CHAPTER 1

Introduction

1.1. Physiographic division of Nepal

Nepal lies in the sub-tropical region between latitudes 26^0 22' to 30^0 27' N and longitudes 80^0 04' to 88^0 12' E. It is sandwiched between India and China. India borders Nepal on east, west and south and China borders on the north. It covers 0.03% of the world and 0.3% of the total of Asia. The total area of the country is 1,47,181 sq. km extending roughly 885 km from east to west in length and 145 km to 241 km from north to south. The altitudinal variation is from about 60 m above mean sea level in the southern plain (called Terai) to the Mount Everest (8848 m) in the north east.

Generally, Nepal is divided into five major physiographic zones.

i. Terai

It elevates from 60 m to 200 m. It is relatively low and flat region in the south. This region is characterized by heavy jungle in the northern part known as ‘Charkose Jhadi’. Most of the lands in this region are used for agricultural production which fulfills most of the country's demand.

ii. Churiya (Siwalik Range)

This region elevates from 200 m to 1500 m. It is small hilly region after terai towards north. In this region, forests are less dense and slopes are highly eroded.

iii. Middle Mountain (Mahabharat Range)

This region elevates from 1500 m to 2500 m. It is characterized by moderately High Mountain with rugged terrain. This region contains densely populated area with valley like Kathmandu, Pokhara. This region is also favorable for agriculture.

iv. High Mountain

This region elevates from 2500 m to 4000 m. This region is characterized by High Mountain larger than Middle Mountain with steep slopes. Sometimes snow falls at the top of high mountains.

v. High Himalaya

It elevates from 4000 m to 8848 m. This region is characterized by series of several mountains chain including world's famous highest peak Mt. Everest. In the lower parts of this region, semi-desert like condition prevails and in the upper parts, snow covers almost all the times.

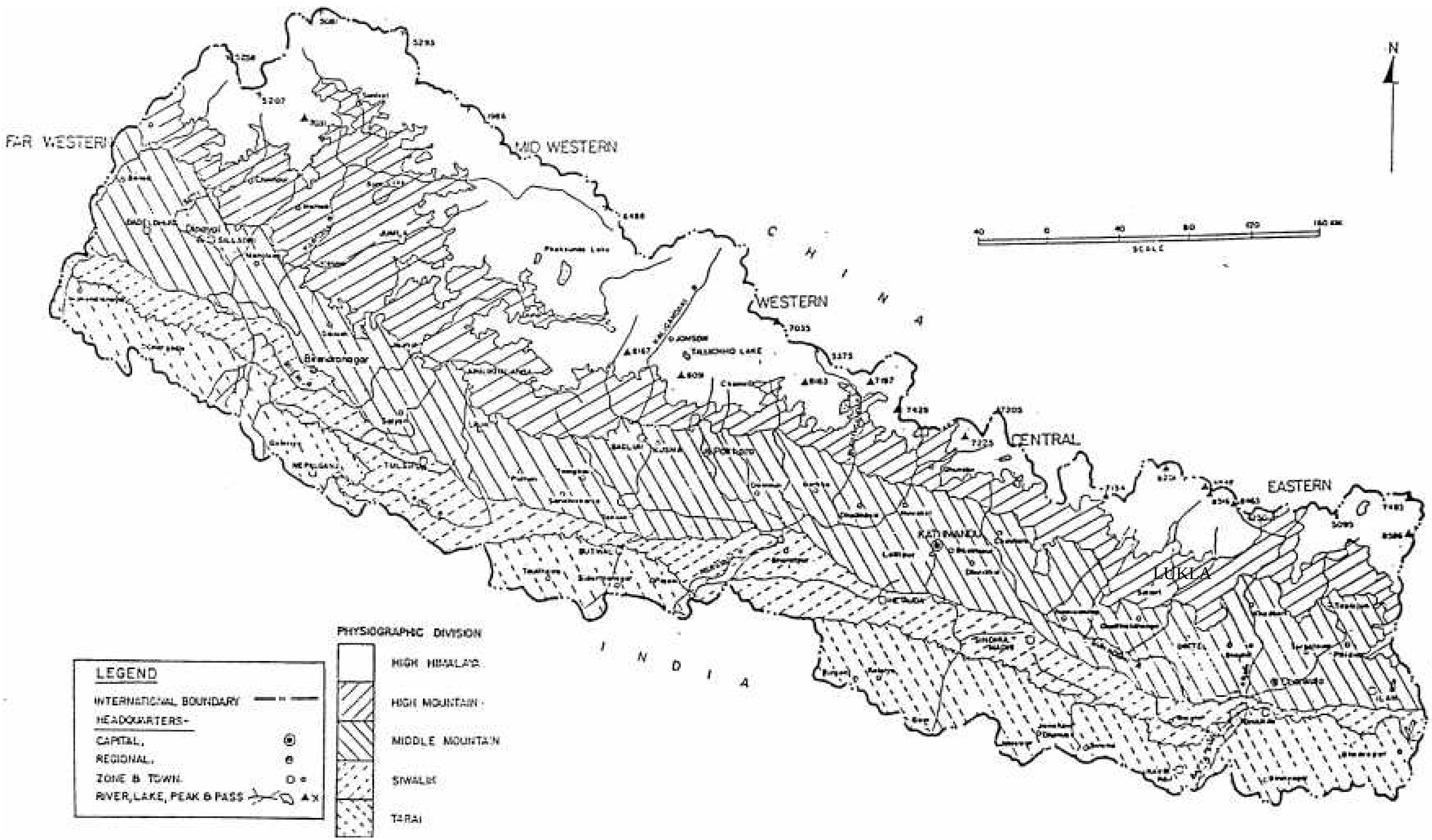

Fig. 1.1: Physiographic Division of Nepal

1.2. General Climatic Feature

Physiographic feature, altitude geographic location controls the whole Nepal in climatic aspect. Generally five different types of climate have been identified in Nepal.

i. Tropical

This type of climate is found in Terai and Siwalik range. In this type of climate summer is hot and humid whereas winter is mild and dry.

ii. Warm-temperate

This type of climate is found in Mahabharat range. It is characterized by warm summer and cool winter. Snow falls occasionally in higher altitude of the hills.

iii. Cool-temperate

This type of climate is found in Mid-mountain region. It is characterized by cool summer and cold winter. Winter precipitation is in the form of snow at high altitude.

iv. Alpine

This type of climate is found in lower Himalayas. It is characterized by cool summer and frosty winter. Snow covers most of the time.

v. Tundra

This type of climate is found in high Himalayas. It is characterized by permanent cover of snow and ice. Precipitation is in the form of snow only.

1.3. General Seasons in Nepal

Nepal is strongly influenced by the monsoon circulation. On the basis of monsoon phenomena, there are four seasons in Nepal.

i. Pre-monsoon season (Hot weather season)

It is the period from March to May. This season is characterized by hot and dry weather with almost cloudless sky. In this season, due to the highest temperature, pre-monsoon thunderstorm activity with some rain, violent dust storms activity known as 'Chaite Huri' could be experience.

ii. Summer monsoon season (Rainy Season)

The period from June to September is considered as summer monsoon season. Summer monsoon establishes with cloud and rain activities, domination of easterly wind flow, lowering in temperature than pre-monsoon season, precipitation amount is more than 80% of annual total. In case of failure of monsoon, more intense drought or drought like condition prevails.

iii. Post-monsoon Season (Mild Weather Season)

It is the period from October to November. Domination of westerly wind flow with falling in temperature gradually as an indication of pre-winter activity, negligible amount of rainfall occurs in this period.

iv. Winter Season (Cold Weather Season)

It is the period from December to February. This season is characterized by dry, cool weather, lowest temperature and less amount of rainfall (but greater than post monsoon season). Above 4000 m, precipitation falls as snow. Some areas also get frosty night.

1.4. Four factors that determine climate

Climate differs from one location to another because of differences in

- latitude, the angular distance north or south from the equator
- altitude, the height above sea level
- continentality, the distance from the sea
- exposure to regional circulations, including winds and ocean currents

1.4.1. Latitude

The latitude of a given site determines the length of the day and the angle of incoming sunlight and therefore the amount of solar radiation received at that site. Seasonal and diurnal (day-night) variations in the amount of solar radiation received cause seasonal and diurnal variations in the weather. Near the equator, the days of the year are all about the same length, and the noon sun is nearly overhead year-round. Because day length and solar angle change little with the season, there is little seasonal variability in the weather.

In the polar regions, on the other hand, the sun does not rise at all in the winter, and in the summer it never sets, although it remains low in the sky. Thus, polar weather has a high seasonal variability, but a low diurnal variability. In the mid-latitudes, the climate is characterized by both seasonal and diurnal changes. Except at the equator, day length varies throughout the year. In the Northern Hemisphere, the longest day of the year is at the summer solstice (June 21), the shortest day of the year is at the winter solstice (December 21), and the day is 12 hours long on the vernal and autumnal equinoxes (March 20 and September 22). The altitude angle of the sun also varies throughout the year, with an increase of about 47° from winter to summer. The more direct summer sunlight produces more heating than the slanted rays of the winter sun.

The latitude of a given site affects its climate not only because it determines the angle of solar radiation and the length of a day, but also because it determines the site's exposure to latitudinal belts of surface high and low pressure that encircle the earth. High pressure belts (i.e., zones where high pressure centers are often found) are associated with sinking motions, or subsidence, in the atmosphere, clear skies, dry air, and light winds. Low pressure belts are associated with rising motions, or convection, in the atmosphere, cloudiness, precipitation, and strong winds. Belts of low pressure occur in the equatorial (0-20° latitude) and sub-polar (40-70°) regions and alternate with belts of high pressure that form in the subtropical (20-40°) and polar (70-90°) regions.

1.4.2. Altitude

Temperature, atmospheric moisture, precipitation, winds, incoming solar radiation, and air density all vary with altitude. Up to an altitude of 7 miles (11 km), temperature generally decreases with altitude. The rate of decrease is typically 3.5^0F per 1000 ft or 6.5^0C/km. Thus, locations at high elevations generally have a cooler climate than locations at lower elevations.

Incoming solar radiation increases with altitude. As radiation passes through the earth's atmosphere, a small fraction is absorbed by the atmosphere, resulting in a minor increase in the air temperature. Another small fraction is scattered by atmospheric constituents and redirected into space. Because solar radiation reaches higher elevation land surfaces before lower elevation land surfaces, there is less depletion of the solar beam through absorption and scattering at higher elevations than at lower elevations.

