# CLIMATOLOGY NOTES

**DONE BY CHINEMBIRI.O for A and O Level lessons around HARARE**

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## Structure and Composition of the Atmosphere

### *Vertical Stratification*

**1 Troposphere**

Bottom most layer, 80% of atmosphere, most of water vapour and dust

Vertical mixing of air, all weather occurs here

Temperature decreases with height, 15C near ground, -57C at tropopause

Due to lesser terrestrial radiation and less dense air

Reaches 16 km in tropics and 9km in polar regions

**2. Stratosphere**

19.9% of atmosphere

Temperature constant until 20km, starts to increase until stratopause at 50km

Due to ozone layer, which absorbs UV rays. Lesser UV with decreasing height.

No precipitation – particulates and aerosols stay for long time

Strong winds (jet streams) distribute aerosols all over the world

**3. Mesosphere**

Temperature decreases with height because of decreasing density.

At 3000km (strong winds), about -90C

**4. Thermosphere**

Minimal mass of atmosphere, pressure about one-millionth of at sea level

Temperature increases with height due to insolation. Temprature reach 1500 c.

Atomic oxygen is highly concentrated in this layer.

## ATMOSPHERIC GASEOUS COMPONENTS

Atmosphere composed of mixture of gases and microscopic solid particles and water droplets

Permanent Gases

Nitrogen and oxygen make up 99% of clean, dry air

Variable Gases

Water Vapour

Source is evaporation from Earth's surface, decreases rapidly with height

Most found in lowest 5km of atmosphere

Source of moisture to form clouds, also a greenhouse gas

Carbon Dioxide

0.037% of the atmosphere

Supplied via the carbon cycle – decomposition, respiration, photosynthesis

Absorbs radiation, greenhouse gas

# Earth's Energy Budget OR the heat budget

Variation in solar energy received from places results in movement of air from one place to another, resulting in atmospheric circulation and oceanic circulation

## Global Energy Balance

Balance exists between insolation and outgoing terrestrial radiation so that the net radiation is zero

Insolation is scattered by gas and dust particles in the atmosphere, reflected by clouds and Earth's surface, and absorbed by atmosphere and Earth's surface

Outgoing radiation is released from the atmosphere and Earth's surface, and removed by wind and condensation

### *Scattering*

Some insolation will be scattered in all directions until it reaches Earth's surface or returns to space

This occurs because radiation travels in a straight line

## ***Reflection***

- Most important loss of insolation
- Light coloured and smooth, shiny surfaces will reflect more radiation
- Albedo – snow is about 90%, at low angles water is nearly 80%
- In general albedo of sea lower than land

## ***Absorption***

- Some gases in atmosphere absorb certain wavelengths of radiation, which gain energy and heat up
- Insolation is converted to long-wave radiation to warm atmosphere
- Greenhouse gases, oxygen, ozone, carbon dioxide, water vapour

## ***The Greenhouse Effect***

- Terrestrial radiation has longer wavelengths due to being cooler that sun
- Atmosphere absorbs long wave easier than short wave – greenhouse effect, gases absorb and re-emit radiation to warm the Earth

DIY…. effects of human activities

Mitigations

## ***Poleward Heat Transfer***

- Net radiation for Earth is zero, but not in all latitudes. Most energy received at tropics.
- In poles, low angle of sun, greater albedo of snow and ice, thicker atmosphere for sun to pass through, insolation is much smaller.
- Major equalising factor is transfer of heat by air movement, creating air currents, winds and ocean currents and weather

# Factors affecting Temperature

***The geographical and seasonal pattern of air temperature is affected by several factors. These include the following:***

- Insolation
- Latitude
- Altitude
- Cloud cover
- The nature of the surface
- Land and water contrasts
- Continentality
- Ocean currents
- Topography
- Aspect

## Insolation

The rate at which solar radiation is received at any point determines the temperature of that particular point. These vary considerably over time and space. Some of them the solar constant, distance from the sun, altitude of the sun, land and sea, the seasonal changes and length of day and night.

## Latitude

Areas that are close to the equator receive more heat than places that are close to the poles. This is because:

- Incoming solar radiation (insolation) is concentrated near the equator, but dispersed near the poles
- Insolation near the poles has to pass through a greater amount of atmosphere where there are greater chances of it being reflected back to space (Nagle 2000).

***Fig below shows the influence of latitude on temperature***

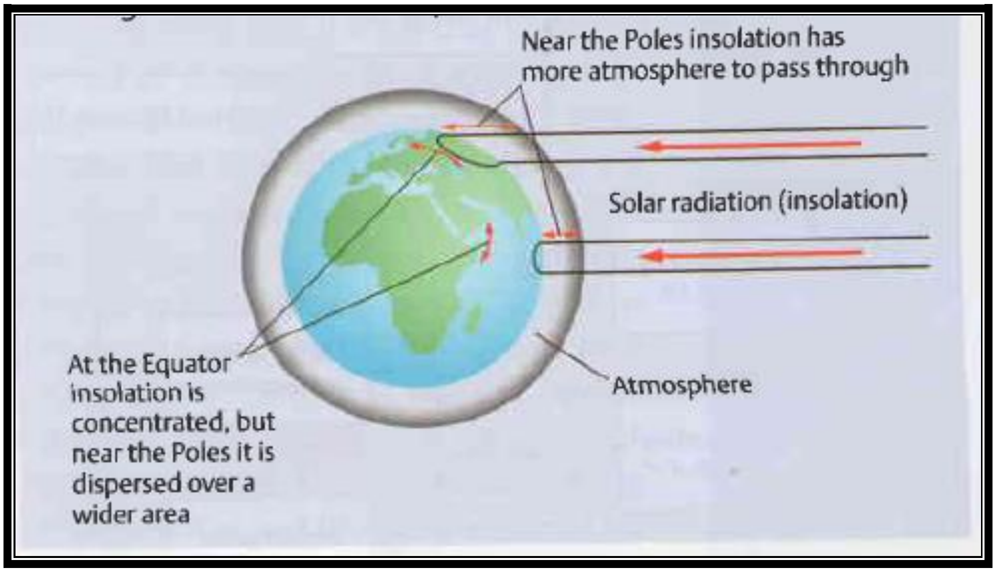



Fig 5.3 The influence of latitude on temperature (After Nagle, 2000)

## Altitude

Temperature decreases with altitude. On average it drops about 0.6C for every 1000m because at higher altitude air is thinner and less dense. Fig below shows temperature decreasing at an average of 100C for every 1000m

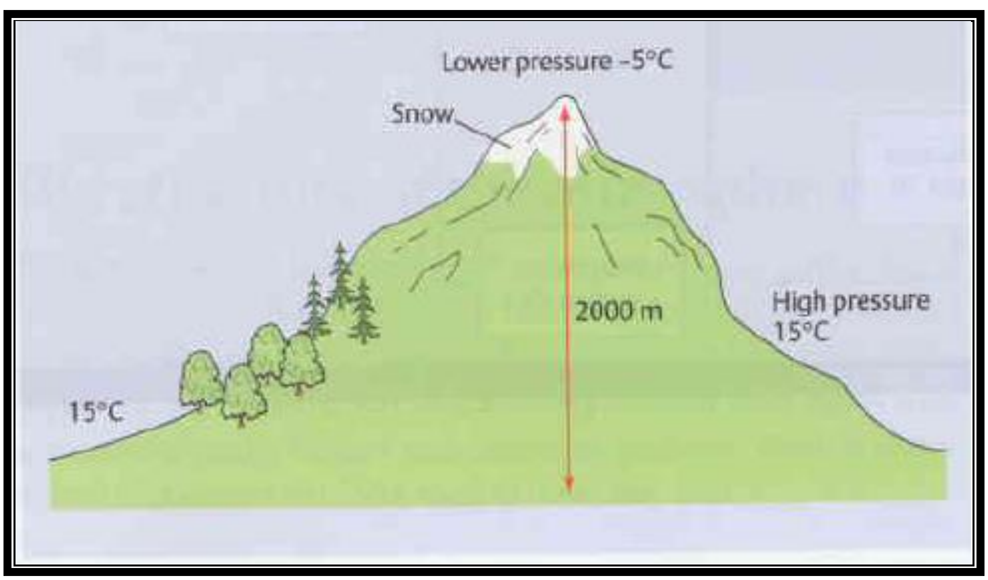


Fig 5.4 Influence of altitude on temperature (After Nagle, 2000)

## Cloud Cover

Clouds limit the inflow of direct heat from the sun and outflow of indirect heat through terrestrial radiation. The impact of cloud cover varies with the cloud type, cloud depth and area of cover. For example, deep cumulonimbus clouds in the Equatorial zone restrict incoming radiation and the outgoing one. As a result temperatures are high (30 0C) during the day and at night (29 0C) (Buckle 1996). Whereas in deserts such as the Sahara, the cloud free skies permit high day-time global radiation totals and rapid night-time heat loss. This causes the desert to be very hot during the day and cold at night.

## Nature of the Surface

Air is heated from the ground, and as such the nature of the surface greatly influences the temperature of the area. Different surfaces have varied conductivity and specific heat capacities. For example, dry sandy soils heat up and cool down rapidly since air is a poor conductor of heat. Low conductivity and low specific heat capacity account for the likelihood of night frost during cold spells in Western Zimbabwe (Buckle, 1996). However, moist clay soils are associated with higher night and day-time temperatures.

Influence of distance from the sea on temperature (after Nagle, 2000).

In much of Africa the mean daily temperature range exceeds the annual range. Latitudinally, the lowest values are in the equatorial zones where cloudiness and high humidity help reduce the daily range by limiting day time insolation and increasing night time counter radiation (Buckle, 1996). The cloud cover acts like a blanket holding day time temperatures down, while keeping night time temperatures up. Continentality is the chief cause of extreme daily ranges

## Aspect

Aspect is the direction that a place faces. On a local scale aspect is very important.

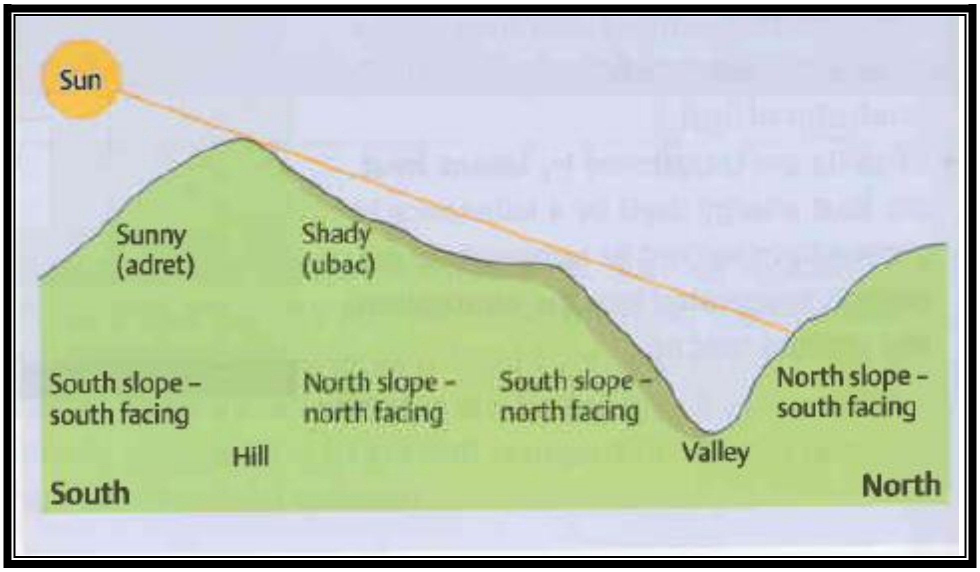
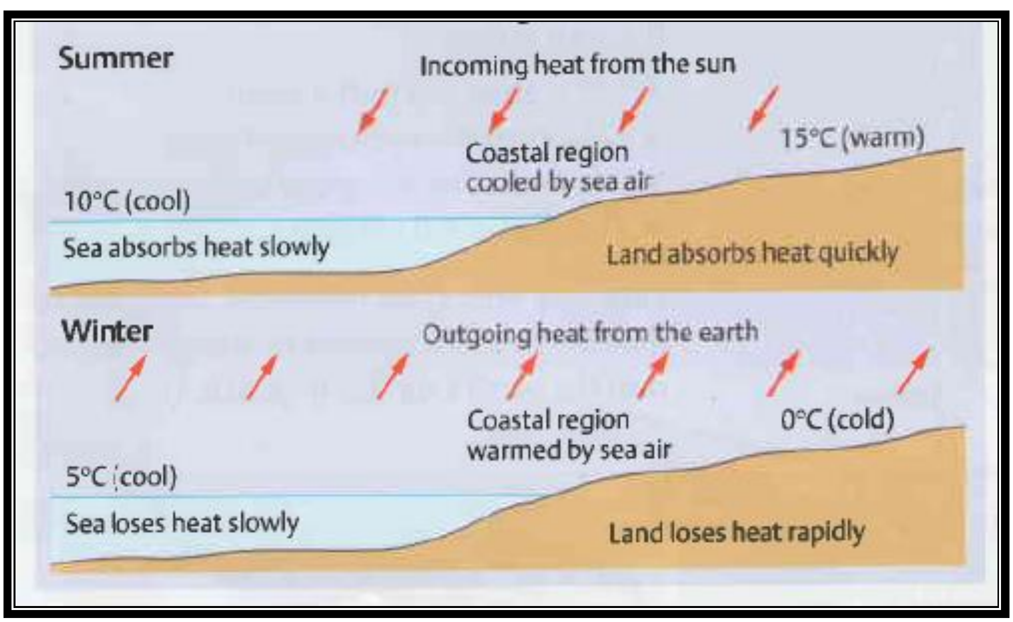



Fig 5.5 Influence of aspect on temperature (After Nagle, 2000)

**Distance from the sea**

It takes more energy to heat up water than it does to heat land. But it takes longer for water to loose heat. Therefore land is warmer than the sea during the day, but colder than the sea during the night. Areas that are close to the sea are cool by day, but mild at night. This effect is reduced with increasing distance from the sea. (See fig below).



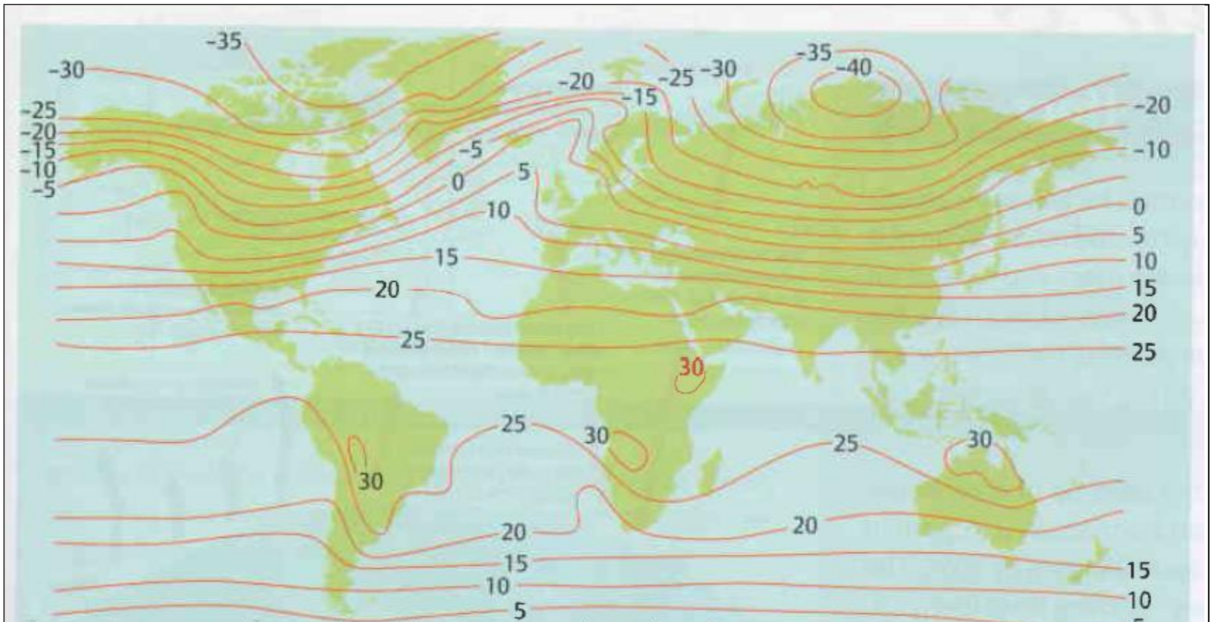
# AIR TEMPERATURE

## The Mean Distribution of Air Temperature at Sea level

Mean monthly temperatures decrease away from the equator. The east-west isotherms are best developed in the southern hemisphere where there are fewer large land masses to disrupt the pattern (Nagle, 2000). There are strong seasonal contrasts in January. In the northern hemisphere ocean currents raise temperatures pole-wards. Temperatures over northern continental interiors are very low with marked contrasts between land and ocean. The isotherms area almost parallel in the southern hemisphere. By contrast in the summer, continental interiors heat up, while coastal areas remain mild.

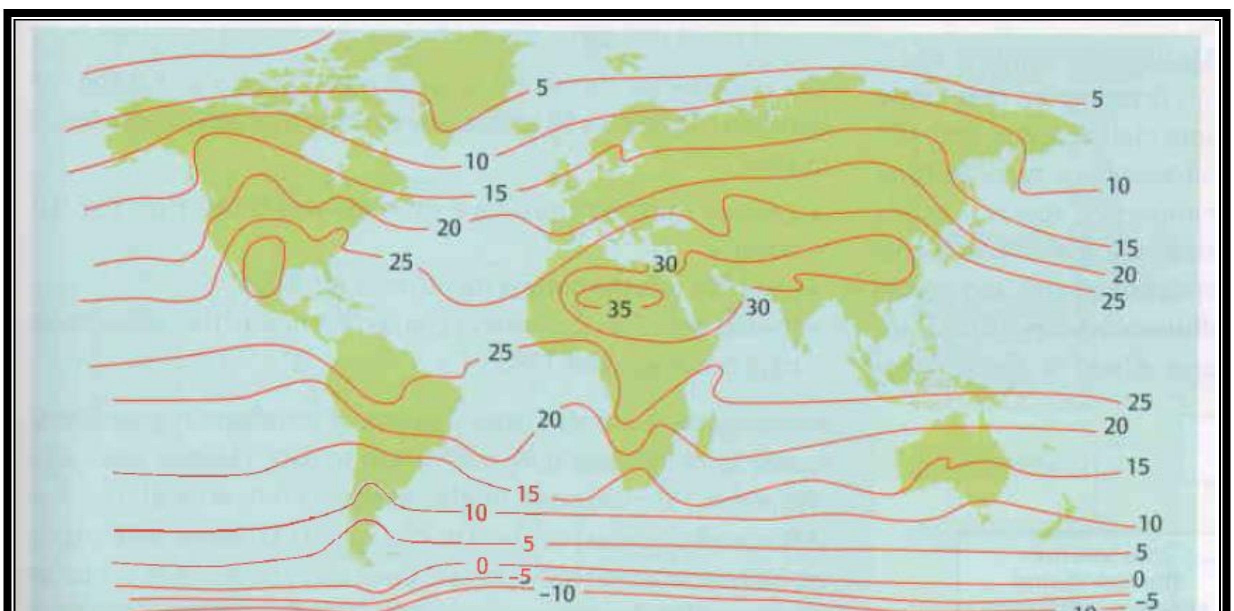
**Two patterns are clear**:

- The difference between the northern hemisphere (with large land masses) and the southern hemisphere (lacking large land masses)
- The contrasts between maritime and continental areas




Temperatures over northern continental interiors are very low with marked contrasts between land and ocean. Isotherms are roughly parallel in the southern hemisphere.

Mean January temperatures (After Nagle, 2000**)**




Temperatures in northern continental interiors contrast with the low January temperatures. The land/ocean contrast in the northern hemisphere is reversed.

Mean July Temperatures (After Nagle, 2000)

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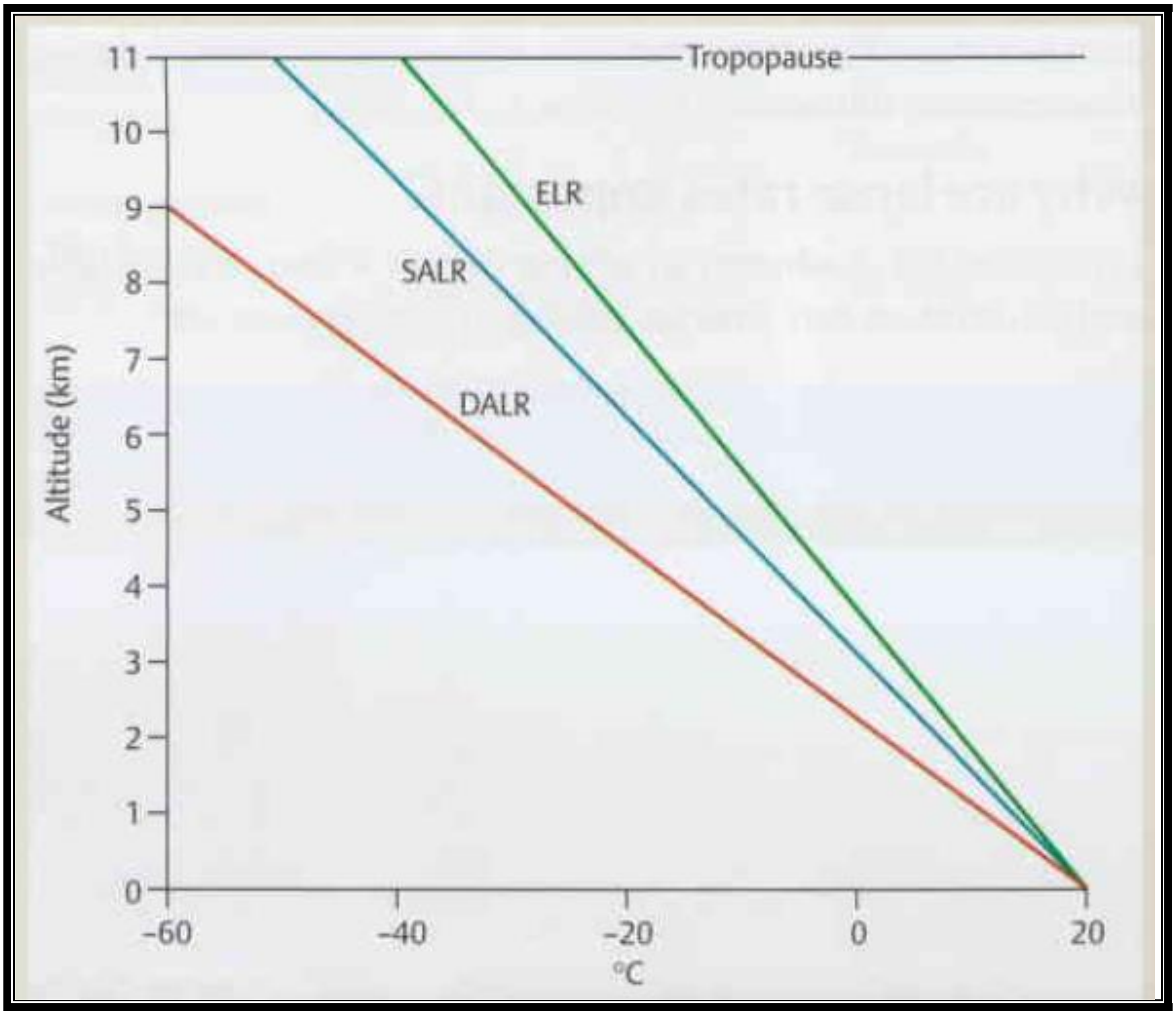
# Air Stability and Instability

Stability and instability are terms used to describe the physical conditions in the atmosphere according to whether the air has any tendency to rise or resist uplift. Parcels of warm air that rise through the lower cool adiabatically. The rate and maintenance of any vertical uplift depend on the temperature-density balance between the rising parcel and the surrounding air (Waugh, 2009).

