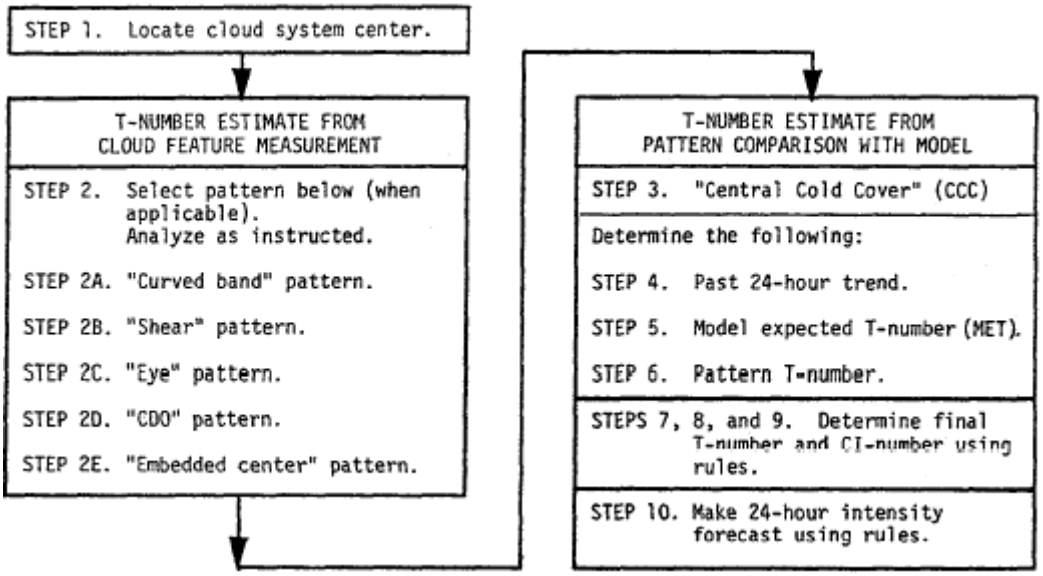


Fig 6.6. Procedures for T-number determination

14. The procedure consists of 10 steps in which you determine the intensity of a tropical cyclone by first locating the CSC (step 1) and then analysing the centre's appearance and its relationship to the dense (cold) clouds of the pattern. The CSC is located at the centre of curvature of a partial wall cloud in developing cyclones, and at the centre of the eye in mature cyclones. If the CSC is not obvious, a model cloud pattern is selected that compares to the one seen in imagery, and the model's expected CSC is used.

15. The first intensity estimate is made in step 2 by measuring the cloud features related to intensity. This is done when the cloud pattern contains cloud features similar to those in the cloud pattern descriptions listed in steps 2A to 2E. Since these patterns are seen in IR imagery, as well as visible imagery, you should know the importance of IR pictures in tropical cyclone analysis and the major differences with regard to intensity analysis. Infrared pictures permit tropical cyclone monitoring day and night. In fact, the day-by-day changes in the cloud patterns are often followed in IR pictures with much the same quality as they are in visible pictures. Visible and IR images of tropical cyclones are very similar; however, there are two major differences. First, thin cirrus clouds, which are normally transparent in visible imagery, are often opaque in IR, thereby obscuring the view of lower cloud features. When the cirrus obscures the boundaries of lower cloud features, no intensity measurements are possible or erroneous ones are taken. This means the CDO size and embedded eye distance measurements cannot be interpreted in unenhanced IR pictures.

16. Second, the low cloud lines used to pinpoint the storm centre may not be visible because of the cirrus. With these differences in mind, let us now look at the cloud band pattern types seen in imagery and the type of imagery used to obtain the intensity estimate. Refer to figure 6.7.

17. **The Curved Band Pattern (Visible)**. Intensity estimates obtained from the curved band pattern seen in visible imagery are based on the extent to which the dense overcast cloud band encircles the CSC. At the minimal tropical storm stage, the band is observed to curve about halfway around the centre. When the band coils completely around the centre, the hurricane stage is attained provided that the required minimum length of developmental time has elapsed. The reason for this is that the coiling occasionally occurs

more rapidly than pressure falls are known to occur. Continued strengthening of a hurricane/typhoon results in continued coiling of the curved band or in the formation of a centre or eye embedded in the dense overcast that appears central to the band curvature. When an eye is observed, the intensity determination is based on the eye characteristics, the amount of dense overcast surrounding it, and the amount of outer banding surrounding these central features. The curved band pattern is the most common cloud pattern seen in satellite pictures, and figure 6.8 shows a model of this pattern used in intensity analysis.

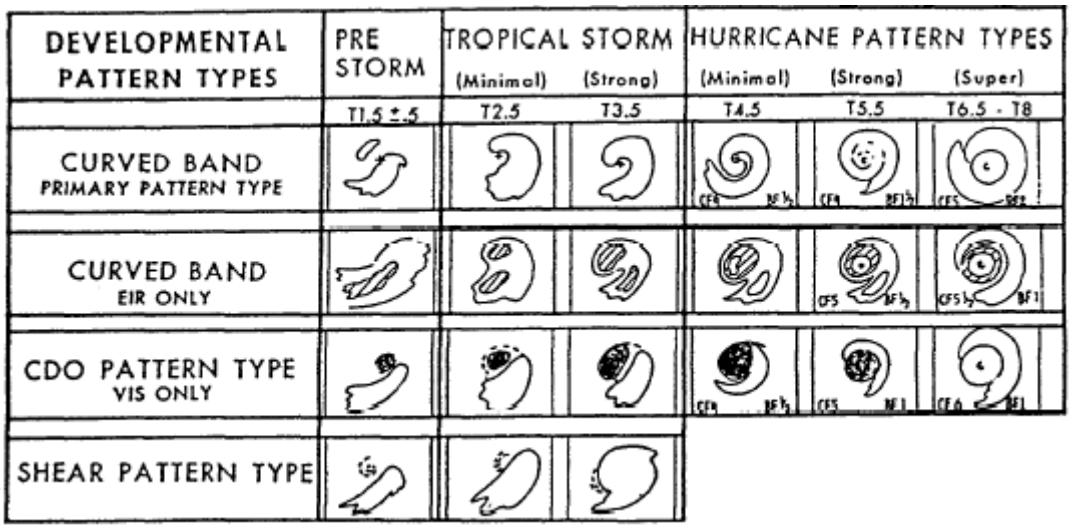


Fig 6.7. Development cloud pattern types used in intensity analysis. Pattern changes from left to right are typical 24-hour changes

18. **<u>Curved Band Pattern (EIR)</u>**. The curved band pattern as seen in EIR imagery is similar to that seen in visible imagery, especially during the early stages of tropical cyclone development. It is during the weak hurricane stage, when the curved band extends once around the centre, that the EIR imagery proves more objective than visible imagery in intensity analysis. It is simpler and more objective and consists primarily of two simple measurements: First, the temperature of the coldest cloud band that completely surrounds the eye, and second, the temperature of the eye itself. Based on these two temperatures, an intensity estimate is made.

19. **<u>Central Dense Overcast (CDO) Pattern</u>**. The CDO pattern is used to determine intensity levels when a dense overcast cloud mass appears over the curved cloud features that define the centre or surrounds the eye. When the CDO pattern is observed and no eye is present, the size of the dense cloud mass relates to the intensity level. The CDO size increases with increasing intensity. An eye usually becomes visible with the CDO before the T5 intensification level is reached. When the CDO contains an eye, the distance the eye is embedded within the CDO determines the intensity estimate.

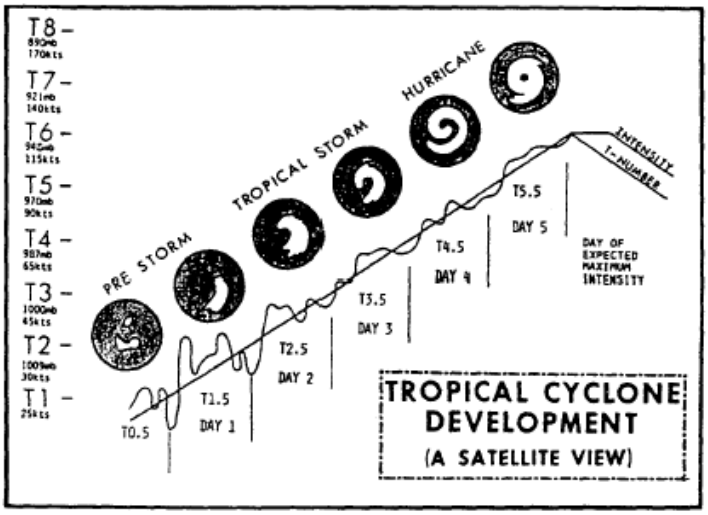


Fig 6.8. Model of tropical cyclone development used in intensity analysis (curved band pattern type)

20. **Shear Pattern**. Vertical wind shear may prevent the dense, upper-level clouds of a tropical cyclone from coiling around the system centre as they do in the curved band patterns. When a shear pattern is observed, it is the curvature of the low cloud lines and their proximity to dense overcast clouds that determine the intensity level. The low cloud centre appears off to the side of the dense overcast. When the low cloud lines are circularly curved, parallel, and near the edge of the dense overcast, minimal tropical storm intensity is indicated. The second intensity estimate is determined in steps 4 through 6 (fig. 6.8). Step 3 will not be discussed at this time because it deals with a cloud pattern type that indicates a tropical cyclone's development has been, or soon will be arrested. Steps 4 and 5 set limits within which measured estimates must fall, and also provide a reasonable intensity estimate when measurements of cloud features are not possible. This estimate, called the "model expected T-number" (MET), is determined by comparing satellite pictures (today's *vs* yesterday's) and deciding whether or not the cyclone has continued on its past trend of development. Using only this simple decision, you can obtain the intensity estimate by extrapolation along the intensity change curve provided in the model that best fits the past history of the cyclone's development. For example, in the curved band pattern type, you are only required to determine whether or not the band has curved farther around the storm centre from one day to the next. Step 6 refines the step 5 estimate of intensity. It is made by comparing the cloud pattern seen in imagery to patterns in the model that correspond to the stage of development indicated in step 5. When the cloud pattern being analysed appears to be obviously stronger or weaker than is expected from its developmental rate, the intensity estimate is adjusted up or down accordingly. This estimate is used whenever the cloud features relating to intensity are distinguishable but not clear-cut enough for measurement.

21. The intensity estimate determined from the cloud features is then examined according to the rules of the technique to see if it falls within specified limits or if it must be adjusted (steps 7 through 9). The rules, in general, hold the change in intensity close to one T-number a day for the pre-storm stage of development and to within one number of the model expected T-number during the later stages of development.

22. The final step in the technique (step 10) provides instructions for making a 24-hour intensity forecast.

23. Step 3, which we passed over previously, is used when the cloud pattern exhibits a central cold cover. When the CCC is observed, the analysis consists of a simple application of the rules given in appendix.

Upper-Air Features seen in Imagery

24. The cloud patterns discussed in this section are associated with upper-level ridges, troughs, and jet streams.

25. **Upper-Level Ridges.** The distribution of middle and high clouds is the key in locating and positioning upper-level ridges. A sharp upper ridge has a narrow cloud band associated with it, and the clouds end abruptly at the ridgeline. See figure 6.5. The cloud pattern associated with medium amplitude ridgelines is broader, and the ridgeline ranges in position from the leading edge of the cloud deck to a few degrees back into it. When the cirrus deck extends beyond the ridge, the ridgeline closely approximates the leading edge of the middle cloud deck. If there are no cirrus clouds present, the ridgeline is positioned at or slightly behind the leading edge of middle clouds. Flat, low amplitude ridges have their axes well back into the cloud pattern and are often related to anticyclonically curved striations. The axis lies very close to the axis of the curved striations seen in the pattern.

26. **Upper-Level Troughs.** Changes in the appearance of frontal bands denote the position of upper-level troughs. A 500-hPa trough is positioned along a line extending from the centre of the upper-level vortex southward to the point on the frontal band where the clouds become ragged and discontinuous in appearance. See figure 6.5 a comma-shaped cloud area (PVA MAX) is present equatorward of the vortex, the trough is drawn to the west of the comma cloud.

TASK-24 : JET STREAM AS SEEN IN SATELLITE PICTURES

1. Satellite pictures shoe extensive cloud formations associated with jet stream, which are probably one of the longest cloud systems seen in satellite pictures. These cloud formations have a characteristic appearance and thus help to reveal the presence of a jet stream in the neighbourhood; some of them may also help to locate the position of the net axis. Four types of jet- associated cloud organisations have been noticed in satellite pictures. They are:

a. Long shadow lines.

b. Sharp edged large Ci-sheet.

c. Longitudinal Ci bands or streams.

d. Transverse bands in Ci formations.

2. The long shadow lines and sharp-edged large Ci-sheets are usually associated with PFJ, The jet axis could be placed parallel to the shadow line or the northern boundary of the ci sheet is either coincident with or slightly to the north of the jet axis. The northern boundary of the cloud sheet is sharp, long and usually anticyclonically curved. The jet cirrus is most often observed where the winds follow a broad anticyclonically turning path; where the jet stream turns cyclonically, the Ci is rarely detectable.

3. Over India and Bangladesh, longitudinal or transverse Ci bands are the types generally let with rather than the shadow lines or the large Ci shields. Corresponding land-based observations have shown that these jet stream clouds are mostly cirri form clouds with a fair amount of alto clouds (mainly Ac) mixed up. Although the STJ over India is a quasi-permanent feature of the daily charts over the country during the winter, still the Ci bands do not form every day. They appear only in association with large amplitude troughs, when certain amount of advection of moist air takes place at high levels over the country, resulting in the formation of clouds. The clouds form only ahead of the trough line.

4. In the case of longitudinal bands or streaks, they are oriented closely parallel to the wind flow or at small crossing angles. The jet stream axis is located just to the north of these streaks, nearly parallel to the streaks. In the case of the transverse bands, the transverse elements are nearly perpendicular to the flow at the jet level. The axis of the jet stream lies just to the north of cloud configuration.

5. Transverse band type cloud configuration is associated with strong jet streams; turbulence is also more with a strong jet stream and it is believed that as a consequence, these transverse clouds develop. It has been found that larger vertical wind shear and smaller lapse rates are associated with longitudinal bands.

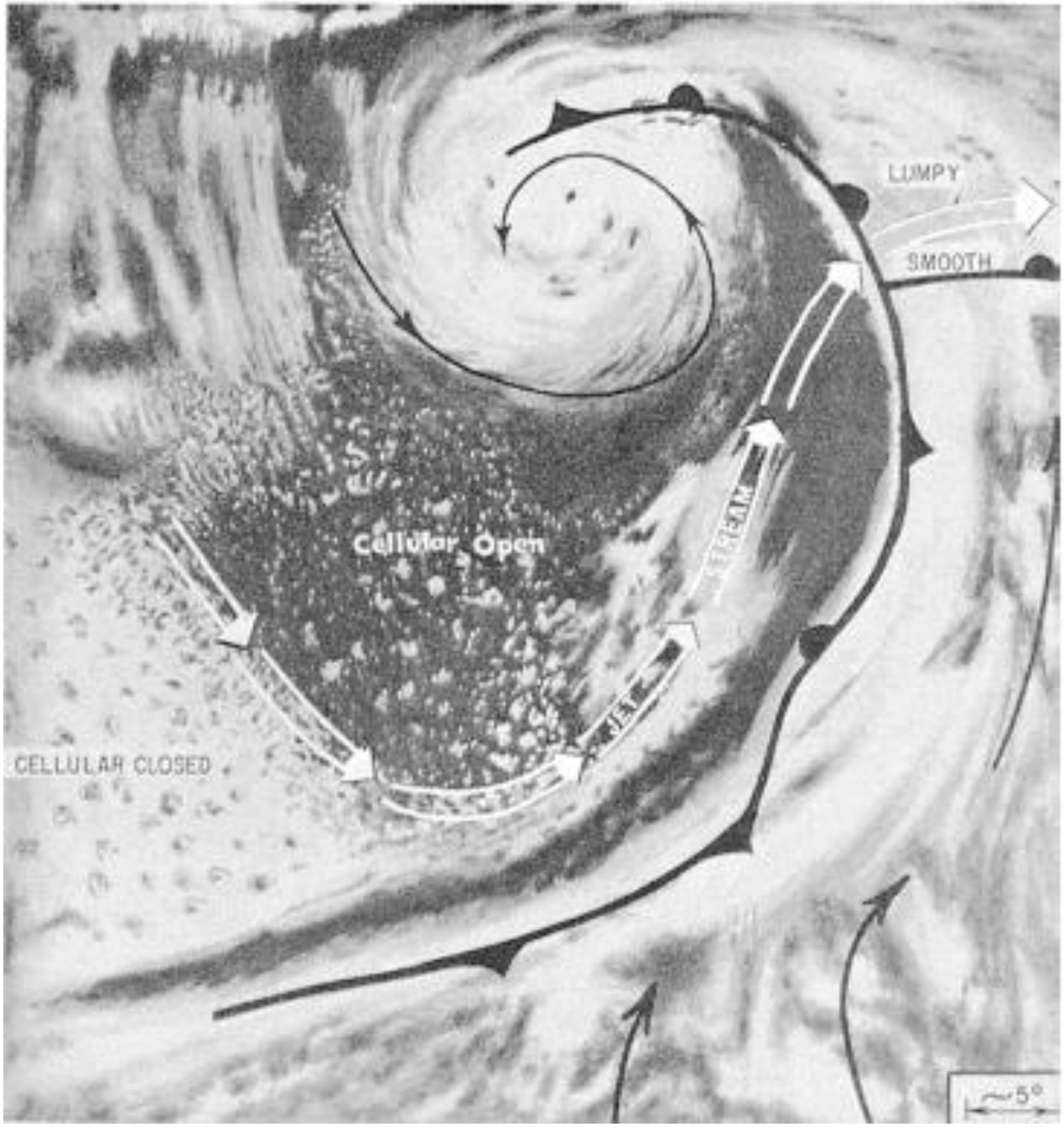


Fig 3.1. The jet stream east of the upper air trough is shown crossing over an occluded front

SECTION-6 : MONSOON CIRCULATION

TASK-25 : MORPHOLOGY OF MONSOON CIRCULATION

Introduction

1. The term 'monsoon' is derived from the Arabic word ‘mausim’, which means season. Since early times, the term monsoon has been used to signify any annual climate cycle with a dominant seasonal wind reversal. Over vast oceanic areas, the tropical circulation undergoes very little seasonal variation, except for some minor latitudinal shifts and intensity changes. On the contrary, the circulation varies substantially and displays a distinct seasonal rhythm over the tropical continents. Over the oceans, evaporation consumes a high proportion of the incident radiation, which is also rapidly absorbed by the water surface and spread over greater depths by waves and turbulence, and dissipated to other latitudes by ocean currents. But over the continents the incoming radiation is mostly used to warm the earth's surface, so that temperatures in these regions reach very high levels owing to the much lower heat capacity of the soil compared to water. This warming over the tropical continents in early summer produces thermal low-pressure areas which gradually take on some functions of the equatorial trough, thus forming a new region of convergence. The trade winds from the winter hemisphere cross the equator, slow down and create a minor area of convergence owing to their change in direction caused by the reversal of the Coriolis force. Subsequently, the air accelerates towards the low-pressure area over the continents. In winter, the continents experience relatively low temperatures and high surface pressure; winds are re-established towards the equator which are similar to the trade winds, though they are not stable in speed and direction.

Monsoon Areas

2. Thus the tropical continents and adjacent oceans experience a semi-annual reversal in wind direction characteristic of monsoons. Ramage (1971) formulated four simultaneous criteria to delineate the monsoon areas:

a. The prevailing wind direction shifts by at least 120 deg between January and July.

b. The average frequency of prevailing wind direction in January and July exceeds 40%.

c. The mean resultant wind in at least one of the months exceeds 3 metre per sec.

d. Fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5 deg latitude-longitude rectangle.

3. Based on these criteria, Ramage delineated the monsoon region of the world as shown in fig 7.1, which is widely adopted. The monsoon region includes almost half of the African continent, South and East Asia and northern Australia. These areas are inhabited by more than half of the world's population for whom the monsoon rainfall is essential to their survival and prosperity.

4. To include in the definition of “monsoon region”, the areas where the surface position of ITCZ has a relatively small annual oscillation in positions like Eastern Pacific, Central and South America and Western Atlantic, Asnani (1993) proposed the following definition:-

"It is the region which lies between the latitude 5 degree north of the northern most (July) surface position of ITCZ, and the latitude 5 degree south of the southern most (January) surface position of ITCZ, on climatological basis."

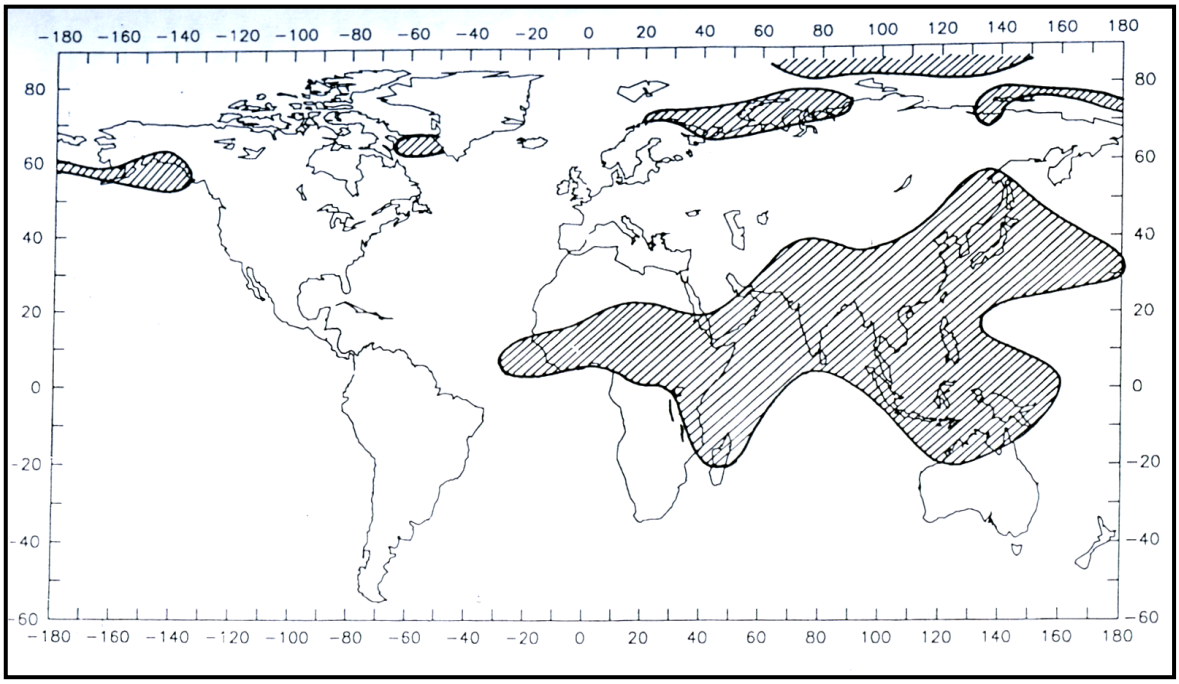


Fig 7.1 The monsoon regions of the world (After Ramage, 1971)

5. **<u>The Asian Summer Monsoon</u>**. Nowhere else on the globe is the annual reversal of wind and rainfall regimes as spectacular as in the North Indian Ocean and the adjoining land areas in summer. Asia is the largest continent on the earth and its tropical zone has the right conditions for the circulation to develop with predominant winds of one direction in winter and of the opposite direction in summer. The monsoon circulation is most pronounced over a wide area, from Pakistan in the west to Japan in the east. It is now a well-established fact that the Himalayan mountain range plays a crucial role in the genesis and development of the Asian monsoons.

6. During the summer season, at lower levels in the atmosphere, the seasonal wind reversal and the associated south-to-north migration of the ITCZ is modified by the land ocean contrasts and orographic barriers. These features accentuate a vigorous monsoon activity over South Asia practically all through the summer. The equatorial trough migrates to its northernmost position (30 deg N) over South Asia during the northern hemispheric summer. This migration results in a pressure gradient from the equatorial region to the region of maximum heating and lowest pressure, which is associated with a quasi-geostrophic westerly current deflected by surface friction towards the low-pressure area in the subtropics. Thus, while the Asian summer monsoon is a consequence of the thermal differences between the land and the sea in general terms, it is primarily due to the seasonal shifting of thermally produced planetary belts of pressure and winds under continental influences.

Monsoon Circulation

7. The south Asian Monsoon circulation is an important component of the global general circulation. It is by now generally recognised that the monsoons arise due to the land-ocean temperature contrasts during summer and winter. Global temperature and pressure fields for both summer and winter at the surface and at 400 hPa level reveal the following important points, especially regarding over northern hemisphere:-

a. Continents are hotter in summer and colder in winter compared to the oceans.

b. At the surface lower pressures prevail over the warmer regions while higher pressures prevail over colder regions, relatively speaking. At upper levels the reverse is true.

c. Over the Afro-Asian land mass region the low and high pressure systems, at the surface during summer and winter respectively, are truly continental in scale.

d. The position of the surface temperature maxima or minima do not coincide exactly with the position of pressure minima and maxima respectively. However the reverse association between surface temperature maps and pressure maps is so striking that we can safely conjecture that the differential temperature field is the casual factor for the pressure field.

e. The temperature gradients are strong both in the north-south and in the east-west directions. Over the tropics the north-south temperature gradient reverses with seasons. Over the middle latitudes east-west temperature gradients show reversal with seasons.

8. The mean sea level pressure fields during January and July over Asia and neighbourhood (eastern hemisphere) show the lowest pressure occurs in July over northwest India-Pakistan. In January the highest pressure occurs over Siberia. Thus the annual cycle of the surface pressure has the largest amplitude over the Asian continents.

9. The wind field (streamlines and isotachs) at 850 hPa and 200 hPa during winter (December, January, February) and summer (June, July, August) show the strongest westerly jet stream is seen over Japan and neighbourhood in winter at 200 hPa. In summer we see the well marked anticyclone over Tibet called the Tibetan anticyclone and easterly jet stream to the south of it over south Asia and Africa, with the strongest winds over India. Thus we see that very large amplitude variations of wind in the upper troposphere occur over Asian continents.

10. The notable feature in the mean wind field over Indian region at 850 hPa and 700 hPa is the southwesterly monsoon winds over the Arabian sea, peninsular India and the Bay of Bengal. The shallow low pressure area lying over northwest India-Pakistan is called the heat low. The pressure trough of cyclonic shear zone extending from the heat low over NW India to the head Bay of Bengal at the surface and 850hPa is called the monsoon trough. The latitude-height section of the mean zonal winds averaged between 50°E and 100°E brings out the lower tropospheric westerlies (maximum 20 kts) and the upper tropospheric easterly jet stream (60 kts). A profile of zonal winds with height at Trivandrum for July-August (1963-1964) brings out the same features.

11. The precise definition of monsoon is difficult. However Kurashima (1968) adopted the following practical definition:

a. Monsoon is a part of the planetary wind system.

b. Monsoon occurs frequently during summers and winters and has sufficient stability to represent winds of the respective seasons.

c. Wind direction in winter and summer are nearly opposite.

12. Based on the above definition Kurashima prepared the map of geographical distribution of monsoon. Following evolved from the maps:-

a. Monsoons are indeed global phenomena. Over Asia monsoon occurs over a wide region extending from the Indian Ocean to Kamchatka region.

b. Monsoons are stronger over the Afro-Asia-European region than over the American region.

13. India is mostly within the tropics. Therefore India will be under the influence of axially symmetric monsoon due to north-south shifts of the Hadley cell which arise due to north-south temperature contrasts. During the monsoon season, the symmetric Hadley cell of the northern hemisphere is highly overshadowed by the symmetric circulation over the monsoon region. Reversal takes place of north-south currents from winter to summer. The meridional winds are of the order of 1 m per sec. During the monsoon season the usual Hadley cell exists over two-thirds of the tropics but over the monsoon region, a reverse Hadley cell extends even up to 30°N exists.

14. India will also be under the influence of the axially asymmetric monsoon due to the east-west temperature gradients produced by east-west land ocean contrast. Thus it is necessary for us to understand the physics of both the theories of symmetric and asymmetric monsoon.

15. The physics of the maintenance of this monsoon circulation and the energy sources for the monsoon wind system has been studied by (a) observational studies (b) theoretical studies and (c) numerical model studies.

16. Keshavamurty (1968) studied the balance of angular momentum in Indian monsoon region. He estimated the different terms in the equation for the region equator to 20°N and 50°E to 100°E (Table 7.1).

17. It is seen from these estimates that the zonal winds are maintained by the Coriolis term. In the lower troposphere the Coriolis term maintains the zonal winds against frictional; dissipation. In the upper troposphere the zonal momentum produced by the Coriolis term and the east-west pressure gradient is taken out of the region by mean flux. The implication is that the mean meridional circulation with southerlies in the low level and northerlies in the upper levels turns into the low level monsoon westerlies and the upper level easterly jet by Coriolis turning, because of earth’s rotation.

Table. 7.1: Budget of Angular Momentum

(Units: 10^{25} gm cm^2 sec^{-2})	Lower Troposphere (Westerly Momentum)	Upper Troposphere (Easterly Momentum)
Ω-Transport or Coriolis Term	6.8	7.0
Frictional Torque	-7.0	****
Zonal Pressure Gradient Term	-0.4	2.6
Mountain Torque	2.1	****
Flux divergence		
East-West Mean Transient	-0.3	-5.0
	-0.4	0.5
North-South Transient Mean	0.3	-4.8
	-0.1	-0.5
Balance	1.0	-0.2

18. The monsoon circulation is basically a heat engine. The heating contrast of the Asian continent with respect to the Indian Ocean sets up a vertical Hadley type circulation with southerlies in the lower levels and northerlies in the upper levels. These result in the observed zonal winds because of Coriolis turning (as the earth is rotating). The basic factors are the differential heating and the rotation. In addition to the heating of the Asian continent, the Tibetan plateau acts as an elevated heat source. The monsoon current also imports moisture evaporated over the Arabian Sea and the Indian Ocean and copious rains are produced in the hilly regions of northeast India and in the low level convergent region of the monsoon trough over north and adjoining central India. Large amounts of latent heat are thus released and this heat enhances the north-south temperature gradient and hence the monsoon meridional circulation and the zonal winds. Thus the monsoon engine imports part of its fuel from large distances.

Intra-Seasonal Variations in Monsoon Circulation

19. The large-scale monsoon rainfall is associated with the seasonal migration of the ITCZ. However, during the summer monsoon season, two convergence zones are often observed in the same longitudinal belt, one in the monsoon zone north of 15°N and the other appearing intermittently in the westerly regime over the equatorial Indian Ocean. As both these convergence zones cannot be termed ‘inter-tropical’, Gadgil (1988) suggested the use of the term ‘tropical’ convergence zone’ (TCZ), preferring to distinguish these zones by the large-scale convergence and organized precipitation associated with them rather than by the confluence of the surface winds. Thus, we have a continental TCZ and an oceanic TCZ prevailing in the same longitudinal belt. As mentioned earlier, within the Indian summer monsoon season, the large-scale monsoon rainfall fluctuates between active spells and break situations. During the active spells, the continental TCZ is almost continuously present over the monsoon zone north of 15°N, while during dry spells or intense breaks it complexly disappears from the monsoon zone. The break situation is usually followed by the northward propagation of the oceanic TCZ onto the continent, similar to the onset phase, leading to the revival of the continental TCZ (Gadgil, 1988).

20. It has been observed that the above sequence of break spells and revival of active spells occurs at intervals of about 30-50 days. Sikka & Gadgil (1980) presented detailed analysis of fluctuations in the cloud cover over the Indian monsoon region, based on satellite cloud photographs. They found organized progression of zonally oriented cloud band from the equatorial regions towards the foot-hills of the Himalaya with a periodicity of around 40 days, which was associated with active and break phases of the monsoon (see also Yasunari, 1980). There has been considerable excitement over the detection of 30-50 day oscillation in the monsoon system, leading to a large number of studies to identify this oscillation in various surface and upper-air parameters of the monsoon circulation. However, analysis of long-period data revealed that the 30-50 day mode has a large amount of inter-annual variability (Singh & Kripalani, 1985; Mehta & Krishnamurti, 1988), in which ENSO as well as east-west propagations might be playing important roles.

21. in this connection, it may be noted that spectral analyses of monsoon circulation parameters have revealed both 10-20 day periodicities involving westward wave propagation (Krishnamurti & Bhalme, 1976), and 30-50 day periodicities involving northward propagation (Yasunari, 1979). The performance of the monsoon in a specific year is a result of the relative phases and amplitudes of these two dominant modes. Soman (1993) also found that, while the 30-60 day mode contributed the maximum variance in daily rainfall in the southern and central parts of India, there is a large inter-annual variability in its amplitude, periodicity and speed of northward propagation. In view of this, the practical predictive potential of this oscillation is limited.

TASK-26 : CHARACTERISTIC FEATURES OF SOUTH ASIAN SUMMER MONSOON

Characteristic Features of Monsoon

1. The Asian summer monsoon is a unique feature of the tropical circulation, which has been visualized by some people as ‘breathing in’ during summer and ‘breathing out’ during winter, over the Asian continent. The unique physiographic setting in the South Asian longitudes, with the vast Asian continent spanning equatorial to polar latitudes in the northern hemisphere contrasted by the extensive water surface of the Indian Ocean spanning equatorial to Antarctic latitudes in the southern hemisphere, primarily helps in the development of intense thermal centres of action due to differential heating of land and water. The resultant pressure patterns and meridional circulation in summer are further accentuated by the presence of huge mountain massifs over South Asia, leading to the establishment of the southwest monsoon or Indian summer monsoon.

2. The summer monsoon circulation, which has an almost astronomical precision in the timing of its annual visit to the region, has completely transformed the seasonal variation of the climate over South Asia. Its importance to the tropical circulation and consequently to the general circulation of the atmosphere is well-recognized; in fact, there has been a suggestion that the ‘monsoon year’ (defined as a year starting shortly before the summer monsoon season) should be taken as the climatic year in the tropics (Yasunari, 1991). In this chapter, we will briefly summarise some important characteristic meteorological features which pervade the region during the summer monsoon season, in contrast to the features during other parts of the year.

Sea Level Pressure Patterns

3. The sea level pressure distribution over South Asia shows a gradual transition, during March to May, from the winter conditions to the characteristic patterns of the summer monsoon circulation. This period is known as the pre-monsoon season, which is marked by high temperatures and extensive heat lows over the land areas. The land low over South Asia is a part of the global-scale low extending from the Sahara to Central Asia across Arabia, Iran, Afghanistan, Pakistan, northwest India, and even up to Myanmar. By May, the continental low-pressure areas completely dominate North Africa and Asia. The main centre over South Asia lies over northwest India and adjoining Pakistan, with an extension as a trough into peninsular India, which becomes more marked in June. The lowest pressure for the whole belt from Africa to Asia is over the central parts of Pakistan. The progressive development of this heat low before and during the monsoon season is considered to be one of the major causative factors of the Indian summer monsoon circulation.

4. Monsoon activity reaches a peak in July when the low-pressure area extending from north Africa to northeast Siberia is most intense, with its main centre over north Baluchistan. A trough lies over north India with its axis running from northwest India to Head Bay (north coast of the Bay of Bengal) which is known as the ‘monsoon trough’. As the monsoon becomes established over South Asia, the subtropical anticyclones over the Indian Ocean are strengthened, thereby creating strong pressure gradients north of the subtropical highs up to the monsoon trough. This facilitates large-scale cross-equatorial flow into the South Asian region. The pressure patterns over South Asia, south of the monsoon trough, are characterised by weak ridges in the Arabian Sea off the west coast of

India and in Bay of Bengal off the Tennaserim coast and over Myanmar, and a weak trough over eastern peninsular India off the east coast. The intensity of the Afro-Asian low starts decreasing in August, with a consequent decrease in the pressure gradient south of the monsoon trough. Towards the end of the monsoon season, the trough over northern India shifts to the Bay of Bengal and the pressure field becomes flat over almost all South Asia and the winter conditions are gradually re-established.

The Heat Low

5. The heat low over the central parts of Pakistan and neighbouring regions, which is a very important causal factor of the monsoon, is generally linked to the regions of maximum heating which are out of reach of the maritime air mass. The tapering shape of peninsular India, the Himalayan barrier to the north and the Assam hills to the northeast, which ;make the maritime air mass pervade over most of South Asia, displace the centre of the low to the extreme northwest of the continent. Blocking of cold air incursions from the north by the Himalaya in the lower troposphere also makes the heat low more intense in this region. The existence of the lowest pressures of Afro-Asian low in this region, in spite of the relatively higher temperatures in the Sahara, is a result of the topographical configuration and the associated adjustments in the wind field.

6. The heat low is shallow, extending up to about 1.5 km and is overlain by a well marked ridge extending to the upper troposphere, which is part of the subtropical high pressure belt. The role of the upper ridge in maintaining the surface heat low is not clear. While the pressure distribution of the upper ridge is inimical to that of the surface low, it is argued that the subsidence in the upper ridge maintains the clear skies required for the heat low; however, this argument cannot hold in the pre-monsoon months when the subtropical ridge is well away from the heat lows over South Asia (Rao, 1976). When the western disturbances, which bring cold air in their wake, move north with the advance of summer, heating continues uninterruptedly in May in the absence of any rain or clouding over the area of the heat low.

7. The intensity of the heat low has been correlated with monsoon activity. Negative pressure anomalies in the heat low region and positive anomalies over peninsular India are regarded as favourable conditions for good monsoon activity.

Monsoon Trough

8. During the summer monsoon season, the most conspicuous signature of the monsoon circulation at the surface is a trough extending from the heat low over Pakistan southeastwards up to Gangetic West Bengal, termed the monsoon trough. The monsoon trough is a major semi-permanent feature of the summer monsoon circulation in the lower troposphere and exerts considerable influence on the summer monsoon activity in South Asia. The monsoon trough is regarded as a part of the equatorial trough of the northern summer in the Indian longitudes and indicates the separation between air of northern and southern hemispheric origin. The monsoon trough is flanked by westerly to southwesterly winds to the south and easterlies to the north. The air mass on the south of the trough line is derived from the Arabian Sea branch of the monsoon, while that on the north is derived from the Bay of Bengal branch. In the daily weather charts the trough can be located up to the 600 hPa level (about 6 km) during periods when the monsoon has advanced into the northern parts of India.

9. It is observed that the trough line tilts southward with height; the slope in the western sector is steep in the lower troposphere and less marked in the middle troposphere, while in the eastern sector the slope is moderate in the lower troposphere and increases appreciably in the middle troposphere. The temperature to the north of the trough is on average about 2°C higher than that to the south, which explains the southward slope of the trough axis with height (Joshi & Desai, 1985). The mean position of the monsoon trough during the month of July is from Ganganagar to Calcutta through Allahabad. The trough axis experiences considerable day-to-day variation in its position, which has a vital bearing upon the monsoon rainfall distribution in the region. No other semi-permanent feature has such a control on monsoon activity. This arises mainly from the different synoptic systems that prevail in association with different positions of the trough. During the break monsoon periods, when there is a temporary 'lull' in monsoon activity, the trough line shifts to the foot-hills of the Himalaya, rainfall over the central parts of India decreases considerably, and there is an increase in rainfall over north India along the Himalayan foot-hills. The monsoon depressions, which originate over the Head Bay of Bengal, move across the Gangetic plains in a northwesterly direction along the trough line. A southward swing of the trough line to the central parts of India is generally associated with the monsoon depressions. Table 18.2 indicates the normal positions of the monsoon trough. It can be seen that the trough is in the normal position only 30-40% of the time, indicating its high intra-seasonal variability. It is also seen that the western part of the trough is more often towards the north than towards the south, while the eastern part shows the opposite trend. This is due to the influence of the mid-latitude disturbances on the northwestern parts of the subcontinent and the formation of monsoon depression over the head of the Bay of Bengal (Srinivasan & Ramakrishnan, 1970). Energy balance studies in the monsoon trough region indicate that the vertical structure of the trough has a large convergence zone in the lower levels and a compensating outflow higher up (Anjaneyulu, 1969). Along the normal position of the trough, the rainfall distribution shows a minimum. This is due to the intense rainfall activity on the southern and northern sides of the trough axis, associated with its north-south excursions.

Table7.2 Mean Positions of Monsoon Trough during 1961-69
(After Srinivasan & Ramakrishnan, 1970)

Month	Longitudinal section (°E)	Normal position of trough (°N)	Percentage frequency of location of trough		
			South of normal	Normal Position	North of Normal
July	77	27-29	27	30	43
	81	25-27	27	45	28
	87	23-25	37	35	28
August	77	27-29	20	33	47
	81	25-27	20	47	33
	87	23-25	34	39	27

Surface Winds

10. During the monsoon season, the northern parts of South Asia, north of the monsoon trough, have weak easterlies at the surface. Southerlies occur over West Bengal and Bangladesh, while the rest of the subcontinent is dominated by southwesterlies and westerlies at the surface. The southwesterlies are relatively stronger over the southwestern Arabian Sea. Nearer the west coast of peninsular India, the winds are west-southwesterlies to westerlies, with speeds of 5-10 m per sec, while the mean speeds do not rise above 5 m per sec inland. This pattern persists throughout the active phase of the summer monsoon, and towards the end of September, as the monsoon withdraws from the subcontinent, there is a large-scale weakening of the wind field consequent to the weakening of the pressure gradient.

Upper Winds

11. The subtropical high-pressure belt, which passes over northern India throughout the year, greatly influences the upper-wind distribution. During the pre-monsoon months, the subtropical ridge over land around 18°N at the 850 hPa level and extends up to 700 hPa. At higher levels, the subtropical ridge gradually shifts southwards to 8°N at the 200 hPa level. At this level, the anticyclone from the east extends up to Sri Lanka. Westerlies strength with height over northern India, reaching maximum speeds of around 20 m per sec near 25°N at 300 hPa and 25-30°N at 300 hPa and 25-30 m per sec near 25-30°N at 200 hPa. The subtropical westerly jet stream in the upper troposphere is found a little north of Delhi with its core at the 200 hPa level. Above these levels, speed decreases slowly. Some upper-level easterlies are found over southern peninsular India and Sri Lanka, but they are generally weak during the pre-monsoon months and gradually start building up towards the beginning of the monsoon season.

12. As the monsoon becomes established over the whole subcontinent towards late June and early July, the lower tropospheric winds (up to 700 hPa) are characteristic of the monsoon trough, dominated by westerlies to the south of the monsoon trough and easterlies to its north. Westerly winds over the peninsula increase with height in the lower layers, reaching a maximum (10-12 m per sec) around 900 hPa in the west and 800 hPa in the east. The monsoon trough tilts southward with height and gradually loses significance around 500 hPa. Above 700 hPa, the subtropical high from the west over-rides the warm surface low over Pakistan and surrounding area. At 500 hPa, this ridge becomes better organized and is flanked by easterlies over northern India and a trough over southern peninsular India and the adjoining sea. However, winds at this level are generally weak. At 300 hPa and above, almost the whole subcontinent is dominated by easterlies. The subtropical westerly jet stream over north India weakens and its core is shifted north of the Himalaya. At levels between 200 and 100 hPa, another ridge (the Tibetan high) develops over northeast India and the easterlies strengthen, reaching a maximum (30-40 m per sec over southern peninsular India) at 100 hPa, becoming concentrated into a core of high winds known as the tropical easterly jet stream. These are the typical upper wind patterns which throughout the monsoon season with remarkable stability.

Temperature

13. The highest temperature over India during the pre-monsoon or hot weather season, which coincides with the spring season (March - May) of northern middle latitudes. During this season, the monthly mean surface air temperatures exceeds 33-35°C in the inland areas of northwest India and adjoining areas. The surface temperatures, particularly the daytime temperature, drops dramatically with the onset of the summer monsoon, the monthly means falling to less then 30°C. This is due to the extensive cloudiness and rainfall associated with the monsoon circulation.

14. During the pre-monsoon season, a thermal high develops at 850 hPa over central India, with temperatures decreasing in all directions. The thermal high shifts southwards and becomes weaker with height and is seen as a weak thermal high over southern peninsular India at about 300 hPa. The thermal field becomes flat at the 200 hPa level, with temperatures varying in the range -52 to -48°C. Above this level, the temperature gradient is reversed, with a marked increase in temperature from the central parts of the peninsula towards the north.

15. During the monsoon season, the thermal field at 850 hPa over south Asia is dominated by a ridge in the north extending form the thermal high over Iran, and two thermal troughs running along the west coast if India and the Myanmar coast, with an intervening ridge near the east coast of India. At 700 hPa and above, the troughs and ridge around peninsular India disappear and the thermal field is dominated by the ridge in the north, which becomes better organized at still higher levels and results in a prominent ridge over the Tibetan region at 200 hPa, which is a very important feature of the monsoon circulation.