Although more incoming solar radiation reaches the ground at higher elevations, the effect on air temperature is minimal, and changes in air temperature from day to night on exposed mountainsides and peaks are smaller than the diurnal changes at lower altitudes. This is explained by the way the earth's atmosphere is heated and cooled and by the decrease of land surface area with elevation. Most of the radiation received from the sun does not heat the earth's atmosphere directly but rather passes largely unimpeded through the atmosphere, is received at the earth's surface, and heats the ground. The ground, in turn, heats the atmosphere from below. Because there is less land surface area at higher elevations, less heat is transferred to the atmosphere during the day.

At night, radiation loss from the ground cools the earth's surface, which then cools the air above it. Air near the surface, for example, on mountain slopes, thus cools more than air at the same elevation in the free atmosphere. Air that cools over mountain slopes flows down the slopes and collects in valleys and basins, resulting in the formation of temperature inversions, layers in which temperature increases with height instead of decreasing. Temperature inversions are commonly seen both in mountainous areas and over flat terrain at night or at all times of day in winter; they are an important exception to the rule that temperature decreases with altitude.

The temperature at a given site within a mountain massif depends not only on the site's altitude, but also on its exposure to incoming solar radiation. South-facing slopes receive more solar radiation than north-facing slopes, for example. Temperature differences within a mountain massif are important because they drive local winds.

Atmospheric moisture also generally decreases with altitude. The earth's surface supplies the atmosphere not only with heat, but also with water vapor through evaporation of water (the oceans are the primary source of atmospheric moisture) and transpiration from plants. As altitude increases, the distance from the source of moisture increases, and, therefore, the amount of moisture in the atmosphere decreases.

Although moisture decreases with altitude, precipitation usually increases. The cooler air at higher altitudes can hold less moisture than the warmer air at sea level. Thus, when warm, moist air from sea level is lifted over a mountain range, the air cools, its capacity to hold moisture decreases, and much of the moisture is released as precipitation.

Wind speeds generally increase with altitude. Wind speeds are lowest at the earth's surface because of friction. Winds measured on mountain peaks tend to be stronger than winds at lower elevations because the peaks extend high into the atmosphere where wind speeds are higher. The limited surface area of the peaks themselves produces little friction to slow the winds. Winds that are carried over mountains or through mountain passes may even speed up because of the influence of the complex terrain. However, valleys, basins, and lee slopes within a mountain area are often sheltered from the generally stronger winds at high altitudes by the surrounding topography.

Air density, the mass of a unit volume of air, decreases exponentially with height. The same amount of heat input results in a greater change in temperature at higher elevations than at lower elevations. The less dense mountain atmosphere also responds more quickly to the input of heat than the denser atmosphere at lower elevations. This quick response to heat input combined with the rapid transport of warm and cold air, clouds, and storms by the strong winds at high elevations contributes to the perception of the high "changeability" of mountain weather. Lower air density also affects the perception of wind velocity. The same wind velocity is perceived as somewhat weaker at higher elevations than at lower elevations because it is transporting less mass and thus less momentum.

1.4.3. Continentality

Locations at the center of a continent experience larger diurnal and seasonal temperature changes than locations on or near large bodies of water because land surfaces heat and cool more quickly than oceans. Interior locations also experience more sunshine, less cloudiness, less moisture, and less precipitation than coastal areas, where the maritime influence produces more cloudiness and precipitation and moderates temperatures. Precipitation is especially heavy on the windward side of coastal mountain ranges oriented perpendicular to prevailing winds from the ocean. As mentioned previously, marine air that is lifted up a mountain range releases much of its moisture as precipitation. As a result, far less precipitation is received on the leeward side of a mountain range.

1.4.4. Regional Circulations

Although latitude, altitude, and continentality are the primary determinants of climate in a mountainous region, exposure to regional winds and ocean currents is also a factor. Regional winds are associated with the semi-permanent atmospheric high and low pressure systems that form in different latitude belts and directly affect climate. In the Northern Hemisphere, winds blow clockwise around high pressure centers and counterclockwise around low pressure centers.

1.5. Diurnal Mountain Winds

Diurnal mountain winds develop over complex topography of all scales, from small hills to large mountain massifs and are characterized by a reversal of wind direction twice per day. As a rule, winds flow upslope, upvalley, and from the plain to the mountain massif during daytime. During nighttime, they flow downslope, down-valley, and from the mountain massif to the plain. Diurnal Mountain winds are strongest when skies are clear and winds aloft are weak.

Diurnal mountain winds are produced by horizontal temperature differences that develop daily in complex terrain. The resulting horizontal pressure differences cause winds near the surface of the earth to blow from areas with lower temperatures and higher pressures toward areas with higher temperatures and lower pressures. The circulations are closed by return, or compensatory, flows higher in the atmosphere.

Four wind systems comprise the mountain wind system, which carries air into a mountain massif at low levels during daytime and out of a mountain massif during nighttime.

- The slope wind system (upslope winds and downslope winds*)* is driven by horizontal temperature contrasts between the air over the valley sidewalls and the air over the center of the valley.
- The along-valley wind system (up-valley winds and down-valley winds) is driven by horizontal temperature contrasts along a valley's axis or between the air in a valley and the air over the adjacent plain.

- The cross-valley wind system results from horizontal temperature differences between the air over one valley sidewall and the air over the opposing sidewall, producing winds that blow perpendicular to the valley axis and toward the more strongly heated sidewall.
- The mountain-plain wind system results from horizontal temperature differences between the air over a mountain massif and the air over the surrounding plains, producing large-scale winds that blow up or down the outer slopes of a mountain massif. The mountain-plain circulation and its upper level return flow are not confined by the topography but are carried over deep layers of the atmosphere above the mountain slopes.

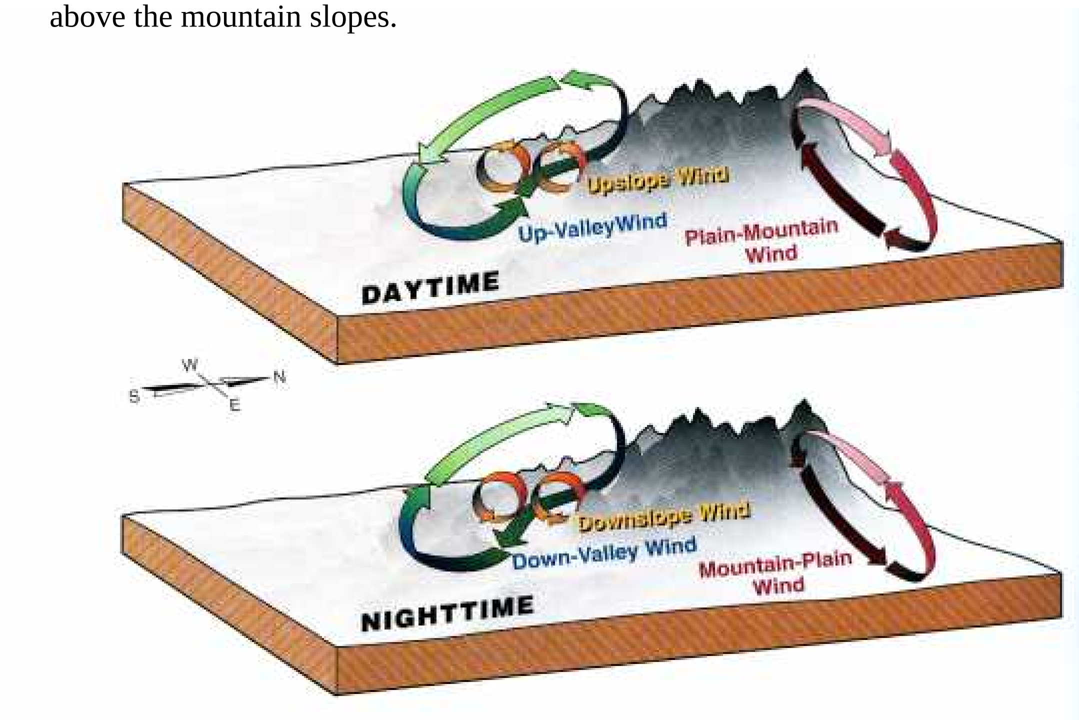

Fig. 1.2: Wind systems over mountainous terrain

1.5.1. The daily cycle of slope and along-valley winds and temperature structure

Because diurnal mountain winds are driven by horizontal temperature differences, the regular evolution of the winds in a given valley is closely tied to the thermal structure of the atmospheric boundary layer within the valley, which is characterized by a diurnal cycle of buildup and breakdown of a temperature inversion. The relationship between winds and temperature structure can best be described by considering the typical diurnal cycle in a representative valley. Figure 1.3 shows the four distinct phases of wind and temperature structure evolution in a valley during the course of a day.