## Stability

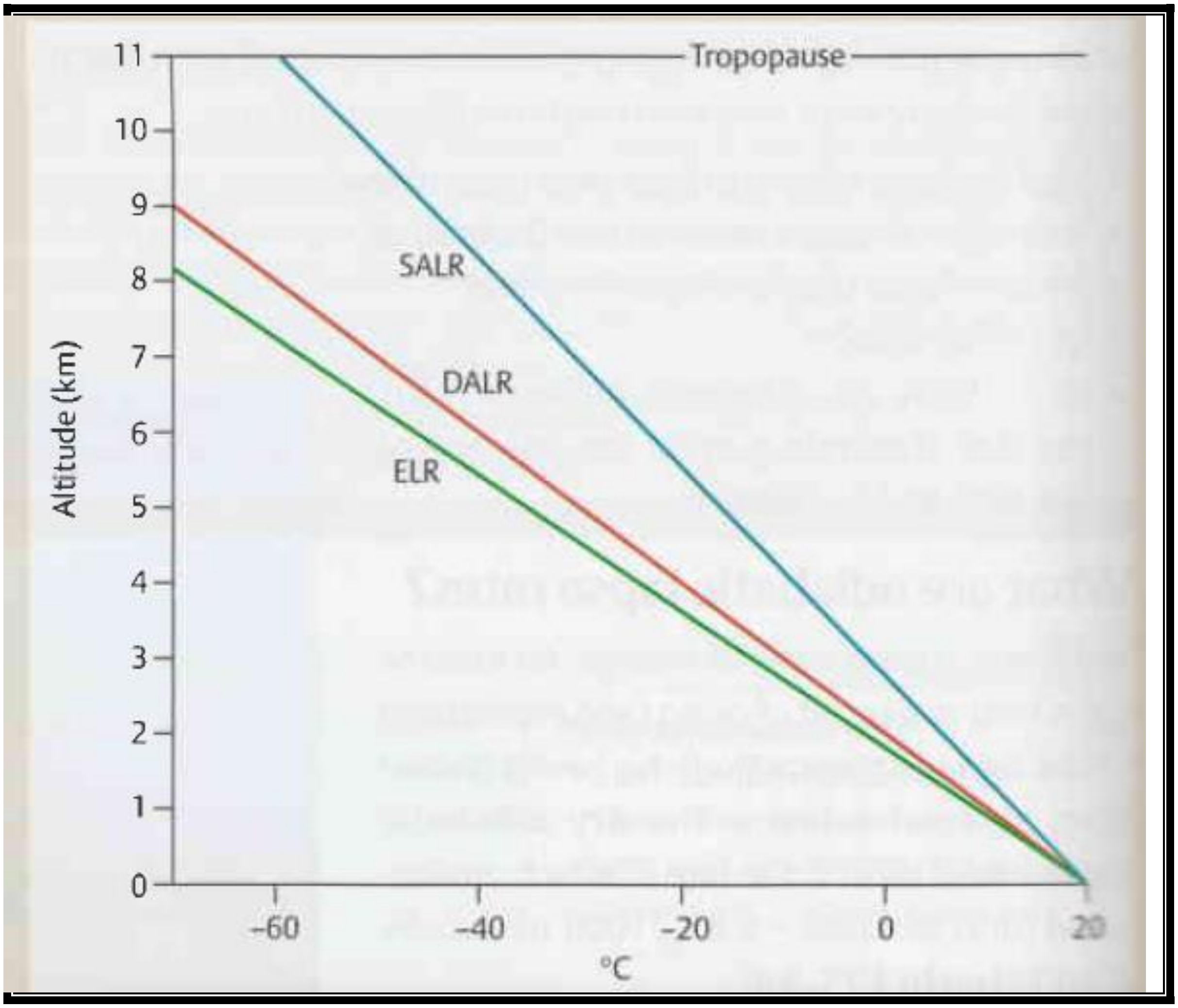
Air that resists vertical displacement is said to be stable. A stable air parcel, forced to rise or sink, will eventually return to its original position. Stability is often linked to anticyclones when convection currents are suppressed by sinking air to give dry, sunny conditions.




**Stability (After Nagle, 2000)**

## *Instability*

However, air that is displaced up or down and continues to rise or sink is said to be unstable. Instability is an important concept in weather studies because uplift and the consequent adiabatic cooling of air leads to deep cloud development and precipitation. It depends on the steepness of the environmental lapse rate (ELR) and the moisture content of the air.  Instability can be absolute, conditional and convective.



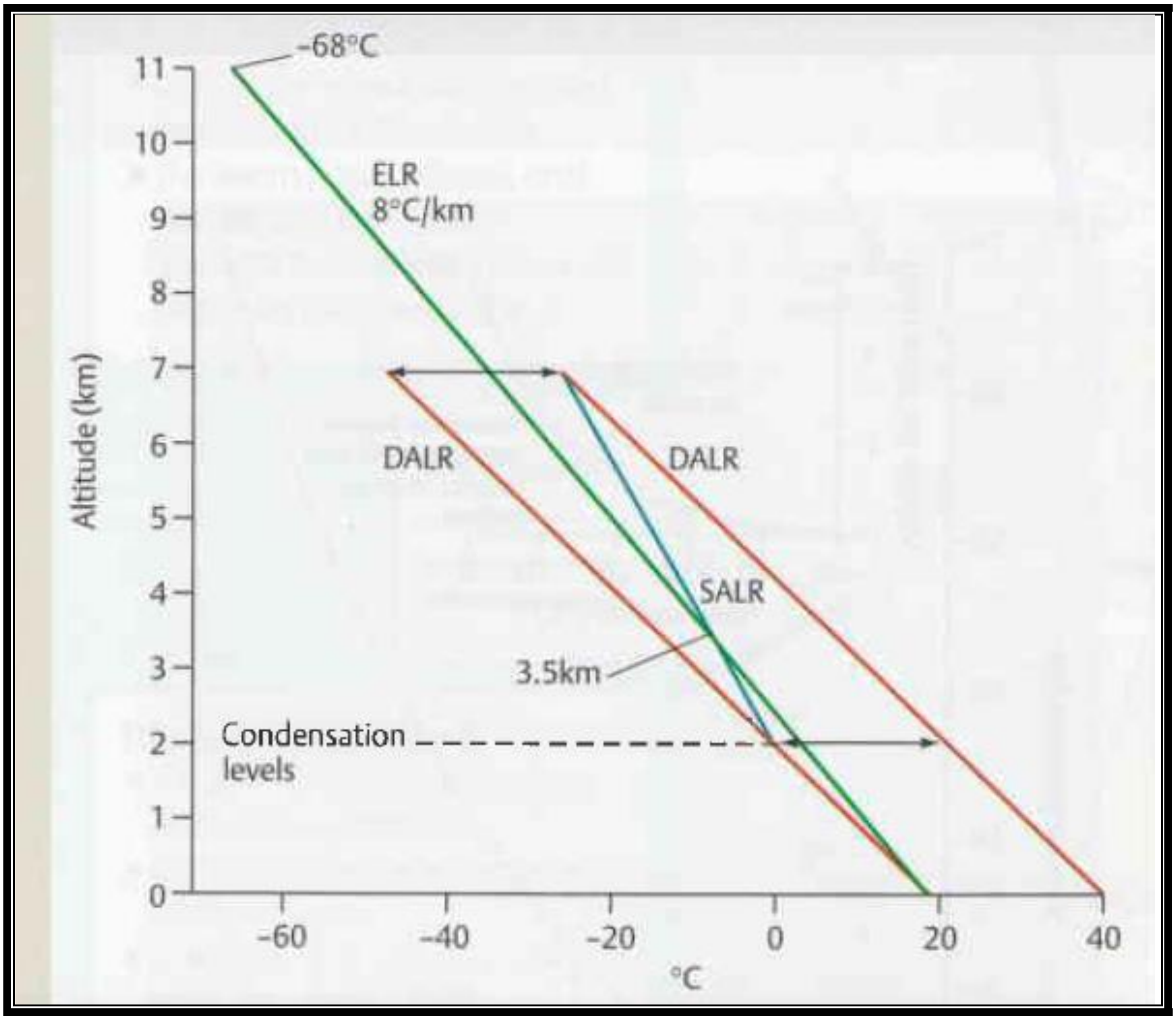

Instability (After Nagle, 2000)

# *Absolute instability*

Absolute instability exists whenever the ELR is steeper than the DALR. A thermal – parcel of warm air that rises from the surface of its own accord- is an example of absolute instability (super-adiabatic).

# ***Conditional Instability***

Conditional Instability happens when cool, dry, stable air is forced up for example by orographic or frontal uplift, far enough to reach saturation and commence cooling at the slower SALR. The air is then unstable and will continue to rise of its own accord until it reaches the temperature of the surrounding environment. The air is described as conditionally unstable because the air is unstable conditional upon it being forced up, becoming saturated, and releasing latent heat- all of which depends upon the amount of moisture in the air (Buckle, 1996). It develops when the ELR lies between the DALR and the SALR.



Conditional Instability (After Nagle, 2000)

## *Convective Instability*

A very stable layer can be converted into an absolutely unstable layer if the lower part is moist and the upper part quite dry. A stable airstream, in which the upper part cools at a faster rate than the lower part, because the upper part is dry and the lower part is saturated, is potentially unstable. The potential instability, realised by the lifting of the whole layer is called convective instability. It is associated with thunderstorms.

# *MOISTURE IN THE ATMOSPHERE*

## Humidity

Humidity is a measure of water vapour content in the atmosphere (Waugh, 2009).It can be represented by three different measures: absolute humidity, relative humidity and vapour pressure. Humidity depends on the temperature of the air. At any given temperature, there is a limit to the amount of moisture that the air can hold.

**Absolute Humidity**

This is the total amount of water vapour that a given volume of air contains. It is measured in grams per cubic metre (g/m3). Specific humidity is similar to absolute humidity but is expressed in grams of water per kilogram of air (g/kg).

**Relative Humidity**

is the amount of water vapour held in a volume of air at a given temperature compared with the maximum amount that could be held at that temperature (Buckle, 1996). Therefore it is the ratio of the air water vapour content to its water vapour capacity. Relative humidity measures how near the air is to saturation at a given temperature. If the relative humidity is 100%, the air is saturated. If it is between 80 and 99%, the air is said to be moist and the weather is humid or clammy (Waugh, 2009). When the relative humidity drops to 50%, the air is dry.

**Vapour Pressure**

is the pressure exerted by water vapour molecules in a given volume of air. It varies from place to place according to the vapour concentration.

**Saturation Vapour Pressure**

The maximum value the vapour pressure can attain at a given temperature is called the saturation vapour pressure. It depends on temperature and not on total pressure.

**Saturation Deficit**

Saturation deficit is the amount of water vapour needed to bring a parcel of non-saturated air to saturation at a given temperature and pressure. It is the amount by which the vapour pressure falls short of the saturation vapour pressure at a given temperature. Saturation is thus brought about by increasing vapour pressure, decreasing temperature or a combination of the two.

**Dew Point**

This is the temperature at which the air becomes saturated by the water vapour it contains and starts to condense to form tiny dew droplets. Dew point varies from day to day and place to place. It is always less than air temperature unless saturation has been reached.

## Hydrological Cycle

The water cycle has inputs, outputs, flow regulators and storage elements just like the solar energy cascade you studied in previous modules. The main inputs to this cycle are from precipitation. The distribution of this input across the world shows a marked relationship to the distribution of factors influencing precipitation.

The main outputs are evaporation and transpiration. The rate of evaporation depends upon a number of factors. Most important is the supply of energy, for evaporation involves the conversion of water to water vapour and this requires considerable inputs of energy. Another important factor is the availability of moisture at the surface. As the surface dries out and moisture becomes less available, rates of evaporation tends to decline. In addition, evaporation is favoured by a moisture gradient between the surface and the air above, and thus rates decline when the atmosphere is moist. Finally wind plays an important part by removing the moist air and maintains a moisture gradient (Briggs and Smithson, 1995).Storage occurs in oceans, the cryosphere (ice-covered areas of the world) and the ground water.

Over vegetated land, much of the moisture is returned to the atmosphere not by evaporation from the ground surface but by transpiration from the plants. By transpiration we mean evaporation from leaves of plants. Other processes involved in the hydrological cycle are percolation, infiltration, overland flow and run-off.

## Condensation

Is the process by which water vapour in the atmosphere is changed into liquid or, if the temperature is below 0 $^{0}$C, a solid (Waugh, 2009). It usually results from air being cooled until it is saturated. Cooling may be achieved by:

## Radiation (contact) Cooling

This occurs in calm, clear nights. The ground loses heat rapidly through terrestrial radiation and the air in contact with it is then cooled by conduction. This may result in radiation fog or dew if the air is moist or hoar frost if the temperature is below freezing point.

**Advection Cooling**

Advection cooling commonly results from warm moist air moving over a cooler land or sea surface. For example, advection fogs in California and Namib Desert are formed when air from the land drifts over cold offshore water currents.

**Orographic Uplift**

Warm moist air is forced to rise as it crosses a mountain barrier

**Frontal Uplift**

Warm moist air is forced to rise as when it meets a colder, denser mass of air at a front.

**Convective or Adiabatic Cooling**

This occurs when air is warmed during the day and rises in pockets as thermals. As the air rises, it expands and uses its energy, and so loses heat and the temperature drops. The air is said to be cooled adiabatically because the cooling is due to the reduction of pressure with height rather than the loss of heat to the surrounding air.

**Evaporation**

Evaporation will take place if:

- There is abundant moisture available
- The vapour content of the air is such that the air is not saturated
- The temperature of the air is high and certainly above that level at which condensation occurs
- The air is in motion so saturated air is quickly removed from contact with the surface and additional non-saturated air can replace it
- There is an energy source to sustain the transformation. This may come in the form of direct sunlight or from advected air containing sufficient heat energy.

**Evapotranspiration**

**Two processes here are involved: evaporation and transpiration. The rate of evapotranspiration depends on:**

- **Temperature** At higher temperatures there will be a larger saturation deficit making evapotranspiration greater.
- **Humidity** Little evapotranspiration will occur if the air is nearer to saturation, but if there is a big saturation deficit, as in the warm dry air, then evaporation will be considerable.
- **Turbulence** Strong convection allows unsaturated air to be nearer a water surface by lifting moist air upwards.
- **Ground surface** Evapotranspiration draws on moisture from below ground, continuing even after the soil surface has dried out. On bare ground the rates are much higher than on vegetated surfaces.

Over the continents evapotranspiration is at a maximum in regions where:

- Temperatures are high
- Humilities low
- Winds fairly strong and
- Vegetation sparse

**Seasonal patterns of Atmospheric Humidity in Africa**

In Africa, atmospheric humidity changes with seasons. The highest annual mean humidity occurs in low-lying equatorial regions. Very low dew points are recorded in the dry tropics. In Western Africa seasonal humidity variations reflect the changing airstreams on the monsoon trough.

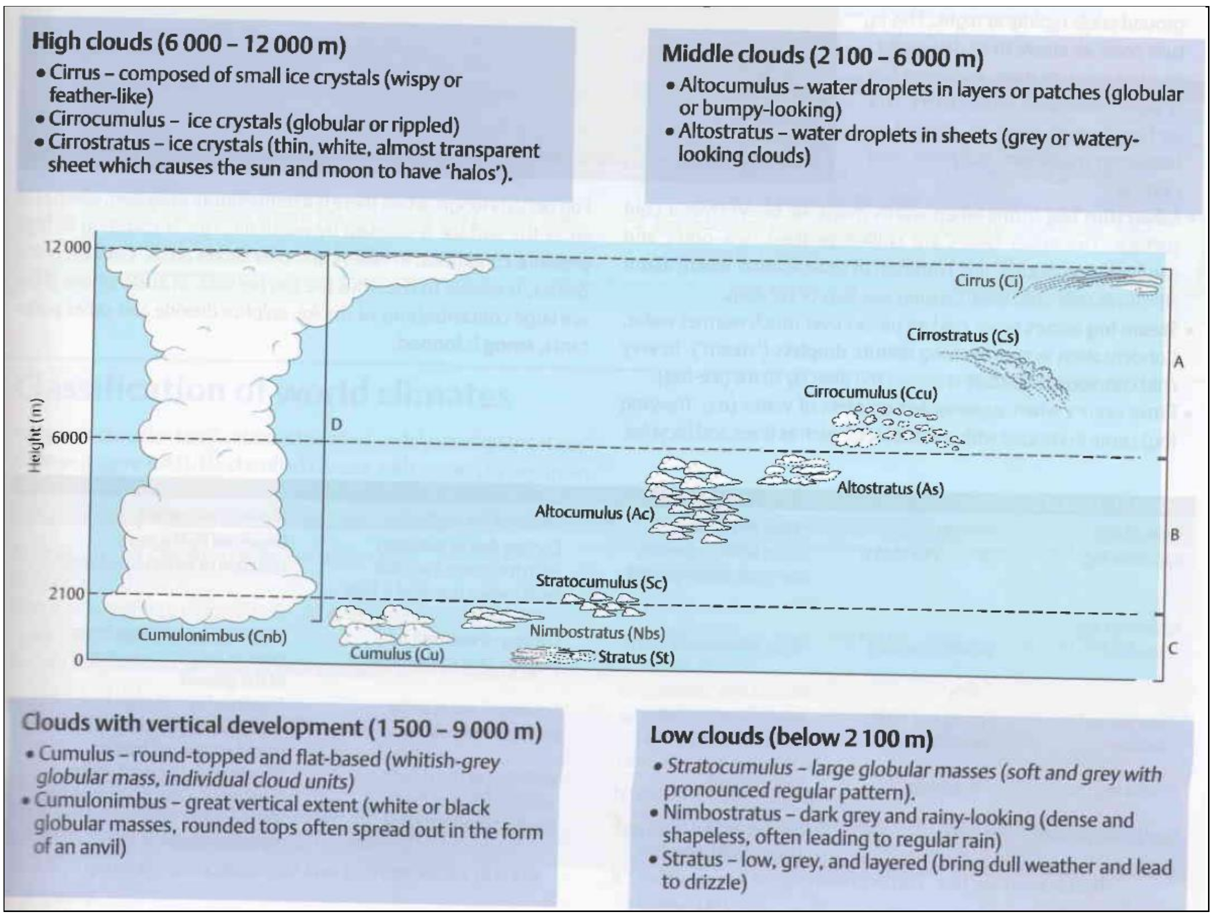
## *CLOUDS*

# CLOUD COVER AND FOG

This unit focuses on types of clouds and fog. The clouds are classified according to their height above sea level and their physical appearance. Types of fog to be considered are radiation, rime, steam, advection and smog.

## Classification of Clouds

Clouds are traditionally classified into ten main types. Some are simple to identify and others are of a composite nature and more complex in appearance.




## Classification of Clouds (After Waugh, 2009)

### High Level Clouds (6 000 - 1 800m)

- **Cirrus (Ci)**

These are fine, feathery, detached streamers, often with hooked ends in high, tangled mass that sometimes forms into long parallel bands converging on the horizon (Buckle, 1996). The cirrus clouds can be thick and dense, especially on a cumulonimbus anvil.

- **Cirrostratus (Cr)**

  The Cirrostratus clouds are thin, filmy veil either giving a complete sky cover or appearing in a series of separate filaments

- **Cirrocumulus (Cc)**

  This is a thin patch or sheet composed of tiny flakes in regular ripple-like pattern, giving a mackerel sky.

**Medium Level Clouds (2 000 - 8 000)**

- **Altocumulus (Ac)**: A layer of small, thin, woolly cloudlets in a pattern of cells or long waves, sometimes in patches and sometimes a complete sky cover.
- **Altostratus (As):** uniform or fibrous sheet with a smooth, level base, sometimes giving a complete sky cover.
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**Low Level Clouds (Up to 2 500m)**

- **Stratus (St)** These are dense, shapeless layers of clouds with a flat base. They either form an unbroken, uniform sheet or a series of ragged patches.
- **Stratocumulus (Sc)** dense, undulating layer, either in patches or a continuous sheet, or with a regular pattern of long wavy rolls or globular cells.
- **Nimbostratus (Ns)** thick, dense, generally shapeless mass with a base near ground level and beneath which hang ragged, wind-driven patches.

**Clouds of Great Vertical Extent**

- **Cumulus (Cu)** detached, fluffy heaps with sharp, clear cut outlines and horizontal bases, which vary in size.

- **Cumulonimbus (Cb)** huge, dense thunderstorm clouds of enormous vertical size, having a dark horizontal base and a turret-like top that is often splayed into a plume or anvil shape ahead of the main cloud mass.

# ***Fog***

Fog is cloud at ground level. It mostly occurs in high pressure (calm) conditions as winds tend to mix and disperse it (Nagle, 2000). In the morning fog usually disappears when the ground is heated causing rising air to lift it. A number of types can be identified:

**Radiation Fog**

Radiation Fog forms when the ground cools rapidly at night. In turn this will cool air above to its dew point, resulting in condensation (Nagle, 2000). Radiation fog is best developed when there is a surface layer of moist air, clear skies favouring maximum radiation, and calm air.

**Steam Fog**

Steam fog occurs when cold air passes over much warmer water. In such circumstances condensation results in minute droplets (steam). In very cold conditions moisture is converted directly to ice (ice-fog).

**Advection Fog**

This is formed when warm moist air blows over a cold surface. The lower layers are chilled to their dew point resulting in condensation. It is common in coastal areas such as in California and Namib Desert where warm air moves (from the desert) over cold water currents. This causes the formation of sea fog or hill fogs when it happens on cold land.