TASK-27 : ONSET AND WITHDRAWAL OF SUMMER MONSOON

Introduction

1. The seasonal transition from the pre-monsoon to monsoon is so sudden that it is often termed the 'burst' of the monsoon over the Indian subcontinent. The most perceptible changes associated with the establishment if the monsoon circulations are surface winds blowing from the southwest, a sudden fall in the surface air temperature, and an increase in cloudiness and rainfall. The southwest monsoon over the Indian peninsula first arrives over the south Indian state of Kerala. While almost simultaneously the Bay of Bengal branch reaches West Bengal and the hills of Assam in northeast India. Kerala is widely known as the 'gateway' of the Indian summer monsoon: its onset over the south Kerala coast is one of the most important meteorological events in the region and is eagerly awaited every year by the public in general as well as the Indian meteorological community. This events heralds the rainy season for the region, ending the days of parching heat and bringing much-needed water for agricultural, industrial and domestic rains over other parts of the subcontinent, which heralds the start of agricultural operations for the main cropping seasons of the years known as *kharif*. The onset, advance and withdrawal of the monsoon are particularly critical to rain fed agriculture during the *kharif* season.

Normal Dates of Onset and Withdrawal

2**.** Although the four month period of June to September is termed the Indian summer monsoon season in a general large-scale sense, the actual rainy period differs widely over different parts of the subcontinent. A sustained increase in rainfall and cessation of rainfall activity have been traditionally used to demarcate the beginning (or onset) and end (or withdrawal) of the monsoon circulation, respectively. The calendar dates of these seasonal demarcations also have considerable inter-annual variability, of the order of 1-2 weeks. The standard deviation of the date of onset of the monsoon over the west coast of India is about 6-7 days (Ramdas *et al.,* 1954) while that around Delhi is 7-8 days (Bhullar, 1952). The mean picture of the onset and withdrawal of the monsoon over different parts of the subcontinent can be obtained by examining long-term mean daily or pentad (5 day) rainfall at individual stations spread over the subcontinent and locating the dates of sharp increases or decreases in rainfall. However, this procedure runs into difficulties in places where there is substantial pre-monsoon thunderstorm activity at the time of onset, such as at Kerala in southern India. In view of this, identifying the dates of onset and withdrawal in individual years and then averaging the dates may give a more reliable mean picture.

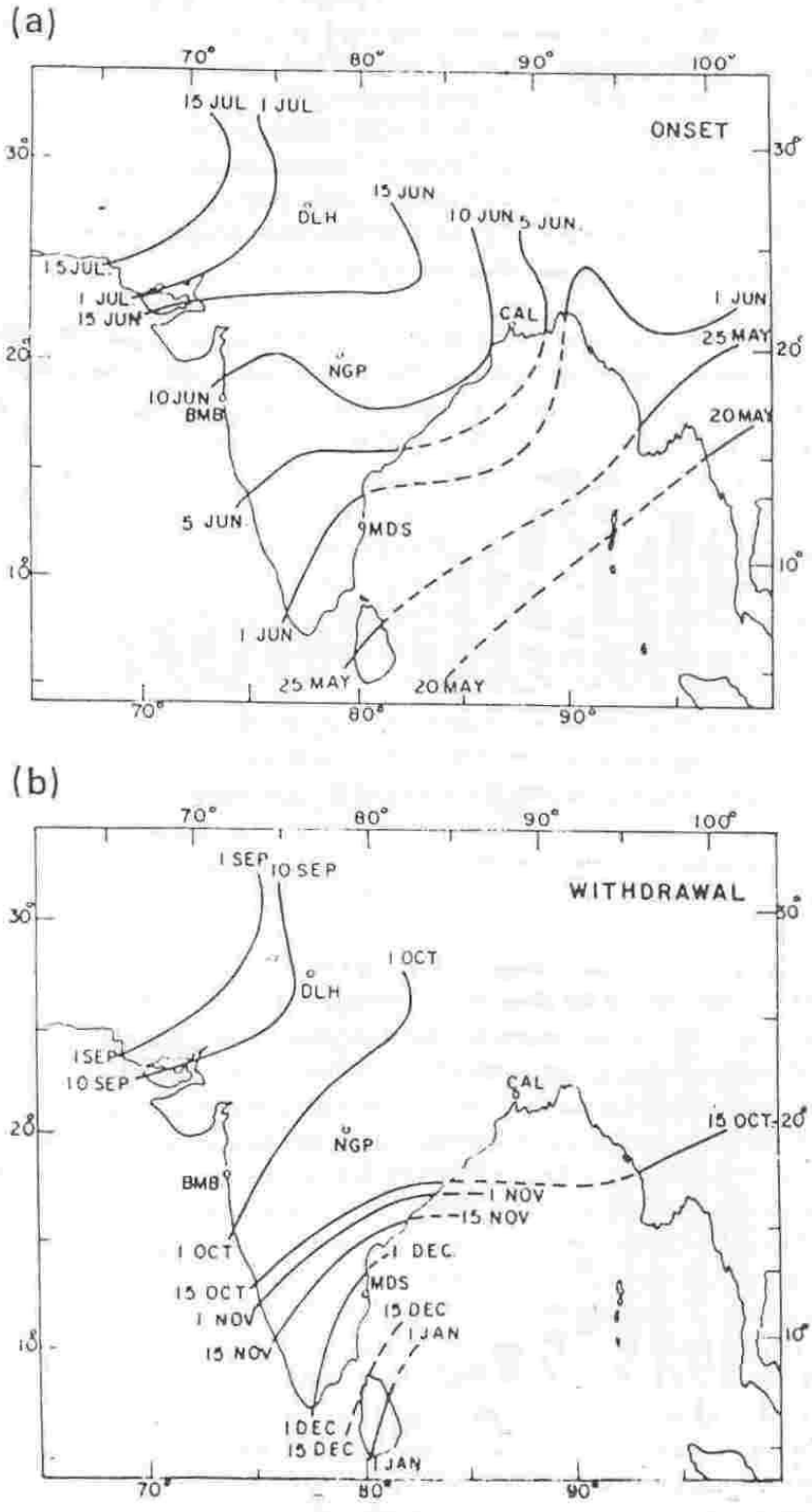


Fig 1.1 Normal dates of onset (a) and withdrawal (b)of the summer monsoon over India

3. Fig 1.1 a, b show isolines of the mean dates of monsoon onset and withdrawal prepared on the basis of long-period data by the India Meteorological Department (IMD, 1943). These dates have been derived from normal pentad rainfall data at different stations using data up to 1940; the middle date of the pentad representing an abrupt rise in rainfall was taken as the onset date. These diagrams, published more than half a century ago, are still widely used as a standard source of reference in Indian as well as in international meteorological literature. Although the approach of identifying the onset dates from normal pentad rainfall has several limitations (Soman, 1993), and despite some attempts (Subbaramayya & Bhanu Kumar, 1978; Subbaramayya *et al*., 1984, 1990) to revise the onset diagram, these diagrams continue to be used, perhaps indicating the controversial nature of the problem and the difficulties involved in identifying a unique set of dates of onset and withdrawal of the summer monsoon. The isochrones (Figure 18.2a) show that the monsoon rains commence around the third week of May over Sri Lanka and the Andaman Islands in the Bay of Bengal, while the monsoon enters the Indian mainland on 01 June over the south Indian state of Kerala. The onset of the monsoon brings with it a dramatic change in the weather situation, associated with a conspicuous cloud cluster over a large area near the western part of the southern tip of the Indian Peninsula. Ananthakrishnan *et al.,* (1981) analysed the pentad rainfall data of six stations in the Andaman Nicobar group of islands for the period 1953-78 and studied the onset of the summer monsoon rainfall over the southeast Bay of Bengal. Their study suggests that the onset dates are ahead of the dates interpolated from the normal onset dates from the men charts. In the island stations the onset of the monsoon from south to north occurs progressively from the last week of April to the first week of May. It normally takes the monsoon a period of about 1½ months to reach northwest India and Pakistan. It proceeds gradually northwards and by the middle of July the whole of the Indian subcontinent comes under the grip of the monsoon. However, the northward progress of the monsoon is not a smooth affair and takes place in surges, interspersed with periods of weakening or stagnation of monsoon activity. The northward advance is usually associated with synoptic disturbances and there is a pronounced tendency for the formation of low-pressure systems at the leading edge of the monsoon current. This is also termed the ‘onset vortex’. Ananthakrishnan *et al.* (1968) found that, on about 75% of occasions, the advance of the monsoon is associated with either a feeble trough in the low level, a depression or a cyclonic storm. Subbaramayya *et al.* (1984) distinguish three phases in the advance of the monsoon over India, interspersed with periods of stagnation: the first affects the southern peninsula, the second reaches the northern peninsula and central India, and the third penetrates to the central parts of north India. Joshi *et al*. (1990), using satellite data, have shown that the onset of the monsoon is preceded by an increase in the tropospheric temperatures over almost the entire region, and these temperature changes are more prominent over the heat low and Tibetan plateau region, compared to other neighbouring areas.

4. The withdrawal of the monsoon begins in the western parts of the; northwest Indian state of Rajasthan in early September (Fig). Thus, the effective duration of the southwest monsoon rains over this area is only about 1½ months. The southward retreat of the monsoon rains continues rapidly until about the middle of October, by which time it withdraws completely from the northern half of the Indian peninsula. The withdrawal phase of the monsoon is associated with the equatorward motion of the ITCZ, causing heavy rainfall in its wake (Subbaramayya *et al.,* 1977). Over the southern part of the peninsula, the withdrawal phase becomes mixed up in a transition to the northeast monsoon type of circulation and cannot be located precisely from rainfall variations. Thus, it is not appropriate to identify the much later decrease of rainfall in south India as the withdrawal of the southwest monsoon.

Onset of Indian Summer Monsoon over Kerala Coast

5. The announcement of the date of arrival of the monsoon over Kerala, though a subjective decision, is based on an overall judgement taking into account the changes in circulation features in the lower and upper troposphere which indicate a seasonal reversal of wind regimes, and a sustained increase in rainfall over Kerala and the island stations over the southeast Arabian Sea. The specification of objective criteria for fixing the onset has several difficulties owing to the considerable inter-annual variation of the meteorological features associated with the onset. Furthermore, the onset is accompanied by significant changes in circulation features, the vertical distribution of moisture, cloud characteristics, the state of the sea etc., which are difficult to quantify in terms of criteria for onset. The problem is further confounded by the fact that Kerala frequently witnesses widespread rainfall activity due to pre-monsoon thundershowers which may be difficult to differentiate from the; monsoon rains. Therefore, the experience and studied judgement of the forecaster contribute to a large extent to the declaration of the date of onset.

6. Since the early days of Indian meteorology, an essential requirement for declaring the; monsoon onset has been a sustained increase of rainfall at observatory stations over Kerala and the Arabian Sea islands. Ananthakrishnan *et al.* (1968) made a pioneering study of the criteria for declaring the date of onset of the Indian summer monsoon over Kerala. Their study revealed considerable differences between the real-time dates issued by the IMD and those refixed by a post-mortem examination of rainfall sequences. This has necessitated a re-examination of past data to decide on the onset date based on some objective criteria, in order to make the onset dates useful for scientific investigations. Although the use of upper-air and satellite data has considerably reduced the subjective errors in deciding on the dates of onset, one has to rely on the rainfall data alone to generate long-period series of onset dates.

7. The onset of the summer monsoon over Kerala is characterized by a transition in the nature of rainfall from light- to heavy-spell category. This can be seen every year, despite appreciable inter-annual variations. The general impression that the pre-monsoon thunderstorm rains merge with the monsoon rains and confound the onset phase of the monsoon, as stated earlier, is mainly drawn from an examination of mean rainfall sequences. However, this is not the case in individual years. The merging occurs mainly because the averaging of calendar date rainfall data causes dispersal in the onset dates over the period of averaging, blunting the abrupt rainfall changes associated with the onset. Ananthakrishnan & Soman (1987) devised some objective criteria for fixing the onset dates in individual years based on the analysis of 80 years data (1901-80) of daily areal mean rainfall series for south and north Kerala. They considered the date of the onset as the first day of the transition from light to heavy rainfall category, with a provision that the average daily rainfall during the first 5 days after the transition should not be less than 1 cm. They also demonstrated the abruptness with which the monsoon sets in Kerala by superimposing the daily rainfall amounts relative to the onset dates (figure 18.3) identified using this criterion. Here, the onset dates are labelled as zero for every year and the rainfall amounts for days 0 and ± n (n = 1, 2, 3,….) are averaged over the 80 year period. The superimposed epoch analysis clearly shows that the rainfall jumps abruptly from about 0.5 - 0.6 cm per day during the pre-monsoon months to over 1.7 cm on the onset day.

8. Based on this objective method, Ananthakrishnan & Soman (1988, 1989) have prepared long-term series of onset dates for south (1891-1990) and north (1870-1990) Kerala. They reported that the mean date of onset for south Kerala is 30 May, with a standard deviation of 8.5 days, while the mean onset date for north Kerala is 01 June, with a standard deviation of 8.4 days. The frequency distribution of the onset dates over south Kerala during the 100 year period 1891-1990 is shown in Figure 18.4. The onset has the maximum frequency in the 3 day interval 1-3 June and the frequency distribution has a skewed character (Soman, 1993). Although the onset of the monsoon over Kerala may appear to be of local character, it is generally considered to be a pointer for advancement of the monsoon over India. For example, Deshpande *et al.* (1986) reported that the anomalies in the dates of onset were similar at Kerala and Bombay on 70% of occasions. An examination of the time series plot of onset dates of the southwest monsoon, as deviation from long-term mean, over south Kerala (Figure 18.5) suggests considerable inter-annual variability. During the last 100 years there have been extremes in the dates of monsoon onset over Kerala of as much as three weeks from the normal. However, no significant long-term trend in the time series of onset dates has been observed. However, the monsoon onset was more often in June than in May during the period 1901-40 while the reverse occurred during 1941-80. Figure 18.5 also shows the long-term changes in the onset date by means of a smoothed curve obtained from the 9-point Gaussian low-pass filtering of the anomalies of onset dates over south Kerala. The smoothed curve shows a prominent bunching of late onset dates for a few decades around 1900 and of early onset around 1960.

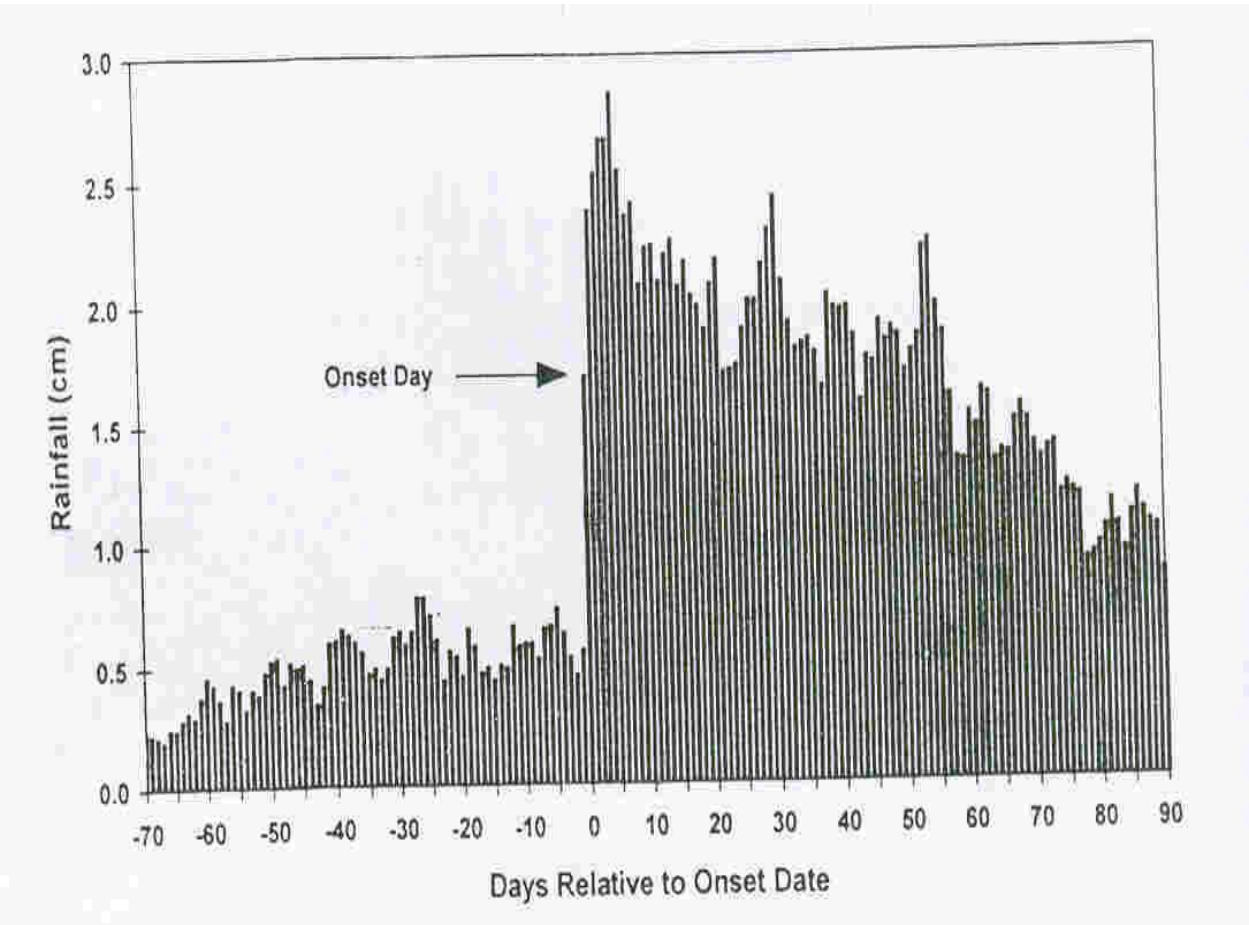


Fig 1.2 Average daily areal mean rainfall for south Kerala obtained by superimposing the onset dates.

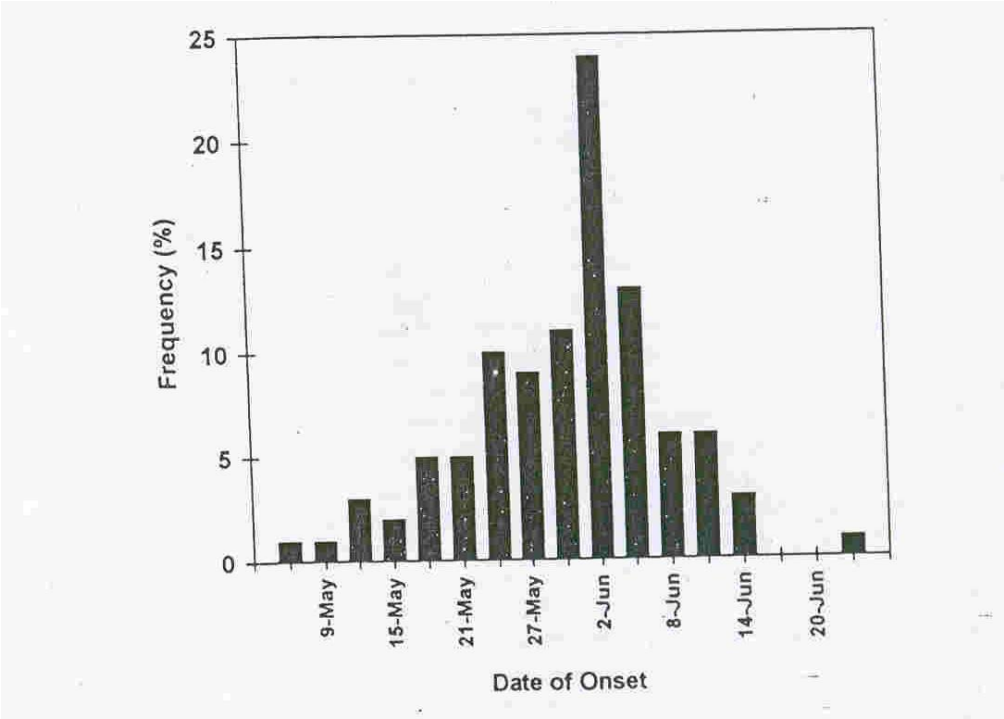


Fig 1.3 The frequency distribution of the onset dates of the southwest monsoon over south Kerala at 3 day intervals for the period 1891-1990.

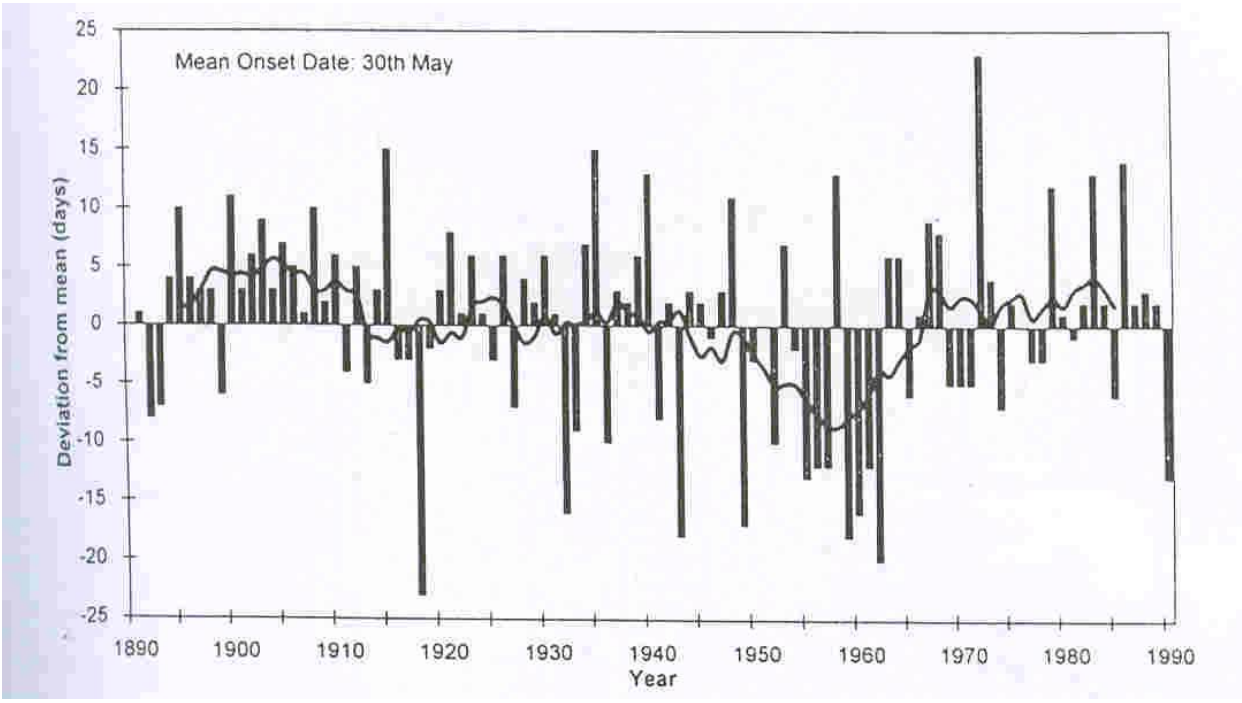


Fig 1.4 Time series of the deviation from the mean onset (30 May) over south Kerala for period 1891-1990.The smoothed curve indicates the 9-point low-pass filtered series

9. Soman & Krishna Kumar (1993) examined the circulation and moisture changes over India associated with onset of the southwest monsoon over south Kerala, by compositing the meteorological parameters such as rainfall, wind, relative humidity, vertically integrated zonal moisture transport and outgoing long-wave radiation (OLR) at several stations. The study is based on 80 years (1901-80) of daily rainfall data and 15 years (1971-85) of daily aerological data at a well-distributed network of stations in India. The following are some of the main features associated with the onset of monsoon as brought out by their study:

a. The rainfall composites at most of the Indian peninsula stations, except some on the east coast, showed an abrupt increase with respect to onset over Kerala, with an appropriate lag based on the normal date of onset at these stations.

b. Subtropical westerlies weaken and shift poleward and tropical easterlies strengthen and spread towards the north with the onset of the monsoon.

c. The composite streamline charts depict a trough in the mid-tropospheric levels between 600 and 400 hPa. This could be regarded as the manifestation of the onset vortex noticed in some years over the eastern Arabian Sea during the onset.

d. Vertical distribution of relative humidity shows rapid deepening of the moist layer a few days before the onset at the respective stations. This large-scale moisture build-up over the Indian landmass before the strengthening of the monsoon flow is similar to the increase noticed over the Arabian Sea.

e. The OLR composites show a convective area developing near the equator in the vicinity of 70°E about 5 days before the onset over south Kerala. The intensification and northward movement of this convection brings the first monsoon rains over Kerala.

TASK-28 : MONSOON DEPRESSION

Introduction

1. Monsoon depressions are important synoptic-scale disturbances in the hierarchy of tropical circulation systems and make major contributions to the monsoon circulation in organizing the low-level convergence. The low-pressure systems and depressions during the monsoon season move along the axis of the monsoon trough, towards the northwest. The depressions are low-pressure areas having two to three closed isobars at 2 hPa intervals covering an area of about 5° latitude by 5° longitude. Monsoon depressions normally form in the Bay of Bengal north of 18°N latitude and move west-northwest across the central and northern parts of India giving widespread rains, with relatively heavy rainfall in the southwestern quadrant of the system. These low-pressure systems are referred to as 'depressions' when the surface wind speeds are between 8-16 m per sec and as 'cyclonic storms' when the speeds are above 16 m/s. in addition to providing beneficial rainfall along and near their tracks, these systems transport heat and moisture upwards and maintain the activity of the monsoon trough. Table 18.3 shows the number of depressions and cyclonic storms during the monsoon season over the South Asian region during the period 1891-1970. Some of the systems in early June, usually associated with the advance of the monsoon, and also in September are not typical of monsoon depressions, particularly in the Arabian Sea. The number of monsoon depressions and cyclonic storms during the course of a single season shows considerable inter-annual variability. While there have been several years without a single depression or cyclonic storm throughout the monsoon season, there have been years which have witnessed as many as 12-15 depressions and/or cyclonic storms in a single season. Monsoon depressions rarely grow into cyclonic storms; they do not acquire hurricane intensity during the summer months as they are not long over oceanic areas before reaching land, and the large-scale vertical wind shear in the troposphere inhibits local accumulation of latent heat release over the disturbance (Cadet, 1970). Cyclonic storms of very severe intensity are formed in the South Asian seas only during the post-monsoon season.

Table1.1 Number of Depressions & Cyclonic Storms during the Monsoon Season (1891-1970) (After Rao, 1976)

Region of Formation	Depressions				Cyclonic Storms			
	Jun	Jul	Aug	Sep	Jun	Jul	Aug	Sep
Bay of Bengal	71	107	132	141	35	38	26	32
Arabian Sea	18	9	2	9	15	3	2	5
Land Areas	12	39	42	21	1	1	0	1

2. Most monsoon depressions form over the warm waters of the northern parts of the Bay of Bengal in association with a series of low-pressure systems and easterly waves that move over the region during the monsoon season across Myanmar. They generally develop out of three types of situations (Rao, 1976):-

a. In almost 50% of cases, a diffuse pressure field develops over the northern By of Bengal and adjoining areas, associated with a southward swing of the monsoon trough, where the pressure gradient to the south of the monsoon trough is fairly strong as a normal feature of the monsoon season.

b. About one-third of the depressions are generated from the diffuse low-pressure systems travelling across Myanmar into the northern Bay of Bengal., which are remnants of the typhoons of the China Seas.

c. About 15% of the monsoon depressions are seen initially as closed circulations in the middle troposphere, which gradually descend to lower levels leading to the formation of surface lows.

3. The frequency of monsoon depressions formed in the Bay of Bengal is about 80% of the total number of depressions forming in the South Asian monsoon region. Relatively few monsoon depressions form in the east Arabian Sea and over land in the monsoon trough region. The average annual frequency of monsoon depressions is seven, of which one occurs in June and two each in July, August and September.

4. The monsoon depressions normally take a west-northwesterly course during the peak monsoon months of July and August, while they move in a northwesterly direction during the transition periods in early June and late September. On average, a monsoon depression travels at the rate of about 300-400 km per day. The depressions generally weaken after crossing the central parts of the country, owing to the cut-off in moisture supply, then move further northwest and merge in the seasonal low over west Rajasthan and adjoining Pakistan, Sikka (1977) found that one-third of the depressions fill up on crossing 85°E and nearly half on crossing 80°E. On a few occasions, when the depressions are fed by fresh moisture supply from the Arabian Sea branch of the monsoon, the depressions may retain their intensity and take a westerly course to reach Gujarat state (India), causing very heavy rains in the region. Quite often, the monsoon advance over northwest India during the latter part of June occurs in association with the movement of a monsoon depression over that region. The anomalies in the frequency and tracks of the monsoon depressions substantially contribute to the spatial and temporal variability in monsoon rainfall and are believed to be associated with the large-scale circulation features of the monsoon system.

5. The surface winds are stronger to the south of the centre of monsoon depressions as a result of the superimposition of the Westerlies, caused by the depression perturbation south of its centre, on the prevailing basic flow of moderate to strong Westerlies. The intensity of the wind circulation is strongest around the 700hPa level and decreases upwards as well as downwards. The cyclonic becomes very weak and generally disappears at around the 300 hPa level. Monsoon depressions are generally cold-core systems below 700hPa and have a warm core higher up; on the rare occasions when they intensify to become cyclonic storms, they have a warm core throughout their depth. The axis of the depression joining the centres are different levels slopes south-westwards with height. The heaviest rainfall associated with a monsoon depression occurs in its south-western sectors (Pisharoty & Asnani, 1957; Koteswaram & George, 1958).

6. When a monsoon depression is located over the head of the Bay of Bengal, almost the whole Indian subcontinent experience widespread rainfall activity. This situation is generally associated with the development of a trough of low pressure along the northern parts of the west coast of India, inducing increased rainfall. Bhalme & Mooley (1980) and Mooley & Parthasarthy (1983) have shown that the monsoon depressions have greater westward penetration during years of excess monsoon rainfall over the Indian region than during years of deficient rainfall. The rainfall associated with monsoon depressions has less intensity than that associated with cyclonic storms however, it covers a much larger area, particularly over the land regions.

7. The cyclonic vortices, lows and depressions embedded in the monsoon flow play a notable role during the monsoon season by helping in organizing the monsoon flow and producing rainfall over a wider area through convergence and associated vertical motion. Owing to the transport of heat and moisture upwards over the lows and depressions, the periodic passage of these lows across the country maintains the normal location and activities of the monsoon trough, which are condusive to good rainfall distribution over a large area. The number of low-pressure systems during good monsoon years is generally large and they traverse central India in succession, contributing subcontinent. There is a year-to-year variation in the number of storms and depressions during the monsoon season over the region of South Asia. There is apparently no direct relationship between the number of these cyclonic systems and the monsoon seasonal rainfall over the region (Mooley & Shukla, 1989). This is partly to the varying intensity and rain-producing potential of these systems. However, Pant *et al.,* (1988) identified a low-frequency signal common to the time series of all-India summer monsoon rainfall and the number of storms and depressions, suggesting the possible influence of some slowly varying large-scale climatic anomalies on the number of monsoon depressions and cyclonic storms during the period 1891-1980 over the Indian land area, with the highest number of 200 being over the head of the Bay of Bengal.

Mid-Tropospheric Cyclones

8. The mid-tropospheric cyclones (MTCs) are upper-air cyclones circulations having the maximum intensity in the middle tropospheric levels, between 600 and 400hPa. These systems, which are either non-existent or rarely seen as a trough on the surface charts, generally occur over the northern parts of the west coast of India and occasionally over the northern Bay of Bengal. They are not a frequent as monsoon depressions and are usually observed during the first half of the monsoon season. Observations show that these systems have a cold core below the 600hPa level and a warm core above, and generally have a horizontal extent of the order of about 500-1000km. They help to accentuate the monsoon activity over the region and sometimes help in the early shift of the monsoon current to the drier western parts of the Indian subcontinent. Similarly systems are observed over other areas in the Asian monsoon region such as China and parts of Indo-China, but with less frequency.

9. A detailed study on MTCs over the Northeast Arabian Sea region was carried out by Miller and Keshavamurty (1968) based on the intensive observations during the International Indian Ocean Expedition. It was observed that the area of convergence and cyclonic circulation was well-marked between the 600 and 500hPa levels, the system extending to 1000 km horizontally and 6-8 km vertically. Unlike the monsoon depressions, MTCs are often quasi-stationary for several days or move slowly westwards. The typical life-span of an MTC is in the range of 3 to 7 days: only rarely do they last more than a week. MTCs generally develop in situ over the northeast Arabian Sea. However, a few monsoon depressions coming from the Bay of Bengal occasionally move westwards and transform into MTCs over western India (Carr, 1977).

Off-shore Trough along West Coast of India

10. During the summer monsoon season, particularly in the onset and advance phase, the existence of a weak sea level trough off and along the west coast of India is a common feature seen in the daily weather charts. Small vortices develop within this trough, known as

off-shore vortices (George.1956) which are shallow zones of convergence causing enhanced rainfall activity along the coastal belt. These systems have a horizontal extent of about 100 km and a vertical extent up to the 600 hPa level. The off-shore troughs form more often near coastal Karnataka, though they may also appear and disappear in situ over any area along the west coast. They often move northward along the coast, with associated northward movement of the belt of heavy rainfall. Pressure gradient increases to the south of the trough, particularly as it shifts to the north. Upper winds are affected only in the lower troposphere, often below one km. Nearly half of the active to vigorous monsoon situations in Konkan and three-quarters of such events in coastal Karnataka are associated with the off-shore trough (Rao, 1976). Heavy rainfall generally occurs in the southern portion of the trough. But the intensity and distribution of rainfall along the coast may vary widely in association with the trough.

TASK-29 : BREAK MONSOON

Introduction

1. Although the mid-monsoon months of July and August contribute most of the seasonal monsoon rainfall, the rainfall distribution is not spatially or temporally uniform over the subcontinent even in these two months. The rainfall undergoes significant variations associated with changes in the synoptic patterns. There are occasions when the monsoon activity abruptly weakens and dry weather prevails over most parts of South Asia. These situations are called 'break monsoon periods' and are characterized by certain distinct circulation features. The breaks occur with varying intensities and durations and are mostly associated with the north-south excursions of the monsoon trough. When the monsoon trough shifts close to the foot-hills of the Himalaya, a sudden decrease in rainfall ensues over most parts of peninsular India but there is a considerable increase along the foot-hills of the mountains in India and Nepal. this is sometimes associated with an increase in rainfall in the southernmost parts of the Indian peninsula and Sri Lanka. During this period, situations characterized by cessation of rain and the disappearance of monsoon-like synoptic conditions may occur.

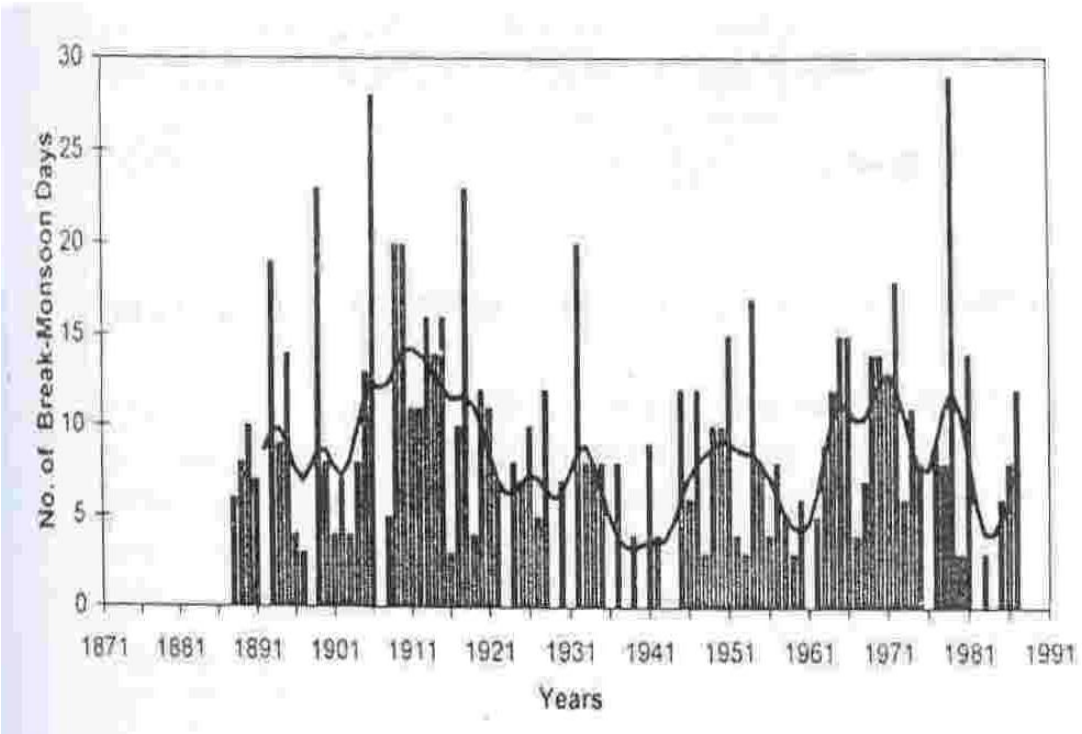


Fig 1.1 Time series of number of break monsoon days for the period 1888-1990; the smoothed curve indicates the 9-point Gaussian low-pass filtered series.

2. The number of break monsoon days in the months of July and August for the period 1888-1967 are shown in Table 1.1 (Ramamurthy, 1969). Ramamurthy (1969) adopted the following criteria for defining the occurrence of a break:

a. The monsoon trough is absent near the surface up to about the 850 hPa level.

b. The duration of such a synoptic situation is at least 2 days.

Table 1.2 Statistics of Break Monsoon Situations During Jul and Aug (1888-1967)

Month	No. of breaks	No. of break days	Average duration (days)	Longest break (days)	Most frequent duration (days)	No. of break-days in		
						First ten days	Second ten days	Remaining period
July	53	306	5.8	17	4	81	117	108
August	55	356	6.5	20	3	115	159	82

3. It is observed that the middle 10 day period of August has witnessed the most frequent occurrence of break monsoon situations, which are also of longer duration than in July. Ananthakrishnan & Pathan (1971) found a distinct minimum around 14-18 August in the normal pentad rainfall at several stations. Ramaswamy (1973) found that, during the period 07-15 August, the daily rainfall is less than the month's daily mean in the central and western parts of the country, while it is more in the Himalayan belt. He calls this the normal period of a large-scale break. During prolonged breaks the troposphere over the region as a whole gets warmer and it appears as if the monsoon has temporarily withdrawn from the region. The breaks in the monsoon should not, however, be confused with the active and weak spells which are seen in rainfall pattern, as well as the convective cloud bands which traverse the region from south to north. While the typical circulation features associated with a break are easily discernible, there are no consistent precursors to its development, which makes the forecast of a break extremely difficult. The break usually ends with the formation of a monsoon depression over the head of Bay of Bengal or a low-pressure area along the west coast of India. In years when the break occurs in the middle of September, the monsoon may withdraw prematurely from the subcontinent.

4. Alexander *et al.* (1978) described the circulation anomalies over India during active and break monsoon by compositing six occurrences of each during the period 1965-73. Strong monsoons are characterized by an anomalous strong cyclonic circulation at low levels centered near 20°N over India and stretching eastward over to Myanmar and Indo-Chine. Break monsoons are characterized by an anticyclonic low-level circulation centered slightly further south and also elongated in an east-west direction. During active periods a cloud mass is present in the monsoon trough over central India and during break periods this northern cloud mass weakens and more cloudiness appears near the equator and over the southern tip of India. Pant (1983) described active and break phases of the monsoon and suggested a mechanism involving eddy dynamical processes to explain the slow northward propagation of anomalies associated with active and break phases of the monsoon.

5. The revival of the monsoon from a break condition perhaps represents the reorganization or restoration of the conductive synoptic situations. There are two distinct modes of transition from a break to an active phase. In the first, revival from the break condition occurs with the formation of a disturbance in the northern part of the Bay of Bengal and its subsequent westward movement over the monsoon zone. The second mode of revival involves northward movement of the oceanic tropical convergence zone on to the subcontinent.

SECTION-7 : INTENSITY AND MOVEMENT OF TROPICAL CYCLONE

TASK-30 : INTENSIFICATION OF TROPICAL CYCLONE

Introduction

1. Dvorak technique makes use of a climatological model of development of TC in which the intensity of any vortex in the tropical oceanic areas is defined in terms of T-Nos (T stands for tropical). Intensity according to this model increases at the rate of one T-no during 24 hours. The checklists are placed at the end of this annexure along with some examples.

2. The intensity is computed on the basis of the changes on the organisation of the convection in the central area of the disturbance defining central feature (CF) and in the bands surrounding the central area that determine the band feature (BF). The T-No ranges between T1 to T8. While T1 and T2 refer to the pre-cyclone stage, T2.5 to T3.5 patterns identify marginal cyclones with sustained winds between 35-55 knots. This includes cyclonic storms and Severe Cyclonic storm stages. T4 onwards is associated with Very Severe Cyclonic storm, whereas T6.5 and above are called Super Cyclones.

3. Four basic patterns types are used which represent a range of intensities. The measurements made to assign the intensity depend on the pattern. The “curved band” pattern portrays information on intensity according to the length of the deep convective cloud band spiraling around the circulation center as already described. With the “shear” pattern, the deep convective cloud is confined to one side of the low-level center as depicted by the shallow cumulus cloud lines. The appearance and location of that center relative to the deep convection is used to assign the intensity. This is the only type of cyclogenesis during the southwest monsoon season over the Indian Sub-continent. Since the upper air easterlies continue to be strong even in the month of October over the Bay of Bengal, the shear cyclogenesis is very common and quite often (90% cases) terminates after the tropical disturbances acquire the stage of marginal cyclones (Kalsi and Jain, 1889).

4. For more intense TCs greater than minimal hurricane intensity (32 m/s), the “Central Dense Overcast (CDO)” or “Eye” patterns are typically observed. The Dvorak method uses a different approach to assign intensity with the “CDO” and “Eye” pattern, depending upon whether VIS of IR images are used. As intensity increases the CDO becomes more circular and colder. In addition the eye becomes more distinct, more centrally located within the CDO and warmer as intensity increases.

Step 1. Locate the Cloud System Centre (CSC)

5. The cloud system centre is defined as the focal point of all the curved lines or bands of the cloud system. It can also be thought of as the point toward which the curved lines merge or spiral.

6. **Procedure**:

   a. The CSC is located at the centre of the eye or at the centre of curvature of a partial eye wall when one of these features is observed.

b. When the CSC is not obvious, locate the model expected CSC. Draw a line along the "curved band axis" through the most dense (coldest) portion of the band. The axis should roughly parallel the concave (inner) over-cast boundary of the band. Locate the model expected centre location in relation the curved band. (See plus symbols in diagram in Step 2A.) The centre is located near the inner (concave) edge of the band on the counterclockwise end (comma head) portion of the band. Locate tightly curved lines, merging lines, or CDO near the point where the centre is expected to fall. The CSC is located at the centre of curvature, near the point of mergence or at the centre of the CDO (for CDO of ≤1½° latitude in size). For large CDO's, the centre is sometimes defined by an arc of overshooting cloud tops or in an isolated cluster of convective tops. When not visible, use (c) below.

c. When features are not visible at the expected CSC, or when the curved band is not apparent, use the circle method. The method consists of first drawing lines following the cloud line curvature or curved boundaries that fall within the curve of the curved band axis, and then fitting circles to the lines with tightest curvature. The CSC is located at the centre of the area common to the circles. For relatively circular embedded centre patterns of >T3.5 intensity, fit a log 10° spiral overlay to the curved band axis to locate centre.

d. When a cloud minimum wedge is visible on the concave side of the band near its middle, the CSC is located at the midpoint of a line drawn between the deepest cloud minimum incursion of the wedge and the counterclockwise extremity of the curved band axis. This method is frequently used with EIR pictures. In EIR pictures, the centre is often located in the tight gradient near the coldest part of the pattern.

e. When the location of the CSC is unclear, or could be placed at different locations, use all the methods above along with an extrapolation from the past track positions in making the final decision.

f. When more than one well-defined CSC is apparent, use the one defined by the strongest appearing, lowest level cloud lines that best fits the past track of the storm. When strong vertical shear is apparent, remember that the upper level (dense) clouds will not be centered directly over the low-level centre, but will be displaced with the CSC on the tight temperature gradient (sharp boundary) side of the dense cloud pattern.