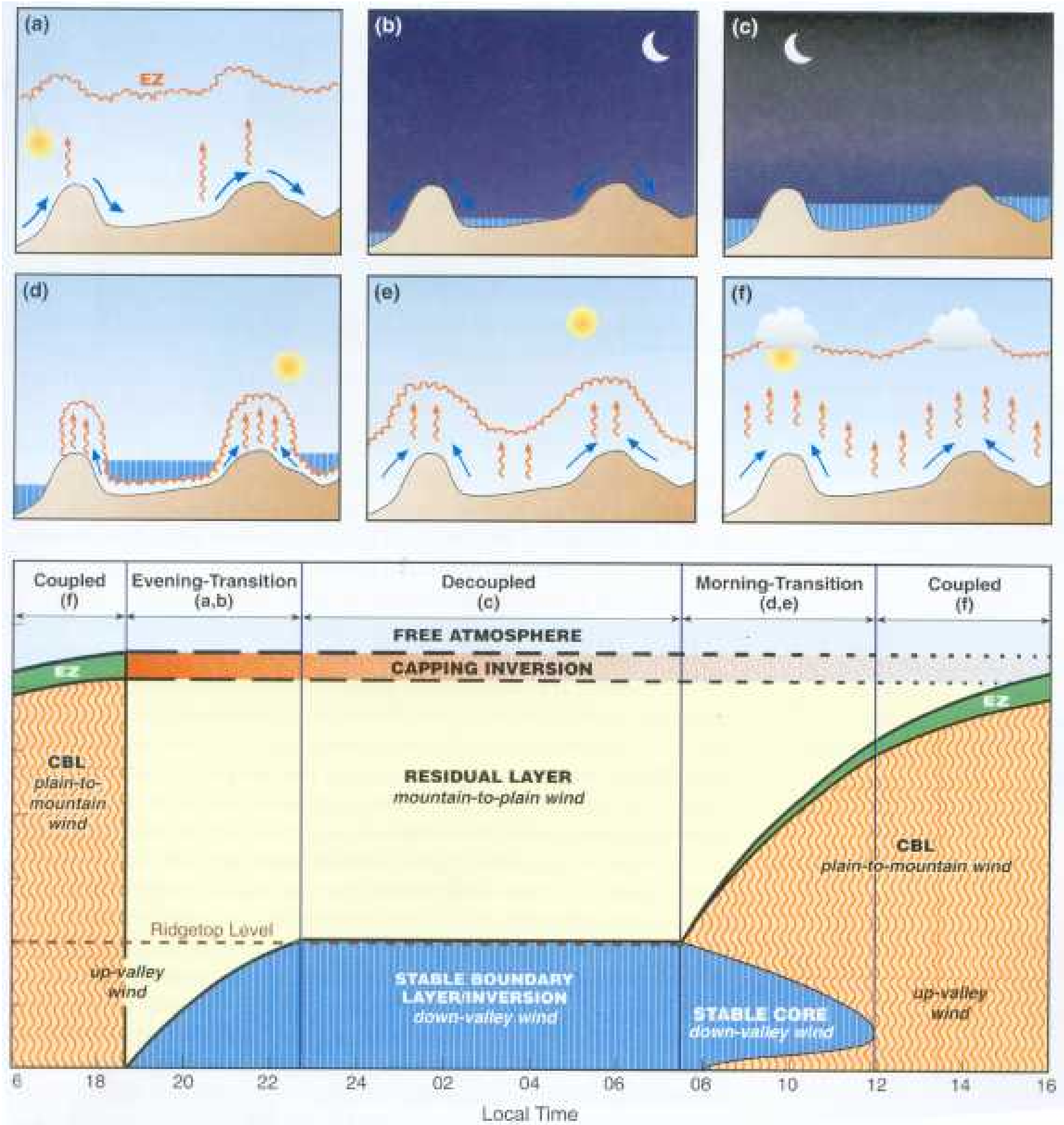

Fig. 1.3: Temperature and wind structure evolution over a valley cross section during the course of a day.

1.5.1.1. Evening Transition Period

The evening-transition phase begins when slope winds reverse from upslope to downslope. The downslope winds drain cold air off the sidewalls into the valley, resulting in the build-up of an inversion and causing along-valley winds to begin to reverse direction from up valley to down-valley. This transition phase ends when down-valley flows prevail through the depth of the valley.

1.5.1.2. Decoupled Period (Night Time)

During the nighttime phase, the valley atmosphere is decoupled from the atmosphere above ridge-top level. Downslope winds blow down the sidewalls, and down-valley winds blow within the inversion layer.

1.5.1.3. Morning Transition Period

The morning-transition phase begins with the reversal of slope winds from downslope to upslope. Convective currents rising from the ground destroy the inversion from below, and along-valley winds reverse from down-valley to up-valley. This transition phase ends when the inversion is destroyed and winds blow up-valley through the depth of the valley.

1.5.1.4. Coupled Period (Daytime Period)

During the daytime phase, the valley atmosphere is coupled with the atmosphere above ridge-top level. Upslope and up-valley winds prevail in an unstable convective boundary layer that extends from the valley floor and sidewalls into the above-valley atmosphere.

1.5.2. The four components of mountain wind system

1.5.2.1. Slope wind system

The slope wind system is a closed circulation driven by horizontal temperature contrasts between the air over a slope and the air at the same level over the center of the valley. The temperature contrasts result from the heating or cooling of an inclined boundary layer over the slope.

Slope flows are typically in the range of 1-5 m/s (2-11 mph). Although the highest daytime temperatures and lowest nighttime temperatures are found at the ground, peak wind speeds occur a few meters above the surface because of frictional drag near the ground. Figure 1.4 illustrates the relationship between wind and temperature structures over a mid slope location during upslope and downslope flows.

Both upslope and downslope wind speeds generally increase with distance. Downslope flows on the long slope of the continental ice dome in Antarctica can reach gale force (28-47 knots) on the periphery of the continent. The strongest downslope flows occur around sunset when slopes first go into shadow. The strongest upslope flows occur at midmorning when the temperature contrast between the warm, sunlit slopes and the valley atmosphere is strongest.

The depth of slope winds also varies over time and space. Downslope flows are shallower than upslope flows. The depth of a downslope flow increases with distance down the slope and can be estimated as 5 % of the drop in elevation from the ridge top. Thus, at 100 m in elevation below the ridge top, a 5-m-deep downslope flow can be expected. Downslope flow depth tends to decrease during the night when an inversion builds up in the valley. Upslope flows increase in depth with distance up the slope and with time, usually attaining a depth of 50-150 m (150- 500 ft) during the first several hours after sunrise.

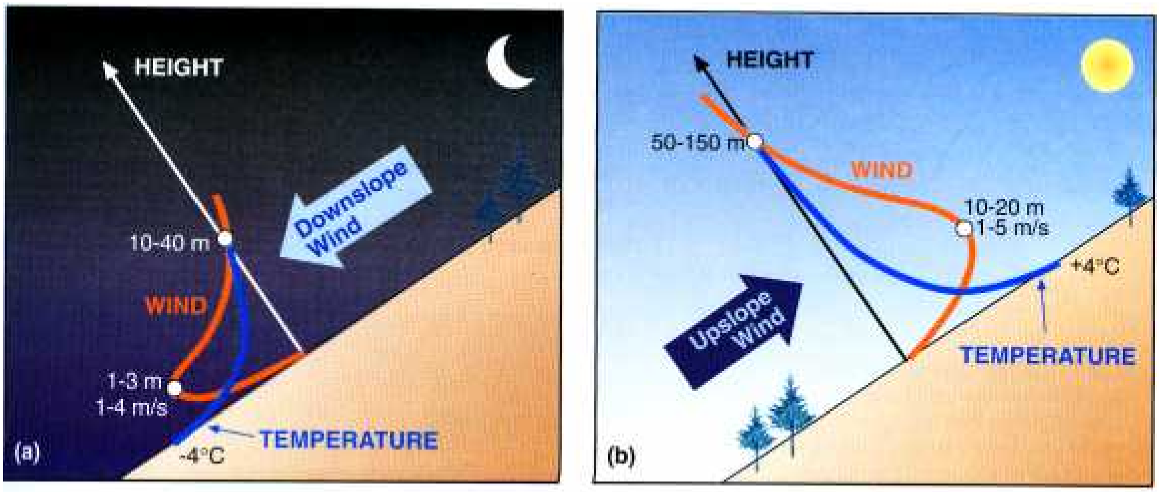

Fig. 1.4: Typical profile of wind and temperature as a function of height above a slope during (a) night time (b) day time

1.5.2.2. Along- valley wind system

Along-valley winds are the lower branch of a closed circulation that moderates horizontal air temperature differences that form within a valley or between a valley and a nearby plain by advecting relatively cold air into an area with relatively warm air. The lower branch of the circulation is illustrated in fig. 1.5, which shows two columns of air-one in the valley and one above the plain. The top of both columns is defined by the elevation of the valley ridge tops, and the bottom of both columns is defined by the elevation of the valley floor. Both columns have the same horizontal area at the top. During daytime, the air in the valley warms more rapidly than the air over the plain, reducing the pressure in the valley atmosphere relative to the plain and driving an up-valley wind from the plain into the valley. The air carried into the valley during daytime at low levels is returned to the plain by the upper branch of the circulation. During nighttime, the valley atmosphere cools faster than the plain atmosphere at the same level, producing higher pressure in the valley column and driving a down-valley flow from the valley to the plain. A return current forms above the down-valley flow and carries air back toward the upper end of the valley. The development of a closed circulation requires rising motions (and cooling) above the warm columns and sinking motions (and warming) above the cold columns. These vertical motions thus also play a role in equilibrating horizontal temperature differences within the valley.