**Rime**

Rime occurs when super-cooled droplets of water come in contact with solid objects such as rocks and trees.

**Smoke Fog**

This forms in areas where there are large concentrations of smoke, sulphur dioxide and other pollutants (Nagle, 2000).

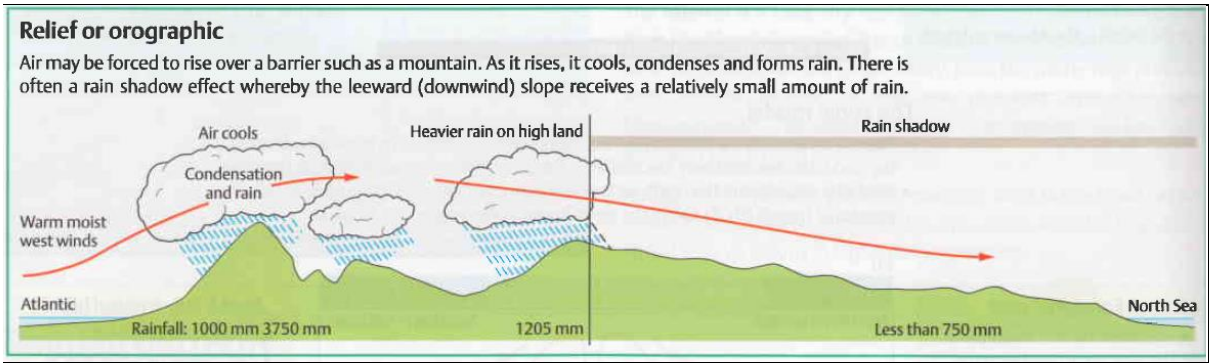
# Types of rainfall

Precipitation refers to the various forms of water droplets and ice particles that fall from the clouds. The forms of precipitation are rain, hail, snow, drizzle and sleet. The form and size of precipitation reaching the ground depend on the humidity and temperature conditions beneath the cloud base, and on the processes within the cloud. Three major types of rainfall are common in Zimbabwe: convectional, relief and frontal or cyclonic rainfall. Certain conditions are needed for the formation of precipitation:

- Cooling air
- Saturated air
- Condensation and cloud formation
- An accumulation of moisture
- Growth of cloud droplets

## Relief rain

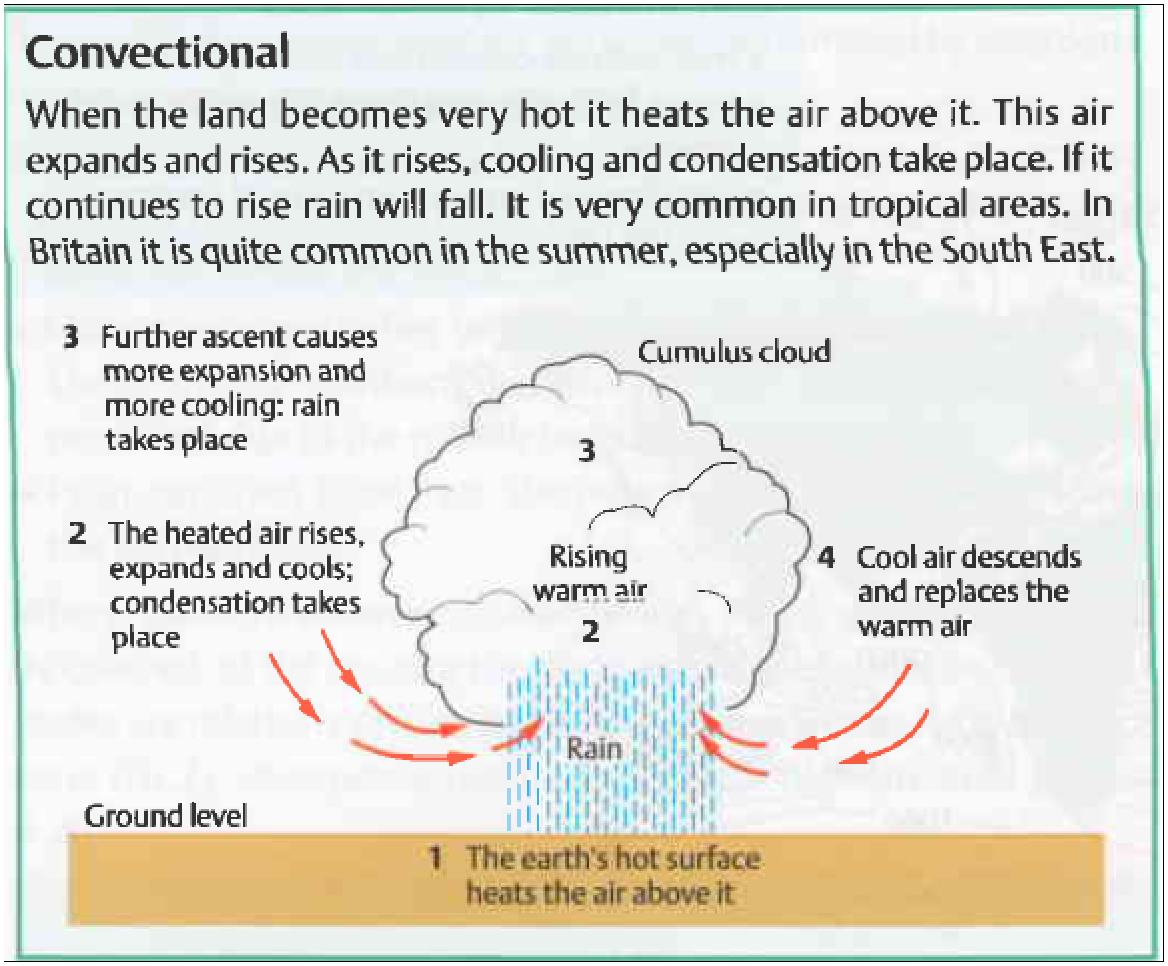
This type of rain occurs when air is forced to rise over a barrier such as a mountain. As it rises it cools, condenses and forms rain. There is often a rain shadow effect whereby the leeward slope receives a relatively small amount of rain. In Zimbabwe this rain occurs in the Manicaland province near Vumba Mountains. Fig below show the formation of relief rain.




**Relief rainfall formation (After Nagle, 2000)**

# Convectional Rainfall

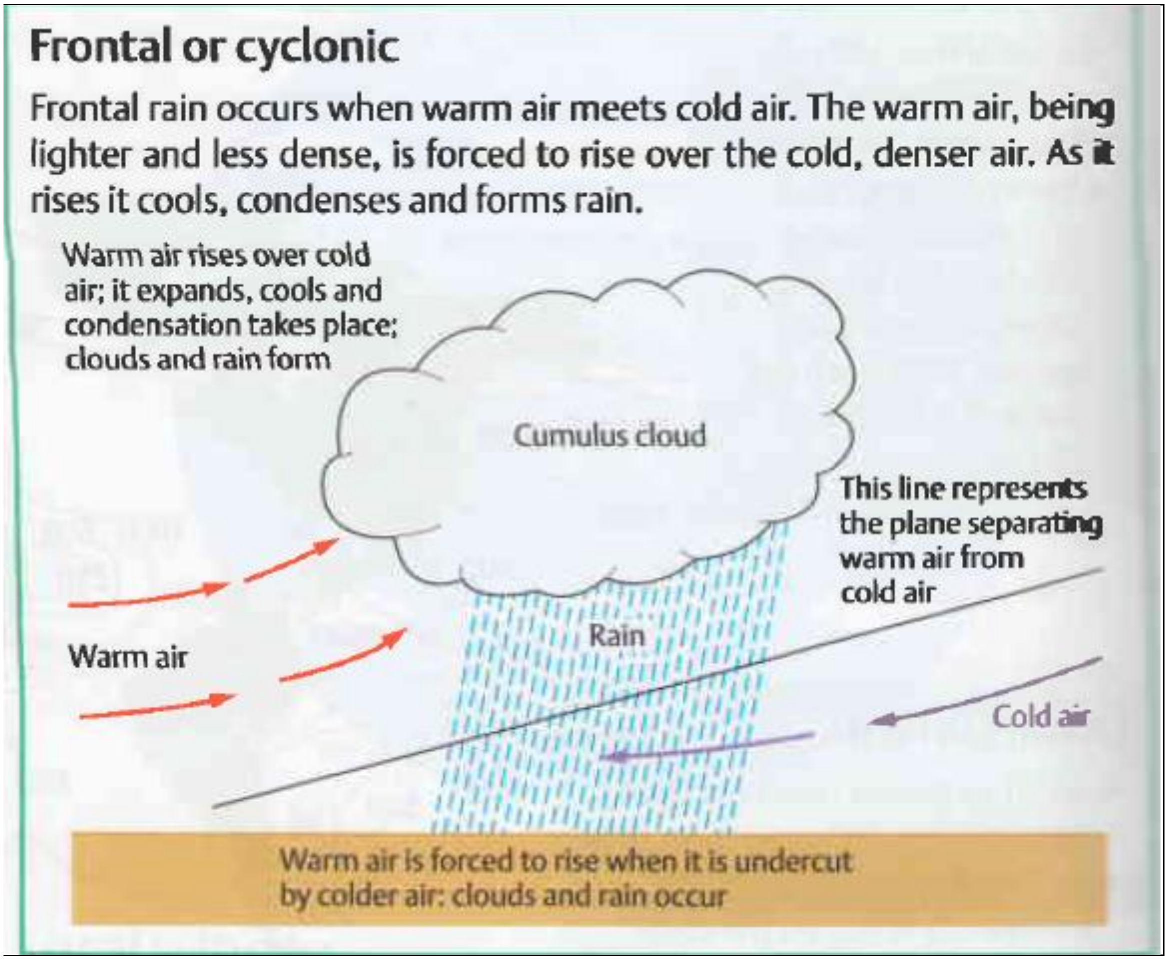
When the land becomes very hot it heats the air above it. The heated air expands and rises. As it rises, cooling and condensation starts to occur. If these processes continue, rain will fall. Convectional rain is common in tropical areas. Fig below shows the formation of convectional rain.




Formation of Convectional rainfall (After Nagle, 2000)

# Frontal or Cyclonic Rain

Frontal rain occurs when warm air meets cold air. The warm air is then forced to rise over the cold air because it is lighter and less dense. As it rises it cools, condenses and forms rain. The diagram below shows the formation of frontal or cyclonic rain.



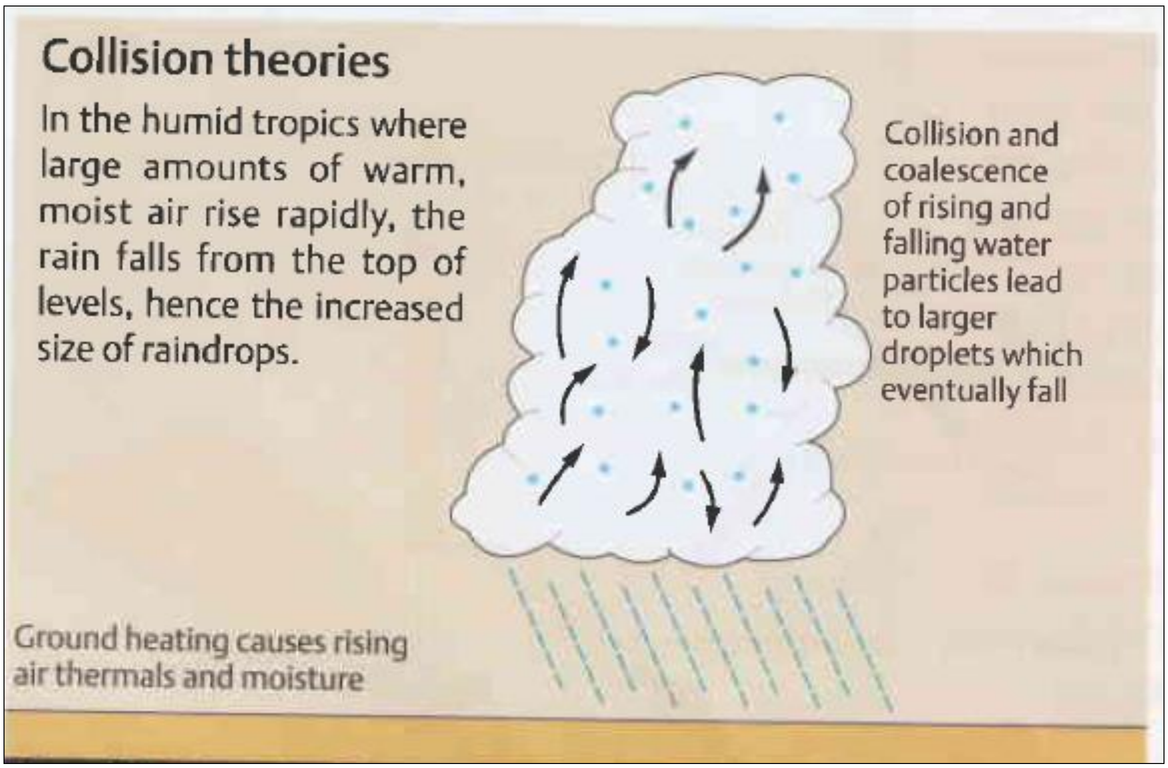

Formation of Frontal or Cyclonic rain (After Nagle, 2000)

# Theories of raindrop Formation

There are two groups of theories that attempt to explain the rapid growth of raindrops. These are the collision and coalescence of small water droplets by the sweeping action of falling drops, and the growth of ice crystals at the expense of water droplets.

**The Collision and coalescence Theory**

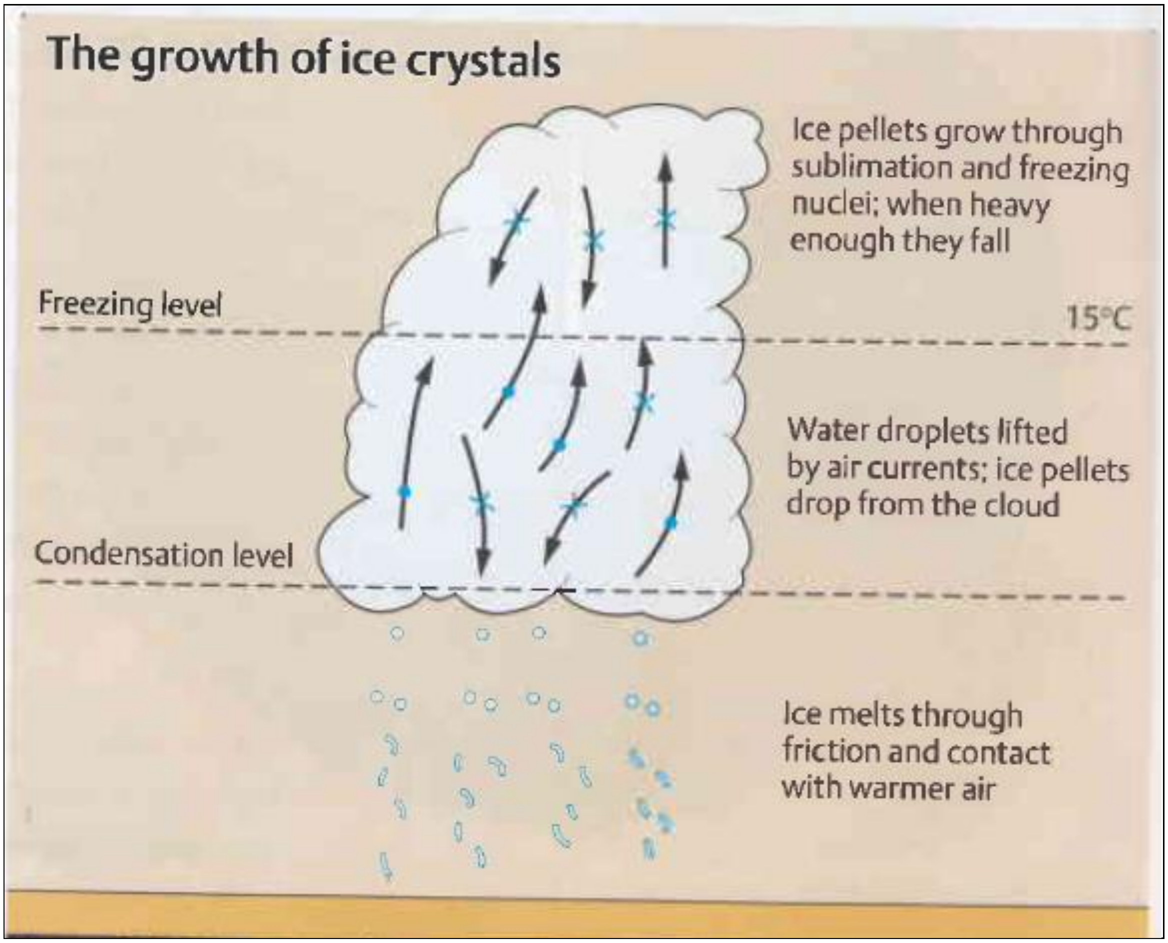
The warm moist air (without crystals) as found in the tropics, contain numerous water droplets of different sizes. These are swept upwards at different velocities, and in doing so, collide with other droplets. Larger water droplets have greater chances of collision and subsequent coalescence with smaller droplets. However, coalescence does not result from collision alone but is due to droplets having different electrical charges. Negatively charged droplets colliding with positively charged droplets have a high chance of merging, because opposite charges attract each other (Buckle, 1996). When coalescing droplets reach a radius of 3mm, their motion causes them to disintegrate and form a fresh supply of droplets. Coalescence is common in tropical regions where temperatures are high; clouds have high water content and therefore a high droplet density.




Collision Coalescence (After Nagle, 2000)

**The Ice Crystal Growth Theory**

The theory is associated with clouds where at least some of the water exists in solid state (ice crystals). This occurs when the temperature of the air is between -5 0C and -25 0C. According to this theory water droplets grow in clouds whose tops rise above the freezing level. Supper cooling takes place when water remains in the atmosphere after temperatures have fallen below 0 0C, usually due to lack of condensation nuclei. The relative humidity of air is ten times greater above an ice surface than on water. This means that the water droplets evaporate and the resultant vapour condenses (sublimates) back onto the ice crystals which then grow into hexagonal-shaped snowflakes. The ice crystal growth theory suggests the formation snow, sleet and hail.




Ice Crystal Growth Theory

## The Inter-Tropical Convergence Zone (ITCZ)

The Inter-tropical Convergence Zone (ITCZ) represents the region of convergence of the two moisture-laden trade wind systems. It is a region of relatively low pressure and great instability containing clusters of deep, convective which produce heavy precipitation (Buckle, 1996). It occurs in a location where low-level convergence is at a maximum and in the vicinity of the latitude of maximum surface temperature. Over the oceans the zone of maximum temperature does not normally exhibit much variability in location except where the ocean currents are influential. Over the land, the conditions are more complex, leading to more of a seasonal migration in its position in conjunction with the migration of the latitude of maximum insolation.

**Convergence**

The approach and meeting of airflows from different direction is known as convergence. This implies a net flow of air into a region, causing an accumulation of air and related increase in density (Buckle, 1996). Regions of convergence are associated chiefly with areas of low pressure where air moves towards a point or spirals inwards, but they take many forms and occur at several different levels at a time. There is also isotach convergence, whereby the airflow narrows or whereby the wind speed decreases in a downwind direction, causing faster flowing air behind to overtake and pile up. The convergence of two surface airstreams of different density is likely to initiate considerable uplift. This would result in instability and perhaps precipitation.

**Divergence**

This is the act opposite of convergence. Air moves away from a place, often in several different directions at the same time causing a net outflow (Buckle, 1996). The resulting depletion of air and consequent reduction in density is typically compensated by vertical motions as air moves to fill the divergence. Divergence may again show variations in form and scale and may occur at any level. It is associated with areas of high pressure and there is also isotach divergence whenever wind speed increases in a downwind direction, or the flow spreads apart. Divergence is linked to the warming of air through subsidence and compression, and as such is a key factor in the creation of a stable atmosphere.

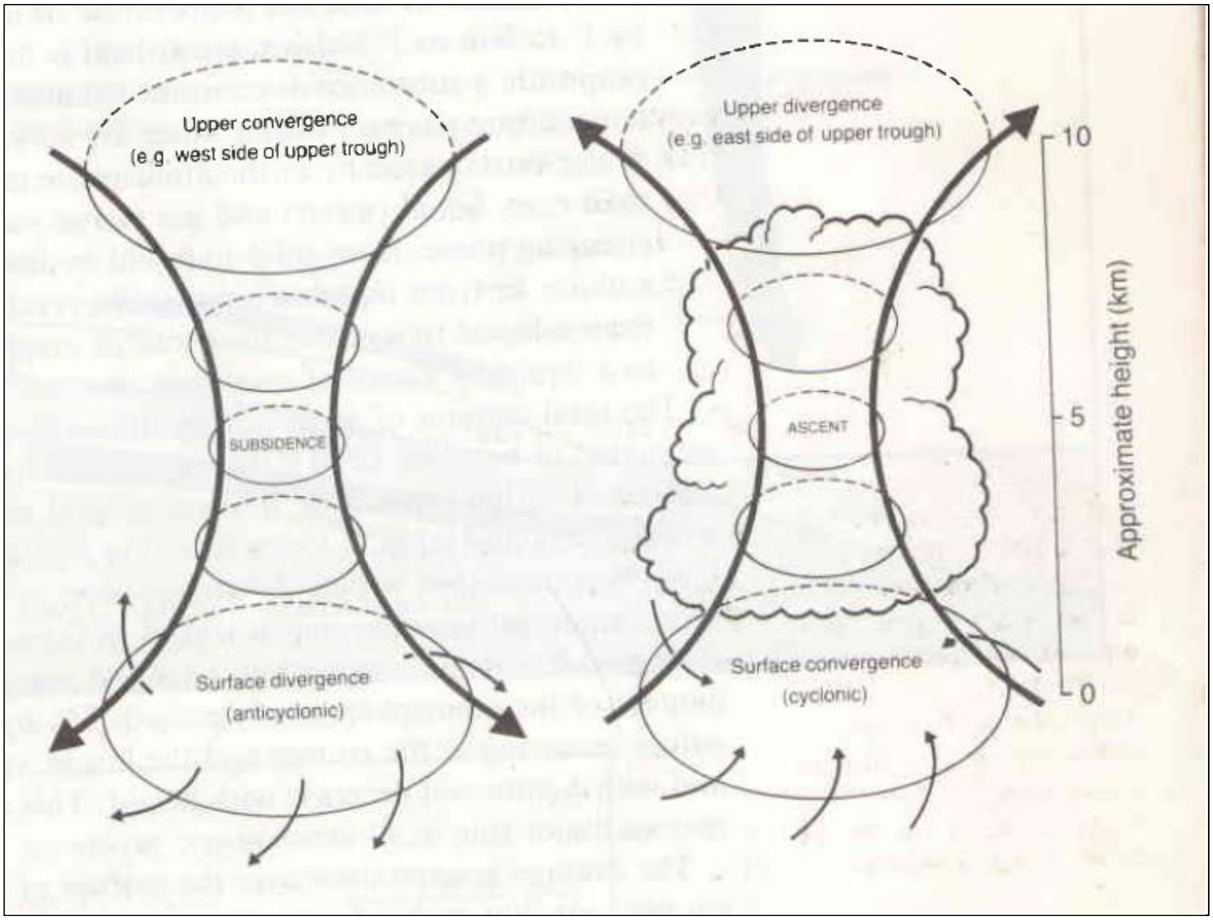



Fig 8.6 Schematic illustration indicating how patterns of divergence and convergence at the surface and in the upper troposphere are linked by vertical motions (After Musk, 1988).