Step 1A. Initial Development

7. The earliest signs of tropical cyclone development are observed about 1½ days before a disturbance reaches tropical storm strength. At this time, the disturbance is classified a T1. A T1 is first used when a cluster of deep layer convective clouds showing line or band curvature has the following three properties:-

a. It has persisted for 12 hours or more.

b. It has a cloud system centre defined within an area having a diameter of 2½° latitude or less which has persisted for 6 hours.

c. It has an area of dense, cold (DG or colder) overcast* of >1½° in extent that appears less than 2° from the centre. The overcast may also appear in cumulonimbus lines the curve around the centre. The cloud system centre will be defined in one of the following ways:-

(1) Curved band, a dense (DG or colder) overcast band that shows some curvature around a relatively warm (cloud minimum) area. It should curve at least one-fifth the distance around a 10° log spiral. Cirrus, when visible, will indicate anticyclonic shear across the expected CSC. (See diagrams, Step 6, PT 1.5 pattern types)

(2) Curved cirrus lines indicating a centre of curvature within or near a dense, cold (DG or colder) overcast. (See Figure 4, Step 6, PT 1.5 (b))

(3) Curved low cloud lines showing a centre of curvature within 2° of a cold (DG or colder) cloud mass. (See diagrams, Step 2B, DT 1.5 pattern.) In many cloud clusters that eventually develop, the northern boundaries show a straightening about 1½ days prior to the T1 classifications. During the organising stage of the T1 pattern, there may be extreme variability in the cloud pattern. In most developments at the T1 stage, strong upper-level horizontal anticyclonic shear will be indicated across the disturbance centre when curved cirrus lines are present to reveal the shear. These upper level clouds may indicate patterns far more advanced than T1 at the time of the initial classification. These patterns do not involve deep tropospheric circulations at this time and will be short lived. This means that the Day-2 data T-number may at times be less the Day-1's, but still development is indicated as long as the DT is 2 or more. There may also be times during the first two days of development when cirrus or convective clouds are almost absent, showing little pattern during the nighttime hours. This usually does not mean the storm is weakening. The rule is to never lower the T-number at night during the first 24 hours of development. A flat boundary rotating clockwise across the north side of the pattern throughout the period is a good sign of development. Note that a classification of T1 forecasts tropical storm

* The amount of cold overcast may decrease during the subsequent nighttime hours making it crucial that the analyst watch for the required amount of overcast when it occurs. Intensity (T2.5) 36 hours after the T1 observation only when the environment is expected to remain favorable. A minus symbol is used after the T1 to indicate a T1 pattern that is not expected to develop. (See step 11)

Step 2. Determine the Pattern Type that best describes your Disturbance and Measure Cloud Features as Indicated

8. The manner in which the cloud system centre is defined determines the pattern to be analysed. The pattern types listed below are described on the following pages. When the cloud pattern being analysed does not resemble one of the patterns, proceed to Step 3.

a. Step 2A. "Curved Band" Pattern.

b. Step 2B. "Shear" Pattern.

c. Step 2C. "Eye" Pattern.

d. Step 2D. Central Dense Overcast (CDO) Pattern.

e. Step 2E. Embedded Centre Pattern.

9. **General Analysis Rules**:

a. When short-interval pictures are available, use the average measurement of all of the pictures with well-defined features taken within the 3-hour period ending at analysis time.

b. When two or more T-number estimates are made from the same picture, use the estimate closest to the MET.

c. When in doubt concerning ambiguous features, bias the analysis toward the MET.

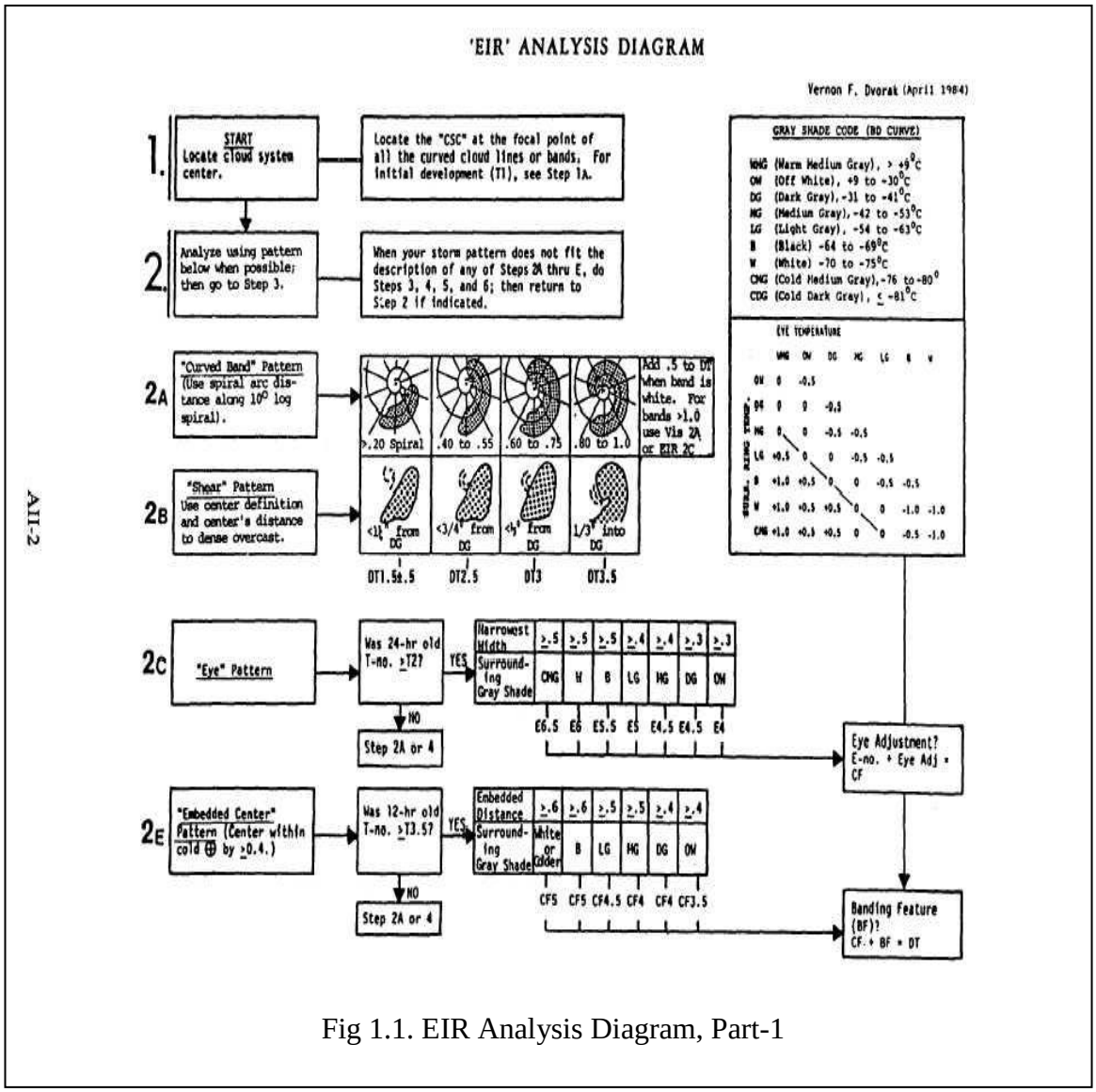


Fig 1.1. EIR Analysis Diagram, Part-1

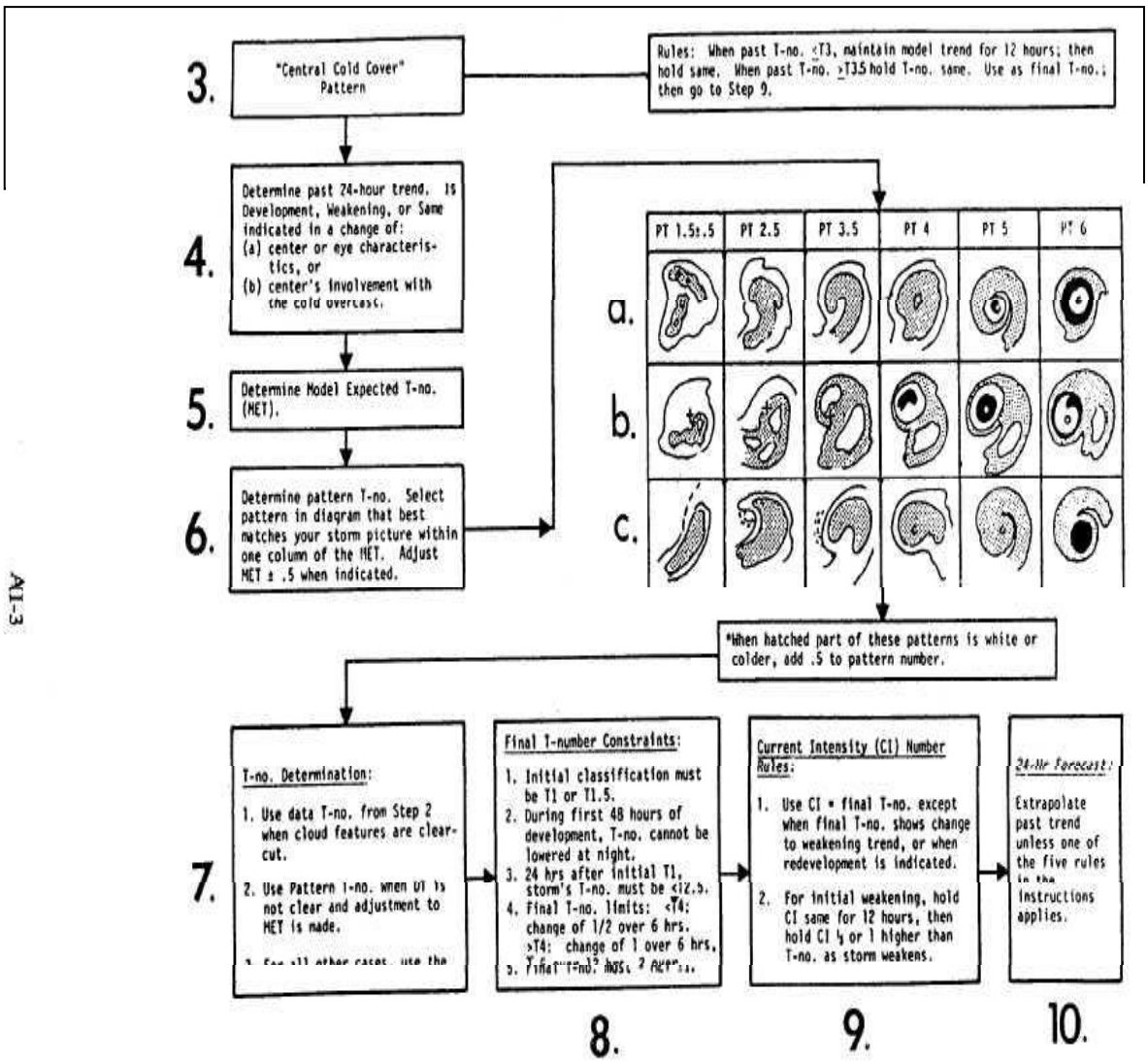


Fig 1.2. EIR Analysis Diagram, Part 2

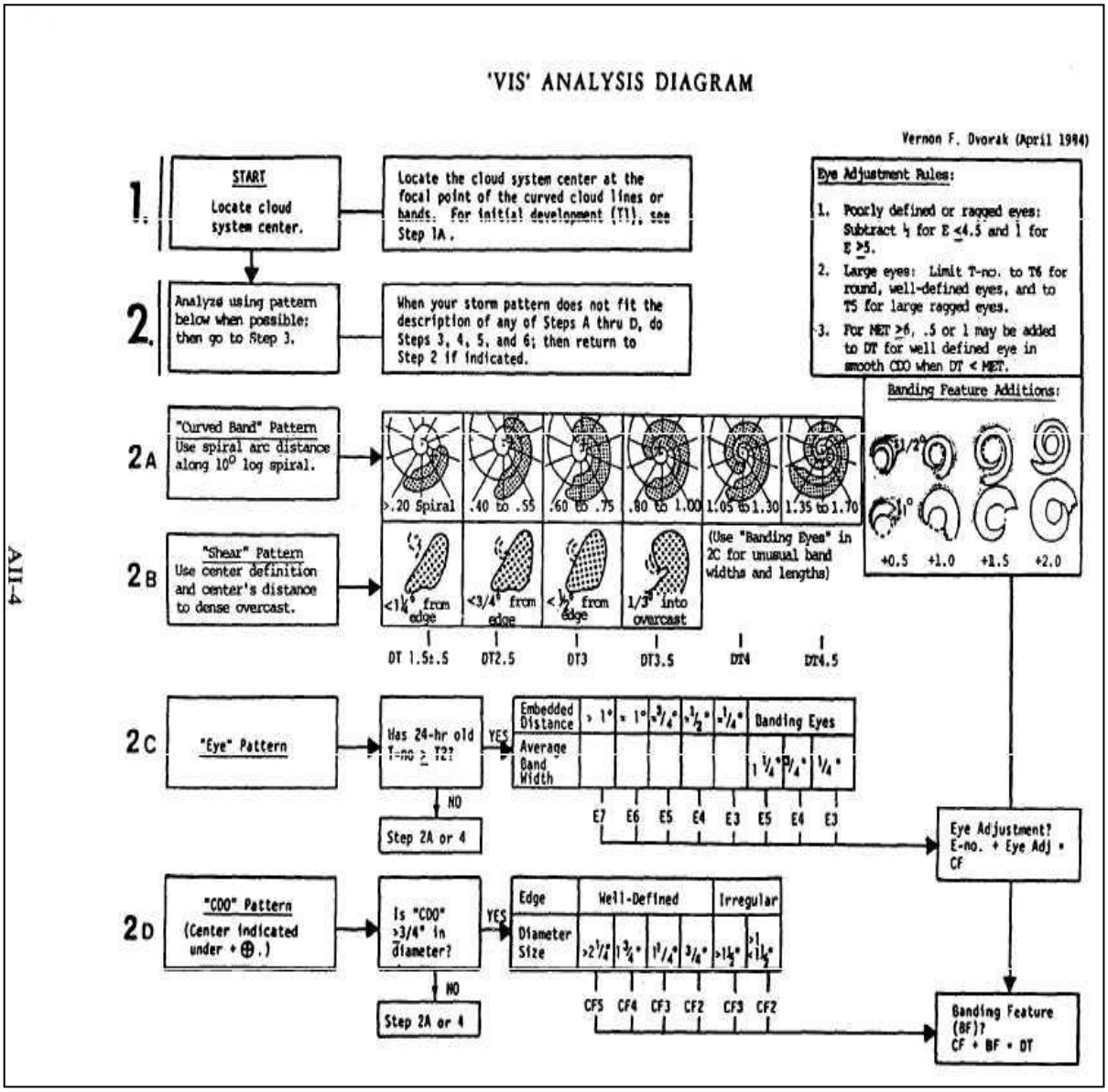


Fig 1.3. VIS Analysis Diagram, Part 1

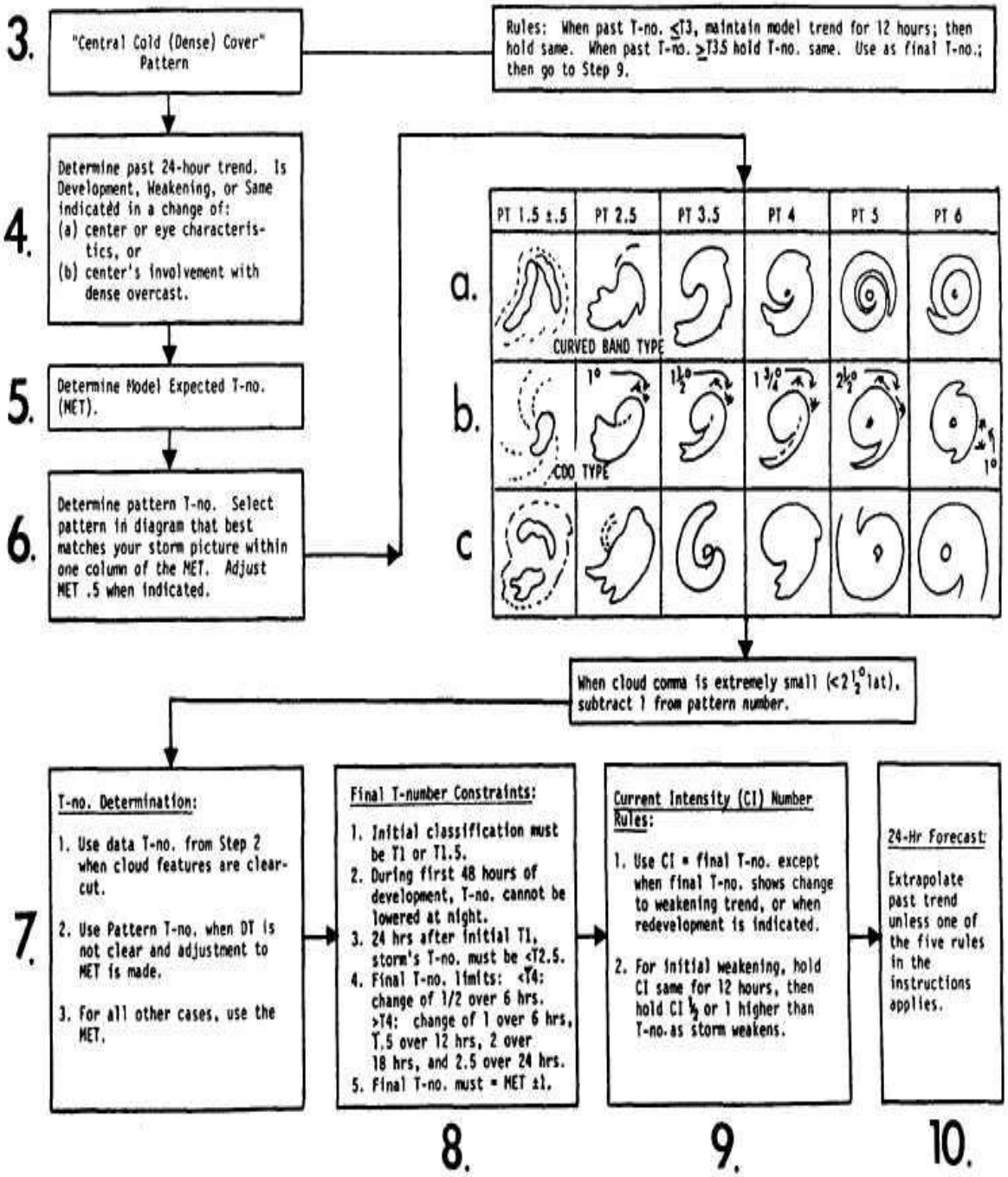


Fig 1.4. VIS Analysis Diagram, Part 2

AII-5

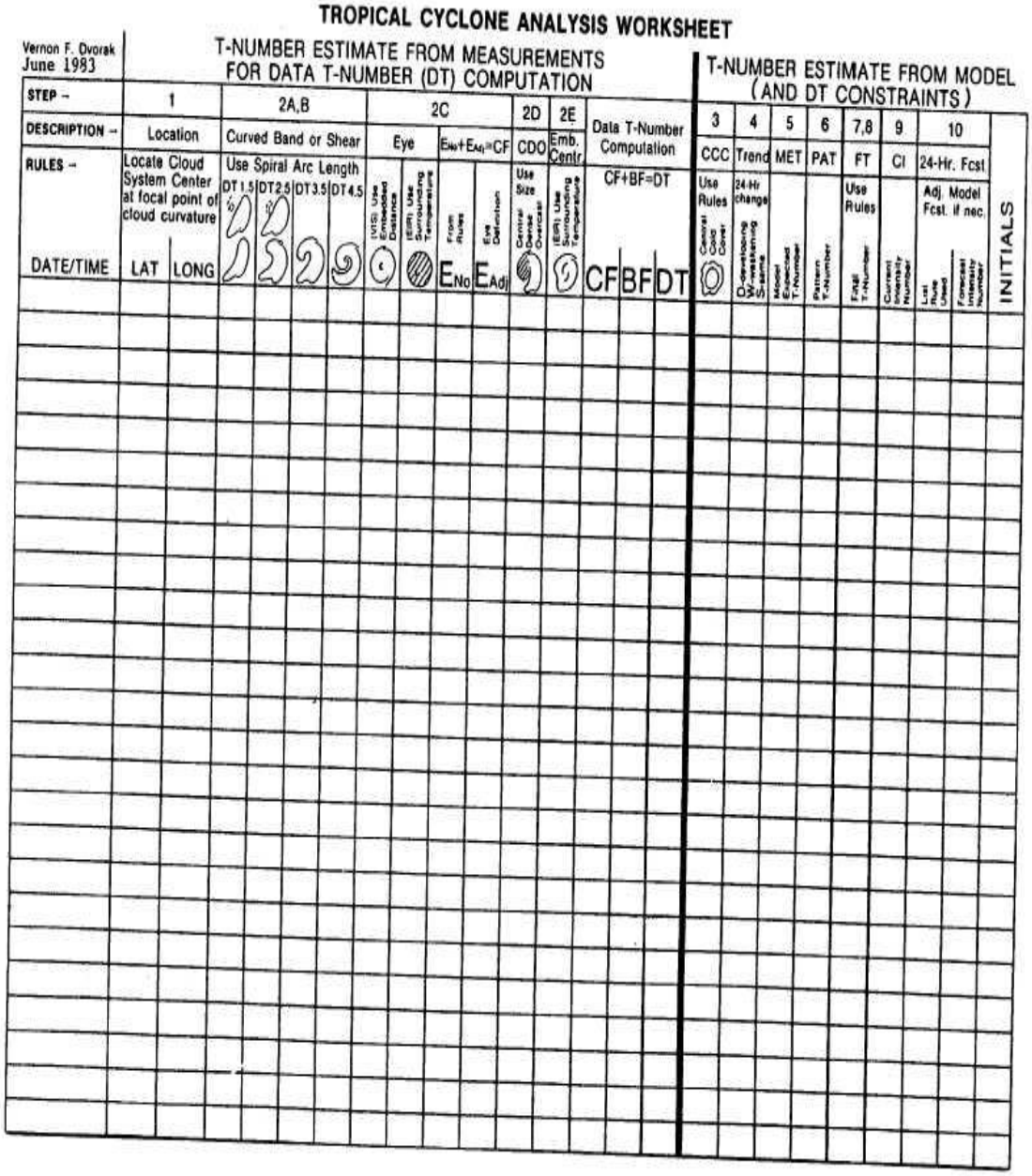


Fig1.5. Analysis Worksheet

Step 2A. Curved Band Pattern

10. The intensity estimate determined from this pattern type is derived by measuring the arc length of the curved band fitted to a 10° logarithm spiral overlay. (A circle will give the same answer most of the time.) The intensity values that relate to the curved band length are given in the analysis diagrams, figure 1, 3. Curved band measurements may be used with both VIS and EIR pictures until an intensity of DT 4.5 is reached. For EIR patterns greater than DT 3.5 use measurements from VIS diagram.

11. The spiral overlay is fitted to the curvature of the dense (cold) band by first drawing a line along the "curved band axis" and then fitting the spiral curve to the line drawn. The curved band axis is defined as the axis of the coldest overcast gray shade (most dense clouds) within the cloud band. The line should roughly parallel the overcast edge on the concave side of the band. When the band indicates two possible axes, use the one with tightest curvature. Cellular cold globs that do not fall in line with the curve of the comma band are ignored when drawing the line. Fit the spiral to the line drawn on the picture and measure the spiral arc length of the dense (cold) band that follows the spiral curve.

12. In EIR patterns (like those in figure 2, Step 6, Row b), the cold comma band will often show warm breaks through its middle. These breaks will appear to be almost clear in the VIS picture, When this occurs, draw the comma axis as though it were continuous through the breaks paralleling the edge of the cloud minimum incursion into the concave side of the band. As the curved band pattern evolves it will usually be defined by the dark gray shade of the BD curve, but may at times appear defined in warmer or colder shades of gray. At times the boundaries of the band must be interpreted from its form in previous pictures.

13. During the first 2 days of development (T1 to T2), the amount of overall band curvature may change excessively, very little, or even decrease somewhat for short periods even though typical development is occurring. For this reason, the tendency should be to raise the T-number by one during the first 24 hours of development as long as the band remains curved enough for T2 and clear signs of weakening or rapid development are not apparent. It is also important to allow at least 24 hours to pass between a T2 and a T4 classification. Even though the coiling process has been observed to be faster than this at times, the surface pressure does not fall accordingly.

14. During the T2.5 or T3 stage, a tightly curved band < 1¼° diameter of curvature observed within the curve of the broad curved band can also be, used as an indicator of tropical storm intensity. This is evidence that the wall cloud is forming. This tight curvature at weak tropical storm intensity is often ragged in appearance but will have deep-layer convective cloudiness on nearly opposite sides of a system centre.

Step 2B. Shear Patterns

15. Shear patterns appear in pre-hurricane stages of development when vertical shear prevents the cold clouds from bending around the cloud system centre as they do in the curved band patterns. The pattern may also appear after the hurricane stage has weakened to a pre-hurricane pattern because of increasing vertical shear.

16. The intensity estimate determined from this pattern type is derived by (a) the way in which the cloud system centre is defined and (b) the distance between the low cloud centre and the dense, cold overcast. For shear patterns associated with tropical storm intensity (T2.5 to T3.5), the centre will be defined by parallel, circularly curved low cloud lines with a diameter of about 1.5° latitude or less. They indicate a centre either near the edge or under the edge of a dense, cold (DG or colder) overcast cloud mass (see patterns in Step 2B, Figures 1, 3. During the weaker stages of development (T1.5 + 0.5), the low cloud centre will either be poorly defined in spiral lines within 1.25° of the cold overcast, circularly defined but some distance (>1.25° latitude) from the cold overcast clouds, or circularly defined near a small amount (<1½° diameter) of dense overcast.

Step 2C. Eye Pattern

17. Eye patterns are analysed in this step only when the eye falls near the point of the expected cloud system centre, and after a T2 or greater pattern has been observed 24 hours prior to the current observation.

18. The eye is defined as one of the following:

a. A warm (dark) spot in a dense, cold (OW or colder) overcast. (When more than one dark spot appears near the CSC, use the centre closest to the expected centre location.)

b. A point in a dense, cold (OW or colder) overcast centered within the curvature of a colder (denser) band that curves at least halfway around the point with a diameter of curvature of 1½° latitude or less.

c. A spiral band wrapped around a relative warm (dark) spot with a diameter of curvature of 1½° latitude or less. The band must curve at least 1.0 the distance around the 10° log spiral curve. (See pattern labeled DT 4 in figure 3 & 2A).

19. The analysis of the eye pattern involves three computations: The eye number (E), the eye adjustment factor (Eye Adj), and the banding feature (BF) Eye Adj. number. The equation is: CF + BF = DT {data T-number), where CF = E No. + 1. EIR only (See figure 2 for VIS)

a. **E (Eye) Number**. To get the E or eye number, first determine the coldest gray shade that surrounds the relatively warm spot. Make certain that the minimum width of this gray shade meets the "narrowest width" requirement shown in the diagram. When a spiral eye is defined, use the average width of the spiral band to determine the narrowest width criteria.

b. **Eye Adjustment Factor**. The eye adjustment factor is determined by using the graph in Figure 6. The graph is a plot of eye temperatures versus the temperature of the coldest ring or spiral that completely encircles the eye. This provides an adjustment of ± 0.5, ± 1, or 0 to the "E" number. No plus adjustment can be made for large eyes (≥ ¾° diameter within the surrounding gray shade) or elongated eyes. When no previous subtraction was made, 0.5 is subtracted for elongated eyes having E numbers of > 4.5. Elongated eyes are defined as those having a short axis of <⅔ the long axis within the surrounding gray shade.

c. **Banding Feature (BF).** The BF addition is used with EIR pictures only when the T-number estimates without the BF is lower than the model expected T-number. It is defined only for patterns of CF4 or more that contain a clear-cut comma tail band that:-

(1) Curves ¼ or more of the distance around the central features or comma head,

(2) Is cold (MG or colder), and

(3) Has a warm wedge (DG or warmer) between the tail and the central features that cuts at least halfway through the pattern for patterns a and b, Figure 7, and at least ⅔ the way for pattern c.

VIS only (See 1 above for EIR)

20. The E (eye) number is obtained by measuring the distance the eye is embedded in dense overcast clouds. The embedded distance of the eye is measured outward from the centre of the eye to the nearest outside edge of the dense overcast for small (< 30nm) eyes. For large eyes, measure outward from the inner wall of the eye. When a banding-type eye is indicated, the arc length of the band around the eye and the average width of the band surrounding the eye are important to the intensity determination, as indicated in the diagram. See analysis diagram (Figure 3, 2C) for the relationships between E-number and embedded distance (eye in CDO), and for band width (banding eye).

SURR. RING TEMP.	EYE TEMPERATURE						
	WMG	OW	DG	MG	LG	B	W
OW	0	-0.5					
DG	0	0	-0.5				
MG	0	0	-0.5	-0.5			
LG	+0.5	0	0	-0.5	-0.5		
B	+1.0	+0.5	0	0	-0.5	-0.5	
W	+1.0	+0.5	+0.5	0	0	-1.0	-1.0
CMG	+1.0	+0.5	+0.5	0	0	-0.5	-1.0

Fig 1.6. Eye Adjustment Graph. Rules: (1) For large or elongated eyes, use values to the right of the diagonal line only; (2) for elongated eye patterns >4.5, subtract .5 when no other subtraction was made

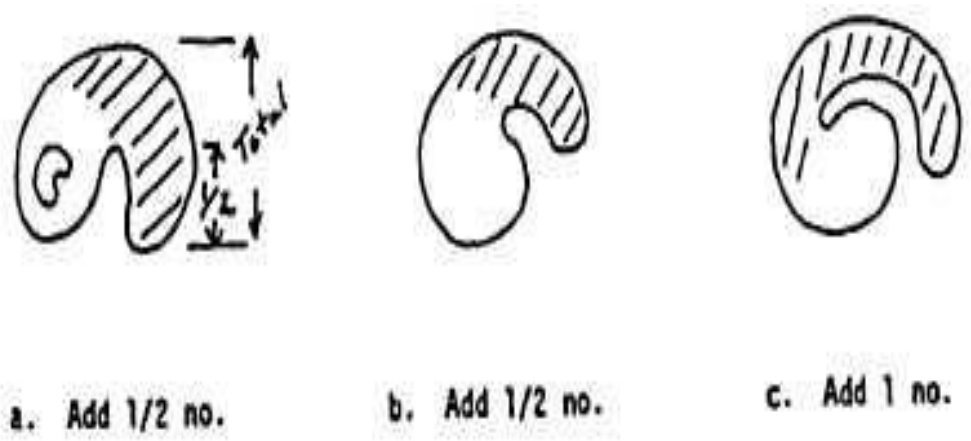


Fig 1.7. EIR Banding Features. Add to the CF only when the DT is lower than the MET

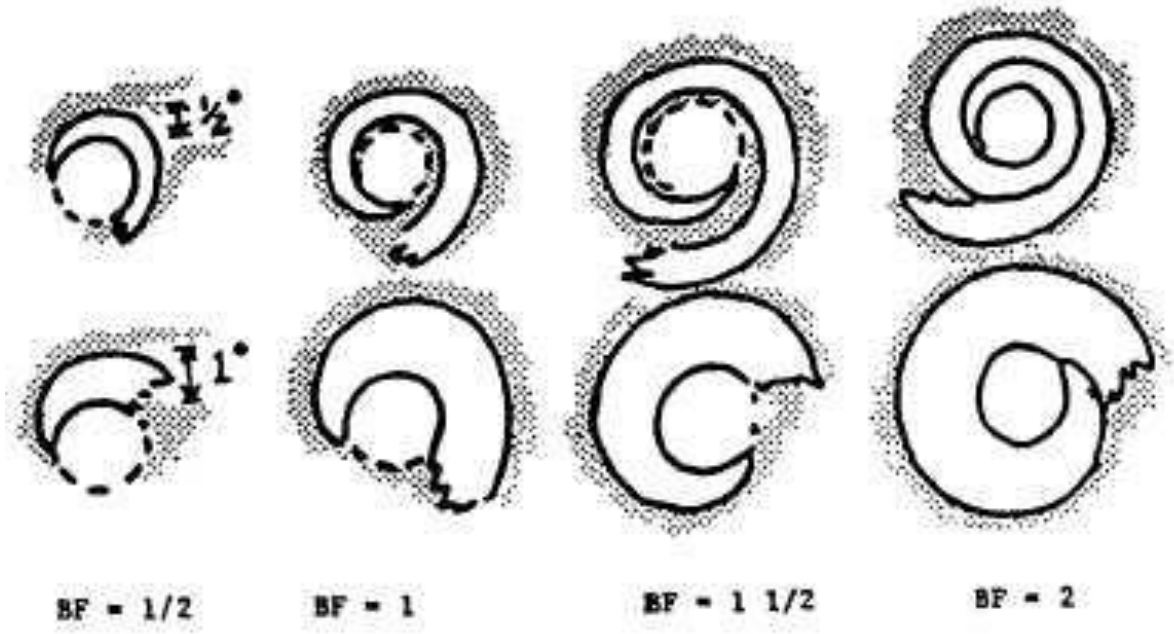


Fig 1.8. VIS Banding Features

21. The eye adjustment factor is determined by the definition, shape, and size of the eye. The eye is well defined by either its blackness or by a well-defined boundary. To be well defined, the eye should be dark or black. Remember that a very high or very low sun angle may reduce the eye definition unrealistically, and that high-resolution pictures may show a poorly defined eye that would not appear in the low-resolution pictures for which the technique was designed. A poorly defined eye is one that is barely visible. A ragged eye is one with a very uneven boundary with little circularity. VIS eye adjustment rules are as follows. (a) For poorly defined or ragged eyes, subtract ½ number for E numbers of < 4.5 and subtract 1 number when E > 5. When analysing patterns with poorly defined eyes especially in high-resolution pictures, also check the CDO size. Use the estimate which is most consistent with the MET. (b) For large eyes, limit the maximum T-number to T6 for round, well-defined eye patterns, and to T5 or lower for all other large-eyed patterns. And, (c) the E-number may also be adjusted upward by either 0.5 or 1.0 when the eye is well-defined, circular and embedded in a very smooth, very dense appearing canopy. The addition is made only when the data T-number is lower than the MET and the storm's past history gives an expected T-number of T-6 or more. The general rule for the eye adjustment factor is: When an adjustment is not clear-cut, use the guidance of the MET to make the final decision.

22. The BF adjustment is often an important factor when VIS pictures are used. It is defined as a dense, mostly overcast band that curves quasi-circularly at least ¼ the distance around the central feature. Bands that curve evenly around an inner BF may also be counted. The amount of the BF term ranges from 0.5 to 2.5. It depends on the width of the band and the amount the band curves evenly around the central features, as shown in figure 8. A BF term is not used for pre-hurricane patterns when the curved dense band concept in Step 2A is used. However, it is still needed for CDO patterns and all hurricane patterns when indicated. For banding eye patterns use the central coil (once around the eye) as the CF and add the BF as indicated. This pattern type is rarely used for DT of greater than 4.5.

Step 2D. CDO Patterns (VIS only)

23. CDO patterns are defined when a dense, solid-looking mass of clouds covers the cloud system centre and lies within the curve of the system's comma band. Both its size and the sharpness of its boundary are important to the analysis. A well-defined CDO has an abrupt edge on at least one side of the cloud mass. An irregular CDO appears within the curve of the comma band but has ragged boundaries and uneven texture. Generally, well-defined CDO's that measure about 1° latitude in their narrowest width are associated with tropical storm intensities while those measuring 2° latitude or more are associated with hurricanes. The size-CF number relationship is given in the analysis diagram, Figure 3. Examples of CDO's are shown in Figure 4, Step 6(b). For CDO patterns, the analysis equation is CF + BF = DT. Banding features (BF) are usually added to the CF term for CDO patterns. The BF's are described above in 2C.

Step 2E. Embedded Centre Patterns (EIR only)

24. Embedded centre patterns are analysed when the storm has had a previous history of a T3.5 or greater intensity and when the CSC is clearly indicated to be within a cold overcast (OW or colder). Curved cloud lines or bands within the cold overcast as well as the outer curved bands will indicate the location of the CSC within the overcast. A 10° logarithmic spiral can often be fitted to the system's pattern to help locate the CSC in patterns of hurricane intensity. (See Step 2A for fitting spiral.)

25. The analysis of this pattern is similar to the eye pattern analysis except that no eye adjustment factor is added. Determine the coldest overcast in which the CSC is embedded the required distance. This yields the central feature number (CF). Then add a banding feature (BF) adjustment when indicated. The equation being CF + BF = DT.

Step 3. Central Cold Cover (CCC) Pattern

26. The CCC pattern is defined when a more or less round, cold overcast mass of clouds covers the storm centre or comma head obscuring the expected signs of pattern evolution. The outer curved bands and lines usually weaken with the onset of CCC. When using VIS pictures, substitute the word "dense" for "cold." It is only rarely that the CCC pattern is used with VIS pictures since the CDO or curved lines are usually visible through the thin cirrus clouds. When the CCC persists (see rules in diagram, Step 3), development has been arrested until signs of development or weakening once again appear in the cloud features. Care should be exercised under the following conditions:-

a. Do not confuse a CCC pattern with a very cold comma pattern. A very cold (usually white) pattern is indicated by a very cold (very smooth texture) comma tail and head with some indication of a wedge in between. Curved cirrus lines or boundaries usually appear around the cold pattern and not around the CCC pattern. The very cold pattern for T-numbers of T3 or less warrant an additional ½ number in intensity estimate and often indicates rapid growth.

b. Do not assume weakening in a CCC pattern when the comma tail begins to decrease in size. It is common to observe the tail decreasing in size at the onset of the CCC. Also the CCC often warms as the eye of the T4 pattern begins to be carved out by a warm incursion into the side of the cold over-cast. This signals the resumption of pattern evolution (intensification) even though some warming is evident.

Step 4. Determine the Trend of the past 24-Hour Intensity Change

27. The trend of the past 24-hour intensity change is determined qualitatively by comparing the cloud features of the current picture with those in the 24-hour old picture of the storm. In general, a disturbance has developed when its centre appears better defined with no change in the relation to the dense clouds of the disturbance or is more involved with dense overcast clouds. More precise definitions for development, weakening or steady state changes are given below.

28. The storm has developed (D). In the Curved Band Pattern, curved band coils farther around the CSC:

a. **CDO Pattern**. CDO becomes larger or an increase in banding features is noted.

b. **Shear Pattern**. CSC becomes more tightly defined in curved cloud lines or appears closer to the dense overcast.

c. **Eye Pattern.** Eye is more embedded, more distinct (warmer), less ragged, or is surrounded by colder (smoother textured) clouds, or more banding features.

d. No significant warming (darkening) of the cloud system is noted. By significant, it is meant that a change that is not diurnal (near sunset), which lasts for more than 3 hours, and is great enough to lower the T-number.) The storm has weakened (W):

29. The storm has weakened when its cloud pattern indicates a persistent trend opposite to those listed in (a)-(e) above. Watch in particular for pat-terns that become sheared out (elongated with time) or for patterns undergoing non-diurnal warming (lowering) of their cloud tops.

30. The storm has become steady state (S):

a. When a central cold cover appears in a T3.5 or greater storm or has persisted for more than 12 hours in a weaker storm.

b. When the CSC's relationship to the cold clouds has not changed significantly.

c. When there are conflicting indications of both development and weakening.

Step 5. The Model Expected T-Number (MET)

31. The MET is determined by using the 24-hour old T-number, the D, S, or W decision in Step 4, and the past amount of intensity change of the storm. When the growth rate has not been established in the case of new developments or reversals in trend, assume a past rate of change of one T-number per day. Equations for determining the MET are given below.

MET = 24-hour old T-number + (0.5 to 1.5) when D was determined.

MET = 24-hour old T-number - (0.5 to 1.5) when M was determined.

MET = 24-hour old T-number when S was determined.

32. Rapid or slow past rates of change are established when two consecutive analyses showing rapid or slow pattern evaluation are observed at 6-hour or more intervals, or when one observation accompanied by signs of strong intensification or weakening is observed (see Step 10).

Step 6. The Pattern T-Number (PT)

33. The pattern T-number is used primarily as an adjustment to the MET when an adjustment is indicated. The PT-number is determined by choosing the pattern that best matches your storm picture from either the model expected T number column or the column on either side of it. When the pattern being analysed looks more like the pattern in the column to the right or left of the MET column, then raise or lower the MET 0.5 to determine the PT.

Step 7. Rules for Determining the T-Number

34. Use the data T-number (DT) when the cloud feature measurements are clear-cut. Use the pattern T-number (PT) when the DT is not clear and the pattern is understandable. When neither the DT or the PT is clear, use the Model Expected T-number (MET).

Step 8. Final T-Number

35. This step provides the constraints within which the final T-number must fall. In other words, when the T-number gotten from Step 7 does not fall within the stated limits, it must be adjusted to the limits. The constraints hold the final T-number change to 1.5 during the first 24 hours of development; to 2 numbers in 24 hours for T-numbers T2 to T4 (i.e. ½ number over a six hour period); and to 2.5 numbers over a 24 hour period for changes in storms of T4 or greater intensity (i.e. 1 number over a six hour period, 1½ numbers in 12 hours, 2 in 18 hours, and 2.5 in 24 hours). In general for storms of hurricane intensity, the final T-number must be within one number of the model expected T-number (MET). The constraints are listed in the diagram. The rules also prohibit the lowering of the T-number at night during the first 48 hours of development because the diurnal changes in clouds often give deceptive indications of weakening at this time.

Step 9. Current Intensity (CI) Number

36. The CI number relates directly to the intensity of the storm. The empirical relationship between the CI number and the storm's wind speed is shown in figure 9.

37. After each intensity analysis, the previous analyses of the storm should be reviewed in the light of the current data. When an error was made in the previous day's analysis, correct the T-number to provide a more-accurate model-expected intensity. The correction may at times alter the current intensity analysis.

CI Number	MWS (Knots)	MSLP (Atlantic)	MSLP (NW Pacific)
1	25 K		
1.5	25 K		
2	30 K	1009 mb	1000 mb
2.5	35 K	1005 mb	997 mb
3	45 K	1000 mb	991 mb
3.5	55 K	994 mb	984 mb
4	65 K	987 mb	976 mb
4.5	77 K	979 mb	966 mb
5	90 K	970 mb	954 mb
5.5	102 K	960 mb	941 mb
6	115 K	948 mb	927 mb
6.5	127 K	935 mb	914 mb
7	140 K	921 mb	898 mb
7.5	155 K	906 mb	879 mb
8	170 K	890 mb	858 mb

Fig 1.9. The empirical relationship between the current intensity number (CI), the maximum mean wind speed (MWS), and the minimum sea level pressure (MSLP) in tropical cyclones. The MSLP values for the NW Pacific

38. The CI number is the same as the T-number during the development stages of a tropical cyclone but is held higher than the T-number while a cyclone is weakening. This is done because a lag is observed between the time a storm pattern indicates weakening has begun and the time when the storm's intensity decreases. In practice, the CI number is not lowered until the T-number has shown weakening for 12 hours or more. The CI number is then held one higher than the T-number as the storm weakens. (Hold the CI number ½number higher when the T-number shows a 24-hour decrease of ½ number.) When redevelopment occurs, the CI number is not lowered even if the T-number is lower than the CI number. In this case, let the CI number remain the same until the T-number increases to the value of the CI number.

Step 10. The 24-Hour Intensity Forecast (FI)

39. The Forecast Intensity (FI) is an extrapolation forward of the past 24 hour change in T-number (not to exceed 1½ T-number per day) unless the cyclone's cloud pattern or its environment indicates a change in one of the following. Remember that the FI number is similar to a forecast CI number in that for a forecast of weakening the FI number is held one or ½ number higher than the forecast T-number.

a. **Step 10A:** Strong unfavourable signs for Future Development within the Cloud Pattern:

(1) Persistent warming of cloud pattern for more than 12 hours even though other features may indicate intensification.

(2) A central cold cover that persists for more than 3 hours.

(3) The storm's convective clouds are becoming involved with a field of stratocumulus clouds in the path of the storm.

(4) The cirrus cloud lines of the storm indicate less curvature because of increasing strong unidirectional flow aloft across the storm.

(5) The cloud pattern is undergoing increasing elongation (deformation) with time.

Rule 10A. When strong unfavourable signs are observed, forecast (either) no further development or reduce the past development rate of storm by half. The changes in the environment listed in 10B should play a role in this judgement.

b. **Step 10B**. Strong unfavourable signs for future development within the environment. The disturbance is entering an unfavourable environment such that the storm will soon become involved with the following:

(1) Stratocumulus clouds

(2) Land

(3) Southward-moving cirrus appearing less than 10° latitude to the north or west of the storm.

(4) Increasing unidirectional flow across the storm pattern.

(5) A southward surge of the westerlies in the upstream environment of a disturbance. The surge is observed as either jet stream cirrus pointing Southward northwest or north of the disturbance or as curved cloud lines or hands bowed toward the disturbance moving southward becoming more convective or remaining convective with time. Watch for areas of increasing convection. The band structure may be very weak at first. The environment is considered unfavourable when a broad-scale cyclonically curved cloud band is within 25° latitude of a westward-moving disturbance. When the clouds of the disturbance indicate signs of upper-level westerly shear the probability of arrested growth is increased.

