STUDIES ON STORM SURGES AND SEA LEVEL VARIATIONS ALONG THE EAST COAST OF INDIA

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CERTIFICATE

This is to certify that this Thesis is an authentic record of research work carried out by Mr. A. Satyanarayana Murty, M.Sc. (Tech.), under my supervision and guidance in the Department of Marine Sciences for the Ph.D. degree of the University of Cochin and no part of it has previously formed the basis for the award of any other degree in any University.

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C O N T E N T S

PREFACE

About 15% of the tropical cyclones in the world originate in the Bay of Bengal and the Arabian Sea in an year (Gray, 1975). Of these, the majority have their genesis over the Bay of Bengal, striking the east coast of India and the coast of Bangladesh. The dangerous part of the cyclones is the surge generated by them when they cross the coast. The magnitude of devastation due to a storm surge can be fathomed and understood from the two notorious surges of 1970 and 1977, where the former took away more than 200,000 lives in Bangladesh and the death toll in the latter rose to more than 10,000 apart from heavy loss of live stock and damages to the crop, buildings etc.

It became so usual for the east coast of India to face atleast 10 to 15 cyclones every year, out of which 3 to 4 may reach the deep depression stage. As a result the east coast of India experiences frequent heavy damages of varying intensities due to storm surges and it is also not unusual to experience a calamitous deluge once in a decade or so. Loss of life and damages can be minimised only if the magnitude of the surge could be predicted atleast a day in advance. Therefore, an attempt to study the storm surges generated by the cyclones that strike the east coast of India and suggest a method of predicting them through nomograms, is made.

The thesis comprises of six chapters with subsections as detailed below:-

Chapter I	Introduction Section I: Storm surges Section II: Sea level variations Section III: Description of the area Section IV: Treatment of the data
Chapter II	Case studies of storm surges along the east coast of India
Chapter III	Sea level variations along the east coast of India.
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CHAPTER I

INTRODUCTION

CHAPTER I

INTRODUCTION

For some years at the turn of the century scientists alluded the cyclone as one of the simplest of atmospheric disturbances as a symmetric circulation of an ideal vortex in a barotropic environment. But for its awesome destructivity it was felt by many to be of trivial academic interest only.

After several decades of observing the cyclones from research aircraft, after nearly a decade of monitoring its environment with weather satellites, and after numerous experiments with numerical simulation models of its circulation, scientists have established the cyclone as one of the most challenging problems in marine meteorology.

Cyclonic circulation is not a simple or ideal vortex. The circulation properties depend mostly on the intimate coupling between air and sea surface through the mixed layer. However, such a coupling is less important in the developmental phase of a tropical cyclone than it is in its mature stage. Sufficient energy in the form of latent heat of condensation is available, from the influx of moist tropical air to set the heat pump in operation, to make fall the surface pressure in the core of the disturbance by 10 to 15 mb and to sustain winds of gale force.

As the organised winds of a tropical storm begin to sweep the sea and disturb the upper few tens of metres, the coupling of the atmosphere and sea plays a decisive role in its development. The storm system cannot acquire hurricane strength and contract the circle of maximum winds tightly around the eye of the storm unless the heat content of the inspiraling tropical air is augmented by 12 to 15%. And this augmentation can only come from a flux of heat directly from the ocean reservoir.

Another aspect of air-sea coupling that is more dramatic and is of great practical significance is the generation and maintenance of the potential for storm surge, a phenomenon that brings disastrous inundation and often responsible for collosal loss of life and property along coastal areas as a deep depression moves inland. The potential for a storm surge is generated over the open seas by a unique and intricate combination of two factors, the first of which is the 'inverted barometer' which is the hydrostatic rise in sea level due to the atmospheric pressure drop in the cyclone area and the other is the large scale swirl or vorticity of water generated by the stresses of inspiraling winds on the water surface.

In deep water, the uplift of the sea surface measures only in tens of centimetres over distances of kilometres. Due to the asymmetry of wind stresses, the maximum vorticity or swirling transport of water occurs to the right and abreast of the storm centre in the northern hemisphere. In deep water, this vorticity extends its influence deep into the thermocline to trigger internal waves, but does not contribute to a significant rise in sea level until the system approaches shallow water. Where water depths decrease to 90 m or less, the deep column of swirling water set in motion by the wind stress begins to sweep the bottom, and the whole column tends to shrink suddenly. Since vorticity is a conservative property, there must be either a divergence of mass or an uplift of the sea surface. As bottom friction retards divergence, the result is an uplift of the sea surface which, combined with the inverted barometer effect, reaches a peak height on the right of the moving centre, near the position of maximum sustained winds. The mound of water thus generated in this manner, is what is known as the storm surge.

A storm surge is usually defined as 'a raising or lowering of sea level produced by wind and by changes in atmospheric pressure over the sea associated with a storm' (Heaps, 1967). The atmospheric pressure usually acts in a statical sense and its dynamic contribution is quite negligible (Heaps, 1967; Prandle, 1975; Ali, 1980a). As the atmospheric pressure on the sea surface falls, the sea level rises through inverted barometer or sucking effect. This rise is generally known as the statical rise. The statical contribution is about 0.01 m per one mb pressure drop which was first examined by Gissler in 1747 for the Gulf of Bothnia. Since then, observations at other regions have shown this to be a common rule (Rossiter, 1962).

Wind is the main cause of surge generation. Wind exerts tangential as well as normal stresses on the water underneath; the effect of the latter being extremely small (mainly because water is considered incompressible) compared with that of the former. The tangential stress drives water in the direction of the wind in the beginning, and later under the influence of the earth's rotation, it is deflected to the right in the northern hemisphere and to the left in the southern hemisphere. Thus, a long water wave or storm wave (i.e., storm surge) is generated, in the sea. When this water movement is impeded by a coast, the surge height usually tends to rise as the coast is approached and results in coastal flooding.

The amplitude of the surge depends directly on the strength of the wind. When the cyclone is in deep water far from the coast, the generated surge is not so dangerous because counter currents in the subsurface layers are immediately set-up and they tend to maintain the sea surface at level. When the cyclone approaches the coast the counter currents are retarded or nearly blocked by the frictional forces at the shallow bottom. As a result, the piling-up of water against the coast takes place.

Implicit in its definition is the fact that a surge can be either positive or negative. A negative surge is defined as the fall of water surface below the mean sea level. It occurs only on the left hand side of the cyclone where the wind is offshore when looking in the direction of the moving cyclone(in the northern hemisphere). Alternatively, a positive surge occurs on the right hand side of a cyclone due to the onshore component of wind. Unless, otherwise stated, a storm surge means here the positive surge, and estimates show that most of the disasters caused by a cyclone are due to positive surges.

A storm surge is sometimes considered under three categories: fore-runner, proper surge and resurgence (Miller, 1957). A fore-runner is the gradual rise in water level along the coast well before the arrival of the cyclone. This occurs when the cyclone is at a considerable distance from the coast. In the pre-satellite days, the fore-runners were used as good indicators of the formation and impending dangers of a cyclone. The 'surge proper' is the substantial but sudden rise in water level that accompanies the cyclone, and is the principal cause of damage. Resurgence is generally attributed to the free motion of water returning to its normal level after the cyclone crossed the coast.

Sometimes storm surges are confused with tidal waves. But the two have different origins. A storm surge, as described earlier, is of meteorological origin, whereas a tidal wave, a misnomer, is caused due to submarine disturbances such as volcanic eruptions, earth-quakes, turbidity currents and the like.

SECTION - I

STORM SURGES :

1.1.1. Factors modifying the surge:

Once a storm surge is generated, its amplitude may be modified by several factors. Some of the important factors are:-

i) depth of the basin and topography of the shelf area

ii) shape of the basin (configuration of the coast line)

iii) size and intensity of the storm

iv) track of the storm (angle of incidence of the storm with respect to the coast line)

v) phase of the astronomical tides

vi) presence of inlets, such as rivers, creeks, canals, estuaries, backwaters and islands at the landfall.

The surge height is inversely related to the depth of water. Consequently surge height gets amplified in a shallow region, as water is driven from deeper to shallower area, water level must rise for continuity reason. At this time, a return current developed underneath is retarded by the strong frictional influence at the shallow bottom. Further, the surge height is accentuated on a converging coast (such as Gangetic Coast of the Bay of Bengal) than on an open coast. Storm surges are usually higher to the right of the track (Das, 1972; Ali, 1979). When the peak surge arrives at the time of high tide, the surge effect becomes disastrous. However, tide and surge do not interact linearly, which aspect makes the surge prediction complicated.

The presence of rivers, canals, creeks, estuaries and backwaters sometimes makes the surge effect worse. Dynamically, a river system is important for two reasons, viz., i) variations in the amount of fresh water discharge may further modify the sea surface elevation resulting from surge and tide, and ii) rivers allow a potentially deep inland penetration of saline surge water leading to inland flooding during the period of cyclone. This latter effect is commonly referred to as 'the tidal bore' which manifests in the running of sea water like a wall against the fresh water discharge. This situation causes two kinds of hazards - flooding in the low-lying areas of the river banks close to the mouth and scarcity of fresh water for drinking and other essential needs of human livelihood. Rivers at the head of the Bay of Bengal are, perhaps, the best examples of such a situation.

The presence of backwaters and estuaries modifies the surge in a different way. As these water bodies are fully or partially enclosed, the sudden incursion of water from the deep sea into the basin abruptly increases the water level of the body and as a consequence, sudden flooding may take place in the hinterland.

Islands help to amplify the magnitude of the surge by constraining the storm surge water as it rushes through the channels between them. As a result of the sudden onrush of water into the channel, the islands are flooded. The rushing of water between the islands may also cause coastal erosion due to the effect of storm surge diffraction.

1.1.2. Major areas vulnerable to surges:

Most of the records of storm surges are from United States, United Kingdom, Finland, Netherlands, Japan, India and a few from Germany. Obviously, these are the countries which suffer most due to storm surges and they are therefore best studied there. Among them the worst sufferers are countries surrounding the North Sea, the Atlantic Coast of the United States, the Japanese Coasts in the Pacific, the coasts of Phillippines and the coastal areas bordering the Bay of Bengal.

Storm surges have also occurred in lakes and gulfs such as Lake Okeechobee, Lake Erie, Lake Michigan, Lake Biwa, the Gulf of Bothnia, the Gulf of Mexico and the Persian Gulf. Among the countries surrounding the Bay of Bengal and Arabian Sea, India and Bangladesh are the worst affected by storm surges.

1.1.3. Historical resume of surge studies:

It may be stated that our understanding of the theory of storm surges has begun along with the theory of waves and tides as described by Airy (1845). Since then, many contributions have been made by numerous authors to the study of storm surges either directly or indirectly. Much of this early information was summarized by Welander (1961), Heaps (1967) and Lisitzin (1974). These early studies were developed in two phases, before and after the World War II.

Important contributions to storm surge studies have been made only after 1900, even though a few references prior to 1900 are available (Reid, 1849; Wheeler, 1895,1896, Oktt, 1897; Henry, 1900). All these workers not only studied tides and waves but also mentioned about the changes in water levels.

Between 1900 and 1930, contributions to the knowledge of storm surges (it was called storm tide at that time) were limited to about 10 to 15% of the total number of papers published, on the subject of waves and tides (Bretschneider, 1967). Chrystal (1906) investigated the problem of seiches in Loch Ern from the point of hydrodyhamics and field surveys and Honda (1915) studied lake seiches in Japan. It appears that some of the more important investigations began with the works of Eliot (1900), Abott (1913), Taylor (1916,1918, 1920), Cline (1920), Petterssen (1921), Jeffreys (1926), Marmer (1926), Proudman and Doodson (1926) and Yamaguti (1929).

Contributions to this subject were quite limited in number between 1930 and 1945, perhaps, because of the war. Nevertheless, some of the more important works on storm surges appeared during this period. For example, Cline (1933) worked on tides and coastal currents developed by tropical cyclones. Montgomery (1935), and Rossby and Montgomery (1935,1936) studied the vertical distributions and wind stresses over water. Arakawa and Yoshitake (1939) investigated the rise in elevation of the sea surface under a moving squall line. The work of Hellstrom (1941) on wind action on lakes brought out the concept of wind setup theory. The application of the concept of hydrodynamics to surges was also initiated during this period by Lamb (1945). Incidentally, the U.S. Army Corps of Engineers began to report data on hurricanes since 1938 that were highly useful for the studies on storm surges. About the same time, the U.S. Weather Bureau began to prepare charts on hurricanes and hurricane surges. Most of this information was documented by Bretschneider (1967).

Perhaps, over 75% of all the contributions on surges have appeared after 1950 and the literature was reviewed by Rossiter (1959a), Welander (1961), Bretschneider (1967), Heaps (1967), Lisitzin (1974) and Flather (1977).

In the past, storm surges have been profitably studied by empirical and statistical methods (Witting, 1918; Doodson, 1924, 1929; Corkan, 1948; Rossiter, 1959b; Keers, 1968). But in recent times, theoretical studies have been actively pursued by numerical integrations in time and space, involving viscosity and bottom friction. One of the first efforts on these lines was that of Kivisild (1954) for the wind tide (storm surge) computation in the Lake Okeechobee. He carried out the computation manually. However, the advent of electronic computer soon accelerated the development of theoretical model studies in this field. Later, Hansen (1956) computed the storm surge in the North Sea induced by 'Holland-Orken' of 1953, and Platzman (1958) computed the surge in the Lake Michigan accompanied by the moving squall line of 26 June 1954 from the standpoint of forecasting. The important feature that these studies have in common is the principle of numerical integration of the relevant dynamical equations. In these, responses of various factors that cause the surge are numerically estimated as functions of known dynamic processes using the local boundary conditions to the integrated hydrodynamic equations. Subsequently, many workers extended storm surge investigation along these lines. Our knowledge has been enhanced and modified through the works of Svansson (1958), Fischer (1959), Uusitalo (1960), Henning (1962) and Platzman (1963). Of late, storm surge studies have been greatly benefitted by the application of numerical models of the second and third order (Heaps, 1971,1973; Clarke, 1974). Much of the literature on numerical model studies appears from United Kingdom, United States and Japan.

1.1.4. Work done in the Indian seas:

A number of general reviews and descriptions of individual cyclones, and the associated surges in the seas around India have been published by the India Meteorological Department. Most of these studies are confined to the Bay of Bengal rather than the Arabian Sea mainly because the east coast of India and the coastal parts of Bangladesh are more vulnerable to the effects of the storm surges and also because more number of cyclones originate in the Bay of Bengal (Table I). Pertinent references from this area are those of Eliot (1900), Dunn (1962), Bay of Bengal Pilot (1966), Bhaskara Rao (1968), Bhaskara Rao and Mazumdar (1966a, 1966b), Ratnam and Nayar (1966), Janardhan (1967), Frank and Hussain (1971), Flierl and Robinson (1972), Das (1972), Das et al. (1974), Flather (1977), Ghosh (1977,1980,1981), Johns (1977), Ali (1979,1980a, 1980b,1980c,1980d,1980e,1981a,1981b,1982), Sharma and Satyanarayana Murty (1979), Johns and Ali (1980), Dube et al. (1981,1982) and Johns et al. (1981,1982). Since these studies span over a period of eighty years and involve several approaches to the study of storm surges, it is felt desirable to survey the more important literature in some detail.

Eliot (1900) described the accumulating water associated with the cyclone at the Hooghly estuary, as the advancing wall of water upstream due to the Calcutta Cyclone of October 1864 and the storm surge of 40 ft caused by the Backer Gunge Cyclone of October 1876. Bay of Bengal Pilot (1966) mentioned that very strong wind generated currents occurring in the Bay of Bengal during the period of cyclone could influence the sea levels to rise when they impinge on the continental shelf. Ratnam and Nayar (1966) suggested that the resonant coupling mechanism of the kind shown by Proudman (1929) occurs when the speed of the travelling atmospheric pressure disturbance matches that of free gravity waves caused by the wind in shallow water.

Bhaskara Rao and Mazumdar (1966b) developed a technique for forecasting storm waves. This technique is based on the evaluation of relative values of different factors that cause storm waves viz., i) the astronomical tide, ii) the inverted barometer effect, iii) the effect of piling up of sea water against the coast by onshore winds of the cyclonic storm, iv) the effect of individual waves and v) the swell of the fore-runners. A few documented cases of past storm waves have been tested on the basis of this technique.

Janardhan (1967) estimated surge heights at Sagar Island on the basis of static wind setup, assuming a balance between wind stress and surface slope, and came out with some success. The mean percentage deviation of the computed values from the observed, for six instances studied, was found to be only 10% which suggests that the wind pile-up effect explains nearly 90% of the variation from the predicted tidal height.

All other investigations pertain to the simulation of the numerical models of surges in the northern part of Bay of Bengal. The early effors made on these lines in the Bay of Bengal were those of Das (1972) and Flierl and Robinson (1972).

Das (1972), in his prediction model, idealized the storm using predictive equations for pressure and radial wind speeds, and assumed a uniform core speed for it. This was fed to equations of motion and continuity to derive the movement of water mass. Frictional force of the wind was considered to be the driving force and bottom stress components were evaluated assuming a steady Ekman Spiral. For such a system, frictional stress is proportional to the total transport of water. His computations showed a maximum surge of 2 to 3 m at a distance of 40 km southwest of Chittagong. Though the computed amplitude is higher than the observed, the author pointed out that considering the uncertainties in computing the frictional stress and ambiguity in tide gauge data readings, the computed amplitudes were of the right order of magnitude.

Flierl and Robinson (1972) constructed graphs which could estimate the maximum surge height as a function of maximum wind speed, the speed of translation along the track, the orientation of the track and the distance of landfall from Chittagong. Using this graph, the authors studied the characteristics of the surge by examining the sea level response to a number of cyclones moving across the Bay along varied paths. According to them, the geometrical and geographical factors such as the curvature of the coastline, the vast expanse of shallow water over the continental shelf as well as the large astronomical tide range are found to amplify the surge heights. They have considered in detail the relative importance of these factors with reference to the Bay of Bengal.

Das _et al._ (1974) presented the results of their computations for three coastal areas of the Bay of Bengal. While the earlier work (Das, 1972) was concerned with the northeastern sector, this study was extended to the northern and northwestern sectors of the Bay of Bengal. The aim of this paper was to compute the surge generated by a model storm which moves with a uniform speed. It also assumed that the idealized storm has concentric isobars and an inner ring of strong winds, but no cross isobaric flow. By changing the parameters of the model they related the surge to the intensity and speed of propagation of the storm at the time of landfall. The important conclusions of this paper are i) the contribution of storm speed is small compared to the effect of storm intensity on the peak surge at landfall, ii) the maximum surge is generally observed about two hours after landfall, iii) the total sea level elevation depends on the phase of the tidal cycle, iv) the maximum elevation is recorded earlier if the surge coincides with the time of high water and v) superposition of the surge on the astronomical tide leads to an overestimate of sea level elevation by 0.8 to 1.1 m.

Ghosh (1977) has given a scheme of prediction of storm surges on the east coast of India on the lines of the SPLASH model of Jelesnianski (1972). Using this model, nomograms were prepared which accommodate fixed values of pressure drop, radius of maximum wind, vector storm motion and offshore bathymetry. Two sets of nomograms were prepared - one for the northeast coast of India where the continental shelf is extremely shallow and the other for the rest of the east coast where the continental shelf is, in general, steep. By means of these nomograms he has also suggested a method to estimate the total sea level elevation along a coastal stretch for northeast coast of India.

Johns and Ali (1980) developed a numerical model for the simulation of surges, generated by a tropical cyclone in the Bay of Bengal. The area of study covers the region from about 11°N to 22°N in the northeastern sector of the Bay. The surge, developed by a cyclone that originates in the sea around the Andamans can be forecasted with this model in this area much before the landfall on the coast of Bangladesh. This model is nonlinear and also helps to determine the interactive effect between the surge and astronomical tide.

Johns et al. (1981) have simulated the surge, generated by the 1977 Andhra Cyclone by the use of three different numerical models. In each of these, they simulated the effect of three days of wind stress forcing before landfall of the cyclone on the Andhra Coast. The area under investigation in the first model includes the entire Bay of Bengal, north of 6^{o}N and utilizes a curvilinear boundary treatment to represent the coast line. The second one is a coastal zone model extending along the east coast of India utilizing as before a curvilinear treatment of the coast line. In the third model which again covers the entire Bay of Bengal, the coast line is represented by a conventional stepwise procedure. Using the available data of this cyclone, the authors compared the predicted rise in the sea surface derived from each of the models with the estimates that were reported after the disastrous event in that region. It was shown that each of the three models produced a qualitatively similar surge response. The slight quantitative differences were explained in terms of the different boundary treatments. The predicted surge above the mean sea level compared well with the estimated rise of 5 m for that region.

Dube et al. (1981) performed experiments with a numerical storm surge prediction model to simulate the surge generated by the 1977 Andhra Cyclone by using various formulae (Jelesnianski, 1965,1972; Das et al. 1974; Johns and Ali, 1980) that represent the wind distribution in the cyclone. These authors found that most of these formulations do not give a good estimate of the observed wind associated with the Andhra Cyclone. Accordingly, the computed sea surface elevations along the east coast of India do not agree with the observed ones, so a new formulation, for the computation of the surface winds, was suggested. They concluded that the winds in the central region of the storm do not contribute significantly, whereas the far distant winds play an important role in the generation of surges along the coast.

Dube et al. (1982) also studied the effect of coastal geometry on the location of peak surge. Using a forcing wind stress distribution, representative of a severe cyclonic storm with a core of hurricane winds, a comparison of the results is made for different tracks of a cyclone striking at three different places which have three different types of coastal configurations, along the east coast of India between Nagapatnam and Paradip, viz., a straight line coast, a crescent coast and a slant coast. The results of the experiment showed that the location of the highest surge is sensitive to the coastal geometry and the direction of motion of the storm relative to the coast.

Johns *et al.* (1982) have given a finite difference method for the treatment of continuously deforming lateral fluid boundary in the computer simulations of the systems involving the shallow water equations. The technique was described in the context of a numerical storm surge model for the east coast of India. Using a forcing wind stress distribution, representative of the 1977 Andhra Cyclone, a comparison was made between simulations using models both with and without inland intrusion of water along the Andhra Coast.

From the foregoing, it is obvious that the various numerical models have given different types of results which have their limitations. The nomograms prepared by Ghosh (1977) appear to have more practical utility. According to Dube *et al.* (1981), the winds in the central region of the storm do not contribute significantly whereas distant winds play an important role in the generation of surges along the coast. Perhaps, it is because of such a condition that the nomograms of Ghosh (1977) have a limited reliability as he considered the fixed values of pressure drop and the radius of maximum wind. To overcome this difficulty it is proposed to present the nomograms giving the surge generated when cyclones of various intensities and dimensions cross the coast at different orientations. The values of the surges thus obtained, are compared with the case studies and a calibration curve is prepared with the aim of incorporating corrections for the assumptions involved in the preparation of nomograms.

1.1.5. Objectives of the study:

The main objectives of the present investigation are

i) to compute the surge generated by four representative cyclones that crossed the coast at vulnerable points and compare them with the reported values

ii) to prepare the nomograms by computing the surge developed by the varying pressure drops of a model cyclone with different radii and different angles of incidence for the purpose of prediction

iii) to study the mean sea level variations

iv) to compare the computed surges obtained from the nomograms with those of the case studies for arriving at a correction factor based on which a valid prediction can be made for an impinging cyclone of known dimensions and orientation.

v) to compare and contrast the results of the present investigation with those of earlier works.

SECTION - II

SEA LEVEL VARIATIONS

By definition, 'sea level' is the height of the boundary between sea and air, measured relative to a fixed point (bench mark) on land and hence is susceptible to changes with time. The term 'mean sea level' is also in use and the raw material for this is a continuous record of heights of tide referred to a stable bench mark. The determination of mean sea level is achieved by various methods, the simplest being a direct numerical averaging, usually applied to hourly heights of tide. More sophisticated numerical methods frequently used are by employing filters which discriminate different species of tide. The best known numerical filters probably are those of Doodson (1928) and Groves (1955). Integration of a tide chart by planimeter is also used for the determination of mean sea level. In order to avoid the labour, involved in the actual determination of mean sea level, 'mean tide level' is sometimes computed instead of mean sea level, which is defined as the average of the observed high and low waters. For all practical purposes, however, mean tide level must be considered a poor substitute for mean sea level because the short period tidal contributions in mean tide level are not adequately eliminated.

The factors that contribute to the variation of mean sea level are: i) the meteorological and hydrographic parameters, ii) the variation in the distance of the tidal bench mark from the centre of the earth also known as the isostatic adjustment of the earth's crust (secular changes), iii) the changes in the total water balance of the seas and oceans (eustatic change), and iv) the effect of long period tides (Chandler motion and nodal cycle).

1.2.1. <u>The meteorological and hydrographic contribution to mean sea level:</u>

The most important meteorological and hydrographic parameters that contribute to sea level variation are pressure, winds, evaporation and precipitation, and water density and currents.

Changes in atmospheric pressure influence the sea level which, atleast theoretically, reacts like a reverse barometer. In a stationary case the sea surface should be depressed by 1 cm for every rise in atmospheric pressure of 1 mb. While this theoretical statical law is invariably assumed as a sufficiently close approximation to the real situation, a conclusive confirmation is still lacking. Doodson (1924) has shown that for day to day variations in sea level, the theoretical statical coefficient is approached to within 15% for British Waters. But in real practice, the individual effects of wind and pressure cannot be obtained and so a method was proposed by Witting (1918) for the study of Baltic Sea level. Doodson (1960) has used the technique to study the variations in annual mean sea level in English Channel by taking three stations roughly forming an equilateral triangle and correlated the variation of annual mean sea level with the annual mean air pressure for the data of 27 years. Lisitzin (1961) prepared the chart showing the distribution of the average sea level differences caused by atmospheric pressure and found that the contribution of pressure effect on sea level is of the order of 30 cm.

The process of evaporation involves a decrease of the total water in the oceans and seas, while an increase is connected with precipitation and river discharge. Since the total water budget of the oceans remains almost constant due to the concept of hydrologic cycle, the three factors mentioned above must be in balance and so no significant contribution to long term variations in sea level is possible.

The variation of ocean temperature leads to the variation of density and gives rise to steric changes. Steric changes are defined as those resulting from changes in density of a water column without change in its mass. According to Fairbridge (1962), if the temperature of the entire sea water column upto the sea bottom were raised by $1^{o}C$, the sea level would rise by 60 cm. However, according to Fairbridge and Krebs (1962) the variation of mean ocean temperature is very slow and is of the order of thousands of years. But there may be a seasonal rise and fall of shelf and estuary water temperatures. It may come to $10^{o}C$ or so over a 10 m depth range and the consequent steric variation would be a little more than 1.6 cm.

To understand the effect of geostrophic currents on sea level, it is a common practice to prepare the sea surface topography using the concepts of dynamic height and level of no motion with the help of the equation

$$Z = \frac{1}{g} \int_{p_a}^{p_o} \delta \, dp$$

where Z is the change in steric level, p_a the atmospheric pressure, p_o the pressure at the reference level and the specific volume anomaly in the pressure interval dp.

1.2.2. Secular changes:

One of the main causes of movements in the fixed reference mark from the centre of the earth is the isostatic adjustment of the earth's crust. The vertical movements of the earth's crust are attributed to the presence of ice masses which covered large parts of the earth during the Glacial period and their subsequent disappearance. The ice masses exert tremendous pressure causing the earth's crust to subside, but the crust starts to regain its original position as soon as the ice load begins to reduce. This process continues even after the disappearance of the ice. This apparent variation of sea level due to the vertical movement of the crust is known as secular change. The question of land uplift and land subsidence has been studied for a number of coastal regions. Some of these studies are of geological character and do not come within the scope of the present study.

At present it is only in areas where the continental ice sheet of Pleistocene time melted and removed its weight from the land that sea level shows a rapid and persistent local change. In the northern Baltic Sea and the Gulf of Bothnia the deglaciated land is rising at a rate of approximately one metre per century

(Lisitzin, 1972). Similarly, the basins of the Great Lakes of North America are observed to be tilting toward the south as a consequence of the elastic and plastic recoil of the continent, which was relieved some 12,000 years ago, of the massive load of the continental ice sheet. The consequence of land uplift in these areas appears to cause the lowering of the sea level. The approximate value for the decrease in sea level for the entire Baltic Sea amounts to 135 cm per century (Lisitzin, 1974). In most other parts of the world, the average sea level is probably rising slowly because of general melting of land-born ice.

1.2.3. Eustatic changes

The eustatic changes in mean sea level are generally ascribed to a large group of phenomena such as folding of the seabed, sedimentation covering the sea floor and melting or formation of continental ice masses. In other words, eustatic changes are due to changes in the volume of the basin. The first of the phenomena mentioned above is more or less sporadic in character, while the second occurs continuously. However, an estimate of the effect of these factors upon changes in mean sea level is hardly feasible at the present time. But it must be kept in mind that estimates of the changes in the volume of the Antarctic ice sheet have a decisive effect upon the sea level, since these ice masses are the largest in the world. It has been estimated that if the total Antarctic ice sheet were melted, the sea level in the oceans and seas would rise by approximately 100 m.

A numerical estimate on the eustatic rise in mean sea level was first given by Gutenberg (1941). According to him, the value is 1.1 mm per year. Later, many workers estimated and gave different values (Kuenen, 1950; Doodson et al., 1954). However, it was generally agreed that the contribution of eustatic factor on mean sea level during this century is a little over 1 mm per year (Lisitzin, 1965).

In this connection it is interesting to note that 5 or 10 year moving averages of annual mean sea levels showed a close correlation to sun spot cycles (Fairbridge and Krebs, 1962). Averages of world temperature changes of the last century have been prepared by Mitchell (1963) showing approximately a 1°C rise for the period, 1900-1950. This rise in temperature might have been the cause for the rise in mean sea level of about 50 mm during the period 1900-1950 (Fairbridge, 1961,1963).

Apart from glacio-eustatic changes, two other eustatic changes are recognised: Sedimento-eustatic - due to accumulation of sediment in ocean basins and tectono-eustatic - due to modification in the shape of ocean basins because of tectonic action. Both these changes are related to the dimensions of the basin, while glacio-eustasy related to the volume of the basin.

1.2.4. Chandler motion and nodal cycle

The Chandler effect is understood to mean variations in sea level arising as a result of the variable force which is the consequence of the movement of the instantaneous axis of the earth's rotation. Chandler motion gives rise to the so called pole tide and results in a cycle, affecting the sea level with an average period of 429 days, or approximately 14 months (Baussan, 1951). According to Maximov (1952,1956a,1956b), the distribution of the amplitudes and phases of this 14 month cycle in sea level indicates that the movement of the earth's axis produces an asymmetric circumpolar wave from west to east. The most characteristic feature of this wave is that the phases are opposite in the northern and southern parts of the Atlantic and the Pacific. In other words, a maximum sea level in the

North Atlantic due to the pole tide, coincides with a minimum in the North Pacific and the currents have opposite directions in these oceanic areas. The maximum amplitude in the middle latitudes of the two hemispheres, of this 14 month constituent is of a magnitude of 4 cm. The nodal line runs north of the equator in the Atlantic, but south of it in the Pacific and Indian Oceans (Lisitzin, 1963).

Nodal tide

While the semi-diurnal and diurnal constituents constitute the short period fluctuations of tide, the nodal tide is considered as a long period oscillation. According to Schureman (1941) and Doodson (1921), there are not less than 99 long period, 157 diurnal and 129 semi-diurnal tidal constituents. However, a large number of the long period tidal constituents have no practical significance to the nodal tide, as their relative

The nodal tide brings about changes in the slope of the water surface and these changes have a fairly pronounced influence upon the variations of the intensity of the western boundary currents and, consequently, on the temperature of the water masses in that region (Maximov and Smirnov, 1965).

SECTION - III

DESCRIPTION OF THE AREA

1.3.1. Physical features

The Bay of Bengal is a marginal sea and is essentially a northward extension of the Indian Ocean to the eastern side of the peninsular India. It is demarked between 5°N and 22°N, and 80°E and 96°E. Its borders are the east coast of India on the west, Sundarbans on the north and the Burmese - Malayasian Coast on the east. The depth of the Bay decreases from about 4000 m in the south to less than 2000 m towards north beyond the continental shelf.

The east coast of India extends over four states, West Bengal, Orissa, Andhra Pradesh and Tamilnadu, from north to south. The Tamilnadu Coast is almost straight and parallel to the meridians. The coastline of Andhra Pradesh, while being larger than those of the other three states, has varying curvatures, and is curved like a crescent moon at the mouths of the rivers, the Krishna and the Godavari. The Orissa Coast may be considered straight with a general southwest to northeast orientation, while that of the Gangetic West Bengal is almost parallel to the latitudes.

Major rivers - the Ganges, the Mahanadi, the Krishna and the Godavari flow into the Bay of Bengal. Excessive siltation of their estuaries and the formation of sand bars provide ideal conditions for the propagation of surges. The shelf width varies considerably along the coast (Fig.1 a and b), with a maximum extent of 300 km in the northern region of the Bay and 50 km along the Andhra and Tamilnadu Coasts. Although the shelf width off the Andhra Pradesh Coast is smaller, shallow depths across the mouths of rivers Godavari and Krishna, and the abrupt indentations of the coast at their two mouths provide favourable conditions for surge inundation. Rameswaram in Tamilnadu and Paradip in Orissa are also vulnerable to surges, the former because of the shallow shelf and the latter due to its location nearer to the mouth of the Mahanadi Estuary. Gangetic Coast surges are much more intense than those in the south because of the much wider shelves and presence of islands and deltas that are formed by the Ganges and its tributaries.

The bathymetry of Bay of Bengal is shown in Fig. 2.

Fig.1. DISTANCE OF 10 AND 100 FATHOM DEPTH CONTOURS FROM COAST LINE BETWEEN NAGAPATNAM AND NARSAPUR.