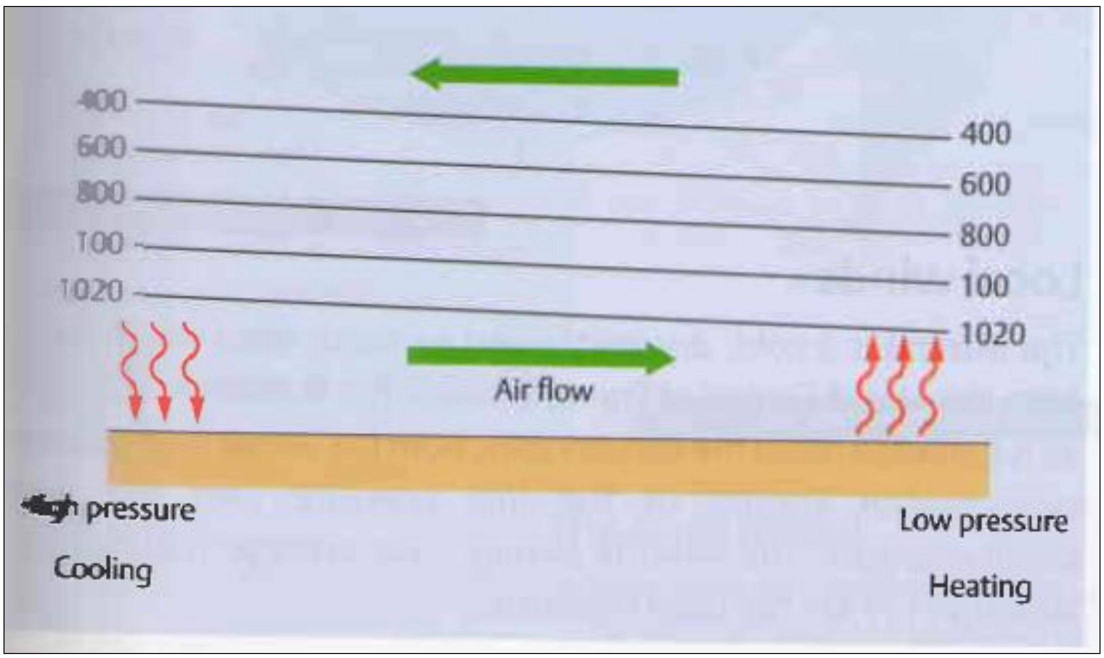
# ATMOSPHERIC MOTION

## Factors which Control Atmospheric Motion

Atmospheric motion is governed by the combination of the following forces: the pressure gradient force, the Coriolis force, the geostrophic wind and the surface friction.

## Pressure Gradient Force

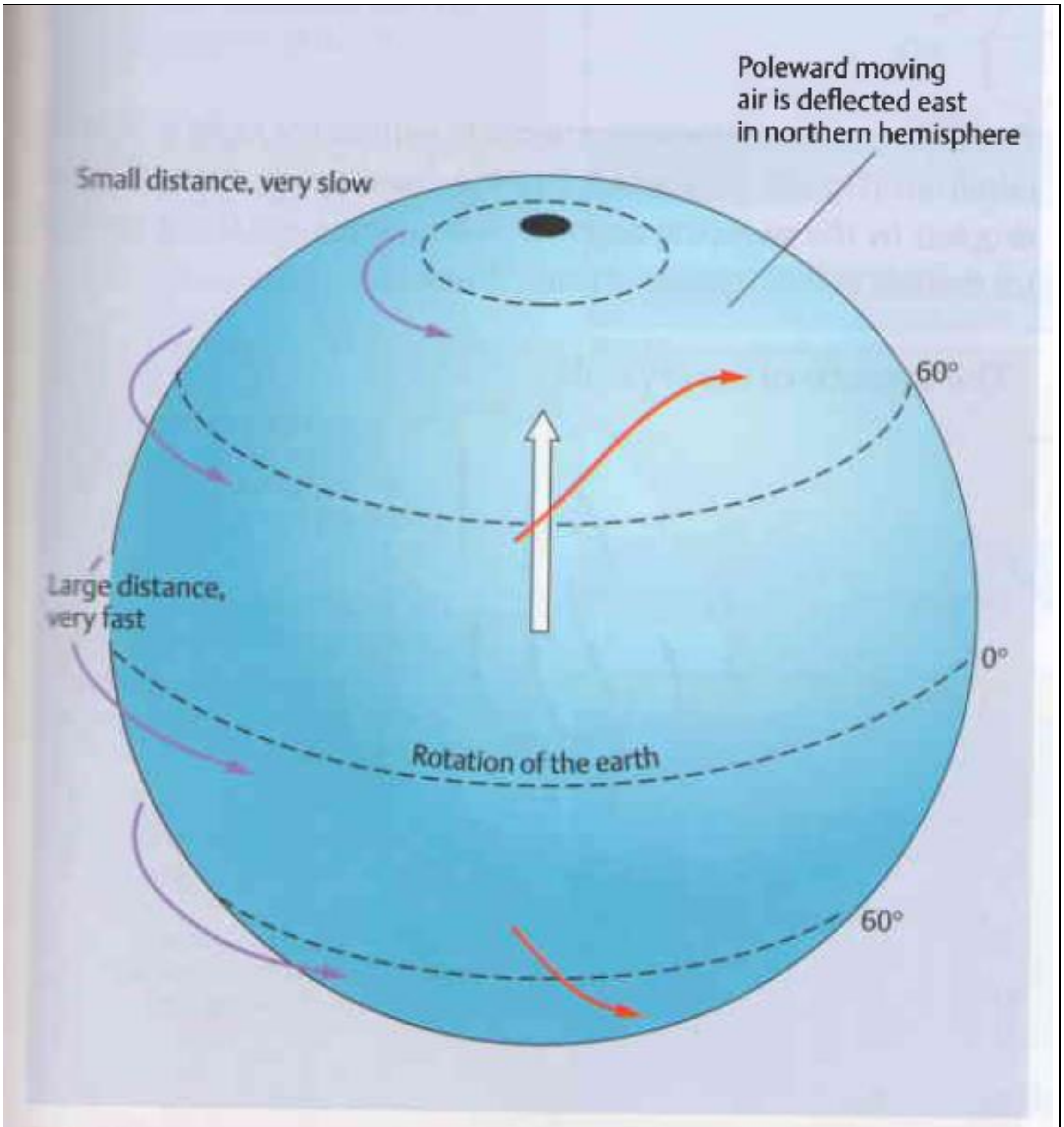
This is the force that acts on air as a result of its location in a pressure gradient. Air moves from high to low pressure, and the wind speed and direction are controlled by pressure differences and the steepness of pressure gradients (Buckle, 1996). High pressure areas are regions of air divergence while low pressure areas are regions of air convergence. Wind speed increases with the value of the pressure gradient or barometric slope.




Pressure Gradient Force (After Nagle, 2000)

## Coriolis Effect

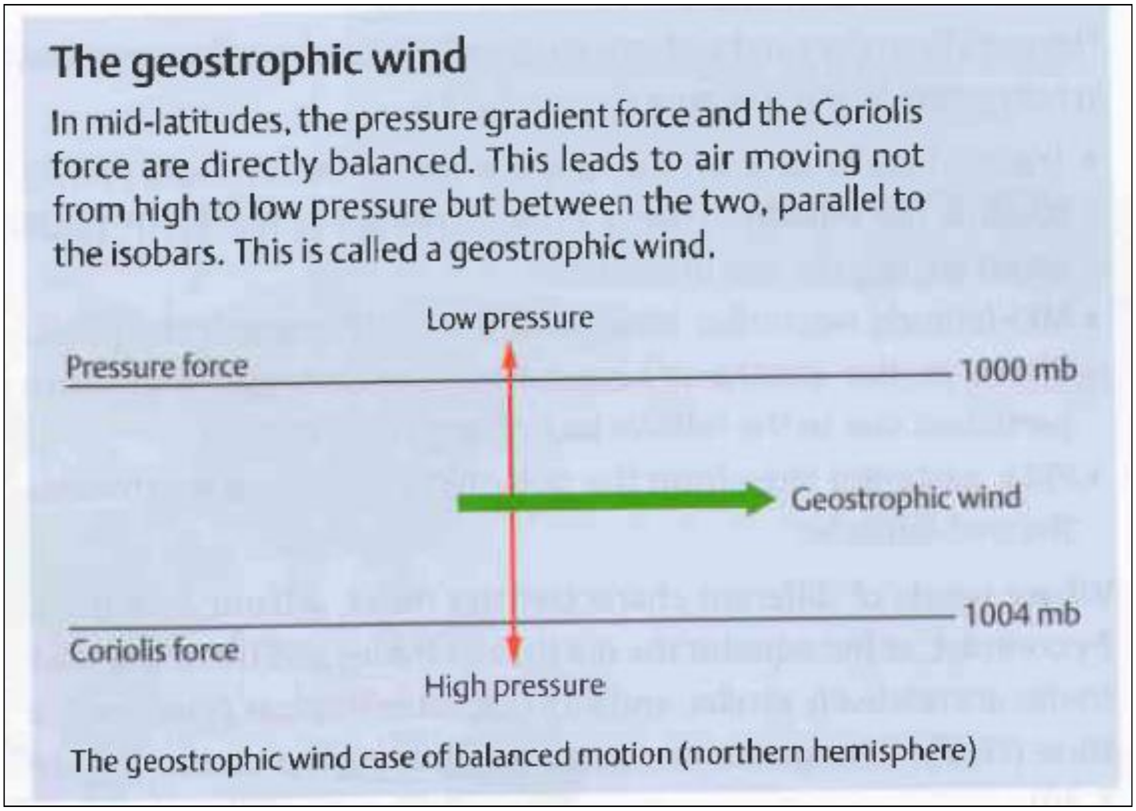
This is an apparent acceleration of air as result of earth rotation. The rotation causes winds to be deflected to the right in the northern hemisphere and to the left in the southern hemisphere. Rotation also causes an acceleration which may take the form of a change in sped or direction, or both (Buckle, 1996). As such in the northern hemisphere, winds blow with low pressure to the left while in the southern hemisphere it blows with low pressure to the right. The magnitude of the Coriolis force increases as the wind speed increases reaching a maximum at the poles. At these points, rotation causes a point on the horizon to pass through a full circle. Conversely at the equator there is no deflection because the horizontal surface over which the wind blows does not rotate since at the equator it lies parallel to the axis of Earth spin. Differences in the earth's rotational velocity also cause the Coriolis deflection to vary with latitude. The deflection caused by the Coriolis Effect gives rise to the Centrifugal effect, the tendency for an orbiting object to fly off in a straight line. Centrifugal force increases with speed, causing air moving east in line with earth rotation to be reduced in apparent weight, so pulling air equator wards. Westward moving air, by contrast, has less centrifugal force than the earth, thereby causing a pole-ward deflection.



Coriolis Effect (After Nagle, 2000)

## *Geostrophic Wind*

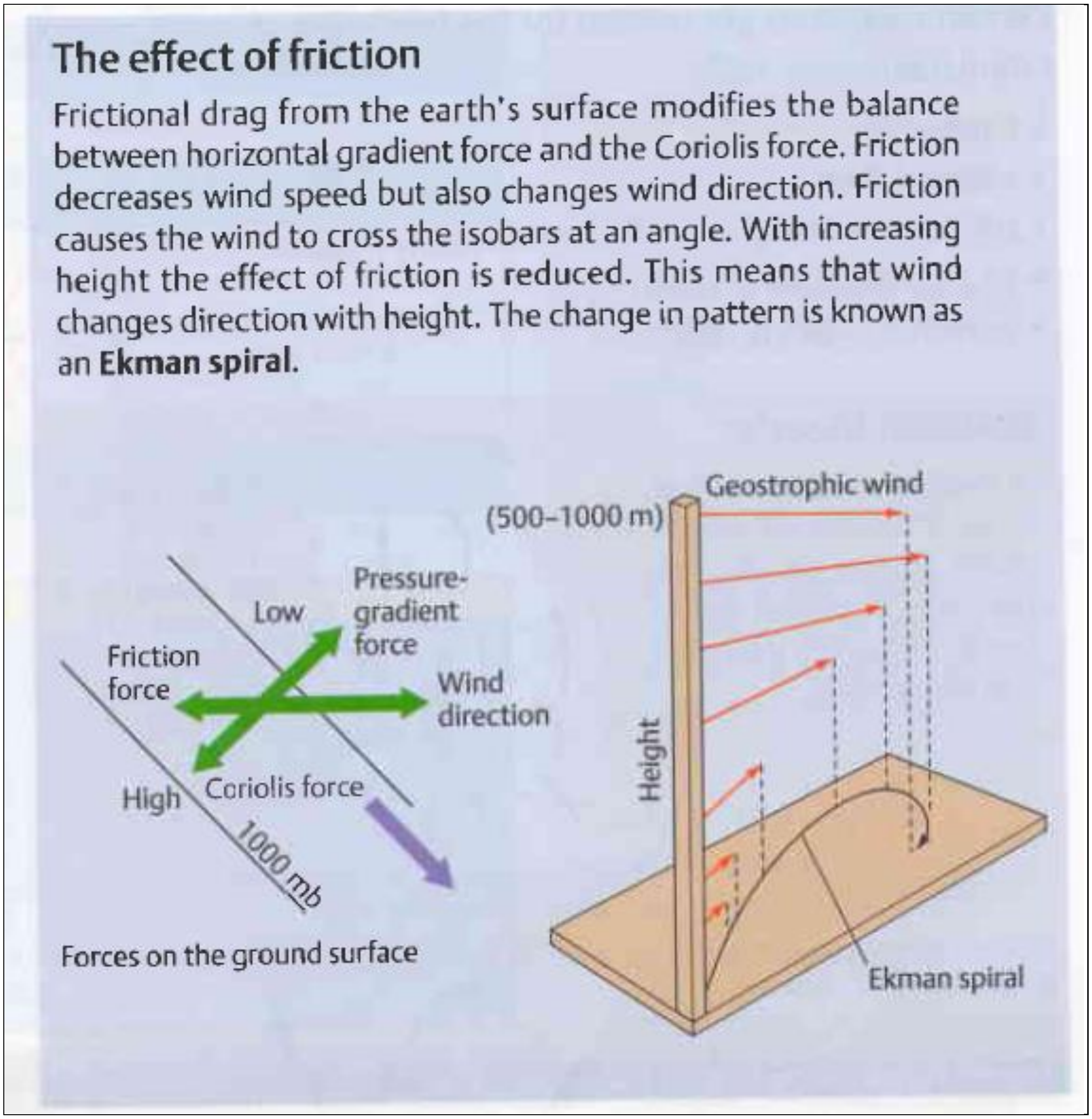
Winds in the upper troposphere, unaffected by friction with the earth's surface, show that there is a balance between the forces exerted by the pressure gradient and the Coriolis deflection Waugh, 2009). The result is the geostrophic wind which blows parallel to isobars. This is because pressure gradient force directs air across the isobars from high to low pressure, but the Coriolis force deflects the wind to the left or right, depending on the hemisphere.




Geostrophic Wind (After Nagle, 2000)

## Surface Friction

The ground surface exerts a frictional drag on wind-flow, causing deceleration in the boundary layer. This reduces wind speed and changes the direction of the wind. These two vary with surface roughness. The friction can come from vegetation, buildings or mountains among other things.




Surface Friction (After Nagle, 2000)

# *The Major Surface Wind System*

At the surface global winds exhibit a latitudinal pattern. Zonal flow follows the parallels of latitudes. This gives rise to the formation of tropical easterlies, mid-latitude westerlies, equatorial westerlies and polar westerlies.

## The Tropical Easterlies/ Trades

These are the winds which blow from the subtropical highs towards the equatorial troughs. Their speed and direction are constant throughout the year. They are diverted to the right in the Northern Hemisphere (NH) and to the left in Southern Hemisphere (SH). The trades give way to a strong monsoon circulation in the Indian and Pacific Oceans. These winds are better developed over the oceans than the land where they are interrupted by winds from other directions and tropical cyclones. Their strengths vary with seasons, reaching a maximum when the subtropical high is most intense. The seasonal shift of the anticyclone causes the wind belt to migrate annually North and South through about 7degrees of latitude.

## The Mid-latitude Westerlies

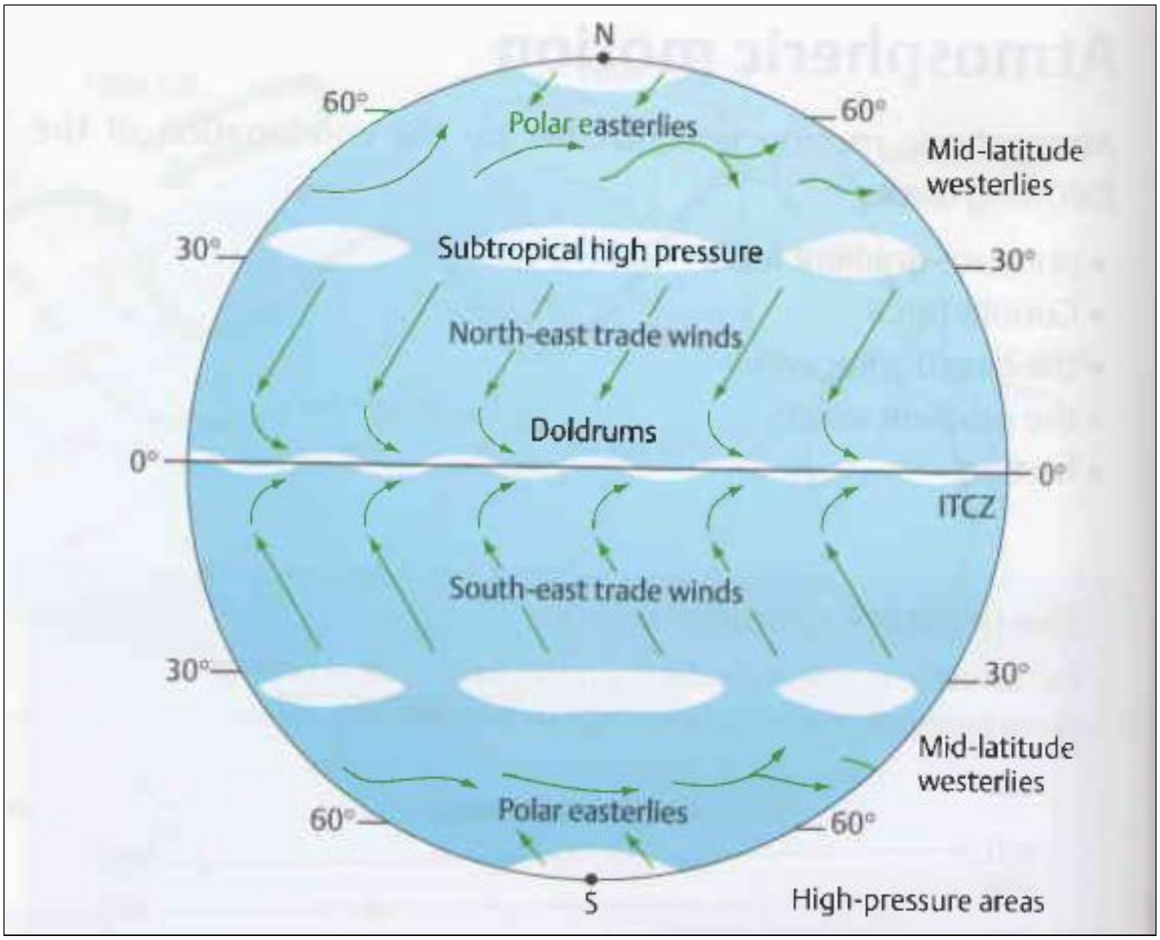
These winds blow pole-wards from the subtropical highs, between 35 and 65 degrees. They are again affected by the rotation of the earth whereby they blow from SW in the NH and NW in the SH. These winds are well developed in the SH where there are vast areas of oceans. In the NH the Mid-latitude Westerly winds vary in direction owing to the passage of migratory high and low pressure centres which disrupt their regular westerly flow.

## The Polar Easterlies

These are dry, cold air flows from the polar region to about 65 degrees N and 65 degrees S. The easterlies are relatively weak and do not extend into the troposphere. Their direction and speed as more persistent in the SH than in the NH due to the permanent anticyclones in the Antarctica.

## The Equatorial Westerlies

The Equatorial Westerlies are variable surface winds that only appear in a few parts of the equatorial zone and during certain times of the year. The winds are seasonal, fluctuating back and forth with the path of the overhead sun, and developing as easterly air, crossing the Equator is deflected to the West by the reversal of the Coriolis Effect.




Global wind patterns (After Nagle, 2000)

# ***The Local Winds***

Winds are the currents in the air, huge swirling streams of invisible energy blowing storms around the globe, building giant waves at sea, and raising dust in the desert. Winds drive the ocean currents and the major agents for the transfer of latent heat and sensible heat.

The examples of meso-scale circulations induced by local temperature differences are the land and sea breeze, the mountain valley wind systems, the harmattan and the monsoon.