Rule 10B. When the storm is entering an unfavourable environment, forecast slow development (½T-number per day) for developing disturbances or when signs are strong in both the disturbance (10A) and the environment forecast no development.

c. **Step 10C**. Strong favourable signs for future development within the cloud pattern:

(1) Two successive observations of 24-hour changes that indicate rapid development. (The observations should be at least 6 hours apart).

(2) One observation of rapid development and either (a) a cold pattern (white or colder); (b) cold (dense) comma tail pattern such as those shown in Figure 2, Step 6b; (c) or signs of two or three strong outflow channels. That is, cirrus bands extending some distance out from the disturbance. This is also observed as a fanning out of the cirrus to the south or east of rapidly developing storms.

Rule 10C. When strong favourable signs are observed, forecast rapid development in a developing disturbance of ≤ T5.5 that is not expected to peak or enter one of the environment listed in Step 10B. Never forecast an intensity greater than in T7.

d. **Step 10D**. Favourable signs for future development within the environment.

The disturbance is moving away from the conditions described in step 10B. A disturbance leaving an unfavourable environment as described in 10B (v) is indicated by jet stream cirrus pointing more northward than previously or as curved cloud bands in the broad-scale environment to the north of the storm bowing more away from the disturbance or dissipating in time.

Rule 10D. When a storm is leaving an unfavourable environment and it had weakened as a consequence of the unfavourable environment, forecast rapid development until the storm reaches its previous intensity and then forecast the previous rate of change. If the rate of development had decreased because of the unfavourable environment, forecast an increase to one T-number/day in the rate of development as the storm moves into the more favourable environment.

e. **Step 10E. Signs of "Peaking".** Most storms reach their maximum intensity 4 to 6 days after the T1 classification has been made. The day of peaking often depends on the direction of motion of the storm; northward in 4 days, westward moving in 6 days, and all others in 5 days. The signs of peaking observed in a storm's cloud pattern are a general warming of the cloud tops (less smoothness in texture), or a more or less circular pattern having an absence of peripheral convective clouds or bands.

Rule 10E. If the signs of peaking are observed at or after the time of expected peaking, forecast no change in intensity.

Step 11 (Optional)

40. Encode the intensity estimate using the code in Figure 10 below. The code is self-explanatory except for the PLUS and MINUS (indications of ongoing change). These are used only when the cloud pattern of a disturbance or its environment indicates a change in trend will occur during the succeeding 24-hour period, or when rapid change is forecast. The PLUS is used either to forecast development when a past trend of W or S is indicated in the code or to forecast rapid development when a D was shown. When the past trend as indicated by D, S, or W is expected to continue for the next 24 hours, the space is left blank.

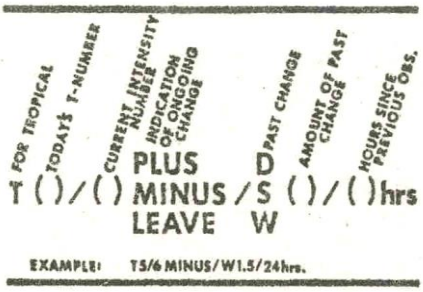
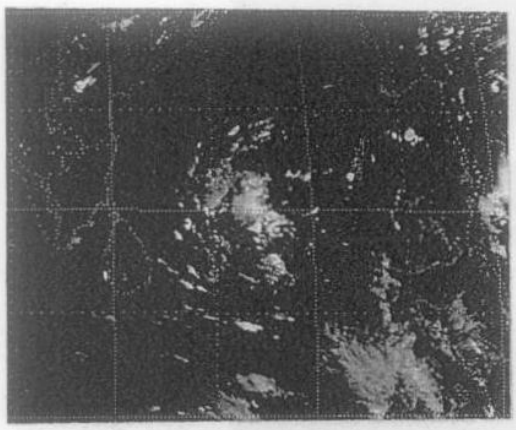
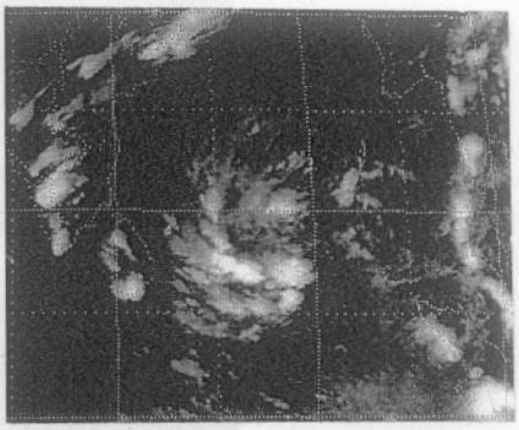


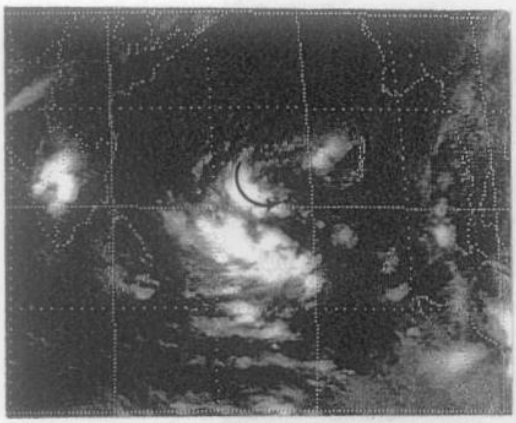
Figure 10. Code to be used for communicating satellite intensity estimations and forecast.



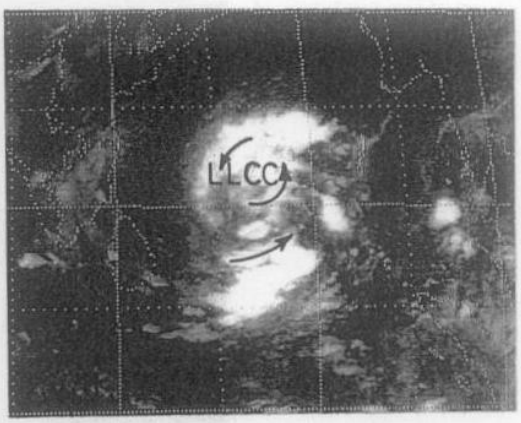
03.05.1990 06 UTC VIS



03.05.1990 12 UTC VIS

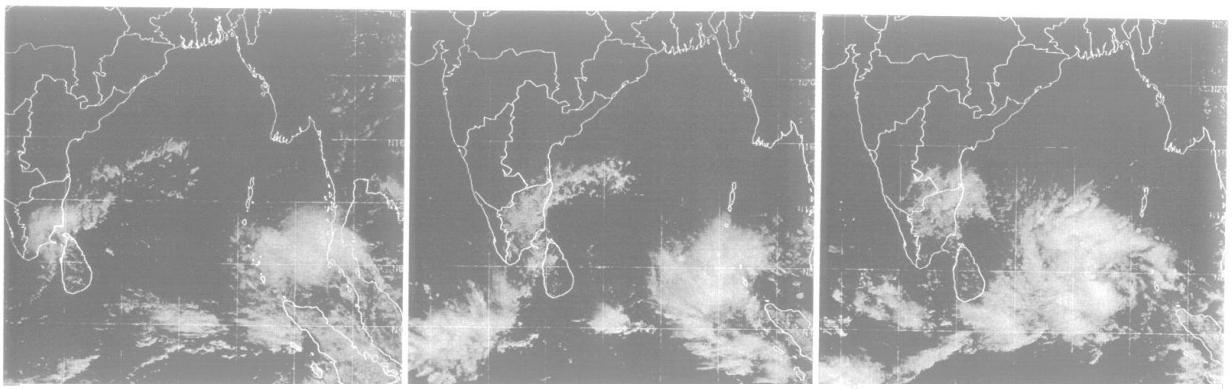


03.05.1990 18 UTC VIS



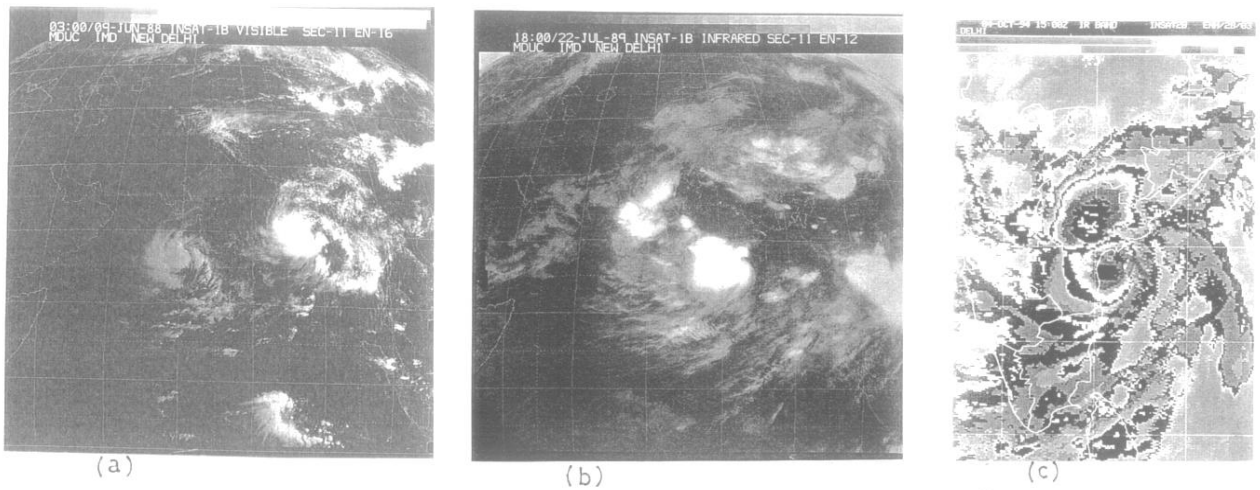
04.05.1990 00 UTC VIS

Pre-depression stage of MPT Super TC on 3 May 1990. There is evidence of LLCC around 11.5°N latitude/87.5°E longitude between 18 UTC/03 May and 00 UTC of 04 May 1990.

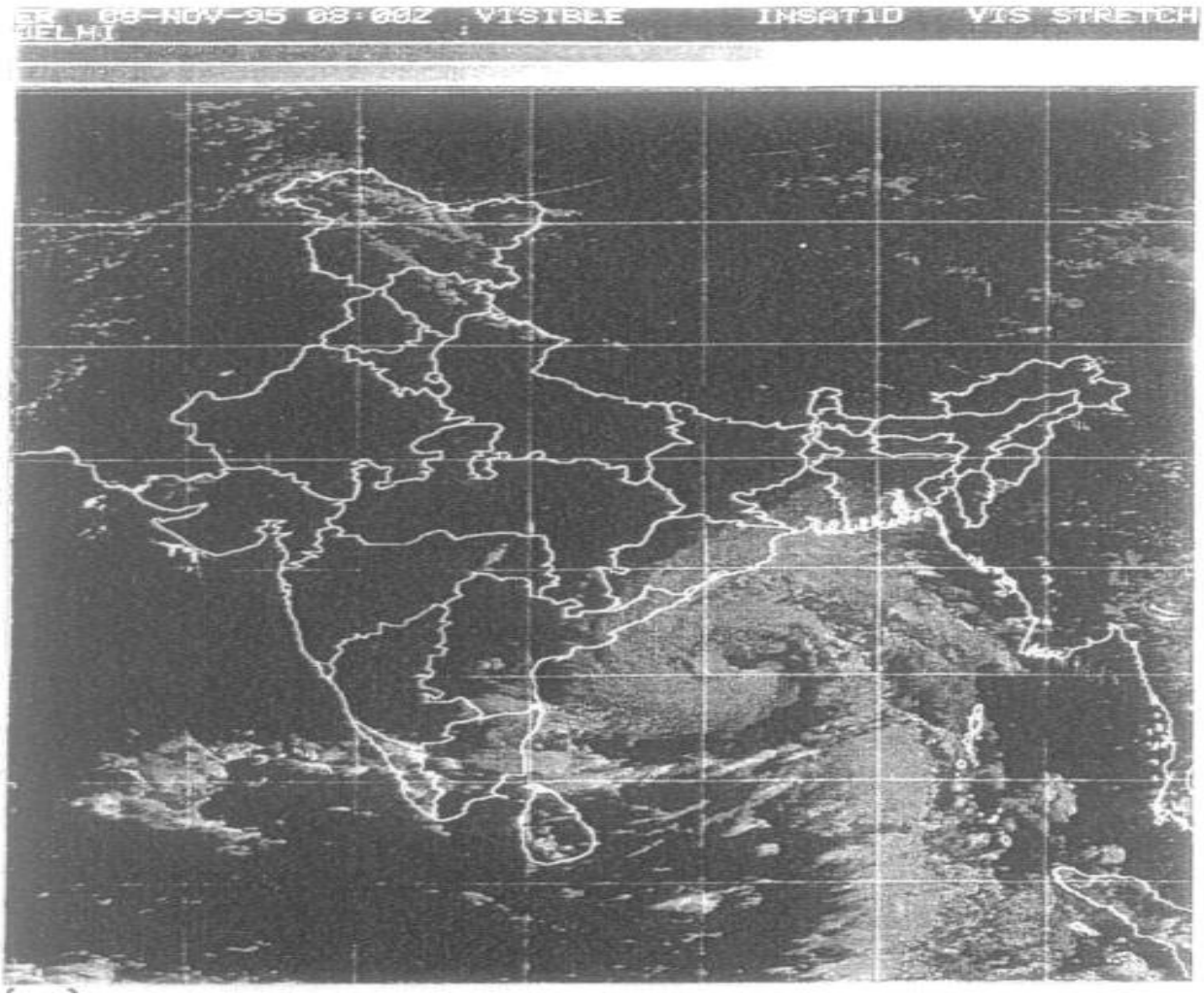


(a) 29.11.1993 06 UTC (b) 30.11.1993 06 UTC (c) 01.12.1993 06 UTC

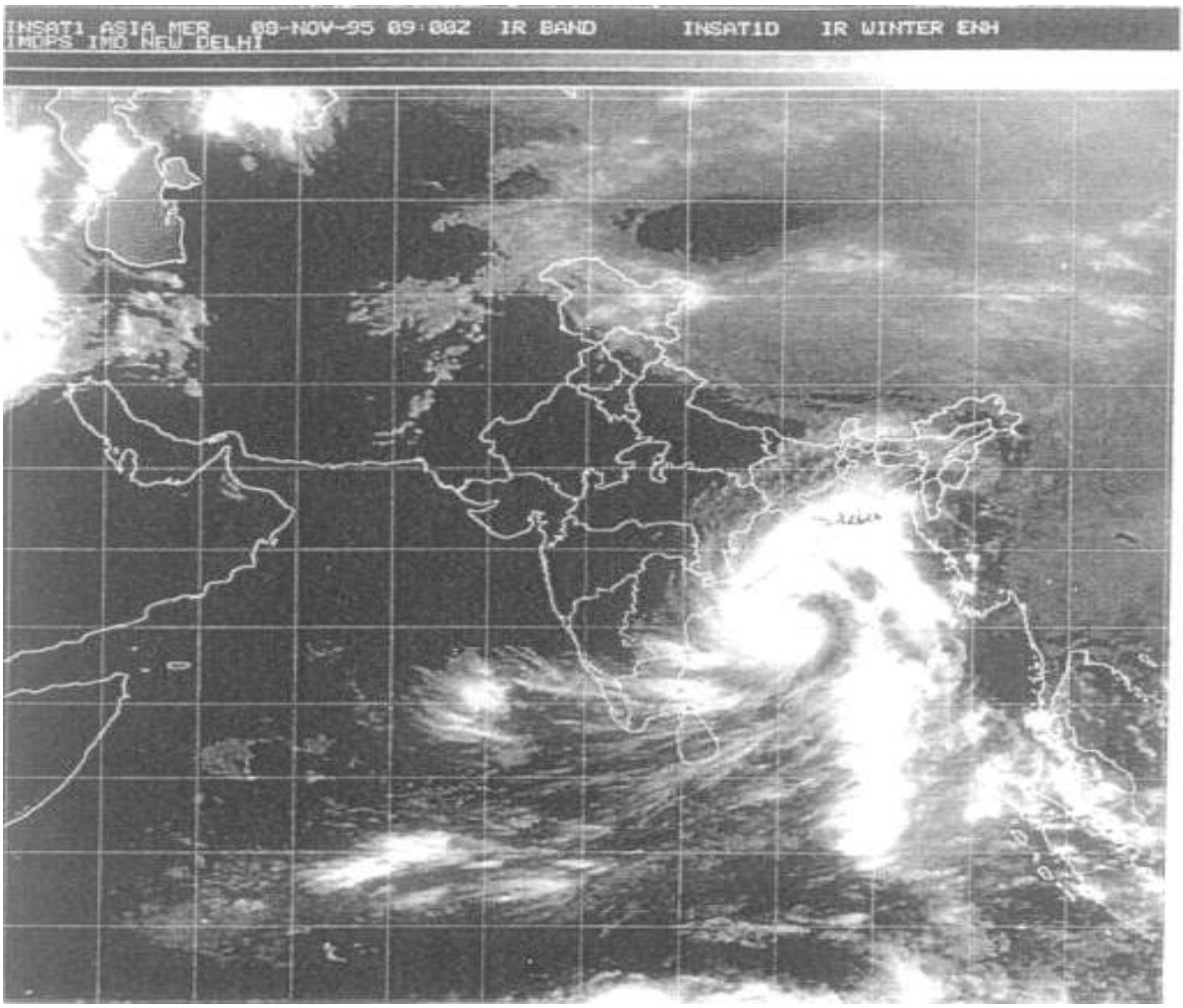
(a) & (b) Transformation of initial innocuous cloud cluster on 29 Nov 1993 that crossed Malay Peninsula and emerged in the Andaman Sea on 30 Nov (c) Depression centre defined by low level lines on 01 Dec 1993



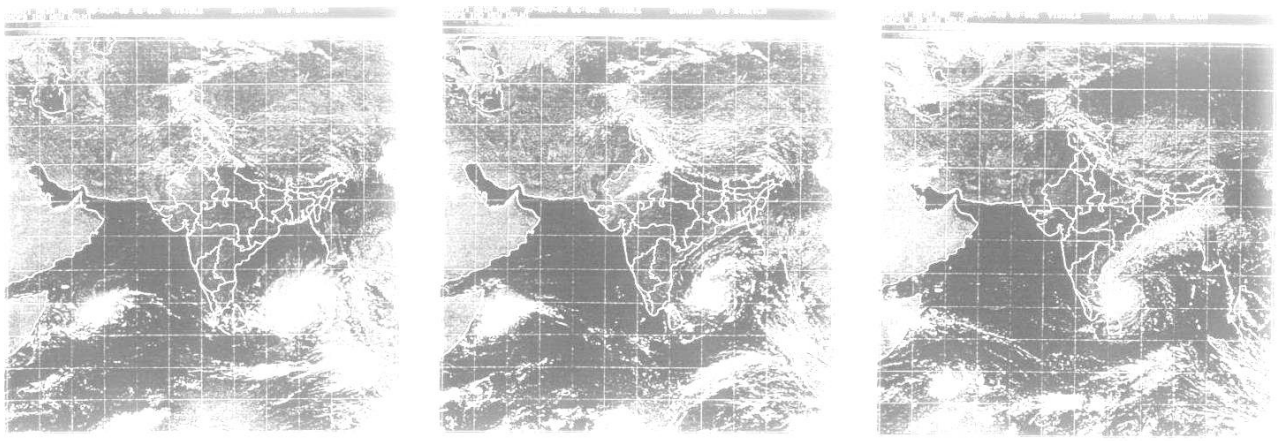
Depression as Marginal Cyclones near the coast. (a) Deep depression on 09 Jun 1988 with LLCC near the eastern edge of deep convection (T2.5). Pressure fall of 14 hPa & departure of minus 15 hPa near the centre indicate marginal TC strength. (b) CSC close to intense cloud cluster over the coast. Pressure fall of 6-9 hPa bur pressure departure of minus 13-15 hPa indicates minimum TC. (c) CSC again close to the eastern edge indicates T2.5 intensity.



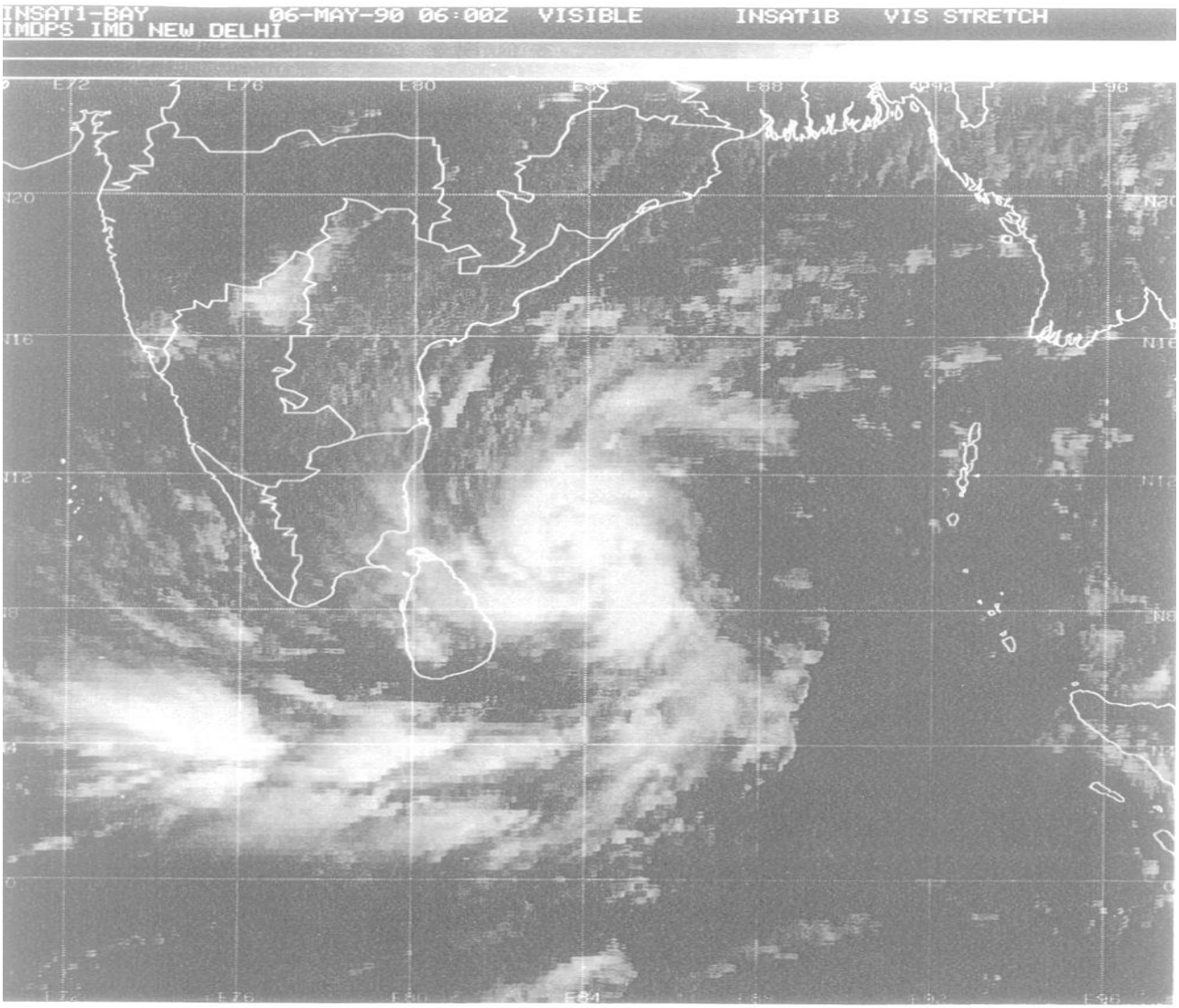
Curved band spiraling more than 3 quadrant (T3.5) on 08 Nov 1995. VIS image showing eye in the dark area enclosed by wall cloud.



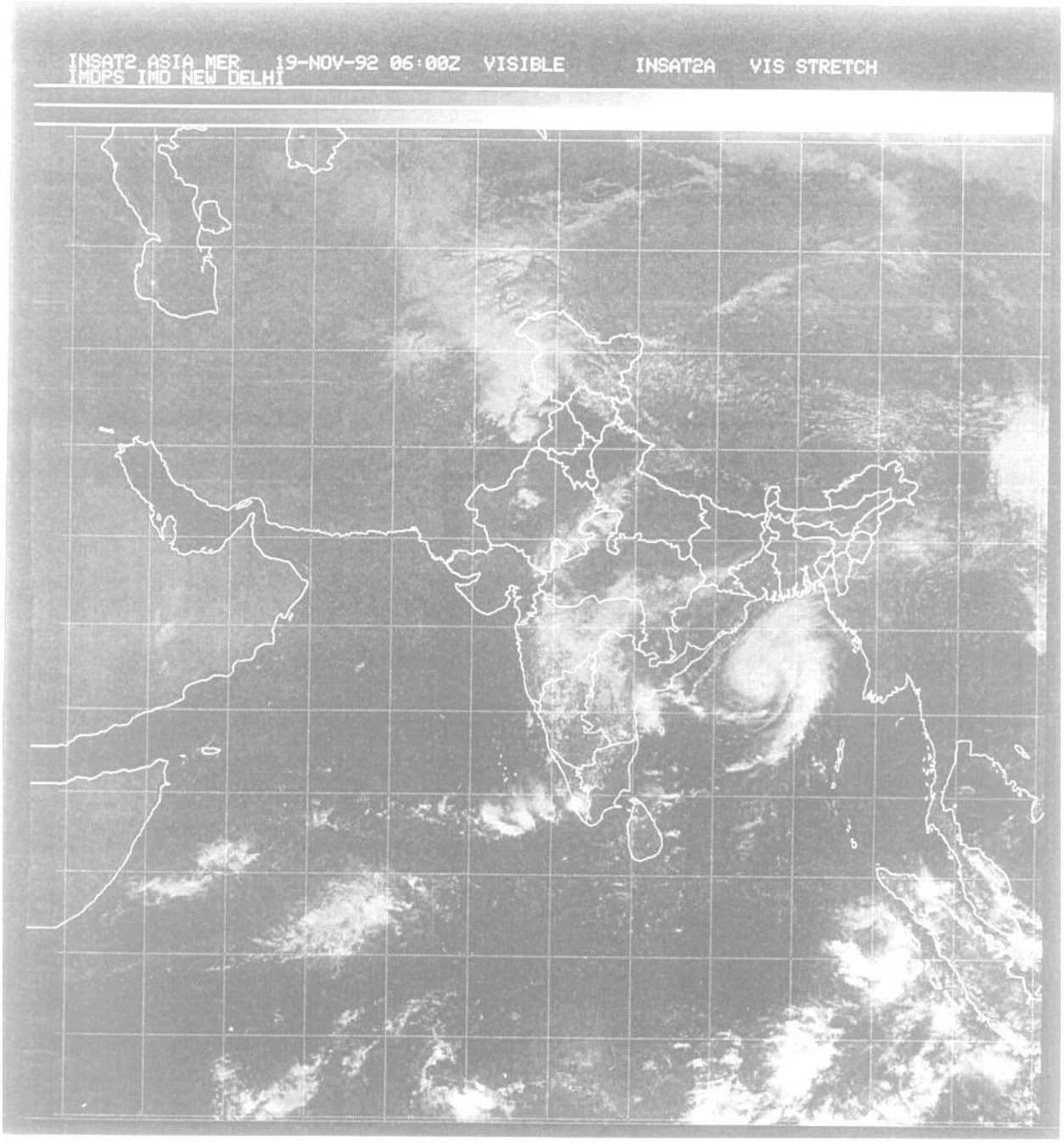
IR image on 08 November 1995 (T3.5) in which central area is filled with cirrus clouds. Moat is seen around the core convection.



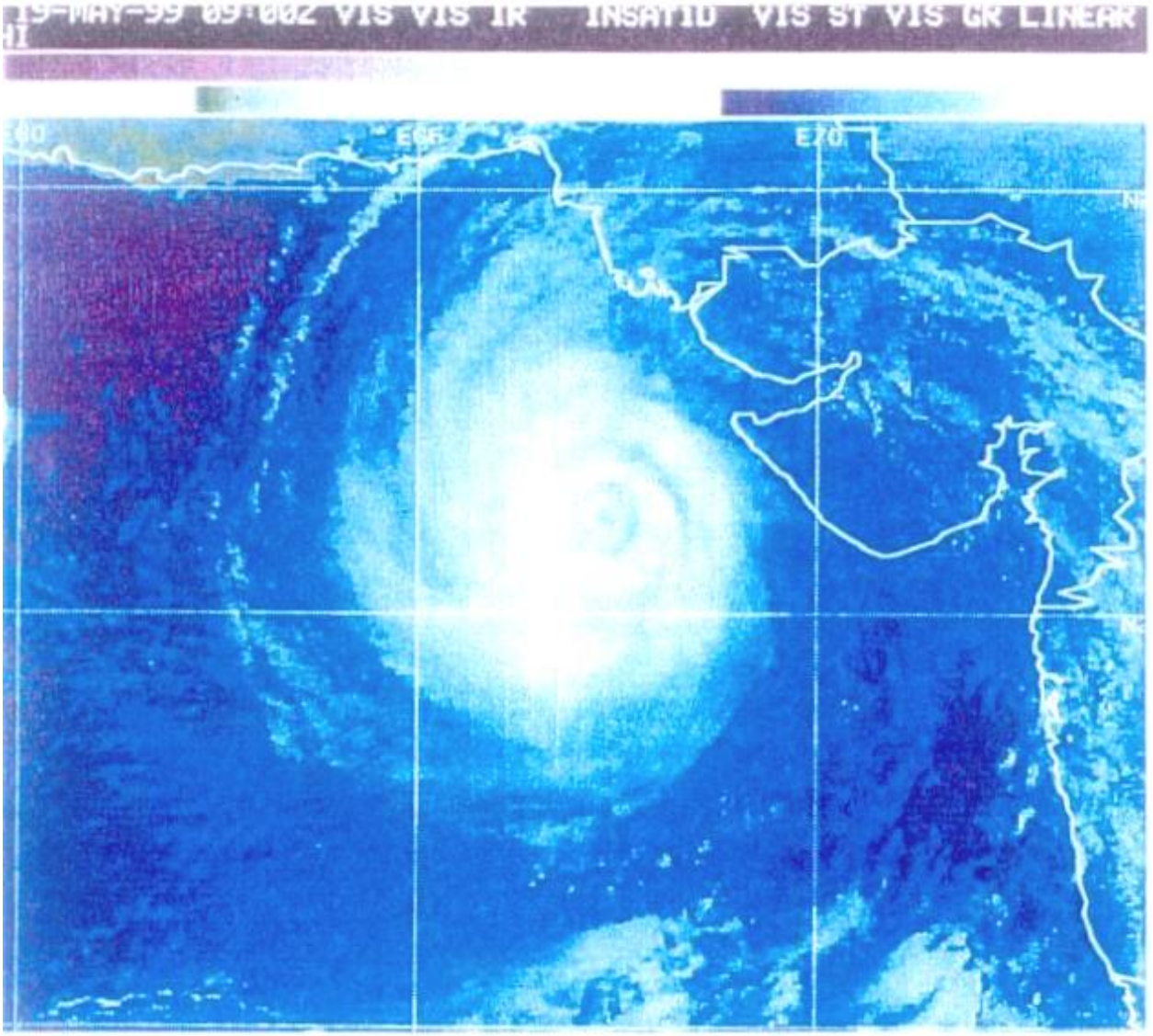
CDO pattern development from 26 Nov 2000 CDO of about 2 deg associated with a severe cyclonic storm on 28 Nov (T4.0)



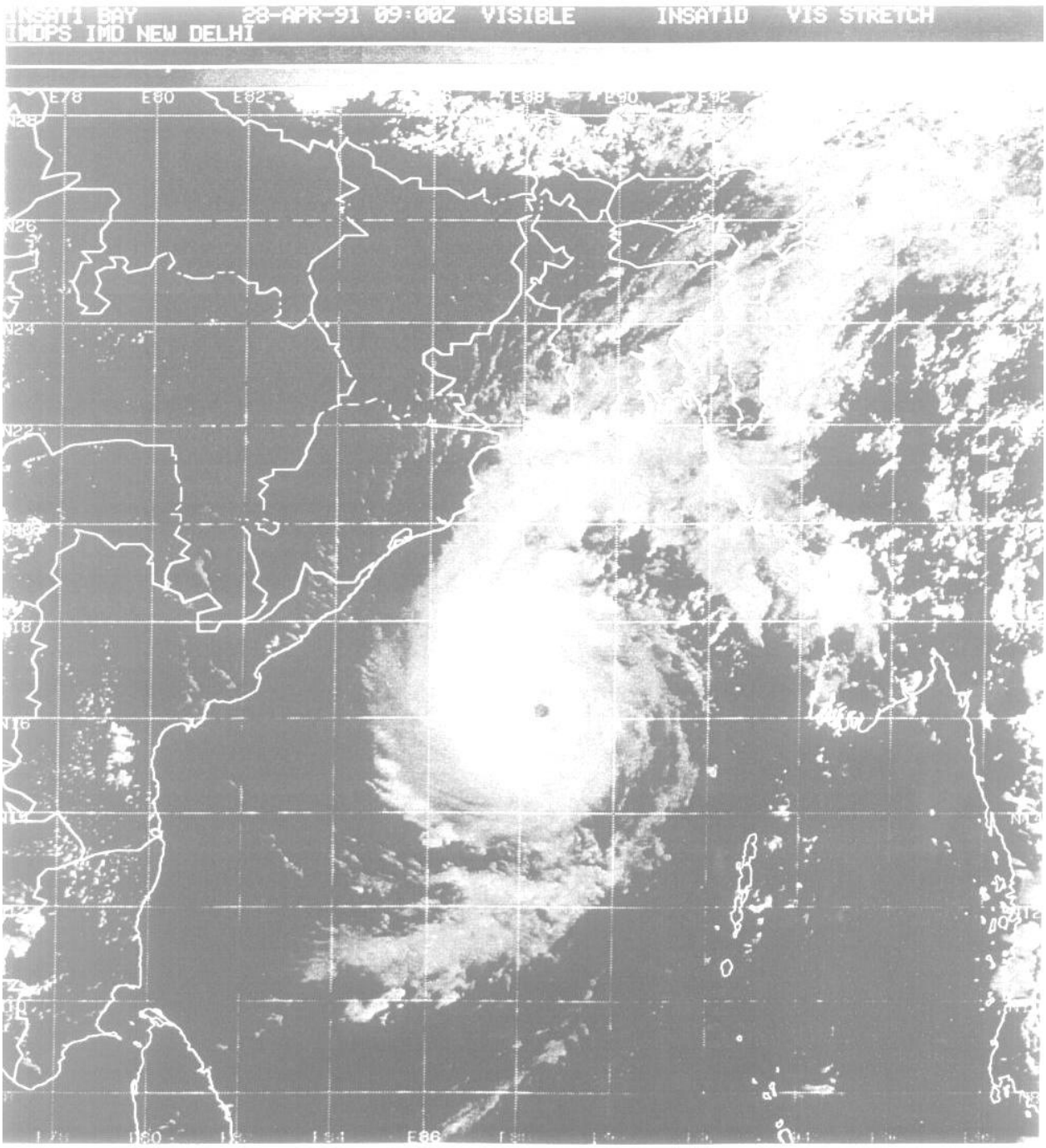
Banding eye pattern (T4.0) showing very severe cyclonic storm stage at 0600 UTC of 06.05.1990 (VIS). Curved band spirals fully around the eye. This pattern is usually absent in the IR imagery at this stage



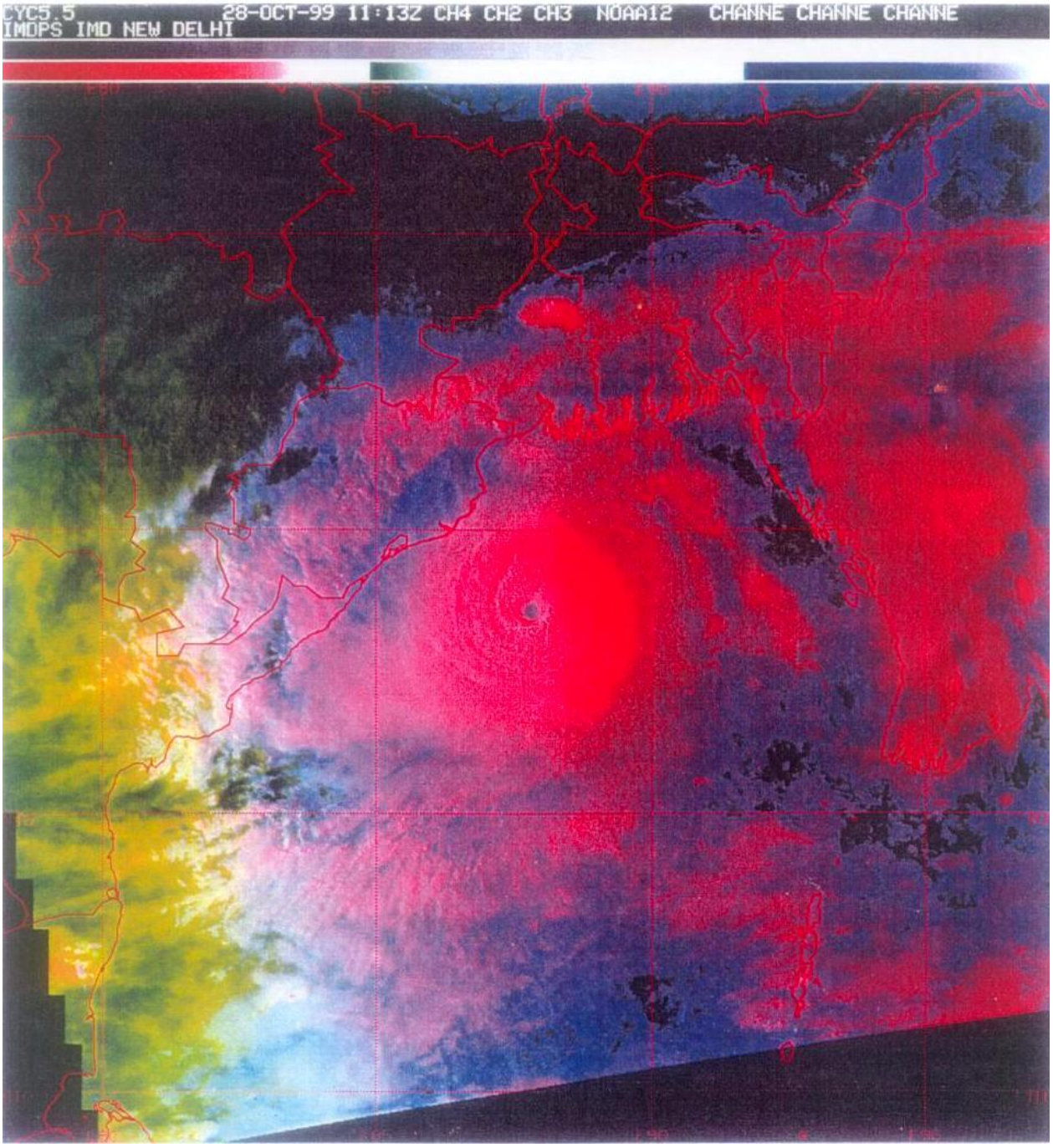
Comma cloud with eye pattern at 06 UTC of 19 November 1992 embedded distance of eye more than 75 kms. Intensity T5.5



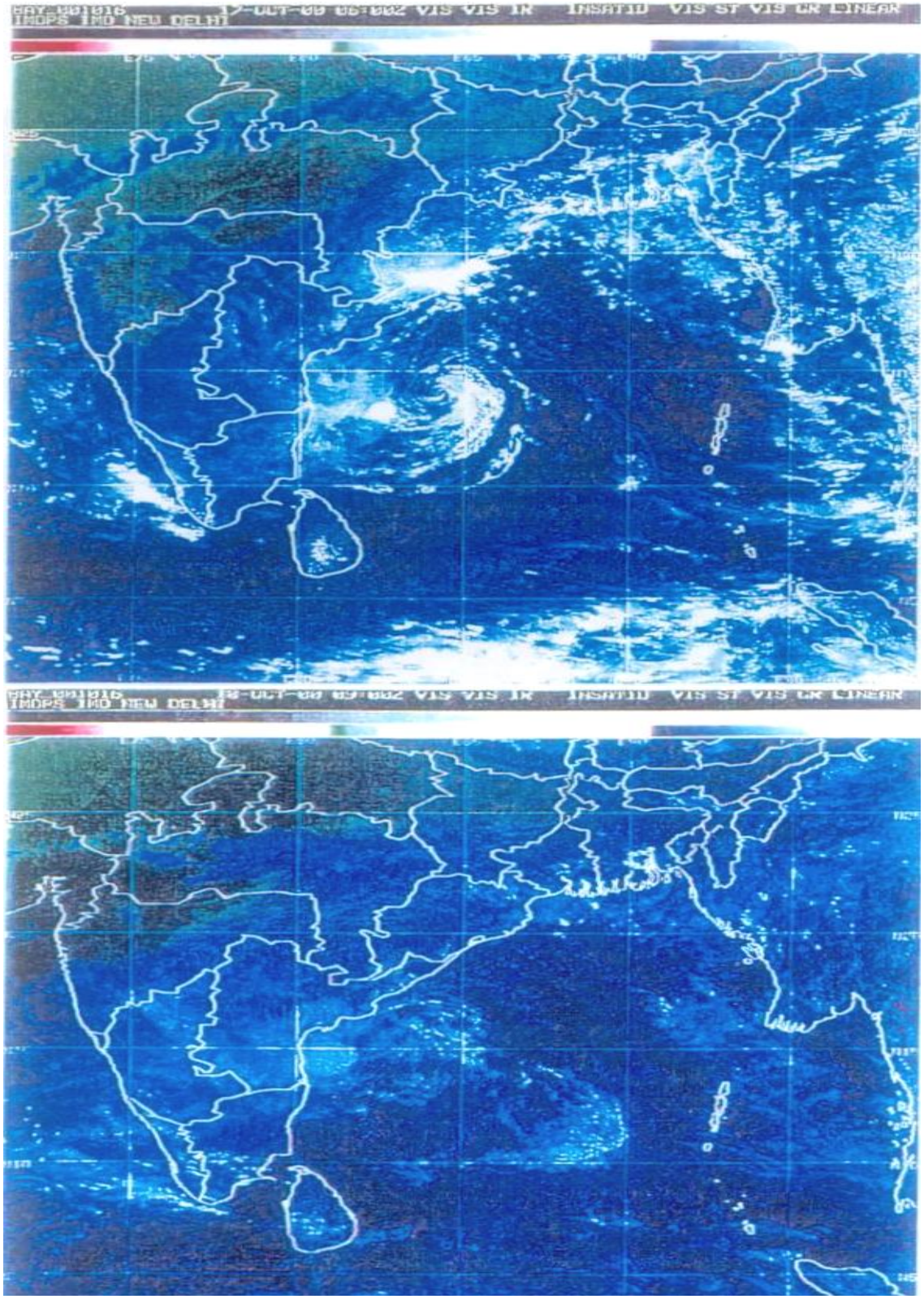
Multiple eye walls showing very severe cyclonic storm stage (T5.5) at 0900 UTC of 19 May 1999. A general impression from Dvorak technique suggests correspondence of this structure with Super Cyclonic Storm.



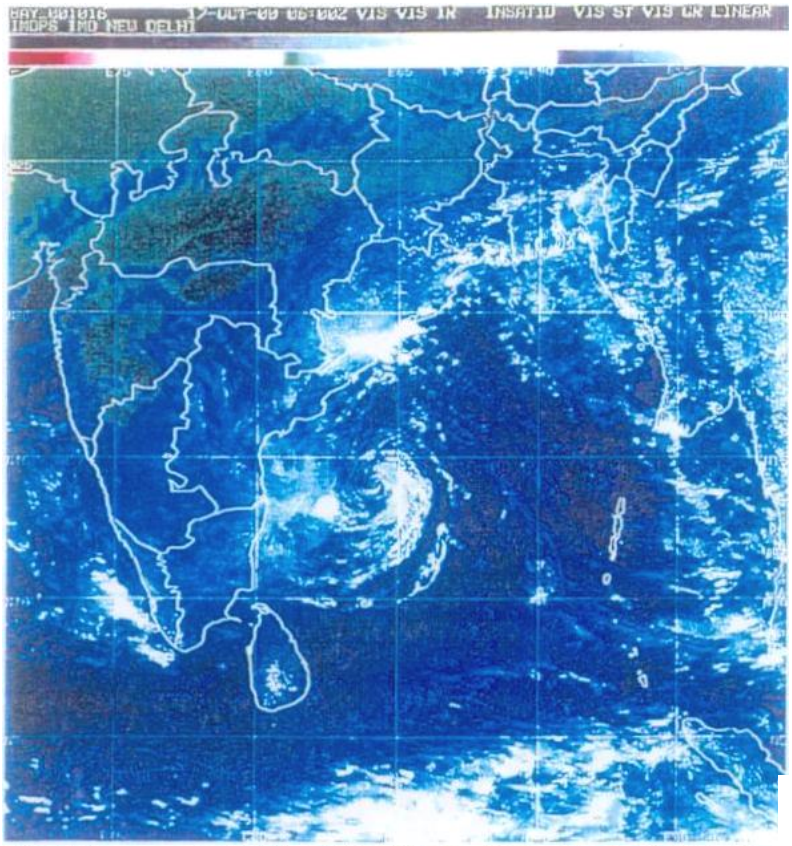
Three concentric eyewalls on 28 April 1991. CI No. T6.5. According to Dvorak double eyewalls are associated with super TCs.