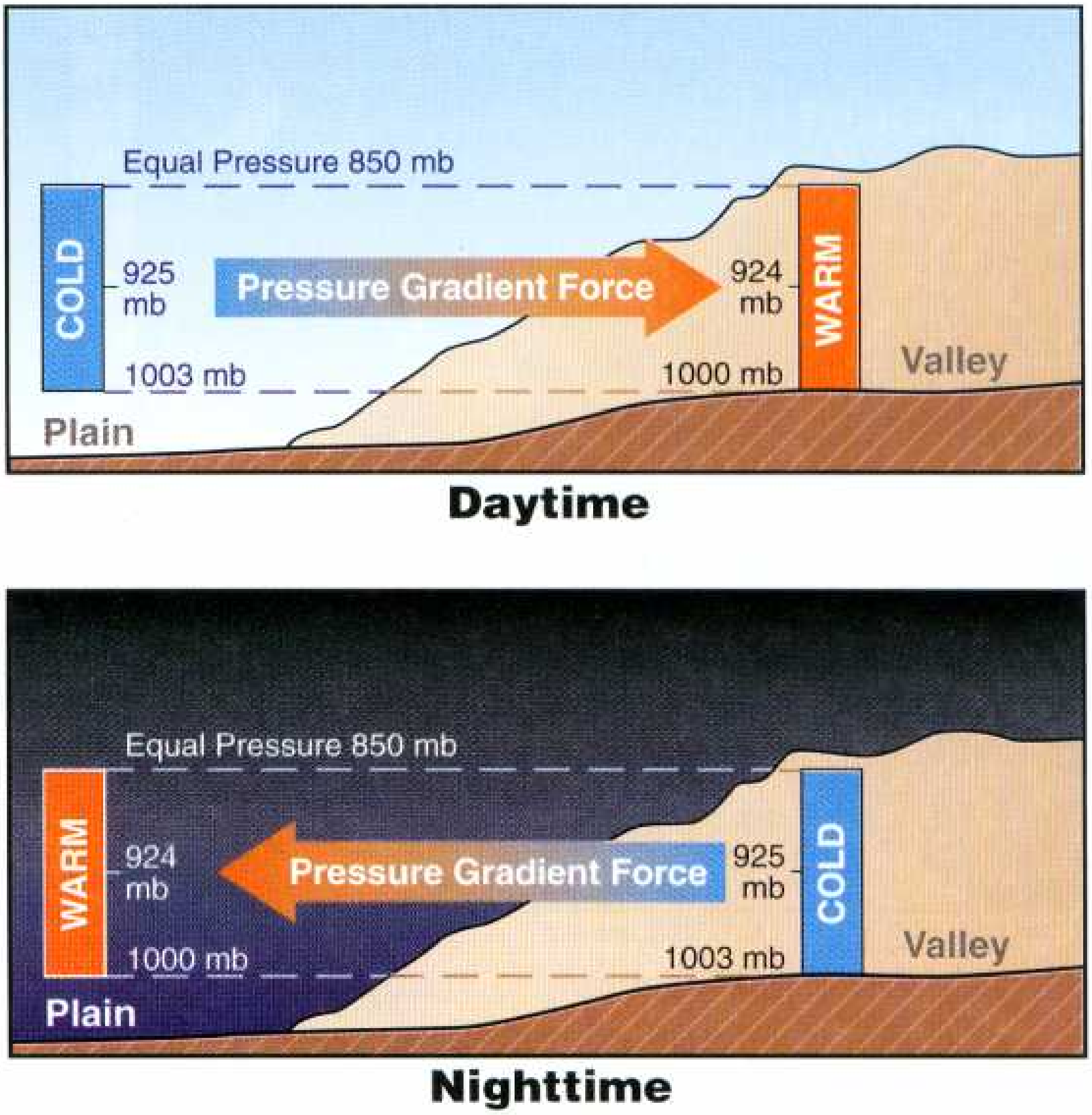

Fig. 1.5: Pressure difference developed between a valley and the adjacent plain

1.5.2.3. Cross-valley winds

Cross-valley winds blow across to the valley axis, when the air above one of the valley sidewalls becomes warmer than the air above the other sidewall. The near-surface flow is toward the more strongly heated sidewall. Compensatory flows in the opposite direction occur at higher altitudes within the valley.

Cross-valley winds are generally weak (2 m/s or less, or 4.5 mph or less). The strength of the flow depends directly on valley width and on the cross-valley temperature gradient, with weaker winds in very wide valleys and in valleys where the temperature differences between the opposing sidewalls are small.

Cross-valley winds are stronger during the day, when the sidewalls are heated unequally by the sun, than during the night, when the outgoing long wave radiation is more evenly distributed across a valley. The strongest cross-valley flows occur early or late in the day in north-south-oriented valleys when one sidewall is in shade while the other sidewall is in full sunlight.

1.5.2.4. Mountain-plain wind system

The mountain-plain wind system is a closed circulation that develops above the slope of a mountain massif in response to temperature differences between air over the mountains and air over the nearby plains. The temperature difference produces a horizontal pressure difference between the air over the plains and the air in the atmospheric boundary layer that forms over the entire mountainous region.

1.6. Cumulonimbus clouds (Cb)

Nepal experiences a series of thunderstorm during the pre-monsoon season (April-May). Cumulonimbus clouds (Cb) is necessary for the origination of such thunderstorm.

1.6.1. Formation

To create a cumulonimbus cloud, three ingredients are required.

1. Plenty of moisture.
2. A mass of warm unstable air.
3. A source of energy to lift the warm, moist air mass rapidly upward.

Typically, the clouds form around front lines, near oceans where sea breezes provide the storm energy, or over mountains which push the air upwards.

When the warm air rises above the typically cooler air above it, it starts to cool and the water vapour condenses into water droplets. This condensation heats the surrounding air by releasing latent heat, thus continuing the rise of air.

As the air mass continues to rise, the water droplets continue to cool and form ice crystals. Gravity causes these droplets and crystals to start to fall, causing a downward movement to compete with the upward lift. Instability between the updrafts and downdrafts causes static electrical charges to build up in the cumulonimbus cloud. The discharge of this electricity causes thunder and lightning. During the spring and summer, cumulonimbus clouds are more likely to form in the afternoon, due to the heating of the earth's surface. However, they can also form along a cold front when the warm buoyant air is forced upward by the heavier cold air mass that cuts under the warmer air like a wedge. This can happen at any time of the year, as demonstrated by thunderstorms that happen in conjunction with snowstorms in the winter.

1.6.2. Appearance

The base of a cumulonimbus can be several miles across, and it can be tall enough to occupy middle as well as low altitudes: though formed at an altitude of about 3,000 to 4,000 metres (10,000 to 12,000 feet), its peak can reach up to 23,000 meters (75,000 feet) in extreme cases. Typically, it peaks at a much lower height.

Cumulonimbus are also characterized by a flat, anvil-like top (the anvil dome), caused by straight line winds at the higher altitudes which shear off the top of the cloud, as well as by an inversion over the thunderstorm caused by rising temperatures above the tropopause. This anvil shape can precede the main cloud structure for many miles, causing anvil lightning.

1.6.3. Effects

Cumulonimbus storm cells can produce heavy rain (particularly of a convective nature) and flash flooding, as well as straight-line winds. Most storm cells die after about 20 minutes, when the precipitation causes more downdraft than updraft, causing the energy to dissipate.

If there is enough solar energy in the atmosphere, however (on a hot summer's day, for example), the moisture from one storm cell can evaporate rapidly -- resulting in a new cell forming just a few miles from the former one. This can cause thunderstorms to last for several hours.

Cumulonimbus clouds contain severe convection currents, with very high, unpredictable winds, particularly in the vertical plane (updrafts and downdrafts).They are therefore extremely dangerous to aircraft. Smaller propeller-driven planes cannot cope with the conditions and must fly around them; larger jet aircraft fly over the smaller ones. Larger planes are also equipped with weather radar and wind shear detectors to help guide them through, in the event that they need to pass through such clouds to land.

The air convection can also form mesocyclones, which can cause hail and tornadoes.

Fig. 1.6: Cumulonimbus cloud at various locations

1.7. Thunderstorm

Thunderstorms are violent rain storms that produce thunder, lightning and frequently hail. The World Meteorological Organization defines it as "One or more sudden electrical discharges manifested by a flash or light (lightning) and a sharp or rumbling sound (thunder)"[B. J. Retallack, 1970]. They are usually of short duration, seldom over 2 hours. They originate from cumulonimbus cloud. Presence of warm and humid air in the lower layers of the atmosphere is an essential pre-requisite for the development of a thunderstorm. Atmospheric instability and intense convective activity are other important requirements for their origin and growth.

Thunderstorms form when air close to the ground is warm and humid. When this warm air lifts, it becomes cooler and the water vapour in the air condenses forming a cloud. If the cloud is warmer than the surrounding air, then the cloud will continue to rise. This causes the cloud to accelerate upward in the form of turbulent bubbles, giving the cloud its characteristics cumulus shape. A variety of conditions can cause the lifting needed to initiate these clouds, including the heating of the ground, wind blowing up and over a mountain, sea breezes, cold fronts and tropical low pressure systems. The greater the temperature difference between the relatively warm cloud and its surrounding air, the more vigorous the thunderstorm will be. If the wind speed and wind direction change significantly with height, the thunderstorm can rotate. These rotating thunderstorms provide the circulation that when concentrated in a small area result in a tornado.

Since thunderstorms require warm, moist air, they occur most frequently in the tropics. In temperate latitudes, they are more likely to occur during the hot summer than during the cooler seasons. Over land, thunderstorms occur most frequently in the afternoon and early evening because land surfaces heat up dramatically during the day and cool down at night. In contrast, thunderstorms at sea are equally likely to occur at all hours because large water surfaces maintain an even temperature throughout the day.

The dew point temperature in the boundary layer (Fig. 1.7) is a fairly good indicator of the fuel supply for the storm.

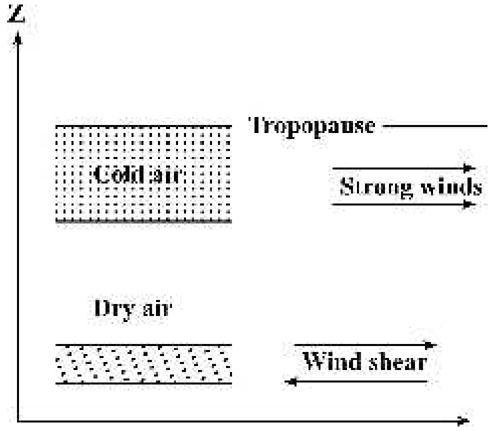

Fig. 1.7: An environmental conductive to severe thunderstorm

Larger dew points imply a warmer, moister boundary layer and favor more intense thunderstorms.

It is estimated at any given moment nearly 2,000 thunderstorms are in progress over the Earth's surface, and lightning strikes the Earth 100 times each second. There are about 45,000 thunderstorms daily and 16 million annually around the world.

Fig. 1.8: Thunderstorm picture from the space

1.7.1. Life cycle of thunderstorms

Thunderstorms undergo a three-stage life cycle (fig. 1.9).

(a) Cumulus stage

In this stage, warm, moist air rises in a buoyant plume or in a series of convective updrafts. When the air becomes saturated, a convective cloud begins to grow. As the warm air plume continues to rise, more water vapour condenses, releasing the latent heat of vaporization. This heat enhances convection, and cloud turrets form. The cloud edges during this stage are sharp and distinct, indicating that the cloud is composed primarily of water droplets. The convective cloud continues to grow upward, eventually growing above the freezing level where super-cooled water droplets and ice crystals coexist.

(b) Mature stage

The stage of maturity is characterized by the presence of both updrafts and downdrafts within the cloud. The downdrafts are initiated by the downward drag of falling precipitation. Air is entrained into the rain shaft, and the downdraft is strengthened by evaporative cooling as the precipitation falls into the sub-saturated air below the cloud base. The cold descending air in the downdraft often reaches the ground before the precipitation. If the freezing level is sufficiently high, the precipitation melts and reaches the ground as rain; otherwise, graupel showers occur. Cloud to ground lightning usually begins when precipitation first falls from the cloud base. As the top of the cloud approaches the tropopause, it starts to flatten out, forming an anvil shape.

(c) Decaying stage

The decaying stage of the thunderstorm is characterised by downdrafts throughout the cloud. Decay often begins when the supercooled cloud droplets freeze and the cloud becomes glaciated. Glaciation typically appears first in the anvil, which becomes more pronounced in this stage. The glaciated cloud appears filmy, or diffuse, with indistinct cloud edges. The cloud begins to collapse because no additional latent heat is released after the cloud droplets freeze and because the shadow of the cloud reduces insolation at the ground. Sinking motions occur with the fall of precipitation throughout the cloud.