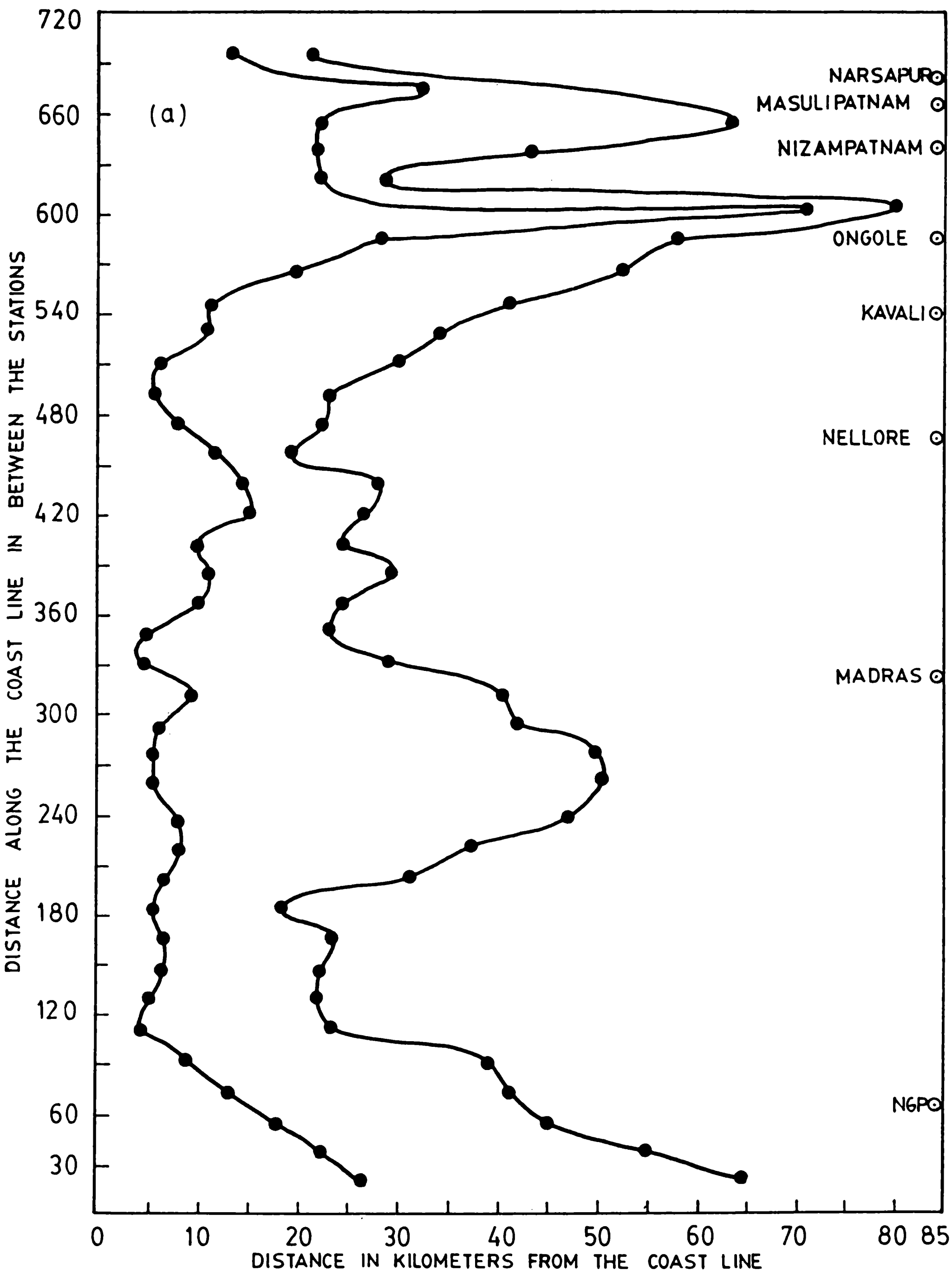


Fig. 1. DISTANCE OF THE 10 AND 100 FATHOM DEPTH CONTOUR FROM THE COAST LINE BETWEEN KAKINADA AND CONTAI.

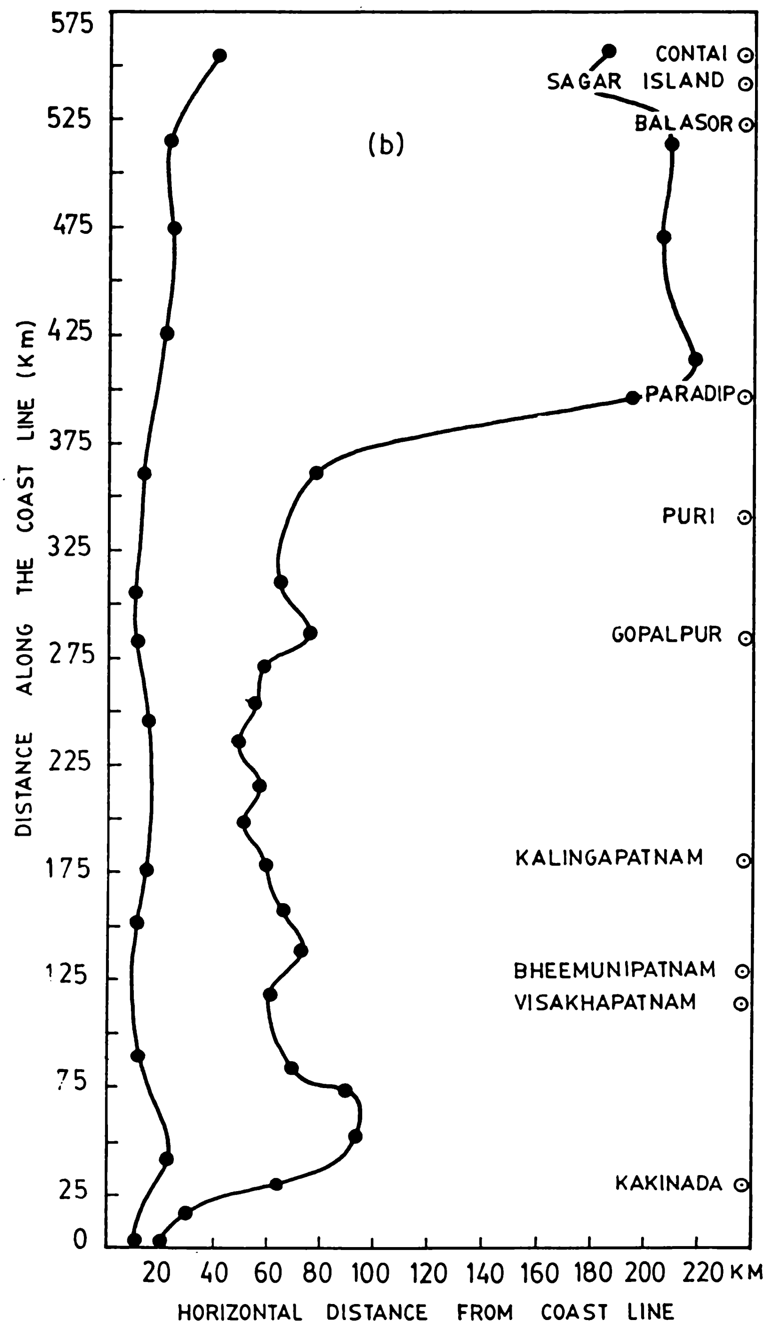
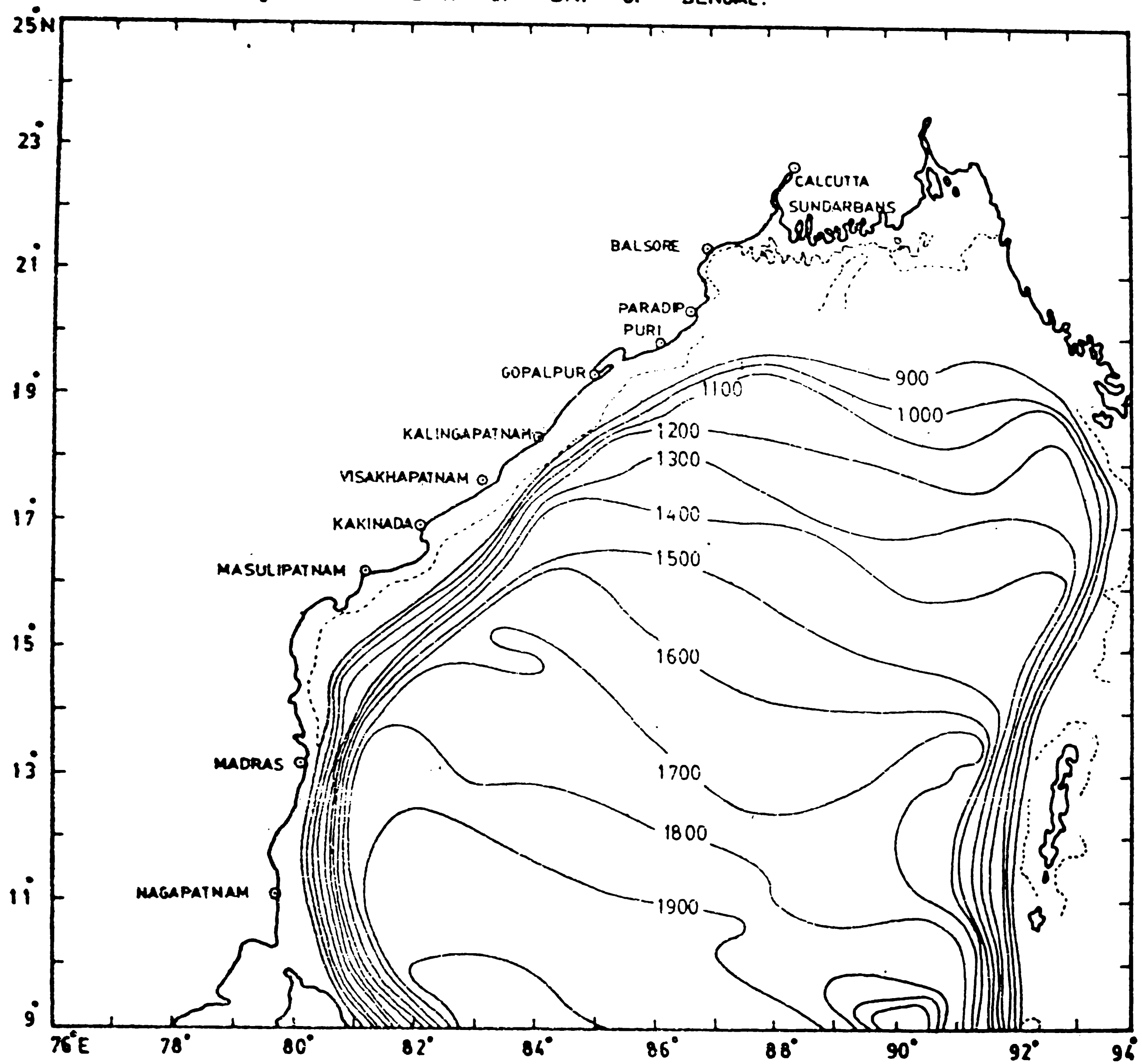


Fig. 2. BATHYMETRY OF BAY OF BENGAL.



1.3.2. Winds and Cyclones

The winds associated with a tropical cyclone in the Indian region, usually, range between 30 and 65 kt. Table 2 gives the nomenclature adopted by the India Meteorological Department for the classification of cyclones, based on wind speed. Table 1 gives the monthly average number of storms and depressions formed in the Bay in different months. It may be seen from the table that out of a total of 13 cyclones in an year, only a third develop to intensities of cyclonic storms and only a ninth develop into severe cyclonic storms. Tables 3a and 3b summarize the information on the total number of storms that affected the various ports on the east coast of India for little over a century between 1877 and 1980. It may be noted that the highest frequency of severe cyclonic storms was recorded at Diamond Harbour (Calcutta) and Paradip. The next highest, which is about half of the above, was recorded at Madras, Masulipatnam and Nagapatnam. The frequency of landfall of the tropical cyclonic storms in each of the coastal districts of the four states along the east coast of India is shown in Fig.3.

Table I. Average number of storms originated in Bay of Bengal and Arabian Sea in each month (source: I.M.D.)

	Jan	Feb	Mar	Apr	May	Jun	July	Aug	Sept	Oct	Nov	Dec	Total
						Bay of Bengal							
All cyclones	0.19	0.04	0.07	0.37	0.80	1.33	1.89	2.07	2.16	1.89	1.46	0.74	13.00
Cyclonic storms	0.06	0.01	0.06	0.26	0.40	0.49	0.54	0.36	0.39	0.76	0.30	0.37	4.49
Severe cyclonic storms	0.01	0.01	0.03	0.10	0.26	0.06	0.10	0.01	0.11	0.27	0.33	0.13	1.43
						Arabian Sea							
All cyclones	0.07	-	0.01	0.10	0.29	0.43	0.16	0.03	0.14	0.51	0.50	0.11	2.36
Cyclonic storms	0.03	-	-	0.07	0.19	0.19	0.04	0.01	0.06	0.24	0.30	0.04	1.17
Severe cyclonic storms	-	-	-	0.06	0.16	0.11	-	-	0.01	0.10	0.23	0.01	0.69

Table II. Range of wind speeds associated with the intensity of tropical cyclones (source: I.M.D.)

Sl. No.	Name	Wind speed range
1.	Low pressure area	upto 16 kt.
2.	Depression	between 17 and 27 kt.
3.	Deep depression	between 28 and 33 kt.
4.	Cyclonic storm	between 34 and 47 kt.
5.	Severe cyclonic storm	between 48 and 63 kt.
6.	Severe cyclonic storm with a core of hurricane winds	64 kt and above

1 kt = 1 knot = 1 nautical mile/hr = 1.85 km/hr

Table IIIa. Number of storms affecting the different ports on the Bay Coast (1877-1980)

Sl. No.	Port	Jan	Feb	Mar	Apr	May	Jun	July	Aug	Sept	Oct	Nov	Dec	Annual
1.	Trincomalee	2	0	2	0	2	0	0	0	0	2	12	12	32
2.	Nagapatnam	5	0	2	1	6	0	0	0	0	5	23	11	53
3.	Madras	3	0	2	3	9	0	0	0	0	15	31	8	71
4.	Masulipatnam	1	0	0	2	7	1	0	1	2	17	17	1	49
5.	Visakhapatnam	1	0	0	0	3	5	1	2	7	18	9	0	46
6.	Paradip	0	0	0	0	7	22	32	28	27	22	12	2	152
7.	Diamond Harbour	0	0	0	0	11	19	31	20	21	16	6	3	127
8.	Chittagong	0	0	0	3	16	8	5	1	2	16	10	6	67
9.	Akyab	0	0	0	5	19	6	2	2	0	9	9	4	56
10.	Sandoway	0	0	0	3	11	0	0	0	1	3	7	1	26

Table IIb. Number of storms severely affecting the different ports on the Bay Coast (1877-1980)

Sl. No.	Port	Jan	Feb	Mar	Apr	May	Jun	July	Aug	Sept	Oct	Nov	Dec	Annual
1.	Trincomalee	0	0	0	0	1	0	0	0	0	0	3	7	11
2.	Nagapatnam	0	0	0	1	3	0	0	0	0	2	9	6	21
3.	Madras	1	0	1	0	5	0	0	0	0	4	14	2	27
4.	Masulipatnam	0	0	0	0	3	0	0	0	1	8	9	1	22
5.	Visakhapatnam	0	0	0	0	0	2	0	0	0	2	2	1	7
6.	Paradip	0	0	0	0	2	8	9	7	8	8	5	0	47
7.	Diamond Harbour	0	0	0	0	6	8	13	8	8	6	2	0	51
8.	Chittagong	0	0	0	0	4	4	1	0	0	8	6	0	23
9.	Akyab	0	0	0	2	4	1	0	1	0	0	5	1	14
10.	Sandoway	0	0	0	2	6	0	0	0	0	3	1	1	13

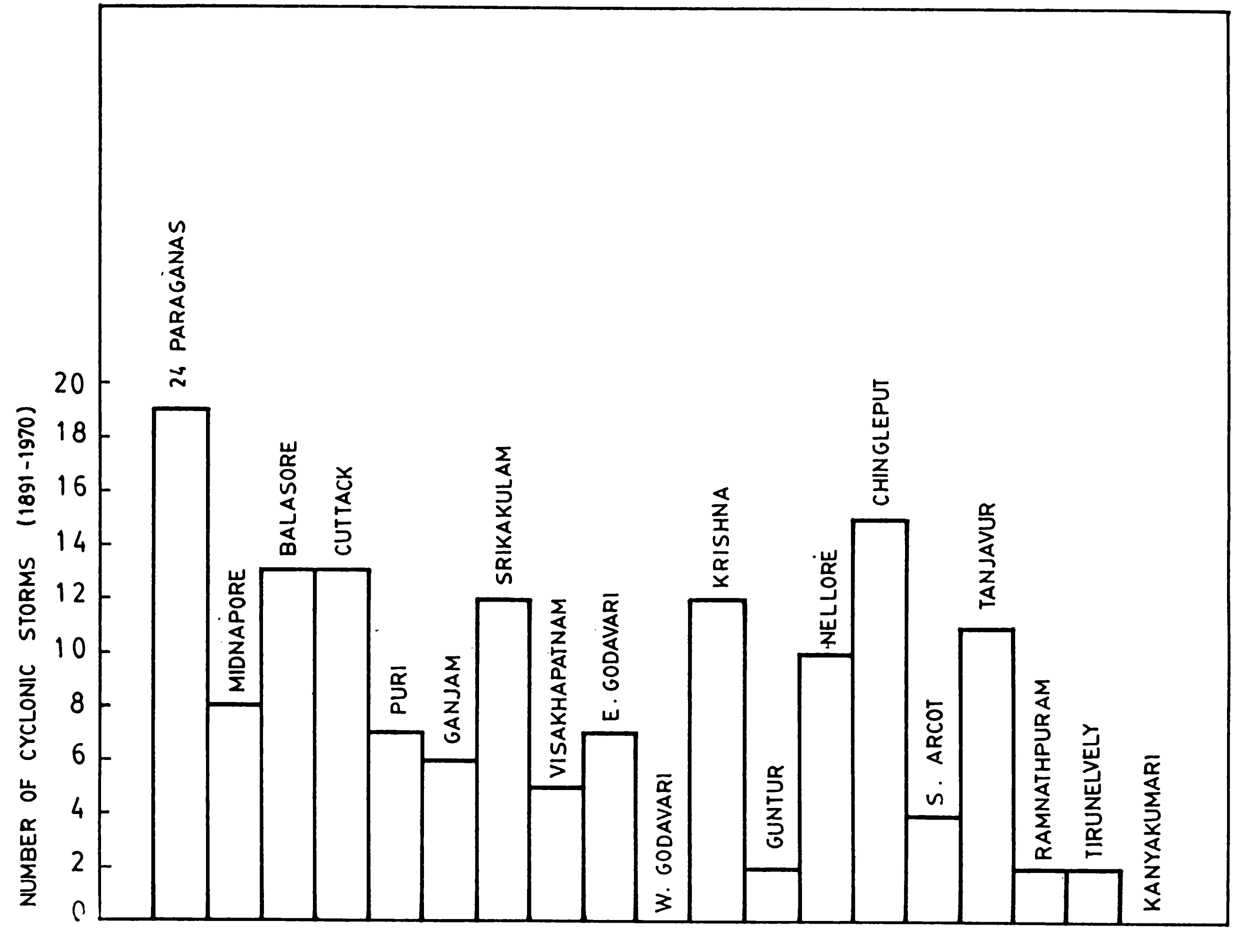


Fig.3. NUMBER OF CYCLONIC STORMS CROSSED EACH DISTRICT ALONG EAST COAST OF INDIA DURING THE PERIOD 1891-1970.(FROM INDIA METEOROLOGICAL Dept).

1.3.3. Annual variation of tropical cyclones in the Bay of Bengal

Tropical storms in the Indian seas originate between 5°N and 22°N. The places of origin shift to north and south depending on the season. The cyclonic storms are rare in the Indian seas during the period January to March. The few that form over the Bay of Bengal originate between 5°N and 8°N. The region of formation shifts to the latitudinal belts 8°N to 13°N in April, 10°N to 20°N in May and 16°N to 21°N in June. During July and August, the storms form between 18°N and 20°N. With the retreat of the southwest monsoon, the origin of the cyclones shifts southward between the latitudinal belts of 15°N and 21°N in September, 8°N and 13°N in October and 6°N and 14°N in November and December.

In April and May, the storms move initially northwest or north and later recurve northeastward. In April, the recurvature takes place over the seas and the storms strike the Arakan Coast. In May, the whole east coast of India, coastal Bangladesh and the Arakan Coast are vulnerable to the incidence of storms. Most of the storms in June, move northwest and cross the north Andhra, Orissa and West Bengal Coasts. In July and August, they move westnorthwest across the Orissa and West Bengal Coasts.

The tendency to recurve begins to increase from September. Most of the storms recurve in the latitudinal belt of 18°N to 21°N in October and between 15°N and 18°N in November. Andhra Pradesh, Orissa and West Bengal are more prone to the incidence of storms in October while Tamilnadu and Andhra Pradesh are more vulnerable in November. Thus, the Andhra Pradesh Coast is vulnerable in October as well as in November. The storms that form in December over the southwest Bay move westnorthwest and cross the Tamilnadu Coast while those that form further east move northwest, later recurve to the north and weaken over the sea (Rai Sarcar, 1956,1958; Ananthakrishnan, 1964; Jayaraman et al., 1966).

SECTION - IV

TREATMENT OF THE DATA

1.4.1. Basic theory of the analysis of surges

The height to which the water level near the coast rises consequent to the passage of a storm depends on various factors. Bhaskara Rao and Mazumdar (1966 b) discussed these factors and formulated an equation for the final height (H):

$$H = A + B + P + X + F \quad \ldots\ldots\ldots\ldots \quad (1)$$

Where A is the contribution due to astronomical tides, B is the static rise due to barometric pressure drop commonly known as inverted barometer effect, P is the rise due to piling-up of water against the coast by winds, X is the mean height of the crests of the individual waves superimposed on the general rise of the sea level and F is the effect of the fore-runners.

The effect of waves (X) is important only for seasonal sea level variations but not to the transient surges. The effect of waves on sea level will be of the order of a few centimetres while the transient surges are of the order of few hundred centimetres. The last factor (F), the fore-runner, serves as a warning to the coastal people about an impending surge when the storm is located a few hundred kilometres from the coast. A fore-runner is a slow and gradual change in water level beginning several hours before the arrival of the storm but the elevation of the level is very small compared to the storm surge (Groen and Groves, 1962).

The contribution of the astronomical tide to the surge can be found from the predicted tide tables published by the Survey of India.

The term B, the inverted barometer effect as applied to storm surges is given by the simple hydrostatic relationship:

$$B = - \frac{\Delta P_a}{g} \times 10^3 \qquad (2)$$

Where ΔP_a is the atmospheric pressure deficiency in millibars, g is the acceleration due to gravity and B is in centimetres. Roughly the pressure deficiency in millibars gives the numerical value of B in centimetres. While the storm surge usually elevates the water level by a few metres, the maximum elevation due to the inverted barometer effect is of the order of only a few centimetres. Therefore, the contribution of the inverted barometer effect is negligible.

In equation (1) the final height is thus contributed mainly by the astronomical tide and the piling-up of water against the coast. Therefore, to predict the storm surge, it is enough if the value of P can be estimated.

1.4.2. Piling-up of water due to winds

The actual method of working out the value of P for a given wind velocity is as follows:

The tractive force of the wind that causes the slope of the water surface, $\frac{dz}{dx}$, is determined from the equation (Welander, 1957,1961; Lisitzin, 1974)

$$\frac{dz}{dx} = \frac{\lambda \tau}{g \rho_w h} \qquad (3)$$

Where Z is the elevation of the water surface, X is the horizontal co-ordinate in the direction of the wind, τ is the wind stress, g is the acceleration due to gravity, ρ_w is the density of water, h is the average depth of the basin and λ is the Manning's frictional factor (Bretschneider, 1967). The factor λ varies between the limits 1 and 1.5 depending on the depth and frictional conditions at the sea bottom (Lisitzin, 1974). If the coastal sea basin is very deep it tends to unity and if it is too shallow it tends to the higher value of 1.5.

As the coastal region of the Bay of Bengal is neither too shallow nor too deep, the mean value of 1.25 is assumed in the present computations.

Wind stress τ in the equation (2) is computed using the bulk aerodynamic equation

$$\tau = \rho_a C_D W^2 \qquad (4)$$

Where τ is the wind stress in dynes cm^{-2}, ρ_a is the density of air in gm cm^{-3}, C_D is the drag coefficient and W is the wind speed in cm sec^{-1}. The value of ρ_a is taken as 1.3×10^{-3} gm cm^{-3}. The selection of the drag coefficient which is a dimensionless quantity is dependent on the roughness of the water surface. Various studies have been carried out by different workers for the selection of the drag coefficient and its variation. Munk (1947) assumes a jump in the drag coefficient at Beaufort 4. Hidaka (1958) uses a value 0.8×10^{-3} for winds having speed below 6.6 m sec^{-1} and 2.6×10^{-3} for winds having speed above 6.6 m sec^{-1}. Hellerman (1967) in his updated estimate of wind stress over the world ocean, adopts a value with 40% jump at Beaufort 4. However, many investigators used a constant drag coefficient. Duing (1970) used the value of 2.6×10^{-3} for the Indian Ocean. Weiler and Burling (1967) found no systematic decrease or increase of drag coefficient in the range of wind speed 1.5 to 10.5 m sec^{-1}. In the cyclonic storms originating in the Bay of Bengal the average wind speeds are much in excess of 10.5 m sec^{-1} and the sea condition will be highly turbulent during the passage of the cyclone. Therefore the value of the drag coefficient should be higher than what is usually taken. Hence in the present computation C_D is taken as 2.6×10^{-3}.

Substituting the values of $\rho_a = 1.3 \times 10^{-3}$ gm cm^{-3}, $C_D = 2.6 \times 10^{-3}$, $\lambda = 1.25$, $g = 980$ cm sec^{-2} and $\rho_w = 1.02$ gm cm^{-3}

$$\frac{dz}{dx} = 3.95 \times 10^{-9} \frac{(W^2)}{h} \qquad (5)$$

The value of P in equation (1) is obtained by the summation of the slopes generated by the wind stress. Therefore, P is calculated by stepwise integration from the initial position to the landfall. As the slopes generated in the deep ocean beyond continental shelf are of no significance, only those in the shelf region are computed using the formula given below:

$$P = \Sigma\, dz = 3.95 \times 10^{-9}\, W^2\, \Sigma \frac{dx}{h} \qquad (6)$$

1.4.3. Computation of winds for case studies

One of the important aspects to be considered in computing the piling-up of water against the coast, is the evaluation of wind distribution. The best way is, perhaps, to use the observed values. But it is almost impossible to observe the wind at each and every grid point for short intervals of time during the actual occurrence of a cyclone. Therefore, sometimes it becomes necessary to assume a model cyclone to get the winds at the required grid points. Alternatively, they can be computed from the pressure distribution using either a cyclostrophic or geostrophic or gradient wind equation (Graham and Hudson, 1960). In the present investigation, winds are computed from the pressure distribution that is reported in the India Meteorological Department charts, using gradient wind equation. Whenever possible, the Marine Mercantile Vessel reports are utilized for comparison to check the validity of the wind computations.

1.4.4. Nomograms

For the preparation of nomograms, model cyclones of different dimensions and intensities having concentric isobars, are assumed. The pressure drop here indicates the difference in pressure at the central point and the outermost closed isobar, and the radius of curvature, the normal distance between the centre and the outermost closed isobar. Winds are computed for various pressure drops and radii using the gradient wind equation. Making use of the actual bathymetry off Nagapatnam, Madras, Masulipatnam, Visakhapatnam and Paradip, the wind setup is computed for different orientations of incidence of cyclones at the above mentioned places. Graphs are drawn for varying pressure drop against winds and wind setup for different radii and different angles of incidence.

Almost all the cyclones that originate in the Bay of Bengal cross the Indian Coast at angles of incidence between 80° and 145° from the north. Therefore, the surges are worked out for the angles of incidence of 90° and 135° from the north, in the present study.

1.4.5. Residual Tide

Residual tide is considered as the surge produced by meteorological and oceanographical factors. It has already been concluded that the main contribution to the surge is from the wind. It is proposed to compare the time of occurrence of the mean residual tide and its magnitude with the wind induced surge.

The tide analysis is made by drawing continuous curves using the data available in tide tables of 1976 and 1977 that were published by the Survey of India. The hourly residual tides are obtained by subtracting the predicted tides from the observed. The residuals are then smoothened (to eliminate the minor oscillations, if any, due to the effect of seiches) by the method of 5 hour moving average time section analysis (Rossiter, 1959; Janardhan, 1967).

1.4.6. Sea level variations

In order to understand the seasonal and longterm variations of sea level along the east coast of India, monthly and annual mean heights of sea level variations at Madras, Visakhapatnam, Sagar Island, Kidderpore and Diamond Harbour are studied. The trend of variation of sea level is estimated by taking the five year moving averages.

1.4.7. Sources of data

The hourly observations of wind and pressure, ship observations are reported by Marine Mercantile Vessels, tracks of cyclones as observed by radars and satellite pictures of the cyclones are obtained from the India Meteorological Department. The hourly observations of the tide gauge data at various ports are collected from the Survey of India. Bathymetry of the coast of study is worked out from the bathymetry charts of various sources.

For the study of the sea level variations, the monthly and annual mean heights of sea level, supplied by the Permanent Service for Mean Sea Level, Birkenhead, England, are used.

1.4.8. Limitations

The author is conscious of the limitations involved in the computation of storm surges from the meteorological data which by themselves not fully justified due to imperfect methods of observation. The computation of winds at the grid points from the isobaric distribution using the gradient wind equation involves subjectiveness of individual analysis and extrapolation at the grid points across which the isobars need not necessarily run. To circumvent these limitations efforts are made to the extent possible, to compare the computed values with those reported by the Marine Mercantile Vessels.

The computation of wind using the pressure drop between the central point and the outermost closed isobar (for the nomograms), gives only an average wind speed but not the *in situ* winds. In general, the winds in a cyclone are maximum near the wall cloud which is slightly away from the eye, while it is almost calm at the eye, and relatively lower at the outer periphery of the cyclone. Some workers, as cited earlier, used the maximum wind for the computation of the surge. But Dube *et al*. (1981) reported that the surge generated, is mostly controlled by the overall effective wind rather than the maximum wind. Nevertheless, to make the nomograms more realistic and practicable for prediction, a graph is drawn between the surge values obtained in the case studies and the corresponding values from the nomograms for same pressure drops. This graph would serve as a correction factor for the prediction of surges.

CHAPTER II

CASE STUDIES OF STORM SURGES ALONG THE EAST COAST OF INDIA.

CHAPTER - II

CASE STUDIES OF STORM SURGES ALONG THE EAST COAST OF INDIA

Four cyclones, that crossed the coast at Chirala, Masulipatnam, Kavali in the south and at Contai in the north have been studied following the methods described earlier. Wind induced sea level (surge) was calculated along three directions with reference to the track of the cyclone using the synoptic map of the pressure distribution of the cyclone and bathymetry of the area off the landfall of cyclone. The three directions AB, CD and EF respectively denote the left hand side, along, and the right hand side of the track of the cyclone.

The progress of tide and surge, and their interaction were examined in each case with the help of the tide data. The non-astronomical sea level obtained from the residual sea level of observed and predicted tides was compared with the computed surge.

2.1. Andhra Cyclone of 14-20 November 1977

2.1.1. Life history of the cyclone

A low pressure area moved westward across Malaysia and South Andaman Sea into the extreme southeast Bay of Bengal and concentrated into a deep depression with its centre near 6°N, 92°E on 14th morning. Car Nicobar reported easterly winds of 25 to 35 kt at 0.6 and 0.9 km above mean sea level on that morning and satellite picture showed organised cloud bands associated with this system. Moving westward, the system rapidly intensified into a cyclonic storm by 15th morning and crossed the coast near Nizampatnam on the evening of 19th between 1630 IST and 1730 IST.

2.1.2. Ship and other observations

Ship ELTP (ALRAZIQ) close to the storm centre reported a pressure of 995 mb at 0650 IST on 15th and located the centre of the storm by the radar near 6°N, 90°E at that time. Ships ATJZ and ATMI about 400 km away to the northwest and northeast of the storm centre respectively, reported surface winds of 35 kt on 15th morning which suggested that this storm was covering a large areal extent. The satellite pictures showed that the storm had become severe by the evening of 15th. It was classified as T4/4 on Dvorak scale which gives the maximum wind associated with this storm as 65 kt on 15th.

Continuing to move westward and intensifying, the storm laid near 7°N, 85.5°E on the morning of 16th and had developed a core of hurricane winds. The satellite picture on this morning showed the system in class T5/5 which gives the maximum wind as 90 kt.

On 16th, ship ATJZ reported northerly winds of 40 kt about 150 km to the west of the storm centre at 0530 IST.

Ship Jagatswamini (ATFY) recorded many crucial and very valuable observations close to the storm centre on 17th and transmitted them to the Storm Warning Centre at Madras. She also went right into the eye of the storm on the evening, while requesting some other ships in the neighbourhood (ships ATCA and S2LU) to record the observations of pressure, wind and state of the sea and report. The reports of this ship on 17th, especially, the maximum wind of 105 kt and the minimum pressure of 941 mb between 1530 and 1630 IST at the storm centre (Fig.4) are highly useful. Some of the important and significant data recorded by this ship on 17th are given in the table IV. The satellite imagery received from NOAA on 17th is shown in Fig.5a.