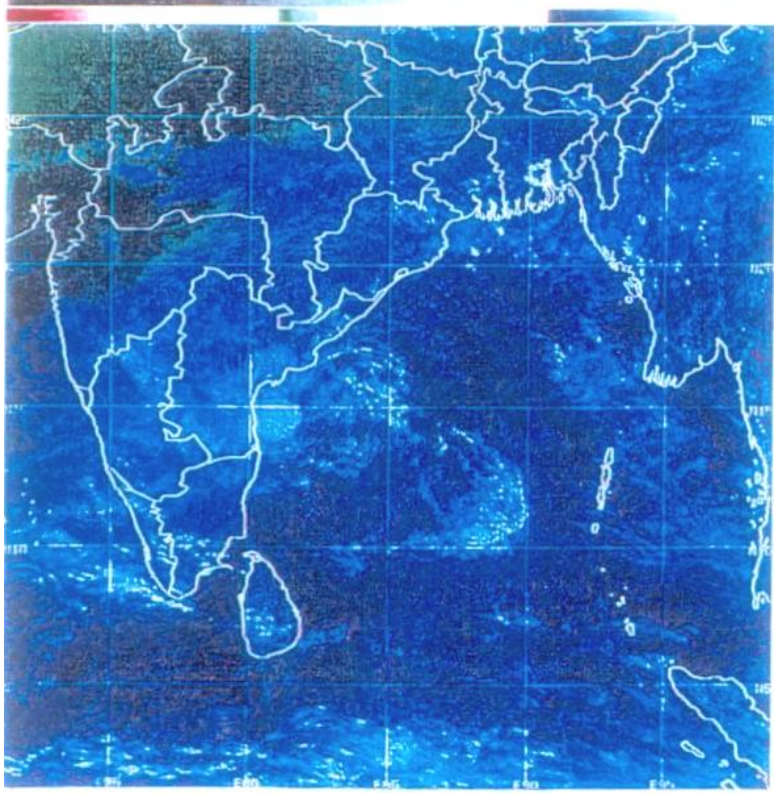


Signs of tightly overlapping eye walls in Orissa super Cyclone at 1130 UTC of 28 Oct 1999.



Weakening of TC due to easterly wind shear. Exposed LLCC of a TC east of deep convection which has since abated.





Weakening of TC due to easterly wind shear. Exposed LLCC of a TC east of deep convection which has since abated.

TASK-31 : MOVEMENT OF TROPICAL CYCLONES

Introduction

1. Tropical cyclones usually move with a direction and speed that closely approximates the tropospheric current that surrounds them. Logically, therefore, charts of the mean flow of the troposphere should be used as a basis for predicting the movement of tropical cyclones, but lack of observations generally precludes this approach. Generally there is a tendency for tropical cyclones to follow a curved path away from the Equator; however, departures from this type of track are frequent and of great variety,

2. Tropical cyclones move toward the greatest surface pressure falls and toward the area where the surface pressure falls increase fastest with time. Calculation is necessary for this rule to be used.

3. Numerous theories have been advanced to explain the cyclone tracks of the past and to predict those of the future. Observational data have never been sufficient to prove or disprove most of them.

4. Tropical storms move under the influence of both external and internal forces. The external forces are a result of air currents that surround a storm and carry it along. The internal forces appear to produce a tendency for a northward displacement of the storm (which probably is proportional to the intensity of the storm), a westward displacement that decreases as latitude increases, and aperiodic oscillation about a mean track.

Movement of Cyclone

5. **Initial Movement of Intense Cyclones.** The movement of cyclones that are undergoing or have just completed initial intensification (about 24 hours) will be as follows:

a. Storms developing in westward moving wave troughs in the easterlies move toward the west and also with a pole ward component given by an angle of approximately 20° to the right of the axis of the trough looking downstream. (See fig. 1.1.)

b. The motion of storms developing from preexisting vortices can be extrapolated from the track of these vortices.

c. Storms developing on the eastern edge of polar troughs initially move up the trough.

d. Pacific storms forming on equatorial shearlines usually are most active in the southern and southwestern portions at the start and they move slowly northeast. After 1 to 2 days, the influence of the trades becomes dominant and the storms turn back toward a direction ranging from west to north. (This appears to be true mainly for cyclones deepening near or west of 130°E.)

e. If a storm is discovered without information as to its past movement and the upper air current, it is best to start it along the climatological mean track and then

secure the data necessary to determine the steering current.

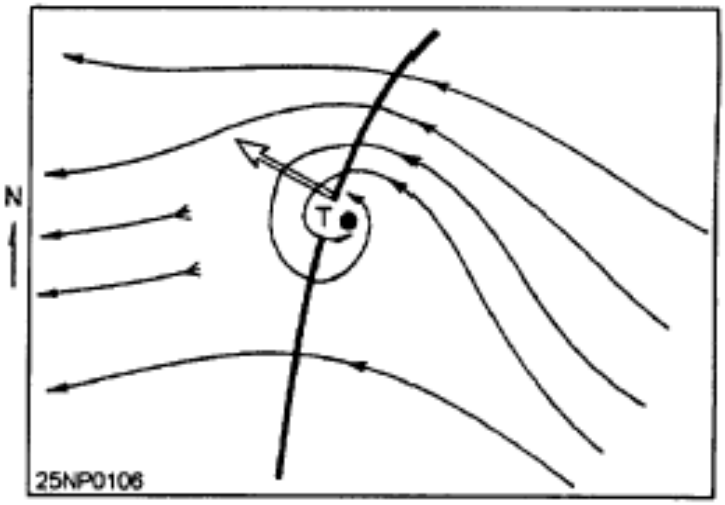


Fig 1.1 Initial movement of tropical cyclones forming in a wave trough In the easterlies

6. **Steering.** The movement of a tropical cyclone is determined to a large extent by the direction and speed of the basic current in which it is embedded. This concept appears to work well as long as the cyclone remains small and remains in a deep broad current. By the time a tropical cyclone has reached hurricane intensity, these conditions seldom exist. It then becomes necessary to integrate the winds at all levels through which the cyclone extends and in all quadrants of the storm to determine the effective steering current. Although the principle is not fully understood and has not yet received universal acceptance as a valid rule, it does have practical applications. For example, changes in winds atone or more levels in the area surrounding the cyclone can sometimes be anticipated, and in such cases, a qualitative estimate of the resulting change in the movement of the cyclone can be made. Conversely, when it appears likely that none of the winds in the vicinity of the cyclone will change appreciably during the forecast period, no change in the direction and speed of movement should be anticipated.

7. Further, this rule applies to situations of nonrecurvature and only some cases of recurvature. It is difficult to determine the steering current, since the observed winds represent the combined effects of the basic current and disturbances. As most storms extend into the high troposphere, it is better to calculate a "steering layer" than a steering level, since presumably the wind throughout most of the troposphere influences the storm movement. One writer recommends an integration of the mean flow between the surface and 300 hPa, over a band 8° in latitude centered over the storm. Another writer indicates that for moderate and intense storms the best hurricane steering winds would be found in the layer between 500 and 200 hPa and averaged over a ring extending from 2° to 6° latitude from the storm centre.

8. **Streamline Analysis and Vector Averages**. Other practical applications of the steering concept to short-range tropical cyclone motion use differing approaches in attempting to measure the basic current. One is by streamline analysis of successive levels to find a height at which the vertical circulation diminishes to a point such that the winds are supposedly representative of the undisturbed flow. Another method is to take vector averages of reconnaissance winds near the zone of strongest winds in the storm.

9. **Use of Observed Winds Aloft**. When sufficient data are available, it has been found that the use of streamline analysis of successive levels usually gives valuable indications of tropical storm movement for as much as 24 hours in advance. However, since wind observations are usually scarce in the vicinity of a hurricane, their analysis is necessarily rather subjective. Some forecasters claim dependable results in using this concept when data were available to high levels near the storm. The technique is not based on the assumption that wind at any single level is responsible for steering the storm, since the forces controlling movement are active through a deep layer of the atmosphere.

10. However, as successive levels are analysed, a level is found at which the closed cyclonic circulation of the storm virtually disappears. This *steering* level coincides with the top of the warm vortex and varies in height with different stages and intensities of the storm. It maybe located as low as 20,000 feet or, in the case of a large mature storm, as high as 50,000 feet. It has been found in analysis that most weight should be given to the winds in advance of the storm within a radius of 200 to 300 miles in preference to those in the rear quadrants. The hurricane generally moves with a speed of 60 to 80 percent of the current at the steering level.

11. **The Direction of Movement.** The direction of movement is not always exactly parallel to the steering current, but has a component toward high pressure that varies inversely with the speed of the current, ranging from almost 0° with rapid movement to as much as 20° with speeds under 20 knots. In westward moving storms, a component of motion toward high pressure could result from the poleward acceleration arising from the variation of the Coriolis parameter across the width of the storm. This would indicate that, to the extent that this effect accounts for the component of motion toward high pressure, northward moving storms would fit the direction of the steering current more closely than westward moving ones. The tendency for poleward drift would be added to the speed of forward motion in the case of a northward moving storm so that it would approach more closely the speed of the steering current. Empirical evidence supports this hypothesis.

12. Corrections for both direction and rate of movement should be made when this is indicated by the windflow downstream in the region into which the storm will be moving. For prediction beyond several hours, changes in the flow pattern for a considerable distance from the storm must be anticipated. It should also be remembered that intensification or decay of a storm may call for use of a higher or lower level, respectively, to estimate the future steering current.

13. **Recurvature.** One of the fundamental problems of forecasting the movement of tropical cyclones is that of recurvature. Will the cyclone move along a relatively straight line until it dissipates, or will it follow a track that curves poleward and eastward? When recurvature is expected, the forecaster must next decide where and when it will take place. Then, he or she is faced with the problem of forecasting the radius of the curved track. Even after the cyclone has begun to recurve, there are a great variety of paths that it may take. At any point, it may change course sharply.

14. The most common recurvature situation arises when an extratropical trough approaches a storm from the west or when the storm moves west to northwest toward a stationary or slowly moving trough.

15. **Indicators of Possible Recurvature.** Some of the indicators of possible recurvature are as follows:-

a. If the base of the polar westerlies lowers to 15,000 to 20,000 feet west of the storm's latitude, and remains in this position, recurvature may then be expected to occur.

b. However, if there is the building of a dynamic high or an eastward movement of this high to the rear of the advancing trough, and the westerlies dissipate in the low latitudes, the storm will move past the trough to its south and continue its westward path.

c. The above rule also holds true in cases where the polar trough moves from the west against a blocking high. The higher latitude portion of the trough continues to move eastward while the southern segment of the trough is retarded and is no longer connected with the upper portion of the trough.

d. Recurvature may be expected when an anchor trough is about 500 miles west of the storm and when the forward edge of the westerlies is from 500 to 700 miles to the west. Correlated with this parameter, RJ Shafer found that a spot value of the thickness between 850 to 500 hPa 7.5 degrees of latitude to the northwest of the storm was one of the most valuable of all parameters on recurvature or as an indication of future movement of Atlantic hurricanes. This thickness reflects the relative strength of cold troughs to the west essential for recurvature. Low values of thickness, 14,000 feet (approximately 4,270 meters) or less, almost always indicate recurvature; and high values, 14,200 feet (4,330 meters) or more, generally indicate continued westward motion.

e. The major trough west of the storm (in the westerlies) is slowly progressive.

f. Long waves are stationary or slowly progressive.

g. There is a rapid succession of minor troughs aloft.

h. The climatological mean track indicates recurvature (use with caution).

j. When the neutral point at the southern extremity of the trough in the westerlies at the 500-hPa level lies at or equatorward of the latitude of the cyclone, recurvature into the trough will usually occur. In this situation, the cyclone would normally be under the influence of southerly winds from the upper limits to a level well below the 500-hPa level while approaching the trough.

k. When the subtropical ridge at the 500-hPa level is broad and consists of large anticyclones, recurvature usually occurs. This case represents a low index situation in which the cyclone remains under the influence of a single, large, slow-moving anticyclone for a relatively long time.

l. Weak troughs exist between two separate subtropical high cells. Sometimes tropical storms move northward through these very weak breaks in subtropical highs.

16. **Nonrecurvature.** The following flow patterns are associated with nonrecurvature:

a. Strong subtropical anticyclone or ridge to the north of the storm with the mean trough in the westerlies located far to the west of the longitude of the storm. If this pattern develops strongly over the western oceans or continents, a storm will generally be driven inland and dissipate before it recurves into the western end of the ridge.

b. Flat (small-amplitude waves) in the westerlies at latitudes near or north of the normal position. A narrow subtropical ridge separates the westerlies from the tropical trough. In many cases the mean trough in the westerlies may be located at the same longitude of the storm.

17. **Motion during Recurvature.** The following rules apply during recurvature only:

a. When the radius of recurvature of a storm is greater than 300 miles, it will not decelerate and may even accelerate. The storm will slowdown if the radius of recurvature is less than 300 miles. In general, when the radius of recurvature is large, it is usually very uniform. A small radius will occur along a brief portion of the track. (See fig)

b. A large radius of recurvature is to be expected if the high northeast of the tropical cyclone has a vertical axis (fig). This will occur when long waves are stationary.

c. When the high slopes south to southeast with height (fig), the cyclone is transferred rapidly from the influence of upper easterlies to that of the upper westerlies and the track has a short bend. This occurs when long waves are progressive.

18. The following rules refer to short-term (24 hours or less) forecasting:

a. Tropical cyclones move toward the area of greatest surface pressure falls (12- to 24-hour pressure change).

b. Tropical cyclones move toward the area where the surface pressure falls increase most rapidly with time. For calculation, 3-hourly reports are necessary. Take, for example, the 24-hour pressure change at 1800 and subtract it from the 24-hour pressure change at 1500. This gives the acceleration of the pressure fall. If this quantity is computed for a network of stations and isolines are drawn for a suitable interval, the negative centre helps to pinpoint the expected cyclone position. This rule is used at intervals of 12 hours or less, and is especially helpful in determining the precise place of landfall for a storm that is just offshore.

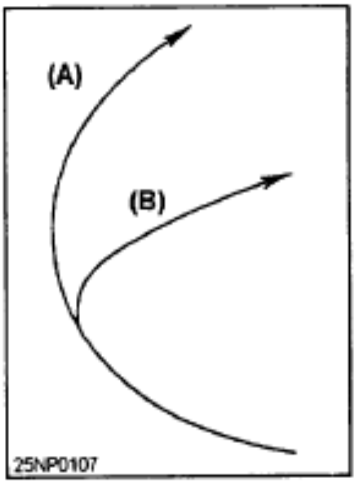


Fig 1.2. (A) Recurvature at a constant speed;
(B) First decelerating and then accelerating

19. **<u>Changes in Intensity during Movement out of the Tropics.</u>** During movement out of the Tropics the storm comes under influences different from those in its origin and tropical path. The following rules and observations can be helpful in forecasting the storm's changes in intensity:

a. A storm drifting slowly northward with a slight east or west component will preserve its tropical characteristics.

b. Storms moving northward into a frontal area or area of strong temperature gradients usually become extratropical storms. In this case the concentrated centre dies out rapidly while the area of gale winds and precipitation expands. The closed circulation aloft gives way to a wave pattern and the storm accelerates to the northeast.

c. If the tropical cyclone recurves into a trough containing a deep, slowly moving surface low, it normally will overtake this low and combine with it. Temporary intensification may result.

d. If the tropical cyclone recurves into a trough containing an upper cold core low, it appears to move into the upper low in most cases.

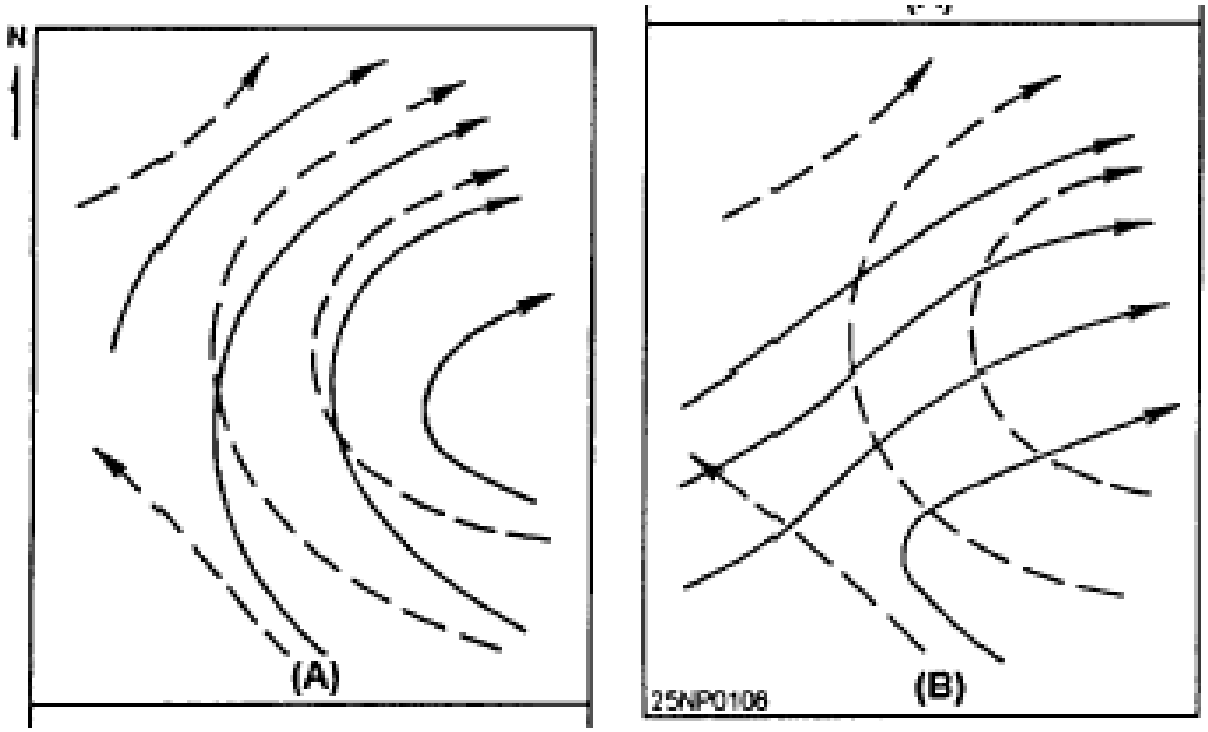


Fig 1.3. Surface flow pattern (dashed line) and 300- to 200-hPa flow pattern (solid lines).

SECTION-8 : CLIMATOLOGY

TASK-32 : CLIMATOLOGY AND WORLD WEATHER

Introduction

1. One of the major tasks of the meteorologists and oceanographers is providing long-range weather information and predictions based on recognised meteorological occurrences in a particular area or region of the world.

2. Air Force and Naval exercises are planned months and sometimes years in advance. To carry out these exercises successfully, we must have an idea of the normal weather conditions for the operational area at that time of year. It is both dangerous and unwise to conduct costly training exercises if the weather conditions for the area are known to be adverse at that time of year.

3. During wartime, an extensive knowledge of weather conditions can be a decisive advantage. Naval and land forces can use their knowledge of weather to surprise the enemy and predict when the enemy will strike. Historically, man wages war when the weather permits. When Napoleon invaded Russia, his defeat was not due to the wisdom of his opponents, but rather to his lack of knowledge of the severe Russian winters. He was beaten by the weather. As you gain more experience, your job will include the preparation of long-range weather forecasts based on climatological studies. You must prepare charts, tables and/or graphs that include sky cover, temperatures, winds, sea conditions, etc. This climatological information is needed for long-range naval exercises, ship deployments overseas, and actual combat operations.

Climatology as Related to Other Sciences

4. Three prefixes can be added to the word climatology to denote scale or magnitude. They are micro, meso, and macro and indicate small, medium, and large scales, respectively. These terms (micro, meso, and macro) are also applied to meteorology.

5. **Microclimatology**. Microclimatological studies often measure small-scale contrasts, such as between hilltop and valley or between city and surrounding country. They may be of an extremely small scale, such as one side of a hedge contrasted with the other, a plowed furrow versus level soil, or opposite leaf surfaces. Climate in the microscale may be effectively modified by relatively simple human efforts.

6. **Mesoclimatology**. Mesoclimatology embraces a rather indistinct middle ground between macroclimatology and microclimatology. The areas are smaller than those of macroclimatology and larger than those of microclimatology, and they may or may not be climatically representative of a general region.

7. **Macroclimatology**. Macroclimatology is the study of the large-scale climate of a large area or country. Climate of this type is not easily modified by human efforts. However, continued pollution of the earth, its streams, rivers, and atmosphere, can eventually make these modifications.

8. Climate has become increasingly important in other scientific fields. Geographers, hydrologists, and oceanographers use quantitative measures of climate to describe or analyse the influence of our atmospheric environment. Climate classification has developed primarily in the field of geography.

Ecology

9. Ecology is the study of the mutual relationship between organisms and their environment. Ecology is briefly mentioned here because the environment of living organisms is directly affected by weather and climate, including those changes in climate that are gradually being made by man. During our growing years as a nation, our interference with nature by diverting and damming rivers, clearing its lands, stripping its soils, and scarring its landscape has produced changes in climate. These changes have been on the micro and mesoscale and possibly even on the macro scale.

Climatic Elements

10. The weather elements that are used to describe climate are also the elements that determine the type of climate for a region. This chapter presents a brief explanation of the importance of these elements. The climatic elements of temperature, precipitation, and wind are not the only parameters included in a climatology package; however, they are the most significant elements used to express the climate of a region.

Temperature

11. Temperature is undoubtedly the most important climatic element. The temperature of an area is dependent upon latitude or the distribution of incoming and outgoing radiation the nature of the surface (land or water); the altitude; and the prevailing winds. The air temperature normally used in climatology is that recorded at the surface. Moisture, or the lack of moisture, modifies temperature. The more moisture in a region, the smaller the temperature range, and the drier the region, the greater the temperature range. Moisture is also influenced by temperature. Warmer air can hold more moisture than can cooler air, resulting in increased evaporation and a higher probability of clouds and precipitation. Moisture, when coupled with condensation and evaporation, is an extremely important climatic element. It ultimately determines the type of climate for a specific region.

Precipitation

12. Precipitation is the second most important climatic element. In most studies, precipitation is defined as water reaching Earth's surface by falling either in a liquid or a solid state. The most significant forms are rain and snow. Precipitation has a wide range of variability over the Earth's surface. Because of this variability, a longer series of observations is generally required to establish a mean or an average. Two stations may have the same amount of annual precipitation, but it could occur in different months or on different days during these months, or the intensity could vary. Therefore, it often becomes necessary to include such factors as average number of days with precipitation and average amount per day. Precipitation is expressed in most studies in the inches, but throughout the rest of the world, millimeters are normally used. Since precipitation amounts are directly associated with amount and type of clouds, cloud cover must also be considered with precipitation. Cloud climatology also includes such phenomena as fog and thunderstorms.

Wind

13. Wind is the climatic element that transports heat and moisture into a region. The climate of an area is often determined by the properties of temperature and moisture that are found upstream of that region. Climatologists are mostly interested in wind with regard to its direction, speed, and gustiness. Wind is therefore usually discussed in terms of prevailing direction, average speeds, and maximum gusts. Some climatological studies use resultant wind, which is the vectorial average of all wind directions and speeds for a given level, at a specific place, and for a given period.

Classification of Climate

14. The climate of a given region or locality is determined by a combination of several meteorological elements and not by just one element. For example, two regions may have similar temperature climates but very different precipitation climates. Their climatic difference, therefore, becomes apparent only if more than one climatic factor is considered.

15. Since the climate of a region is composed of all of the various climatic elements, such as dew, ice, rain, temperature, wind force, and wind direction, it is obvious that no two locations can have exactly the same climate. However, it is possible to group similar areas into what is known as a climatic zone.

Climatic Zones

16. The basic grouping of areas into climatic zones consists of classifying climates into five broad belts based on astronomical or mathematical factors. Actually they are zones of sunshine or solar climate and include the torrid, or tropical zone, the two temperate zones, and the two polar zones. The tropical zone is limited on the north by the Tropic of Cancer and on the south by the Tropic of Capricorn, which are located at 23½° north and south latitude, respectively. The temperate zone of the Northern Hemisphere is limited on the south by the Tropic of Cancer and on the north by the Arctic Circle located at 66½° north latitude. The temperate zone of the Southern Hemisphere is bounded on the north by the Tropic of Capricorn and on the south by the Antarctic Circle located at 66½° south latitude. The two polar zones are the areas in the polar regions which have the Arctic and Antarctic Circles as their boundaries.

17. Technically, climatic zones are limited by isotherms rather than by parallels of latitude (fig. 1.1). A glance at any chart depicting the isotherms over the surface of the earth shows that the isotherms do not coincide with latitude lines. In fact, at some places the isotherms parallel the longitude lines more closely than they parallel the latitude lines. The astronomical or light zones therefore differ from the zones of heat.

Climatic Types

18. Any classification of climate depends to a large extent on the purpose of the classification. For instance, a classification for the purpose of establishing air stations where favourable flying conditions are important would differ considerably from one for establishing the limits of areas that are favourable for the growing of crops. There are three classifications that merit particular attention. They are the classifications of CW Thornthwaite, W Köppen, and GT Trewartha. Thornthwaite's classification of climates places a great deal of emphasis on the effectiveness of precipitation. Effectiveness of precipitation refers to the relationship between precipitation and evaporation at a certain locality. Thornthwaite classified climates into eight main climatic groups; five groups give primary emphasis to precipitation and the other three groups are based on temperature.

19. Köppen's classification includes five main climatic types. They are tropical rain, dry, warm temperate rainy, cool snow forest (Boreal), and polar climates. These main types are further divided into climatic provinces. The Köppen classification is based mainly on temperature, precipitation amount, and season of maximum precipitation. Numerical values for these elements constitute the boundaries of the above types, which were selected primarily according to their effect on plant growth. Fig 1.2 shows Köppen's climatic types.

20. Köppen's climatic types are still considered valid today. His climatic zones, like others, are by no means static. Climatic zones shift with long-range weather patterns. The most noticeable shifts in these climatic zones have been observed over the northern portions of North America and Asia and over Africa. Russia and Canada, for example, have been able to farm land at higher latitudes over the past 200 years due to milder temperatures. Recent studies, however, indicate a general return of cooler temperatures at high latitudes, and now the growing region is gradually moving southward again where temperatures are more moderate. In Africa, desert regions have made notable shifts southward due to decreasing precipitation.

21. Trewartha is the most recent classifier of climate. Initially, his climatic classifications were based on Köoppen's; however, over the years, he has made significant changes and is now recognised for developing his own six climatic groups. These six groups are tropical, dry, subtropical, temperate, boreal, and polar. Five of these groups are based on temperature and one is based on precipitation (see Table 1.2). Trewartha's climatic groups, like Köppen's, are also further broken down into climatic types and subtypes.

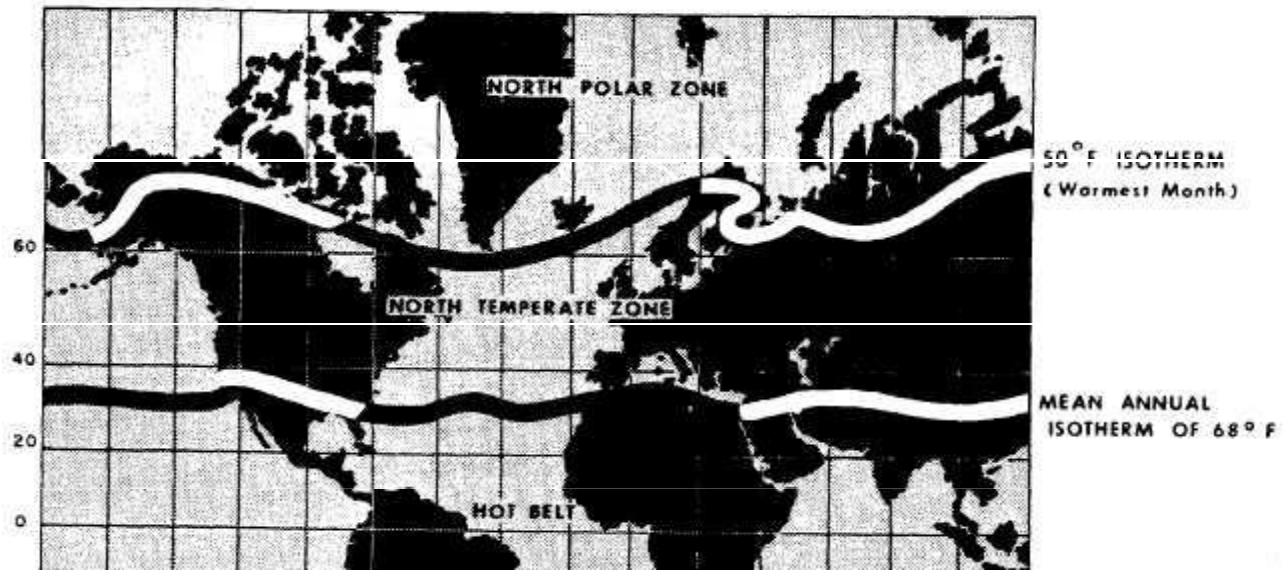
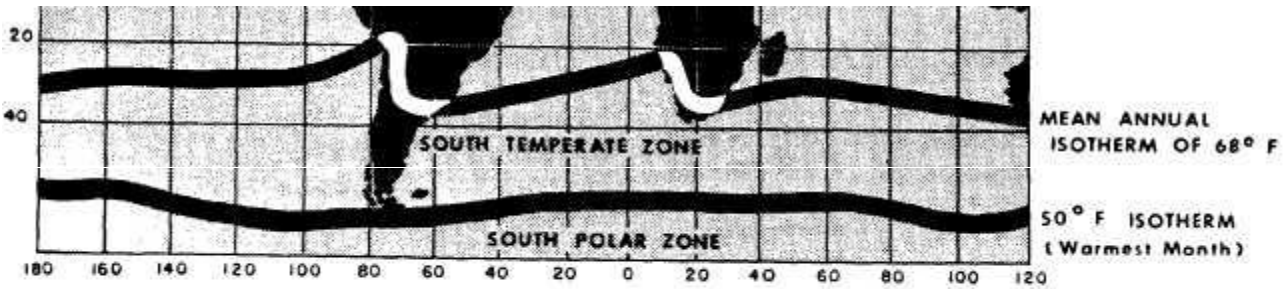


Fig 1.1. Temperature Zones



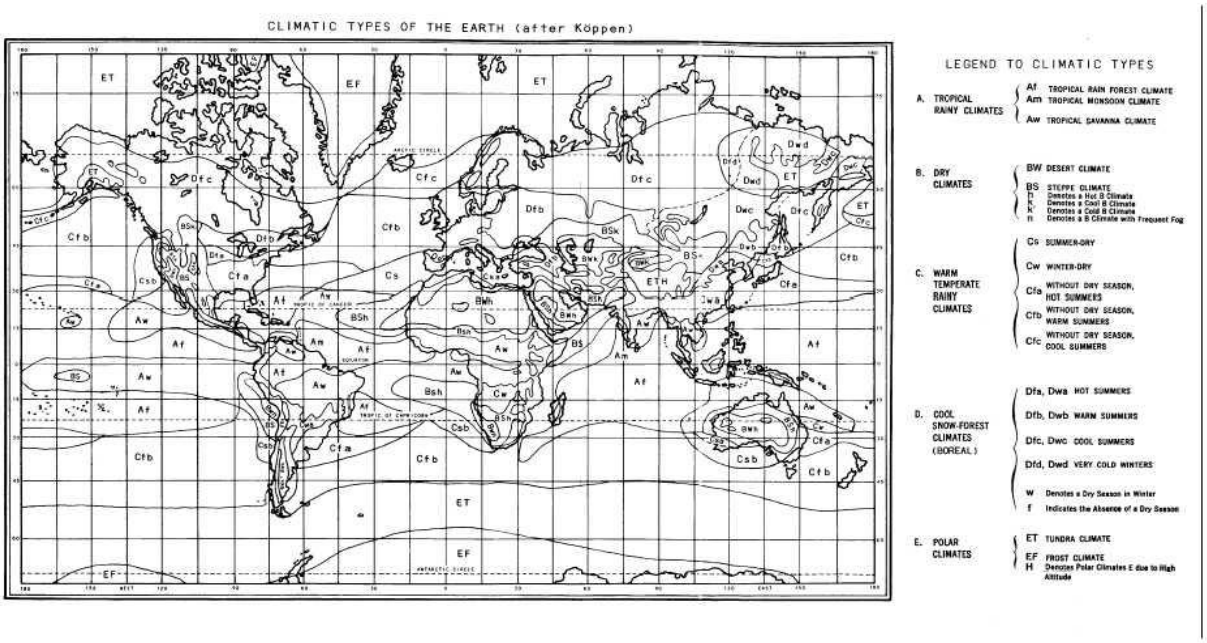


Fig 1.2. Köoppen's Climatic Types

Climatic Controls

22. The variation of climatic elements from place to place and from season to season is due to several factors called climatic controls. The same basic factors that cause weather in the atmosphere also determine the climate of an area. These controls, acting in different combinations and with varying intensities, act upon temperature, precipitation, humidity, air pressure, and wind to produce many types of weather and therefore climate. Four climatic controls largely determine the climate of every ocean and continental region. These controls are latitude, land and water distribution, topography, and ocean currents. Another factor which is now significant in determining a region's climate is man. Man's influence on climate through pollution, deforestation, and irrigation, is now considered a climatic factor.

Table 1. 1 Trewartha's Climatic Groups and their Poleward Boundaries

Basis for Classification	Climatic Group	Poleward Boundary
Temperature	A. Tropical	Frost line over continents and 65ºF (18ºC) over oceans (coolest months).
Precipitation	B. Dry	Bounded by the outer limits where potential evaporation is equal to precipitation.
Temperature	C. Subtropical	50ºF (10ºC) or above for 8 months of the year.
Temperature	D. Temperature	50ºF (10ºC) or above for 4 months of the year.
Temperature	E. Boreal	50ºF (10ºC) or above for 1 month of the year.
Temperature	F. Polar	Below 50ºF (10ºC) entire year.

Latitude

23. Perhaps no other climatic control has such a marked effect on climatic elements as does the latitude, or the position of Earth relative to the Sun. The angle at which rays of sunlight reach Earth and the number of Sun hours each day depends upon the distance of the Sun from the equator. (See fig. 1.3) Therefore, the extent to which an air mass is heated is directly influenced by the latitude. Latitude influences the sources and direction of air masses and the weather they bring with them into a region.

24. The importance of latitude as a climatic control can be shown by comparing an equatorial area to a polar area. In the former, the Sun is close to being directly overhead during the day throughout the year. Therefore, there is little difference between mean temperatures for the coldest and warmest months. In the polar area, however, the Sun never rises far above the horizon; that is, the angle of the Sun to Earth's surface is always acute. The radiant energy received per unit area is therefore slight, and the warming effects of the Sun are relatively weak. The average world surface temperatures are represented on two world charts for January and July in fig 1.1. These are mean charts and are not meant to be an accurate portrayal of the temperatures on any one particular day. Note that in general the temperatures decrease from low to high latitudes.

Land and Water Distribution

25. Land heats and cools about four times faster than water. Therefore, the location of continents and oceans greatly influences Earth's pattern of air temperature as well as the sources and direction of movement of air masses.

Influence on Air Temperature

26. Coastal areas assume the temperature characteristics of the land or water that is on their windward side. In latitudes of prevailing westerly winds, for example, west coasts of continents have oceanic temperatures and east coasts have continental temperatures. These temperatures are determined by the wind flow.

27. Since the upper layer of the ocean is nearly always in a state of mixing, heat losses or heat gains occurring at the surface are distributed throughout a large volume of water. This mixing process sharply reduces air temperature contrasts between day and night and between winter and summer over oceanic areas.

28. Over land, there is almost no redistribution of heat by turbulence; also, the effect of conduction is negligible. Thus strong seasonal and diurnal contrasts exist in the interiors of continents. During the winter, a large part of the incident solar radiation is reflected back toward space by the snow cover that extends over large portions of the northern continents. For this reason, the northern continents serve as source regions for dry polar air.

29. The large temperature difference between the land and water surfaces, which reverses between the two seasons, determines the seasonal weather patterns to a great extent.

30. The isotherms over the Northern Hemisphere are more closely spaced and parallel in winter than in summer. In the Southern Hemisphere, the temperature gradient does not have as great a seasonal change as it does in the Northern Hemisphere. These conditions are due to the unequal distribution of land and water on the two hemispheres. Since the Southern Hemisphere has less land and more water surface than the Northern Hemisphere, the change due to the greater water surface is less with consequently more nearly uniform isotherms. Also, the continents of the Southern Hemisphere taper toward the poles and do not extend as far poleward as do those in the Northern Hemisphere. The nature of the surface affects local heat distribution. Colour, texture, and vegetation influence the rate of heating and cooling. Generally, dry surfaces heat and cool faster than moist surfaces. For instance, plowed fields, sandy beaches, and paved roads become hotter than surrounding meadows and wooded areas during the day. During the night, however, the situation is reversed.

31. The distribution of water vapour and clouds is another important factor influencing air temperature. Although areas with a high percentage of cloud cover have a high degree of reflectivity, the energy which is not reflected is easily trapped in the lower layers due to the greenhouse effect. Thus, areas of high moisture content have relatively high temperature.

Influence on Air Circulation

32. The higher mean temperature of the Northern Hemisphere is an effect not only of its higher percentage of land, but also of the fact that its oceans are also warmer than those in the Southern Hemisphere. This is partly due to the movement of warm equatorial waters from the Southern Hemisphere into the Northern Hemisphere caused by the southeast trades crossing the equator. Another factor conducive to higher mean temperatures in the Northern Hemisphere is the partial protection of its oceans from cold polar waters and arctic ice by land barriers. There is no such barrier between the Antarctic region and the southern oceans.

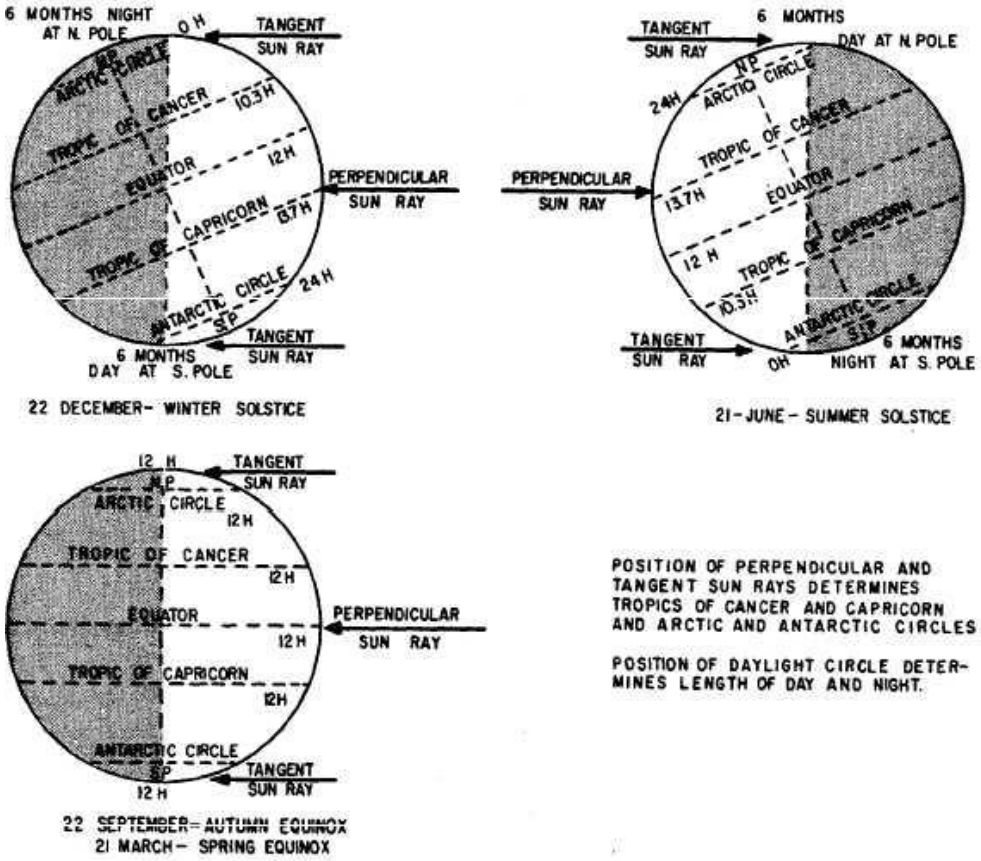


Fig 1.3. Latitude differences in amount of Insolation

Topography

33. Climates over land may vary radically within very short distances because of the elevation and variations in land forms. Therefore, topography plays an extremely important role in determining the climate of a region.

34. The height of an area above sea level exerts a considerable influence on its climate. For instance, the climate at the equator in the high Andes of South America is quite different from that found a few feet above sea level at the same latitude. All climatic values are affected by surface elevation.

35. An important influence on climate is mountainous terrain, especially the long, high chains of mountains that act as climatic divides. These obstacles deflect the tracks of cyclones and block the passage of air masses at the lower levels. If the pressure gradients are strong enough to force the air masses over the mountains, the forced ascent and descent modifies the air masses to a great extent, thus modifying the climate on both the windward and leeward sides of the range. The orientation of the mountain

range may block certain air masses and prevent them from reaching the lee side of the mountains. For example, the Himalayas and the Alps, which have east-west orientations, prevent polar air masses from advancing southward. Therefore, the climates of India and Italy are warmer in winter than are other locations of the same latitude. The coastal ranges in North America, running in a north-south line, prevent the passage of unmodified maritime air masses to the lee side.

36. The most noted effect of mountains is the distribution of precipitation. The precipitation values, level for level, are much higher on the windward side than on the leeward side. In regions where the prevailing circulation flows against a mountain barrier, the amount of precipitation increases more or less uniformly with elevation on the windward side of the range. This steady increase normally occurs up to elevations of about 10,000 feet. However, in the trade wind zone (such as at the Hawaiian islands), precipitation increases only to about 3,000 feet and then decreases gradually. Even with this decrease in amount, more rain is received at 6,000 feet than at sea level.

37. Another important topographical feature is the presence of lakes. The lake effect can be notable for large unfrozen bodies of water. The lee sides of lakes show considerable diurnal and annual modification in the form of more moderate temperatures; increased moisture, clouds, and precipitation; and increased winds (due to less friction) and land and sea breeze effects.

Ocean Currents

38. Ocean currents play a significant role in controlling the climate of certain regions. Ocean currents transport heat moving cold polar waters equatorward into warmer waters and moving warm equatorial waters poleward into cooler waters.

39. Currents are driven by the major wind systems; therefore, cold southward-moving cur-rents flow along the west coasts of continents, and warm northward moving currents flow along the east coasts of continents. This is true in both hemispheres. Basically, this results in cooler climates along the west coasts and warmer climates along the east coasts.

40. A brief explanation of the effects of ocean currents is presented here.

Effects on the West Coasts

41. The northern portions of the west coasts of continents generally have cool summers and warm winters. The summers are cool because of the presence of cold northern waters along their shores. However, the winters are generally mild because of the transport of warm ocean waters to these latitudes. For example, the south and southeast coasts of Alaska and the west coasts of Canada, Washington, and Oregon have relatively warm currents flowing along their shores. These currents are the Aleutian and North Pacific currents, which are branches of the warm northward-flowing Kuroshio

Current. The currents flow along the west side of the Pacific high and bring warm water into southern Alaska and the Pacific Northwest.

42. As these currents merge and flow southward along the British Columbia coast, they move into warmer waters and become the cold California current.

43. The southern portions of the west coasts of continents generally have cooler climates than do the east coasts of the same latitude. For example, during summer, the cold California current flows southward along the shores of California. Due to the Pacific high, the winds normally flow either across the cold current toward shore (on-shore) or parallel to the coastline. This results in cool air being advected inland allowing cities such as San Francisco and Seattle to enjoy relatively cool summers. Unfortunately, when the warm, moist air from the Pacific high does move over the underlying cold current, extensive fog and stratus develop which also move inland. This situation is typical along the southern portions of the west coasts in both hemispheres.