# The land and Sea Breeze

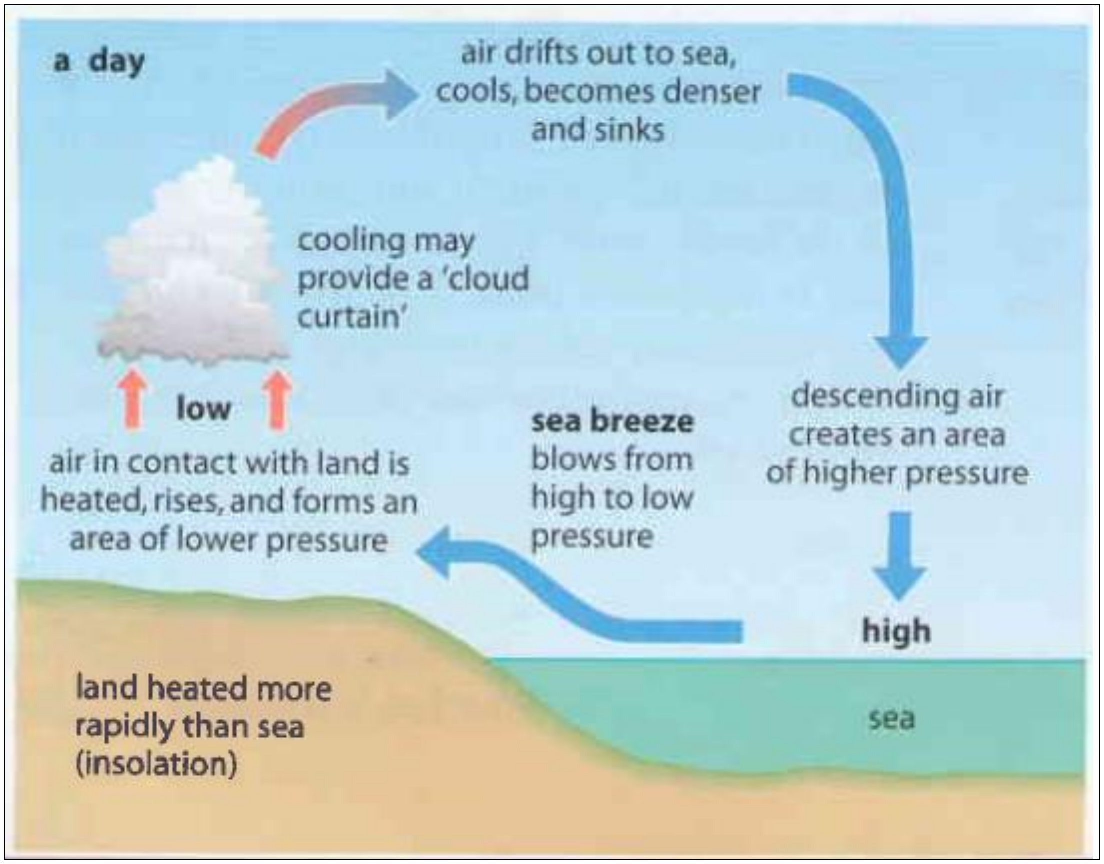
The land and water surfaces have different thermal responses to the same amount of insolation and most land surfaces have a greater daily temperature range than water bodies. The water bodies tend to exhibit much more uniform temperatures throughout the day (Musk, 1988). The following are the reasons for these differences:

- Whereas the insolation is used to heat up the land surface, the same amount of energy is absorbed by a large volume of water.
- The volume of water being heated by the insolation is further increased by the natural motion of water and the convection within it.
- Much of the energy is used as latent heat (through the evaporation of water) rather than direct heating of the water. Evaporation has the effect of cooling the water surface, which further encourages surface mixing.
- The thermal capacity of water is exceptionally large.

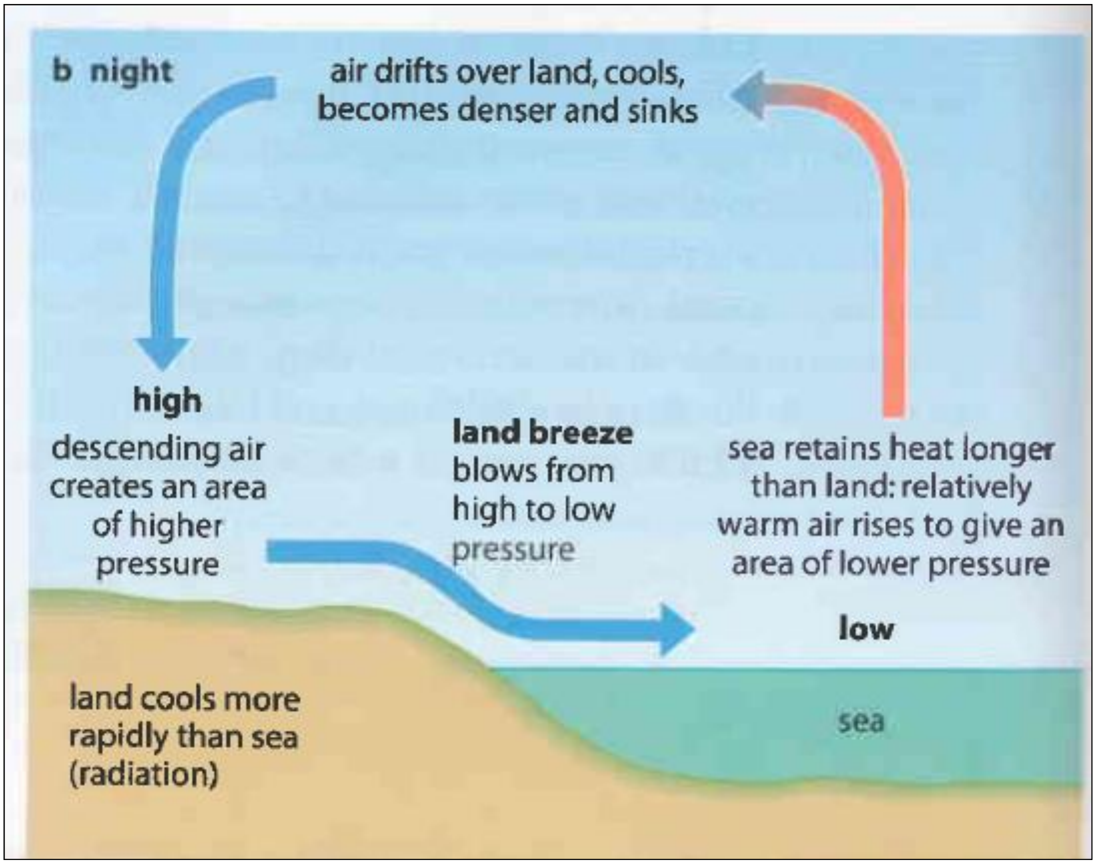
For these reasons, temperature gradients are set up between the land and the sea, and these have a diurnal reversal (the land is warmer than the sea during the day and the sea warmer than the land during the night).

The sea breeze blows from the sea to the land during the day, with a compensating offshore flow aloft. The land breeze blows from the land to the sea at night (again with a weak return flow aloft). Similar circulations can develop in the vicinity of large inland water bodies such as Lake Kariba and the Mazvikadei Dam.

The structures of the sea and land breezes are shown schematically below.



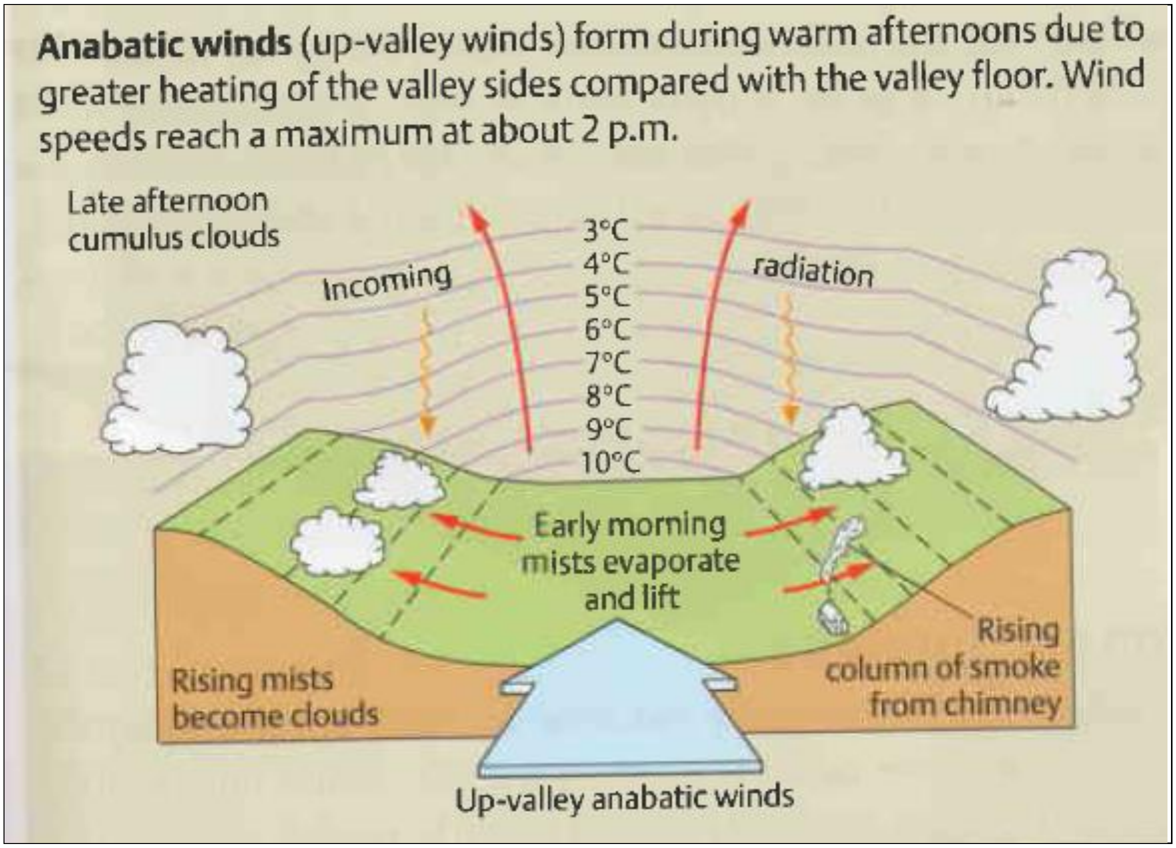

Sea Breeze (After Waugh, 2009




Land Breeze (After Waugh, 2009).

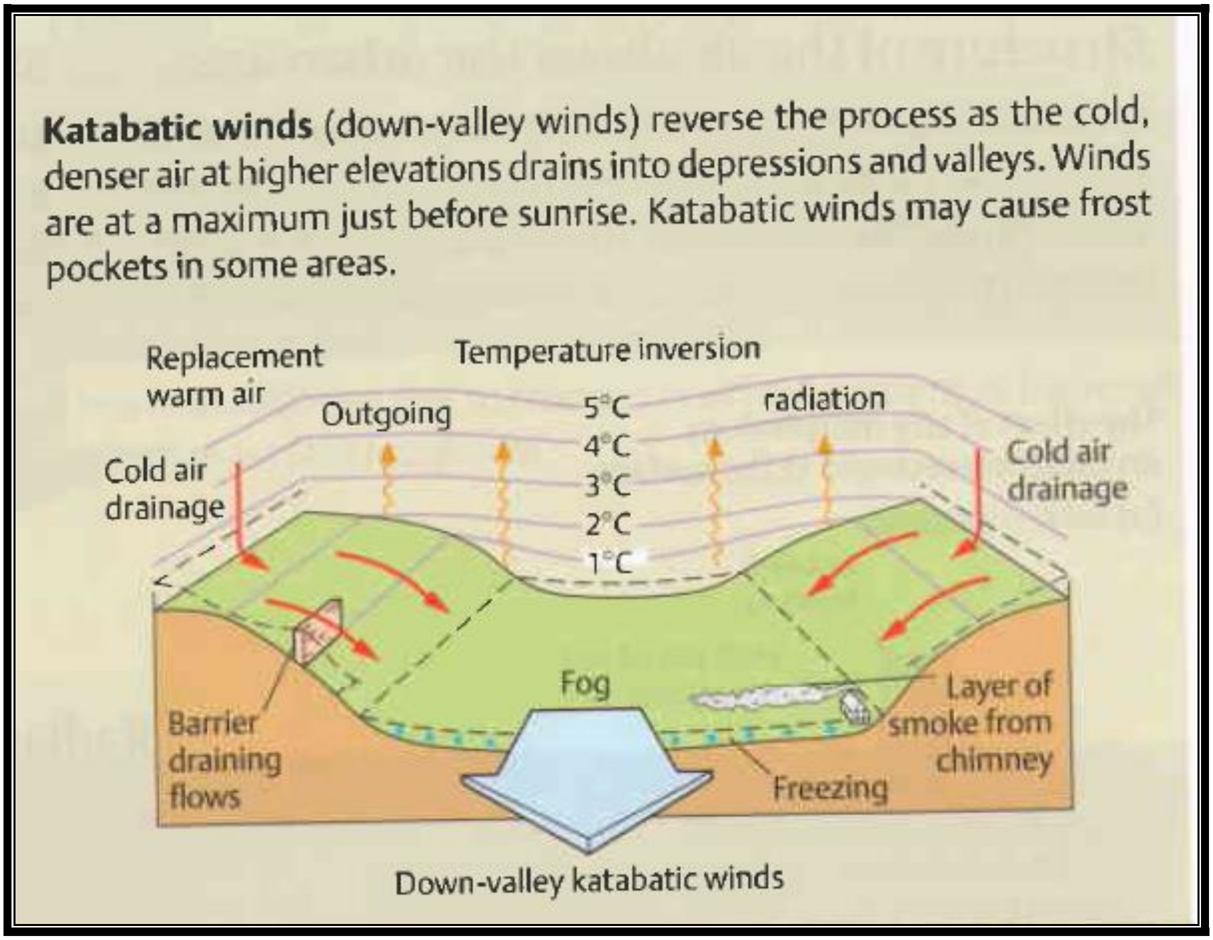
# *The Mountain and Valley Winds*

The differential heating of the mountains and valleys may generate its own local mountain valley wind circulations with the daily rhythm. At night the valley surface radiate long-wave energy to space, and cool down, chilling the air in contact with them. This cool, stable air drains downhill into the valley bottom because of its density, producing a downslope *katabatic flow*, known as the *mountain wind* (Musk, 1988). The accumulation of cold, dense air in hollows, depressions and valley bottoms can sometimes lead to frost in the low-lying land, or valley fog when the air is sufficiently moist.




Anabatic Wind (After Nagle, 2000)

During the day, mountain slopes facing the sun are heated; the air in contact with the slopes expand and rises up the valley and the mountain slopes because of its buoyancy as an *anabatic flow* or *valley wind.* This may lead to the formation of convective clouds over the mountains during the afternoon, with clear conditions over the valley (Musk, 1988). The winds associated with this anabatic flow are greatest at around 2pm, the time of the greatest influence from surface heating. The flow is thermally direct, reversible circulation similar to the land and sea breeze.




Katabatic Wind (After Nagle, 2000)

Musk (1988) argues that the intensity of the circulation in a particular valley depends on the following:

- Orientation of the valley (south-facing slopes are heated more effectively than north-facing ones in the northern hemisphere).
- The magnitude of the relief (i.e. Difference in height of the valley bottom and adjacent ridge tops).
- The type and amount of vegetation cover (bare rock heats up more rapidly than green vegetation, while the later retards surface flows more because of the effect of friction, particularly on forested slopes).

- The geometry of the valley (whether straight or meandering in form, whether the valley sides are steep or gently sloping; and whether there are constrictions or obstacles to surface flows in the valley).
- The surface conditions (the circulations are bets developed when the ground is dry rather than wet, but a snow cover will accentuate the nocturnal down-valley flow).

## The Harmattan

This is the prevailing wind of West Africa during the dry season (November to March). It originates from the Saharan Anticyclone and dominates North Africa. The wind is very dry, with humidity levels below 10%. It causes dust haze which reduces visibility and disrupts air traffic. The wind encourages skin dryness, throat irritation, watery eyes and bronchitis.

# THE GENERAL CIRCULATION MODELS OF THE ATMOSPHERE

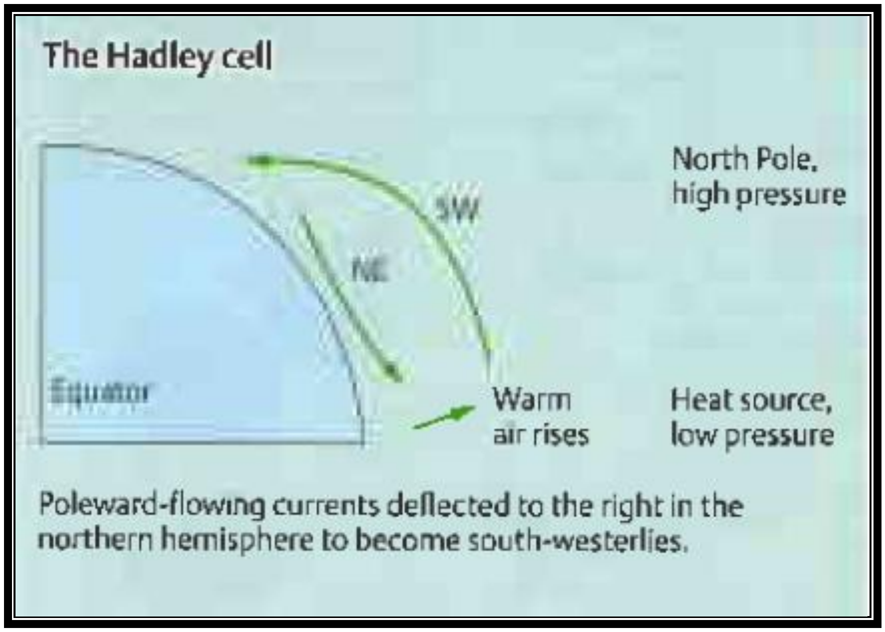
## The General Circulation Models

The fundamental factor that results in atmospheric circulation is the unequal heating with latitude. There is a surplus of energy between 38 $^{0}$S and 38 $^{0}$N, whereas there is a deficit of energy pole-wards. Energy is transferred from low latitude to high latitude to balance this unequal heating (Nagle, 2000). This gives rise to convectional cells which then form circulation models. Two of these are described below: the Hadley and the Ferrel Models.

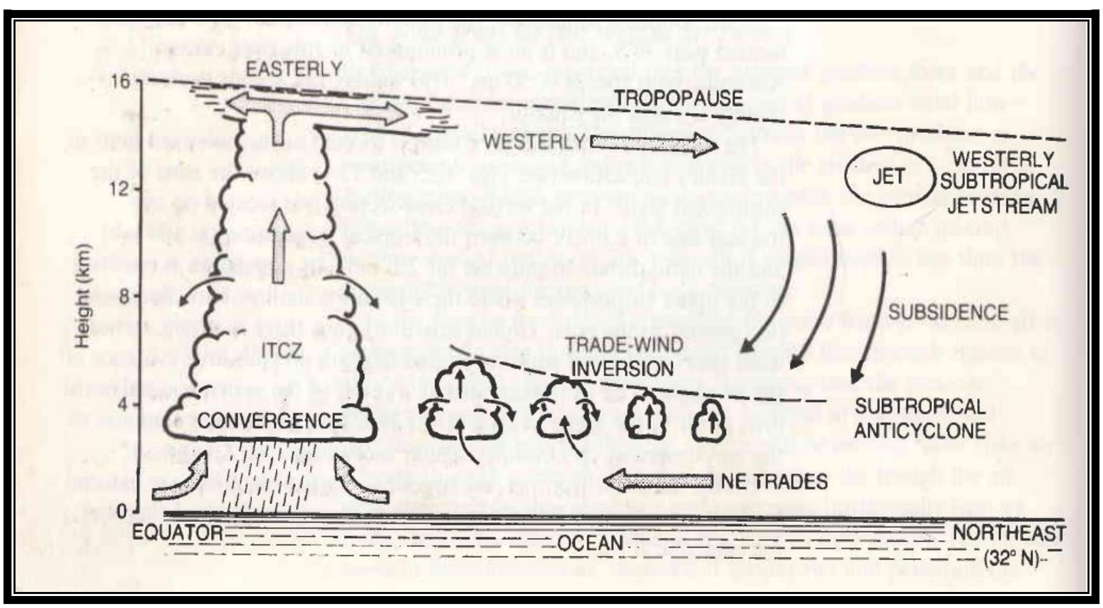
## The Hadley (Single Cell) Model

This single-cell model was proposed by Halley (1686) and Hadley (1735). The model is an earliest one which explains the atmospheric circulation of air. It suggest that intense heating over the Equator leads to rising air which cools and spread out to the north and south with subsequent descent in polar regions. Warm air rising at the Equator creates low pressure in that region. The rising air then descends in the poles, creating high pressure there. The circulation is then completed by air returning to the Equator at lower levels (Fig. 2.1). Figure 2.2 shows a schematic cross-section along the axis of the trade winds indicating the main components of the Hadley circulation over the oceans of the northern hemisphere. This model was based on the following assumptions:

- The earth was not rotating
- The earth was not tilted
- The surface of the earth consisted entirely of land or water only.




Hadley (Single Cell) Model (After Nagle, 2000)




A schematic cross-section along the axis of the trade winds indicating the main components of the Hadley circulation over the oceans of the northern hemisphere (After Musk, 1988)

# ***Ferrell model***

The model explains the general circulation of the atmosphere. It was proposed by Ferrell (1856) and Rossby (1941). As rising air in the Equator cools to the temperature of the surrounding environmental air, uplift ceases and it begins to move away from the Equator. Further cooling, increasing density and diversion by the Coriolis force cause air to slow down and subsides forming the **Hadley Cell**.

About thirty degrees North and South of the Equator, there is some convergence of this pole-ward moving air resulting from the decreasing circumference of the globe over which the air is flowing. The upper air convergence leads to a rise in surface pressure in form of subtropical high pressure belt (horse latitude) with its clear skies and dry stable conditions. Some of the air returns to the Equator and is deflected to the right in the northern hemisphere and to the left in the southern hemisphere.

Air moving equator-wards from the poles converge with air moving pole wards around 60 $^{0}$N and 60 $^{0}$S to form another circulation called the **Ferrell Cell.**

The remaining cell- **Polar Cell** is primarily a consequent of intensely cold air settling over Polar Regions. The cold air produces a surface zone of high pressure from which air diverges in the form of the Polar easterlies. This diverging air meets tropical air along the polar front and after rising, some return at higher levels to subside over the poles. Fig 2.4 shows the cross section of the Ferrel Model from the Equator to the North Pole in the northern hemisphere.