The satellite reports indicated that the storm attained its peak intensity on 18th (Fig.5b). The Joint Typhoon Warning Centre at Gaum estimated the maximum wind associated with this storm as 115 kt on 18th on the basis

PRESSURE (mb) AND WIND (kt) OBSERVATIONS
NEAR 10·8°N, 84·0E ON 17-11-77 BY SHIP JAGATSWAMINI

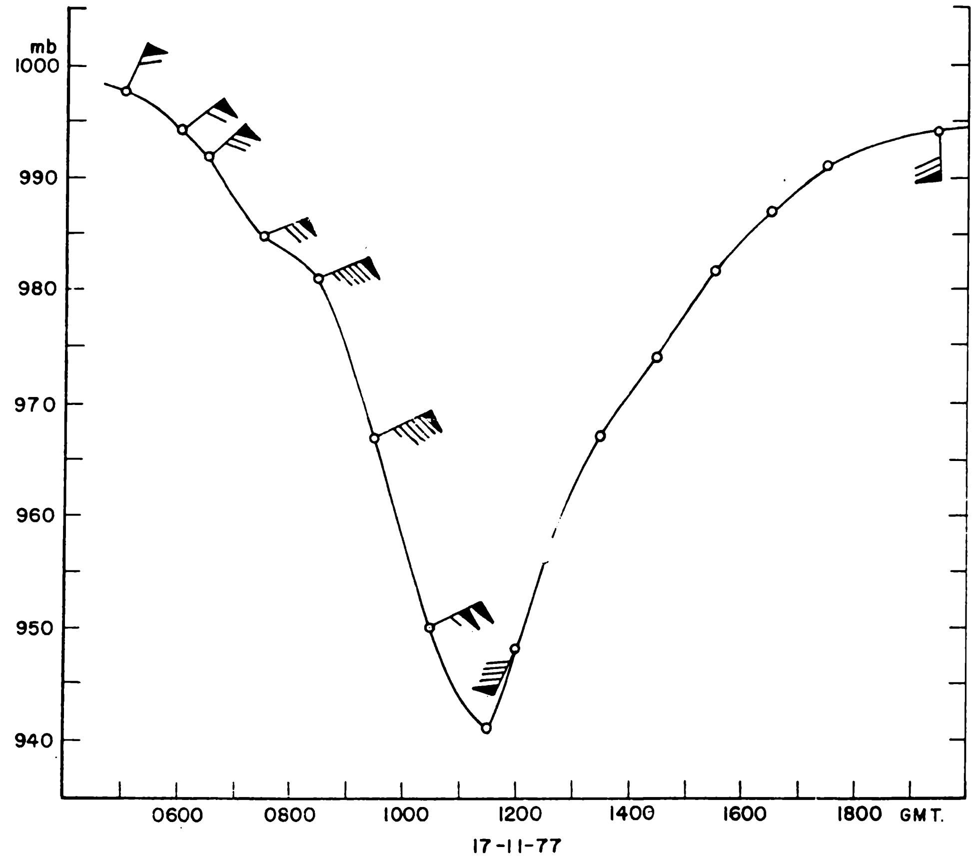
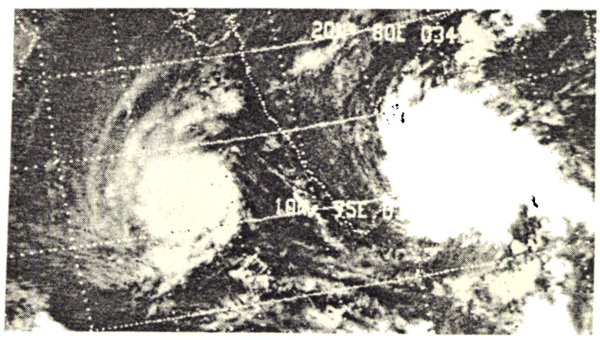


FIG. 4.

Table IV. Observations recorded by ship 'Jagathswamini' (ATFY) on 17 November 1977

Time (IST)	Position	Wind (kt)	Pressure (mb)	Remarks
1030	10.8°N, 83.8°E	NNE/60 kt	997.5	Tendency last one hour 5 mb decreased; very rough sea and confused mountainous swell.
1130	10.8°N, 83.9°E	NE/60 kt	995.5	Tendency last one hour 4 mb pressure drop; very rough sea and mountainous swell.
1415	10.8°N, 84.0°E	NE/E(90 kt)	979.0	Tendency last one hour 7 mb pressure drop; wave height 15-20 m.
1630	10.8°N, 84.1°E	- /90 kt	-	Circular formation of clouds 18 miles in diameter, probably the eye of the cyclone is discernable in radar.
1730	10.7°N, 84.1°E	SSW/90 kt	947.9	Previously passed through calmer and brighter area which might be the eye of the storm;minimum pressure is 941 mb.

Fig 5 a. SATELLITE PICTURE OF ANDHRA CYCLONE RECORDED BY NOAA AT 1600 HRS IST ON 17 NOV. 1977.



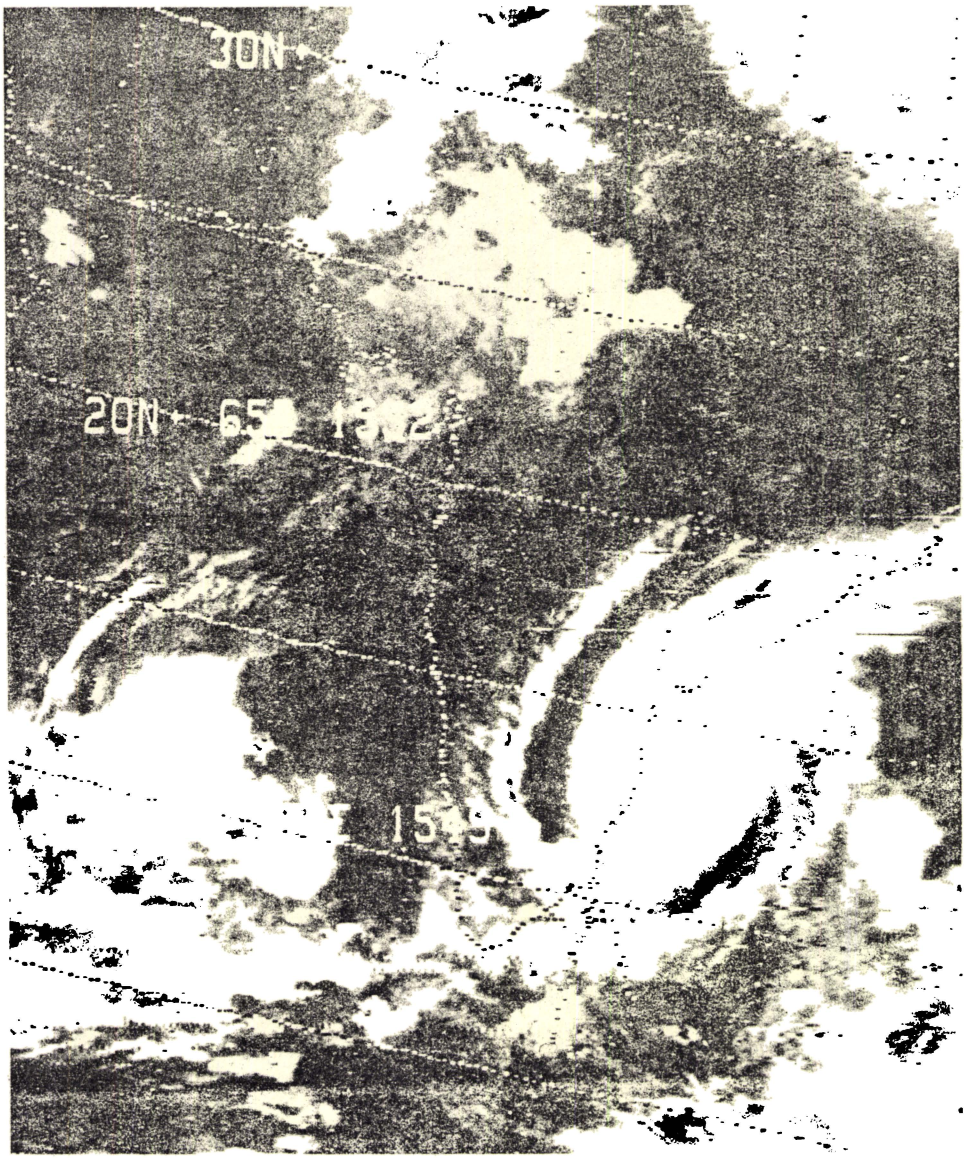


FIG. 5b.Satellite View of Andhra Cyclone at 1550 GMT on 18·11·77 showing clear 'eye'. The Arabian Sea Cyclone is also seen.

of satellite pictures, whereas the National Environmental Satellite Centre at Washington, D.C. reported a maximum wind speed of 140 kt. The radar picture of this storm on 18th is reproduced in Fig.6, which on Dvork scale gives rise to the maximum wind of 125 kt (Pant *et al.*, 1980). The hourly observations of wind and pressure on this day recorded at Nellore are shown in Fig.7.

The storm centre passed a point about 60 km east of Ongole and 50 km west of Masulipatnam between 1430 IST and 1930 IST and was very close to Gannavaram at about 2230 IST on 19th as seen by Fig.8. The hourly values of wind and pressure recorded at Gannavaram, ~~and~~ Ongole and Masulipatnam (Figs. 9a and 9b) show that these stations experienced maximum winds of 70-75, 50-60 and 60-70 kt, and the lowest pressures of 972, 993 and 987 mb respectively on 19th. From the track of the cyclone (Fig.10), it can be inferred that the cyclone crossed the coast between 1630 and 1730 IST.

Taking the maximum wind associated with this storm as 125 kt, the lowest pressure works out to be 943 mb which agrees well with the lowest pressure of 941 mb reported by the ship 'Jagatswamini' at the centre of the storm on 17th. The diameter of the eye of the storm was estimated as 30 km by 'Jagatswamini'.

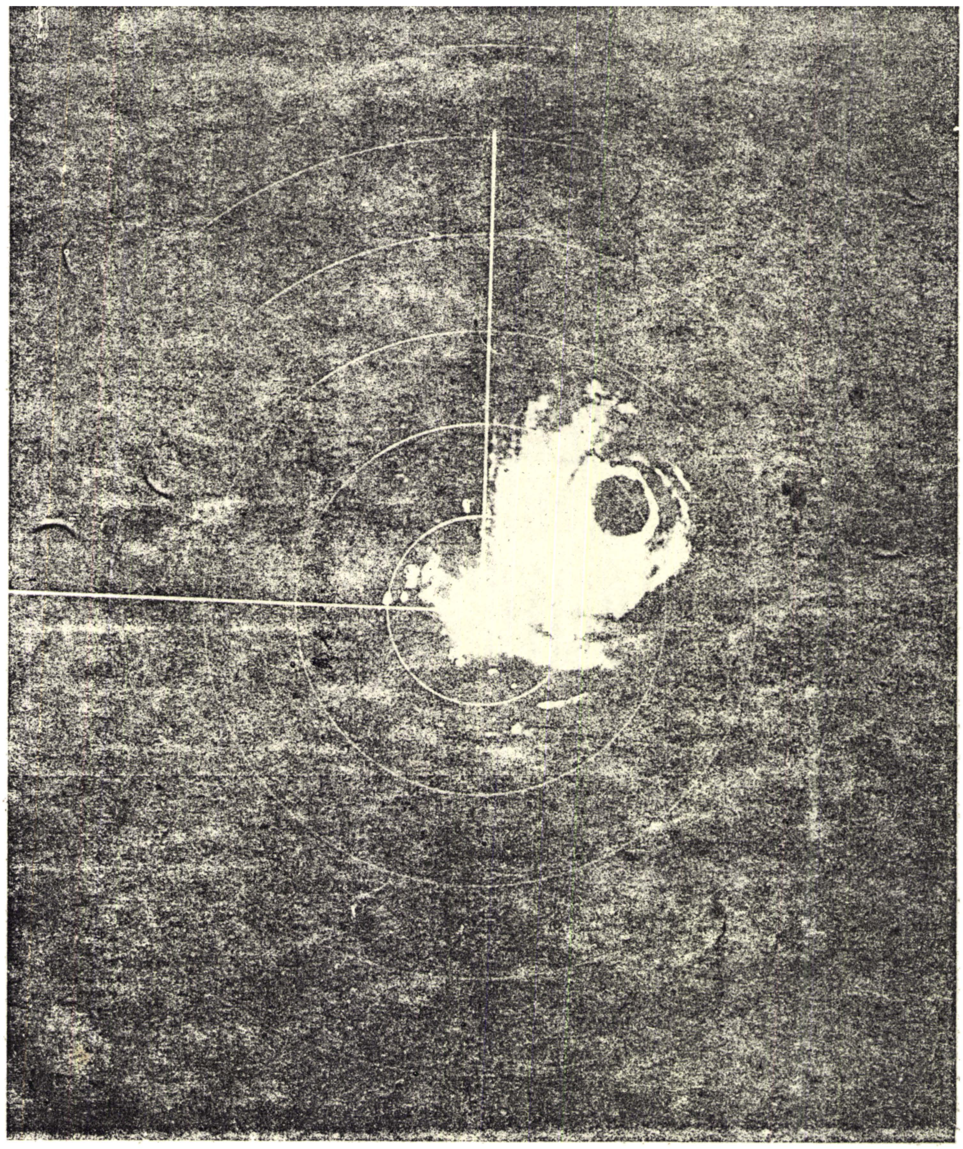


FIG. 6. View of the Andhra Cyclone as seen by Madras Radar at 2116 GMT on 18.11.77 centred about 200 km northeast of Madras.

Fig. 7. HOURLY OBSERVATION OF WIND AND PRESSURE AT NELLORE ON 18-11-1977.

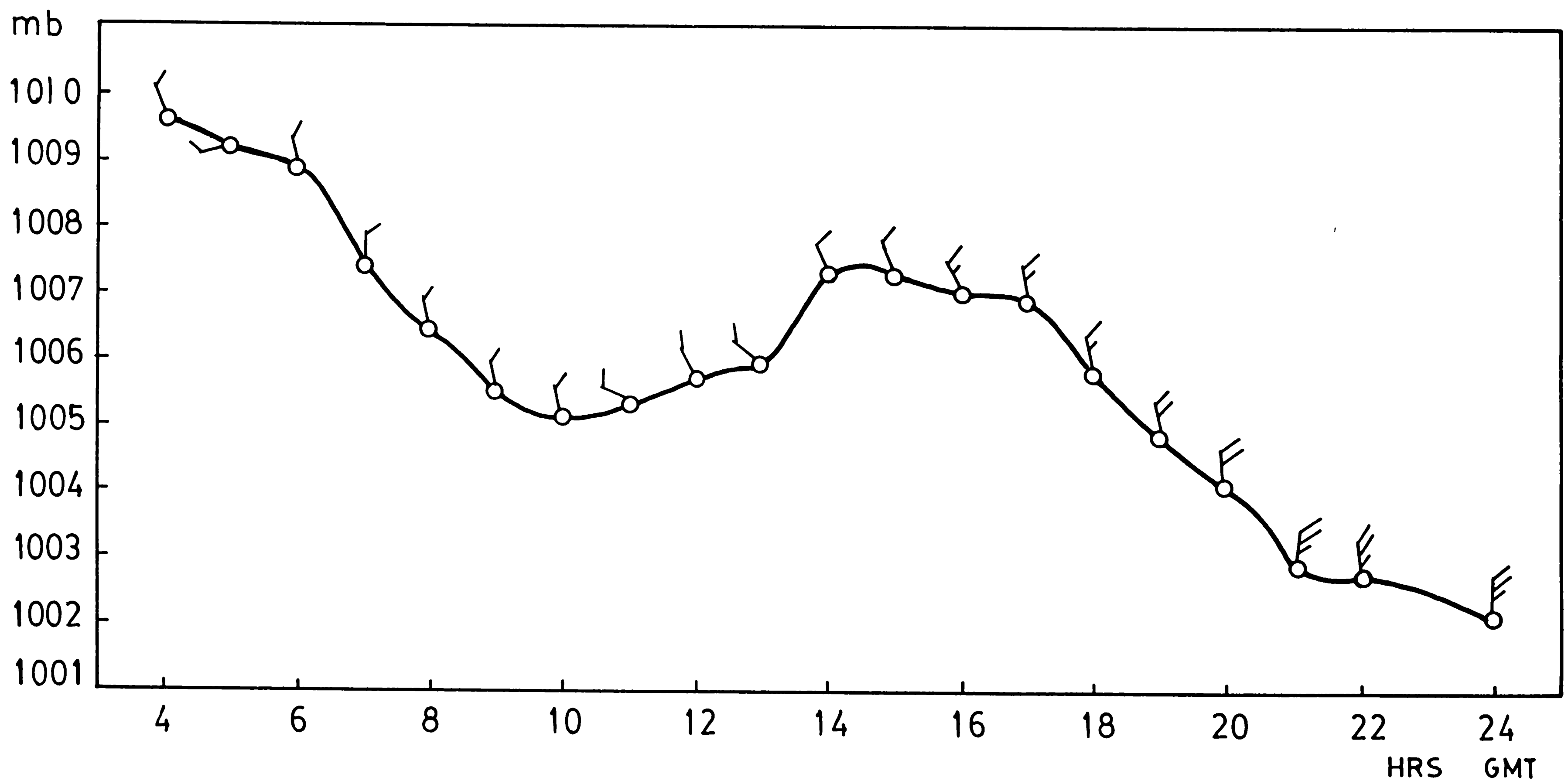


Fig. 7. HOURLY OBSERVATION OF WIND AND PRESSURE AT NELLORE ON 18-11-1977.

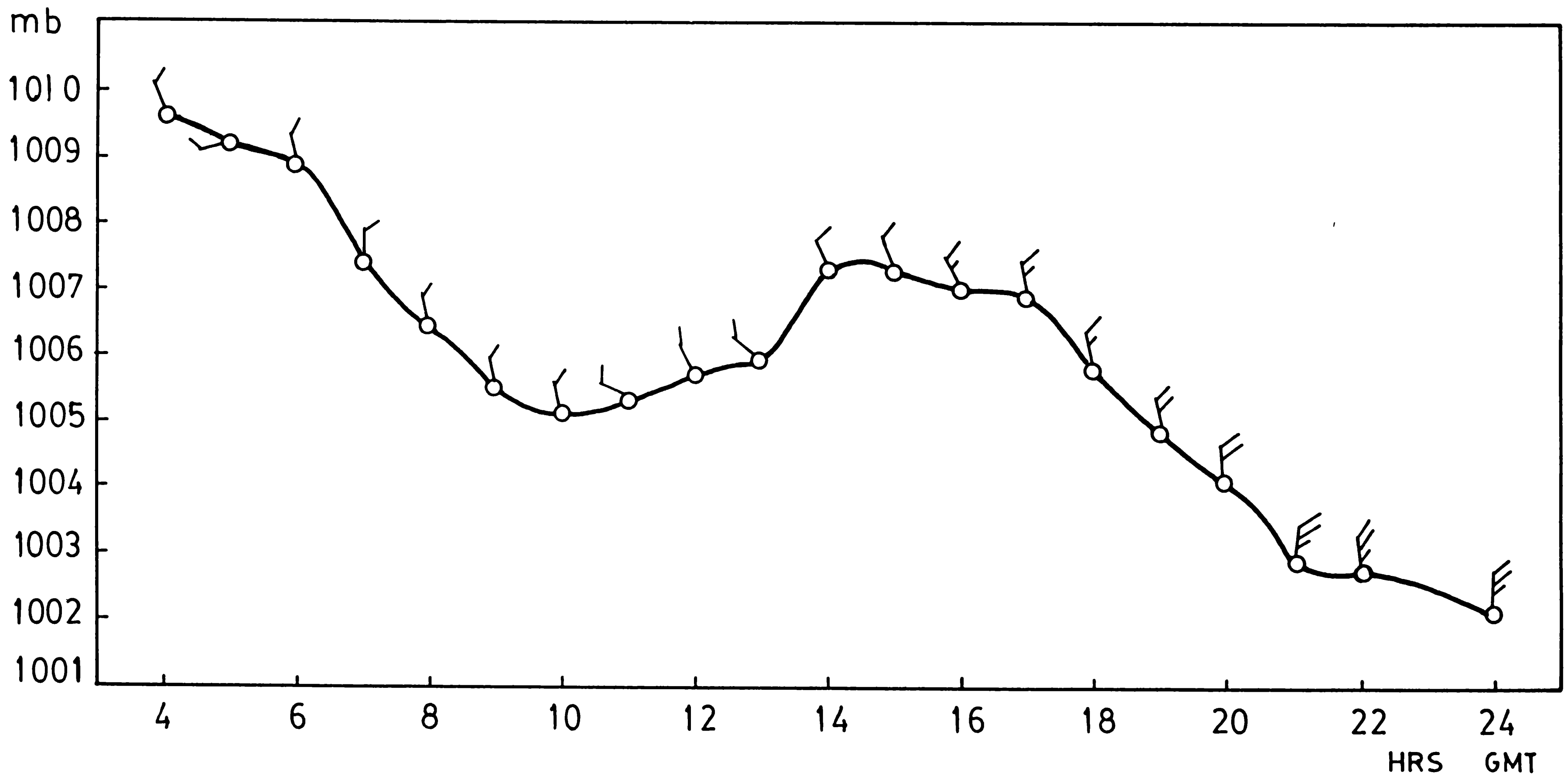


Fig.8. TRACK OF ANDHRA CYCLONE 14-20 NOVEMBER 1977

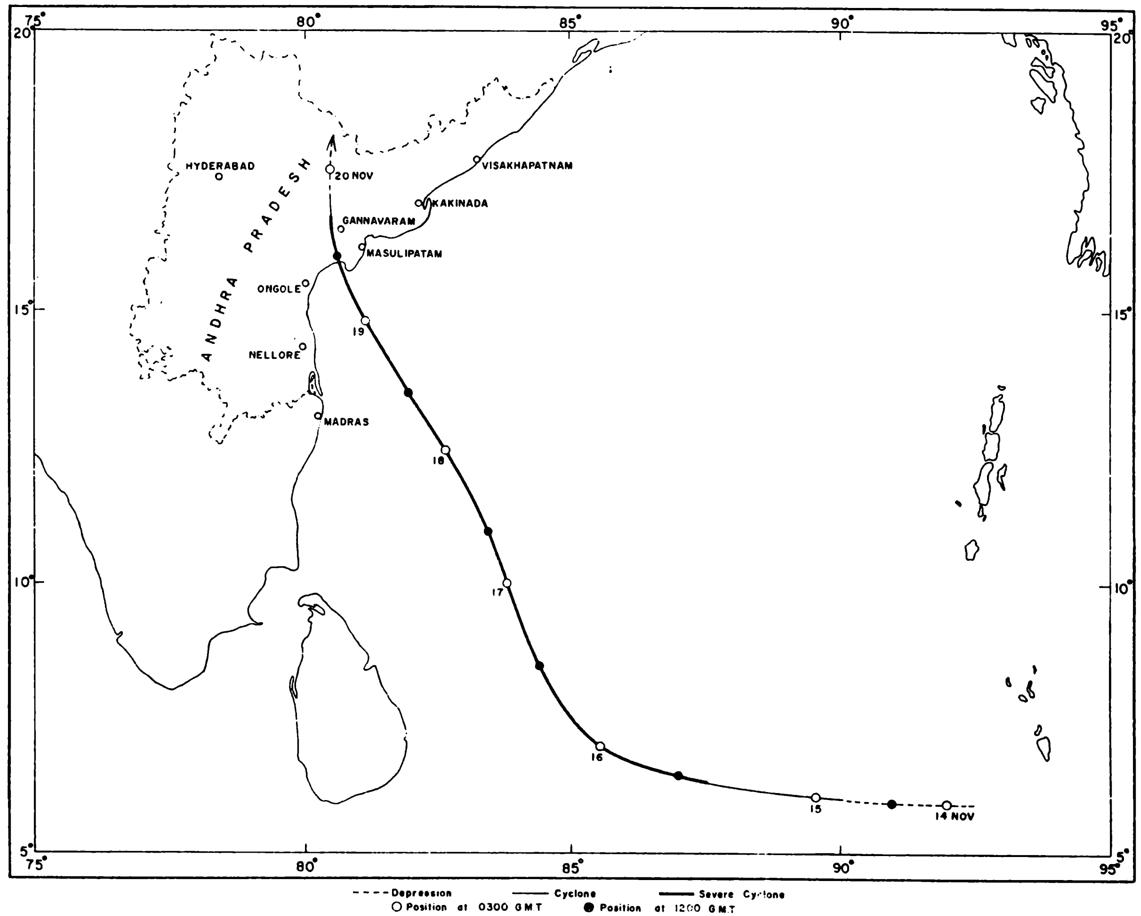


Fig.9. OBSERVATIONS OF WIND (kt) AND PRESSURE (mb) AT GANNAVARAM

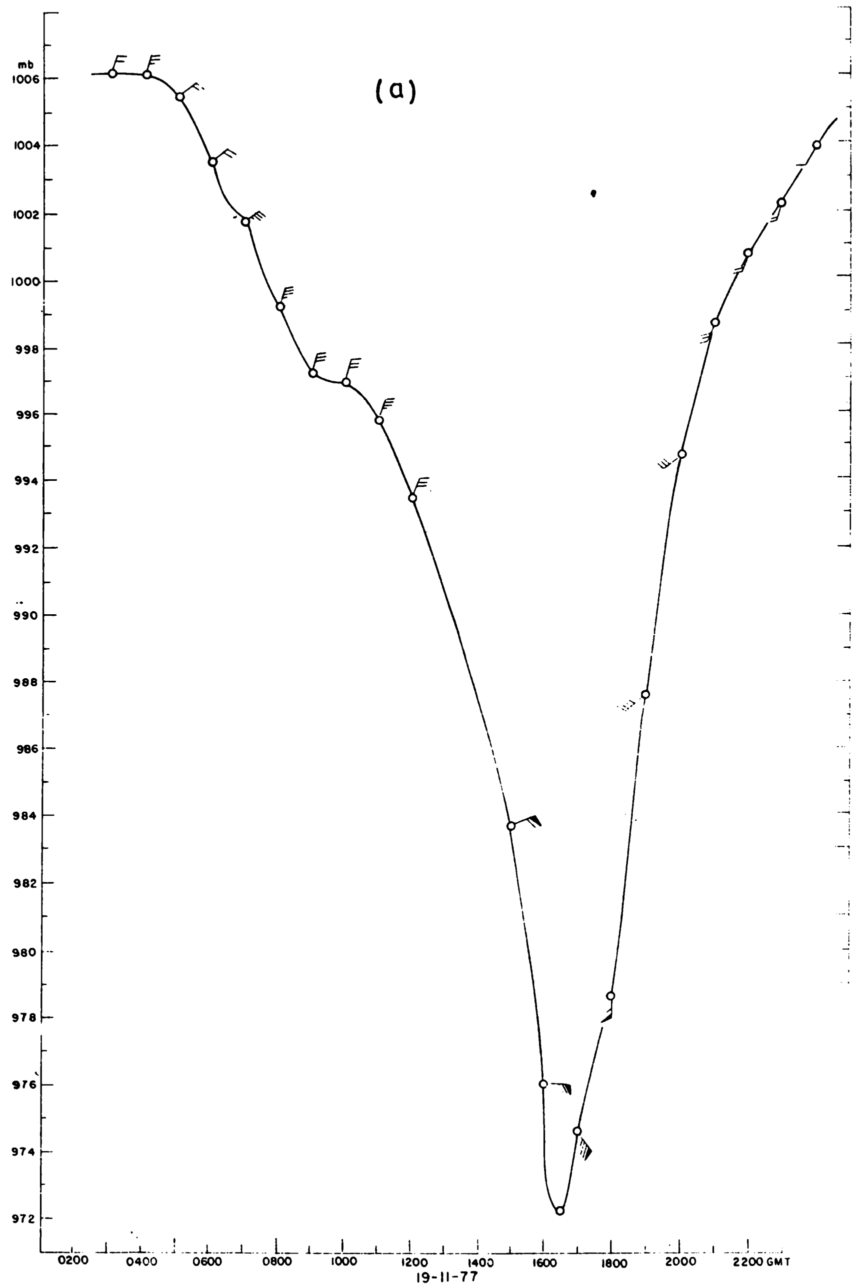


Fig. 9. HOURLY OBSERVATIONS OF WIND AND PRESSURE AT MASULIPATNAM AND ONGOLE ON 19-11-1977.

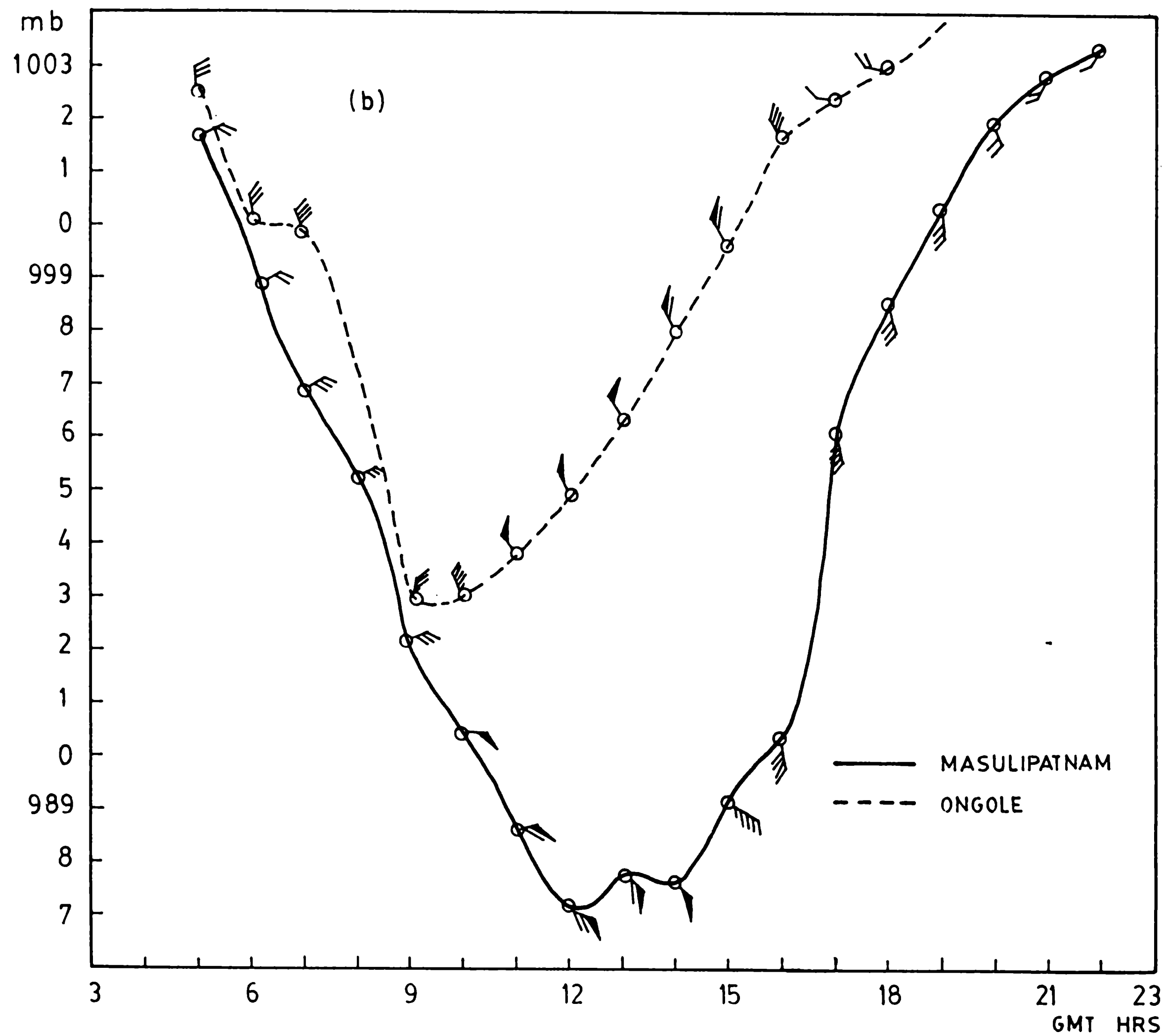
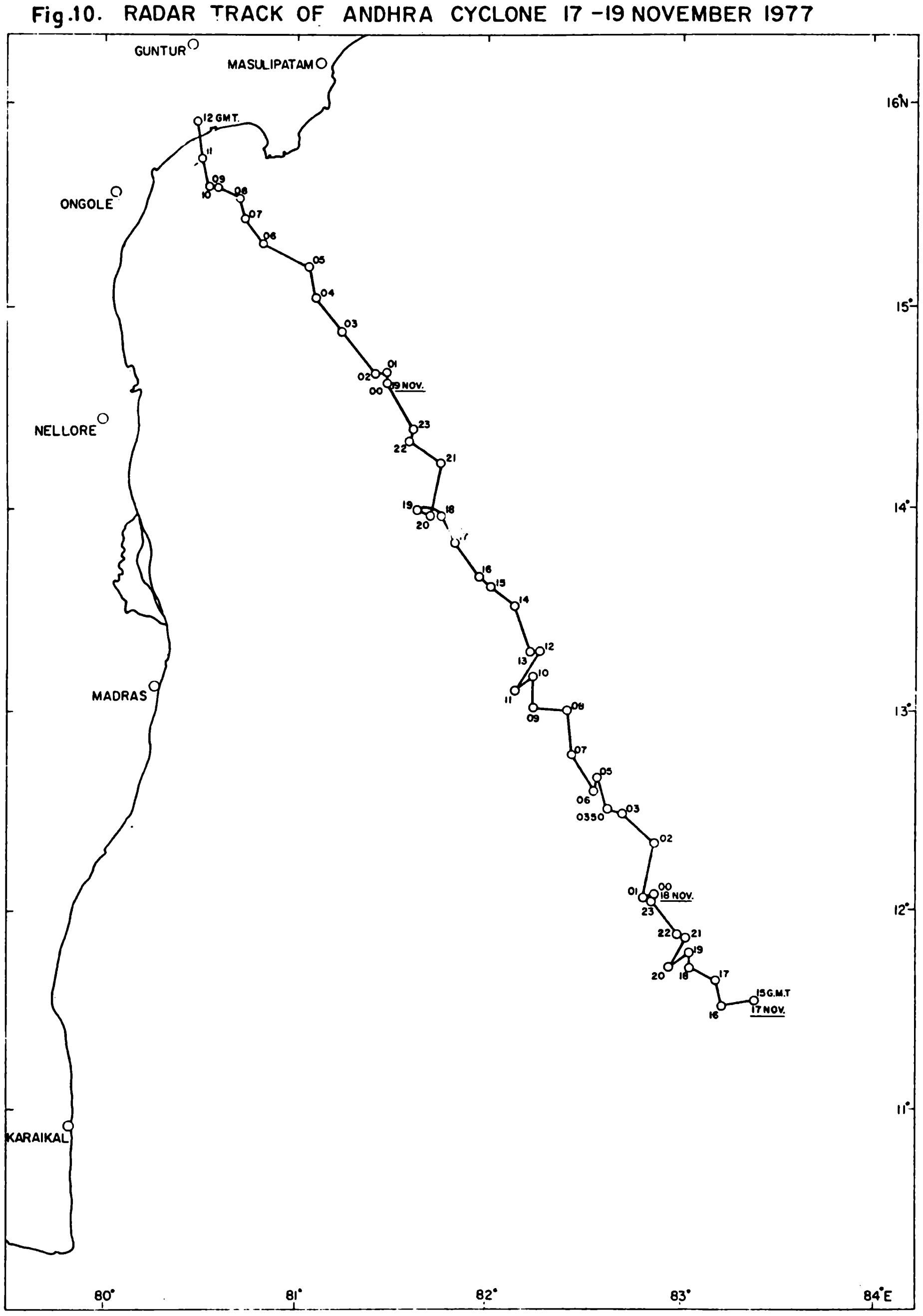


Fig.10. RADAR TRACK OF ANDHRA CYCLONE 17 -19 NOVEMBER 1977



2.1.3. Surface wind structure of the cyclone

With a view to ascertain the distribution of surface winds in different sectors of the cyclone, all available reports of the ships in the field of the cyclone from 16th to 19th were composited by Pant *et al.* (1980) as shown in Fig.11. The winds at different distances from the centre and for different times were plotted in this diagram. The strongest winds exceeding 100 kt occurred in northeast sector unto about 80 km from the centre. Gale force winds have extended 400 to 450 km from the centre in this sector, while they have extended only to 200 to 250 km in the southwest sector. From Fig.11, it may be inferred that the average radius of the cyclone is about 450 km.