44. Another factor affecting west coasts is upwelling. Upwelling is the process by which cold subsurface waters are brought to the surface by wind. It occurs in areas where the wind causes the surface water to be transported away from the coast. The surface water is then replaced by the colder subsurface water. In the Northern Hemisphere, upwelling is common where the wind blows parallel to the coast and the surface water is transported away from the coast. In the process of upwelling, the exchange of water takes place only in the upper layers.

45. Generally, the following statements are true regarding the effects of ocean currents along the west coasts of continents:-

a. The west coasts of continents in middle and higher latitudes are bordered by warm waters which cause a distinct maritime climate characterised by cool summers and relatively mild winters with small annual range of temperatures (upper west coasts of the United States and Europe).

b. The west coasts of continents in tropical and subtropical latitudes (except close to the equator) are bordered by cool waters, and their average temperatures are relatively low with small diurnal and annual ranges. There are fogs, but generally the areas (southern California, Morocco, etc.) are arid (dry).

Effects on the East Coasts

46. The effects of currents along the eastern coasts of continents are less dramatic than those of the west coasts because of the west-to-east flow of weather. The effects, however, are just as significant.

47. In the tropical and subtropical regions, warm ocean currents introduce warm, rainy climates, especially on the windward sides of mountainous landmasses. As the warm currents progress northward into middle latitudes, warm, moist air produces a hot, humid

climate with frequent rain showers during the summer. Winters are relatively moderate (but still cold) along the coast due to the transport of warm water. The higher latitudes along eastern shores normally have cold waters flowing southward from the polar region; warm ocean currents rarely extend very far north. The regions where the two currents meet have cool summers and cold winters with extensive fogs. This is especially true along the Grand Banks of Newfoundland and the Kamchatka Peninsula of eastern Asia.

48. The following general statements are true regarding the effects of ocean currents along the eastern coasts of continents:-

a. The east coasts in the tropics and sub-tropical latitudes are paralleled by warm currents and have resultant warm and rainy climates. These areas lie in the western margins of the subtropical anticyclone regions (Florida, Philippines, and Southeast Asia).

b. The east coasts in the lower middle latitudes (leeward sides of landmasses) have adjacent warm waters with a modified continental-type climate. The winters are fairly cold, and the summers are warm and humid.

c. The east coasts in the higher middle latitudes typically experience cool summers with cool ocean currents paralleling the coasts.

Other Effects

49. Ocean currents also affect the location of primary frontal zones and the tracks of cyclonic storms. Off the eastern coast of the United States in the winter, two of the major frontal zones are located in areas where the temperature gradient is strong and where a large amount of warm water is being transported into the middle latitudes. The fact that these frontal zones are located near large amounts of energy suggests that cyclones developing in these regions along the primary front may be of thermodynamic origin. The main hurricane tracks in the Atlantic and Pacific also appear to follow warm waters. Extratropical cyclones also tend to occur in warm waters in fall and early winter.

Climatic Factors

50. Human activity and vegetation can have marked effects on the climates of local areas. Eventually man’s activities could affect larger areas and ultimately whole continents.

51. It has been known for years now that urban areas and industrial complexes have an influence on climate. Atmospheric pollution is increased, for example, and the radiation balance is thereby altered. This change affects the daily maximum and minimum temperatures in cities, where they tend to be generally higher than in nearby suburbs. A higher concentration of hydroscopic condensation nuclei in cities results in an increased number of fogs. Also, with the greater heat source found in cities, increased convection gives rise to greater amounts of cloudiness and precipitation. An apparent benefit of this increased heat is a slight decrease in severe weather occurring in large cities (Chicago, for example) as compared to adjacent areas.

52. Areas of heavy vegetation generally have distinct climates which may differ considerably from climates of nearly open areas. Falling precipitation caught in trees before reaching the ground may be evaporated, but precipitation which reaches the ground does not evaporate or run off readily. Heavily forested areas can absorb and store considerable quantities of water. Snow in forests can be protected from direct insolation by the trees and may stay on the ground for much longer periods than snow on open, exposed surfaces. In forests, temperature maximums and minimums are higher than over open land at the same latitude. Relative humidities are also higher and wind speeds are considerably lower.

Climatological Data

53. Climatological records are based on the meteorological observations that are taken at a particular locality. This information may be presented in a number of ways. Temperature records generally include the following temperature values: daily maximums and minimums by months; the extremes; the average temperature by year and month; the mean monthly and annual temperature; the mean monthly maximum and minimum temperature; and (sometimes) the monthly and seasonal degree-days. Of great climatic significance is the range between the mean temperature of the warmest month and the coldest month. Other temperature data are sometimes given. These may include the number of days with the following temperatures: maximum of 90°F and above; maximum of 32°F and below; minimum of 32°F and below; and minimum of 0°F and below.

54. Precipitation records include the mean annual and monthly totals. The range between the highest and the lowest annual rainfall for a locality is the best indication of the dependability of the precipitation. The records often show the absolute maximum rainfall and snowfall for a 24-hour period by months, as well as the maximum and minimum precipitation for each month.

55. Climatic records usually show data on winds. Such information indicates the mean hourly speed and the prevailing direction by month. Also shown are the speed and direction of the strongest wind for the 12 months and the year in which it occurred. Data on cloudiness, humidity, thunderstorms, and heavy fog are often included. Other helpful data would be the frequency and distribution of cyclones and anticyclones; passage of fronts; pro-portion of rainfall and snowfall received from cyclonic storms and local, air mass thunderstorms; and climatological data on upper air conditions.

Methods of Presentation

56. Climatological information is presented in many different ways. Tables are frequently used. Maps are particularly useful in presenting climatic information in cases where geography is an important factor. Wind data can be given by means of a device called a wind rose, which presents information on the prevailing wind directions. (See fig. 1.4) Graphs are usually divided into bar and line graphs, or the graph may be a combination of the two. Figure 1.5 is an example of a bar graph and a line graph showing the same information. Figure 1.6 shows a combination of a bar and line graph used to depict cloud cover.

Availability of Data

57. Every Air Force unit activity should have climatological records available for their area and for such other areas as may be necessary to provide climatological support at the local unit level.

Weather Summary

58. Weather summaries are prepared for all Air Force stations from monthly meteorological records (MMRs). Each weather summary is for a specific station. Frequency distributions for various parameters are presented by time of day, month, and year. Weather summaries are revised every 5 years.

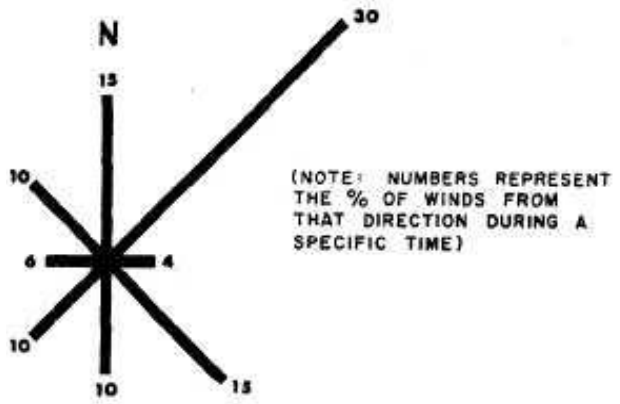


Fig 1.4. A Wind Rose

Local Climatological Data (LCD) Summary

59. The LCD summary is prepared only for selected civilian stations in the continental U.S.A. It consists of means and extremes (temperature, precipitation, wind, etc.) by month, mean temperature and total precipitation by month for specific years of record, and monthly and seasonal degree-days. The LCD is revised annually.

Cross-Wind Summary

60. The Cross-Wind Summary presents the percentage of occurrence of cross-winds for a given location. It is produced only on request.

Interpretation

61. Climatological records must be interpreted correctly to gain the needed information. Proper interpretation requires that all of the meteorological elements be studied so they present a composite picture. One meteorological element alone may

mean very little. For instance, it is possible to conclude that Cairo, Egypt, and Galveston, Texas, have about the same kind of weather based solely on the temperature, since the yearly and monthly means and annual range are approximately the same. However, Galveston has about 40 times as much precipitation. Thus, their weather conditions over the year differ greatly.

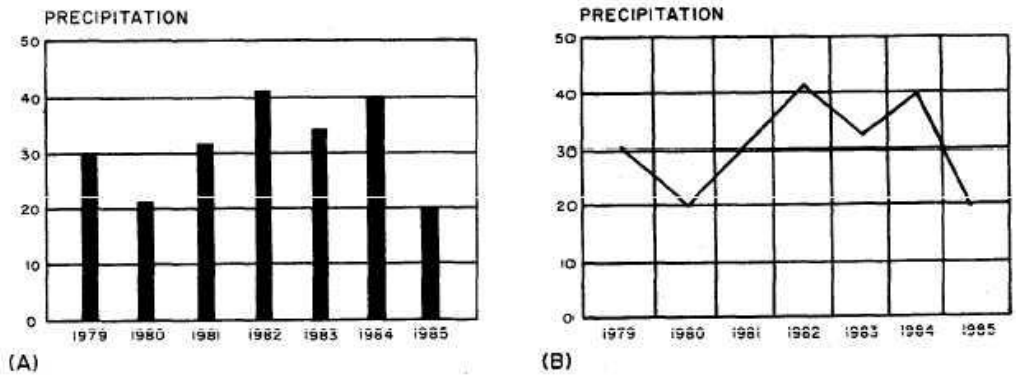


Fig 1.5. A comparison of the bar and line graph method of showing the variable annual precipitation in a time series. (A) Bar graph; (B) Line graphs.

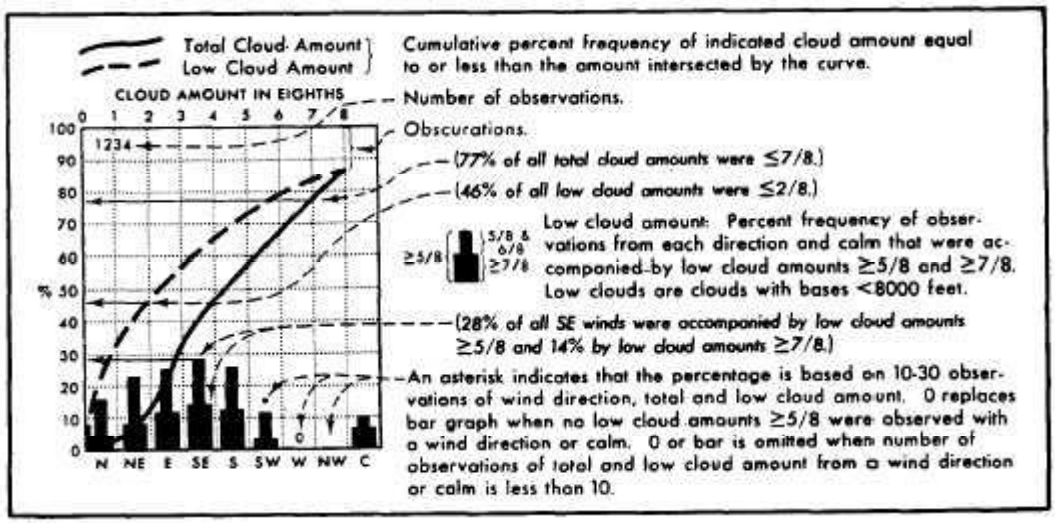


Fig 1.6. Combination bar and line graph (with legend and instructions for use)

62. To interpret just one meteorological element requires a study of several factors. For example, the temperature of a particular locality must be studied from the standpoint not only of the mean but also of the extremes and the diurnal and annual ranges. The effectiveness of precipitation also depends on several factors, such as amount, distribution, and evaporation. The mean precipitation for a particular month for a locality may be several inches, but the interpreter may find from a study of the locality's records that in some years the precipitation for that month is less than an inch, possibly not even a trace.

Application to Weather Prediction

63. Climatology is introduced where operational planning is required for a length of time beyond the range covered by weather-forecasting techniques. A study of the climate of an area or region may well foretell the general weather pattern to be expected.

64. A more direct application of climatology can be made by both the experienced and the inexperienced forecaster and assistant forecaster. Those personnel having personal experience at a particular station can use climatology as a refresher for the overall weather patterns that can be expected for the ensuing season. This knowledge can help them to be more perceptive in their everyday analyses, to be alert for changing patterns with the seasons, and to produce a higher quality forecast.

65. The personnel who have had no experience at a particular station must rely on climatology as a substitute for their experience.

66. Forecasters cannot be expected to become familiar overnight with the weather peculiarities of their new area of responsibility. The station certification period can be greatly reduced if the new people are furnished with "packaged experience" in a form that can place them more nearly on a par with those forecasters already experienced at that station. The Local Area Forecaster's Handbooks are good examples of this type of packaged information. In using the data, however, it must be clear that climatology has its limitations in the field of meteorology. It may be put this way. Climatology is an essential supplement to meteorology, but it must never be considered a substitute for the meteorological situation that constitutes current weather conditions.

World Weather

67. Meteorologists are stationed, and may travel, around the world. Ships and aircraft are constantly in global transit. Therefore, the meteorologists must have a general knowledge of types of weather encountered during various seasons in regions all around the world. This knowledge also increases insight into atmospheric circulation, weather development and movement, weather effects on the environment, and credibility as a knowledgeable analyst, interpreter, and briefer.

Note: You will find that a world atlas can be extremely useful and informative if used in conjunction with the information that follows.

Oceanic Weather

68. Some general considerations of the weather encountered over ocean areas are discussed in this lesson. Because land and water heat and cool at different rates, the location of continents and oceans greatly affects the Earth's pattern of air temperature and therefore influences the weather. The upper layers of the ocean are almost always in a state of motion. Heat loss or gain occurs at the sea surface and is distributed

throughout large volumes of water. This mixing process sharply reduces the temperature contrasts between day and night and between winter and summer.

Oceanic Weather Control

69. It has long been recognised that the ocean plays an important part in climate and weather, particularly in the realms of temperature, humidity, and precipitation. This is only natural, since three-fourths of Earth's surface is covered by water.

70. The two climatic extremes that relate to water and land distribution over Earth are Maritime and Continental. Continental climate is generally evidenced by a wide range in annual and diurnal temperatures, little cloudiness, and little precipitation. Continental climate is a product of a minimal influence from the oceans. Maritime climate prevails over the oceans and is characterised by a small temperature range, both annual and diurnal, and considerable precipitation and cloudiness.

71. Water vapour is considered one of the most important variables in meteorology. The state of the weather is largely expressed in terms of the amount of water vapour present and what is happening to the water vapour. Two principal elements of climate, precipitation and humidity, are dependent upon water vapour. Since the oceans are the main source of water vapour, it follows that weather is largely controlled by the oceans

Effects of Air-Sea Interchange

72. The atmosphere and the oceans have tremendous effects on each other. These effects are principally in the realms of temperature and water vapour. The heat balance of the oceans is maintained by the processes of radiation, the exchange of sensible heat, and the evaporation and condensation of water vapour on the sea surface.

73. The amount of radiant energy absorbed by the sea depends upon the amount of energy reaching the surface and the amount of reflection by the surface. When the Sun is directly overhead, the amount of its energy reflected amounts to only about 3 percent. Even when the Sun is 30° above the horizon, the amount of reflection is just 6 percent. However, there is a reflection of about 25% of the energy when the Sun is 10° above the horizon. (See fig. 1.7) Reflection loss is especially great in the presence of waves when the Sun is low.

74. Much of the insolation is absorbed in the first metre of seawater. This is true of the clearest water as well as of quite turbid (opaque) water. In water that is extremely turbid, the absorption is in the very uppermost layers. Foam and air bubbles are two major causes of a proportionately greater amount of absorption in the uppermost meter of the sea. However, due to vertical mixing, the heat absorbed in the upper layer is carried to great depths of the ocean, which acts as a great heat storage reservoir.

75. There is an exchange of energy between the oceans and the atmosphere. The surface of the oceans emits long-wave heat radiation. The sea surface at the same time receives long-wave radiation from the atmosphere. Although some of this incoming radiation from the atmosphere is reflected from the surface of the oceans, most of it is absorbed in a very thin layer of the water surface. The difference between the incoming long-wave atmospheric radiation and the outgoing long-wave radiation from the sea surface is known as the effective back radiation. The effective back radiation depends primarily on the temperature of the sea surface and on the water vapour content of the atmosphere. The time of day and the season have little effect on effective back radiation, since the diurnal and annual variation of the sea-surface temperature and of the relative humidity of the air above the oceans is slight.

76. For conduction to take place between the oceans and the atmosphere, there must be a temperature difference between the ocean surface and the air immediately overlying it. On the average, the temperature of the surface of the oceans is higher than that of the overlaying air.

77. It might be expected that all of the ocean's surplus of heat is either radiated or conducted to the atmosphere. This is not the case. Only a small percentage of the ocean's surplus heat is actually conducted to the atmosphere. About 90 percent of the surplus is used for evaporation of ocean water.

78. Due to the processes of radiation and mixing, the oceans act as a thermostat relative to the atmosphere. The energy stored at one place during one season may be given off at another locality and during a later season. Hence, there seems to be a constant effort by the atmosphere and the oceans to keep their temperatures in balance by an interchange of heat.

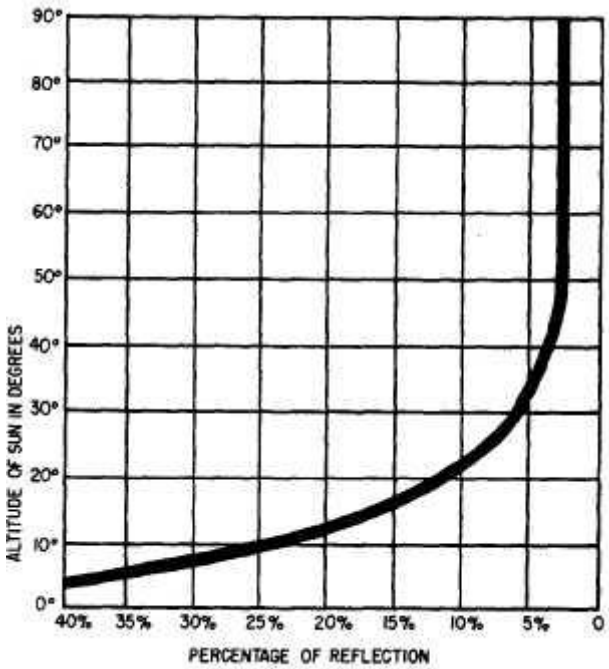


Fig 1.7. Percentage of reflected radiation

Stability

79. The deciding factor of most weather phenomena is the stability of the atmosphere. Air masses may become more stable or less stable as they move over ocean surfaces. The temperature contrast between the ocean surface and the lowest layers of the overlying air determines whether the ocean will promote stability or instability.

80. When the air moving over the ocean has a higher temperature than that of the ocean surface, the lower layers of the air become stable in time. On the other hand, when the air mass is colder than the ocean surface over which it is moving, instability results. As the colder air is warmed by the ocean, convective activity eventually develops. If the warming is sufficiently intense, thunderstorms develop.

Moisture Content

81. The interchange of moisture between the atmosphere and the oceans is one of the most important features of the whole meteorological picture. Without this interchange, weather, as we know it, could not exist; there would be no clouds and no precipitation. The oceans are by far the greatest source of moisture for the atmosphere. Other moisture sources are negligible in comparison. Whether the atmosphere gives up some of its moisture to the ocean or vice versa depends greatly upon vapour pressure. Vapour pressure is the pressure exerted by the molecules of water vapour in the atmosphere or over the surface of liquid water. When the vapour pressure of a liquid is equal to that of the atmosphere above the liquid, there is little or no apparent interchange of moisture. In other words, at equal vapour pressure, just as many molecules escape from the liquid to the atmosphere and vice versa. This is the case when air becomes saturated. The saturation vapour pressure increases with increasing temperature.

82. If the temperature of the surface water is warmer than that of the air, the vapour pressure of the water at its surface is greater than that of the air. When this condition exists, there can be abundant evaporation from the ocean surface. This evaporation is aided by the turbulence of the air brought on by the unstable condition of the lower layers. It follows, then, that the greatest evaporation takes place when cold air flows over warm ocean waters.

83. Let us consider the opposite condition-warm air flowing over a relatively cold body of water. When this happens, there is stable stratification in the lower layers of the atmosphere. The vapour pressure of the air soon reaches a state of equilibrium with that of the water surface. Evaporation stops. However, if the warm air is quite moist, it is possible for the moisture in the air to condense on the water surface. Contact of the warm air with the cold water may result in the formation of fog by lowering the air temperature to the dew point.

84. The direct interchange of moisture from the atmosphere to the oceans occurs through precipitation and, to a lesser extent, condensation. The direct interchange, however, is not as important meteorologically as the indirect interchange. The indirect interchange is a sequence of events beginning with the evaporation of water from the ocean surfaces and ending with the subsequent condensation and precipitation over land areas. Generally, precipitation occurs more frequently over land than over the oceans. Though the oceans are a source of abundant moisture, they normally lack the required precipitation mechanisms, such as vertical mixing, strong temperature contrasts, and orographic lifting.

Equatorial and Tropical Weather

85. In the temperate zone, where westerly winds predominate, pressure patterns move in an easterly direction. In the tropics, however, weather usually moves in the opposite direction. Normally, a moist layer, 5,000 to 8,000 feet deep exists in this region. During unfavourable weather, this layer deepens to more than 12,000 feet. Convergence occurs in opposing trade wind streams, northward flowing air, and areas of cyclonic curvature. The presence of a deep, moist layer and convergent winds account for the weather in equatorial and tropical regions.

Air-Mass Weather

86. Flying weather is usually best in tropical maritime air, at its source, within the sub-tropical highs. Scattered cumulus and patches of stratocumulus clouds may develop, but the sky is almost never overcast. Scant precipitation falls in scattered showers and variable, mild winds prevail.

87. The excellent flying weather in these mT source regions commonly extends through the moving air masses some distance from the sources. Cloudiness in the mT air increases with an increase in distance from the source. On flights from Hawaii or from the Azores northward, through northward-moving mT air, stratiform clouds increase. On flights from Hawaii or the Azores southward, through southward-moving mT air (or the northeast trades), cumuliform clouds increase. Here we are considering only Northern Hemisphere situations; however, a comparable pattern exists in the Southern Hemisphere.

88. A typical breakdown of the weather conditions you may encounter in air masses around the subtropical highs (fig. 1.8) is as follows:-

a. **North of a Subtropical High**. Any mT air that moves northward becomes cooled over the cool ocean surface. A stratus overcast may form, and drizzle may fall. Farther north, low ceilings (usually below 1,000 feet) may reach the surface, producing fog. The mT air surges farthest north in summer because subtropical highs are best developed and polar fronts lie farthest north. This mT air brings most of the summer fogginess to northern seas and coasts. It brings the greatest fogginess in the Atlantic where it blows from the warm Gulf Stream over the cold Labrador Current (near Newfoundland), and in the Pacific where it blows from the warm Kuroshio Current over the cold Oyashio current (near the Kamchatka peninsula).

b. **East of a Subtropical High**. Along the California coast, and along the Atlantic coast of North Africa, the mT air blows from the west and the northwest. This air tends to remain stable for the following reasons:

(1) It is coming from the northern, cooler portion of the source region.

(2) Its surface layers remain cool because it moves over cold ocean currents.

(3) Its upper portions warm adiabatically because of subsidence.

Throughout the year, airways are smooth. The skies are clear to partly cloudy. Clouds are generally patches of stratocumulus, and rain is rare. The chief flight hazard in this air is coastal fog, which often hides the California or European coastal land. Stratus and stratocumulus clouds may cause the sky to be overcast, develop low ceilings, and produce drizzle that reduces visibility.

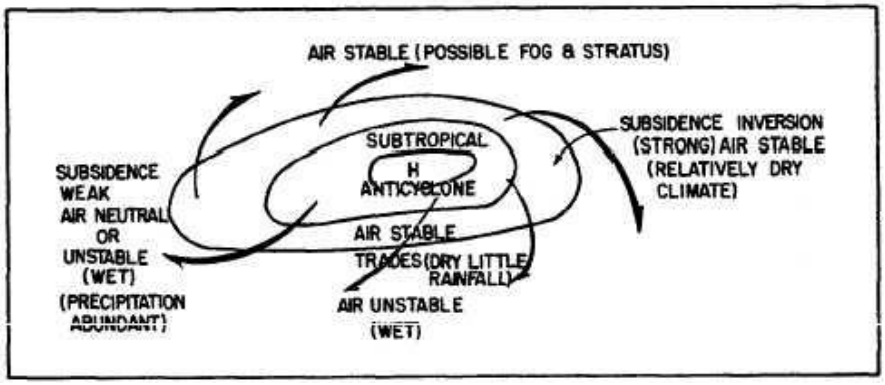


Fig 1.8 Weather, winds, and stability conditions around the subtropical high

c. **South of a Subtropical High.** Where the mT air moves southward or southwestward (as trade winds), its lower layers are warmed by the tropical ocean surface. This produces scattered cumulus. Near the equator, after absorbing much moisture and being heated, this air may develop cumulonimbus.

d. **West of a Subtropical High**. This mT air blows from the east and the southeast. Since it flows over warm water all of the way, the air neither cools or warms. Over the ocean near the Philippines (and near Florida and the West Indies), this trade wind brings good flying weather—clear or scattered cumulus clouds. When it is moving over land, this warm, moist air becomes unstable and turbulent and is a source of thunderstorms. When it moves over cold land (for example, southeastern United States in the winter), it becomes stable and produces stratus clouds or fog. Over cold ocean surfaces, such as the Sea of Japan and the Kamchatka and Labrador currents, it develops the persistent low stratus and fogs characteristic of these areas.

Arctic and Antarctic Weather

89. Geographically, the arctic zone is north of the Arctic Circle (66.5°N) and the Antarctic zone is south of Antarctic Circle (66.5°S). The Arctic is extremely important to the military defense of Canada and the United States and is the subject of ever-increasing military operations. Therefore, meteorologists must familiarise themselves with the prevailing weather and peculiarities of these regions.

Arctic Weather

90. The Arctic is the aerial crossroads of the world. This is not only due to the shorter arctic routes between some of the major cities of the world, but also because flying weather over the Arctic is generally better than that encountered over the familiar ocean routes.

91. To understand some of the important weather and problems of the Arctic, you must understand the broad underlying causes of the arctic climate.

92. **Seasonal Temperature Variations**. From our previous discussion of climatic controls, we have seen that the most important factor that determines the climate of an area is the amount of energy it receives from the Sun. During the winter much of the Arctic receives little or no direct heat from the Sun. The cold winter temperatures common in the Arctic result from a lack of the Sun's energy.

93. The Sun is not the only factor responsible for the arctic climate. Two other factors, the land-sea-ice distribution and mountain barriers, contribute to the tremendous variation in climate at different points of similar latitude.

a. **Land-Sea-Ice Features**. In the Northern Hemisphere, the water features include the Arctic, North Atlantic, and North Pacific oceans. These bodies of water act as temperature moderators since they do not have large temperature variations. A major exception occurs when large areas are covered by ice in winter. The land features are the northern continents of Eurasia, North America, the island of Greenland, and the Canadian Archipelago. As opposed to the water areas, the land areas tend to show the direct results of the extremes of seasonal heating and cooling by their seasonal temperature variations.

b. **Mountains**. The arctic mountain ranges of Siberia and North America are factors which contribute to the climate and airmass characteristics of the regions. These mountain barriers, as in mid-latitudes, restrict the movement of air from west to east. During periods of weak circulation, the air is blocked by the ranges and remains more or less stagnant over the area. It is during these periods that the air acquires the temperature and moisture characteristics of the underlying surface. Thus, these areas are air-mass source regions, and they are particularly effective as source regions during the winter when the surface is covered with snow and ice.

94. The Greenland ice cap is essentially a mountain range more than 10,000 feet above mean sea level. It restricts the movement of weather systems, often causing low-pressure centres to move northward along the west coast of Greenland. Some of the largest rates of falling pressure in the world (other than hurricanes and tornadoes) are recorded here. The deep, low centres that move along the west coast of Greenland are primarily responsible for the high winds that are recorded occasionally in that area. At times, winter temperatures in the Attic are unusually high. This situation is brought about by deep, low centres moving into the Arctic, coupled with compression of air (the Foehn effect) as it often blows down off the sloping edges of the ice caps, primarily the Greenland ice cap.

95. **Arctic Air Masses**. The moisture content of air masses that originate over land is low at all altitudes in the winter. The distinction between air masses almost disappears during the summer because of the nearly uniform surface conditions over the arctic and

sub-polar regions. The frozen surface thaws under the influence of lengthened or continual daylight, the snow melts from the glaciers and pack ice, the ice melts in the lake areas in the Arctic, and the water areas of the polar basin increase markedly. Thus, the polar area becomes mild, humid, and semi-maritime in character. Temperatures are usually between freezing and 50°F. Occasionally, strong disturbances from the south increase the temperature for short periods. Daily extremes, horizontal differences, and day-to-day variabilities are slight.

96. During the winter months, air masses are formed over areas that are completely covered by ice and snow. The air masses are characterised by very cold surface air and a large temperature inversion in the lowest few thousand feet. Since the amount of moisture the air can hold depends on the air temperature, the cold arctic air is very dry (low absolute humidity). The air mass that originates over oceans does not have a surface temperature inversion in the winter, the surface air temperature is warmer, and there is a corresponding increase in the moisture content of the air. It is during movement inland of moist air from the warmer waters that most of the rather infrequent arctic cloudiness and precipitation occurs during this season.

97. During the summer months, the large expanse of open water and warmer temperatures in the Arctic result in increased moisture. Consequently, the largest amount of cloudiness and precipitation occurs during these summer months.

98. **Arctic Fronts**. The weather associated with fronts in the Arctic has much the same cloud structure as with polar fronts, except that the middle and high cloud types are generally much lower, and the precipitation is usually in the form of snow.

99. Periods of maximum surface wind usually occur during and just after a frontal passage. This strong wind flow often creates hazards, such as blowing snow and turbulence, which make operational flying difficult.

100. The best flying weather in the Arctic over land usually occurs in midsummer and midwinter; the worst (low ceilings and visibilities) is during the transitional periods between the two seasons. Winter is characterised by frequent storms and well-defined frontal passages, but because of the dryness of the air, cloudiness and precipitation are at a minimum. In the summer, there are fewer storm passages and fronts are weaker; however, the increased moisture in the air results in more widespread clouds and precipitation. Over the sea areas the summer weather is very foggy, but winds are of lower speeds than in the winter.

101. During the transitional periods of spring and fall, operational flying conditions are usually the worst. Frontal systems are usually well defined, active, and turbulent. Icing may extend to high levels.

102. **Temperatures in the Arctic**. Temperatures in the Arctic, as one might expect, are very cold most of the year. But contrary to common belief, the interior areas of Siberia, northern Canada, and Alaska have pleasantly warm summers with many hours of sunshine each day. There are large differences in temperature between the interior and coastal areas.

103. In the interior during the summer days, temperatures climb to the mid 60s or low 70s and frequently rise to the high 70s or low 80s, occasionally even into the 90s. Fort Yukon, Alaska, which is just north of the Arctic Circle, has recorded an extreme high temperature of 100°F, while Verkhoyansk in north central Siberia has recorded 94°F.

104. During the winter, the interior areas of Siberia, northern Canada, and Alaska act as a source region for the cold arctic air that frequently moves southward into the middle latitudes. The coldest temperatures on record over the Northern Hemisphere have been established in Siberia. In the northern areas of the interior regions, temperatures are usually well below zero during the winter months. In fact, during these long periods of darkness and near darkness, the temperature normally falls to –20°F or –30°F and in some isolated areas the normal daily minimum temperature may drop to – 40°F. In north central Siberia the normal minimum daily temperature in the winter is between –45°F and –55°F. The arctic coastal regions, which include the Canadian Archipelago, are characterised by relatively cool, short summers. During the summer months the temperatures normally climb to the 40s or low 50s and occasionally reach the 60s. There is almost no growing season along the coasts, and the temperatures may fall below freezing during all months of the year. At Point Bar-row, Alaska, the minimum temperature rises above freezing on no more than about 42 days a year.

105. Over the Arctic Ocean, the temperatures are very similar to those experienced along the coast; however, the summer temperatures are somewhat lower.

106. Winter temperatures along the Arctic coast are very low but not nearly as low as those observed in certain interior areas. Only on rare occasions does the temperature climb to above freezing during the winter months. The coldest readings for these coastal areas range between –60° and –70°F.

107. These figures may seem surprising, since at first one might think that the temperatures near the North Pole would be lower than those over the northern continental interiors. Actually the flow of heat from the water under the ice has a moderating effect upon the air temperature along the coast.

108. **Cloudiness**. Cloudiness over the Arctic is at a minimum during the winter and spring and at a maximum during the summer and fall, again due to the low-moisture capacity of cold air. The average number of cloudy days for the two 6-month periods on climatic charts shows a general decrease in cloudiness in the entire arctic area during the winter months. The greatest seasonal variation is found in the interior, and the least is found along the coasts.

109. During the warm summer afternoons in the interior regions, scattered cumulus form and occasionally develop into thunderstorms. The thunderstorms are normally widely scattered and seldom form continuous lines. Along the arctic coast and over the Arctic Ocean, thunderstorms occur infrequently. Although tornadoes have been observed near the Arctic Circle, their occurrence is extremely rare. In these areas, summers are quite cloudy, with stratiform clouds predominating.

110. Seasonal changes in cloudiness take place quite rapidly. Winters are characterised by extensive cloudiness in the coastal regions. These clouds are associated with migratory lows and generally disperse inland as the systems lose their moisture.

111. **Winds**. Wind speeds are generally light in the continental arctic interior throughout the year. The strongest winds in the interior normally occur during the summer and fall. During the winter, the interior continental regions are areas of strong anticyclonic activity that produce only light surface winds.

112. Strong winds occur more frequently along the arctic coast than in the continental interiors. The frequency with which these high winds occur in coastal areas is greater in the fall and winter than in the summer. These winds frequently cause blowing snow.

113. Very strong wind speeds have been observed at many arctic coastal stations. Strong winds are infrequent over the ice pack, but the wind blows almost continuously because there are no natural barriers (such as hills and mountains) to retard the wind flow. As a result, the combination of wind speed and low temperatures produces equivalent wind chill temperatures that are extreme and severely limit outdoor human activity.

114. **Precipitation**. Precipitation amounts are small, varying from 5 to 15 inches annually in the continental interior and 3 to 7 inches along the arctic coastal area and over the ice pack. The climate over the Arctic Ocean and adjoining coastal areas is as dry as some of the desert regions of the mid-latitudes. Most of the annual precipitation falls as snow on the Arctic Ocean and adjacent coastal areas and ice caps. On the other hand, most of the annual precipitation falls as rain over the interior.

115. **Restriction to Visibility**. TWO factors make the visibility in the polar regions a very complex matter. Arctic air, being cold and dry, is exceptionally transparent, and extreme ranges of visibility are possible. On the other hand, there is a lack of contrast between objects, particularly when all distinguishable objects are covered by a layer of new snow. Limitations to visibility in the Arctic are primarily blowing snow, fog, and local smoke. Local smoke is serious only in the vicinity of larger towns and often occurs simultaneously with shallow radiation fogs of winter.

a. **Blowing Snow**. Blowing snow constitutes a more serious hazard to flying operations in the Arctic than in mid-latitudes because the snow is dry and fine and is easily picked up by moderate winds. Winds in excess of 8 knots may raise the snow several feet off the ground, and the blowing snow may obscure surface objects such as runway markers.

b. **Fog.** Of all the elements that restrict flying in the Arctic regions, fog is perhaps most important. The two types of fog most common to the polar regions are advection fog and radiation fog. Fog is found most frequently along the coastal areas and usually lies in a belt parallel to the shore. In the winter, the sea is warmer than the land, and relatively warm, moist air is advected over the cool land causing fog. This fog may be quite persistent. In the summer, warm, moist air is

advected over sea ice, which is now melting, creating the same situation which is found over land in winter.

c. **Ice Fog.** A fog condition peculiar to Arctic climates is ice fog. Ice fog is composed of minute ice crystals rather than water droplets of ordinary fog and is most likely to occur when the temperature is about –45°C (–50°F) or colder but can occur when temperatures are as warm as –30°C (–20°F).

d. **Sea Smoke or Steam Fog.** The cold temperatures in the Arctic can have effects which seem peculiar to people unfamiliar with the area. During the winter months, the inability of the air to hold moisture results in an unusual phenomenon called sea smoke. This is caused by open bodies of comparatively warm water existing simultaneously with low air temperatures. Actually, this phenomenon is similar to that of steam forming over hot water. In the case of sea smoke, the temperatures of both the air and the water are quite low, but the air temperature is still by far the lower of the two, causing steam to rise from the open water to form a fog layer. This fog occurs over open water, particularly over leads (navigable passages) in the ice pack and is composed entirely of water droplets.

e. **Arctic Haze**. This is a condition of reduced horizontal and slant visibility (but good vertical visibility) encountered by aircraft in flight over arctic regions. Colour effects suggest this phenomenon to be caused by very small ice particles. Near the ground, it is called arctic mist or frost smoke; when the sun shines on the ice particles, they are called diamond dust.

116. **Arctic Weather Peculiarities**. The strong temperature inversions present over the Arctic during much of the winter causes several interesting phenomena. Sound tends to carry great distances under these inversions. On some days, when the inversion is very strong, human voices can be heard over extremely long distances as compared to the normal range of the voice. Light rays are bent as they pass through the inversion at low angles. This may cause the appearance above the horizon of objects that are normally below the horizon. This effect, known as looming, is a form of mirage. Mirages of the type that distort the apparent shape of the Sun, Moon, or other objects near the horizon are common under inversion conditions.

117. One of the most interesting phenomena in the Arctic is aurora borealis (northern lights). These lights are by no means confined to the Arctic but are brightest at the arctic locations. Their intensity varies from a faint glow on certain nights to a glow which illuminates the surface of the Earth with light almost equal to that of the light from a full moon. The reactions resulting in the auroral glow have been observed to reach a maximum at an altitude of approximately 300,000 feet.

118. The amount of light reflected from a snow-covered surface is much greater than the amount reflected from the darker surfaces of the middle latitudes. As a result, useful illumination from equal sources is greater in the Arctic than in lower latitudes. When the sun is shining, sufficient light is often reflected from the snow surface to nearly obliterate

shadows. This causes a lack of contrast which, in turn, results in an inability to distinguish outlines of terrain or objects even at short distances. The landscape may merge into a featureless grayish-white field. Dark mountains in the distance may be easily recognised, but a crevasse immediately ahead may be obscured by the lack of contrast. The situation is even worse when the unbroken snow cover is combined with a uniformly overcast sky and the light from the sky is about equal to that reflected from the snow cover. In this situation, all sense of depth and orientation is lost in what appears to be a uniformly white glow; the term for this optical phenomenon is whiteout.

119. Pilots have reported that the light from a half-moon over a snow-covered field is sufficient for landing aircraft at night. It is possible to read a newspaper on occasions by the illumination from a full moon in the Arctic. Even the illumination from the stars creates visibility far beyond what one would expect elsewhere. It is only during periods of heavy cloud cover that the night darkness begins to approach the degree of darkness in lower latitudes. In lower latitudes, south of 65° north latitude, there are long periods of moonlight, since the Moon may stay above the horizon for several days at a time.

Antarctic Weather

120. Many of the same peculiarities prevalent over the arctic regions are also present in the Antarctic. For instance, the aurora borealis has its counterpart in the Southern Hemisphere, called aurora australis. The same restrictions to visibility exist over the Antarctic regions as over the Arctic. Some other characteristics of the Antarctic regions are as follows:

a. Precipitation occurs in all seasons, with the maximum occurring in summer. The amount of precipitation decreases poleward from the coast.

b. Temperatures are extremely low. The lowest temperature in the world, –127°F, was recorded at Vostok, Antarctica. In the winter, temperatures decrease from the coast to the pole, but there is some doubt that this is true in the summer. The annual variation of temperature as indicated by Macmurdo station shows the maximum in January and the minimum in early September. A peculiar, and to date unexplained, feature of Antarctic temperature variations during the Antarctic night is the occurrence of maximum temperatures on cloudless days in the early hours after mid-night. On cloudy days, however, the day is warmer than the night.

United States Weather

121. The weather in the United States, with minor exceptions, is typical of all weather types within the temperate regions of the North American, European, and Asiatic continents. The general air circulation in the United States, as in the entire temperate zone of the Northern Hemisphere, is from west to east. All closed surface weather systems (highs and lows) tend to move with this west-to-east circulation. However, since

this is only the average circulation and weather systems move with the general flow, the fronts associated with the migratory lows also tend to move southward if they are cold fronts and northward if they are warm fronts. Surface low-pressure centres, with their associated weather and frontal systems, are referred to as cyclones. Knowledge of the mean circulation in the temperate region makes it possible to observe and plot average storm tracks and to forecast future movement with a reasonable degree of accuracy.

122. Certain geographical and climatic conditions tend to make specific areas in the United States favourable for the development of low-pressure systems such as west Texas, Cape Hatteras, central Idaho, and the northern portions of the Gulf of Mexico. Once a low has formed, it generally follows the same mean track as the last low that formed in that area. The average, or mean paths are referred to as storm tracks.

123. These storms (lows) are outbreaks on the polar front or the generation or regeneration of a storm along the trailing edge of an old front. The low pressure along these fronts intensifies in certain areas as the front surges southward ahead of a moving mass of cold polar air. Much of the weather, especially the winter weather, in the temperate zone is a direct result of these storms.

124. Airmass weather also affects temperate climates. Air-mass weather is the name given to all weather other than the frontal weather in the temperate region. Air-mass weather is the net effect of local surface circulation, terrain, and the modifying effect of significant water bodies.

125. There are many subdivisions of weather regions in the United States. For the purpose of this discussion, we have divided the continental United States into seven regions as indicated in fig 1.9.

Northwest Pacific Coast Area

126. The northwest Pacific coast area has more precipitation than any other region in North America. Its weather is primarily the result of frontal phenomena, consisting mainly of occlusions which move in over the coast from the area of the Aleutian low and orographic lifting of moist, stable maritime air. Predominant cloud forms are stratus and fog, which are common in all seasons. Rainfall is most frequent in the winter and least frequent in the summer.

Southwest Pacific Coast Area

127. The southwest Pacific coast area experiences a Mediterranean-type climate and is distinctively different from any other North American climate. This climate occurs exclusively in the Mediterranean and southern California in the Northern Hemisphere. In the Southern Hemisphere, it occurs over small areas of Chile, South Africa, and southern Australia.

128. This climate is characterised by warm to hot summers, tempered by sea breezes, and by mild winters during which the temperatures seldom go below freezing. Little or no rainfall occurs in the summer and only light to moderate rain in the winter.

129. Cold fronts rarely penetrate the southwest Pacific coast region. The weather over this region is due to the circulation of moist Pacific air from the west being forced up the slope of the coastal range. In the summer, air is stable, and stratus and fog result. In the winter, unstable air which is forced over the mountain ranges causes showers or snow showers in the mountains.

Intermountain West Central Area

130. The intermountain west central area includes the Great Plains region. This region is located east of the Cascade and coastal ranges, west of the Mississippi Valley, and north of the southwest desert area. The climate is generally cold and dry in the winter, and warm and dry in the summer. Most of the region is semiarid. The western mountain range, which acts as a climatic barrier, has an extreme drying effect on the air in the westerly circulation.

131. Maximum rainfall occurs in the spring and is due mainly to the predominance of cyclonic storm passages during this season. In midwinter a cold high is generally centered in this region which prevents the possibility of storm passages. Annual precipitation is normally light.

Southwest Desert and Mountain Area

132. The southwest desert and mountain area includes lower California and some of southeast California as well as the southern portions of Arizona, New Mexico, and Texas. It is an area almost completely surrounded by high mountains and is either very arid or actual desert. Annual rainfall seldom exceeds 5 inches. The more northerly sections have cold winters, and all parts have extremely hot summers. The chief flying hazard results from a predominance of summer and spring thunderstorms caused mainly by maritime tropical air being forced aloft at the mountains. For this reason nearly all significant peaks and ranges have thundershowers building over them in the spring and summer. The thunderstorms are generally scattered and are almost always severe; however, pilots can usually avoid them by circumnavigating them.