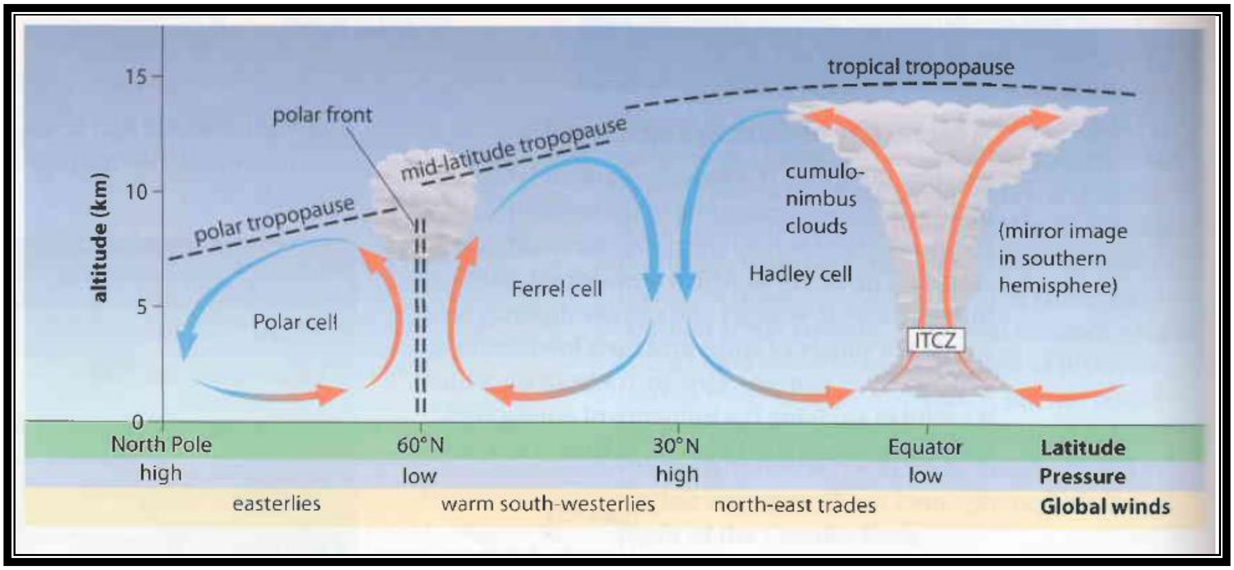



Fig 4.3 Schematic cross-section from the Equator to the north-pole indicating the main components of the Ferrel model of the northern hemisphere (After Waugh, 2009)

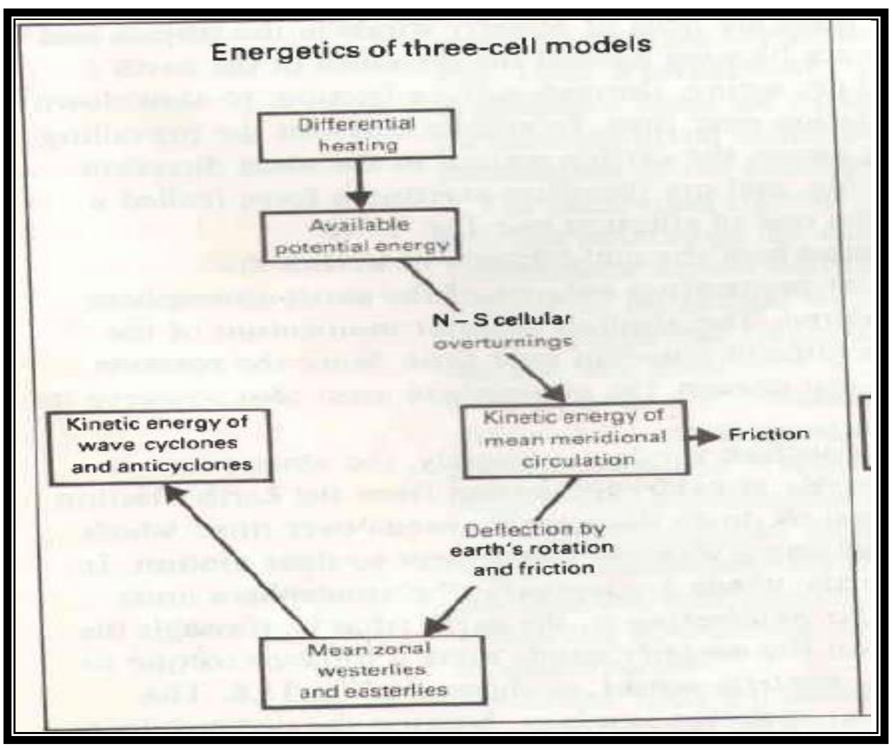
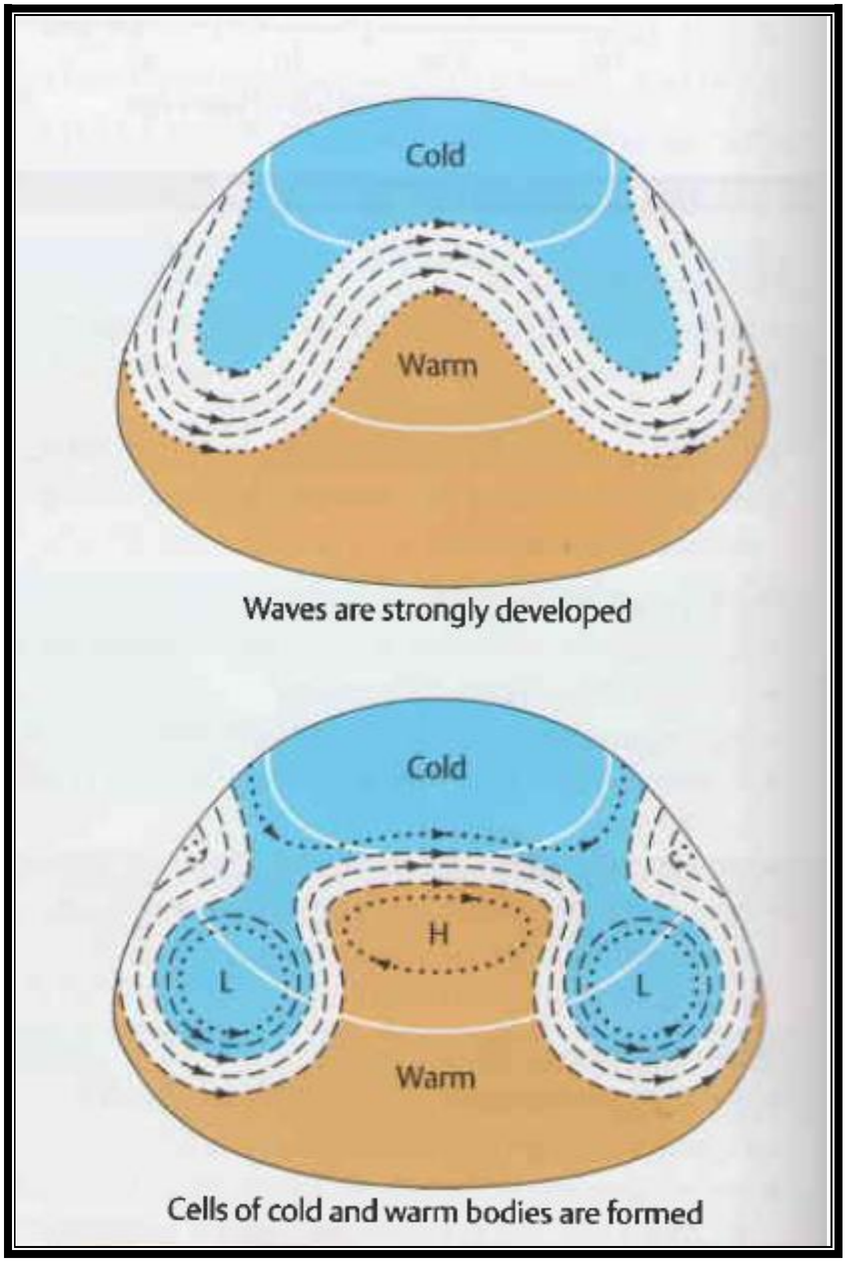


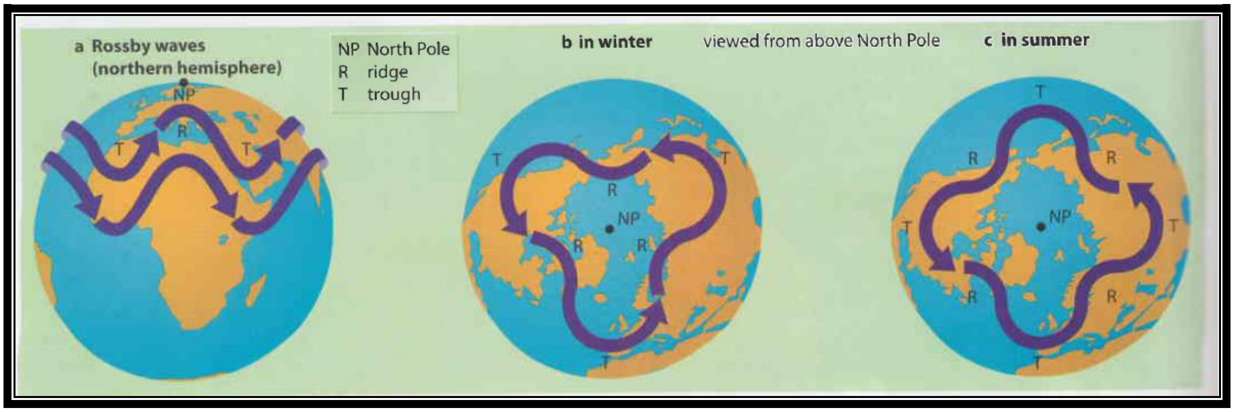

Fig 4.4 Energetics of a three cell model (After Musk, 1988)

## The Rossby Waves

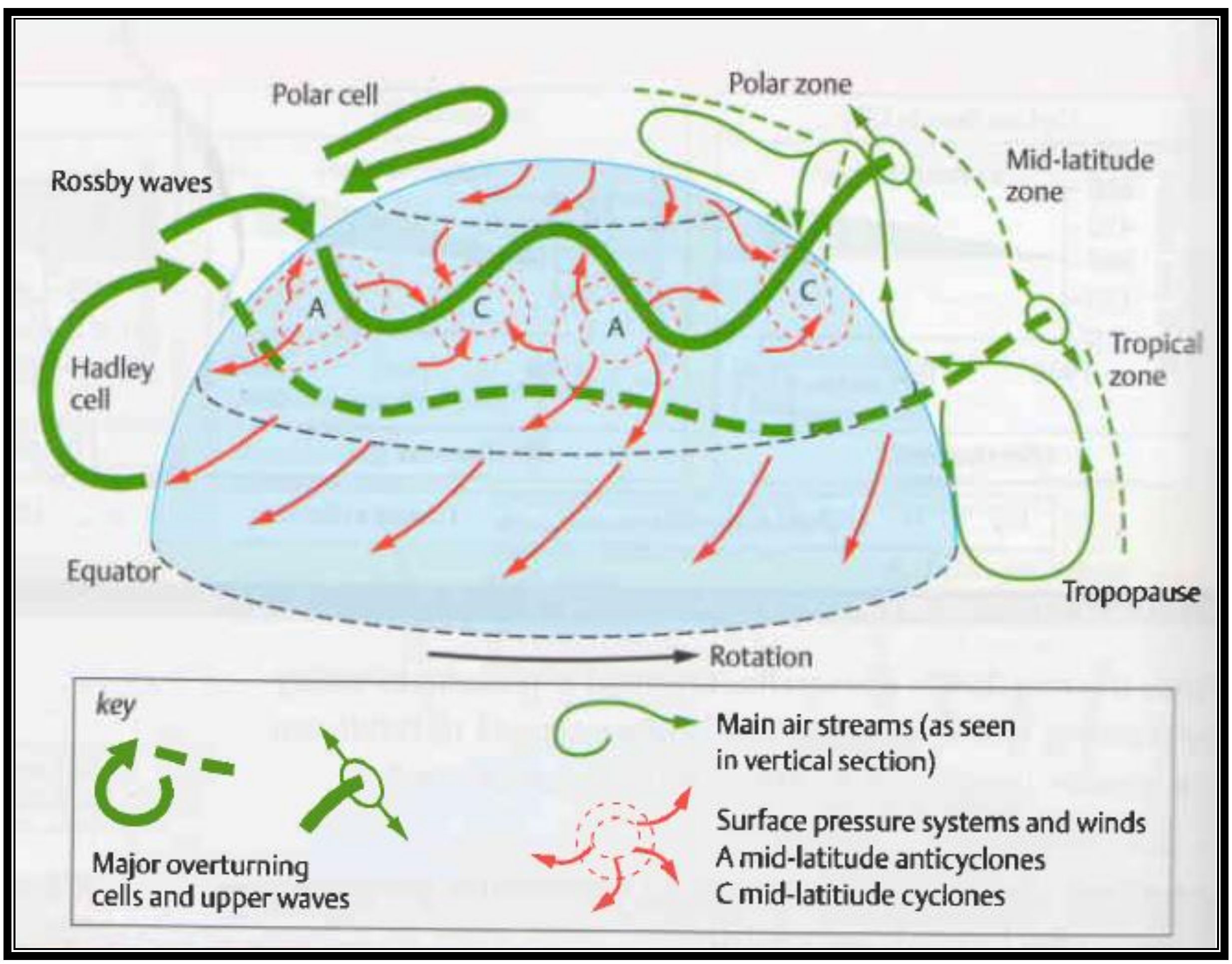
The Rossby Waves are a belt of upper-air westerly which often follows a meandering path. The number of waves varies according to seasons, with usually 4-6 in summer and 3 in winter. They form a complete pattern around the globe. The Rossby waves can be very large indeed with wavelengths of up to 8 000km. The waves circulate the earth in the upper-level westerly flow. Fig 2.4 below shows the development of Rossby waves whereas Fig 2.5 illustrates fully developed Rossby waves in the northern hemisphere during the winter and summer seasons.




Formation of Rossby Waves (After Nagle, 2000)




Rossby Waves (a) in the Northern Hemisphere, (b) in winter and (c) summer as viewed from above the North Pole (After Nagle, 2000).




The generalised model of global circulation (After Nagle, 2000)

# Jet Streams

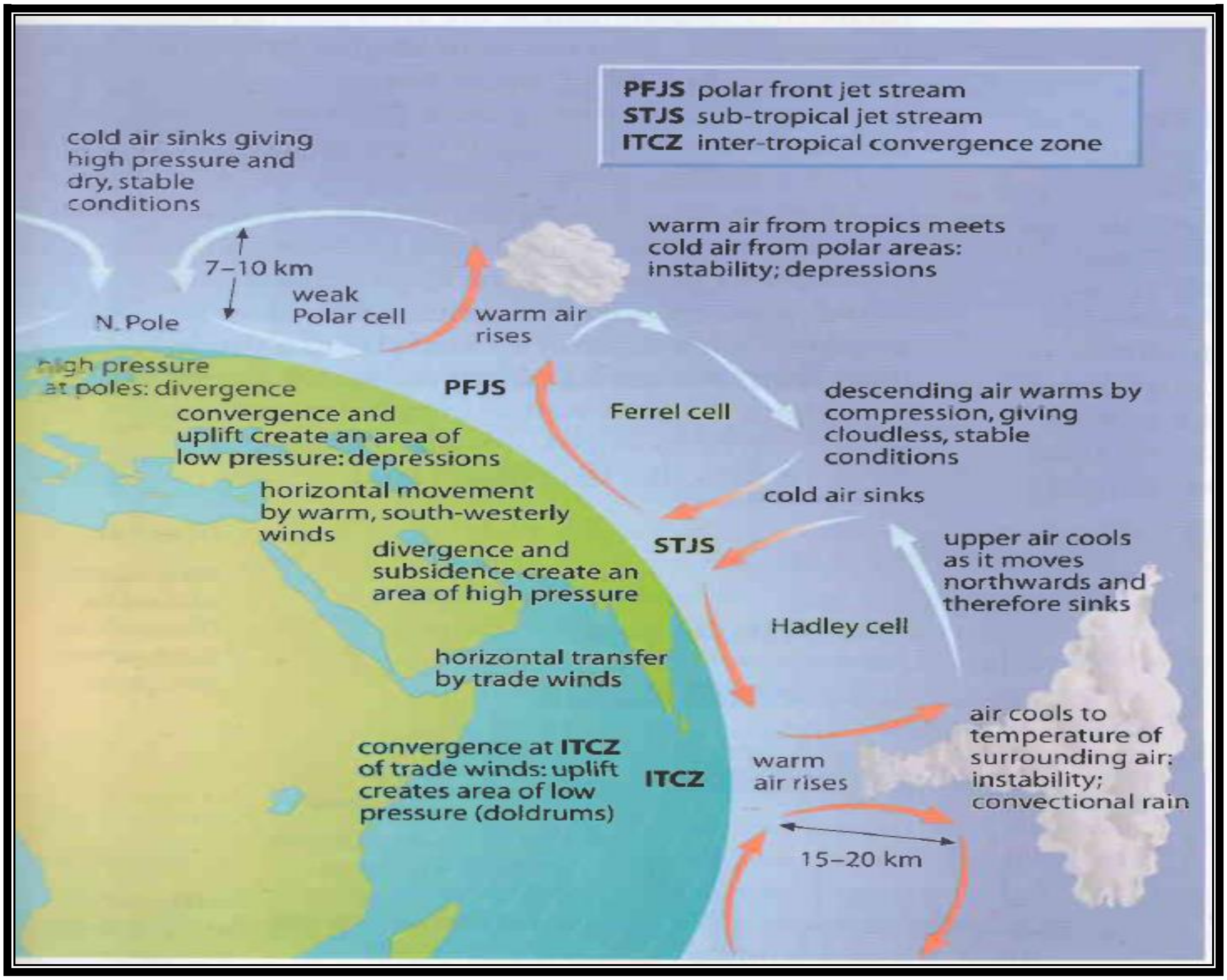
Jet Streams are narrow currents of extremely fast-moving air within the Rossby waves. Their speed exceeds 230Km/h or 50-100m/s in the low density air near the tropo-pause. Above the tropo-pause, temperature and pressure gradients change, causing the strength of the westerly airflow (Rossby Waves) to decline and the jets to disappear. The speed maxima within the jets are called jet streaks. The length of the jet streams can be of several thousand Km, their width- hundreds of Km and the thickness or depth is rarely more than a few Km.

## Types of Jet Streams

Three types of jet streams are identified on the earth. These are:

- The Polar Front Jet Stream (PFJS)
- The Sub tropical Jet Streams
- The Tropical Easterly Jet

the formation of the Ferrel Circulation.




Ferrel (Three-Cell) Model (After Waugh, 2009

# Air Masses

An airmass is a large body of air extending to thousands of square kilometres, whose temperature and humidity are fairly uniform. A cold air mass from the South Pole normally brings cold weather in Zimbabwe during the winter season. Air masses are classified according to:

- **The physical properties of air**. This determines the temperature of the air mass. An air mass can be cold or warm
- **The latitude of origin** e.g. polar air mass, tropical or equatorial air mass

- **The surface over which the air mass develops** which affects their moisture content. This gives rise to maritime and continental air masses.

The above classification is summarised in Table 9.1 below.

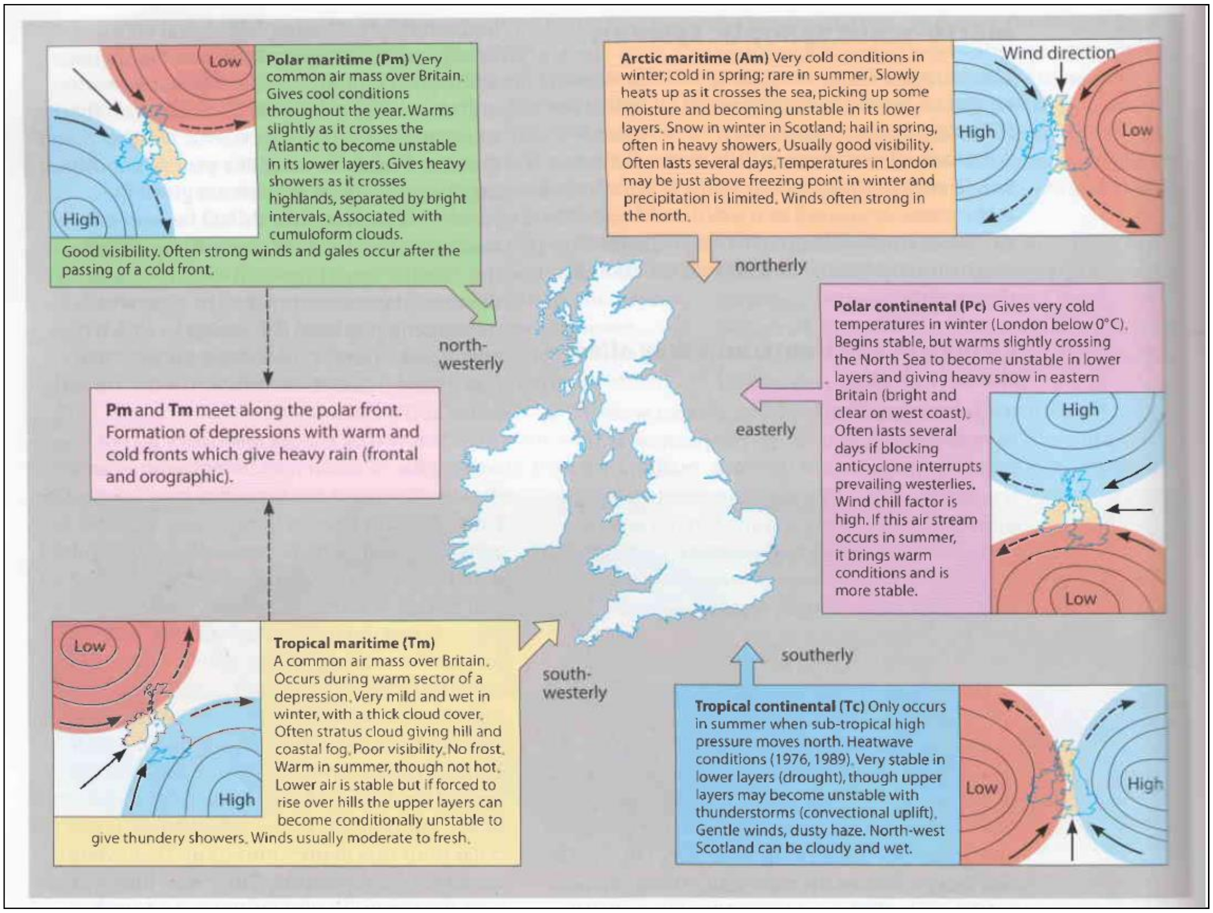
## Classification of Air masses (After Henderson-Sellers and Robinson, 1996)

| MAJOR GROUP | SUBGROUP | SOURCE REGIONS | PROPERTIES AT SOURCE |
|---|---|---|---|
| Polar (P) | Maritime Polar (mP)<br><br>Continental Polar (cP)<br><br>Arctic (A) or Antarctic (AA) | Oceans pole-ward of approx. $50^0$<br><br>Continents in vicinity of Arctic Circle; Antarctica<br><br>Polar regions | Cool, rather damp, unstable<br><br>Cold and dry, very stable<br><br>Cold, dry, stable |
| Tropical (T) | Maritime Tropical (mT)<br><br>Continental Tropical (cT)<br><br>Maritime Equatorial (mE) | Trade wind belt and subtropical oceans.<br><br>Low latitude deserts, chiefly Sahara and Australian deserts<br><br>Equatorial oceans | Moist and warm, stability variable: stable on east side of oceans, rather unstable on west.<br>Hot and very dry, unstable<br><br>Warm, moist, generally slightly stable |

# Air masses affecting the British weather

- Polar maritime air mass from northern Canada and the Arctic Ocean
- Arctic maritime air from the Arctic Ocean
- Polar continental air from Siberia high pressure area
- Tropical continental air from the Saharan sub-tropical high pressure area
- Tropical maritime air from the Azores high pressure area

**The diagram below shows these air masses affecting the British weather.**




Air masses and the British Weather (After Waugh, 2009)

# Airstreams affecting Zimbabwe

- The Zaire airflow/North-west monsoon
- North-east monsoon
- South-east trades

# ANTICYCLONES

An anticyclone is a large mass of subsiding air which produces an area of high pressure on the earth's surface. The source of the air is the upper atmosphere, where amounts of water vapour is limited. An anticyclone tends to be of types: warm and cold, depending on whether its origin is thermal or dynamic.