The decay of a thunderstorm can also be initiated when the precipitation within the storm becomes too heavy for the updrafts to support, when the source of moisture is cut off, or when lifting ceases. The convection necessary for the growth of the thunderstorm can be interrupted by convective overdevelopment of the thunderstorm, that is, the storm itself becomes so large that it cuts off insolation. However, it is more common for thunderstorm to decay in the late afternoon or early evening when solar input diminishes, convection wanes, and sinking motions begin to occur over the entire mountain massif. Not all air mass thunderstorms decay in the late afternoon or early evening. Intense thunderstorms can form their own circulations, which feed water vapour into the cloud and convert it into water and ice, thus releasing latent heat that allows the storms to persist into the night.(Whiteman C. D., 2000)

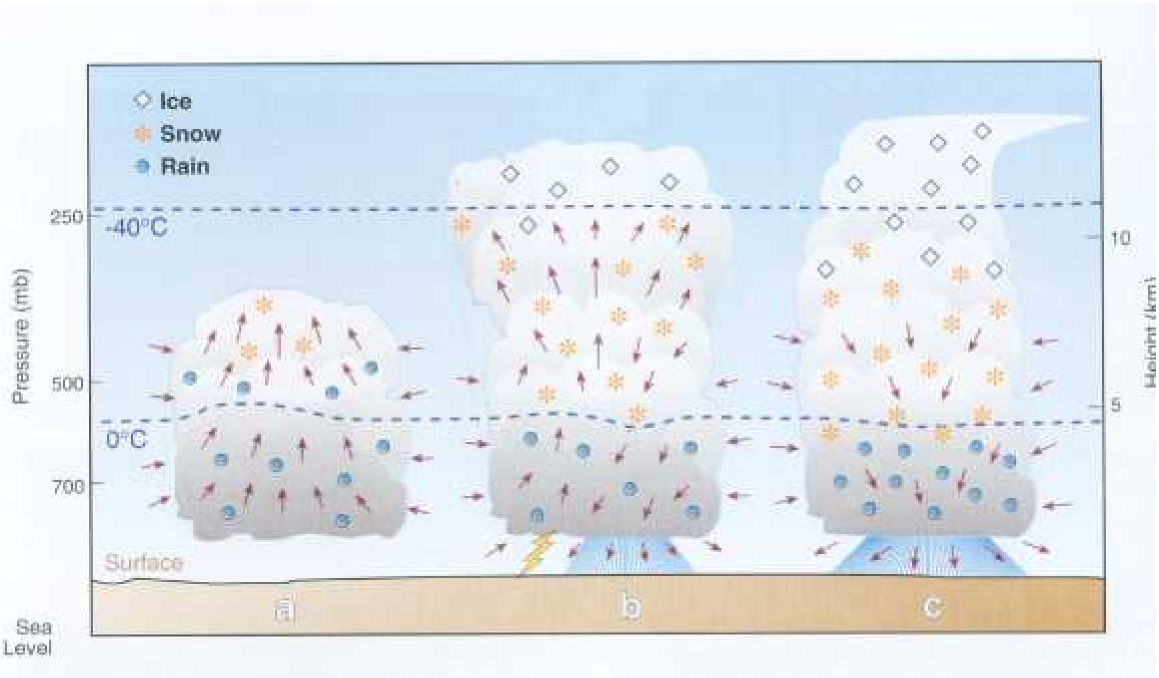

Fig. 1.9: Three stages of life cycle of thunderstorm

Fig.1.10: Cumulus stage of thunderstorm

Fig.1.11: Mature stage of thunderstorm

Fig.1.12: Dissipating stage of thunderstorm

1.7.2. Lightning

Lightning is a large spark, or burst of electric charge, that travels through the air from one charged region of a cloud to another charged region. It occurs when one region of the atmosphere becomes highly, positively charged and another region becomes highly, negatively charged, or when a large charge difference exists between the bottom of the cloud and the ground. These large charge differences are caused by repeated collisions of small ice pellets, called graupel, within the cloud. When the charge differential becomes large enough, a bolt of lightning occurs, carrying electrical charge from one region to the other and neutralizing the charge differential. The lightning can occur between regions of the cloud, between the cloud and the surrounding air, or between the cloud and ground. As the lightning travels through the air, it heats the air to temperatures as high as 30,000° C (54,000° F). Thunder, the loud noise that accompanies lightning, results from the explosive expansion of the air as it is heated.

Lightning is a major cause of forest fires. Based on studies in Montana (Fuquay 1980), the US Forest Service has developed a guide to identify and forecast lightning activity level (LAL) in connection with fire hazard. This is shown in abbreviated form in table 1.1.

The unit of area considered is a square approximately 80 km on a side, corresponding to the largest area over which lightning activity can be effectively monitored from a fire lookout point. Storms become more intense up to LAL 5, although the area covered by the storm does not increase at the same rate; even with LAL 5, measurable precipitation usually affects less than half of a forecast area. The relationship between lightning and precipitation is not well documented, but table 1.1 indicates that lightning activity is related to cloud development, as measured by the maximum height of radar echoes. (Barry R. G., 1992)

LAL cloud conditions	*Average cloud-ground lightning rate (min^{-1})*	*C-G lightning density (6500 km^{2})$^{-1}$*	*Maximum radar echo height (m a.s.l.)*
1. No thunderstorms	-	-	-
2. Few towering Cu	™1	20	< 8500
3. Scattered Cu, occasional Cb	Max. 1-2	40	7900-9700
4. 1-3/10 Cu, Cu congestus	Max. 2-3	80	9100-11000
5. Extensive Cu congestus, moderate-heavy rain with Cb	Steady flashes at some place during storm; max. > 3	-	-
6. Scattered towering Cu, high bases; virga common	™0.5	-	-

Table 1.1: US Forest Service guide to lightning activity level (LAL)

1.7.3 Formation of hail

Hail forms when the upward movements of the air in a thunderstorm, called updrafts, are strong enough to keep graupel in the upper, subfreezing part of the cloud for a relatively long time (tens of minutes). In this cold environment, the graupel will enlarge as water freezes to its surface. When the graupel becomes larger than 5 mm (0.2 in), it is called hail. In severe thunderstorms, hail that is larger than grapefruit may form.

The northeast corner of the Indian sub-continent during the pre-monsoon period is characterized by vigorous thunderstorm activity (Rao, 1981; Pant and Rupa Kumar, 1997; Laing and Fritsch, 1993; Mandhar et al 1999) the result of the establishing of a surface heat low over the north Indian state of Bihar which draws warm, humid air from the Bay of Bangal (Rosoff et al 1999) and superposition of the cold 500 and 200 mb flows from the west. These surface and upper air flows combine to set up ideal conditions for thunderstorms.

Objectives of the study

Using surface and upper air data this study attempts to explain why the thunderstorm occurred and why thunderstorm did not occur. The main purpose of this study is to develop procedures for forecasting severe thunderstorm activity in the Khumbu Himal of the eastern region. For this purpose we have chosen the weather station data of Lukla. In this season many thunderstorms were observed in the past. Lukla is the gateway for the Mt. Everest is directly air-linked with the capital city.

CHAPTER 2

Data Base and Methodology

This study is based on the analysis of meteorological data and weather charts. The following data were used for the period of April 24, 2005 to May 18, 2005.

The surface meteorological data was collected from the automatic weather station (Rainwise Inc. USA). The system in the instrument is configure to measure temperature, barometric pressure, wind speed and direction, rainfall and intensity, humidity, dew point temperature and solar radiation. The station is located near the Lukla airport (Lat-27^0 42’ N, Long-86^0 44’ E, Altitude about 2700 m) of Solukhumbu district. Location of the weather station has been shown in Fig. 2.1. This is the station of Central Department of Hydrology and Meteorology. For this special purpose the data of four (4) minutes intervals was chosen.

For the comparison, surface weather data published by DHM were also used.

Rawinsonde data from the NOAA internet site (raob.fsl.noaa.gov/) of Silguri, Gorakhpur, Patna, Luknow, and Lhasa in Tibet have been used.

Reanalysis synoptic charts of the Indian sub-continent available on the NOAA site at www.arl.noaa.gov/ready

Cloud bases were observed manually and the height were estimated by considering the mountain peaks as the reference frame.

Convective Available Potential Energy (CAPE) value was estimated by using RAOB programme.

Fig. 2.1: Location of the weather station

CHAPTER 3

Result and Discussion

Fig.3.1 (a) shows the variation of temp. and dew temp. with time on 13 May 2005. On that day, air temp was 7°C till 04:00 NST and the dew temperature was 3 °C. After 04:00 NST, temp and dew temp decreased slightly till 07:00 NST. After 07:00 NST temp gradually increased till 11:20 NST. Within this time dew temp was decreasing and reached -1 °C at 08:24 NST. The maximum difference in temp and dew temp was seen around 1100 NST. After 1120 NST. temp gradually decreased till 15:36 NST reaching 5°C. But the dew temp increased till 13:24 NST and became 7°C. After 13:24 NST, there was fluctuation in temp and dew temp till 20:56 NST and the temp and dew temp was constant after 20:56.

Here, the greater difference between temp and dew temp was seen around 11:00 NST. It is the favourable condition for the development of thunderstorm since the greater difference between temp and dew temp, the more vigorous the thunderstorm will be.

But On May 01, 2005, the temperature and the dew temp were less and almost the same (fig.3.1 (b)). It is not the favourable condition for the development of thunderstorm.

Fig.3.2 shows the variation of solar radiation with time on May 13, 2005. On that day, solar radiation was 0 till 06:12 NST and after that time it was gradually increasing. The maximum solar radiation was received around 10:56NST (about 810 w/m^2). After 10:56 NST, there was fluctuation on solar radiation. Around 14:00 NST, solar radiation was almost 0.

Presence of warm and humid air in the lower layers of the atmosphere is an essential pre requisite for the development of a thunderstorm. On 13 May, the incoming solar radiation caused the heating of the surface which yield the lower layer of atmosphere warm, which is one of the favourable conditions for the development of thunderstorm.

Fig.3.3 shows the variation of pressure with time on 13 May 2005. On that day, barometric pressure was 727 mb till 02:28 NST. After 02:28 NST, pressure decreased reaching 726 mb and was constant till 03:56 NST. After that time there was fluctuation in pressure till 04:56 NST. After 04:56 NST pressure was 727 mb till 07:48 NST. After 07:48 NST, pressure started to decrease till 10:12 NST and reached 724 mb and was constant till 13:08 NST. After that time there was fluctuation in barometric pressure till 18:32 NST. After 18:32 NST, pressure was increasing and reached 728 mb at 23:56 NST.