Synoptic weather maps from 15th to 20 th at 0530 IST (Figs.12 to 17) are drawn from the special data of IMD and are used to compute winds. A comparison of the computed and observed winds is given in the table V and they agreed very well. These computed winds are used in the grids wherever observed winds are unavailable.

2.1.4. Surge at landfall

The Fig.18 depicts grids and the three paths AB,CD and EF in which the wind setup (surge) is calculated. The bathymetry used is shown in Fig.19.

Fig.11. SURFACE WIND (Kts) IN DIFFERENT SECTORS OF THE CYCLONE.

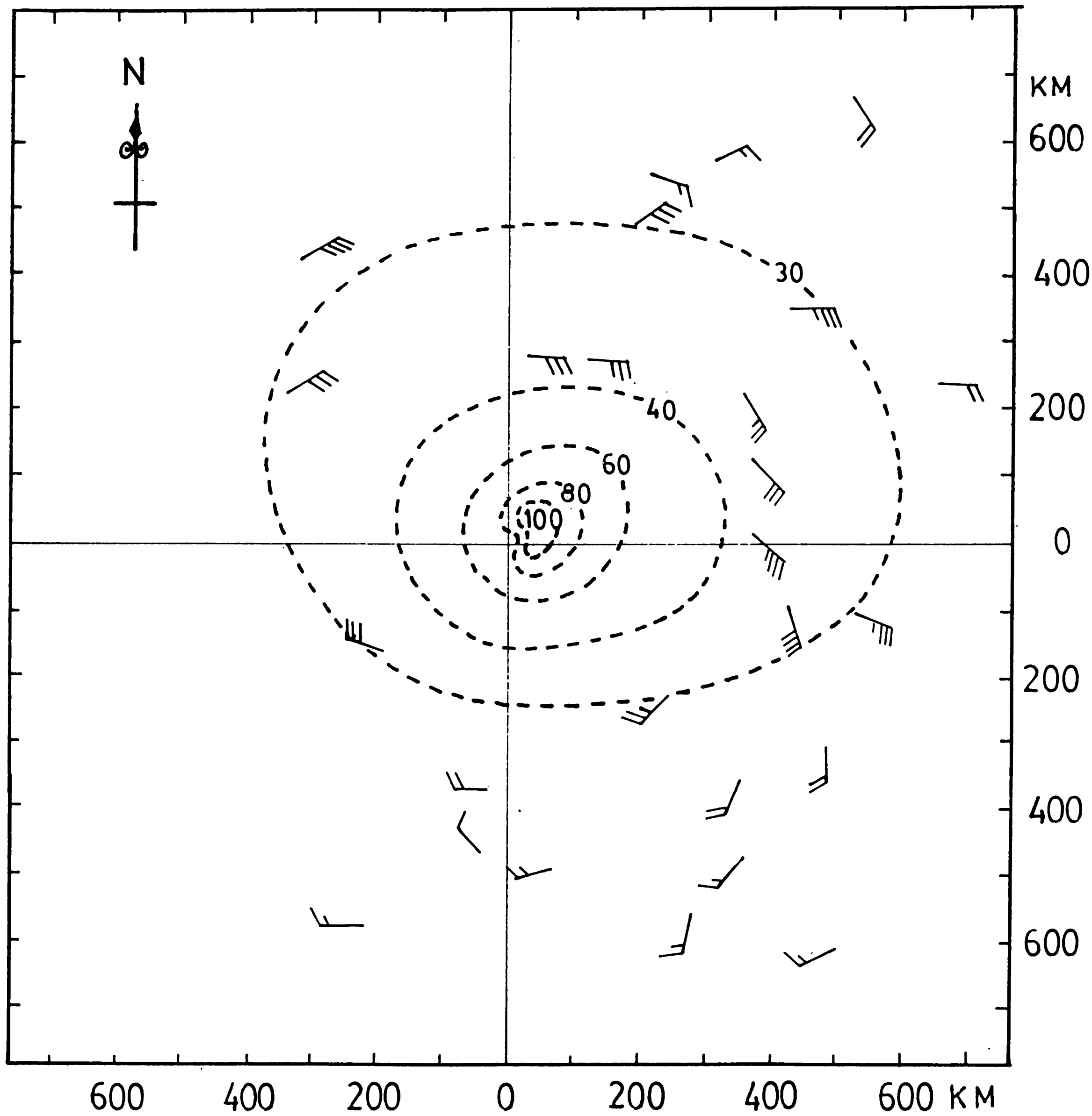


Fig.12. 0530 IST POSITION ON 15 NOV. 1977.

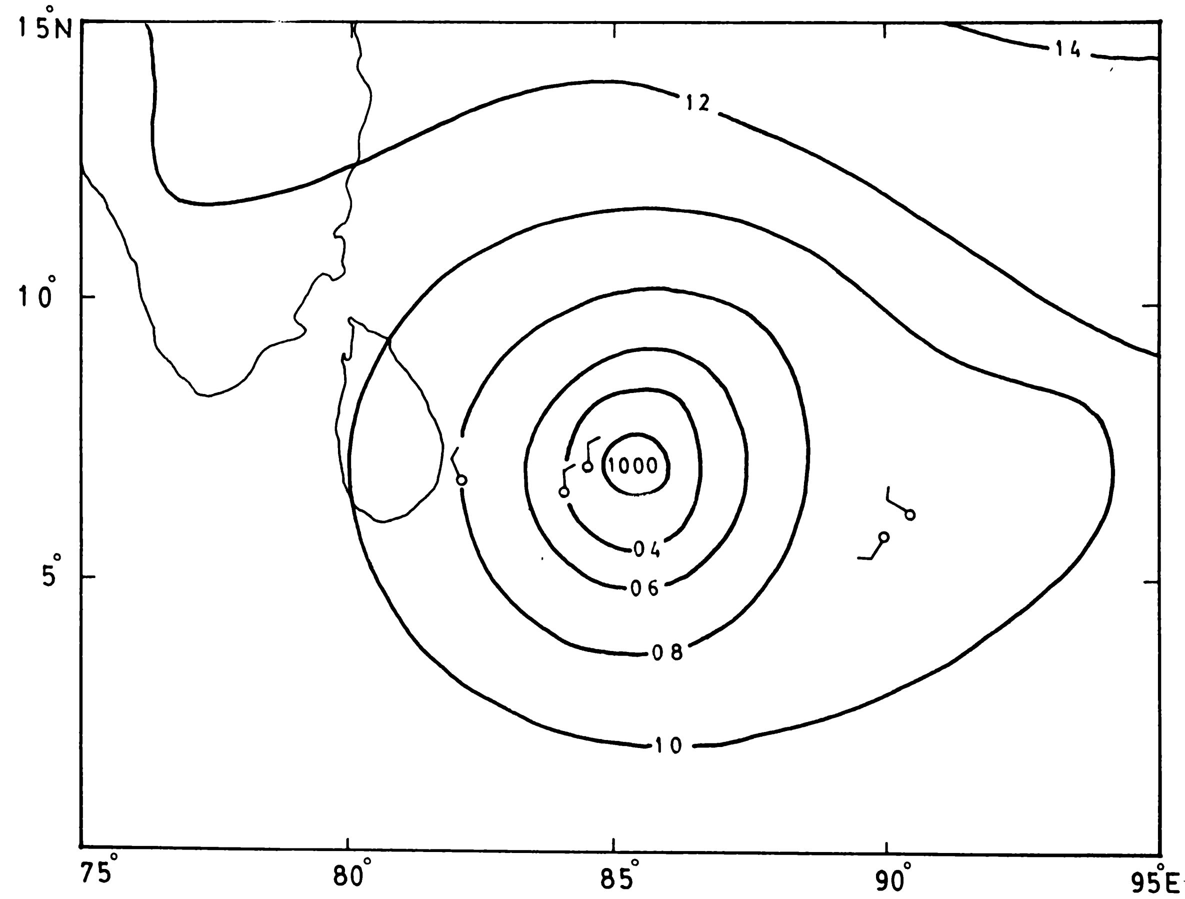


Fig.13.0530 IST POSITION ON 16 NOV. 1977.

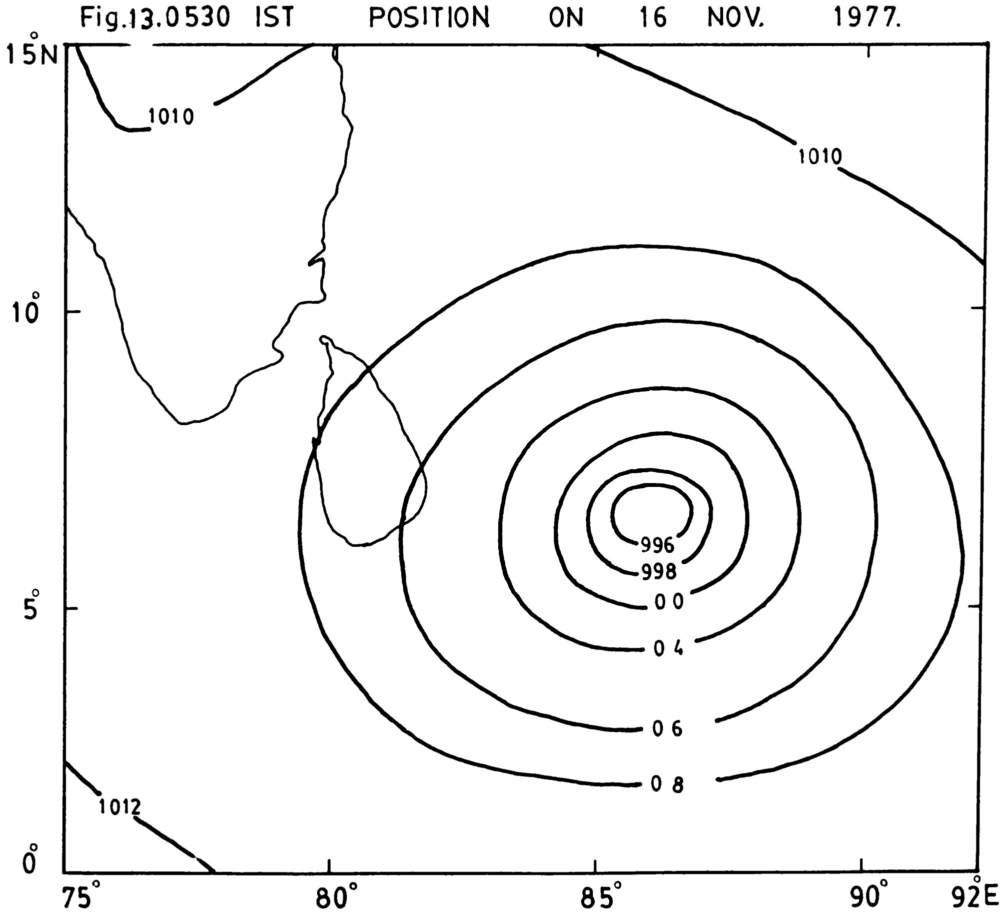


Fig. 14. 0530 IST POSITION ON 17 NOV.1977.

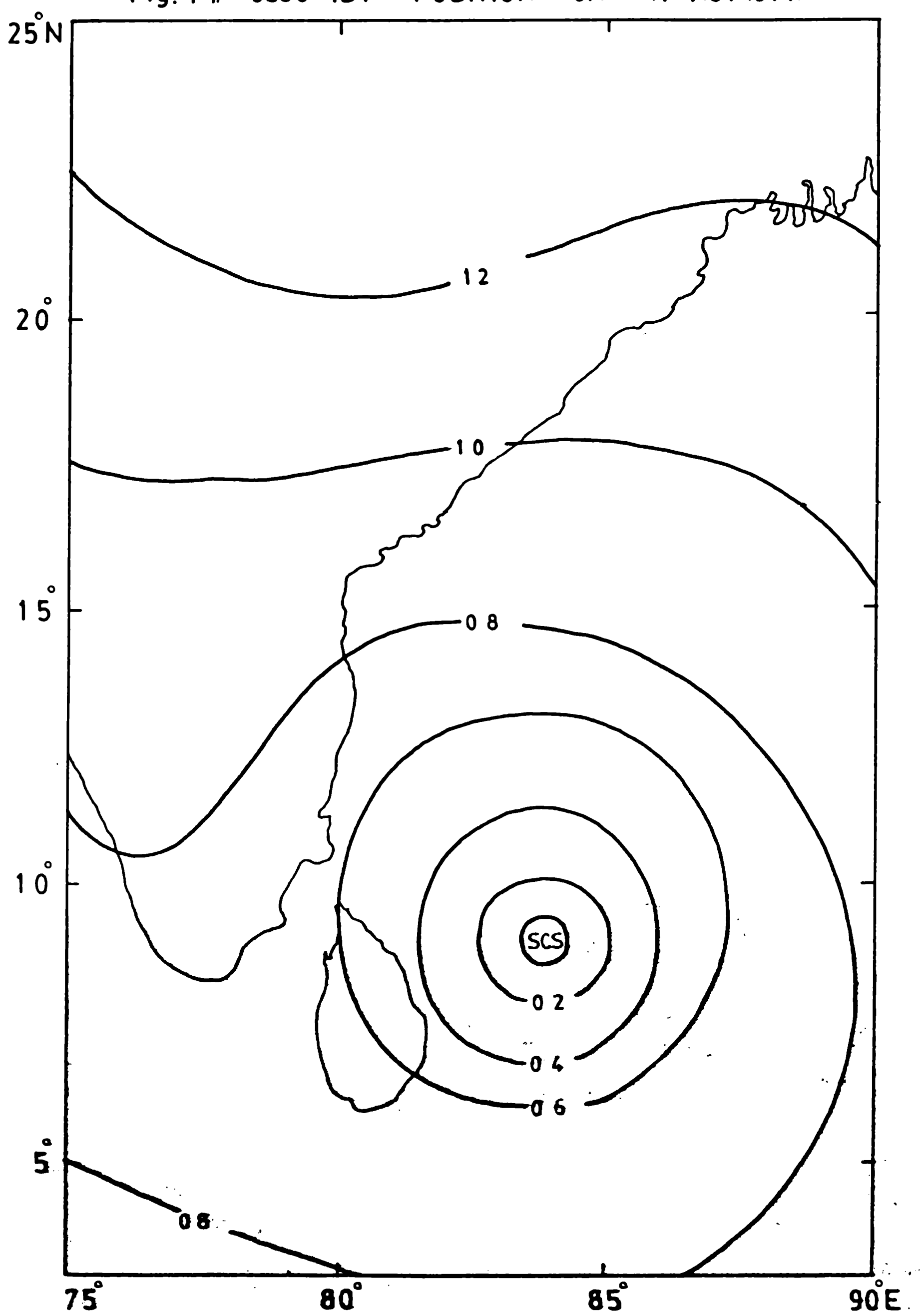


Fig. 15. 0530 IST POSITION ON 18 NOV. 1977.

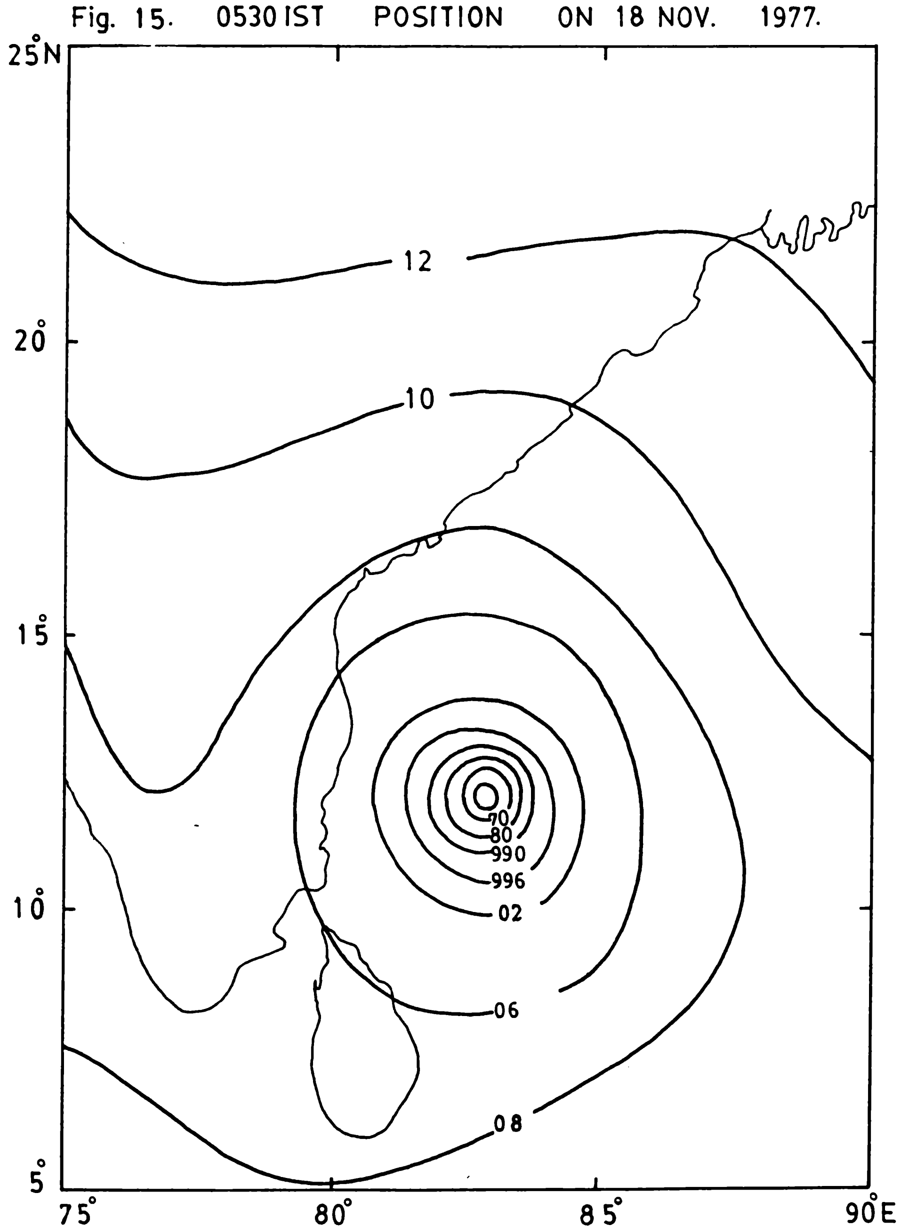


Fig.16. 0530 IST POSITION ON 19 NOV. 1977.

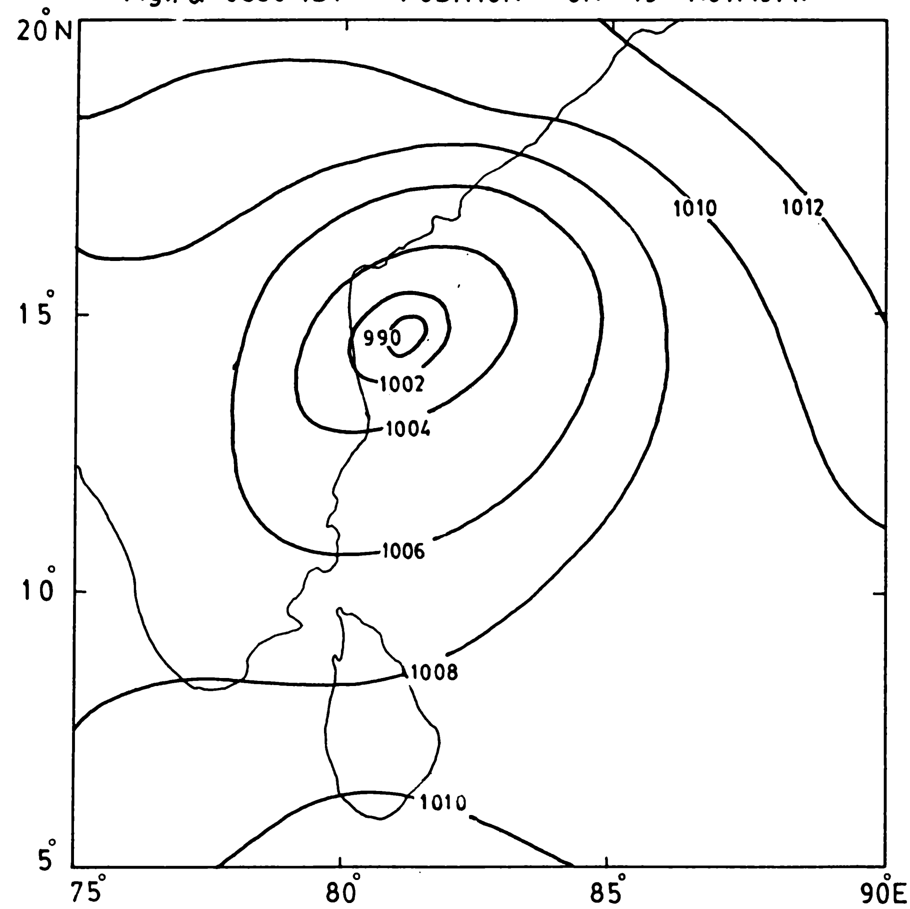


Fig.17. 0530 IST POSITION ON 20 NOV. 1977.

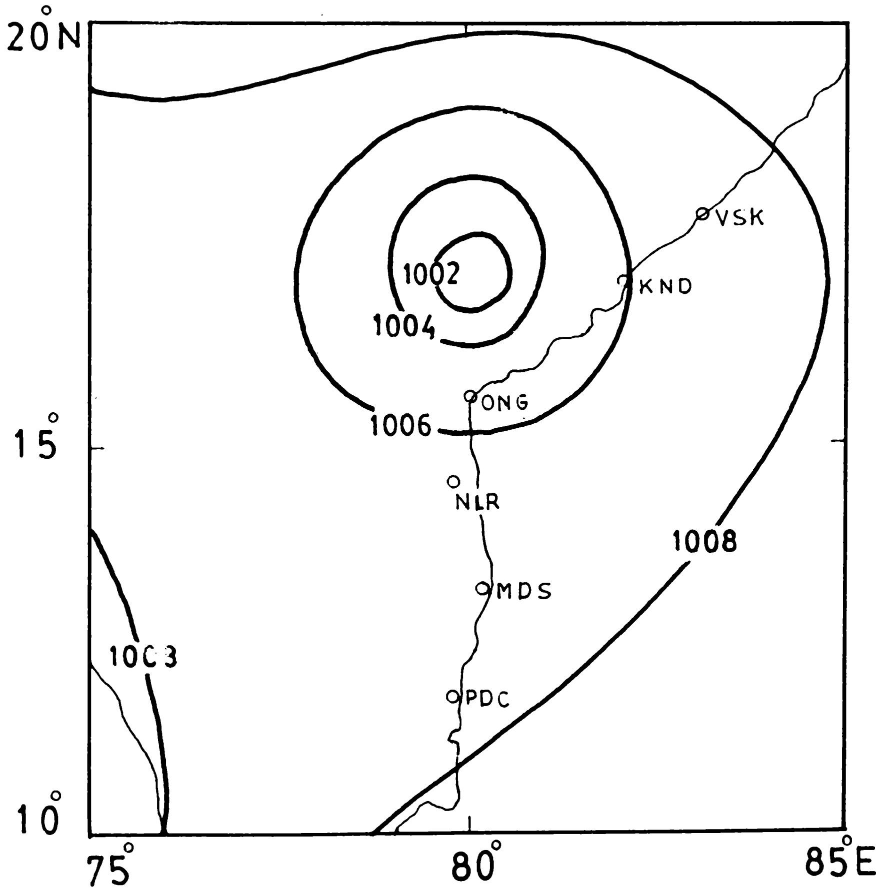


Table V. Observed and computed winds during the period 15 - 18 November 1977

Sl. No.	Observed values								Computed values							
	Date of observation	Time (I.S.T.)	Ship name /Code	Lat °N	Long °E	Pressure (mb)	Obs. wind		Date and time of weather chart		Grid points		P (mb)	N (km)	R (km)	V_g (kts)
							Direction	Speed (kts)	Date	Time	Lat °N	Long °E				
1.	15.11.1977	0930	-	9.2	82.5	1005.2	ENE	20	15.11.'77	0530	9-10	82-83	1.0	77.7	432.9	28.79
2.	16.11.1977	0730	-	9.3	82.3	1005.5	-	30	16.11.'77	0530	9-10	82-83	1.0	83.25	521.7	32.13
3.	16.11.1977	1100	-	7.5	86.0	1006.5	NE	40	16.11.'77	0530	7-8	86-87	2.0	38.85	122.1	40.48
4.	16.11.1977	1300	-	7.0	84.5	1004.5	N	40	16.11.'77	0530	6-7	84-85	2.0	55.5	177.6	40.31
5.	17.11.1977	1100	ATOA (Mahapriya)	12.5	85.0	1004.4	E	BF 6-7 (24-30 kts)	17.11.'77	0530	5-6	85-86	1.0	166.5	499.5	23.47
6.	17.11.1977	1100	S2LU (Banglaraasha)	10.4	85.6	998.0	ESE	70	17.11.'77	0530	12	82	5.0	44.4	166.6	68.99
7.	18.11.1977	0730	VWQQ (Vishva Mangal)	11.8	81.8	996.5	NNW	33	18.11.'77	0530	12-13	87	1.0	66.6	499.5	30

Fig.18 GRID USE FOR THE CYCLONE OF 14-20 NOV 1977.

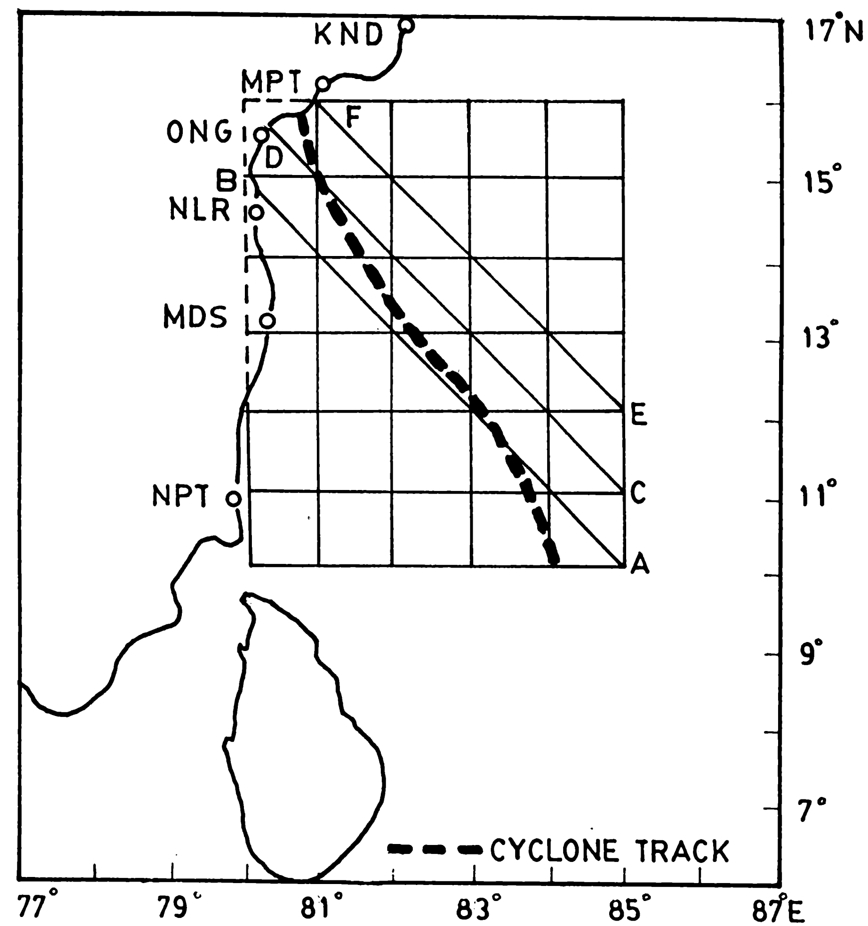
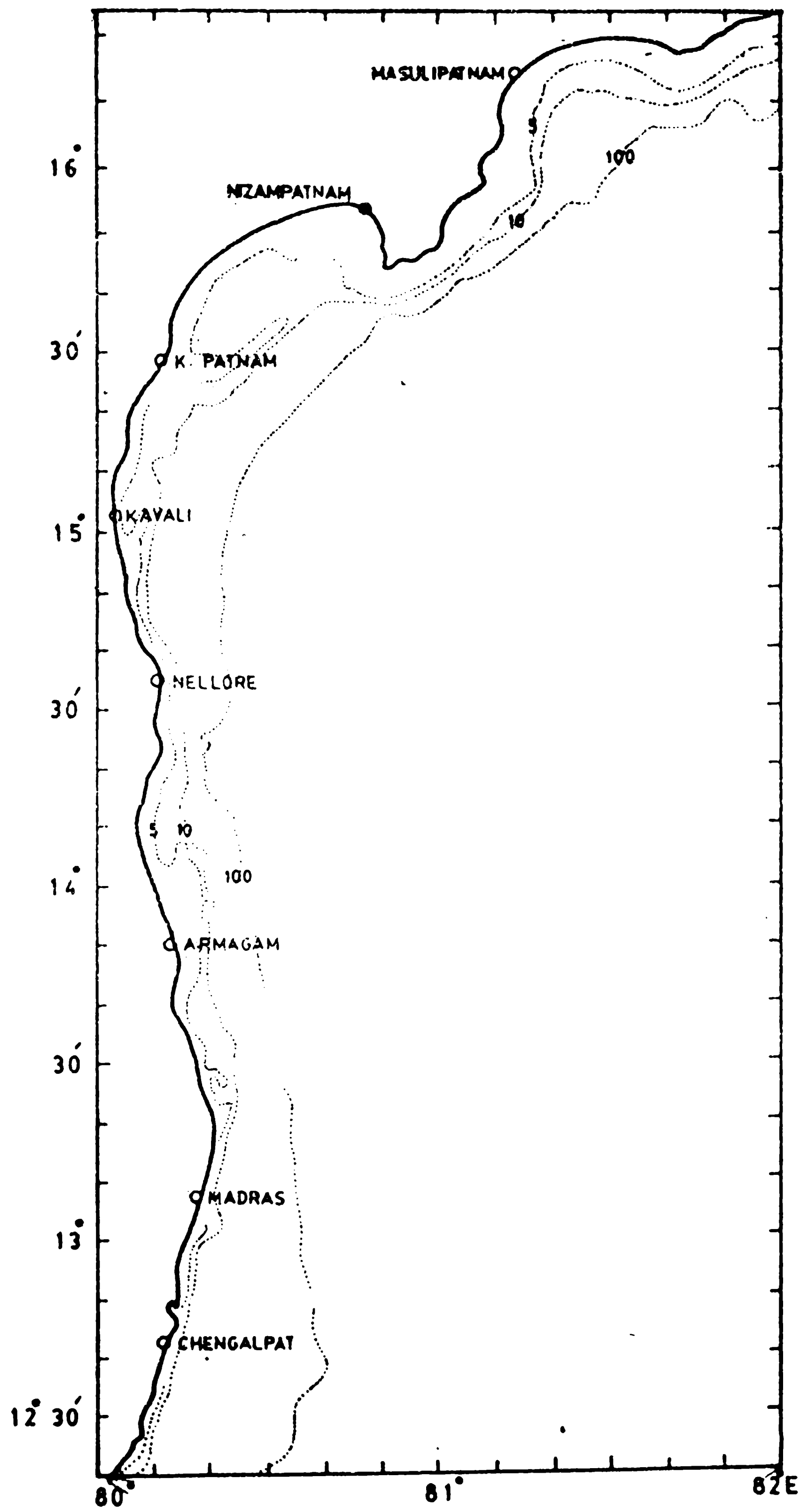


Fig.19. BATHYMETRY BETWEEN MADRAS & MASULIPATNAM.