## COLD ANTICYCLONES

Cold anticyclones are thermal in origin, developing over continental interiors such as Siberia, Greenland and northern Canada in winter, and for the rest of the year over the poles (Musk, 1988). They are caused by persistent radiation cooling of the land surface and chilling of the atmosphere. The lower troposphere is cold, with normal temperatures at higher levels. Inversions are common and the air is very stable.

## WARM ANTICYCLONES

These are formed from the convergence of in the upper troposphere and subsidence beneath. This results in warmer than normal temperatures in the middle and lower troposphere. Mainly they develop in the subtropics and mid-latitude regions. Examples of them include the Azores, Sahara and North Pacific highs in the northern hemisphere and the South Atlantic, South pacific and South Indian Ocean highs in the southern hemisphere. The major characteristics of the cold and warm anticyclones are summarised below:

Table 9.2 Major characteristics of cold and warm anticyclones (After Musk (1988)

| *Feature* | *Cold Anticyclone* | *Warm Anticyclone* |
|---|---|---|
| Characteristics | Shallow anticyclone with circulation extending up to 2-3 km | Deep anticyclone, slow-moving and associated with blocking in higher latitudes |
| Origin | | Convergence in the upper troposphere with subsidence |

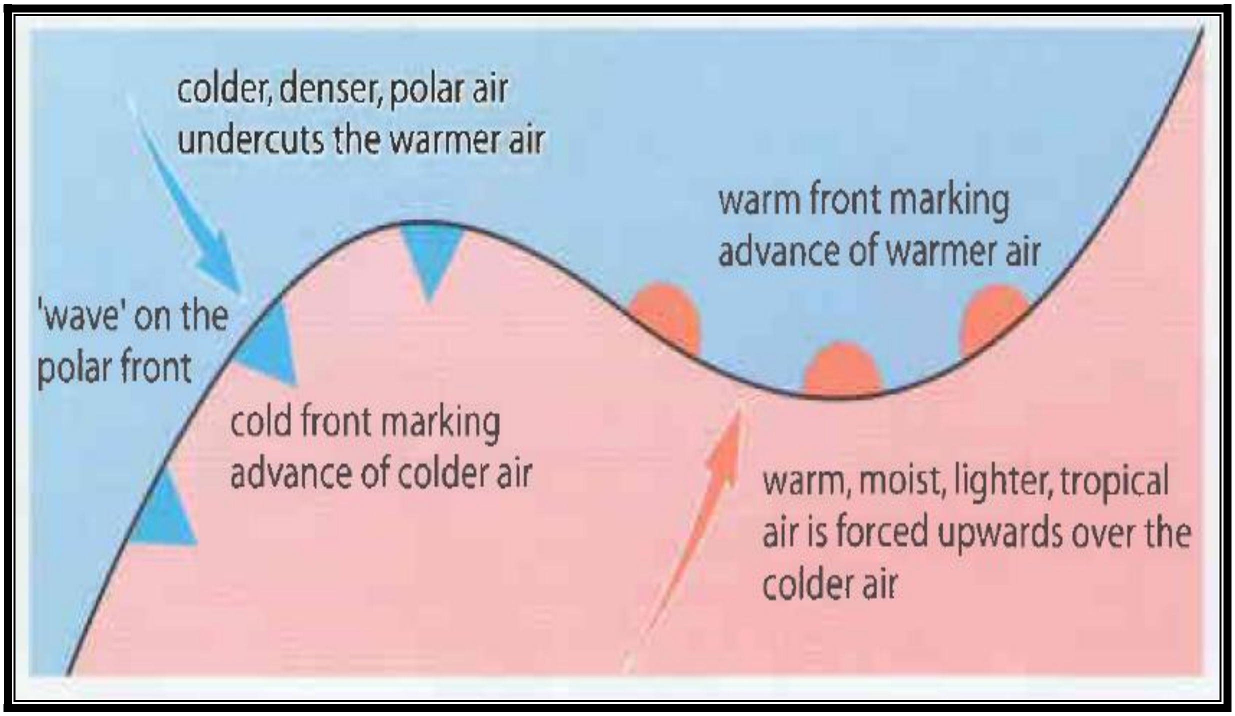
| | | |
|---|---|---|
| | Persistent radiation cooling of land surface and chilling of the overlying atmosphere | beneath throughout the depth of the troposphere |
| Intensity | Decreases with height | Increases with height |
| Region of formation | High latitude continental interiors, e.g. Siberia and Greenland | Subtropical belt 30-40 latitude e.g. Azores, Sahara and southern hemispheric subtropical oceanic anticyclones, often with ridges extending pole-wards |
| **Troposphere** | Cold | Warm |
| Tropopause | Low | High |
| Stratosphere | Warm | Cold |
| Persistence | More mobile in location and less persistent than warm anticyclones | Days, weeks or months |

# DEPRESSION

A depression is a polar front boundary between cold and warm air (Nagle, 2000). Depressions follow a life cycle in which the main stages can be identified as: embryo, maturity and decay.

## Embryo

This stage begins as a small wave on the polar front. The warm, moist tropical air (Tm) meets colder, drier polar air (Pm). The convergence of the two air masses results in the warmer, less dense air being forced to rise in a spiral movement. The upward movement results in less air at the Earth's surface creating an area of low pressure. The developing depression at this time moves in a north-easterly direction under the influence of the upper westerlies, i.e. the polar front jet stream (PFJS) (Waugh, 2009).

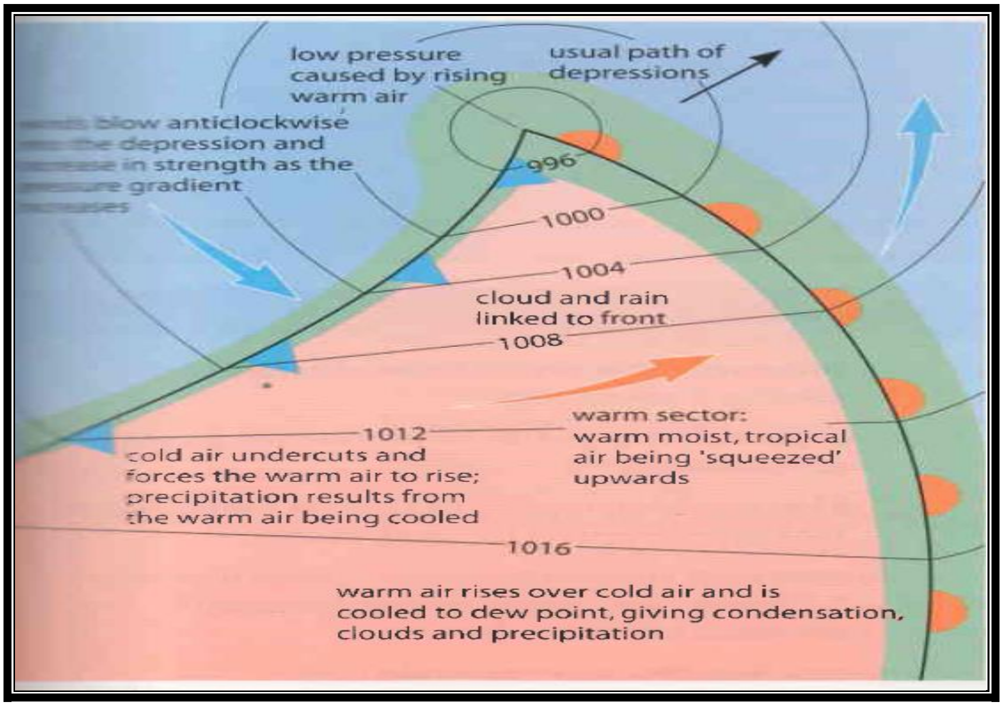



**Embryo stage of a depression (After Waugh, 2009**).

## *Maturity*

Pressure continues to fall as more warm air in the warm sector, is forced to rise. As pressure fall, the pressure gradient steepens the inward blowing wind increase in strength. Due to the Coriolis force these anticlockwise blowing winds come from the south west. Continued uplift of warm air will result in cooling and precipitation as the clouds become both thicker and lower (Waugh, 2009).

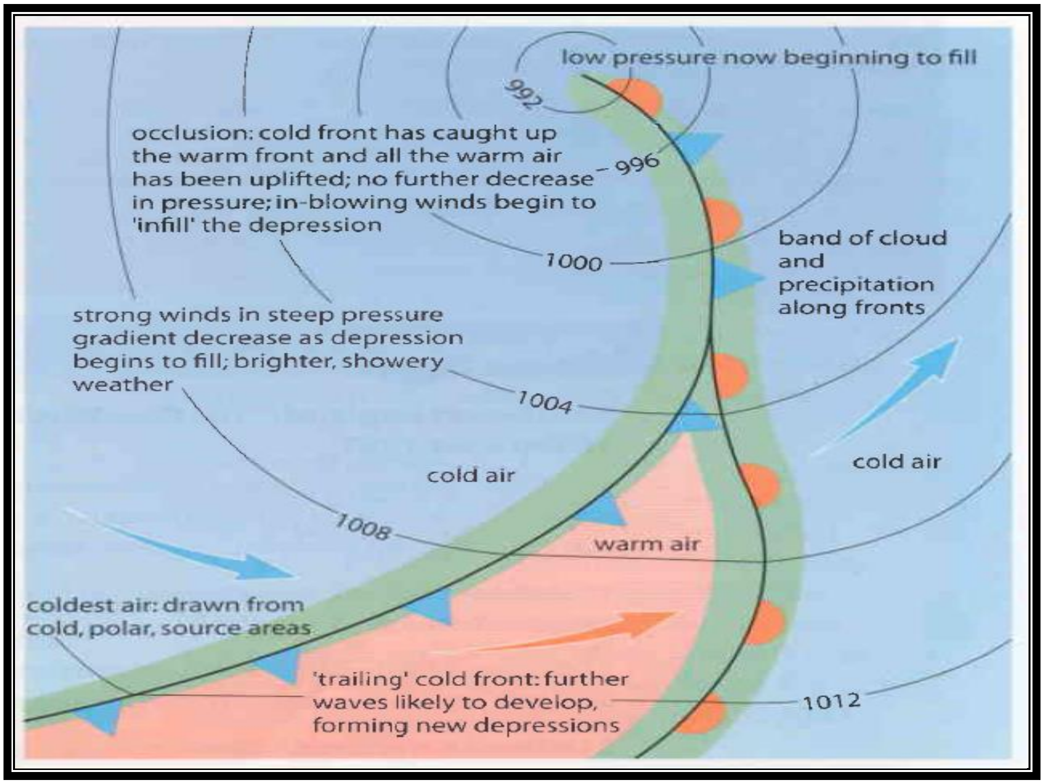
As temperature rise and the uplift of air decrease within the warm sector, there are less chances of precipitation and the low cloud may break to give some sunshine.




Maturity Stage of a Depression (After Waugh, 2009).

## *Decay*

The depression begins to decay when the cold front catches up the warm front to form an occlusion or occluded front. By this stage the tropical air (Tm) will have been squeezed upwards leaving no warm sector at the ground level. As the uplift of air is reduced, so too are the amount of condensation, precipitation and cloud cover. Pressure rises and wind speeds decrease as the cold air replaces the uplifted air and 'in fills' the depression (Waugh, 2009).




**Decay stage of a depression (After Waugh, 2009).**

Anticyclones have characteristics opposite to (or 'anti') those found in a depression or cyclone. These contrasts are summarised in the table below:

The Contrasting features of anticyclones and depressions (After Musk, 1988)

| **Feature** | **Anticyclone** | **Depression** |
|---|---|---|
| Surface pressure | High | Low |
| Wind direction | Anti-cyclonic (clockwise*) | Cyclonic (anticlockwise*) |
| Airflow | Divergent at surface (convergent aloft) | Converges at surface (diverges aloft) |
| Vertical air motion | Subsides | Rises |
| Wind speed | Weak | Moderate to strong |
| Precipitation | Generally dry | Wet |

| | | |
|---|---|---|
| Cloudiness | Stratus or no cloud | Cloudy |
| Stability | Stable air, with a subsidence inversion aloft | May be unstable |
| Temperature gradient | Little temperature contrast across the high | Strong temperature contrasts, especially at the fronts |
| Speed of movement | Slow- moving or stagnant | Generally mobile, moving west-east |

*in the northern hemisphere

# Tropical Cyclones

A cyclone is an intense, rotating, circular storm at the core of which lies closed low pressure centre  and which in excess of 50m/sec (Buckle, 1996). The pressure can be as low as 950mb, but they are different in character and intensity. Tropical cyclones are the most devastating and frightening of all natural phenomena. They are known as hurricanes in North America and the Caribbean, as typhoons in the western North Pacific and cyclones in Africa.

The cyclones are strictly oceanic phenomena, and tend to die out over the land. They form only in those oceans where sea temperatures are at least 26$^{o}$C, and where there is a reasonably deep layer of warm water down to 60-70m or more. Cyclones only occur in barotropic atmospheric conditions (where temperature, pressure, lapse rate and humidity are fairly uniform over large areas) during late summer and early autumn (Musk, 1988). This happens between latitudes 5$^{o}$C and 20$^{o}$C north and south of the equator. They are hardly found within 5$^{o}$C of the equator, where the latitudinal value of the Coriolis parameter is of insufficient magnitude to allow balanced geostrophic flow to develop.

For the hurricane to move, there must be a continuous source of heat to maintain the rising air currents; there must be a large supply of moisture to provide latent heat, released by condensation, to drive the storm and provide heavy rainfall.

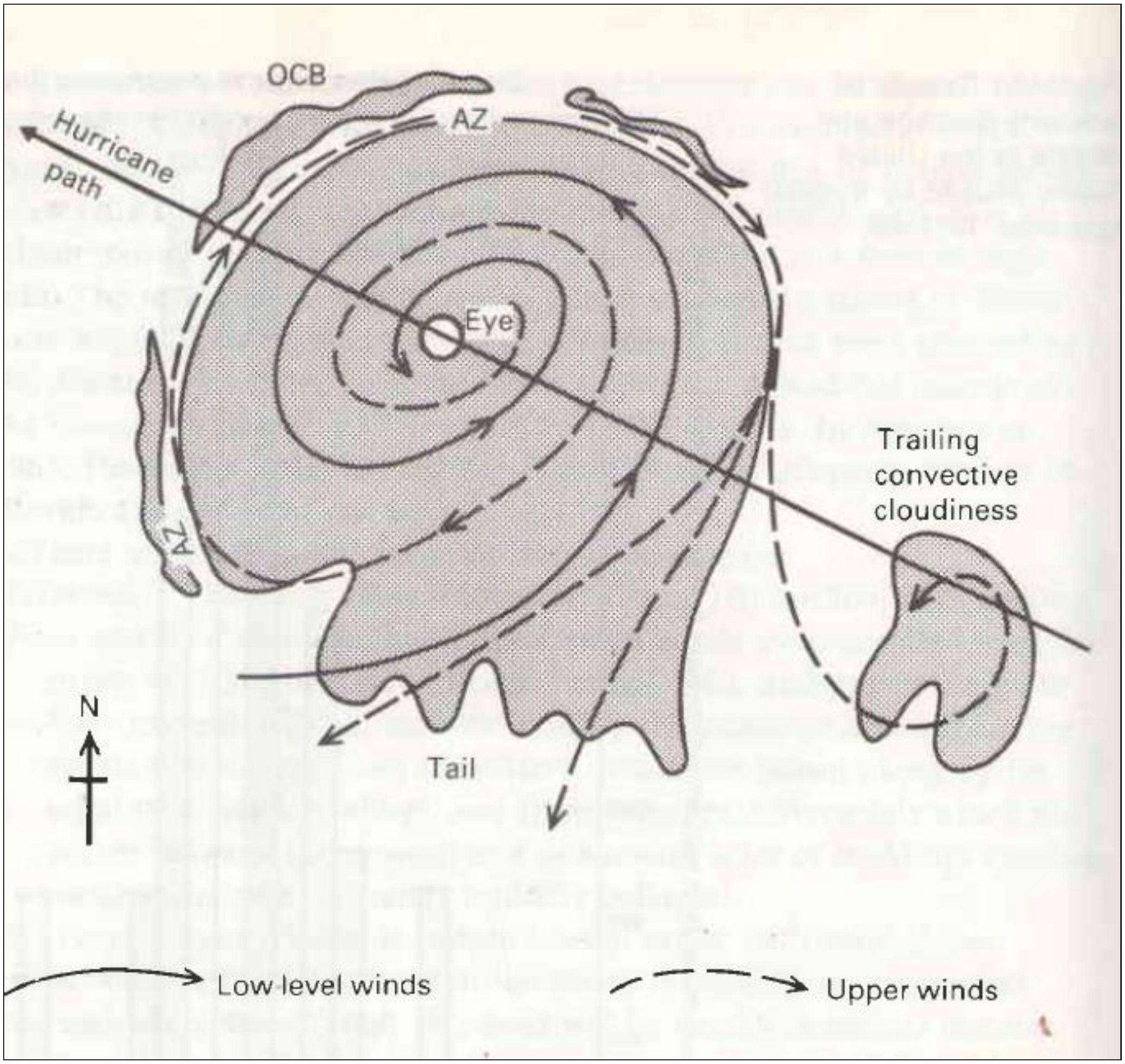
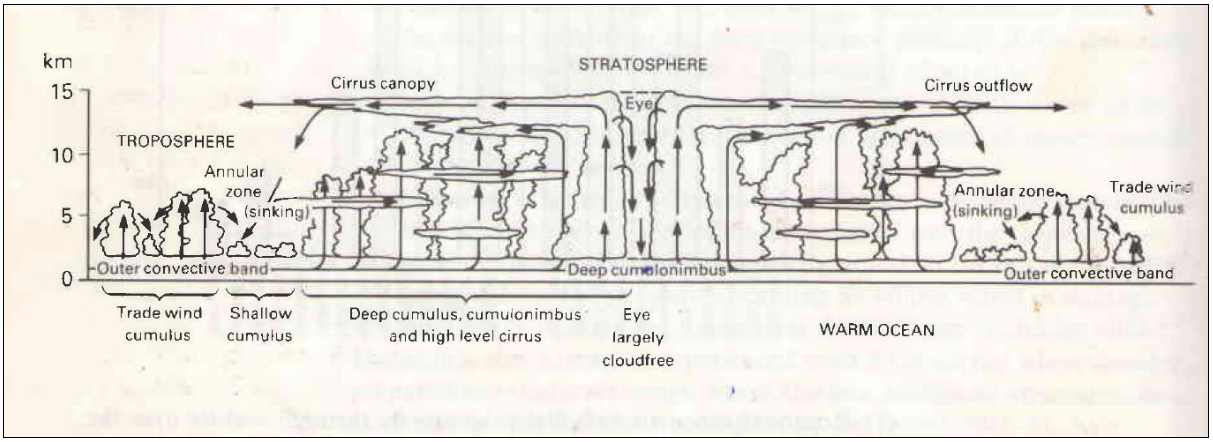



Fig 9.5 Plan view of a mature tropical cyclone, showing the wind directions and main cloud features. (OCB: Outer Convective Band; AZ: Annular Zone) (After Musk, 1988).




A Vertical Cross-section through a mature tropical cyclone (After Musk, 1988).

There are four main causes of hurricane damage:

- Wind
- Storm surges
- Flooding
- Landslide

# Tornadoes

Tornadoes are violent, rotating columns of air, diameter 50-500m, extending down like a dark funnel cloud from the base of a cumulonimbus cloud. They have a narrow vortex, winds reaching speeds greater than 400km/h and a deep low pressure core (100mb). Tornadoes are short-lived, rarely surviving more than 30 minutes. They form in the powerful up draughts of mature thunderstorms. They are usually unpredictable. They form whenever warm, fairly humid air is overlain by colder, drier air creating unstable and severe thunderstorms. They are common in spring and summer where maritime polar air mass lifts warm, moist tropical air on a cold front.

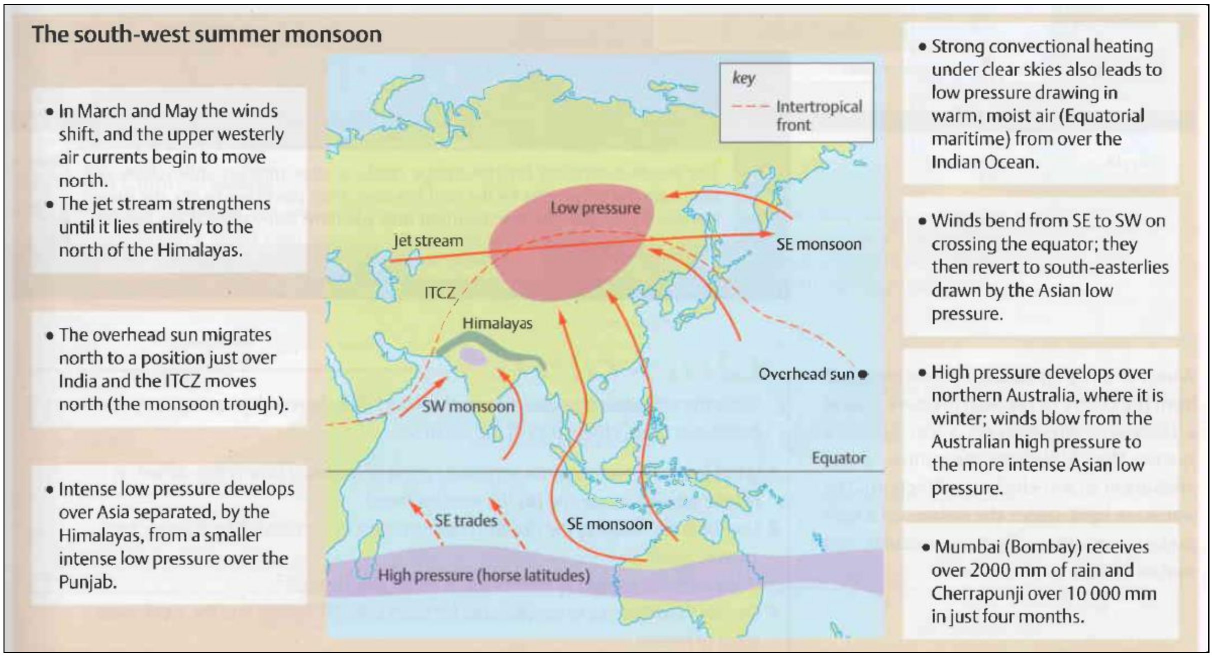
# The Monsoon

The word monsoon is an Arabic term for ‘a season’. In meteorology this term is used to denote a seasonal reversal of wind direction (Buckle, 1996). This phenomenon is common in south-east Asia where it results from three factors:

- The extreme heating and cooling of large land masses in relation to the smaller heat changes over the adjacent sea areas. This in turn affects atmospheric pressure and wind direction.
- The northward movement of the ITCZ during the northern hemisphere summer.
- The uplift of the Himalayas which are sufficiently high to interfere with general circulation of the atmosphere (Waugh, 2009).