It is seen that the barometric pressure has decreased which cause the low pressure centre. It can cause the lifting needed to initiate the cloud which is responsible for the development of thunderstorm.

Fig.3.4 shows the variation of wind speed and wind direction with time on 13 May 2005. On that day, wind direction was north easterly till 04:04 NST and the wind speed was 1 to 2 m/s. After 04:04 NST, wind direction was easterly till 07:00 NST. But the wind speed was still 1 to 2 m/s. After 07:00 NST, wind direction was again north easterly till 08:20 NST. After 08:20 NST, wind direction was westerly till 13:08 NST. After 13:08 NST, there was fluctuation in wind speed and wind direction till 18:04 NST. After 18:04 NST, the wind direction and wind speed both were nearly constant.

It is seen that the wind direction was westerly from 08:20- 11:28 NST. This suggests that on May 13, the moisture necessary for the thunderstorm initiation was transported from the west 5 hrs earlier.

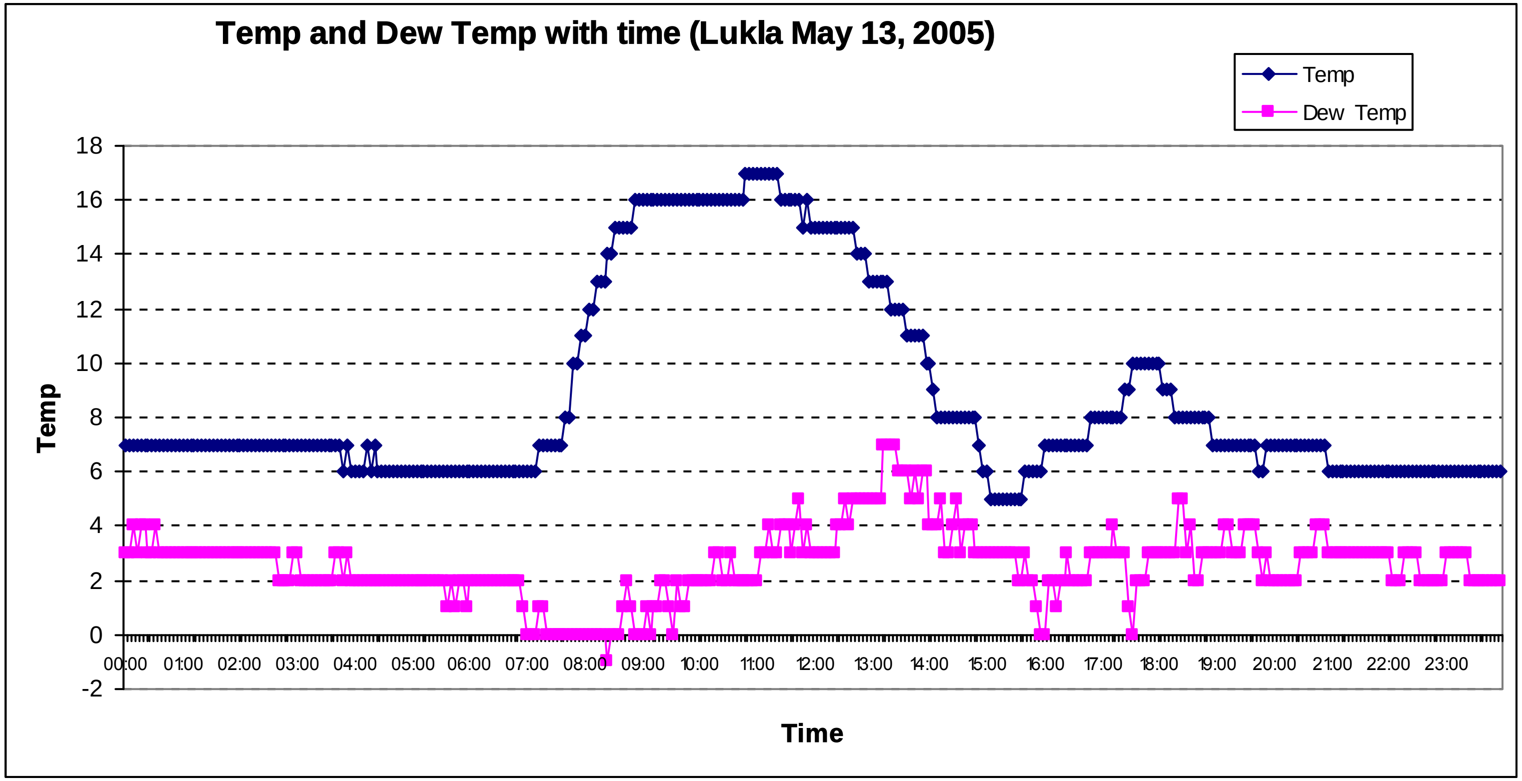

Fig. 3.1 (a): Variation of temperature and dew temperature with time

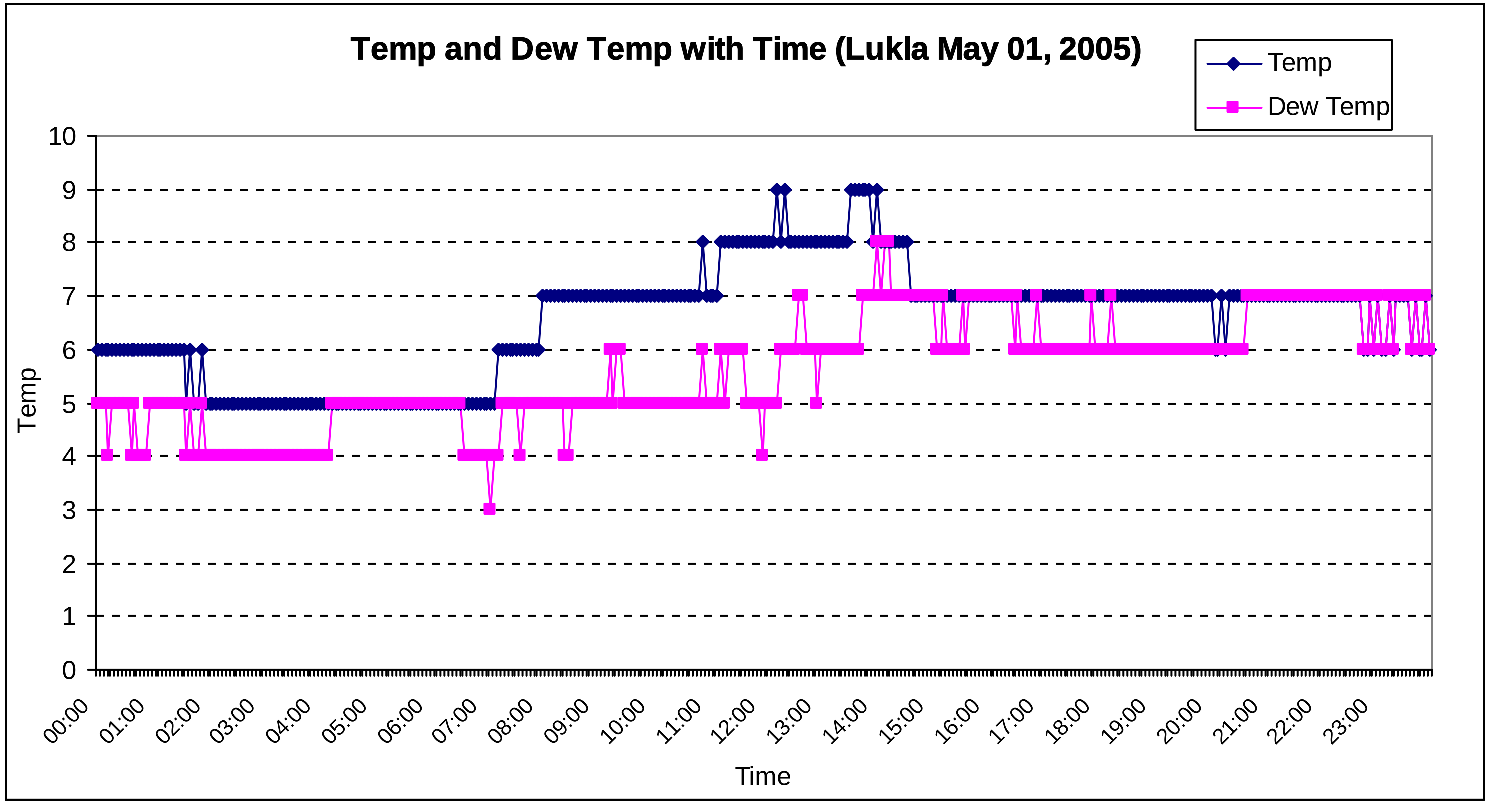

Fig. 3.1 (b): Variation of temperature and dew temperature with time

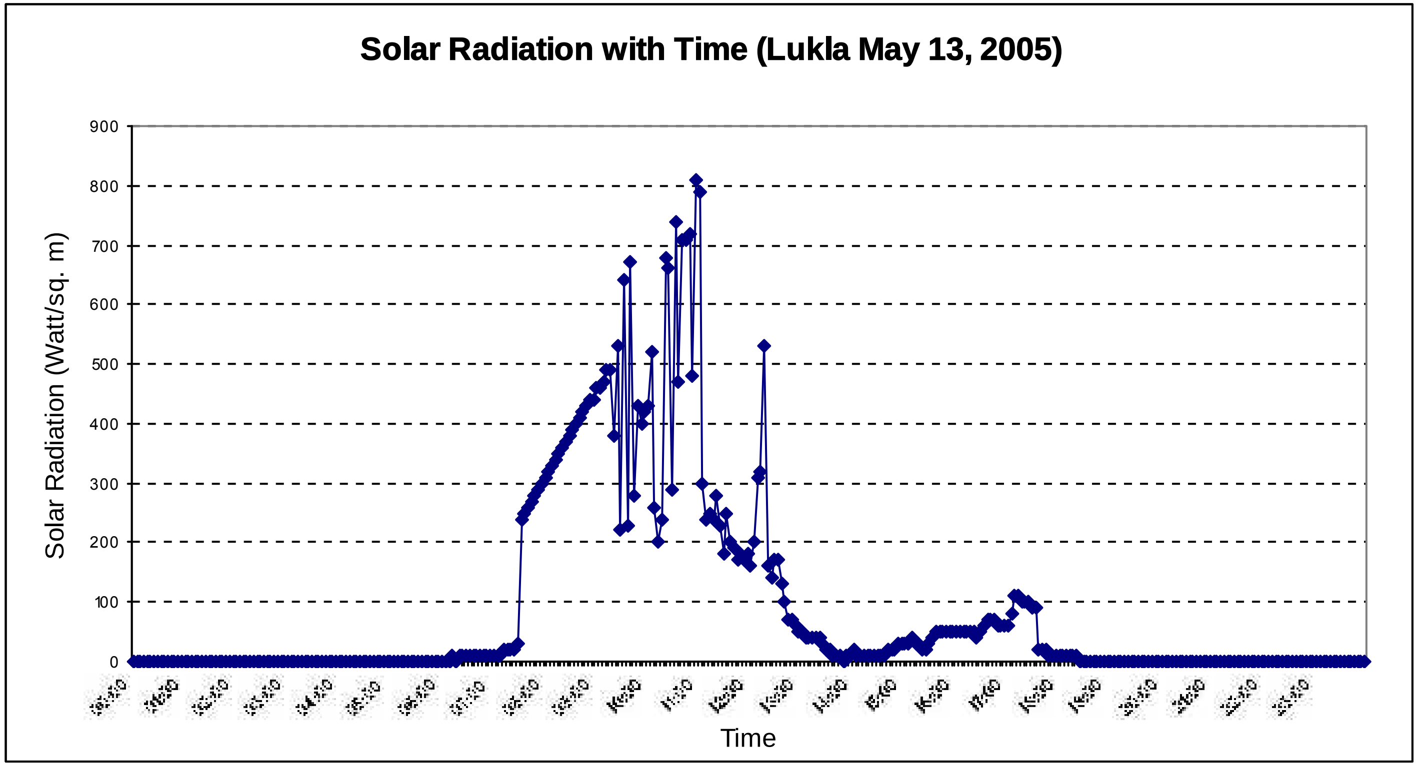

Fig. 3.2: Variation of solar radiation with time

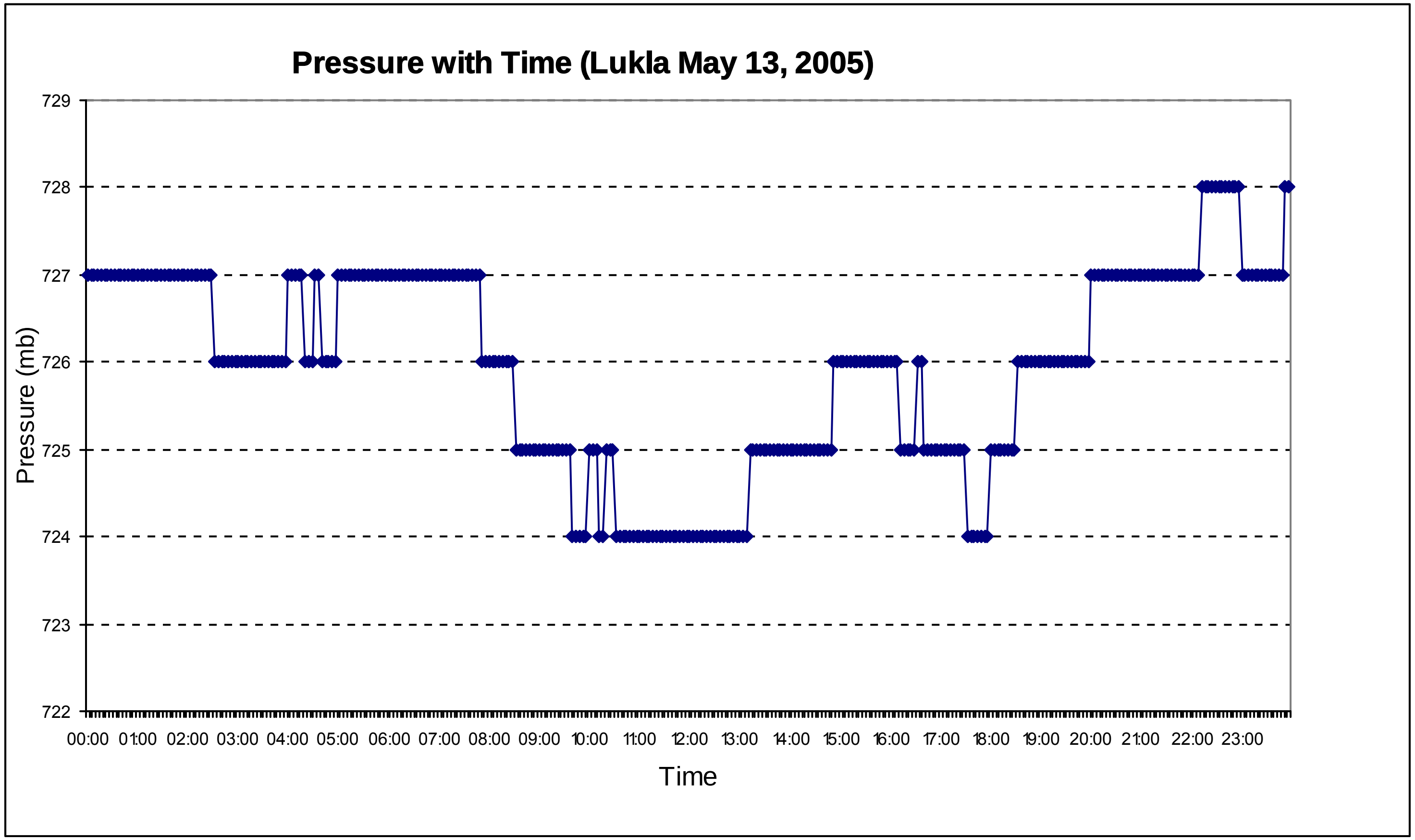

Fig. 3.3: Variation of pressure with time

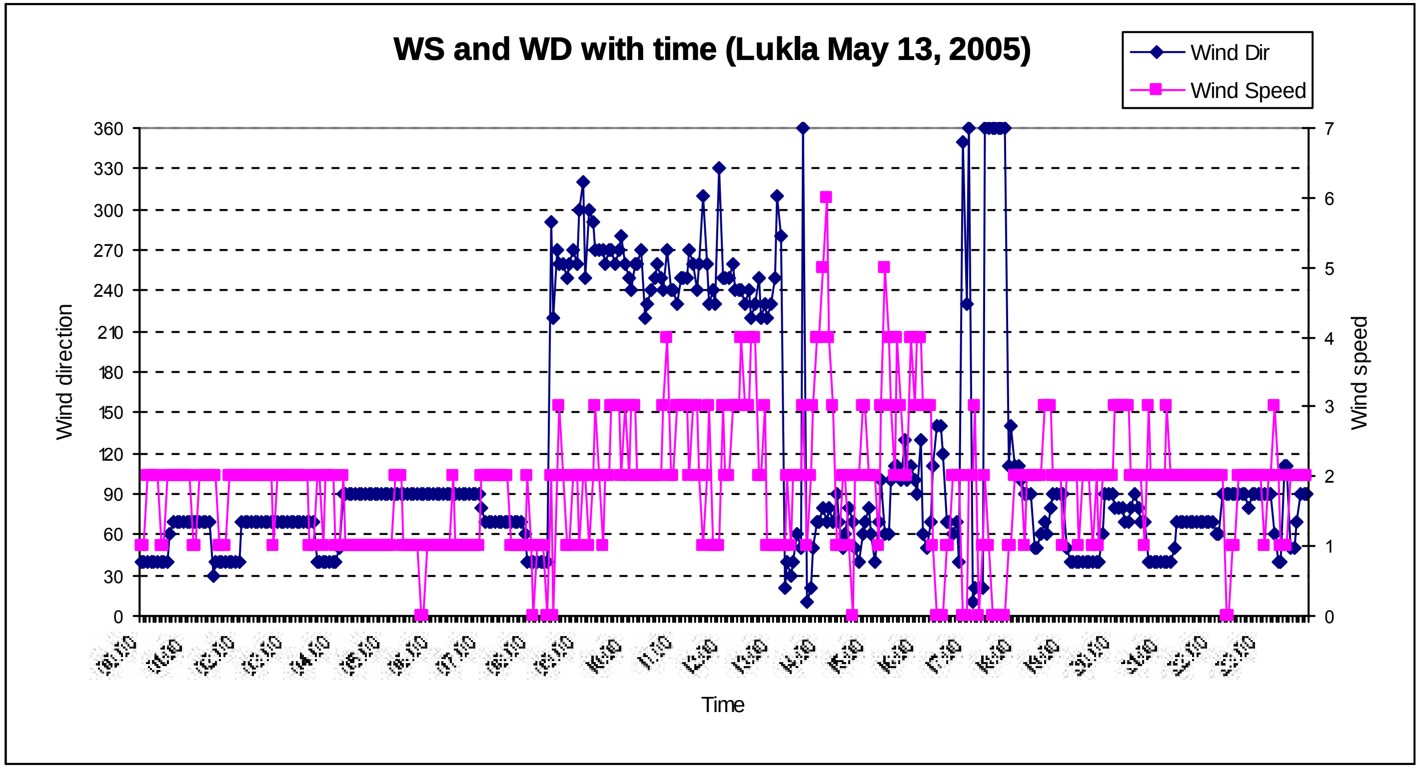

Fig. 3.4: Variation of wind speed and direction with time

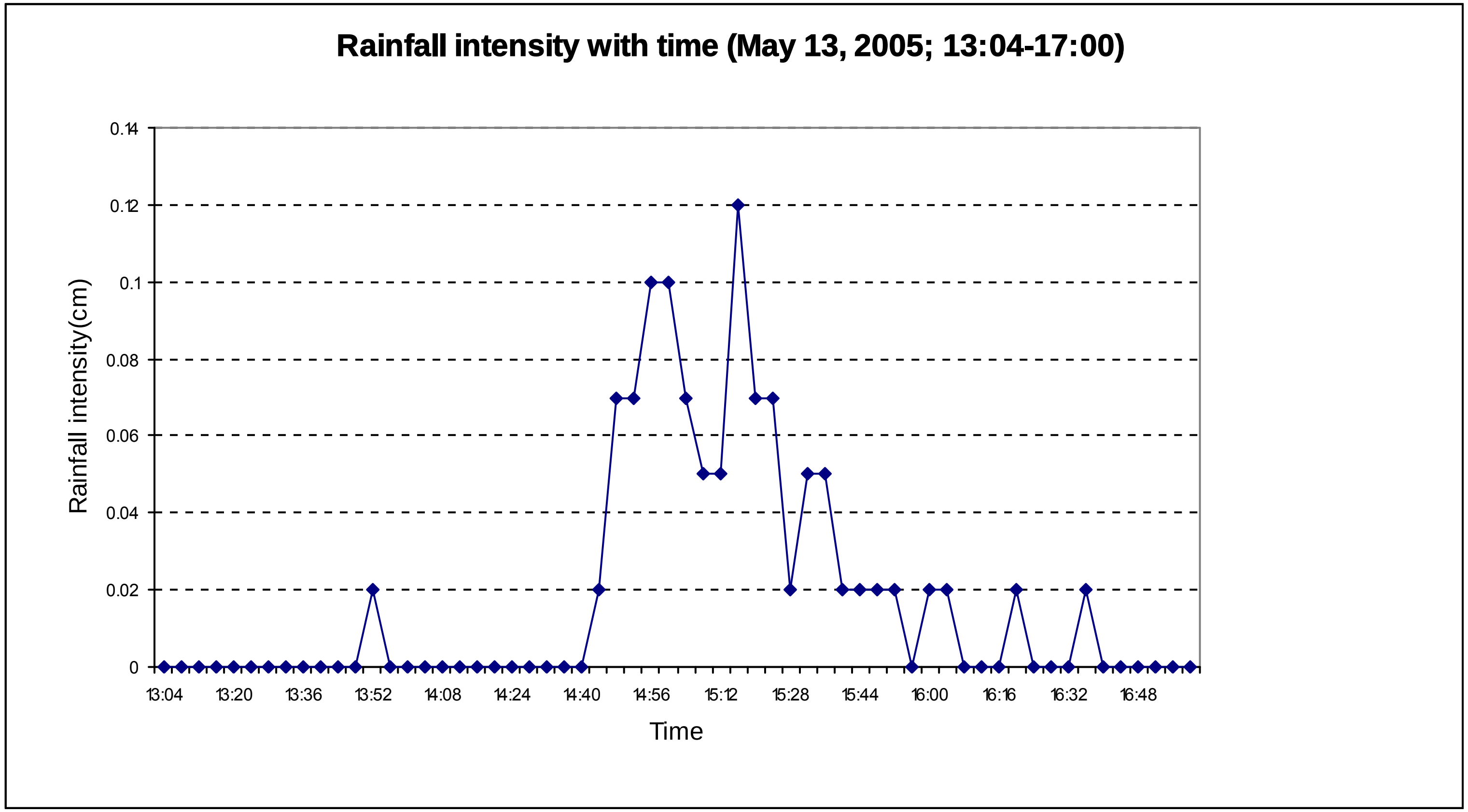

Fig.3.5: Variation of rainfall intensity with time

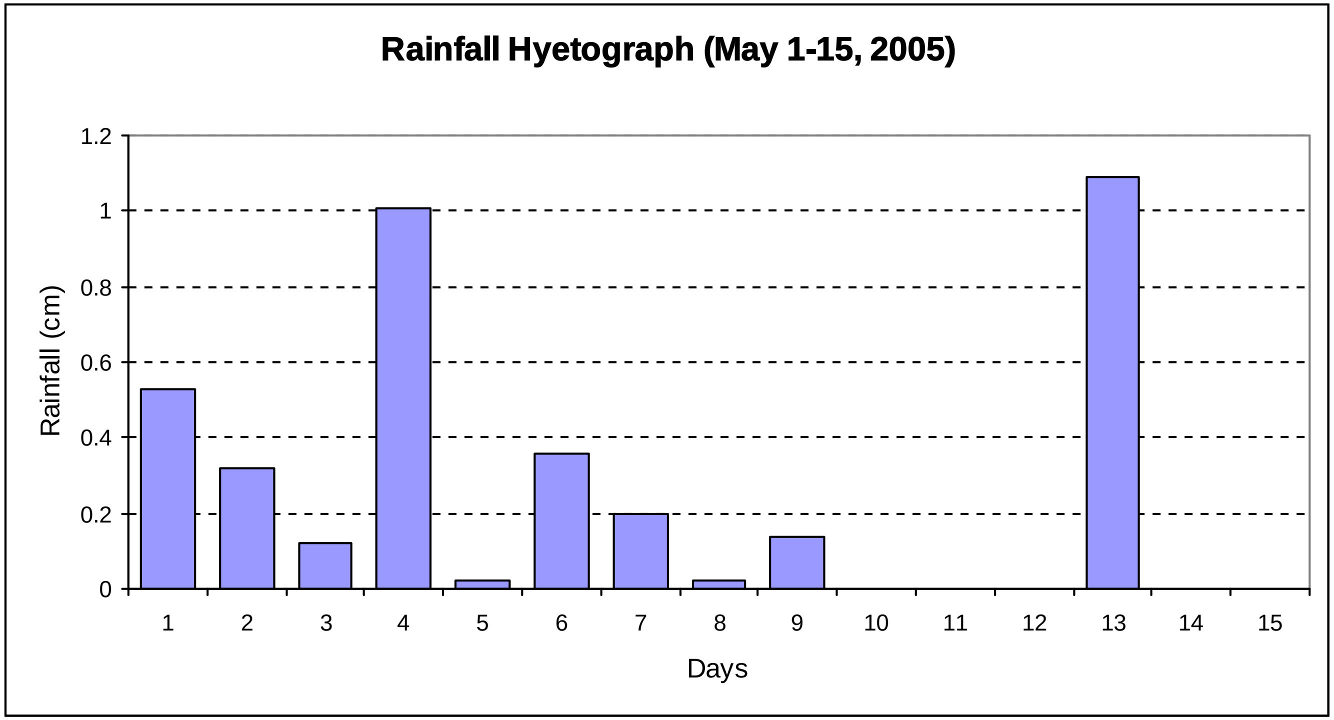

Fig. 3.6: Rainfall hyetograph

From the analysis made by Rosoff et al (1999) showed that combining the interpolated soundings with the actual surface temperatures and dew points and then calculating a lifted index does not give a true indication of thunderstorm potential in the Khumbu. Probably due to the lack of sufficient data the authors could not show the exact relationship.

In this study we found that the range of Temperature and dew point temperature become high before the period of thunderstorm.

Since this study endeavors to find procedures for thunderstorm prediction, we are forced to use other methods to identify severe weather potential in the Khumbu.