It is seen in Fig.20 that the wind setup is the lowest along AB and highest along EF whereas it is intermediate along CD. While the peak value along AB is only 1.5m the peak values along CD and EF are 6.2 and 6.9 m respectively. The increase of surge potential from left of the storm track to right hand side of the track is thus 5.4 m for this cyclone. The wind setups from 0800 IST on 18th to 0800 IST on 19th along AB and CD are almost same (0.35 m), while the wind setup along EF during this period is much higher (around 1.8 m). The fall in sea level after 0000 IST on 20th along AB is gradual, whereas along CD and EF it is more abrupt. ~~In other words the time taken for the sea level to return to normal along AB is 8 hours and along CD and EF, it is 4 to 5 hours.~~ After 0800 IST on 20th the sea level came to almost normal along the three paths. The duration of the surge along AB and CD is about 24 hours, whereas, along EF it is 7 hours longer. The surge, as per the computed value comes to 6.6 m and occurs between 1700 and 2300 IST on 19th. Pant et al. (1980) reported that a storm surge, about 5 m high above mean sea level, affected Divi, Edurumondi and Eletidibba Islands at the Krishna Estuary and along the southern coastal parts of Masulipatnam. This height was visually estimated. According to them, the maximum flooding occurred between 1700 and 2300 IST on 19th. The orders of magnitude of the

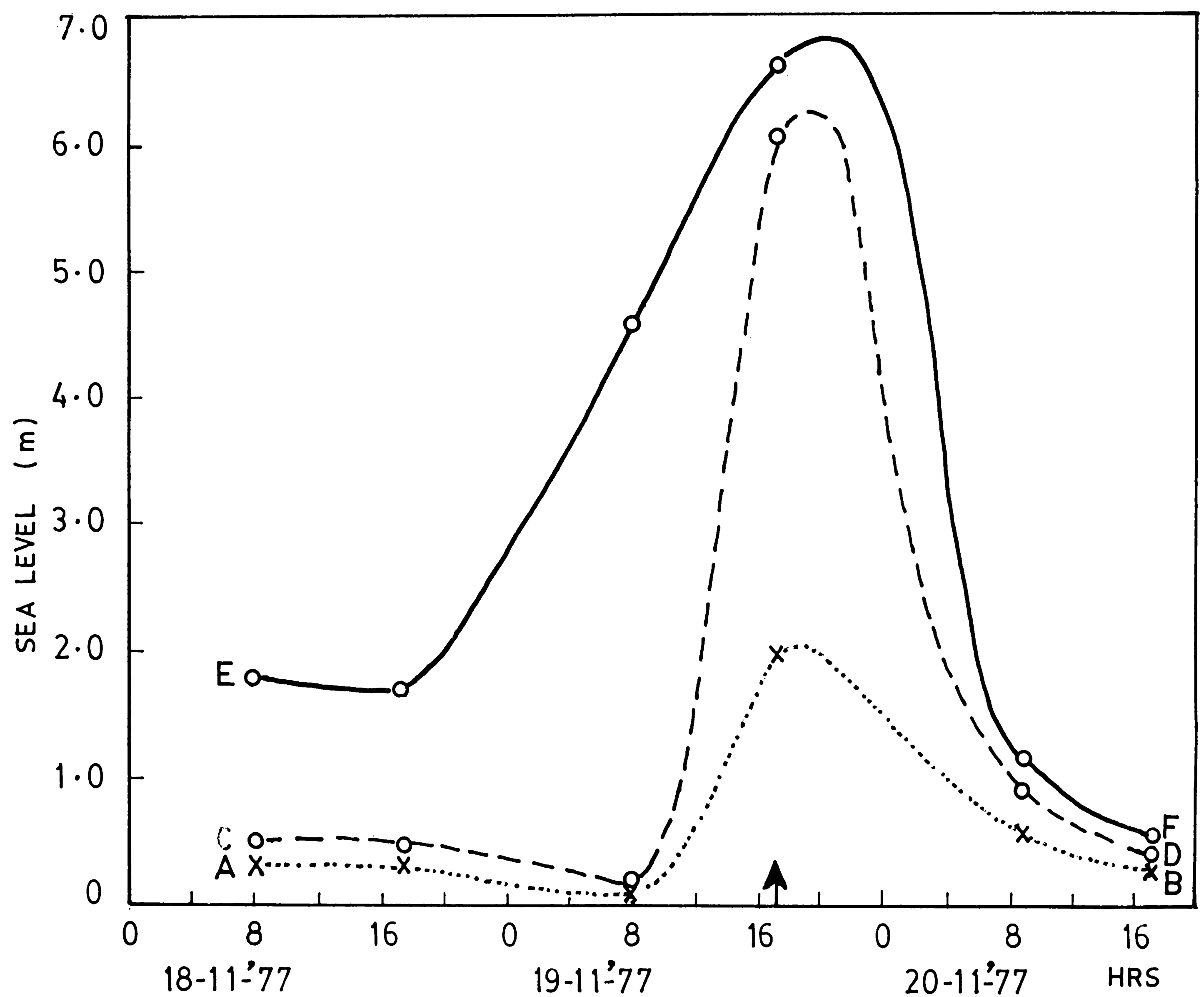


Fig.20. SURGE AT THE LAND FALL BETWEEN 18-11-'77 TO 20-11-'77.

computed and observed surges and their occurrence coincide very well. In the case of a visually observed surge, the height may be an underestimate. Johns et al. (1981) computed the surge generated by this cyclone using three different numerical models. They got a maximum surge of 5.3 m at Kavali at t = 50 hrs that corresponds to the afternoon of 18th November, and then a maximum surge of 4.8 m at Divi Island at t = 55 hrs corresponding to the morning of 19th November. In the extreme northeast coast of Andhra Pradesh at Visakhapatnam, they got the maximum surge elevation of about 3.3 m at t = 45 hrs. According to them the highest peak surge occurred at Kavali and not at Divi and claimed that the predicted and estimated maximum elevations are in good agreement at Divi. The model also appears to produce the peak elevation some hours in advance of the reported time of flooding. Obviously, the numerical model of Johns et al. (1981), seems to be not conforming to reality.

2.1.5. Tide at Madras and Visakhapatnam

The tide gauge data available for the period 18-20 November 1977 for the ports of Madras and Visakhapatnam, located north and south of the landfall are analysed and represented in Figs.21a and b respectively. While the average rise in sea level at Madras was 0.5 m,

Fig.21. TIDE AND SURGE OF MADRAS FROM 18 - 20 NOV. 1977.

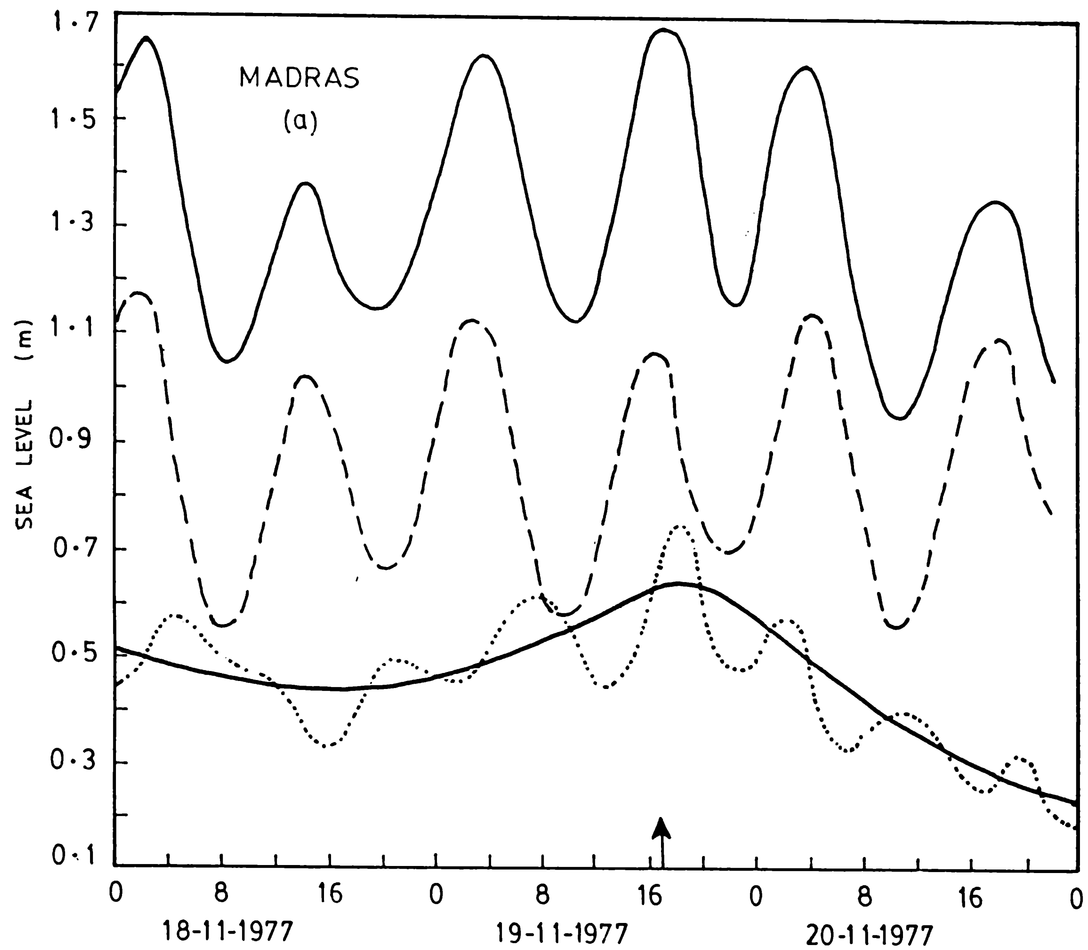
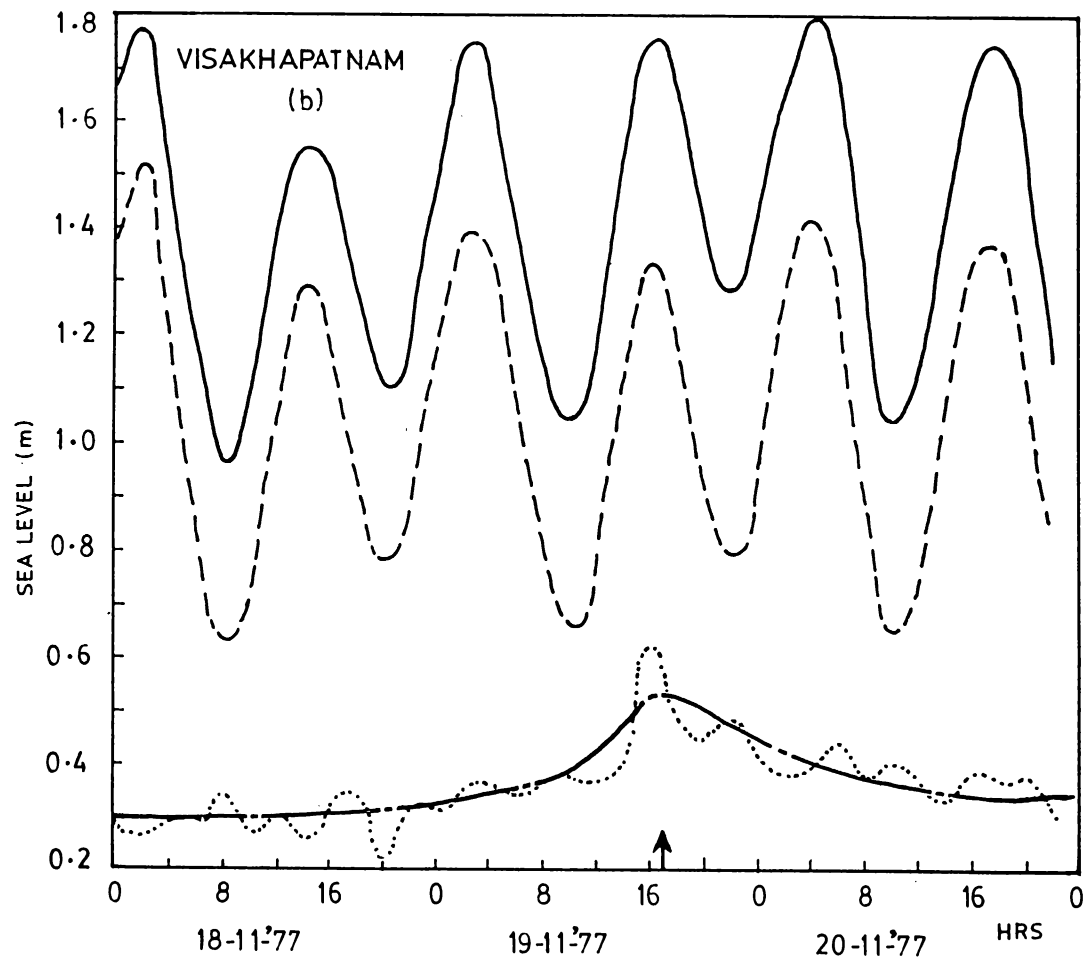


Fig. 21. TIDE AND SURGE OF VISAKHAPATNAM FROM 18-11-1977 TO 20-11-1977.



the peak of 0.6 m occurred at 1800 IST on 19th. At Visakhapatnam the average rise is 0.3 m and the peak of 0.53 m occurred at 1600 IST. Thus the maximum sea level at Visakhapatnam and Madras occurred before and after landfall at Chirala, which is located in between these two stations. Hence it can be inferred that the surge progressed from north to south. The duration of the high sea level at the two tide gauge stations and that of the surge at the land fall is almost the same.

As per the tide tables the high tide at Kakinada predicted for 19th is 1.44 m above datum level (0.57 m above mean sea level) at 1607 IST and low tide is 0.89 m above datum level (0.02 m above mean sea level) at 2211 IST. However, the total surge of over 6 m occurred at the landfall (from Fig.20), which is not very far away from Kakinada, implies that the contribution to the storm surge from astronomical tide is very low and that the major contribution is from the cyclone (meteorological).

2.2. Masulipatnam cyclone of 3-6 November 1976

2.2.1. Life history of the storm

A low pressure area moving westward from the Gulf of Siam across Andaman Sea, concentrated into a depression on the morning of 3rd over the Central Bay with its centre near $13^{o}N$, $87^{o}E$. Thereafter, moving in a northwesterly direction, the depression progressively intensified and became a severe cyclonic storm by the evening of 4th. The cyclone crossed the coast just north of Masulipatnam around 2030 IST on 4th as indicated by the satellite picture (Fig.22). On 5th evening this system was still of storm intensity over north interior Karnataka. The cyclone weakened into a depression by 5th night and moved westnorthwestwards across Maharashtra.

2.2.2. Wind and pressure reports of the cyclone

On 4th evening the low level winds over coastal Andhra Pradesh were only 20-25 kt before the cyclone crossed the coast. This suggests that the system was a cyclone of narrow core (less than 100 km in diameter). When it was moving close to Masulipatnam and Gannavaram, these stations reported surface winds of 40 kt at 2030 IST, and 55 kt at 2330 IST respectively. The hourly observations of wind and pressure (Figs.25 a and b) show the lowest pressures of 985 mb at 2100 IST and 987 mb at 2340 IST respectively at Masulipatnam and Gannavaram on 4th. The winds are computed using the synoptic weather maps of 0830 IST and 1730 IST on 4th (Figs. 23 and 24). Table VI gives the comparison of observed and computed winds and they agreed very well. These computed winds are used for the computation of surge wherever required.

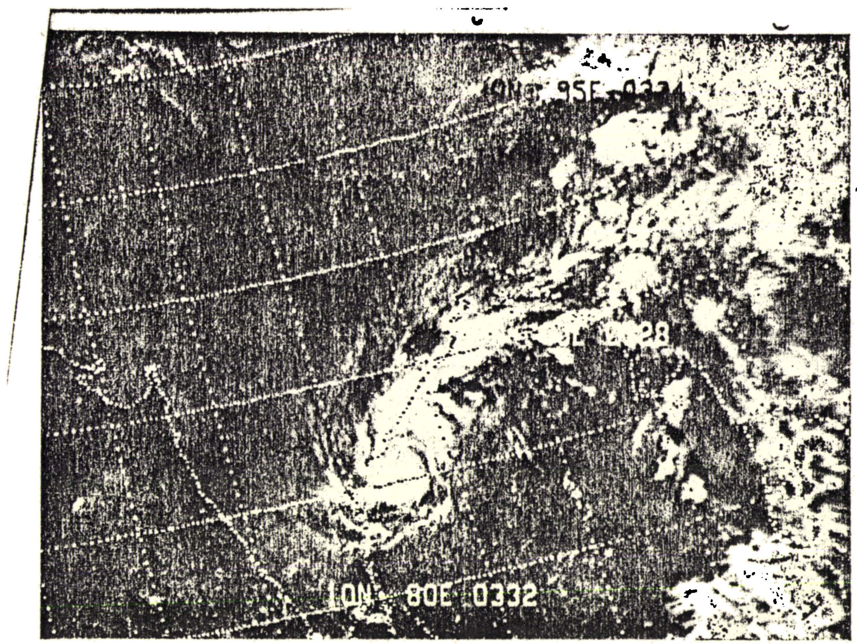


Fig 22. Satellite view of Masulipatnam cyclone at about 0300 GMT on 4 November

Fig. 23. 0830 IST POSITION ON 4-11-1976.

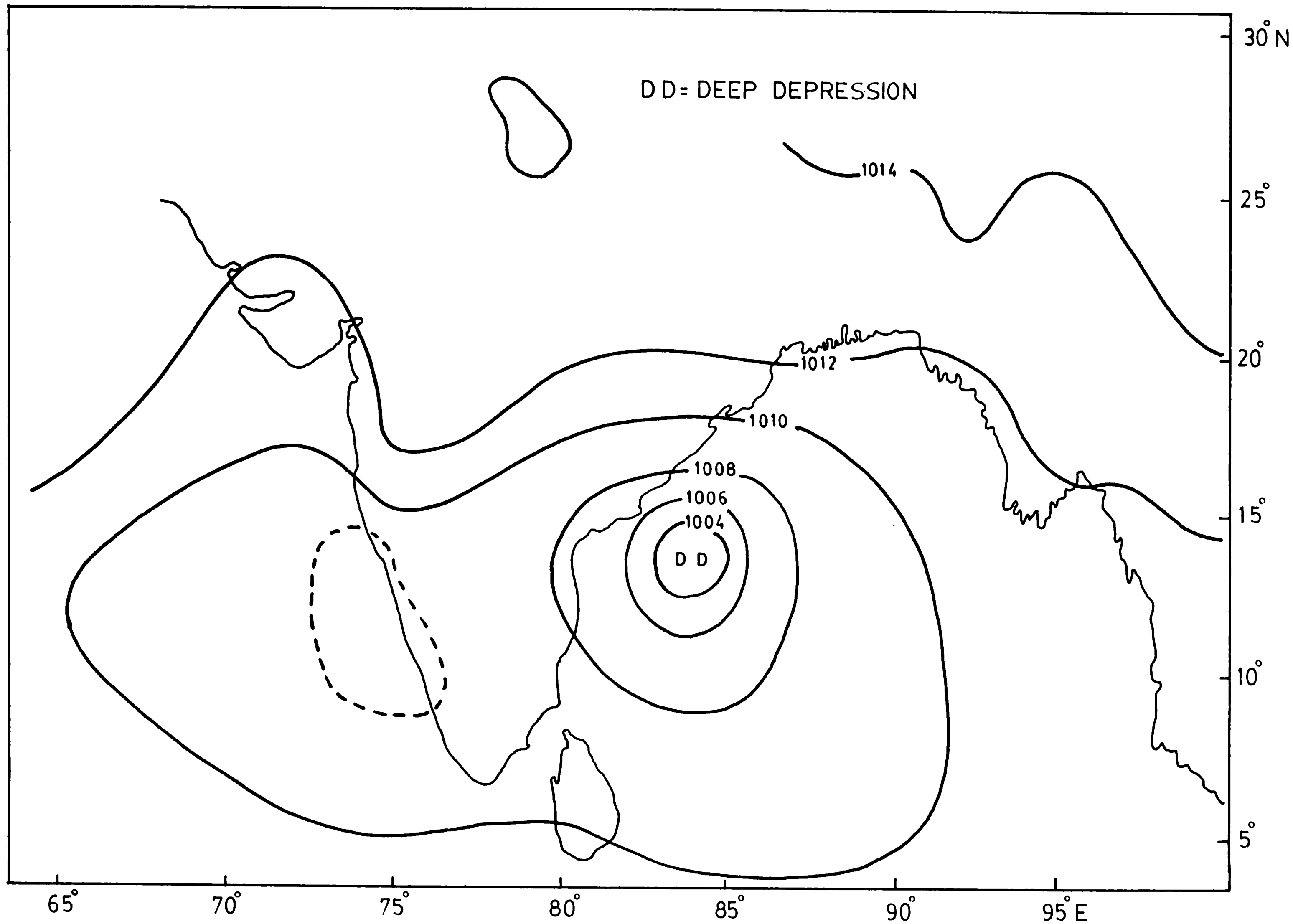
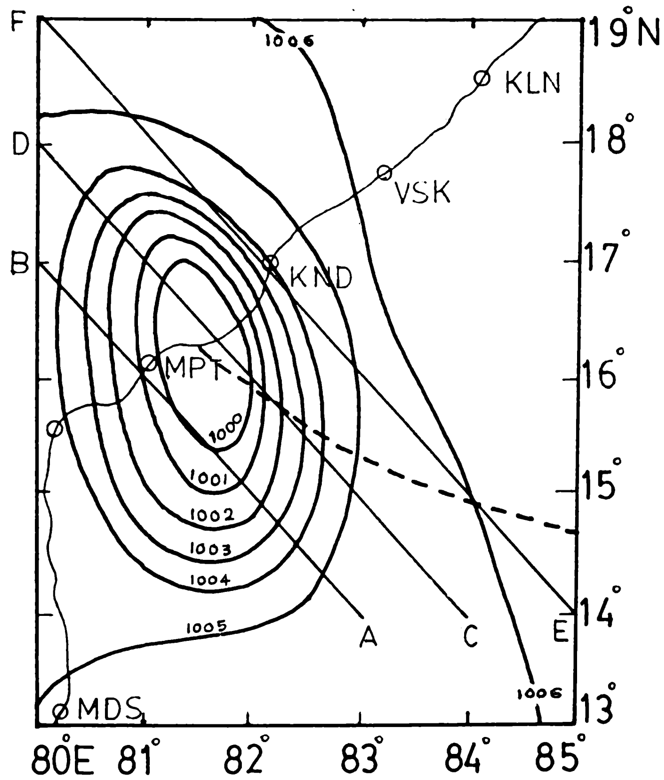


Fig.24.1730 IST POSITION ON 4-11-76.

(– – – –) CYCLONE TRACK



OBSERVATIONS OF WIND (kt) AND PRESSURE (mb)
AT MASULIPATAM ON 4 NOVEMBER 1976

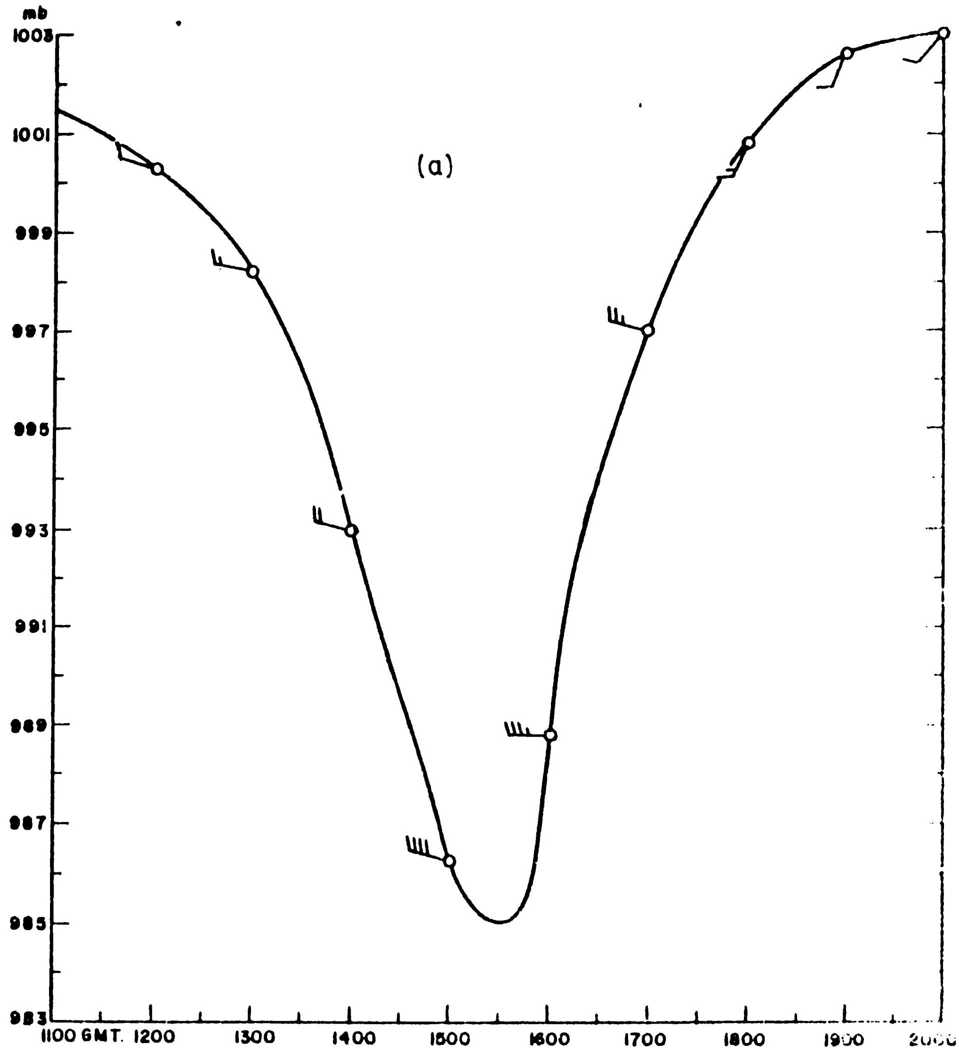


FIG. 25.

OBSERVATIONS OF WIND (kt) AND PRESSURE (mb)
AT GANNAVARAM ON 4 NOVEMBER 1976

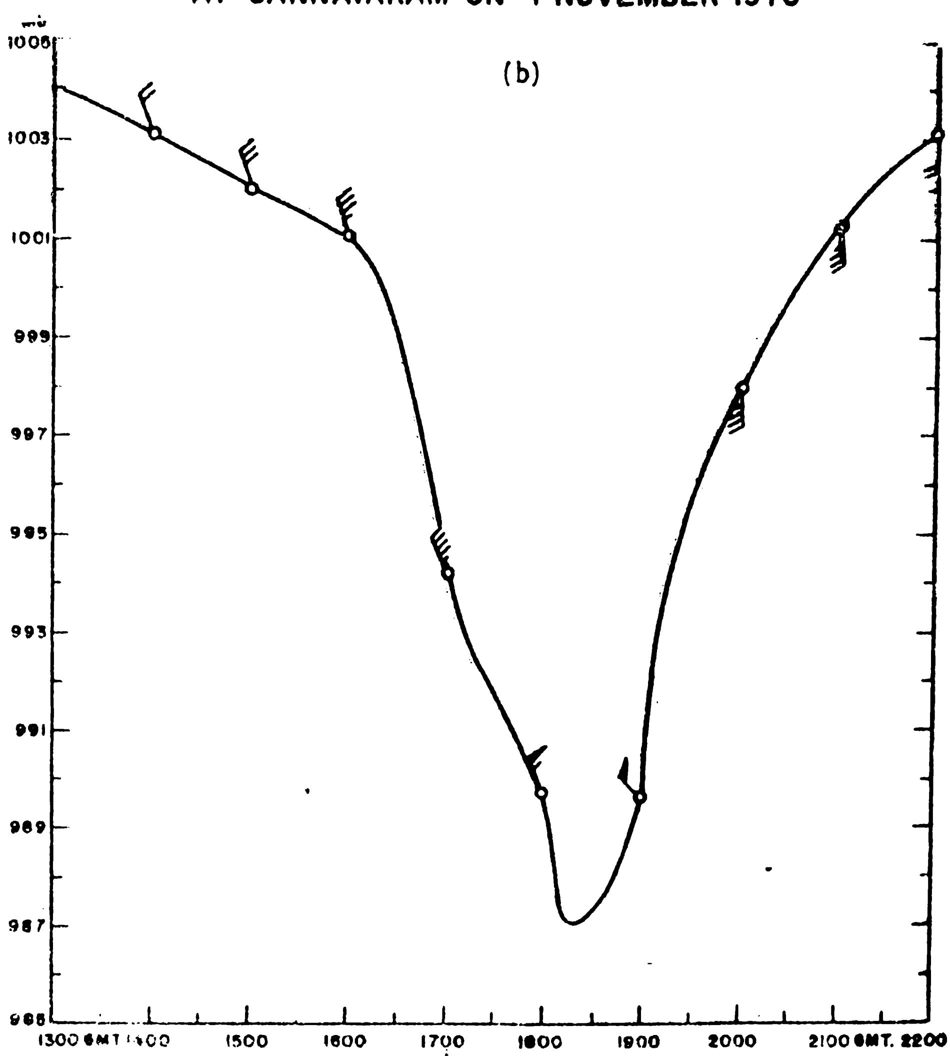


FIG. 25.

Table VI. Comparison of observed and calculated winds on 4.11.1976.

Observed values				Computed values						
Time of observation (IST)	Ship/or coastal station	Wind Direction	Wind Speed kt	Time of weather chart (IST)	Grid position Lat °N	Grid position Long °E	ΔP mb	ΔN km	R km	Vgr kt
1930	Gannavaram	NW	20	1730	14-15	82-83	1	30	210	19.75
1730	Ongole	E	15	1730	15-16	81-82	1	40	140	13.95
2030	Glasgow (Ship)	SW	20	1730	14-15	83-84	1	30	210	19.75
1830	ATIV	SW	20	1730	15-16	82-83	1	20	140	23.50
0630	SHIP	SSE	20	1730	14-15	84-85	1	150	350	26.08
1730	ATNG	E	30	1730	16-17	82-83	1	20	100	35.40

2.2.3. Surge at landfall

The computed wind setup at the landfall along AB, CD and EF is presented in Fig.26. It can be noticed that the wind setup along AB is the lowest and the highest along EF, with intermediary values along CD. The maximum wind setup along AB, CD and EF are 0.62, 0.70 and 0.90 m respectively. The peak occurs almost at the time of landfall along AB and consequently little later along CD and EF respectively. The wind setup along AB and CD appear to be more abrupt than along EF. Similarly, the fall in the sea level also shows an abrupt variation along AB and CD. A comparison of the three curves shows that the higher surge level along EF is maintained for a longer period. As there is no available reported data about the development of the surge, it could not be possible to make a comparative study of the computed with the observed. The storm surge associated with this cyclone could not be reported probably because of its small magnitude of less than 1 m.

2.2.4. Tide at Madras and Visakhapatnam

The available tide gauge data on either side of the land fall viz., Madras and Visakhapatnam are analysed and shown in Figs.27 and 28 respectively.

Fig. 26. SURGE AT THE LAND FALL POINT BETWEEN 3-11-'76 TO 6-11-'76.

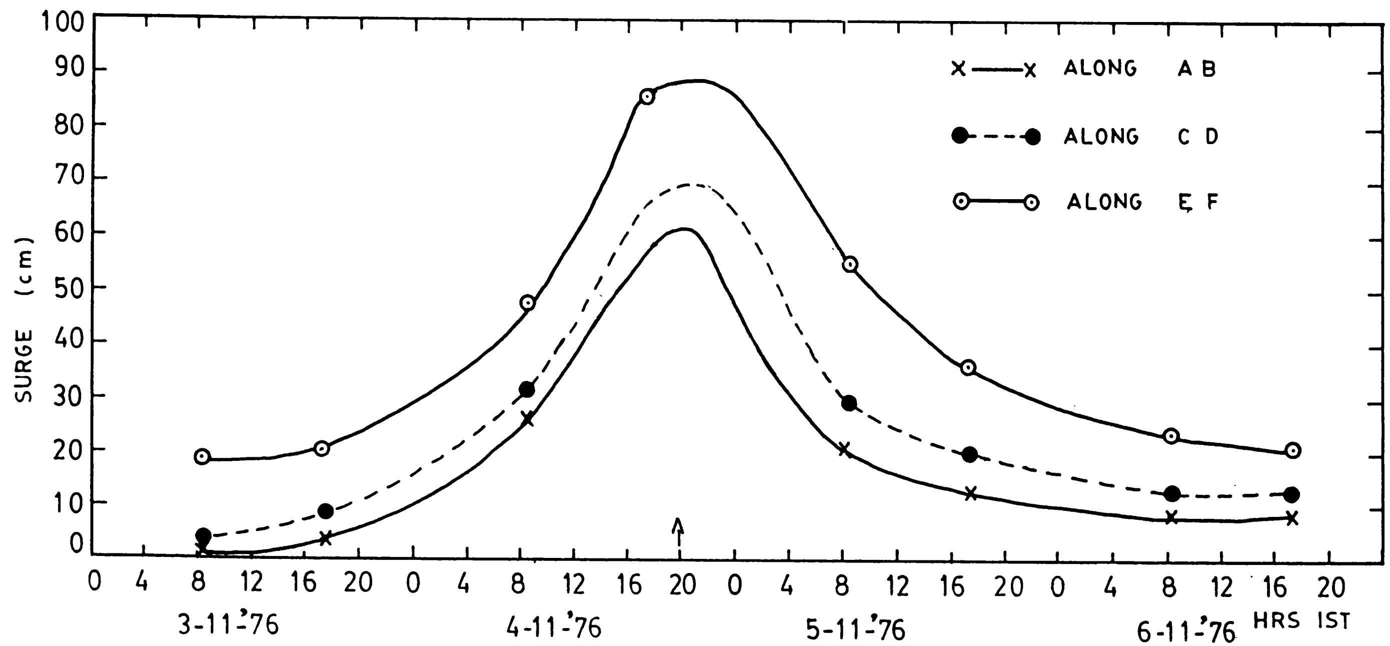


Fig. 27. TIDE AND SURGE OF MADRAS FOR THE PERIOD 3-11-'76 TO 6-11-'76.