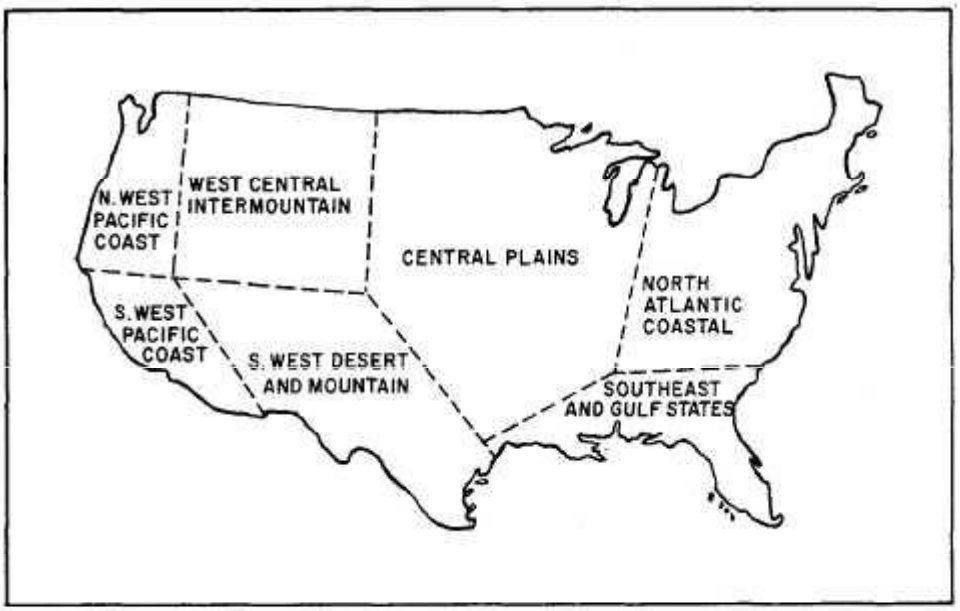


Fig 1.9. United States weather regions

Central Plains Area

133. The central plains area includes the continental climate regions of the Great Plains, Mississippi Valley, and Appalachian Plateau between the Rocky Mountains to the west, the Appalachians to the east, and the gulf states to the south. The western section is generally drier than the eastern section. The main weather hazards are caused by wintertime outbreaks and associated wave phenomena along polar fronts. Convective air-mass thunderstorms, which are prevalent over this area in summer, also pose a threat to flying. Frontal passages, both cold and warm, and associated weather are common in this area. Thunderstorms are usually of convective origin and are most violent if they have developed in maritime tropical air. This occurs often in the spring, and tornado activity becomes a climatic feature due to its frequency.

Southeast and Gulf States Area

134. The southeast and gulf states area includes all the states bordering on the Gulf of Mexico as well as South Carolina and Georgia. Stagnating south bound cold fronts, rapidly moving squall lines, air-mass thunderstorms, and stratus clouds occur in various combinations to make this area an especially complex one for the forecaster. Frontal passages can be expected only in the late fall, winter, and early spring. A circulation phenomenon known as gulf stratus affects this area. In the winter, when the circulation near the surface is southerly, the warm, moist gulf air is cooled from below to saturation. When this occurs, fog and the gulf stratus may form and may persist over the area for several days. The southerly circulation in summer causes warm, moist air to be heated from below, and convective thunderstorms are common. Since the air is generally quite moist and unstable, these storms are generally severe.

North Atlantic Coastal Area

135. The North Atlantic coastal area is an area of storm track convergence, and cyclonic storm activity is frequent in winter. Moreover, these storms are intensified by the heating and addition of moisture to the air over the Great Lakes. The lake effect is directly accountable for the large amounts of snowfall often found over this area in the winter. Generally good weather prevails in summer due to the predominant influence of the Bermuda high.

European Weather

136. Most of Europe has a relatively mild climate which is largely due to its oceanic exposure to the north, west, and south. The east-west orientation of the mountains in Europe normally prevents extremely cold arctic air from penetrating southward to the Mediterranean. As a result, very cold weather is limited to the northern limits. The southern coast and Mediterranean countries enjoy moderate temperatures year round because relatively warm maritime air masses move inland from the Atlantic and because of the moderating influence of the Mediterranean Sea. However, this inflow of maritime air also brings frequent cloudiness, considerable precipitation, and high humidity.

137. When continental air masses dominate, Europe is subjected to low-temperature extremes, low humidity, and clear skies much the same as North America. This is especially true north of the Alpine Mountains. South of this region, somewhat normal migratory patterns do exist. The end result is relatively dry summers and wet winters characteristic of the western coastal region of North America and Canada. Temperatures are highest in Europe during the summer; Athens, Greece averages 80°F; Granada, Spain 77°F; Greenwich, England 63°F; and Paris, France 65°F farther north, summer temperatures average as much as 20 to 25 degrees less. During the winter, the Mediterranean temperatures average in the upper 40°F to low 50°F range while the extreme northern sections average 10°F or less. The Atlantic coastal countries with their predominantly maritime climates maintain far less temperature extremes between summer and winter. Precipitation in the form of rain and drizzle is common along the European Atlantic coast and near the Mediterranean Sea. Snow does occur at times in areas east of Spain and north of the Mediterranean Sea. At higher elevations inland, snow is common and frequently abundant. Central Spain and southern Russia, by contrast, experience semiarid and arid climates.

Asiatic Weather

138. Asia's climate is predominantly continental. The only exceptions are the heavily populated coastal areas that have tropical and maritime climates during the summer. This primarily continental climate results in limited precipitation and large temperature ranges both daily and seasonally.

139. Asia is a huge continent with large expanses of land extending far northward. The Himalaya Mountains stretch across the southern portion in an east-west direction; mountains also parallel the eastern coast. These geographical features often contain continental arctic and polar air inland, resulting in the most extreme temperature ranges found in the Northern Hemisphere. Northeastern Siberia's temperatures often range from –60°F in the winter to above 60°F in the summer. Extremes range as high as 98°F and as low as –90°F. The large interior of Asia also results in extreme pressure difference. In the winter the continent is dominated by a cold high-pressure area and in the summer by

a warm low-pressure area. This accounts for the northeast winter monsoons and southwest summer monsoons.

140. In the winter the interior is dry, receiving less than 1 inch of precipitation. Coastal areas under maritime influence receive normal amounts (about 8 inches) of precipitation. In the summer, precipitation is plentiful except well inland. Rain is so abundant in some regions, such as India, that the yearly rainfall average (425 inches or more) is among the highest in the world.

141. The extreme south and southeast regions of Asia differ sharply from its northern neighbors. These southern regions enjoy the tropical and maritime climates that feature only minor seasonal temperature variations. Eastern Asia enjoys a climate very similar to that found along the eastern coast of North America from the Florida Keys to eastern Canada. East and Southeast Asia, like the eastern and southeastern United States, is also subject to an occasional tropical cyclone (typhoon) in the summer and in the fall.

South American Weather

142. South America has a variety of climates but lacks the severe weather of North America. Continental polar air does not exist here because the continent tapers sharply from north to south. The larger northern area is close to the equator and does not experience the influx of cold maritime polar air from the south. Tropical climates prevail over much of the continent. Yet, due to the high Andes Mountains along the western coast, there are areas that are extremely dry and others that are extremely wet.

Northeastern Climate

143. The South American northeast's climate consists mainly of high temperature and humidity and copious rainfall throughout the year. September is the warmest month with average temperatures of around 82°F. January is the coolest month with average temperatures of around 79°F. Nighttime temperatures rarely fall below 65°F. Rainfall averages 87 inches annually with 12 inches falling in June and just over 2 inches falling in October. The higher elevations of northeastern South America have greater ranges of temperature, humidity, and precipitation; however, these ranges are not extreme.

Southern Climate

144. In the southern region, below 20° south latitude, South America has distinct seasons very similar to those in the southeastern United States. These seasons, however, are reversed. The warmest month is January, which averages 74°F; July, which averages 49°F is the coolest month. Precipitation occurs fairly evenly throughout the year and averages 38 inches. There is no distinct rainy season.

145. Below 40° south latitude, the climate is progressively drier and cooler. However, the extreme southern tip of South America is characterised by year round cold and damp climate due to a strong maritime influence.

West Coast Climate

146. The west coast, from northern Peru to the middle of Chile, is a desert. North and south of this desert midsection, the climate is quite humid. The northwest coast has a typical tropical climate with wet and dry seasons. Below central Chile, the climate again shows atypical Southern Hemisphere seasonal reversal of that found in North America. The weather in this region is similar to that found along the northwest coast of North America. The climate is generally rainy and cool. Summer does not seem to exist as we know it. Yet, winter temperatures average above freezing.

African Weather

147. Africa's climate is unlike that of any other continent for several reasons. The most important is the fact that the entire continent is within the tropical zone. The equator bisects the continent; therefore, in the areas north and south of the equator, the climates are similar, yet they differ because the region north of the equator is much larger than the southern region. Since the northern area is so broad in the east-west direction, maritime effects inland are minimal. Also, an extensive low-pressure area develops inland due to extreme land mass heating. The southern section, however, with its maritime influences is dominated by a belt of high pressure during winter and by low pressure during summer.

148. Another factor is the cold currents which exist along its western shores. These currents allow an influx of cool winds and associated weather to the west coast. The final factor involves the lack of high mountain ranges common to other continents. Since there are no prominent mountain ranges, the various climate types in Africa blend together, showing no sharp distinctions.

149. The most important climatic element in Africa is precipitation. Precipitation is greatest near the equator (60 to 80 inches to over 120 inches in places). It decreases sharply to the north (less than 10 inches), and decreases gradually south of the equator (average of 20 to 40 inches). Because Africa is in the tropical zone, the precipitation belt of the inter-tropical convergence zone (ITCZ) moves with the seasons. This belt of precipitation moves northward in the summer and southward in the winter.

150. Africa does have distinct climatic regions. Air-mass movement and influences allow for a division of eight climatic regions.

Northern Region

151. The northern region includes the great Sahara desert. The desert is a source region for dry continental-type air masses. While maritime air may transit the area, the air masses are highly modified and often exhibit continental properties after moving inland. This desert region is extremely hot during the day throughout the year but is very cool at night due to a lack of moisture; hence, strong, radiational cooling.

Southwestern Region

152. The southwest region is an arid to semiarid area which is known as the Kalahari Desert. The temperatures are not as extreme as in the Sahara because the land area involved is much smaller.

North Central Region

153. The north central region is a semiarid area located along the edge of the Sahara. While the temperatures are similar to those of the neighboring desert (50°F in winter to well above 80°F in summer), this area occasionally gets precipitation in the winter. The source of this precipitation for the northern area is maritime air from the Mediterranean; in the south, it is the spotty rainfall provided by the meandering ITCZ.

Sub-Equatorial Region

154. The sub-equatorial region extends toward the equator from the semiarid region in the north. The region is marked by seasonal rainfall associated with the position of the ITCZ. The region is wet for about 5 months (Nov-Mar) and dry during the rest of the year. Temperatures show little seasonal variation (68°F to 86°F) because of the close proximity to the equator. The only exception to this temperature stability occurs in the western portion which, during the winter, is occasionally influenced by cool weather from the north.

Equatorial Region

155. The equatorial region includes the southwest tip of northern Africa and the region between 5° north and south latitudes, extending from the west coast to Lake Victoria. It is the wettest climate in all Africa. These areas have two distinct rainy seasons associated with the northward and southward movement of the ITCZ. Rainfall averages over 120 inches annually in some areas. Throughout the rest of the year, precipitation remains plentiful because of the influx of maritime air from the west. There are no significant mountains in the region to prevent this maritime air from migrating inland. Temperatures are moderate year round.

Southeast Coastal Region

156. The southeast coastal region has a humid subtropical climate. This region has rainfall all year (45 inches on the average) and temperatures remain generally moderate all year, ranging from an average maximum of 72°F in winter (July) to 89°F in summer (January).

Southeastern Interior Region

157. This region has a wet-and-dry type of maritime climate; however, it is considered temperate because of the lower temperatures common to the higher elevation.

Australia and New Zealand Weather

158. Australia has a generally mild climate with cool winters in the south and warm winters in the north. Summers are warm along the coasts and generally hot in the interior. Freezing temperatures are infrequent. Australia's climatic zones are relatively uncomplicated due to the lack of high mountain ranges.

159. The northern one-third of Australia is located within the tropical zone. The region has a rainy season that runs from January to April. Annual precipitation is greatest

(nearly 100 inches) in the extreme north and tapers off to the south and inland toward the semiarid interior. The interior, along the Tropic of Capricorn, is very hot and dry in the summer with average maximum temperatures at or above 90°F. In the winter, average maximum temperatures in some areas drop to 68°F.

160. The southern two-thirds of Australia is under the influence of the high-pressure belts of the Southern Hemisphere as well as of the migratory lows found farther southward. The southwest and southern portions of this region have rainy winters and near-drought conditions in the summer similar to the Mediterranean climate. Temperatures average 80°F in January and 55°F in July. The climate of the southeast corner is very similar to the southwest region except it experiences a shorter winter and less annual precipitation.

161. New Zealand is located southeast of Australia. It is a very narrow country with a southwest to northeast orientation and is exposed to the prevailing westerlies. Therefore, the climate is moderate and predominantly maritime with moderate precipitation occurring throughout the year. The northern part of New Zealand has a subtropical climate; however, winter frost and occasional snow can occur at locations farther south in highland areas. Fog is often widespread and very persistent over much of the country in advance of approaching frontal systems. Precipitation averages 49 inches in the northern half of the country and up to 170 inches in the southern half. Temperatures range from an annual average of 59°F in the north and 55°F in the central region to 50°F in the south.

TASK-33 : CLIMATOLOGY OF BANGLADESH

Chapter objectives

After reading this chapter, you should be able to:-

- Assimilate the Physical Features, Climatic Elements and different Climatic Regions of Bangladesh.
- Understand how the Pressure , Temperature, Rainfall and Humidity varies during different seasons in different regions.
- Understand the different seasons of Bangladesh and Aviation weather Hazards.

Structure

1. Introduction
2. Season in Bangladesh
3. Pre–Monsoon Season
4. SW Monsoon.
5. Post Monsoon.
6. Aviation weather Hazards

Introduction

I. Bangladesh is the darling child of the nature and it is situated at the bank of Bay of Bengal in the northern hemisphere. It lies between 20° 34' N and 26° 38' N Lat as well as 88 ° 01' E and 92° 41' E Long. The Tropic of Cancer, the very important geographical line on the globe, passes over central part of Bangladesh.

2. Conical shape of Bay of Bengal in the southern side, the Himalayas in the northern side and a vast body of landmass in the west-north-west side are responsible for copious rainfall in Bangladesh. Bangladesh has a typical monsoon climate with the same threefold division of the year that occurs in India, but the cool season is generally warmer than India. Generally, annual average rainfall varies from 1500 to 2500 mm, Although temperature during the hot season are rather lower than in some parts of India, the heat is made uncomfortable by the high humidity. This damp, muggy season continues throughout the main rainy season, but the heat is rarely dangerous. It is, however, very unpleasant for the unacclimatized visitor.

Seasons in Bangladesh

3. Like India, Bangladesh has got the following seasons:

a. NE Monsoon or Winter Monsoon.

b. Pre-Monsoon or Spring Transition.

c. SW Monsoon or Summer Monsoon.

d. Monsoon or Autumn Transition.

North East Monsoon or Winter Monsoon (December to February)

4. This season is characterised by an anticyclonic pressure system dominating the country except in February. Mist and Fog are the main weather in December and January. Wind blows from northwest or north direction. Cold waves blow from almost north to south over Bangladesh. Mean temperature ranges from 18-21º C in the northern and central districts and lowest temperature come down occasionally 4-5º C over these districts. Rainfall during winter is scanty but western disturbance may cause 3-10mm of rainfall over the northern districts.

5. The main weather conditions during this period are:-

a. Poor visibility due to mainly fog, mist and haze.

b. Low clouds associated with fog lifting.

c. Upper air jet.

6. In Bangladesh the advection-radiation fog is more common in the landmass other than coastal districts of Chittagong where advection fog is more common.

Pre-monsoon or Spring Transition (March to May)

7. During this season temperature increases rapidly due to normal sunshine approaches to our area from south to north. The mean temperature remains within 23°C to 30°C. April and May are the hottest months of the year. The highest temperature ranging from 44°C to 45°C is attained in the northern and northwest districts. Over rest of the country it ranges from 41°C to 43°C. Local severe storms Nor'wester, which move generally from northwest direction that are responsible for the welcome of precipitation.

8. Pre-monsoon or spring transition has got the following main weather activity:-

a. Norwester.

b. Tornado.

c. Tropical Cyclone.

Nor'wester

9. **Definition.** In the month of March, April and May, that is during pre-monsoon period thunderstorms that usually originate in West Bengal and Chotanagpur, some of which reach severe thunderstorm or tornado violence, are termed as Nor'wester because of their pro-tendency to move from north-west. In Bengali Nor'wester are popularly known as 'Kal Baishaki' due to their black appearance and period of formation in the Bengali month Baishak.

10. **Climatology of Nor'wester.** Maximum frequency of Nor'wester occurs in the month of March, and decreases in the month of April and May. In Bangladesh, Nor'wester or thunderstorm may occur in the middle of February due to early withdrawal of winter and may continue till the onset of monsoon. During this period the severe thunderstorm turns into tornado and causes havoc to the man and material, specially in the central parts of the country.

11. **Classification of Nor'wester.** According to the place of origin, movement and frequency Nor'wester can be classified as Type 'A', Type 'B', Type 'C' and Type 'D'. They can be differentiated as follows:-

a. Type 'A'. Type 'A' Nor'wester occurs simultaneously over a large area and generally in the afternoon. 70% to 80% Nor'wester are of this type. They usually start in West Bengal and Chotanagpur in the afternoon and proceed from northwesterly direction with a speed of 48-64 kph.

b. Type 'B'. They originate in the sub-mountainous districts of North Bengal and move southwards mostly during night and early morning. They move slowly at a speed of 16-32 kph (10-20 mph) to Meghna estuary from north. About 10% Nor'wester are of this type.

c. Type 'C'. They originate over hills of Nagaland, Manipur and Mizoram and travel westward. They are very rare and come from northeasterly direction.

d. Type 'D'. They are very similar to Type 'B' but their place of origin is near the Khashi hills. They move usually from north to south. They are about 5% only.

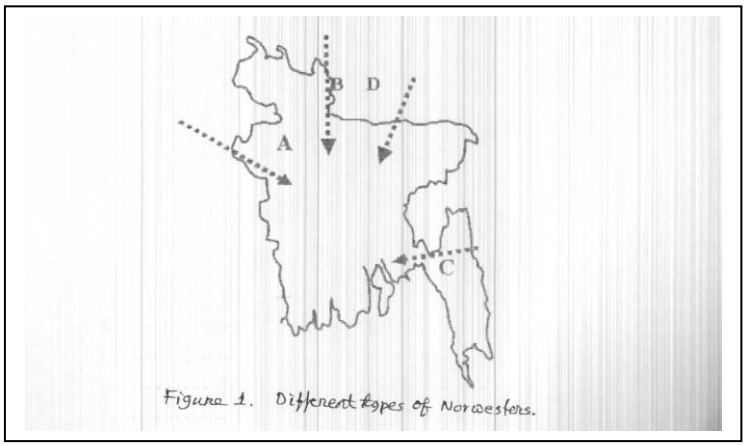


Fig 2.1

12. **Frequency of Nor'wester**. The frequency of Nor'wester with their duration is classified as:

a. In 30% cases, duration ranges from 1h to 1.5h.

b. In 20% cases, duration ranges from 1.5h to 2h.

c. In 40% cases, duration ranges from 2h to 4h.

d. In rest 10% cases, duration ranges more than 4h.

Tropical Cyclone

13. The tropical cyclones are the most destructive nature's phenomenon. Most of them are violent in nature and *few* are weak and die out before they intensify into a matured storm. Many tropical cyclones have visited Bangladesh since long and killed many people in the coastal areas. The recent tropical cyclones that visited Bangladesh as giant killer are November 12 Cyclone, 1970, May 25 Cyclone, 1985, November 29 Cyclone, 1988 and April 29 Cyclone, 1991. The most destructive one hit Cox's Bazaar on 29 April 1991 with a wind speed of 225 km/hour and maximum water surge of 20 feet. The cyclone killed 1,38,00,00 people and damaged many valuable physical infrastructures including very costly aircraft in BAF Base Zahurul Haque, Chittagong. In the recent years, the awareness of mass people about cyclone and development in forecasting cyclone signals has saved many people and material.

14. Tropical cyclone is called as killer weather in Bangladesh. Post monsoon tropical cyclones affects more than that of Pre-monsoon. The track of tropical cyclones in the month of May is as shown in the Fig 2.2.

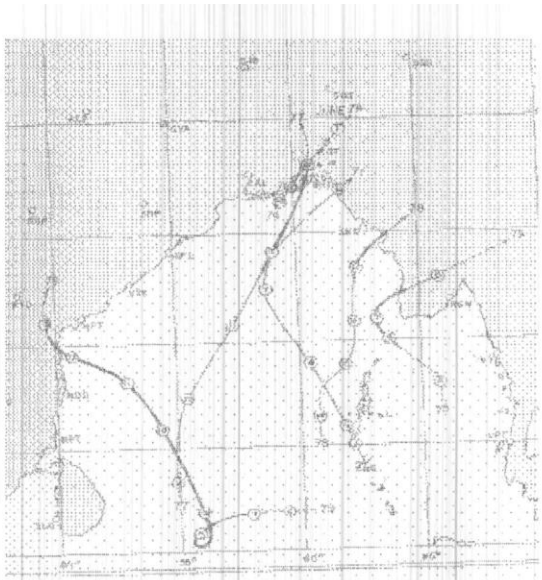


Figure 2: The track of tropical cyclone over Bay of Bengal during Pre-monsoon season.

Fig 2.2

Southwest Monsoon or Monsoon (June to September)

15. During the month of June the surface wind changes to southerly direction over the southern and central district and south-easterly to easterly over northern district of Bangladesh. Maximum rainfall occurs in this season. The monsoon usually sets in the south eastern corner of Bangladesh on 1st June and gradually approaches to the northwest direction with heavy to very heavy rain. During the first two months the rainfall is usually 700mm - 850mm per month over Sylhet and south eastern districts, 450mm - 600 mm per month over northern and southern districts, and the rainfall intensity is less over western part of Bangladesh compared to other regions. Monsoon depressions cause the copious rainfall over Bangladesh.

Onset and Withdrawal of Southwest Monsoon over Bangladesh

16. The onset and withdrawal dates of summer monsoon over Bangladesh are shown:

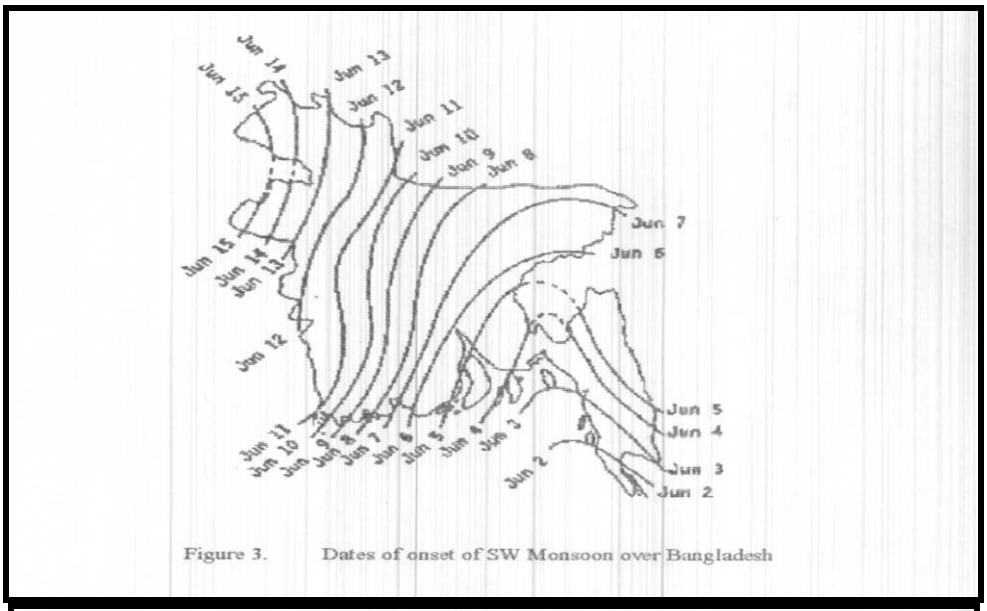


Figure 3. Dates of onset of SW Monsoon over Bangladesh

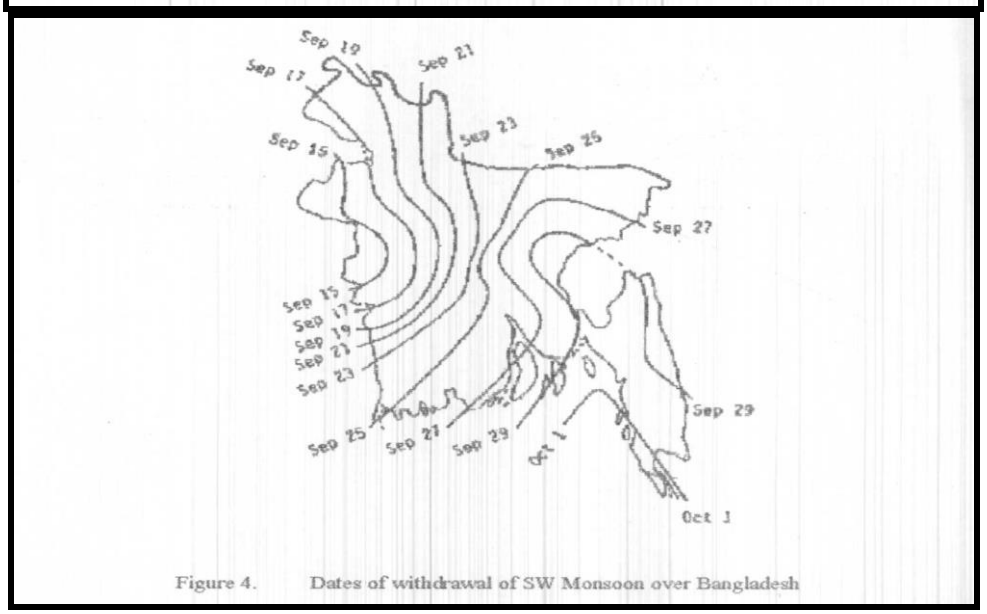


Figure 4. Dates of withdrawal of SW Monsoon over Bangladesh

Fig 2.3

Weather in Southwest Monsoon

17. The weather in SW Monsoon is mainly:

a. Poor visibility due to heavy to very heavy rain.

b. Low ceiling with maximum amount of 6-8 oktas at 400 ft or below.

c. Strong wind in coastal districts with speed of 30-40 kts.

Aspects of Monsoon

18. The following three features are very important in respect of SW Monsoon:

a. Break monsoon which causes flood over northern part of the country.

b. Monsoon depression which provides sufficient amount of rainfall over south part of the country, this part would have insufficient rainfall.

c. Cyclones do not form during the south west monsoon period.

d. Retreat mainly accompanied by northerly or northwesterly wind.

Post monsoon Season or Autumn Transition (October to November)

19. It is the transitional period from summer to winter. During this period rainfall, temperature and speed of wind decreases while atmospheric pressure increases. During October, Sylhet gets 200-250 mm of rain and rest of the country gets 100-170 mm of rain. The mean temperature falls 28-29° C in November.

20. The following are the main weather phenomenon during the Post Monsoon or Autumn Transition:

a. Tropical Cyclone.

b. Thunderstorms.

c. Poor visibility with the fog, mist and haze.

Tropical Cyclone

21. Both the Pre-Monsoon and Post-Monsoon period are the favourable for the formation of tropical cyclones over Bay of Bengal. However, the topical cyclones that forms in Post-monsoon periods affects much than that of Pre-monsoon period. The number for storms and tropical cyclones formed in the Bay of Bengal for the period 1891-1998 are shown in Table-1. It also shows the intensification of storms into tropical cyclones in the month of November and second maximum in October.

Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
16	3	6	29	76	114	150	182	172	173	148	69
5	1	4	21	43	38	41	29	35	71	111	35

Table 1. Storms and cyclones over Bay of Bengal for the period of 1891-1998

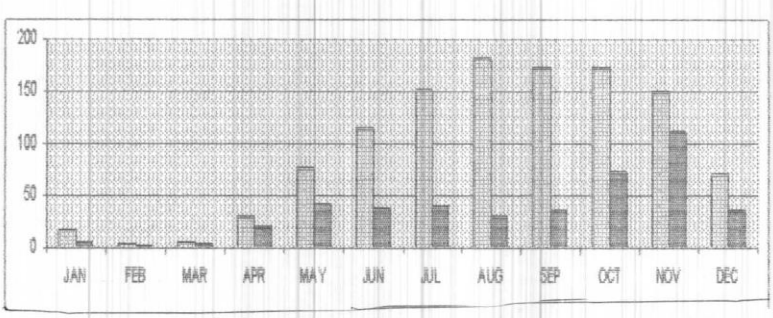


Table 2: Bar graph of storms and cyclones over Bay of Bengal for the period of 1891-1998.

Fig 2.4

22. During the Post-monsoon period the intensification of storms into tropical storms are more than that of Pre-monsoon season, and more so Bangladesh is mostly affected during the Post-monsoon period specially the Chittagong regions. The track of tropical cyclones in the Bay of Bengal during Post-monsoon period is shown in Fig 2.5.

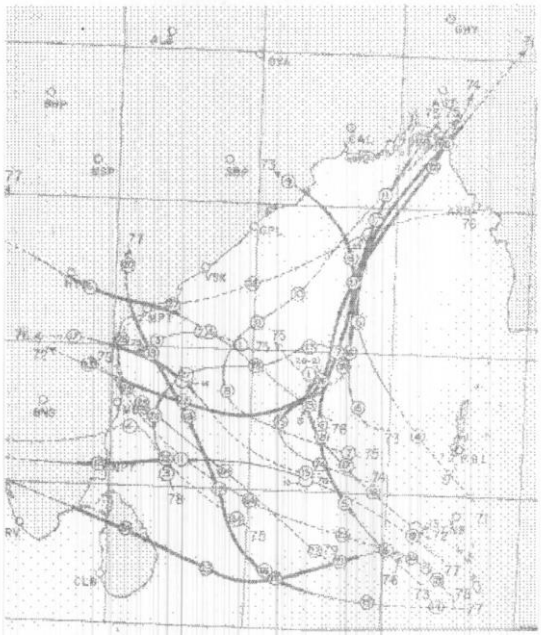


Figure 5. The track of tropical cyclone over Bay of Bengal during Post-monsoon season

Fig 2.5

Weather and Bangladesh

23. Bangladesh has one of the worst records in the world for natural disasters. Being a small country, it has all the variability and peculiarity of weather activities. The killer weather is tropical cyclones. Tornado and thunderstorm also bring disaster for the country. Hundreds of thousands of people have been killed by floods and cyclones, as well as tidal surges, tornado, droughts and even cold spells. Continuous heavy rainfall for few weeks especially in the south-eastern part of the country is very common. Poor visibility condition with visibility less than 1000m is usual in the morning of winter months over all the country.

Aviation Weather Hazards

24. The following are the aviation weather hazards in Bangladesh:-

a. Low ceiling during SW Monsoon especially over Chittagong regions.

b. Poor visibility during winter monsoon and Post-monsoon season almost over whole country.

c. Strong/gusty/squally wind with Nor'wester during Pre-monsoon season and strong wind with SW monsoon period over southeastern parts of the country.

d. Thunderstorms and hail storms over central and western part of the country.

TASK-34 : CLIMATOLOGY OF TIBET & HIMALAYAS

Chapter objectives

After reading this chapter, you should be able to:-

- Assimilate the Physical Features, Climatic Elements and different Climatic Regions of Tibet & Himalayas.
- Understand how the Pressure , Temperature, Rainfall and Humidity varies during different seasons in different regions of Tibet & Himalayas.
- Understand the different seasons of Tibet & Himalayas and Aviation weather Hazards.

Introduction

1. Politically autonomous Tibet is the land of Buddhist people, as the monastery of Dalai Lama is situated in Tibet. Tibetan Plateau is famous among the meteorologists also as it has a very prominent role in governing the climate of south Asia. The plateau is surrounded by large mountain ranges. The availability of data over the region is very meager. The primary impact of Tibetan Plateau is on thermodynamics of the atmosphere. In addition to the thermal influence, this plateau exerts major mechanical blocking influence on both the zonal and the meridional flow, which results in strong horizontal and vertical perturbations in the atmospheric flow. The plateau is also the major region of maximum dissipation of atmospheric kinetic energy and momentum. It modifies the intensity and speed of synoptic scale systems moving near the plateau.

Physiography

2. On the northern slope of Central and Eastern Himalayas lies the Tibetan Plateau, the highest and the most extensive plateau of the world covering an area of more than one million square kilometer. Tibet is the highest plateau in the world with average elevation of 5,000 m above sea level. The Chinghai-Tibet Plateau covers Chingha province, including Chamdo region, portions of Western Szechwan and Western Yunnan from the northern part of the Himalayas.The surface of Central Asia consists of two primary elevations separated by a trough like depression. The northern elevation is the Tienshan Plateau with an average elevation of 3,300 m and the southern elevation is the Tibetan Plateau. The intervening depression is the Tarim Basin. The Kunlun Mountain Systems form the northern boundary of the Tibetan Plateau, while the Central and Eastern Himalayas form its southern boundary. Some famous rivers like Brahmputra, Mekong, Yantze and Hwang Ho originate from Tibet. Nair Singh range, Bayan Kara Sangh and Naan Ling Shan are few ranges on the plateau.

Climate

3. It is a region of extreme climate, with large seasonal and diurnal variations of weather elements. Near the ground, Tibet has lowest temperature, lowest pressure, smallest absolute humidity and greatest wind speeds when compared to other regions of the earth's surface at the same latitude. In free atmosphere at the same height, Tibet is marked by highest temperature, lowest pressure and maximum absolute humidity in summer, highest pressure in winter and minimum wind speed in annual average compared to the free atmosphere conditions at the same height over the other regions in the same latitudinal belt. Following are some of the parameters of the climate of Tibet:

a. **Rainfall.** Tibetan Plateau is generally classified as an arid or semi-arid region. Along the slopes of the plateau, particularly along the southern slopes, rainfall is plentiful. In the western and northern sectors of the plateau, arid and semi-arid climates prevail. In the interior plateau, the annual amounts of rainfall rapidly decrease from 60 cm in the eastern part to 5 cm in the western part. Rainy season extends from April to October but most of the rainfall occurs during the period from June to August. Each year, about two to three heavy rain spells are observed over the plateau.

b. **Planetary Boundary Layer.** The PBL of Tibetan Plateau is the highest in the world by virtue of the height of the plateau. The thermal Low of summer and the cold High of winter over the plateau are formed in this layer and these constitute the world's highest PBL pressure systems with usual thermodynamic characteristics. The summer thermal Low may reach 400 hPa. This may be compared with thermal Low in North Africa reaching only to 850 hPa level and in Iraq reaching 700 hPa level. Over the Indian Plains, the thermal Low reaches to 500 hPa. During winter, the cold High over the Tibetan Plateau may reach to 500 hPa. These boundary layer systems become important sources for transients in the pressure and wind systems of the middle and lower troposphere in the northern hemispheric sub-tropical belt.

c. **Thunderstorms.** During the year, there are more thunderstorm days (about 90) over the Tibetan Plateau than elsewhere in this zonal belt, as compared to 5-10 days in Iran & North Africa, 10-20 days over the Atlantic, 40-60 days in China & North America and 5-10 days over the Pacific. Also, there are more hail days, about 55 annually. As many as 64 hail days were observed in the year 1957.

d. **Seasonal Oscillation of the Sub-Tropical Westerly Jet Stream.** During the northern summer, the sub-tropical westerly jet stream is relatively weak and lies entirely to the north of the plateau, easterlies prevailing to the south of the plateau in the upper troposphere. During the northern winter, it shifts southwards and intensifies considerably. Before reaching the plateau from the western side, it splits into two branches; the primary stronger branch running south of the plateau and the secondary, weaker branch running north of the plateau and sometimes getting mixed up with the polar front jet stream.

Seasons

4. In a year, Tibet gets following four seasons:

a. Winter (Dec, Jan, Feb)

b. Spring (Mar, Apr, May)

c. Summer (Jun, Jul, Aug, Sep)

d. Autumn (Oct, Nov)

5. Winter & summer are the main seasons:

a. **Winter**:

(1) **Pressure Pattern**. As the Siberian High establishes itself to the north of Tibet, steep pressure gradient is seen over Tibet with pressure values 1020-1026 hPa.

(2) **Upper winds**. Upper winds are westerlies with a speed reaching 75 Kts at 200 hPa level (Jet core is to the south of Tibet i.e. over India).

(3) **Weather.** This is the coldest season of the year. During this season temperature reaches up to -10˚c. Precipitation occurs in the form of both rain and snow due to passing western disturbances. Visibility deteriorates due to occurrence of fog. Following are some of the parameters:

PARAMETER	DEC	JAN	FEB
Mean Temperature	-01.9	-02.3	0.80
Max Temperature	15.3	06.8	09.2
Min Temperature	-09.1	-10.2	-06.9
Relative Humidity	35.0	28.0	27.0
Rainy Days	0.20	0.50	0.50
Snow Days	0.40	0.50	0.50

b. **Spring**. This is just a transition season between winter and summer season. Min temp in mar is still -3˚c but in May it touches 12˚c. Rainfall is more in comparison to winters. Thunderstorm & hail occur during this season. As the season progresses, the amount of rainfall in a month increases. Following are some of the parameters observed during this season:

PARAMETER	MAR	APR	MAY
Mean Temperature	04.3	08.3	12.6
Max Temperature	12.0	15.7	19.7
Min Temperature	-03.2	0.90	05.1
Relative Humidity	30.0	35.0	40.0
Rainy Days	01.8	03.7	07.7
Snow Days	01.8	02.2	01.3

c. **Summer:**

(1) **Pressure Pattern**. During summer, as the heat low forms over central Pakistan and adjoining Rajasthan, over Tibet also low pressures are observed. Pressures are of the order of 1000 – 1002 hPa.

(2) **Upper Winds**. Due to northward movement of STR, anticyclone over Tibet is observed. Tibetan High is a large anticyclone that is known to have its largest amplitude near 200 hPa during the summer season.

(3) **Weather.** In summer, the max temp reaches up to 23˚c. During this season, due to SE Monsoon over China, Tibet gets maximum rainfall especially over its eastern regions and the slopes. Following are some of the parameters observed during this season:

PARAMETER	JUN	JUL	AUG	SEP
Mean Temperature	15.5	14.9	14.1	12.8
Max Temperature	22.5	21.7	20.7	19.6
Min Temperature	09.2	09.9	09.4	07.6
Relative Humidity	53.0	67.0	70.0	65.0
Rainy Days	14.4	20.4	21.2	13.8
Snow Days	0.00	0.00	0.00	0.10
Wind	W'ly	E'ly	E'ly	E'ly

d. **Autumn.** This season is again a transition between summer and winter season. Amount of rainfall starts reducing. During this season, temperatures start getting lowered reaching up to even – 5˚c sometimes in November. Following are some of the parameters observed during this season:

PARAMETER	OCT	NOV
Mean Temperature	08.1	01.9
Max Temperature	16.4	11.6
Min Temperature	01.4	-05.0
Relative Humidity	48.0	37.0
Rainy Days	02.9	0.60
Snow Days	0.90	0.70
Wind	E'ly	E'ly

Climatology of Himalayas

6. The Himalayan mountain range is the highest at the same time youngest in the world. It evolved in the orogenesis of the tertiary period only about 70 million year ago. During tertiary period there were three main continents with the Tethys Sea spanning the equatorial regions. The disruptions of Gondwanaland give rise to the peninsular India. At the same time the overloaded heavily sediment bed of Tethys Sea got uplifted into the lofty Himalayas.

7. Study of climate of Himalayas is important because Himalayas serve as a great barrier to the cold continental air. In monsoon months Himalayas drive the rain bearing winds towards the Indian plains. Himalayas represent political boundary.

8. **Physiography**. The geographical limits of the region lay approximately between latitude 27°N and 36°N and longitude 72°E and 100°E. The area lies in the temperate zone. It consists of mountains and plateaus. On the northern side of these mountains lies Tibetan Plateau. The loftiest mountain ranges of the world radiate from the Pamir Knot near the intersection of latitude 38°N and longitude 74°E. It extends over 3000km with a varying length of 250 to 300km

9. To the north of Himalayas and almost parallel to it is the Karakoram range running SE wards from Pamir Knot through the northern parts of Kashmir. Important mountains and peaks are Mt.everest (8848m) K2 (8611) Kanchenjunga (8585m) Daulagiri (8172) Nangaparbat (8126), Nandadevi (7817).

10. The important rivers are the Indus, the Ganges, and the Brahmaputra. The Indus and Brahmaputra start from the Kailash Mountain in Tibet. The Ganges and its tributaries run from the central Himalayas and flow in southerly and easterly direction through the plains of north India.

11. **Climatological Divisions**. Himalayas are broadly divided into following three divisions.

a. Western Himalayas

b. Central Himalayas

c. Eastern Himalayas

Each region, according to elevation is further divided in to three ranges

a. Lower range (below 5000ft)

b. Outer range (5000ft-10000ft)

c. Inner range (above 10000ft)

12. The study of the climate of Himalayas is hampered due to following reasons; remoteness of areas makes it difficult to have weather stations, nature of mountain terrain restricts the local weather of any station to represent only a very small area, difficulties in making standard observations at mountain stations summit, slope and valley bottom.

13. Climatologically the year can be divided into four seasons as follows:-

a. Winter season

b. Spring season

c. Monsoon season

d. Autumn season

Winter

14. The period is characterized by the fall in temperature and a corresponding rise in pressure. The winds are W'ly to NW'ly, which bring in cold dry continental air. The northern portion is extremely cold and dry. The winds above three km are Wly and increase with height. Wly jet stream lies over India at 12 km. severe turbulence occurs over peaks/ ridges in afternoon hours

Western Himalayas

15. **Lower Ranges**. Men daily max temperature varies between 10 – 22°C. There is a marked fall of 2 -4°C in temperature in Jan. Thereafter the temperature gradually increases with the advance of season. The increase in temperature is quite conspicuous in March and is of the order of 4- 6°C. The seasonal and the monthly variation are between 1- 6° C.

16. **Outer Ranges**. Mean daily max temperature varies between 5 and 14°C. The arrival of winter is marked by a fall of 1 - 4°C in max temperature. The departure is indicated by a rise in temperature of 5 -7°C in March. Monthly variation is between1 and 2°C.

17. **Inner ranges**. It varies between 5 to -8°c .The fall in temperature from Dec to Jan is between 3-4°C.The seasonal variation is of the order of 4-8°c

Minimum Temperature

18. Mean daily min temperature in the lower ranges varies between 2-11°C. The temperature falls by 1-2°C from Dec to Jan and then increases with the advance of the season. In the outer ranges the mean daily min temp varies between 6 to-4°C. Murree and Srinagar experience sub freezing temperature in Jan and Feb. The seasonal variation is of the order of 4-7°C. The inner ranges experience sub-freezing mean daily temp throughout the season. It varies between -6 to -23°C.

Rainfall

19. The distribution of rainfall over the ranges is quite complex because of the complexity of the relief. Over the lower ranges the mean monthly rainfall varies between 2and14cm per month. As the western disturbances move northeasterly direction, they give more rainfall on the ranges facing west. Rainfall on all the ranges shows an increase with the advance of the season from Feb to March .The variation in the mean monthly rainfall over the outer ranges is less than that over the lower ranges; it varies 3-12 cm per month. The rainfall is min in Dec. This is due to lower frequency of western disturbances in Dec. Over the inner ranges the Leah is completely shaded from the winter rains and gets only 3 cm of rain during this season. The total rainfall for the season is less than 4cm.

Thunderstorm

20. Most of the thunderstorms occurring during this season are associated with the western disturbances and occur in the cold front. Thunderstorm does not occur over the lower ranges in Dec and Jan. in Feb and March the frequency of thunderstorms is about 1-2 per month. Outer ranges experience more days of thunderstorms with the exception of Srinagar valley .the frequency increases with the advance of season. Thunderstorm does not occur over the inner ranges during the winter season.

Fog

21. Over the outer ranges fog occur more frequently than the lower ranges. The frequency of occurrence of fog is more in Jan and Feb as compared to that in Dec and Mar this is due to the lower temperature that prevail in Jan and Feb.

Central Himalayas

22. **Lower ranges**. The mean daily max temperature varies between 15-27°C. The fall in temperature from Dec to March is of the order of 1-2°C. The seasonal variation is between 6-8°cand monthly variationbetween1-5°C.

23. **Outer ranges**. The mean daily max temperature varies between 10-18°C in Dec 8-17° in Jan, 9-19° in Feb and 13-24° in March .the seasonal variation is of the order of 4-7°C.

24. **Inner ranges**. Over inner ranges the data is scanty. Walachumggola experience max temperature of 14°c in Dec, 6°c in Jan, 7°c in Feb and 9°c in March. The seasonal variation is of the order of 8°c.

Minimum Temperature

25. Mean daily min temp over the lower ranges varies between 3-10°C in Dec, 2-9° C in Jan, 4-10°C in Feb and 7-13°C in March. Lowest min is observed in Jan .Seasonal variation is of the order of 4-7°C.In outer ranges it shows wider variation, varies between -2 to 11°C. Sub-freezing min temperature occur over the inner ranges. Walchumggola experiences -03°C.