Table 3.1: CAPE Values

Date	CAPE (J/Kg)	Thunderstorm
26 April 2006	1436.09	Yes
1 May 2005	2.76	No
11 May 2005	689.27	No
12 May 2005	93.64	No
13 May 2005	1388.03	Yes

Table 3.2: Cloud Cover and occurrence of thunderstorms on different days

Date	Cloud Base AGL	Cloud cover	Thunderstorm
26 April 2005	800 m	Clear upto 1300 NST	Yes (15:00-16:30)
1 May 2005	1200 m	Overcast almost all day.	No
11 May 2005	1800 m	Early morning foggy Clear around 11:00-12:00. Then after scattered and late evening overcast.	No
12 May 2005	1800 m	Cloudy in the morning	No
13 May 2005	800 m	Clear upto 1230 NST	Yes (14:40-16:20)

Table 3.1 shows Convective Available Potential Energy (CAPE) values for all the Khumbu 0545 hrs (NST) sounding from the surface 700 mb. CAPE measures potential thunderstorm severity in Joules/ Kg and is the positive buoyancy of a profile bounded by the sounding and the moist adiabat. For the purpose of comparison, a temperature of 0^0 C was chosen and the parcel was lifted saturated, from the surface. The CAPE comparison isolated 1,11 and 12 May with the values of 3, 689 and 93 Joules/ Kg respectively from the non thunderstorm days. CAPE values for the days of 26 April and 13 May were 1436 and 1388 Joule/ Kg respectively.

Observations of convective cloud formation each day indicated the arrival of humid air. The transport mechanism of this humid air may have been either the upper air flow from the west or the valley flow from the south or a combination of both. Cloud movement was observed to be south to north in lower altitudes but west to east at higher altitudes. Clouds were observed to form at lower elevations several hours earlier before they formed at the Khumbu site and cloud formation was usually very sudden. The cloud base heights at the lower elevations were always observed to be considerably lower than the cloud base heights at the Khumbu site.

At the Khumbu site, the 700 mb and higher flow has been observed to interact with the valley flow. Frequently small cumulus observed rising from lower elevations in the south to north valley flow were either instantly whisked away to the east when they reached higher elevations, or dissipated immediately. Thus the 700 mb flow may represent a source of a "low level" moisture, either enhancing or suppressing cumulonimbus formation, but a lack of adequate dew point data from the India soundings prevented a detailed analysis.