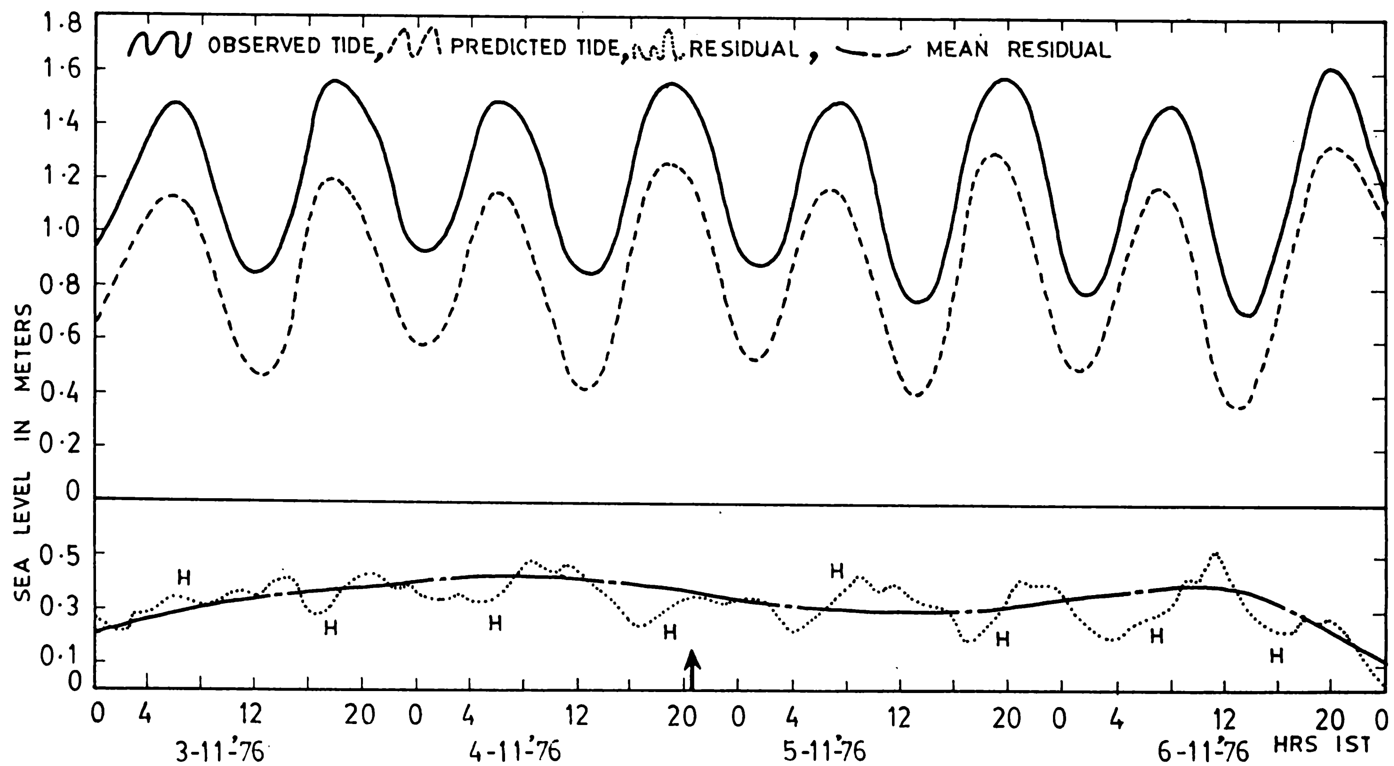
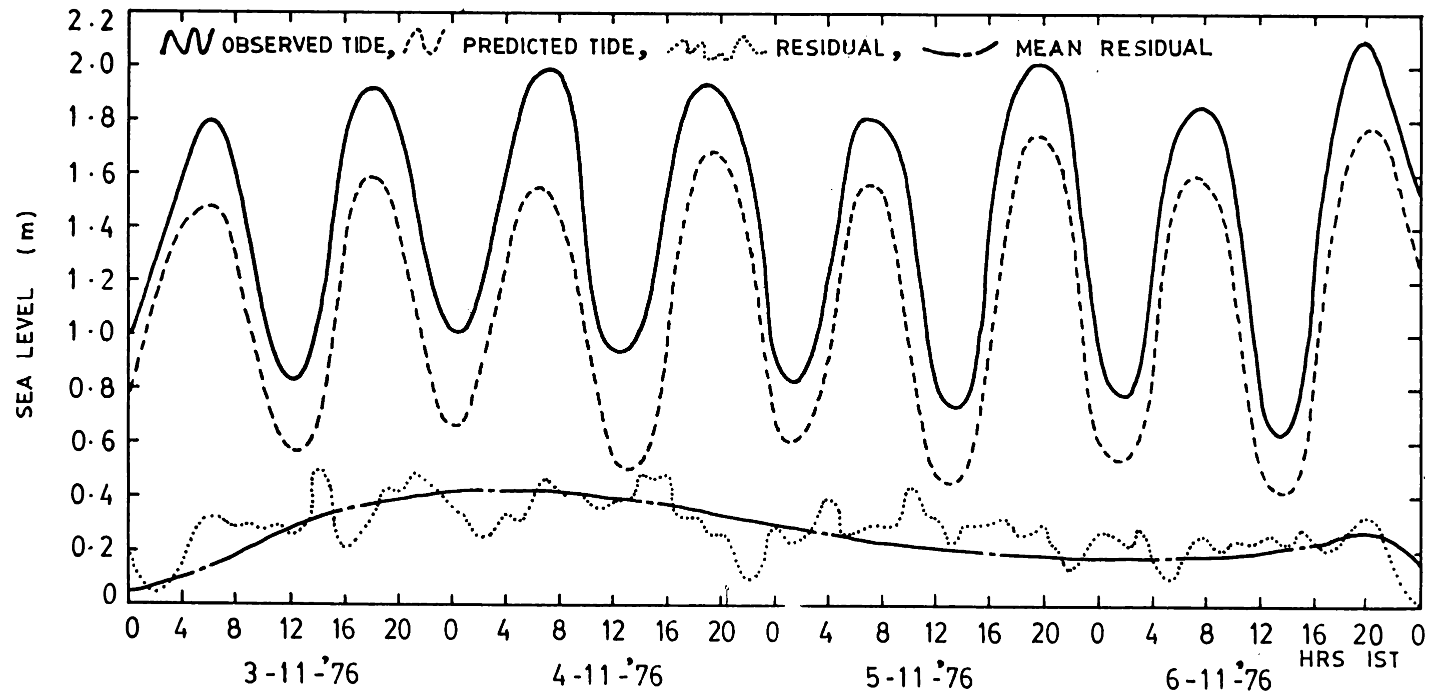


Fig.28. TIDE AND SURGE OF VISAKHAPATNAM FROM 3-11-'76 TO 6-11-'76.



The mean residual (non-astronomical) sea level at Madras during the period of occurrence of the cyclone was found to have an average value of about 0.3 m above the mean sea level with peaks of 0.41 m and 0.40 m occurring at 0800 IST on 4th and 1000 IST on 6th respectively. Evidently, the peaks at Madras occurred before and after the land fall of the cyclone. The transient sea level at Visakhapatnam occurred between 0400 IST of 3rd and 0000 IST of 5th with a peak value of 0.42 m at 0000 IST on 4th. The peak surge at Visakhapatnam was noticed before the occurrence of the landfall. A close examination of the tide gauge data at both the places indicates a similarity of the peak surge occurring around 16 hours before the landfall. At both the places the residual sea level shows a tendency of approaching the mean sea level much before and after the landfall, which means, that the oscillatory nature of sea level during the total duration of the cyclone is transitory.

2.3. Kavali Cyclone of 15-17 November 1976

2.3.1. Life history of the storm

A low pressure area from the South China Sea moved westward across the Andaman Sea on 14th and concentrated into a depression on the morning of 15th with its centre near 10.5°N, 87°E. Moving rather fast in a northwesterly direction, it became a deep depression the same evening and rapidly intensified into a severe cyclonic storm on the morning of 16th when it was centred near 13.8°N, 82.5°E. The severe cyclonic storm crossed south Andhra Coast between Nellore and Kavali (Kavali is about 35 km north of Nellore) around midnight of 16th and weakened into a low over Rayalaseema by 17th. The radar track of the storm (Fig.29) confirms that the cyclone crossed the coast between 2230 and 2330 IST of 16th at a point north of about 22.5 km from Nellore.

2.3.2. Wind and pressure reports of the cyclone

The ship VWLG reported northerly wind near 12°N, 84°E at 1630 IST of 15th and south-southwesterly wind at 1900 IST at the same position and barometric pressure drop during this period was from 1002.2 mb to 996.2 mb on the same day. The wind shift and the fall in pressure suggested that the depression centre passed close to this ship that evening. Another ship UYJE near 12.5°N, 87.0°E reported a wind of 30 kt at 1730 IST of 15th.

On 16th again the ship UYJE near 16.5°N, 83.5°E reported surface wind of 38 kt from east at 1130 IST and the ship ATJW near 15°N, 82°E reported 30 kt wind of ESE at 2030 IST.

RADAR TRACK OF SEVERE CYCLONIC STORM 16 NOVEMBER 1976

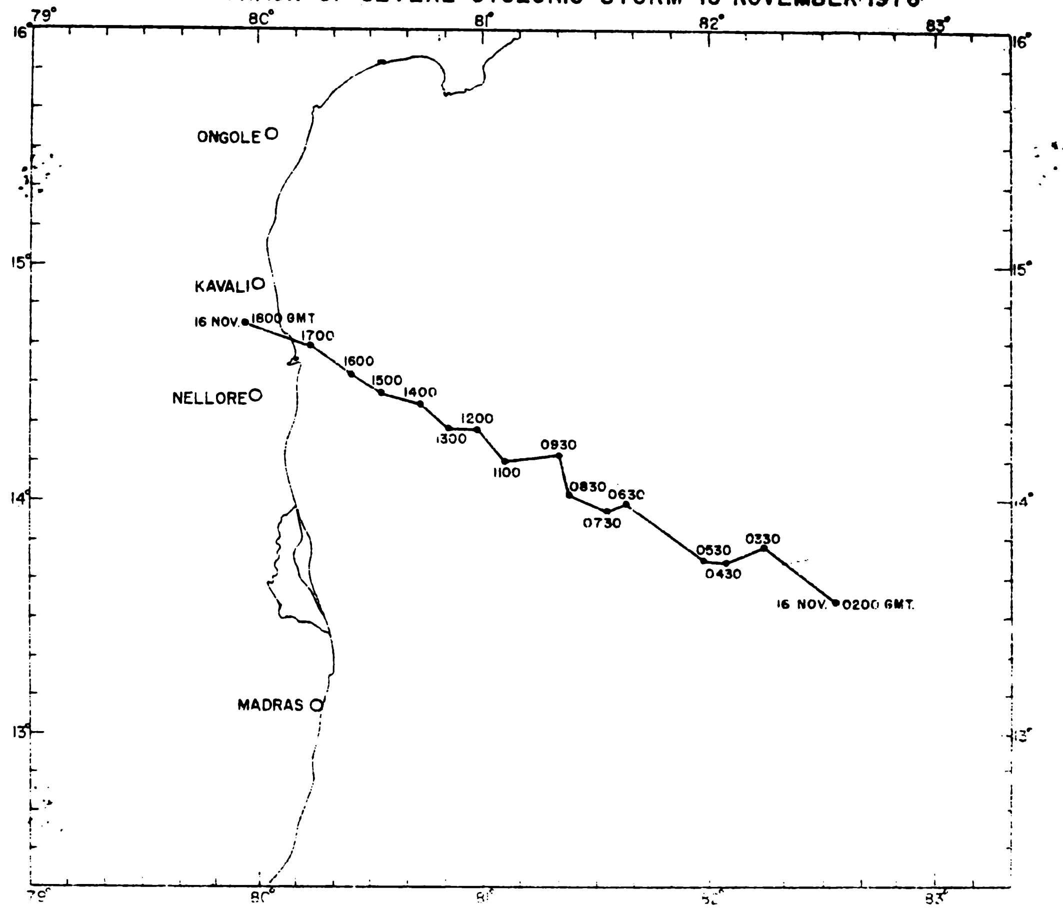


FIG. 29.

The satellite picture (Fig.30) of this storm on 16th, taken when it was crossing the coast, showed a maximum wind of 60 kt on Dvorak scale (Pant et al., 1978). But no such strong wind was reported by any other source of observation. Although, the severe cyclone passed within 30 km off Nellore observatory, the wind at Nellore did not exceed 25 kt at any time on this day (Fig.31). This suggests that the strongest winds did not extend beyond 20 km or so from the storm centre. This was also supported by the report of IMD sources according to which the severe damage extended only to about 20 km to the north and 15 km to the south of the track. A maximum wind of 60 kt would give the pressure of 992 mb at the centre of the storm.

The interesting aspect of this cyclone is that all the stations nearer to the landfall showed two low pressures, one in the morning and the other in the night of 16th (Fig.31). This may be due to the to and fro motion of the cyclone at the landfall area on that day.

Table 7 gives the comparison of the observed and computed values of the wind on 16th. The observed values are obtained from ship reports, nearby coastal stations and from the satellite pictures. The winds are computed using the pressure distribution of the synoptic maps of

Fig. 30. SATELLETE PICTURE KAVALI CYCLONE ON 16th NOV 1976 AT 1152 HRS IST.

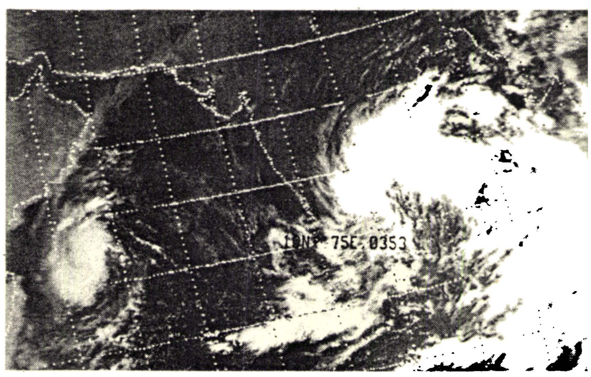


Fig. 31. HOURLY OBSERVATIONS OF WIND AND PRESSURE ALONG SOUTH EAST COAST OF INDIA ON 16-11-1976.

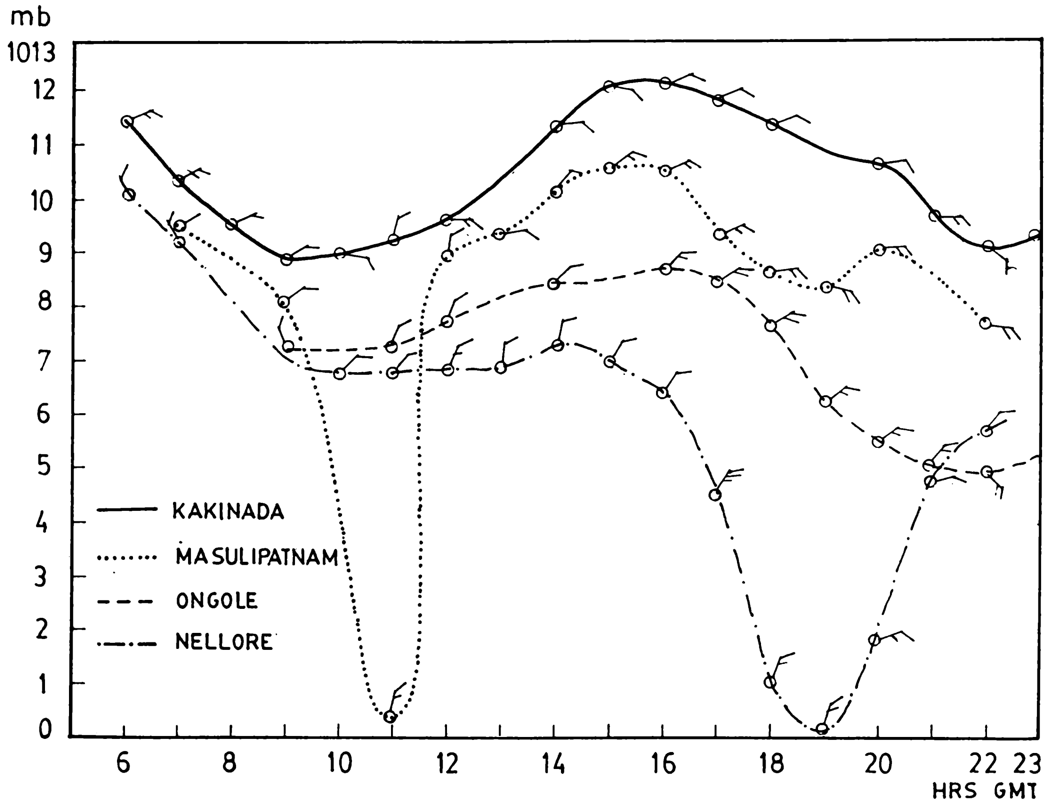


Table VII. Observed and computed winds at 1730 IST on 16 November 1976

Sl. No.	Observed values						Computed values							
	Time (IST)	Lat °N	Long °E	Observed wind		Ship/ Coastal station	Time of weather chart (IST)	Grid points		P (mb)	N (km)	R (km)	V_g (kts)
				Direction	Speed (kts)			Lat °N	Long °E				
1.	2030	15.3	82.0	ESE	30	ATJW	1730	15-16	83-84	2	140	250	24.05
2.	1130	12.5	87.0	NE	30	UYJE	1730	16-15	83-84	2	185	550	24.89
3.	1700	16.2	81.03	ENE	25	Masulipatnam	1730	16-15	81-82	2	90	375	27.66
4.	1700	15.3	80.04	NE	20	Ongole	1730	15-16	80-81	1	60	180	23.49
5.	1700	14.2	80.00	N	20	Nellore	1730	14-15	80-81	2	150	310	25.45
6.	-	14.8	81.00	-	60	Satellite picture	1730	14.8	81.0	2	30	180	52.78

0830 and 1730 IST which are shown in Figs.32 and 33. As the observed winds are not available in all the grids used, the need of computed winds arose and they are used wherever required.

2.3.3. Surge at landfall

Fig.34 depicts the computed wind setup in the area of landfall along AB, CD and EF. The setup increased northward along the coastline with lowest and highest values along AB and EF respectively. The setup along the three paths increases gradually while the fall is relatively abrupt. The peak surge occurred almost simultaneously with the landfall. The peak values along AB, CD and EF are 212, 244 and 265 cm respectively. It is interesting to note that there is a resurgence in the area of landfall particularly on the left hand side of the track.

2.3.4. Tide at Madras and Visakhapatnam

Figs. 35 and 36 show the tides of observed and predicted tides, and residual and mean residual sea levels at Madras and Visakhapatnam respectively during the period 15th to 17th November 1976 when the cyclone was intense.

Fig. 32. 0830 IST POSITION ON 16 NOV. 1976.

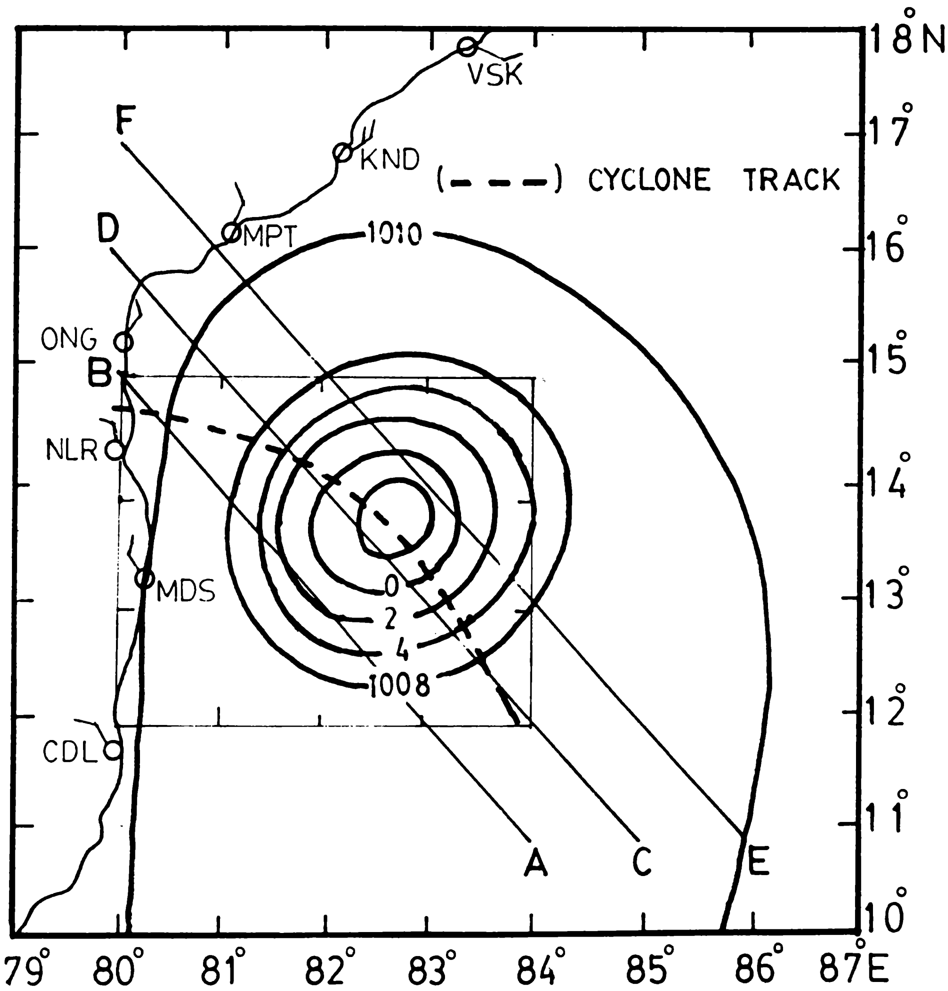


Fig. 33- 1730 IST POSITION ON 16 NOV. 1976.

(– – – – – –) CYCLONE TRACK.

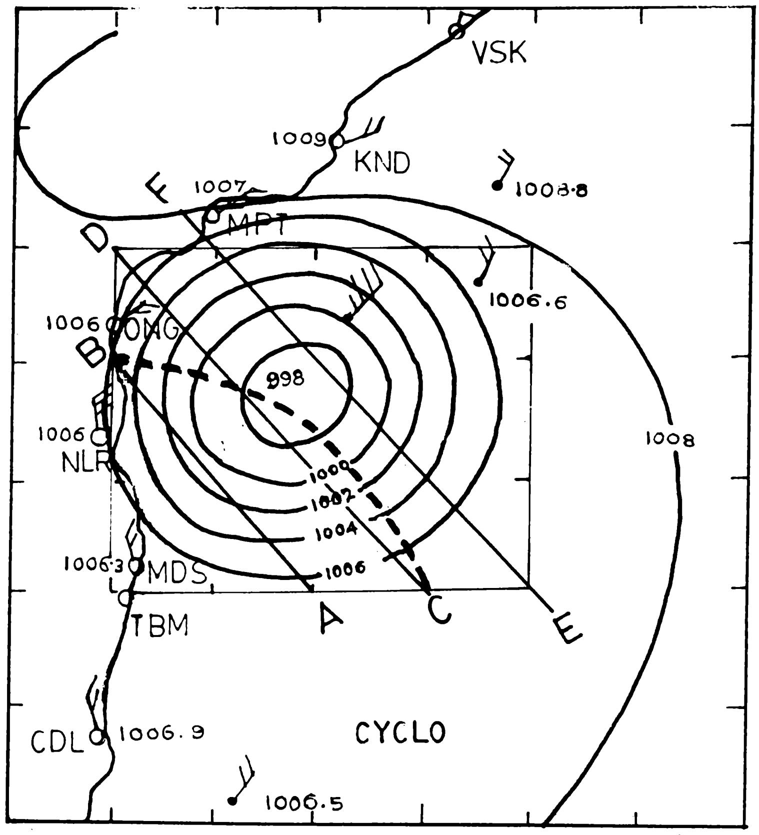


Fig. 34. SURGE ALONG AB , CD AND EF.

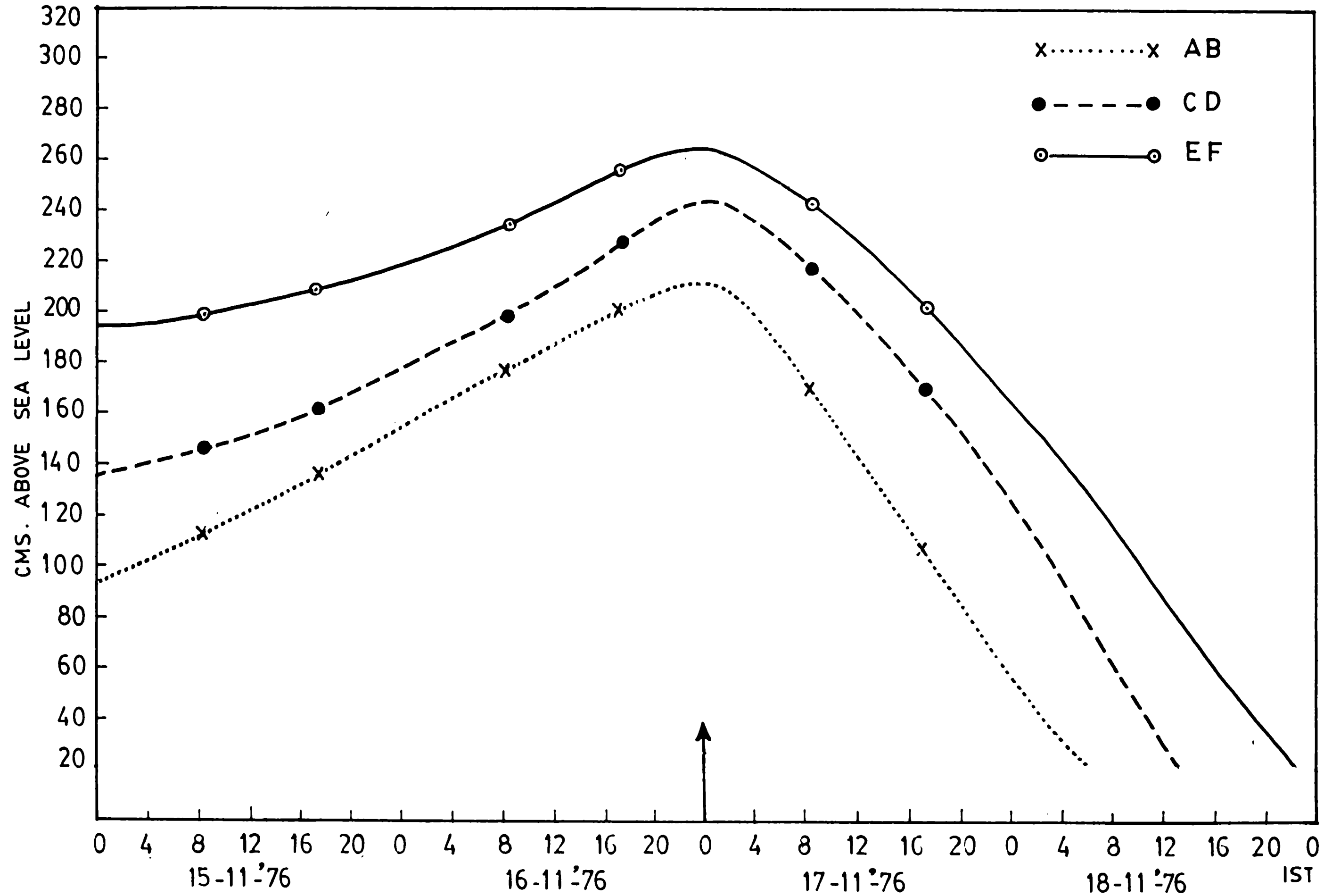


Fig. 35. TIDE AND SURGE OF MADRAS FOR THE PERIOD 15-11-'76 TO 17-11-'76.

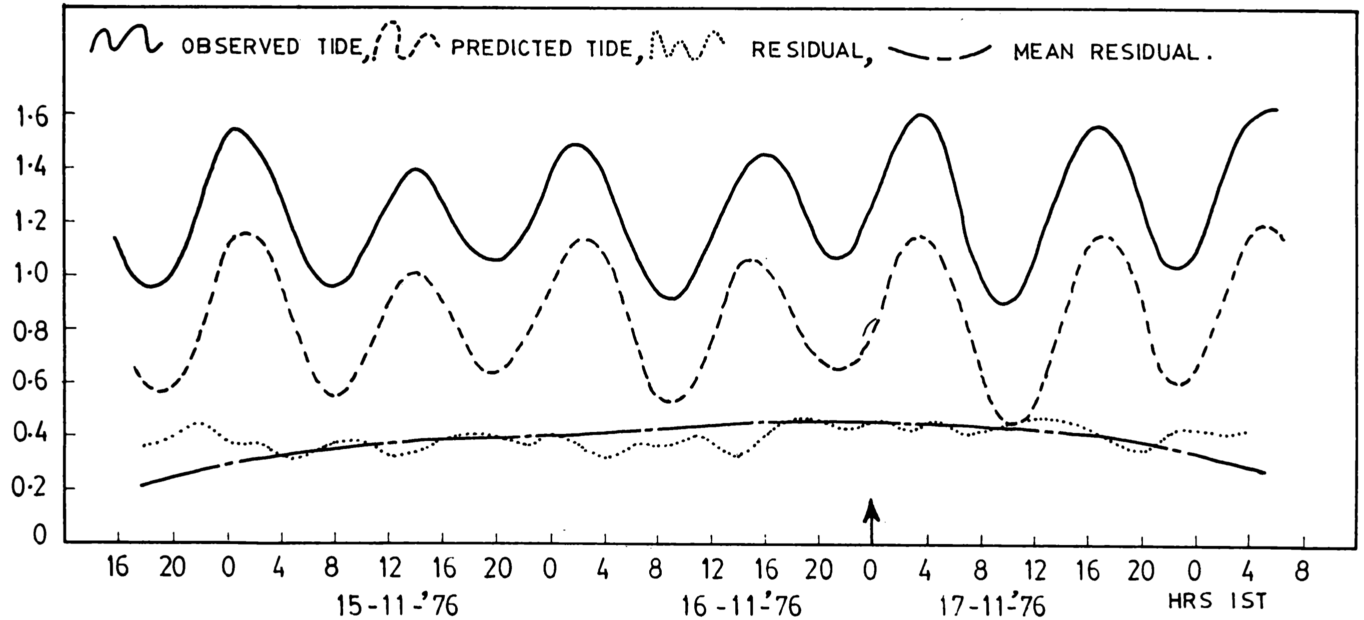
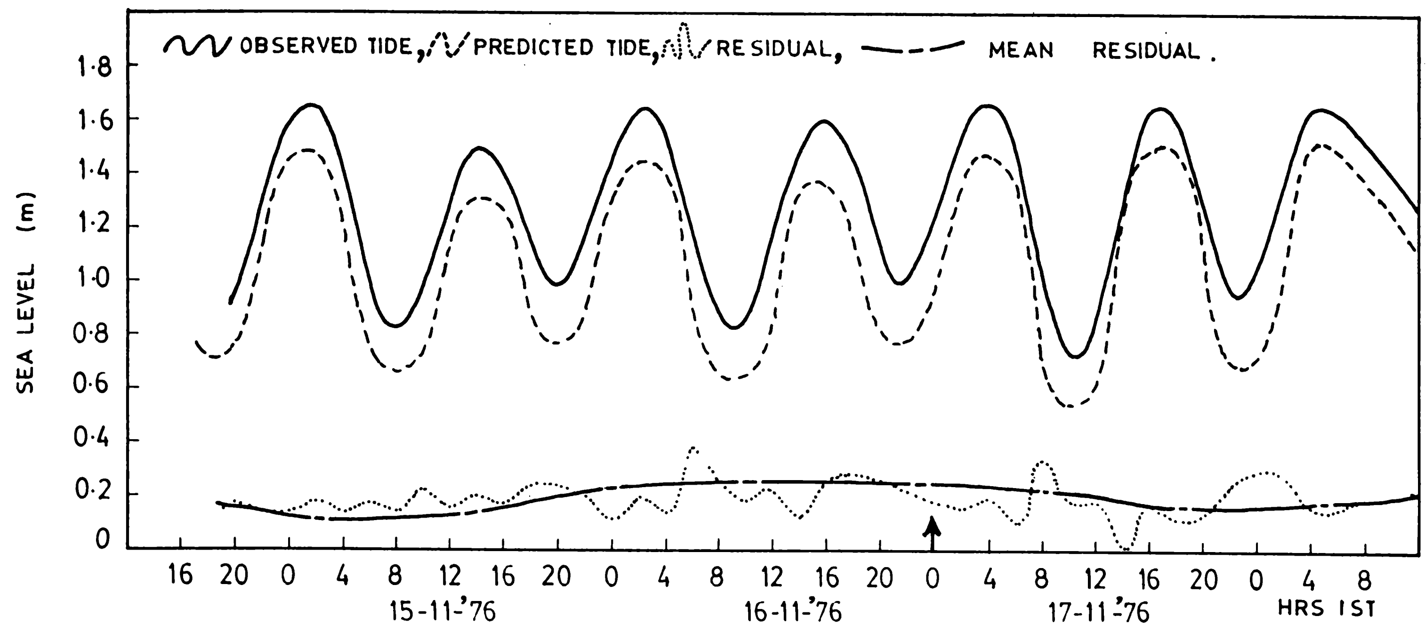


Fig. 36. TIDE AND SURGE OF VISAKHAPATNAM FOR THE PERIOD 15-11-'76 TO 17-11-'76.



The mean residual sea level at Madras increases gradually from a value of 0.2 m to a maximum of about 0.46 m before the landfall and then decreases. At Visakhapatnam the mean residual sea level is relatively much lower than that at Madras and occurs also much in advance of landfall. It can be inferred that the influence of the cyclone is confined only to the smaller area.

2.4. Contai Cyclone of 8-19 September 1976

2.4.1. A well marked low pressure area moved from Gulf of Siam northwestward to east Central Bay off south Arakan Coast on 7th and it concentrated into a depression on the morning of 8th with its centre near 17.5°N, 92.5°E. Moving northwestward it became a deep depression on 9th and further intensified into a severe cyclonic storm on the morning of 10th near 20°N, 90°E (Fig.37). The storm was tracked by the Cyclone Warning Radar at Calcutta from 1830 IST of 10th when it was located 300 km southeast of Calcutta (Fig.38). It can be inferred from the radar track (Fig.38) that the storm crossed the coast near Contai between 1030 IST and 1130 IST of 11th. The NOAA-4 satellite picture (Fig.39) taken at about the same time also confirms this. Although, its

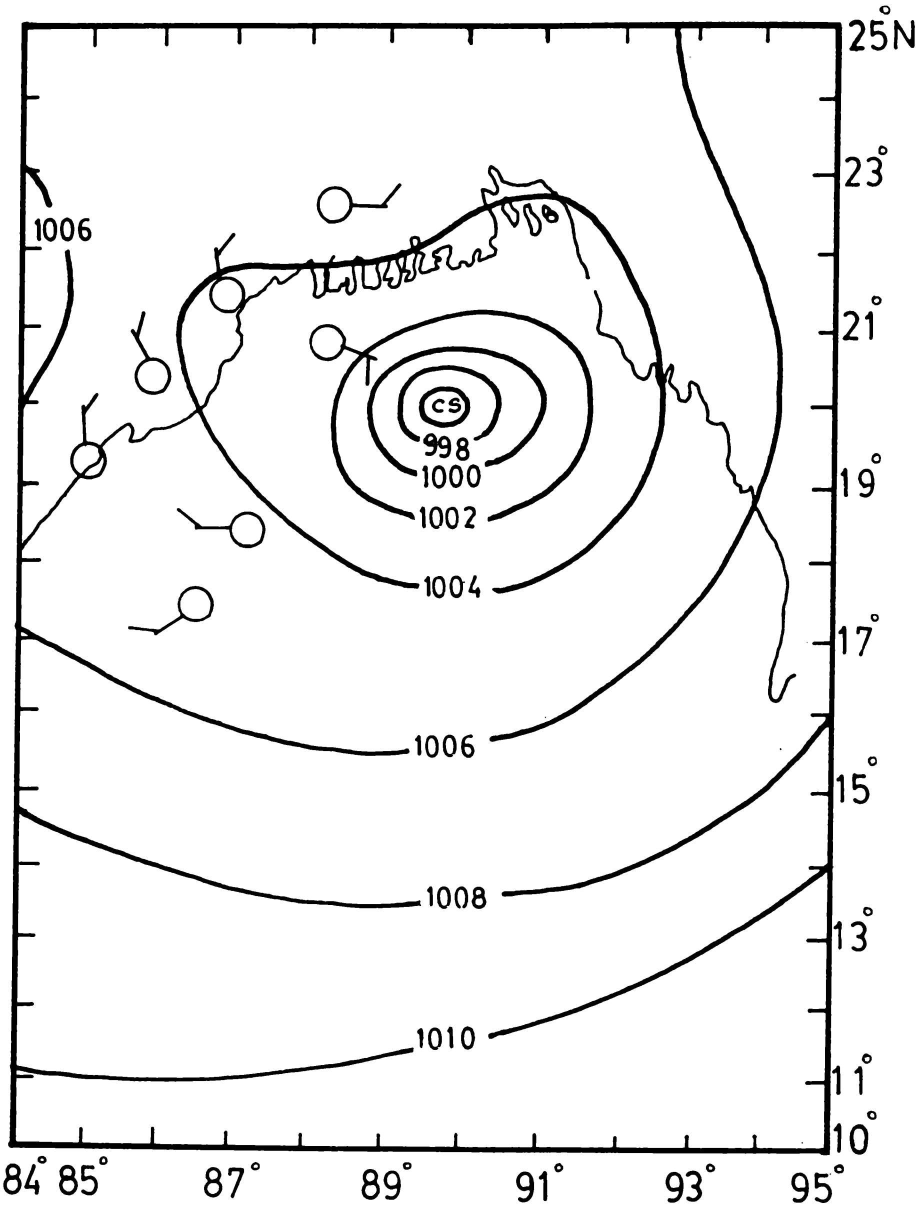


Fig.37.0830 POSITION ON 10-9-1976.