Rainfall

26. Rainfall in this region is derived from the passing western disturbances. Over the lower ranges the rainfall varies from 1-7cm per month there is a tendency for mean monthly rainfall to decrease from west to east. Over the western ranges the western half experiences more rainfall than the eastern half. The rainfall over the inner ranges varies between 2and 5 cm per month max rainfall over this region occur in March

Thunderstorm

27. Thunderstorm occurring during this season is mostly associated with the western disturbances which move across the area. The frequency of occurrence increases gradually with the advance of the season and reaches the maximum in March. The outer ranges in the western half of the region seem to be more susceptible for the occurrence of thunderstorms. The frequency of occurrence varies from 1-6 per month. Thunderstorms do not occur over the inner ranges.

28. Fog over the lower ranges Tehri gets 6 days of fog in Jan and 2 days each in Feb. and March. The outer ranges get fog more frequently. The frequency of occurrence of fog over these regions varies between 1-4 in Dec 3-8 in Jan 1-6 in Feb and 1-5 in Mar.

Eastern Himalayas

29. The max temperature over sub Himalayan area in this season is about 15-20°C and over outer Himalayas it is 5-20°C and that over the inner Himalayas is 0-5°C. In all the ranges January appears the month when the max is lowest.

Minimum Temperature

30. It is seen that the zero deg isotherm in all the month passes through the outer Himalayan region and it is least in Jan. Coldest region in all the ranges in this season is the month of Jan.

Rainfall

31. This is the season of least rainfall over this area. The WDs passing through north or across cause rainfall over the area. It is seen that rainfall amount decreases towards inner Himalayas. Very little rainfall is recorded all along the eastern Himalayas in Dec and Jan. a definite increase in rainfall.

Thunderstorm

32. From the isolines of number of days of thunderstorm for Dec and Jan it is observed that thunderstorm do not occur over the Himalayas or on its northern side during these months. But the frequency of thunderstorm for Feb is 1 to 2 and that for March is 2-5 over sub-Himalayan and outer Himalayan area. Inner Himalayas are still free from any thunderstorm during these two months also. Thunderstorm of March is of destructive nature and is often associated with heavy fall of hail.

Fog

33. The fog over the Himalayas is associated with the western disturbances. Frequency of no of days of fog in different months of season is about 1 to 5 days over all the ranges of Himalayas

Spring

34. This is the period of transition from winter to summer. The pattern of winds and pressure undergoes a remarkable change. The temperature gradually increases during this season and reaches max in May or June. Insolation is intense and sets up convection currents and instability in the atmosphere.

Southwest Monsoon

35. This period is from June to Sept. monsoon sets over eastern Himalayas in first week of June and extends to western Himalayas by July winds are generally easterly as normal position of AMT is south of Himalayas. During break, AMT shifts to the foothills of Himalayas causing heavy rain fall. Long break causes extensive rainfall over upper catchments area of river originating from Himalayas sometimes lead to flood. Winds move north of 35°N. Recurvature of monsoon depression cause heavy rain in lower altitudes and heavy snowfall over higher latitude.

Conclusion

36. Tibet is region of extreme climate, with large seasonal and diurnal variations of weather elements. It plays major role in affecting the monsoon over Indian Subcontinent. The major seasons are summer and winter. June is the hottest month and January is the coldest month. In comparison to other regions of same latitude, Tibet has lowest temperature near ground. It has highest temperature in free atmosphere in summer. It receives more number of thunderstorms also.

TASK-35 : CLIMATOLOGY OF INDIA

Introduction

1. The **climate of India** defies easy generalisation, comprising a wide range of weather conditions across a large geographic scale and varied topography. Analysed according to the Köppen system, India hosts six major climatic subtypes, ranging from desert in the west, to alpine tundra and glaciers in the north, to humid tropical regions supporting rainforests in the southwest and the island territories. Many regions have starkly different microclimates. The nation has four seasons: winter (January and February), summer (March to May), a monsoon (rainy) season (June to September), and a post-monsoon period (October to December).

2. India's unique geography and geology strongly influence its climate; this is particularly true of the Himalayas in the north and the Thar Desert in the northwest. The Himalayas act as a barrier to the frigid katabatic winds flowing down from Central Asia. Thus, North India is kept warm or only mildly cold during winter; in summer, the same phenomenon makes India relatively hot. Although the Tropic of Cancer—the boundary between the tropics and subtropics—passes through the middle of India, the whole country is considered to be tropical. As in much of the tropics, monsoonal and other weather conditions in India are unstable: major droughts, floods, cyclones and other natural disasters are sporadic, but have killed or displaced millions. India's long-term climatic stability may be further threatened by global warming.

History of Indian Climate

3. During the Late Permian (some 260–251 Ma), the Indian subcontinent was part of the vast supercontinent Pangaea. Despite its position within a high-latitude belt at 55–75° S (as opposed to its current position between 5 and 35° N), latitudes now occupied by Greenland and parts of the Antarctic Peninsula, India likely experienced a humid temperate climate with warm, frost-free weather, though with well-defined seasons.[2] Later, India joined the southern supercontinent Gondwana, a process beginning some 550–500 Ma. During the Late Paleozoic, Gondwana extended from a point at or near the South Pole to near the equator, where the Indian craton (stable continental crust) was positioned, resulting in a mild climate favourable to hosting high-biomass ecosystems. This is underscored by India's vast coal reserves—much of it from the late Paleozoic sedimentary sequence—the fourth-largest reserves in the world. During the Mesozoic, the world, including India, was considerably warmer than today. With the coming of the Carboniferous, global cooling stoked extensive glaciation, which spread northwards from South Africa towards India; this cool period lasted well into the Permian.

Tectonic Movement

4. Tectonic movement by the Indian Plate caused it to pass over a geologic hotspot—the Réunion hotspot—now occupied by the volcanic island of Réunion. This resulted in a massive flood basalt event that laid down the Deccan Traps some 60–68 Ma, at the end of the Cretaceous period. This may have contributed to the global Cretaceous-Tertiary (K-T) extinction event, which caused India to experience significantly reduced insolation. Elevated atmospheric levels of sulphur gases formed aerosols such as sulfur dioxide and sulfuric acid, similar to those found in the atmosphere of Venus; these precipitated as acid rain. Elevated carbon dioxide emissions also contributed to the greenhouse effect, causing warmer weather that lasted long after the atmospheric shroud of dust and aerosols had cleared. Further climatic changes 20 million years ago, long after India had crashed into the Laurasian landmass, were severe enough to cause the extinction of many endemic Indian forms. The formation of the Himalayas resulted in blockage of frigid Central Asian air, preventing it from reaching India; this made its climate significantly warmer and more tropical in character than it would otherwise have been.

Semi-arid Wasteland

5. A semi-arid wasteland near Tirunelveli, Tamil Nadu. Monsoon clouds dump torrents of rain on lush forests that are only kilometres away in windward-facing Kerala, but are prevented from reaching Tirunelveli by the Agasthyamalai Range of the Western Ghats (background).

Fig 4.1

A scene in Uttarakhand's Valley of Flowers National Park. In contrast to Tirunelveli, the park receives ample orographic precipitation due to its location in a mountainous windward-facing region wedged between the Zanskars and the Greater Himalayas.

6. As in much of the tropics, monsoonal and other weather conditions in India are unstable: major droughts, floods, cyclones and other natural disasters are sporadic, but have killed or displaced millions. India's long-term climatic stability may be further threatened by global warming.

Fig 4.2

The formation of the Himalayas (pictured) during the Early Eocene some 53 million years ago was a key factor in determining India's modern-day climate; global climate and ocean chemistry may have been impacted.

Indian Climate Zone

Average annual temperatures across India:

	Below 20.0 °C	(< 68.0 °F)
	20.0–22.5 °C	(68.0–72.5 °F)
	22.5–25.0 °C	(72.5–77.0 °F)
	25.0–27.5 °C	(77.0–81.5 °F)
	Above 27.5 °C	(> 81.5 °F)

Climatic zones in India

	Alpine	**E**	*(ETh)*
	Humid subtropical	**C**	*(Cfa)*
	Tropical wet and dry	**A**	*(Aw)*
	Tropical wet	**A**	*(Am)*
	Semi-arid	**B**	*(BSh)*
	Arid	**B**	*(BWh)*

7. India is home to an extraordinary variety of climatic regions, ranging from tropical in the south to temperate and alpine in the Himalayan north, where elevated regions receive sustained winter snowfall. The nation's climate is strongly influenced by the Himalayas and the Thar Desert. The Himalayas, along with the Hindu Kush mountains in Pakistan, prevent cold Central Asian katabatic winds from blowing in, keeping the bulk of the Indian subcontinent warmer than most locations at similar latitudes. Simultaneously, the Thar Desert plays a role in attracting moisture-laden southwest summer monsoon winds that, between June and October, provide the majority of India's rainfall. Four major climatic groupings predominate, into which fall seven climatic zones that, as designated by experts, are defined on the basis of such traits as temperature and precipitation. Groupings are assigned codes (see chart) according to the Köppen climate classification system.

Tropical Wet

8. A tropical rainy climate covers regions experiencing persistent warm or high temperatures, which normally do not fall below 18 °C (64 °F). India hosts two climatic subtypes that fall under this group. The most humid is the tropical wet climate (also known as a tropical monsoon climate) that covers a strip of southwestern lowlands abutting the Malabar Coast, the Western Ghats, and southern Assam. India's two island territories, Lakshadweep and the Andaman and Nicobar Islands, are also subject to this climate. Characterised by moderate to high year-round temperatures, even in the foothills, its rainfall is seasonal but heavy—typically above 2,000 millimetres (79 in) per year. Most rainfall occurs between May and November; this is adequate for the maintenance of lush forests and other vegetation throughout the remainder of the year. December to March are the driest months, when days with precipitation are rare. The heavy monsoon rains are responsible for the extremely biodiverse tropical wet forests of these regions.

9. In India, a tropical wet and dry climate is more common. Noticeably drier than areas with a tropical monsoon climate, it prevails over most of inland peninsular India except for a semi arid rain shadow east of the Western Ghats. Winter and early summer are long, dry periods with temperatures averaging above 18 °C (64 °F). Summer is exceptionally hot; temperatures in low-lying areas may exceed 50 °C (122 °F) during May, leading to heat waves that can each kill hundreds of Indians. The rainy season lasts from June to September; annual rainfall averages between 750–1,500 millimetres (29.5–59.1 in) across the region. Once the dry northeast monsoon begins in September, most precipitation in India falls on Tamil Nadu, leaving other states comparatively dry.

Tropical Dry

10. A tropical arid and semi-arid climate dominates regions where the rate of moisture loss through evapotranspiration exceeds that from precipitation; it is subdivided into three climatic subtypes. The first, a tropical semi-arid steppe climate, predominates over a long stretch of land south of Tropic of Cancer and east of the Western Ghats and the Cardamom Hills. The region, which includes Karnataka, inland Tamil Nadu, western Andhra Pradesh, and central Maharashtra, gets between 400–750 millimetres (15.7–29.5 in) annually. It is drought-prone, as it tends to have less reliable rainfall due to sporadic lateness or failure of the southwest monsoon. North of the Krishna River, the summer monsoon is responsible for most rainfall; to the south, significant post-monsoon

rainfall also occurs in October and November. In December, the coldest month, temperatures still average around 20–24 °C (68–75 °F). The months between March to May are hot and dry; mean monthly temperatures hover around 32 °C, with 320 millimetres (13 in) precipitation. Hence, without artificial irrigation, this region is not suitable for permanent agriculture.

Thar Desert

Fig 4.3

11. Most of western Rajasthan experiences an arid climatic regime. Cloudbursts are responsible for virtually all of the region's annual precipitation, which totals less than 300 millimetres (11.8 in). Such bursts happen when monsoon winds sweep into the region during July, August, and September. Such rainfall is highly erratic; regions experiencing rainfall one year may not see precipitation for the next couple of years or so. Atmospheric moisture is largely prevented from precipitating due to continuous downdrafts and other factors.[17] The summer months of May and June are exceptionally hot; mean monthly temperatures in the region hover around 35 °C (95 °F), with daily maxima occasionally topping 50 °C (122 °F). During winters, temperatures in some areas can drop below freezing due to waves of cold air from Central Asia. There is a large diurnal range of about 14 °C (25.2 °F) during summer; this widens by several degrees during winter.

12. East of the Thar Desert, the region running from Punjab and Haryana to Kathiawar experiences a tropical and sub-tropical steppe climate. The zone, a transitional climatic region separating tropical desert from humid sub-tropical savanna and forests, experiences temperatures that are less extreme than those of the desert. Average annual rainfall is 300–650 millimetres (11.8–25.6 in), but is very unreliable; as in much of the rest of India, the southwest monsoon accounts for most precipitation. Daily summer temperature maxima rise to around 40 °C (104 °F). The resulting natural vegetation typically comprises short, coarse grasses.

Subtropical Humid

13. Most of Northeast India and much of North India are subject to a humid subtropical climate. Though they experience hot summers, temperatures during the coldest months may fall as low as 0 °C (32 °F). Due to ample monsoon rains, India has only one subtype of this climate, *Cfa* (under the Köppen system). In most of this region, there is very little precipitation during the winter, owing to powerful anticyclonic and katabatic (downward-flowing) winds from Central Asia.

14. Humid subtropical regions are subject to pronounced dry winters. Winter rainfall—and occasionally snowfall—is associated with large storm systems such as "Nor'westers" and "Western disturbances"; the latter are steered by westerlies towards the Himalayas. Most summer rainfall occurs during powerful thunderstorms associated with the southwest summer monsoon; occasional tropical cyclones also contribute. Annual rainfall ranges from less than 1,000 millimetres (39 in) in the west to over 2,500 millimetres (98 in) in parts of the northeast. As most of this region is far from the ocean, the wide temperature swings more characteristic of a continental climate predominate; the swings are wider than in those in tropical wet regions, ranging from 24 °C (75 °F) in north-central India to 27 °C (81 °F) in the east.

Montane

Fig 4.4 Pangong Lake in Ladakh, an arid montane region lying deep within the Himalayas.

15. India's northernmost areas are subject to a montane, or alpine, climate. In the Himalayas, the rate at which an air mass's temperature falls per kilometre (3,281 ft) of altitude gained (the dry adiabatic lapse rate) is 9.8 °C/km. In terms of environmental lapse rate, ambient temperatures fall by 6.5 °C (11.7 °F) for every 1,000 metres (3,281 ft) rise in altitude. Thus, climates ranging from nearly tropical in the foothills to tundra above the snow line can coexist within several dozen miles of each other. Sharp temperature contrasts between sunny and shady slopes, high diurnal temperature variability, temperature inversions, and altitude-dependent variability in rainfall are also common. The northern side of the western Himalayas, also known as the trans-Himalayan belt, is a region of barren, arid, frigid, and wind-blown wastelands. Most precipitation occurs as snowfall during the late winter and spring months.

16. Areas south of the Himalayas are largely protected from cold winter winds coming in from the Asian interior. The leeward side (northern face) of the mountains receives less rain while the southern slopes, well-exposed to the monsoon, get heavy rainfall. Areas situated at elevations of 1,070–2,290 metres (3,510–7,510 ft) receive the heaviest rainfall, which decreases rapidly at elevations above 2,290 metres (7,513 ft). The Himalayas experience their heaviest snowfall between December and February and at elevations above 1,500 metres (4,921 ft). Snowfall increases with elevation by up to several dozen millimetres per 100 metre (~2 in; 330 ft) increase. Elevations above 5,000 metres (16,404 ft) never experience rain; all precipitation falls as snow.

Seasons

17. The India Meteorological Department (IMD) designates four official seasons:

a. **Winter**. Winter, occurring from December to early March. The year's coldest months are December and January, when temperatures average around 10–15 °C (50–59 °F) in the northwest; temperatures rise as one proceeds towards the equator, peaking around 20–25 °C (68–77 °F) in mainland India's southeast.

b. **Summer.** **Summer** or **pre-monsoon** season, lasting from March to June (April to July in northwestern India). In western and southern regions, the hottest month is April; for northern regions, May is the hottest month. Temperatures average around 32–40 °C (90–104 °F) in most of the interior.

c. **Monsoon.** **Monsoon** or **rainy** season, lasting from June to September. The season is dominated by the humid southwest summer monsoon, which slowly sweeps across the country beginning in late May or early June. Monsoon rains begin to recede from North India at the beginning of October. South India typically receives more precipitation.

d. **Post-monsoon.** **Post-monsoon** season, lasting from October to December. In northwestern India, October and November are usually cloudless. Tamil Nadu receives most of its annual precipitation in the northeast monsoon season.

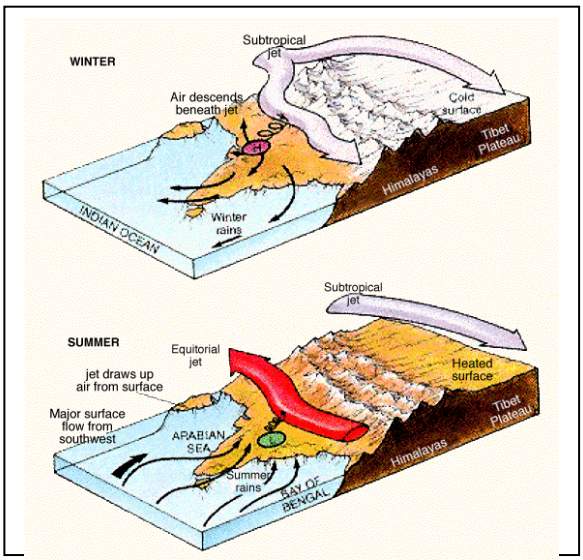


Fig 4.5

18. The Himalayan states, being more temperate, experience an additional two seasons: autumn and spring. Traditionally, Indians note six seasons, each about two months long. These are the spring (*Sanskrit: vasanta*), summer (*grīṣma*), monsoon season (*varṣā*), early autumn (*śarada*), late autumn (*hemanta*), and winter (*śiśira*). These are based on the astronomical division of the twelve months into six parts. The ancient Hindu calendar also reflects these seasons in its arrangement of months.

Winter

19. Once the monsoons subside, average temperatures gradually fall across India. As the Sun's vertical rays move south of the equator, most of the country experiences moderately cool weather; temperatures change by about 0.6 °C (1.08 °F) per degree of latitude. December and January are the coldest months, with mean temperatures of 10–15 °C (50–59 °F) in Indian Himalayas. Mean temperatures are higher in the east and south, where they reach 20–25 °C (68–77 °F).

Fig 4.6 Winter in Jammu and Kashmir

20. In northwestern India, virtually cloudless conditions prevail in October and November, resulting in wide diurnal temperature swings; as in much of the Deccan Plateau, they range between 16 °C (28.8 °F) – 20 °C (36.0 °F). However, from March to May, "western disturbances" bring heavy bursts of rain and snow. These extra-tropical low-pressure systems originate in the eastern Mediterranean Sea. They are carried towards India by the subtropical westerlies, which are the prevailing winds blowing at North India's range of latitude. Once their passage is hindered by the Himalayas, they are unable to proceed further, and they release significant precipitation over the southern Himalayas. The three Himalayan states (Jammu and Kashmir in the extreme north, Himachal Pradesh, and Uttarakhand) experience heavy snowfall; in Jammu and Kashmir, blizzards occur regularly, disrupting travel and other activities.

21. The rest of North India, including the Indo-Gangetic Plain, almost never receives snow. However, in the plains, temperatures occasionally fall below freezing, though never for more one or two days. Winter highs in Delhi range from 16 °C (61 °F) to 21 °C (70 °F). Nighttime temperatures average 2–8 °C (36–46 °F). In the Punjab plains, lows can fall below freezing, dropping to around −6 °C (21 °F) in Amritsar. Frost sometimes occurs, but the hallmark of the season is the notorious fog, which frequently disrupts daily life; fog grows thick enough to hinder visibility and disrupt air travel 15–20 days

annually. Eastern India's climate is much milder, experiencing moderately warm days and cool nights. Highs range from 23 °C (73 °F) in Patna to 26 °C (79 °F) in Kolkata (Calcutta); lows average from 8 °C (46 °F) in Patna to 14 °C (57 °F) in Kolkata. Frigid winds from the Himalayas can depress temperatures near the Brahmaputra River. The two Himalayan states in the east, Sikkim and Arunachal Pradesh, receive substantial snowfall. The extreme north of West Bengal, centred around Darjeeling, also experiences snowfall, but only rarely.

22. In South India, particularly the hinterland of Maharashtra, Madhya Pradesh, parts of Karnataka, and Andhra Pradesh, somewhat cooler weather prevails. Minimum temperatures in western Maharashtra, Madhya Pradesh and Chhattisgarh hover around 10 °C (50 °F); in the southern Deccan Plateau, they reach 16 °C (61 °F). Coastal areas, especially those near the Coromandel Coast, and low-elevation interior tracts are warm, with daily high temperatures of 30 °C (86 °F) and lows of around 21 °C (70 °F). The Western Ghats, including the Nilgiri Range, are exceptional; there, lows can fall below freezing. This compares with a range of 12 °C (21.6 °F) – 14 °C (25.2 °F) on the Malabar Coast; there, as is the case for other coastal areas, the Indian Ocean exerts a strong moderating influence on weather.

Summer

23. Summer in northwestern India lasts from April to July, and in the rest of the country from March to June. The temperatures in the north rise as the vertical rays of the Sun reach the Tropic of Cancer. The hottest month for the western and southern regions of the country is April; for most of North India, it is May. Temperatures of 50 °C (122 °F) and higher have been recorded in parts of India during this season. In cooler regions of North India, immense pre-monsoon squall-line thunderstorms, known locally as "Nor'westers", commonly drop large hailstones. Near the coast the temperature hovers around 36 °C (97 °F), and the proximity of the sea increases the level of humidity. In southern India, the temperatures are higher on the east coast by a few degrees compared to the west coast.

Fig 4.7 A summer view of Khajjiar, a Hill Station in Himachal Pradesh.

24. By May, most of the Indian interior experiences mean temperatures over 32 °C (90 °F), while maximum temperatures often exceed 40 °C (104 °F). In the hot months of April and May, western disturbances, with their cooling influence, may still arrive, but rapidly diminish in frequency as summer progresses. Notably, a higher frequency of such disturbances in April correlates with a delayed monsoon onset (thus extending summer) in northwest India. In eastern India, monsoon onset dates have been steadily advancing over the past several decades, resulting in shorter summers there.

25. Altitude affects the temperature to a large extent, with higher parts of the Deccan Plateau and other areas being relatively cooler. Hill stations, such as Ootacamund ("Ooty") in the Western Ghats and Kalimpong in the eastern Himalayas, with average maximum temperatures of around 25 °C (77 °F), offer some respite from the heat. At lower elevations, in parts of northern and western India, a strong, hot, and dry wind known as the Loo blows in from the west during the daytime; with very high temperatures, in some cases up to around 45 °C (113 °F); it can cause fatal cases of sunstroke. Tornadoes may also occur, concentrated in a corridor stretching from northeastern India towards Pakistan. They are rare, however; only several dozen have been reported since 1835.

Monsoon

26. The southwest summer monsoon, a four-month period when massive convective thunderstorms dominate India's weather, is Earth's most productive wet season. A product of southeast trade winds originating from a high-pressure mass centered over the southern Indian Ocean, the monsoonal torrents supply over 80% of India's annual rainfall. Attracted by a low-pressure region centered over South Asia, the mass spawns surface winds that ferry humid air into India from the southwest. These inflows ultimately result from a northward shift of the local jet stream, which itself results from rising summer temperatures over Tibet and the Indian subcontinent. The void left by the jet stream, which switches from a route just south of the Himalayas to one tracking north of Tibet, then attracts warm, humid air.

27. The main factor behind this shift is the high summer temperature difference between Central Asia and the Indian Ocean. This is accompanied by a seasonal excursion of the normally equatorial intertropical convergence zone (ITCZ), a low-pressure belt of highly unstable weather, northward towards India.[34] This system intensified to its present strength as a result of the Tibetan Plateau's uplift, which accompanied the Eocene–Oligocene transition event, a major episode of global cooling and aridification which occurred 34–49 Ma.

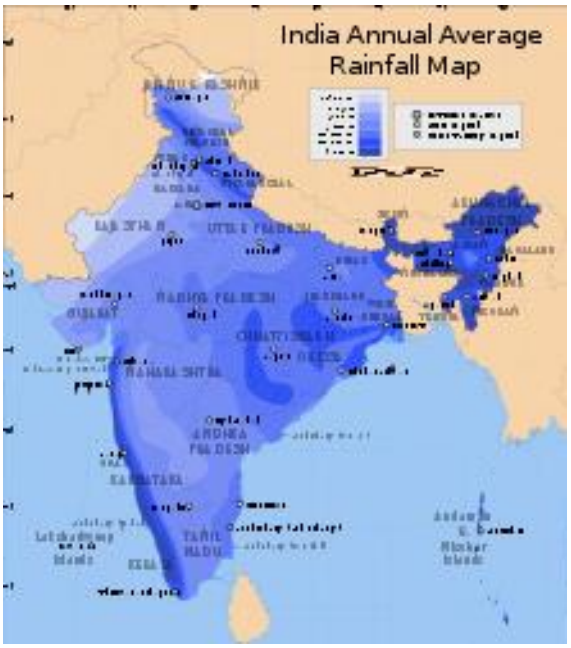
28. The southwest monsoon arrives in two branches: the Bay of Bengal branch and the Arabian Sea branch. The latter extends toward a low-pressure area over the Thar Desert and is roughly three times stronger than the Bay of Bengal branch. The monsoon typically breaks over Indian territory by around 25 May, when it lashes the Andaman and Nicobar Islands in the Bay of Bengal. It strikes the Indian mainland around 1 June near the Malabar Coast of Kerala. By 9 June, it reaches Mumbai; it appears over Delhi by 29 June. The Bay of Bengal branch, which initially tracks the Coromandal Coast northeast

from Cape Comorin to Orissa, swerves to the northwest towards the Indo-Gangetic Plain. The Arabian Sea branch moves northeast towards the Himalayas. By the first week of July, the entire country experiences monsoon rain; on average, South India receives more rainfall than North India. However, Northeast India receives the most precipitation. Monsoon clouds begin retreating from North India by the end of August; it withdraws from Mumbai by 5 October. As India further cools during September, the southwest monsoon weakens. By the end of November, it has left the country.

Fig 4.8

29. Monsoon rains impact the health of the Indian economy; as Indian agriculture employs 600 million people and composes 20% of the national GDP, good monsoons correlate with a booming economy. Weak or failed monsoons (droughts) result in widespread agricultural losses and substantially hinder overall economic growth. The rains reduce temperatures and replenish groundwater tables, rivers, and lakes.

Onset dates and prevailing wind currents of the southwest summer and northeast winter monsoons.



Regional variation in rainfall across India. The monsoon season delivers four-fifths of the country's precipitation.

Fig 4.8

Post-monsoon

30. During the post-monsoon months of October to December, a different monsoon cycle, the northeast (or "retreating") monsoon, brings dry, cool, and dense Central Asian air masses to large parts of India. Winds spill across the Himalayas and flow to the southwest across the country, resulting in clear, sunny skies. Though the India Meteorological Department (IMD) and other sources refers to this period as a fourth ("post-monsoon") season, other sources designate only three seasons. Depending on location, this period lasts from October to November, after the southwest monsoon has peaked. Less and less precipitation falls, and vegetation begins to dry out. In most parts of India, this period marks the transition from wet to dry seasonal conditions. Average daily maximum temperatures range between 28 and 34 °C (82 and 93 °F).

31. The northeast monsoon, which begins in September, lasts through the post-monsoon seasons, and only ends in March, carries winds that have already lost their moisture while crossing central Asia and the vast rain shadow region lying north of the Himalayas. They cross India diagonally from northeast to southwest. However, the large indentation made by the Bay of Bengal into India's eastern coast means that the flows are humidified before reaching Cape Comorin and rest of Tamil Nadu, meaning that the state, and also some parts of Kerala, experience significant precipitation in the post-monsoon and winter periods. However, parts of West Bengal, Orissa, Andhra Pradesh, Karnataka and North-East India also receive minor precipitation from the northeast monsoon.

Temperature and Precipitation

32. Shown below are temperature and precipitation data for selected Indian cities; these represent the full variety of major Indian climate types. Figures have been grouped by the four-season classification scheme used by the IMD; year-round averages and totals are also displayed.

Average temperatures in various Indian cities (°C)													
—	Winter (Jan – Feb)			Summer (Mar – May)			Monsoon (Jun – Sep)			Post-monsoon (Oct – Dec)			Year-round
City	Min	Avg	Max	Min	Avg	Max	Min	Avg	Max	Min	Avg	Max	Avg
Port Blair	23	**26**	28	25	**27**	29	25	27	27	25	26	28	27
Thiruvananthapuram	23	**26**	29	24	**27**	30	28	26	24	29	26	23	26
Bangalore	7	**12**	18	13	18	23	15	**19**	23	8	13	18	17
Nagpur	14	**21**	28	24	**32**	40	24	27	30	16	22	28	26
Bhopal	13	**18**	24	23	**30**	36	23	26	28	16	22	26	25
Guwahati	11	**17**	24	19	25	31	25	**28**	32	17	22	27	24
Lucknow	10	**15**	21	23	**30**	35	24	29	33	15	20	25	25
Jaisalmer	7	**14**	23	24	**33**	40	23	29	35	12	19	27	22
Dehradun	4	**12**	20	14	23	32	22	**26**	30	7	15	23	18
Amritsar	4	**10**	18	13	25	34	25	**28**	32	10	16	24	21
Shimla	1	**5**	9	10	14	18	15	**18**	20	7	10	13	13
Srinagar	−2	**4**	6	7	14	19	16	**22**	30	1	8	16	13

Leh	−13	**−6**	0	−1	6	12	10	**16**	24	−7	0	7	6

Average precipitation in various Indian cities (mm)													
—	**Winter (Jan – Feb)**			**Summer (Mar – May)**			**Monsoon (Jun – Sep)**			**Post-monsoon (Oct – Dec)**			**Year-round**
City	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**	**Oct**	**Nov**	**Dec**	**Total**
Port Blair	40	20	10	60	360	**480**	400	400	460	290	220	150	2,890
Thiruvananthapuram	26	21	33	125	202	**306**	175	152	179	223	206	65	1,713
Bangalore	31	20	61	110	150	212	249	279	**315**	291	210	140	1,962
Nagpur	16	22	15	8	18	168	290	**291**	157	73	17	19	1,094
Bhopal	4	3	1	3	11	136	279	**360**	185	52	21	7	1,043
Guwahati	8	21	47	181	226	309	**377**	227	199	92	25	10	1,722
Lucknow	20	18	8	8	20	114	**305**	292	188	33	5	8	1,019
Jaisalmer	–	–	3	–	7	10	**90**	88	15	–	6	–	219
Dehradun	47	55	52	21	54	230	**631**	627	261	32	11	3	2,024
Amritsar	24	33	48	30	45	27	**231**	187	79	18	6	18	746
Shimla	60	60	60	50	60	170	420	**430**	160	30	10	20	1,530
Srinagar	74	71	91	**94**	61	36	58	61	38	31	10	33	658
Leh	12	9	12	6	7	4	16	**20**	12	7	3	8	116

Fig 4.9

Disasters

33. Climate-related natural disasters cause massive losses of Indian life and property. Droughts, flash floods, cyclones, avalanches, landslides brought on by torrential rains, and snowstorms pose the greatest threats. Other dangers include frequent summer dust storms, which usually track from north to south; they cause extensive property damage in North India and deposit large amounts of dust from arid regions. Hail is also common in parts of India, causing severe damage to standing crops such as rice and wheat.

Disaster-prone regions in India.

Map showing winds zones, shaded by distribution of average speeds of prevailing winds.

Fig 4.10

Floods and Landslides

34. In the Lower Himalaya, landslides are common. The young age of the region's hills result in labile rock formations, which are susceptible to slippages. Rising population and development pressures, particularly from logging and tourism, cause deforestation. The result, denuded hillsides, exacerbates the severity of landslides, since tree cover impedes the downhill flow of water. Parts of the Western Ghats also suffer from low-intensity landslides. Avalanches occur in Kashmir, Himachal Pradesh, and Sikkim.

35. Floods are the most common natural disaster in India. The heavy southwest monsoon rains cause the Brahmaputra and other rivers to distend their banks, often flooding surrounding areas. Though they provide rice paddy farmers with a largely dependable source of natural irrigation and fertilisation, the floods can kill thousands and displace millions. Excess, erratic, or untimely monsoon rainfall may also wash away or

otherwise ruin crops. Almost all of India is flood-prone, and extreme precipitation events, such as flash floods and torrential rains, have become increasingly common in central India over the past several decades, coinciding with rising temperatures. Mean annual precipitation totals have remained steady due to the declining frequency of weather systems that generate moderate amounts of rain.

Cyclones

36. Tropical cyclones, which are severe storms spun off from the Intertropical Convergence Zone, may affect thousands of Indians living in coastal regions. Tropical cyclogenesis is particularly common in the northern reaches of the Indian Ocean in and around the Bay of Bengal. Cyclones bring with them heavy rains, storm surges, and winds that often cut affected areas off from relief and supplies. In the North Indian Ocean Basin, the cyclone season runs from April to December, with peak activity between May and November. Each year, an average of eight storms with sustained wind speeds greater than 63 km/h (39 mph) form; of these, two strengthen into true tropical cyclones, which have sustained gusts greater than 117 km/h (73 mph). On average, a major (Category 3 or higher) cyclone develops every other year.

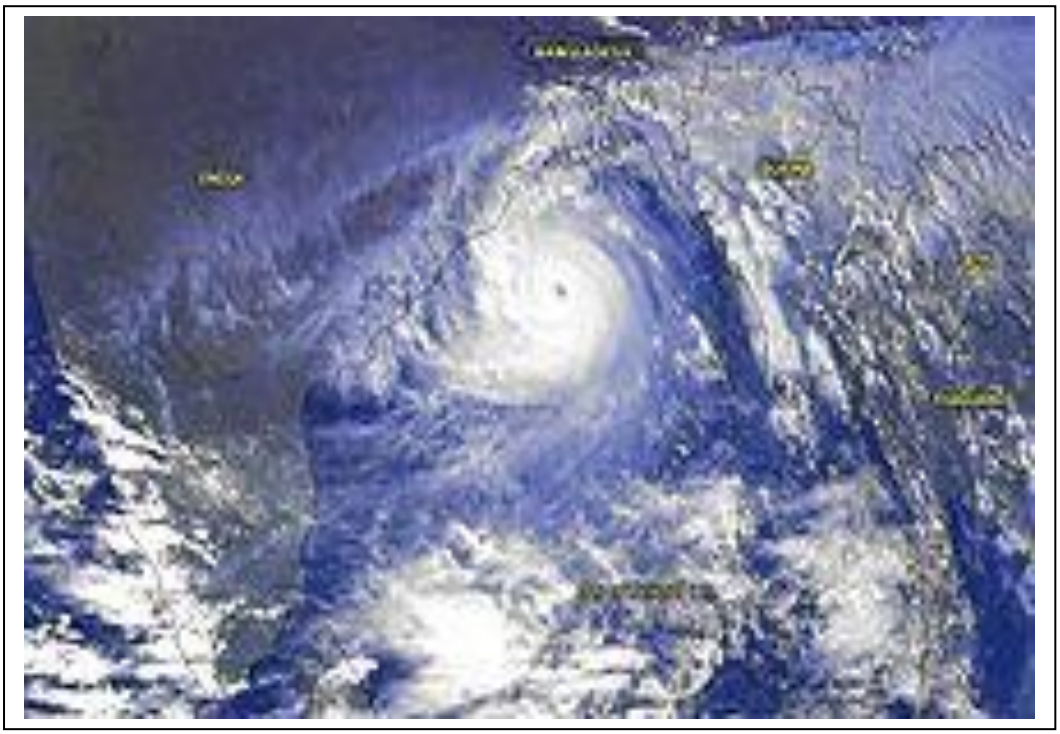


Fig 4.11 Cyclone in the Bay of Bengal

37. During summer, the Bay of Bengal is subject to intense heating, giving rise to humid and unstable air masses that morph into cyclones. The 1737 Calcutta cyclone, the 1970 Bhola cyclone, and the 1991 Bangladesh cyclone rank among the most powerful cyclones to strike India, devastating the coasts of eastern India and neighboring Bangladesh. Widespread death and property destruction are reported every year in the exposed coastal states of West Bengal, Orissa, Andhra Pradesh, and Tamil Nadu. India's western coast, bordering the more placid Arabian Sea, experiences cyclones only rarely; these mainly strike Gujarat and, less frequently, Kerala.

38. Cyclone 05B, a supercyclone that struck Orissa on 29 October 1999, was the deadliest in more than a quarter-century. With peak winds of 160 miles per hour (257 km/h), it was the equivalent of a Category 5 hurricane. Almost two million people were left homeless; another 20 million people lives were disrupted by the cyclone. Officially, 9,803 people died from the storm; unofficial estimates place the death toll at over 10,000.

Droughts

39. Indian agriculture is heavily dependent on the monsoon as a source of water. In some parts of India, the failure of the monsoons result in water shortages, resulting in below-average crop yields. This is particularly true of major drought-prone regions such as southern and eastern Maharashtra, northern Karnataka, Andhra Pradesh, Orissa, Gujarat, and Rajasthan. In the past, droughts have periodically led to major Indian famines. These include the Bengal famine of 1770, in which up to one third of the population in affected areas died; the 1876–1877 famine, in which over five million people died; the 1899 famine, in which over 4.5 million died; and the Bengal famine of 1943, in which over five million died from starvation and famine-related illnesses.

Fig 4.12 The dry bed of the Niranjana River, Bihar

40. All such episodes of severe drought correlate with El Niño-Southern Oscillation (ENSO) events. El Niño-related droughts have also been implicated in periodic declines in Indian agricultural output. Nevertheless, ENSO events that have coincided with abnormally high sea surfaces temperatures in the Indian Ocean—in one instance during 1997 and 1998 by up to 3 °C (5 °F)—have resulted in increased oceanic evaporation, resulting in unusually wet weather across India. Such anomalies have occurred during a sustained warm spell that began in the 1990s. A contrasting phenomenon is that, instead of the usual high pressure air mass over the southern Indian Ocean, an ENSO-related oceanic low pressure convergence center forms; it then continually pulls dry air from Central Asia, desiccating India during what should have been the humid summer monsoon season. This reversed air flow causes India's droughts. The extent that an ENSO event raises sea surface temperatures in the central Pacific Ocean influences the degree of drought.

Extreme Weather

41. India's lowest recorded temperature was −45 °C (−49 °F) in Dras, Ladakh, in eastern Jammu and Kashmir; however, the reading was taken with non-standard equipment. Readings as low as −30.6 °C (−23 °F) have been taken in Leh, further south in Ladakh. However, temperatures on the Indian-controlled Siachen Glacier near Bilafond La (5,450 metres / 17,881 feet) and Sia La (5,589 metres / 18,337 feet) have fallen below −55 °C (−67 °F), while blizzards bring wind speeds in excess of 250 km/h (155 mph), or hurricane-force winds ranking at 12 (the maximum) on the Beaufort scale. These conditions, not hostile actions, caused more than 97% of the roughly 15,000 casualties suffered by India and Pakistan during conflict in the region. The highest reliable temperature reading was 50.6 °C (123.1 °F) in Alwar, Rajasthan in 1955. The India Meteorological Department (IMD) doubts the validity of 55 °C (131 °F) readings in Orissa from 2005.

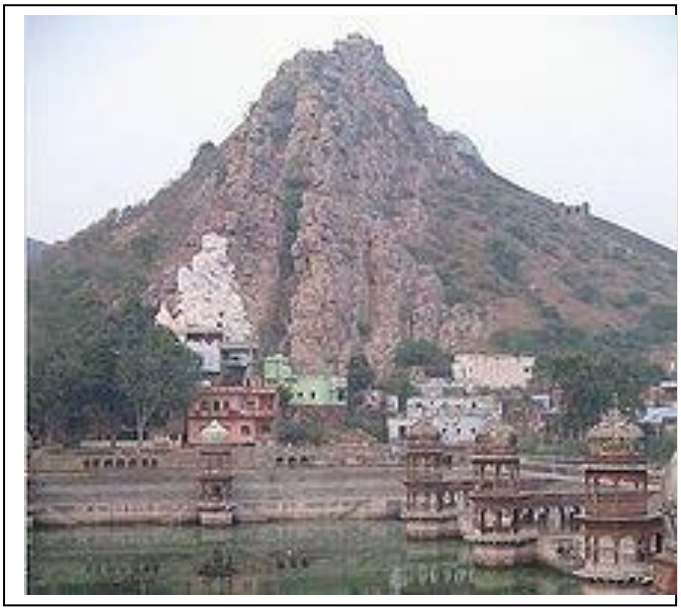


Fig 4.13 Alwar, on the fringes of the Thar Desert, repoted India's max temperature 50.6 °C

42. The average annual precipitation of 11,871 millimetres (467 in) in the village of Mawsynram, in the hilly northeastern state of Meghalaya, is the highest recorded in Asia, and possibly on Earth. The village, which sits at an elevation of 1,401 metres (4,596 ft), benefits from its proximity to both the Himalayas and the Bay of Bengal. However, since the town of Cherrapunji, 5 kilometres (3.1 mi) to the east, is the nearest town to host a meteorological office (none has ever existed in Mawsynram), it is officially credited as being the world's wettest place. In recent years, the Cherrapunji-Mawsynram region has averaged between 9,296 and 10,820 millimetres (366 and 426 in) of rain annually, though Cherrapunji has had at least one period of daily rainfall that lasted almost two years. India's highest recorded one-day rainfall total occurred on 26 July 2005, when Mumbai received more than 650 millimetres (25.6 in); the massive flooding that resulted killed over 900 people.

43. Remote regions of Jammu and Kashmir, such as Baramulla district in the east and the Pir Panjal Range in the southeast, experience exceptionally heavy snowfall. Kashmir's highest recorded monthly snowfall occurred in February 1967, when 8.4 metres (28 ft) fell in Gulmarg, though the IMD has recorded snowdrifts up to 12 metres (39.4 ft) in several Kashmiri districts. In February 2005, more than 200 people died when, in four days, a western disturbance brought up to 2 metres (7 ft) of snowfall to parts of the state.

Global Warming

44. Lakshadweep, comprising tiny low-lying islands, are at risk of being inundated by sea level rises associated with global warming. Current sea level rise, increased cyclonic activity, increased ambient temperatures, and increasingly fickle precipitation patterns are effects of global warming that have impacted or are projected to impact India. Thousands of people have been deplaced by ongoing sea level rises that have submerged low-lying islands in the Sundarbans. Temperature rises on the Tibetan Plateau are causing Himalayan glaciers to retreat, threatening the flow rate of the Ganges, Brahmaputra, Yamuna, and other major rivers; the livelihoods of hundreds of thousands of farmers depend on these rivers. A 2007 World Wide Fund for Nature (WWF) report states that the Indus River may run dry for the same reason.

Fig 4.14 Effects of Global Warming on India

45. Severe landslides and floods are projected to become increasingly common in such states as Assam. Ecological disasters, such as a 1998 coral bleaching event that killed off more than 70% of corals in the reef ecosystems off Lakshadweep and the Andamans, and was brought on by elevated ocean temperatures tied to global warming, are also projected to become increasingly common.

46. The Indira Gandhi Institute of Development Research has reported that, if the predictions relating to global warming made by the Intergovernmental Panel on Climate Change come to fruition, climate-related factors could cause India's GDP to decline by up to 9%. Contributing to this would be shifting growing seasons for major crops such as rice, production of which could fall by 40%. Around seven million people are projected to be displaced due to, among other factors, submersion of parts of Mumbai and Chennai, if global temperatures were to rise by a mere 2 °C (3.6 °F). Such shifts are not new. Earlier in the Holocene epoch (4,800–6,300 years ago), parts of what is now the Thar Desert were wet enough to support perennial lakes; researchers have proposed that this was due to much higher winter precipitation, which coincided with stronger monsoons. Kashmir's erstwhile subtropical climate dramatically cooled 2.6–3.7 Ma and experienced prolonged cold spells starting 600,000 years ago.

Atmospheric Pollution

47. Thick haze and smoke, originating from burning biomass in northwestern India and air pollution from large industrial cities in northern India, often concentrate inside the Ganges Basin. Prevailing westerlies carry aerosols along the southern margins of the steep-faced Tibetan Plateau to eastern India and the Bay of Bengal. Dust and black carbon, which are blown towards higher altitudes by winds at the southern margins of the Himalayas, can absorb shortwave radiation and heat the air over the Tibetan Plateau. The net atmospheric heating due to aerosol absorption causes the air to warm and convect upwards, increasing the concentration of moisture in the mid-troposphere and providing positive feedback that stimulates further heating of aerosols.

Fig 4.15 Clouds of Thick Haze and Smoke Form along the Ganges River Basin.