Table 3.2 shows the observed cloud cover and cloud base heights for the selected study days. The two thunderstorm days shared what appeared to be identical cloud base heights, 800 m AGL, plus clear morning. Cloud base heights of the non-thunderstorm days were higher and ranged from 1200 to 1800 m AGL. The lower the cloud base heights on the thunderstorm days indicate the presence of more humid air.

Also cloud base heights were not reflected in the surface temperature and dew point on the thunderstorm days.

Fig. 3.9: A taxi damaged by hailstone during thunderstorm on 18 May 2005 at Pokhara

CHAPTER 4

Conclusion

Here we try to find the relationship between meteorological data and the possible thunderstorm occurrence. A highly touristic place Lukla has been chosen for this purpose. In order to predict the thunderstorm potential for a particular day we found that the range of Temperature and dew point temperature become high before the period of thunderstorm.

Convective Available Potential Energy (CAPE) values was calculated by using ROAB programme taking the upper air sounding of Siliguri. The CAPE comparison isolated 1, 11 and 12 May with the values of 3, 689 and 93 Joules/ Kg respectively from the non thunderstorm days. CAPE values for the days of 26 April and 13 May were 1436 and 1388 Joule/ Kg respectively. It shows that the greater the CAPE values, the greater potentiality for the occurrence of thunderstorm.

The two thunderstorm days shared what appeared to be identical cloud base heights, 800 m AGL, plus clear morning. Cloud base heights of the non-thunderstorm days were higher and ranged from 1200 to 1800 m AGL. It shows that the lower the cloud base heights, the higher potentiality for the occurrence of thunderstorm. Along this, the clear morning is the mechanism for the prediction of thunderstorm.

Recommendation

- More automatic weather stations should be kept around the study area which is beneficial for the climbers, trekkers and other tourists.
- Data interval is good for 2-5 minutes.
- Upper air meteorological stations should be launched in near future.

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