RADAR TRACK OF SEVERE CYCLONIC STORM 10-11 SEPTEMBER 1976

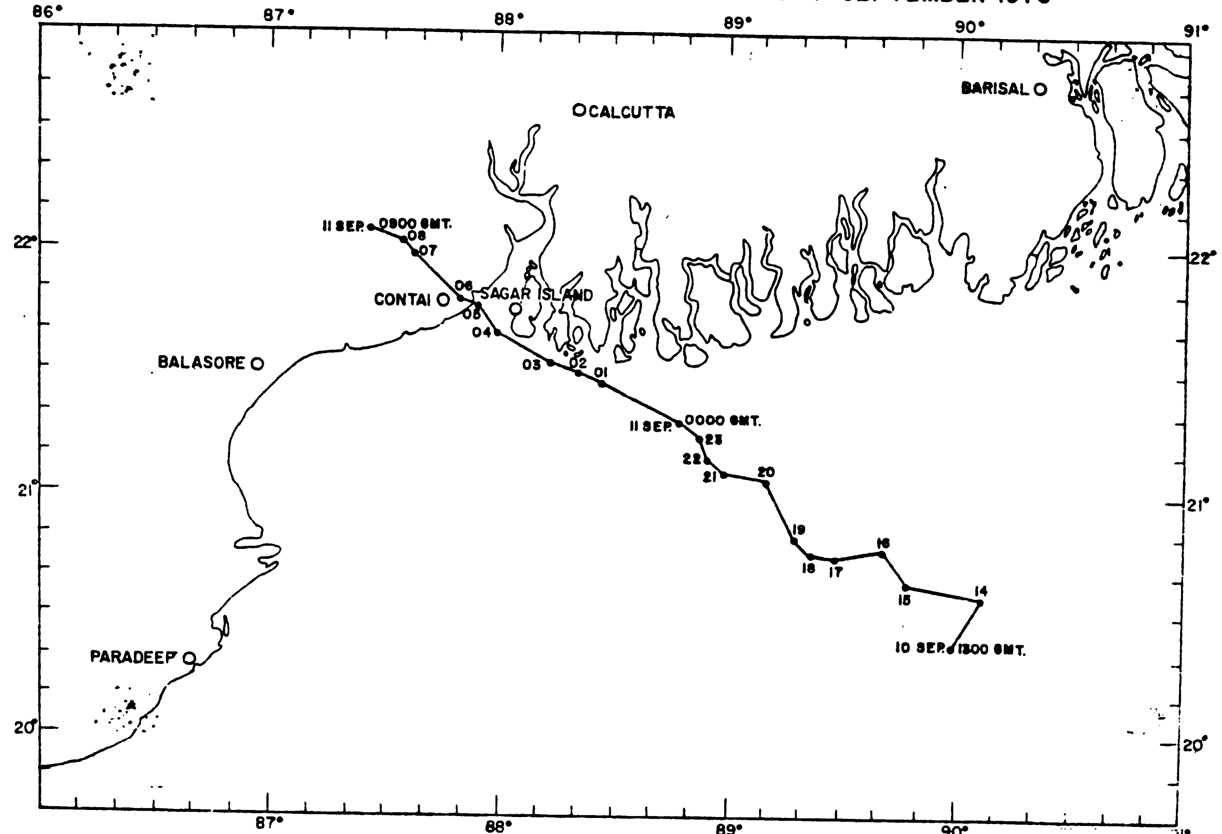


FIG.38.

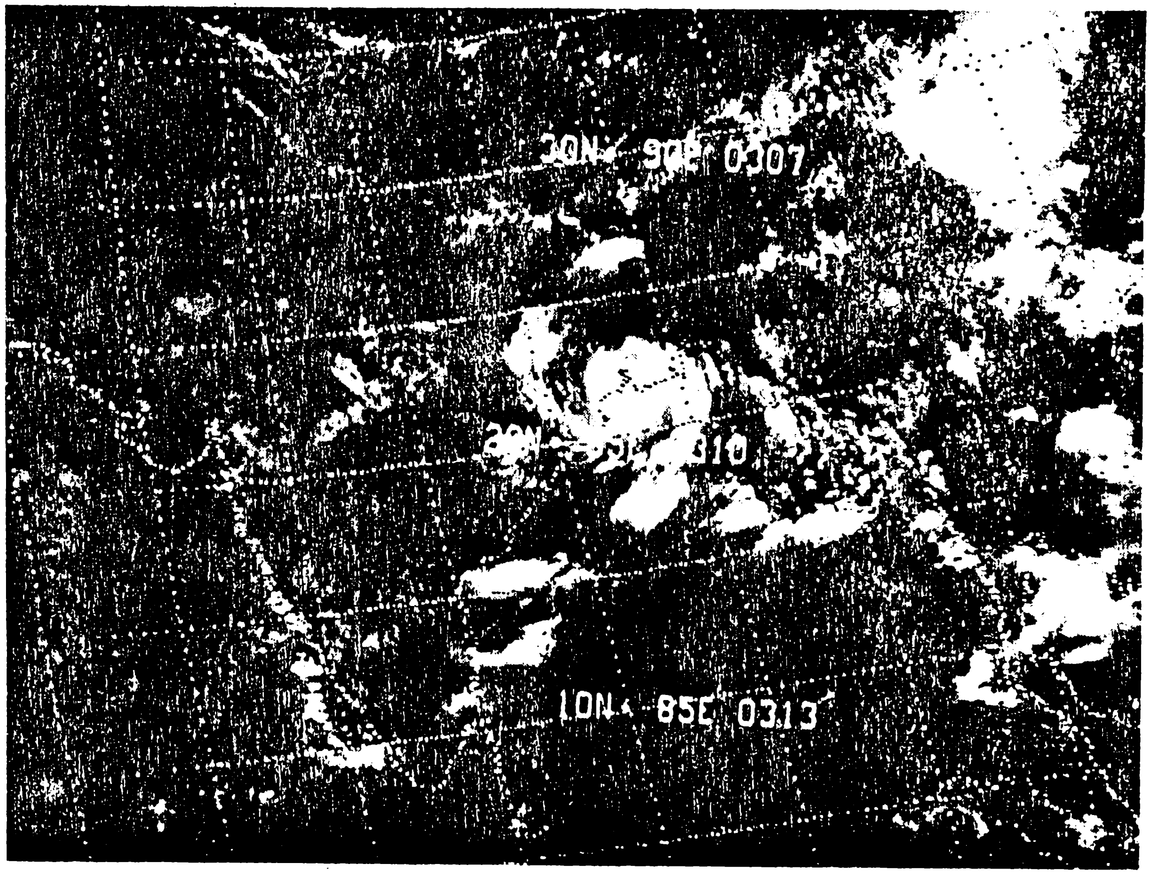


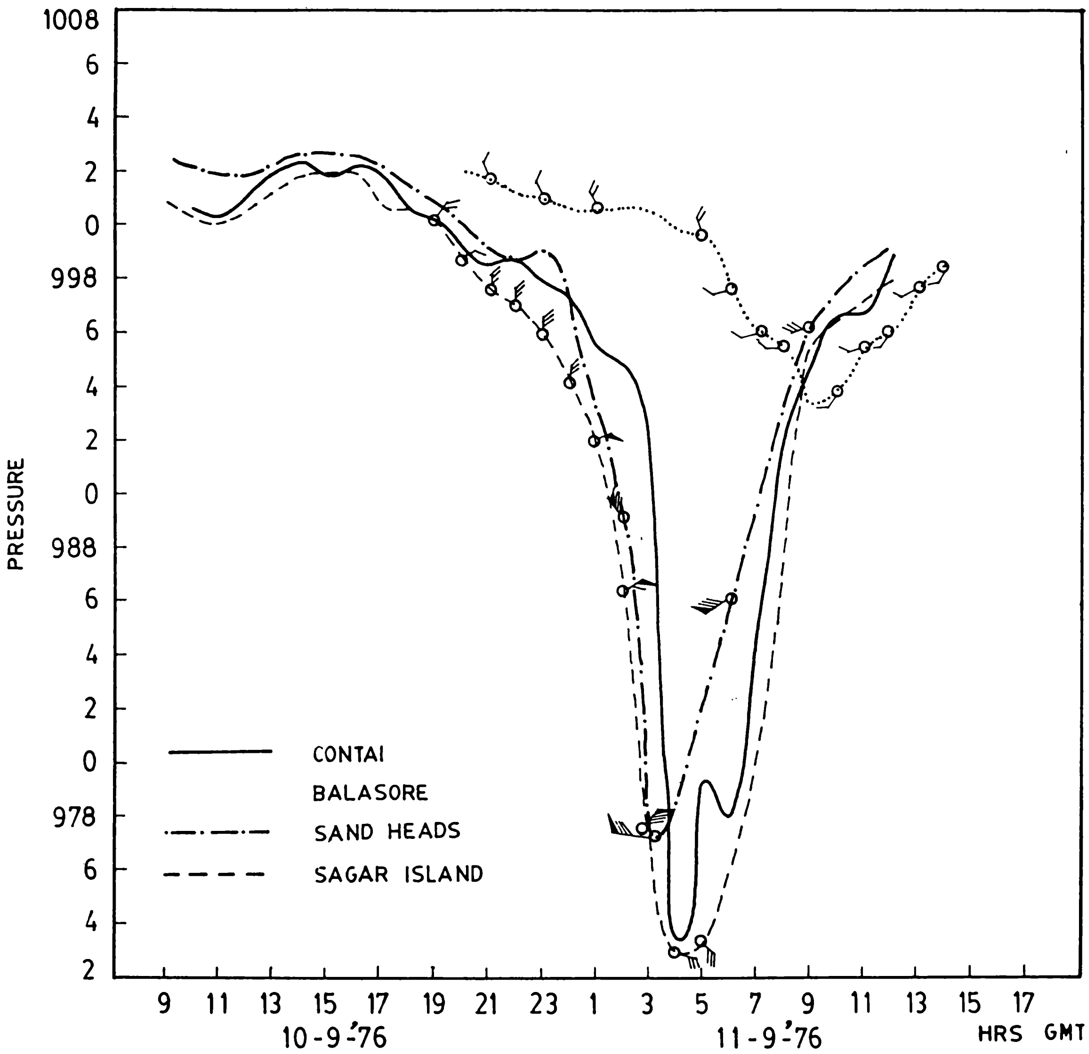
Fig 39.: Satellite view of Contai cyclone at about 0300 GMT on 11 September showing the eye of the cyclone

intensity reduced to cyclonic storm by the night of 11th, it retained that intensity on land upto 13th morning. Thereafter, the storm weakened to deep depression over Bihar Plateau and dissipated on 19th evening.

2.4.2. Wind and pressure reports of the cyclone

The storm had a core of hurricane winds on 11th as revealed by the satellite picture (Fig.39). The hourly observations of wind and pressure reported by Sagar Island, Sandheads, Contai and Balasore during the period 10.9.1976 to 11.9.1976 is shown in Fig.40. Sagar Island and Sandheads reported maximum surface winds of 70 to 80 kt on 11th. The maximum gusty winds reached 85 kt at Sagar Island on 11th by about noon. These two stations also reported the maximum pressure departure from normal as minus 26 to 28 mb on 11th. This cyclone passed close to Sagar Island at 0930 IST on 11th (Fig.38) and the lowest pressure recorded at this place is 973 mb at 0830 IST on this day (Fig.40). The maximum wind estimated from the satellite picture (Fig.39) on Dvorak scale is 70 kt. The pressure at the centre of the storm for a maximum wind of 80 kt works out to be 977 mb (Natarajan and Ramamurty, 1975), while the actual pressure reported was 971.5 mb on 11th by Sagar Island near the storm centre.

Fig. 40. HOURLY OBSERVATIONS OF WIND AND PRESSURE AT GANGETIC COAST FROM 10-9-'76 TO 11-9-'76.



The computed winds are obtained from the pressure distribution of the cyclone at 0830 and 1730 IST positions on 11th which are shown in Figs.41 and 42 respectively. These winds are used wherever required to evaluate the surge. A comparison of the observed and computed winds is shown in the table VIII.

2.4.3. Surge at landfall

Figs.43 and 44 show the grids and bathymetry used to calculate the wind setup along the three paths AB, CD and EF. The magnitude of wind setup at the landfall is shown in Fig.45. The wind setup along AB, CD and EF is almost gradual and similar. The peak is reached slightly earlier along AB and CD than along EF. The occurrence of peak surge at along EF coincides almost with the landfall. The fall of the surge appears to be very abrupt. The magnitude of the peak surge along AB, CD and EF is 113, 126 and 138 cm respectively.

While the computed peak surge (wind setup) is only 1.38 m the reported values are more than 2 m (Pant *et al.*, 1978). During the period of the cyclone there was a recorded rainfall of 29 cm. Perhaps, a tidal bore was induced in Hooghly estuary due to heavy discharge of rain water against the insurgence of the cyclone resulting

Table VIII. Observed and computed winds on 11 September 1976

Sl. No.	Observed values						Computed values						
	Time (IST)	Ship/ coastal station/ other source	Position: Lat. ^{o}N	Position: Long ^{o}E	Observed wind: Direction	Observed wind: Speed (kts)	Time of weather chart	Grid point: Lat ^{o}N	Grid point: Long ^{o}E	p (mb)	N (km)	h (km)	V_g (kts)
1.	0530	TTKA	18.6	87.2	WNW	23	0830	18–19	87–88	2	90	240	29.6[illegible]
2.	0530	ATOQ	18.7	85.6	WSW	32	0830	18–19	88–89	2	60	225	37.2[illegible]
3.	–	From satellite picture	–	–	–	Core of hurricane winds (80 kts)	0830	20–21	89–90	10	45	120	82.3[illegible]
4.	1200	Sagar Island and Sangheads	–	–	–	70–80	0830	20–21	88–89	10	45	75	66.0[illegible]
5.	–	From satellite picture	–	–	–	Core of hurricane winds (80 kts)	0830	20–21	87–88	10	45	150	91.3[illegible]
6.	1200	Sagar Island and Sandheads	–	–	–	70–80	0830	21–22	87–88	10	60	165	81.3[illegible]

Fig. 41. 0830 IST POSITION ON 11 SEPT 1976.

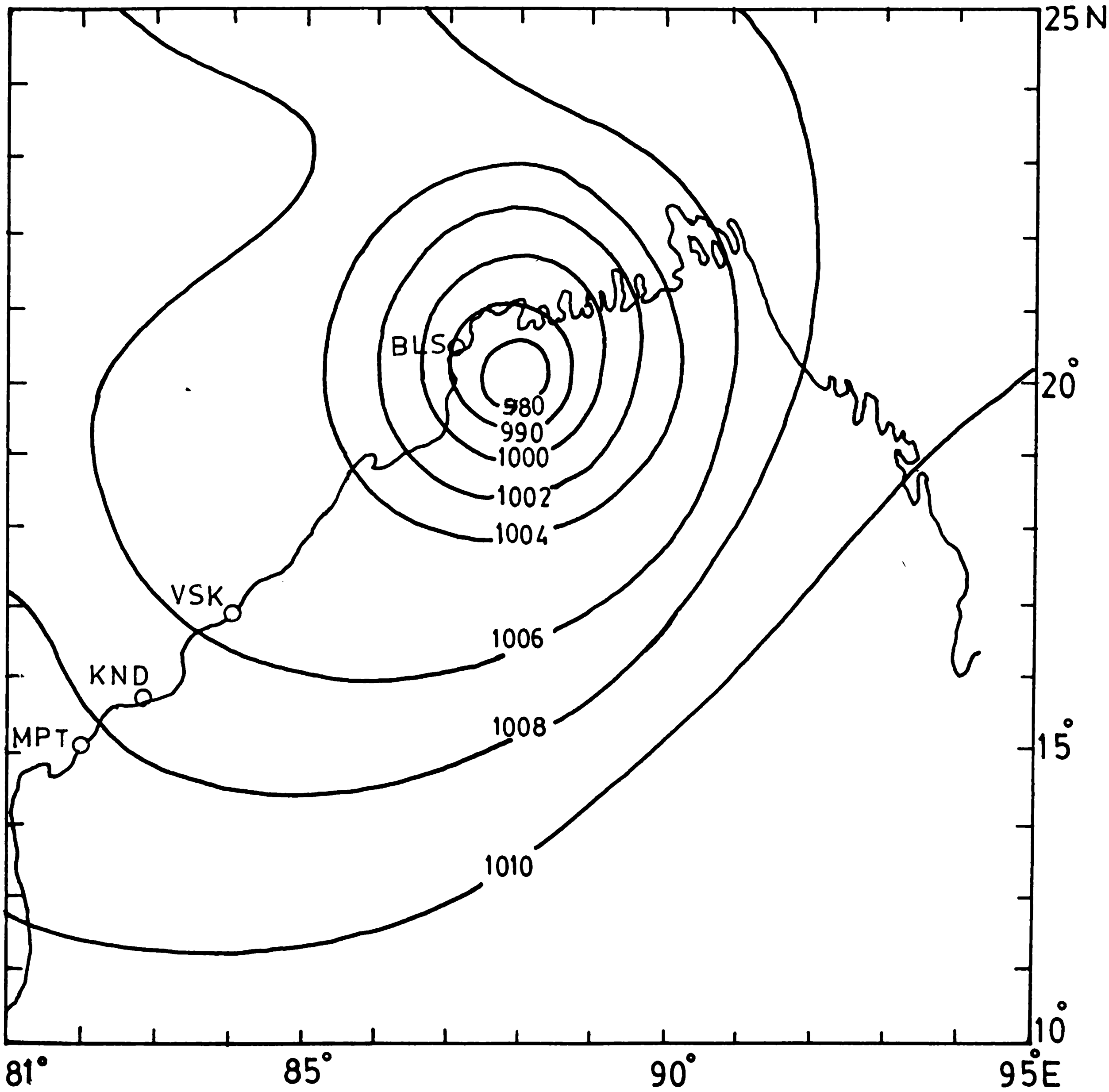


Fig. 42. 1730 IST POSITION ON 11 SEPTEMBER 1976.

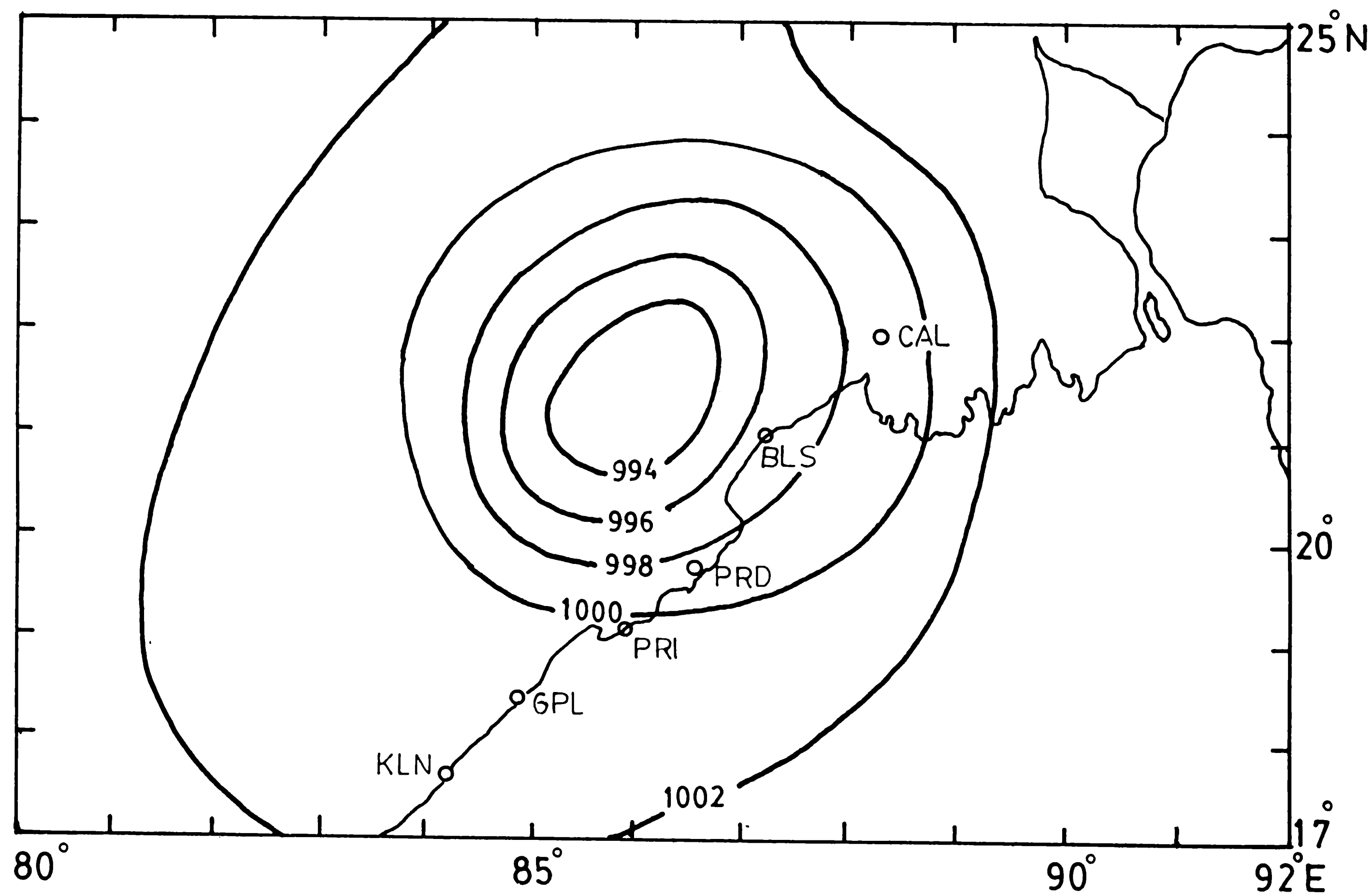


Fig.43.GRID USED FOR THE CYCLONE OF 8-19 SEPT 1976.(- - -) CYCLONE TRACK.

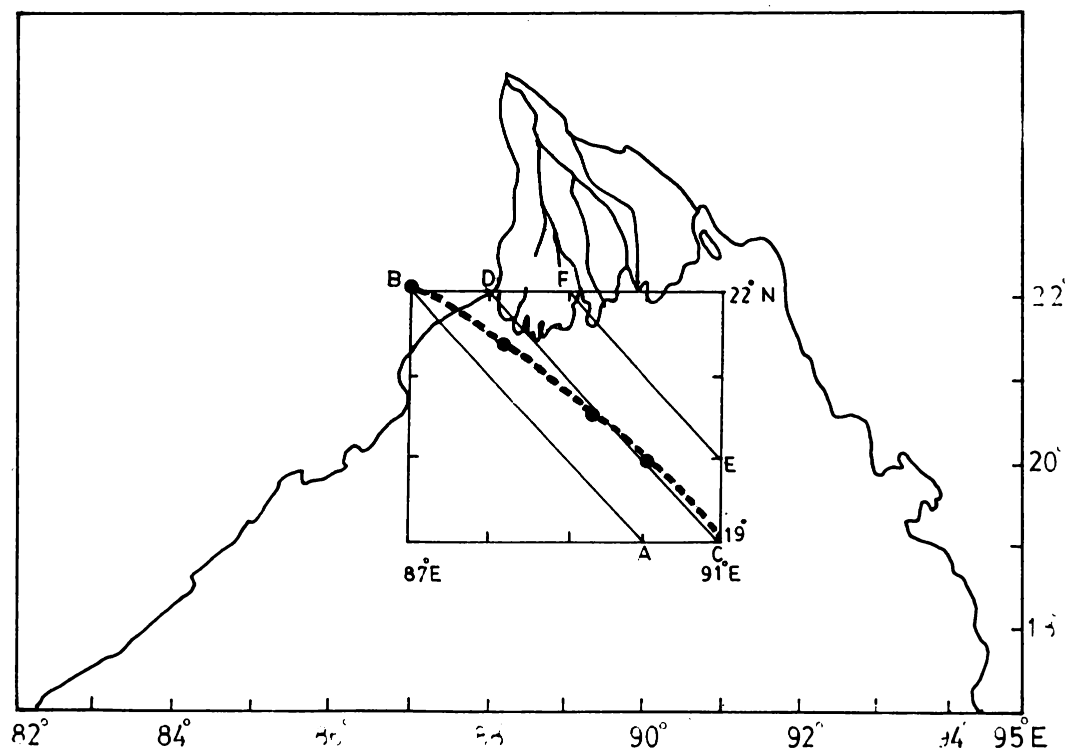


Fig. 44. BATHYMETRY BETWEEN KAKINADA AND CALCUTTA.

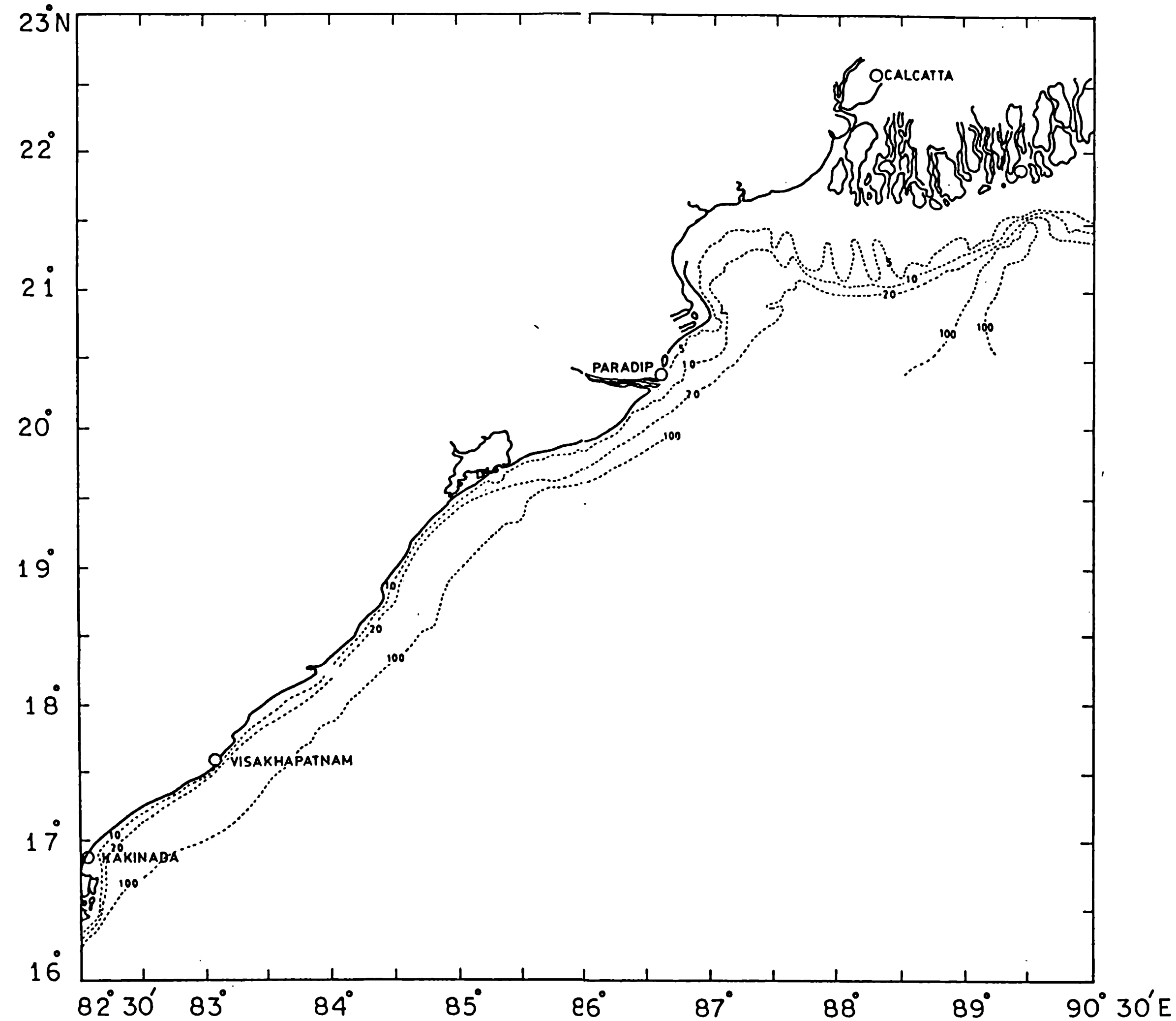
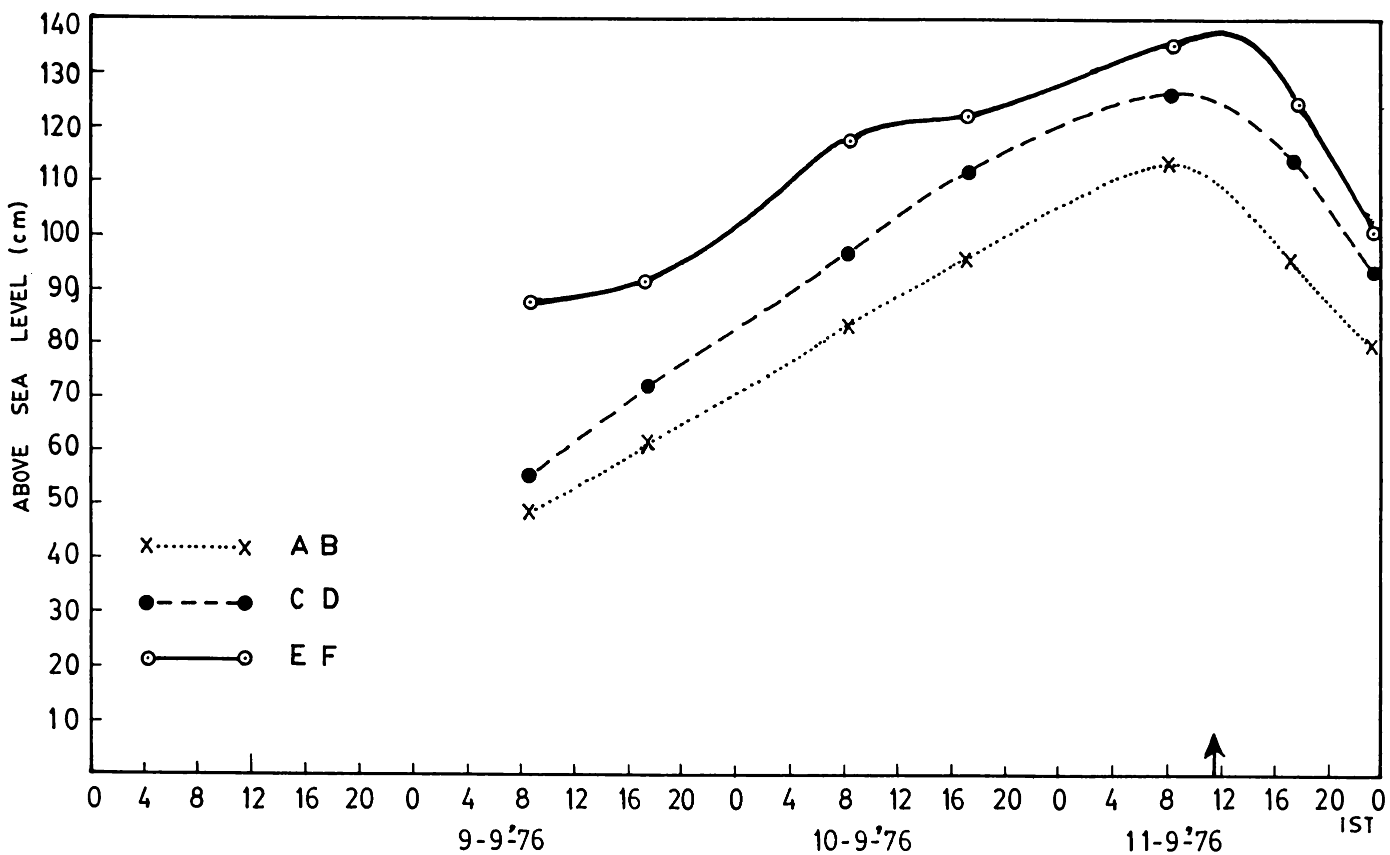


Fig. 45. STORM SURGE ALONG AB, CD AND EF OF 9-9-'76 TO 11-9-'76 CONTAI CYCLONE.



in the overestimate of the surge than that calculated from winds.

2.4.4. Tide and surge at Sagar Island

For the period 8.9.1976 to 19.9.1976 observed, predicted, residual and mean residual tides are presented in Figs.46 to 48. The mean residual sea level shows maximum values of 0.5 m at 2000 IST on 8th and during 0800 to 2000 IST on 11th respectively and a negative sea level of 0.49 m during 0000 to 0800 IST on 10th (Fig.46). The highest rise in sea level of 1.1 m at 0000 IST on 12th (Fig.47) is conspicuous. The sea level rise is not significant during 16th to 19th (Fig.48). It is highlighted to note here that the computed surge at landfall (1.38 m) from Fig.45 and the highest level (1.1 m) obtained from tide curves (Fig.47) agreed very well.

Fig. 46 /TIDE AND SURGE OF SAGOR ISLAND FOR THE PERIOD 8-9-'76 TO 11-9-'76.