## The South-west or Summer Monsoon

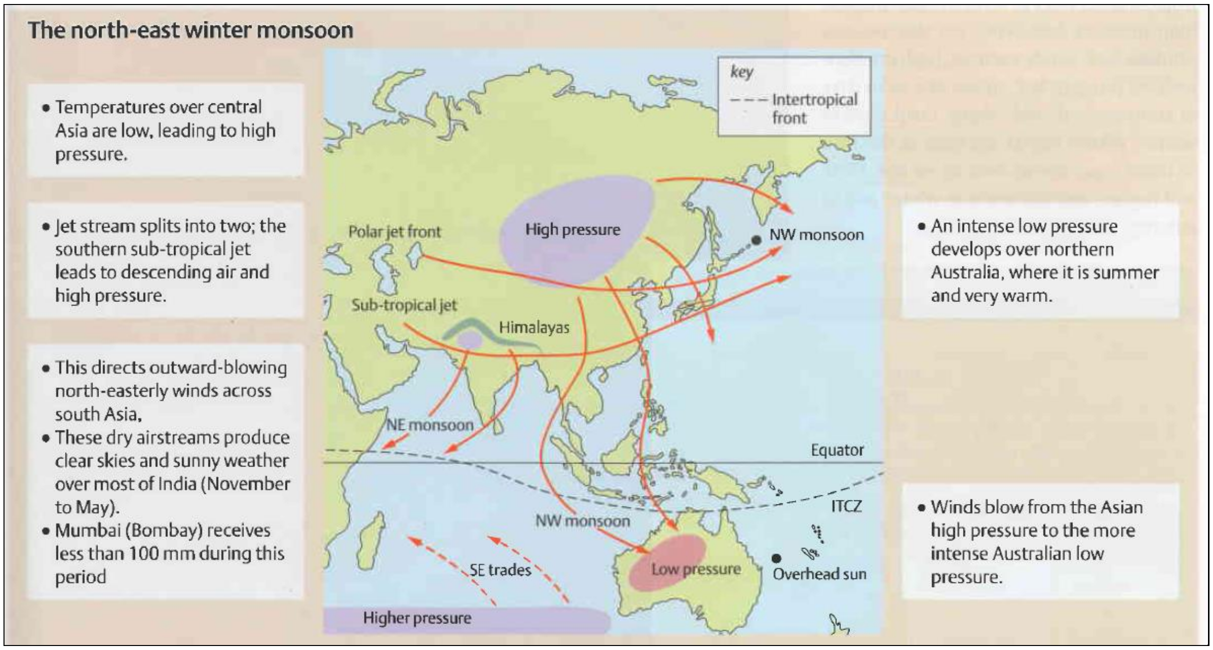
In June the sun is overhead the Tropic of Cancer and so the ITCZ moves northwards over the Asian subcontinent. The increase in insolation and temperature causes the air to rise and create a low pressure area. The warm moist equatorial maritime and the tropical maritime air from the Indian Ocean are drawn northwards and then it is diverted north-eastwards due to the Coriolis force. The air is humid, unstable and conducive to rainfall. Precipitation totals are accentuated as the air rises by both orographic and convectional uplift. In Bombay rainfall reach 2000mm.




South-west or Summer Monsoon (After Nagle, 2000)

# The North-east or Winter Monsoon

In December, the overhead sun, the ITCZ and the subtropical jet stream all move southwards. During this time Central Asia experiences intense cooling which allows a large high pressure system to develop. Airstreams that move outwards from this high pressure area are dry because their source area is semi-desert. They become even drier as they cross the Himalayas and adiabatically warmer as they descend to the Indo-Gangetic plain (Waugh, 2009). During this period Bombay receives less than 100mm of rainfall.




North-east or Winter Monsoon (After Nagle, 2000)

## Summary

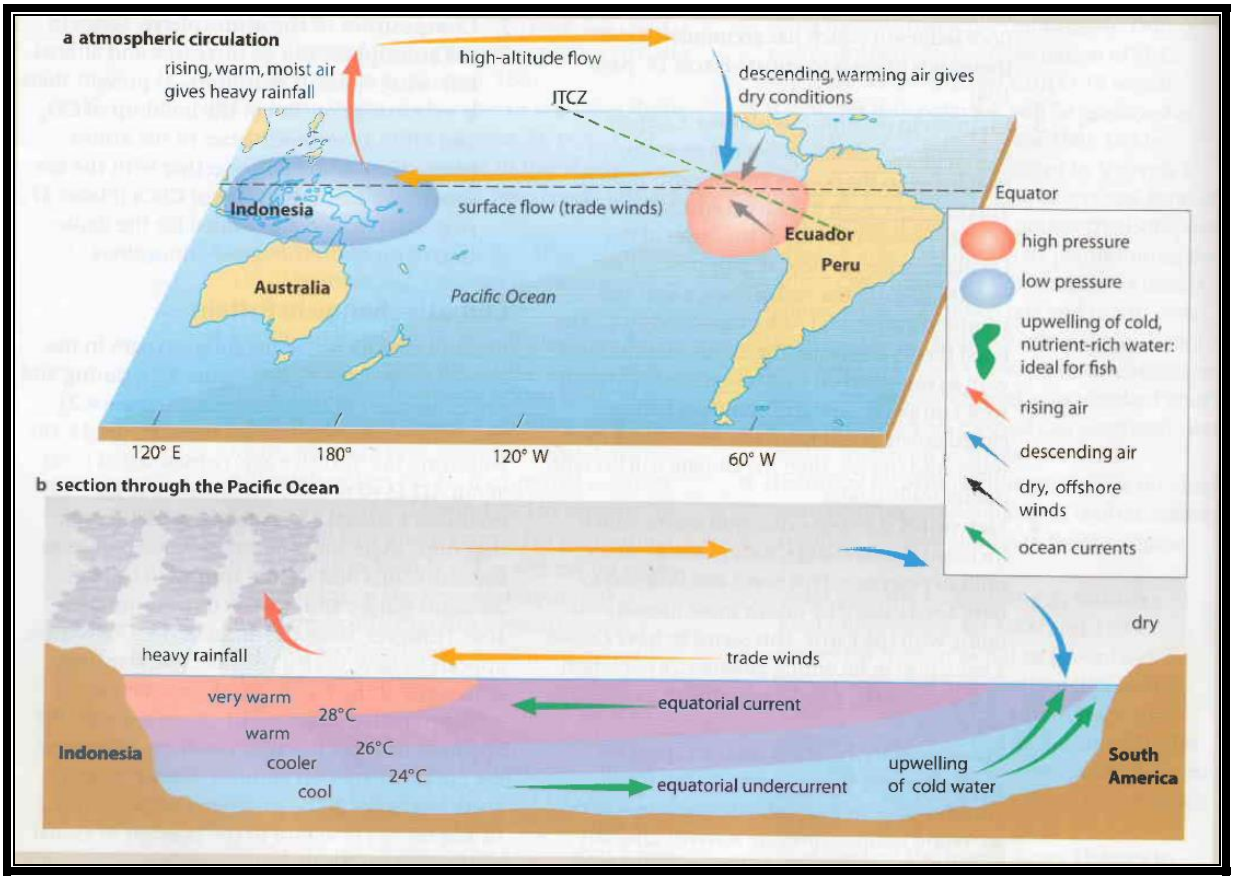
We have shown that air masses are classified according to the physical properties of air, the latitude of origin and the surface over which the air mass develops. An anticyclone tends to be of types: warm and cold, depending on whether its origin is thermal or dynamic. Depressions follow a life cycle in which there main stages can be identified: embryo, maturity and decay. The monsoon phenomenon is common in south-east Asia where it results from three factors:

- The extreme heating and cooling of large land masses in relation to the smaller heat changes over the adjacent sea areas.
- The northward movement of the ITCZ during the northern hemisphere summer.
- The uplift of the Himalayas which are sufficiently high to interfere with general circulation of the atmosphere.

# ***EL NINO SOUTHERN OSCILLATION***

Climate is constantly changing at all scales, from local to global and on varying timespans, both long term and short term (Waugh, 2009). The El Nino and la Nina events are the most interesting examples of the ocean-atmosphere interaction. These occur periodically in the Pacific Ocean. Waugh (2009:250) argues that under normal atmospheric conditions:

- Pressure rises over the eastern Pacific Ocean (off the coast of South America) and fall over the western Pacific Ocean (towards Indonesia and the Philippines).
- The descending air over the eastern Pacific gives clear, dry conditions that create the Atacama Desert in Peru.
- The warm, moist ascending air over the western Pacific gives that region its heavy convectional rainfall. This movement of air creates a circulation cell called the Walker Circulation (Fig).
- The upper air moves from west to east, and the surface air from east to west
- The trades winds than push surface water westwards so that the sea level in the Philippines is normally 60 cm higher than in Panama and Colombia
- The trades also allow flowing water westward as the equatorial current, to remain near to the surface where it can gradually heat. This gives the western Pacific the highest ocean temperature of the world, usually above 28 $^{0}$C.
- The warm air that is pushed away from South America is replaced by an upwelling of cold water, rich in plankton. (See fig 10.1 below)



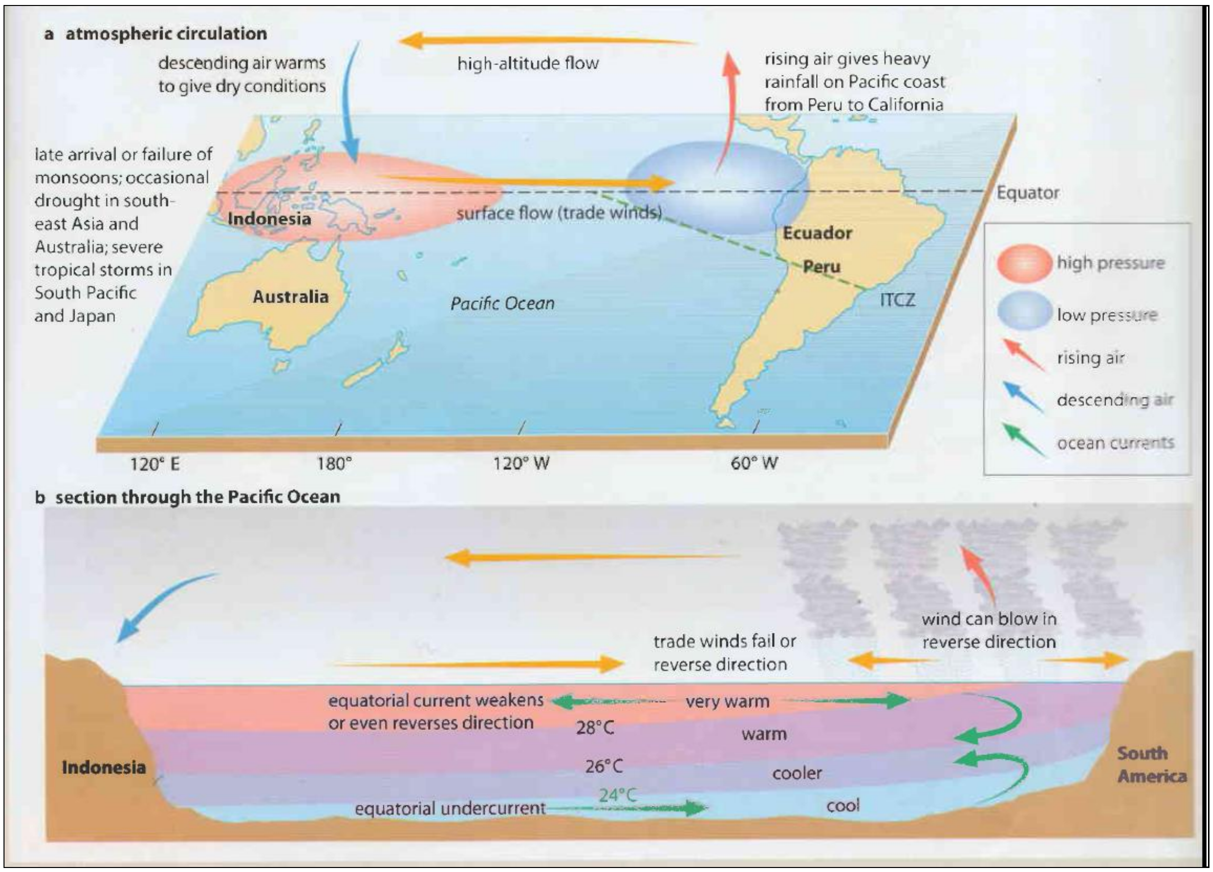

**Walker Circulation Cell (After Waugh, 2009**)

In contrast to normal conditions there is reversal in the equatorial Pacific region, in pressure, precipitation, wind and ocean currents.

- Pressure rises over the western Pacific and fall over the eastern Pacific
- The ITCZ is allowed to migrate southwards and causes the trades to weaken or to be reversed in direction.
- The descending now over south-east Asia, gives that region much drier conditions, even causing drought.
- In contrast the air over eastern Pacific is now rising giving much wetter conditions in places like Peru that normally experience desert conditions.
- The change in direction of the trade winds means that surface water tends to be pushed eastwards so that sea level in south-east Asia falls, while it rises in tropical South America.

- The reversal of the trade wind direction also means that surface water temperatures in excess of 28 $^{O}$C extend much further eastwards
- The upwelling of cold water off South America is reduced allowing sea temperatures to rise by up to 6 $^{O}$C.

Figure 10.2 below illustrate the ocean-atmospheric interaction under the El Nino event.




**El Nino event (After Waugh, 2009**)

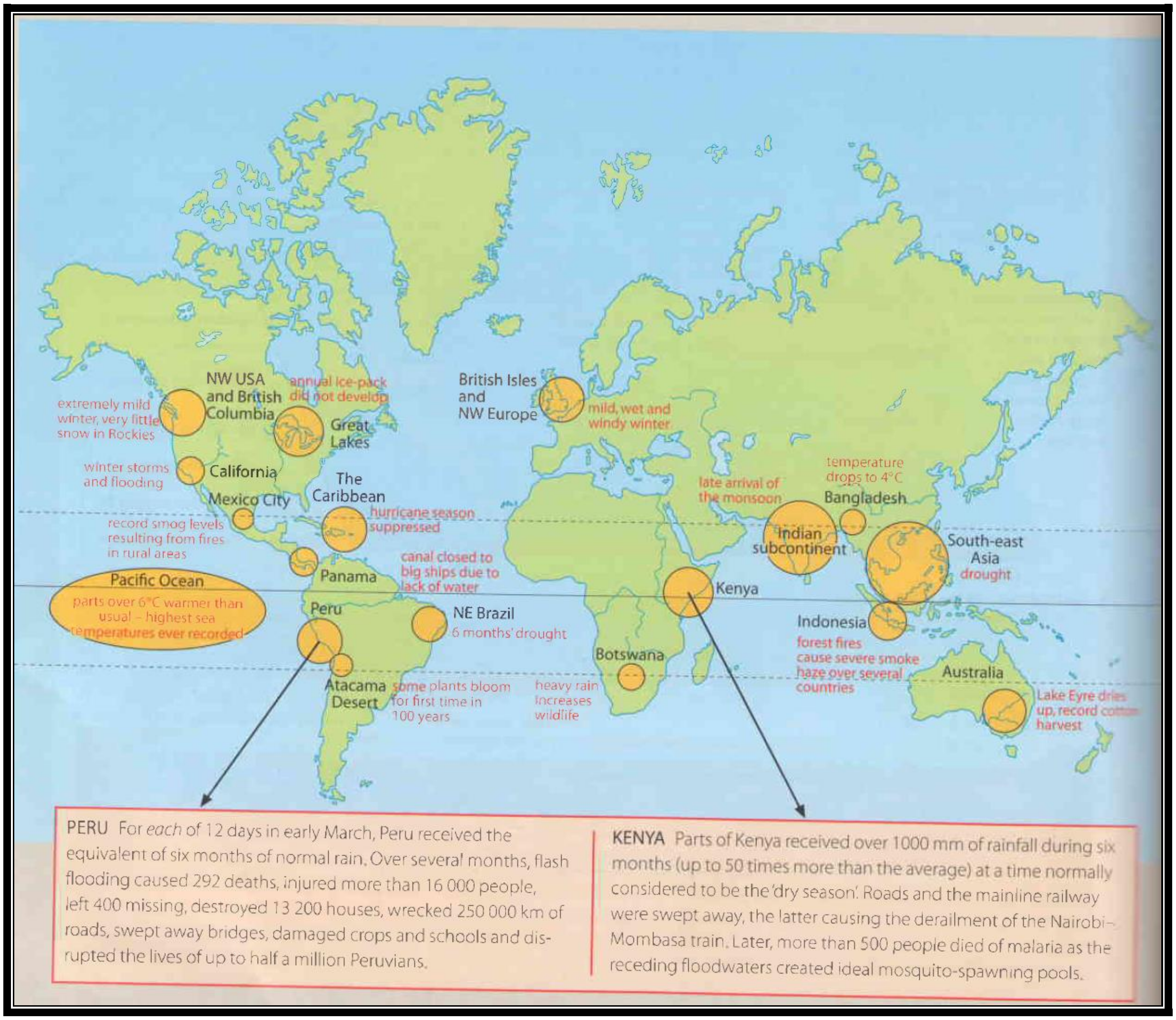
To date the world has experienced the El Nino events in 1982-83, 1986, 1992-93, 1997-98 and 2006-07. The 1997-98 event has been described as the biggest of all (Waugh, 2009). There is much evidence which suggest that the ENSO has had a major climatic effect on places far beyond the Pacific margins. Apart from direr conditions in south-east Asia and wetter conditions in South America:

- Severe droughts have been experienced in the Sahel region, Southern Africa (Zimbabwe included) and across the Indian subcontinent.

- Extremely cold winters were recorded in North America.
- Stormy conditions with floods have been recorded in California.
- Exceptionally wet, mild and windy winters were experienced in Britain and north-west Europe.
- The average global temperatures are high during the El Nino events.
- Rainfall intensities across the globe are increased when temperatures are high due to the effects of increased convective activity.
- A range of hazards have been linked to the El Nino events: floods, famine, drought, wildfires, landslides and disease out-break.

## The 1997/98 El Nino Event

The effects of the 1997-98 El Nino event are shown in the map below:

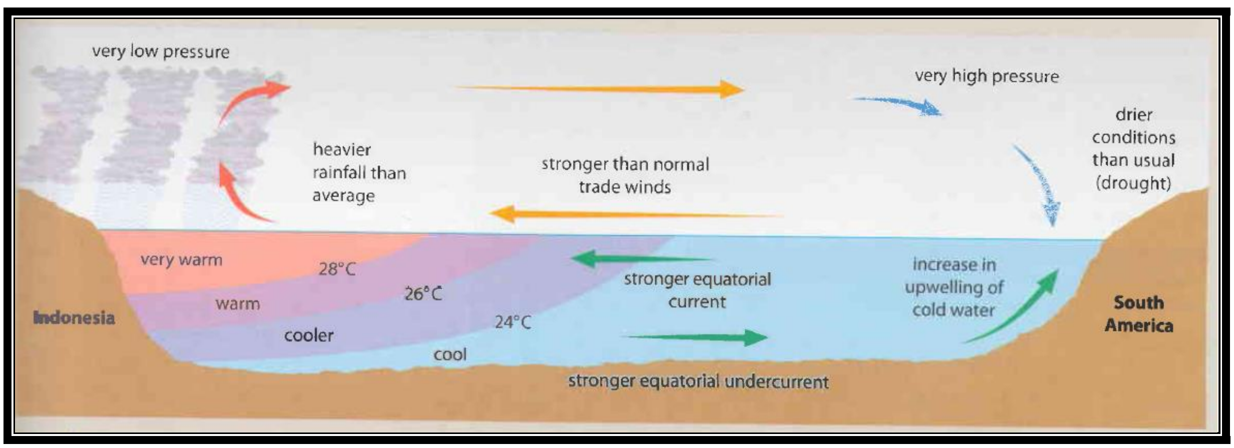



Effects of the 1997/98 El Nino event (After Waugh, 2009

# *La Nina*

La Nina (Little Girl) has climatic conditions that are the reverse of those of El Nino. It appears just before or after El Nino event. In a La Nina event in contrast to the normal conditions in the Pacific Ocean:

- Low pressure over the western Pacific becomes even lower.
- The high pressure over the eastern Pacific becomes even higher.
- Rainfall increases over south-east Asia (e.g. la Nina event of 1988 was responsible for severe flooding in Bangladesh).
- There are severe drought conditions in South America.
- The trade winds strengthen due to increased differences in pressure between the two places.
- The stronger trades push large amounts of water westwards, giving a higher than normal sea-level in Indonesia and the Philippines.
- The stronger trade wind increases the equatorial undercurrent and significantly enhances the upwelling of cold water off the Peruvian coast.
- Increases hurricane activities.
- Can disrupt the jet streams to bring stormy, colder weather to Britain (e.g. The Boxing day Storm of 1998).




La Nina Event (After Waugh, 2009)

# ***Possible effects of a la Nina on World Weather***

- Higher temperature, storms and flooding in South East Asia, Indonesia, Philippines and north of Australia
- Flooding in south-east Africa (Mozambique, Zimbabwe, Swaziland and Lesotho).

- Flooding in north-east Brazil
- Drought in western coast of South America
- Increased hurricane activity in the Caribbean and Florida
- Drought and warmer conditions in western coast of North America
- Increased storm and snow activity in the British Isles.
- Eight degrees Celsius drop in ocean surface temperatures over 5000km stretch in the eastern Pacific Ocean.