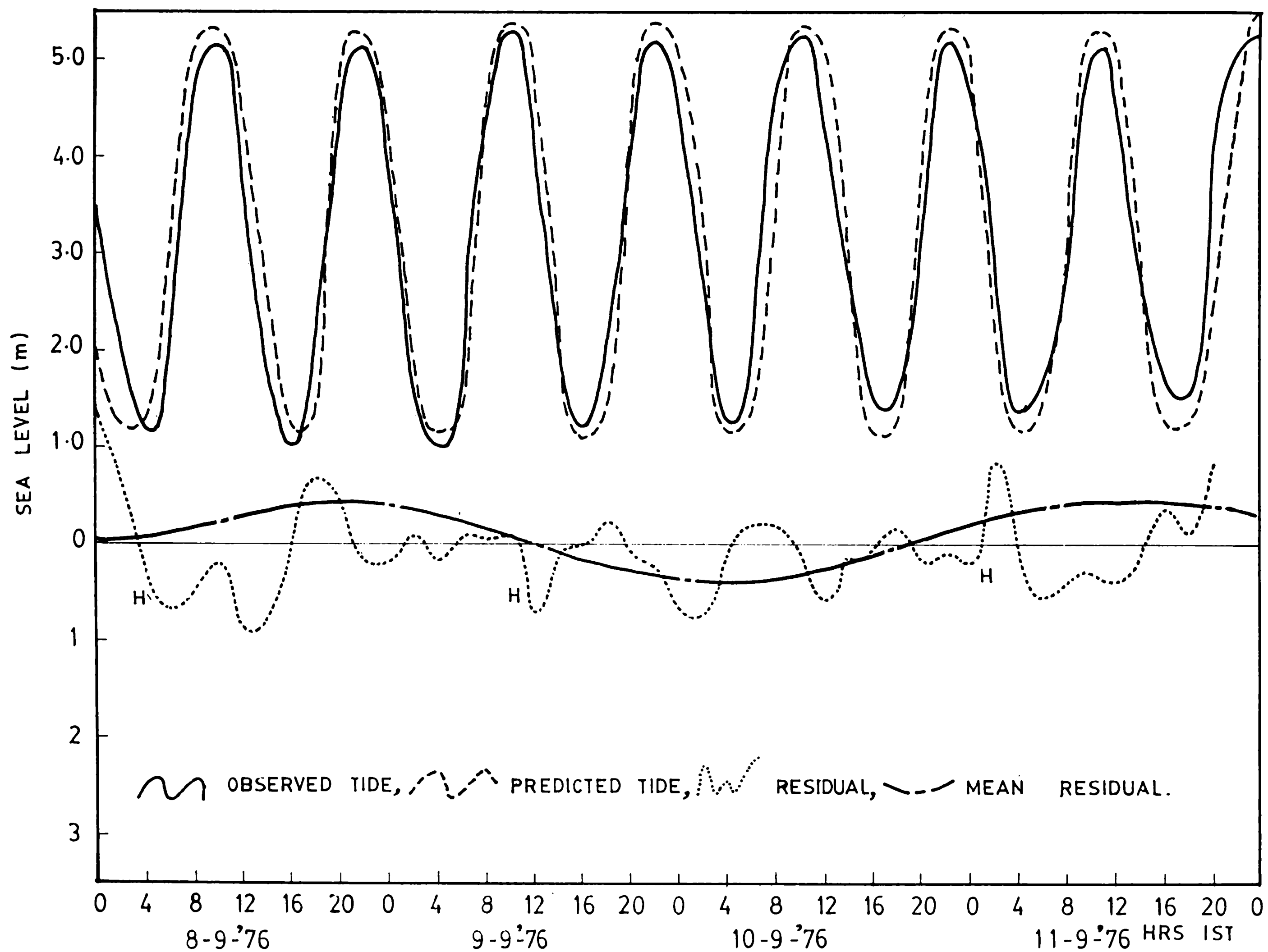


Fig. 47. TIDE AND SURGE OF SAGOR ISLAND FOR THE PERIOD 12-9-'76 TO 15-9-'76.

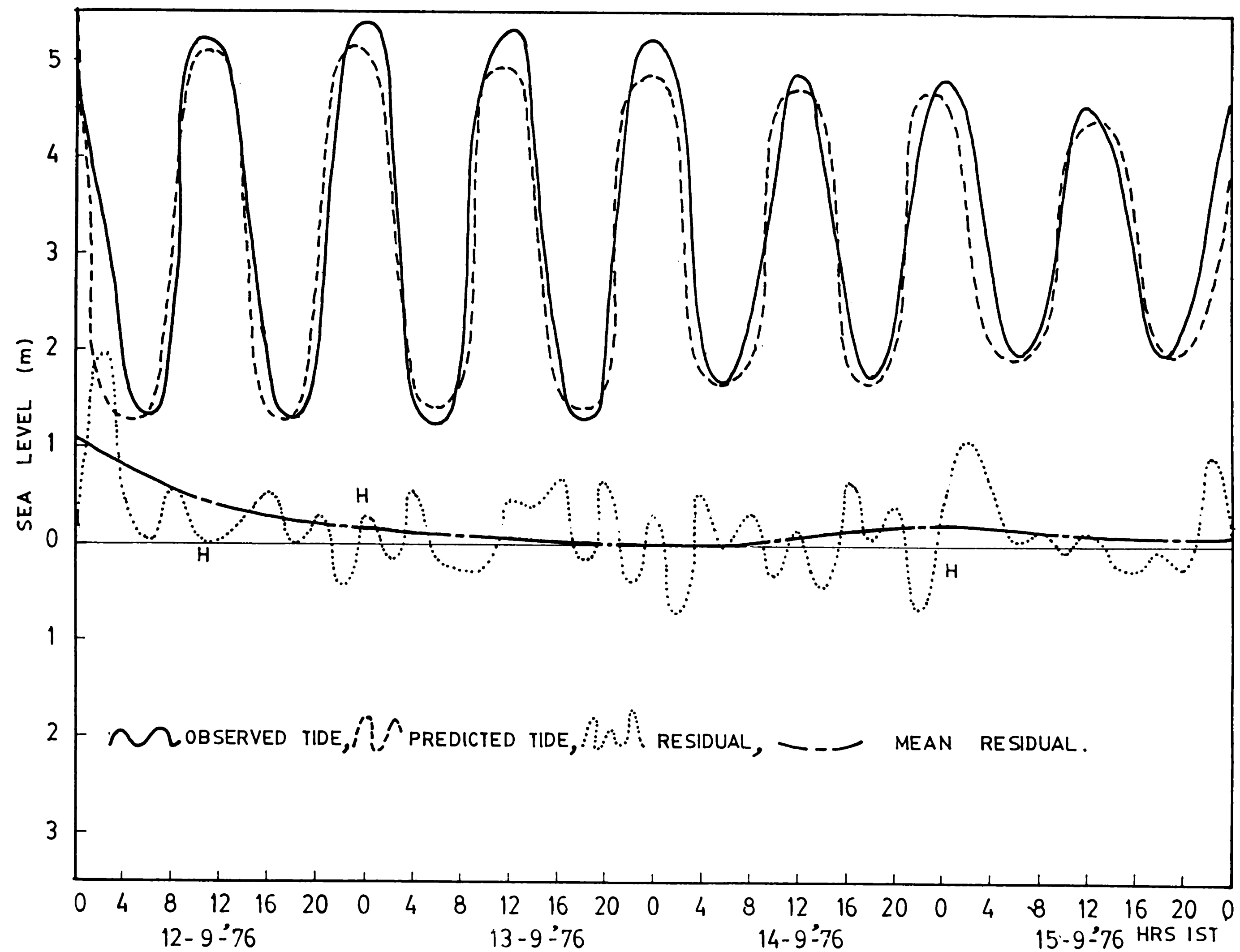
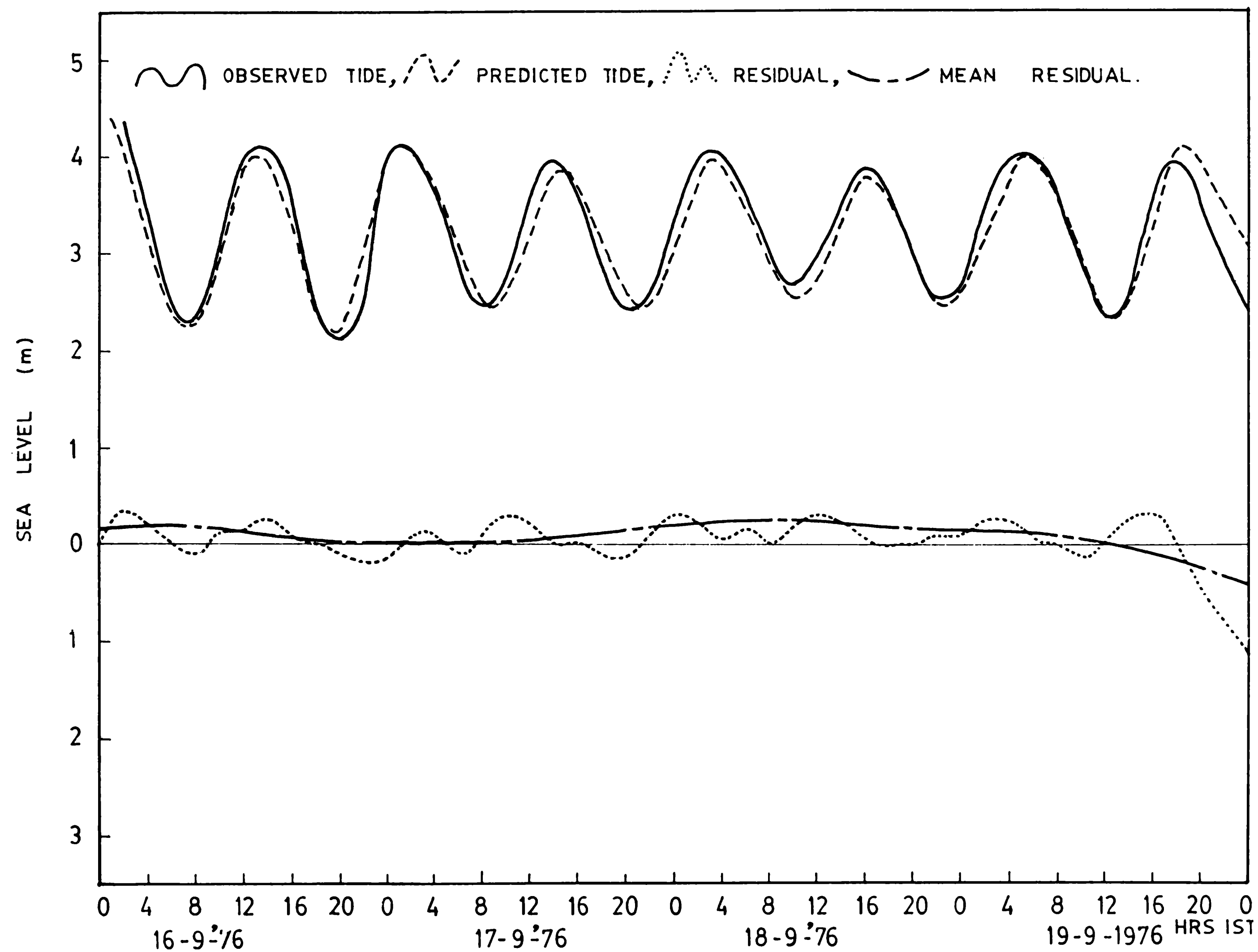


Fig. 48. TIDE AND SURGE OF SAGOR ISLAND FOR THE PERIOD 16-9-76 TO 19-9-1976.



CHAPTER III

SEA LEVEL VARIATIONS ALONG THE EAST COAST OF INDIA.

CHAPTER - III

SEA LEVEL VARIATIONS ALONG THE EAST COAST OF INDIA

Variations in sea level are caused by oceanographic, meteorologic, hydrologic, geologic, seismologic and eustatic factors. The first two factors are important for the seasonal variations, whereas, the others for the long term variations of sea level. A knowledge of the fluctuations in the mean sea level finds application in the accurate prediction of storm surges and basin oscillations, coast and harbour development, and coastal zone management.

Considerable literature is available on the mean sea level and its seasonal and annual changes in many parts of the world oceans (Pattullo *et al.*, 1955; Lisitzin and Pattullo, 1961; Lisitzin, 1963,1972,1974; Rossiter, 1967,1968,1972; Fairbridge, 1961,1962,1963). Whereas the information pertaining to the Indian Ocean on this subject is scanty, the works of Prasada Rao and LaFond (1953,1954), Ramanadham and Varadarajulu (1964), Ramanadham *et al.* (1965), and Varadarajulu and Dhanalakshmi (1975) concern the sea level variations in the Bay of Bengal bordering India and those of Rama Raju and Hariharan (1967), Nair (1974) and Kesava Das (1979) pertain to the west coast. The study of Chugh (1961) deals with the secular variations along both the coasts.

3.1. Available studies in the Indian region

3.1.1. Along east coast

The first attempt in the study of sea level variations in the Indian region was made by Prasada Rao and LaFond (1953,1954). They studied the annual sea level variations of Bay of Bengal and found, the amplitude of the cycle is the largest at Kidderpore with a range amounting to 170 cm (Prasada Rao and LaFond, 1953). The highest values are observed in July and August, and the lowest in January and February. The changes in seasonal sea level at Visakhapatnam are considerable mainly due to the effects of vertical distribution of density of the sea water and the atmospheric pressure difference between Visakhapatnam and Victoria Point (Prasada Rao and LaFond, 1954).

Chugh (1961) presented the annual and secular variations of sea level in the Arabian Sea and Bay of Bengal. The amplitudes and phase lags of the 19 years tide (nodal tide) for four representative ports have also been discussed.

Later, while discussing the fluctuations in the monthly mean sea level at Visakhapatnam, Ramanadham and Varadarajulu (1964) concluded that these fluctuations can be attributed to the heavy concentrated rainfalls, and winds blowing consistently in a particular direction.

Ramanadham *et al*. (1965) indicated that sea level is constant at Waltair over a long period but increases towards the northern stations Sagar Island, Garden Reach and Diamond Harbour along east coast of India amounting to about 30,17 and 12 cm respectively. They also inferred that the sea surface is slightly tilted up along the western Bay of Bengal from Visakhapatnam to north during 20th century, which may be related to the gradual change in water mass characteristics of reduced densities at the head of the Bay of Bengal and increase of density in the southwestern part with a neutral zone off the Visakhapatnam Coast.

Varadarajulu and Dhanalakshmi (1975) explained the monthly and annual mean sea levels off Madras in relation to the physical properties of sea water, wave characteristics and climate in the neighbourhood. According to them the tide range at Madras varies from 15 cm for neaptides to 128 cm for spring tides and these values are relatively smaller than those of the northern stations along the east coast of India.

3.1.2. Along west coast

Dirunal, seasonal and secular variations of sea level along the west coast of India were studied by RamaRaju and Hariharan (1967). According to them the diurnal and seasonal variations of sea level at Cochin showed an annual range of 45 cm (18 inches) with most of the variation occurring during August to December. The short period fluctuations in sea level observed during the southwest monsoon bear a close relation with the associated rainfall and wind. The general trend of secular variation along the west coast of India during this century is a greater increase in sea level in the northern region than in the south. The rates of increase at Bhaunagar and Bombay are 1.77 cm/yr and 0.03 cm/yr, respectively.

Later, the rise in sea level along the west coast was confirmed by Nair (1974) while studying the submerged terraces on the western continental shelf between Bombay and Karwar on the basis of radiocarbon dating. He pointed out that these submerged terraces are of Holocene age and the rise in sea level along the west coast is, possibly, due to tectonic instability, which in turn, may be due to the influence of the downwarping of the Indus Cone on the shelves bordering the Arabian Sea.

The seasonal variation in mean sea level at Mormugao was studied by Kesava Das (1979), who observed a seasonal trend in the monthly sea level, with low values during the southwest monsoon and high values in the fair weather period. The factors responsible for these variations were understood to be the coastal currents, water density and wind; the direct effect of wind for the variation of mean sea level appeared to be relatively insignificant.

3.2. Seasonal variation of sea level along the east coast of India

The monthly mean sea level variations at Visakhapatnam and Madras, Dublet and Sagar, Kidderpore, and Diamond Harbour and Calcutta is presented in Figs.49 to 52. The periods of data used in the estimation of sea level variations for the different stations are indicated at appropriate places.

3.2.1. Sea level variation at Madras and Visakhapatnam

Madras and Visakhapatnam constitute the southern most part of the area of study. At both the places (Fig.49)

Fig. 49. LONG TERM MONTHLY MEAN SEA LEVEL VARIATION.

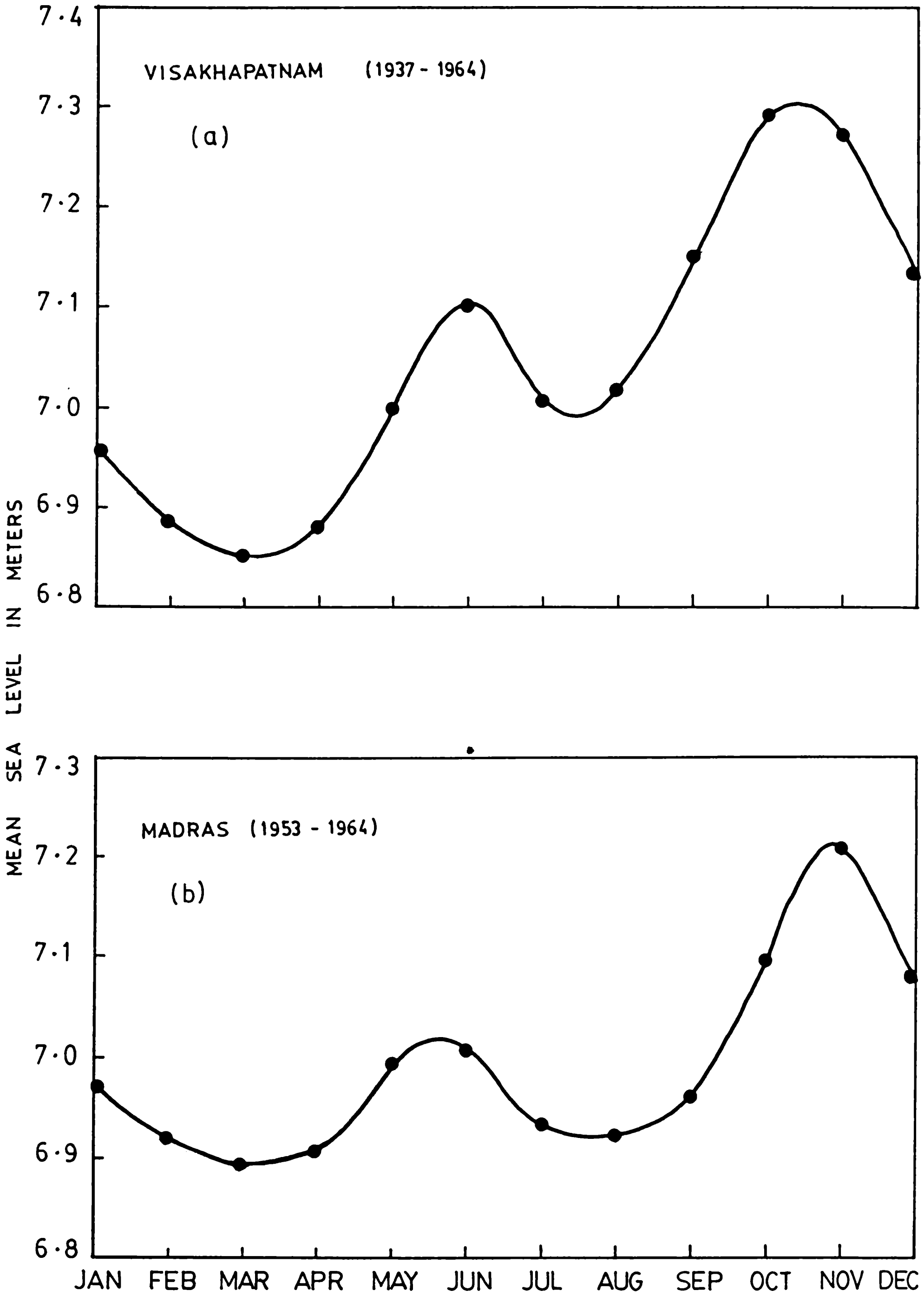


Fig.50.LONG TERM MONTHLY MEAN SEA LEVEL VARIATION FROM 1937-'64.

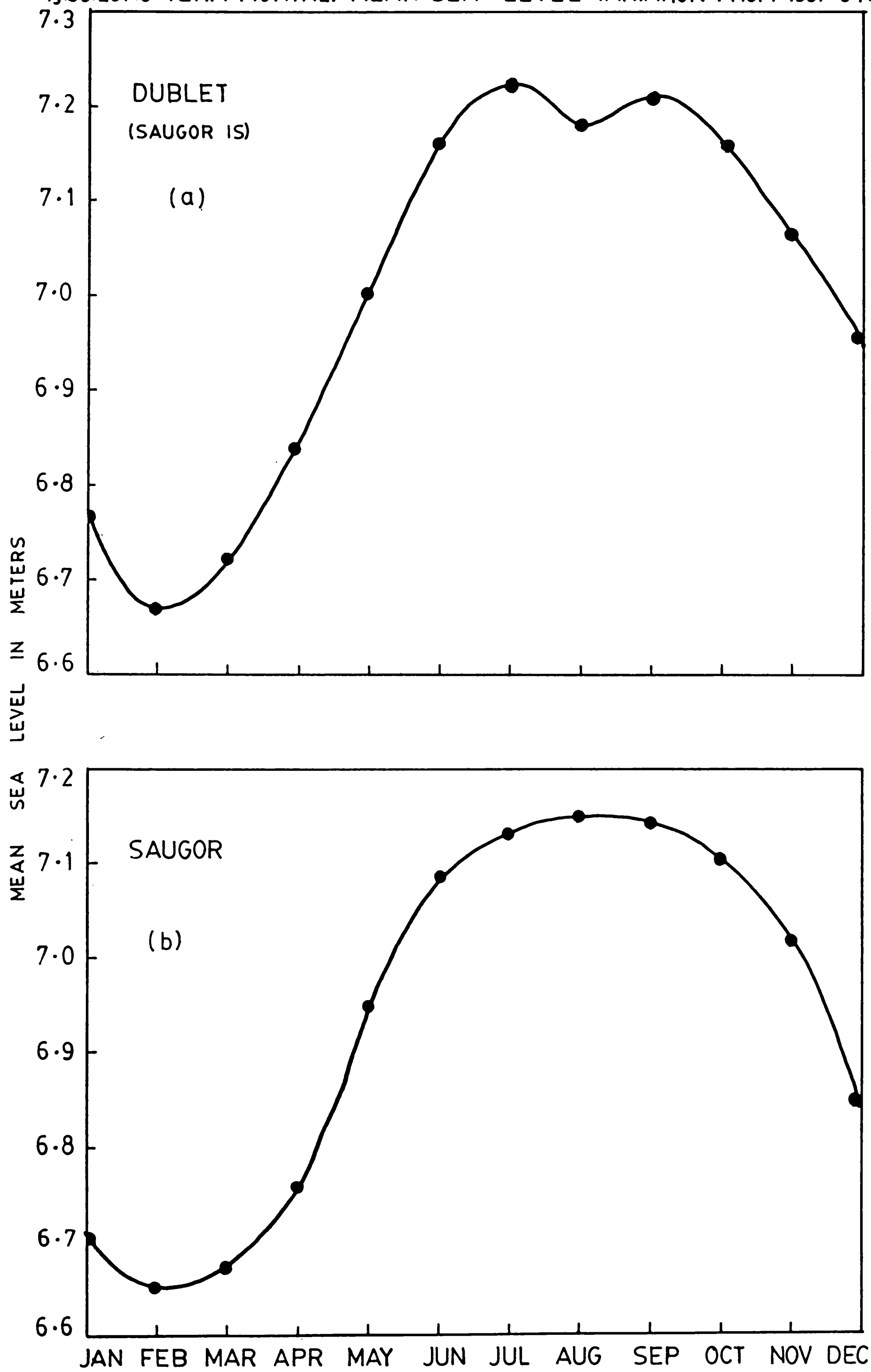


Fig. 51. LONG TERM MONTHLY MEAN SEA LEVEL VARIATIONS AT KIDDERPORE HARBOUR FROM 1881 TO 1931.

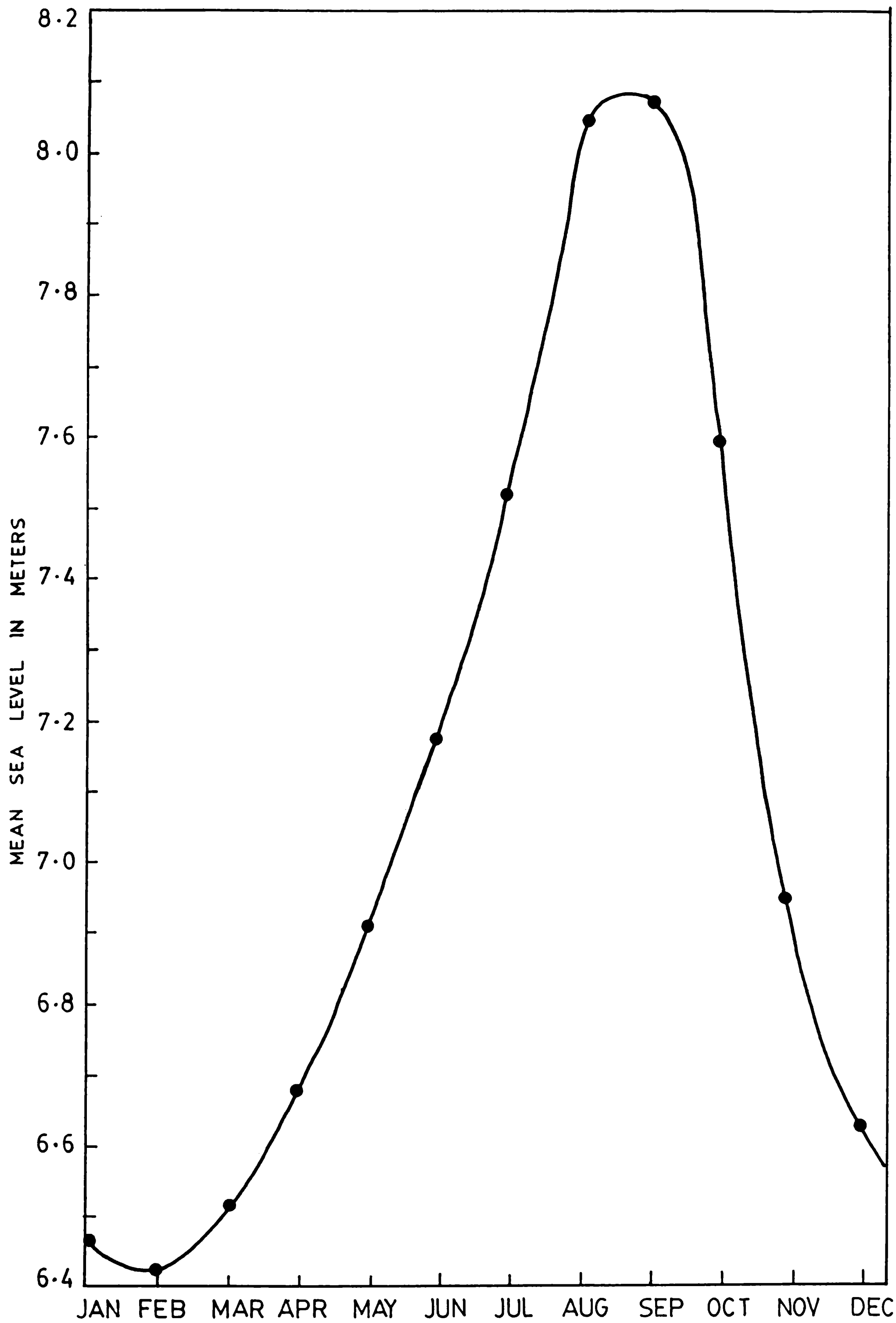
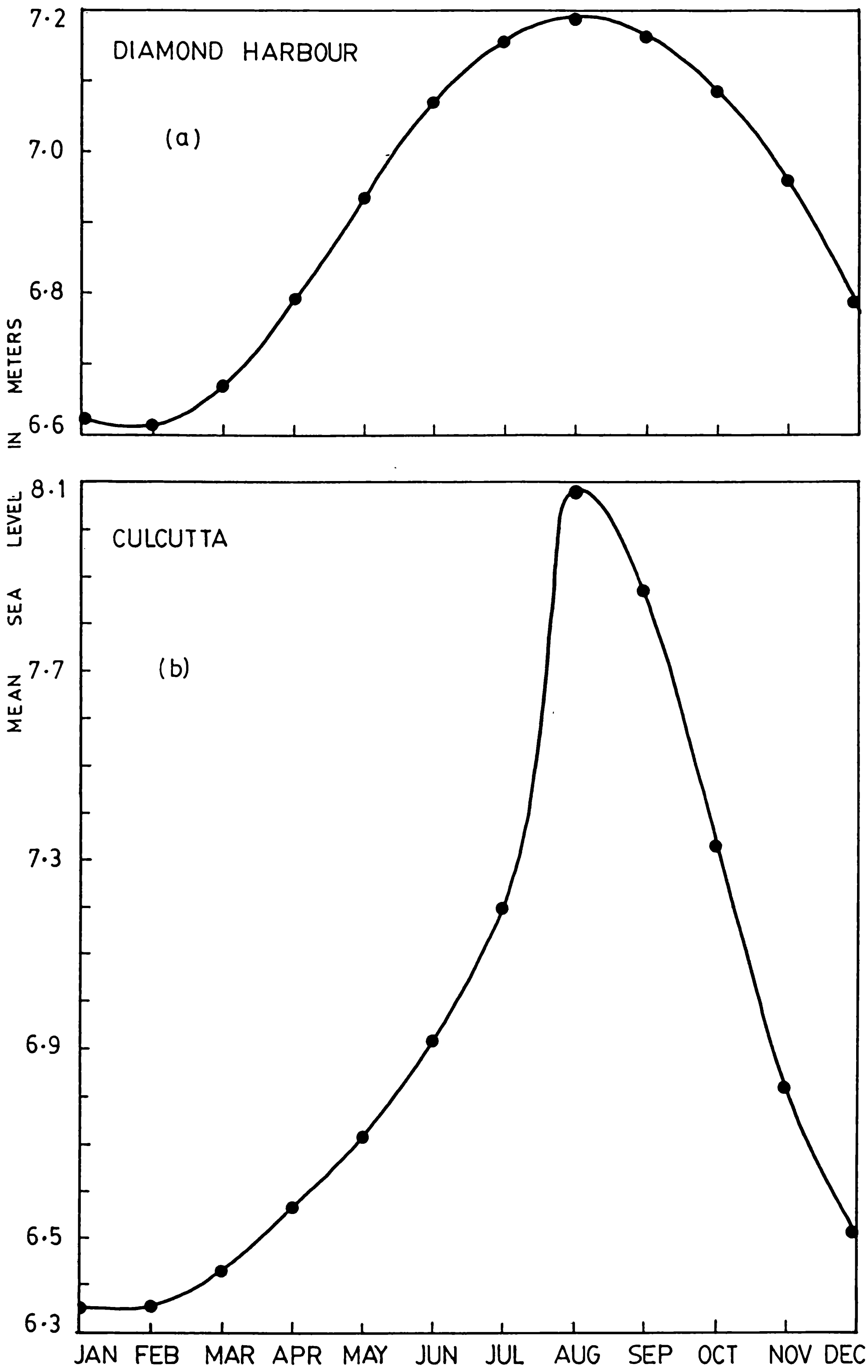


Fig.52.LONG TERM MONTHLY MEAN SEA LEVEL VARIATION FROM 1948-64.



the variation is of bi-annual nature with a primary maximum occurring during October-November and the secondary during May-June. The primary and secondary minima occur during March-April and July-August respectively. It is conspicuous to note that the primary maximum occurs slightly earlier at Visakhapatnam than at Madras while the primary minimum at Madras precedes that at Visakhapatnam. The secondary maxima and minima at both the places occurs almost simultaneously. The primary maxima at both the places is due to the northeast monsoon winds that drive the oceanic waters against the coast, and since they (onset of northeast winds) occur earlier at Visakhapatnam than at Madras, primary maximum occurs at Visakhapatnam. Besides, it is also aided by the process of sinking which commences along the east coast of India by about September. The primary minima at these places is the consequence of the southwest monsoon winds which have an offshore component along the east coast of India. This corroborates with the sequence of occurrence of the primary minima at Madras and Visakhapatnam respectively, as the onset of the southwest monsoon winds cross Madras earlier than Visakhapatnam. The secondary maxima and minima are due to the interruptions in the sea level by steric effects that are caused by upwelling, river discharge and rainfall.

The sea level decreases with upwelling (February-April), while it increases with river discharge and rainfall (May-July and October-December) at Madras and Visakhapatnam (Fig.49).

The range of sea level variation at Visakhapatnam is about 0.45 m and about 0.3 m at Madras (Fig.49).

3.2.2. Dublet and Sagar

At both the places the variation of sea level shows (Fig.50) a sinusoidal nature with higher values during July to September and lower values in February. At Dublet, in spite of the higher values during July to September, there is actually a decrease in sea level from July to August and an increase from August to September. The reason for such a variation is not immediately apparent; however, higher sea level during July-September may be due to the transport of sea water against the land by the onshore southwest monsoon winds and rainfall. The minimum sea level in February at both the places appears to be the consequence of the offshore northeast monsoon winds. The tidal range at both the stations is almost the same with a value of about 0.5 m (Fig.50).

3.2.3. Kidderpore

Fig. 51 indicates a single maximum and a minimum occurring during August-September and February respectively. The annual range of sea level at this place appears to be the highest with a value of 1.65 m. The rate of increase, as well as decrease in sea level, is very high.

3.2.4. Calcutta and Diamond Harbour

The sea level at Calcutta and Diamond Harbour is highest during August while it is lowest in February with a higher rate of increase at Calcutta than at Diamond Harbour (Fig.52).

3.3. Comparison of the seasonal sea level variation at different stations along the east coast of India

While the seasonal sea level variation in the south (Madras and Visakhapatnam, Fig.49) is semi-sinusoidal (bi-annual), it is sinusoidal (annual) in the north (Figs.50 to 52). The sea level is significantly higher at Calcutta and Kidderpore with a maximum value of around 8.1 m. It is also interesting to note that the rate of increase as well as decrease of sea level with time is the highest at these two places. The occurrence of higher sea level shifts gradually from July-August in northern-most to November in the southern most region, under study. While the causative factor for the rise of sea level in the north is the southwest monsoon winds associated with the heavy river discharge and precipitation run-off, it is the northeast monsoon winds that drive the oceanic water against the coast and the associated rainfall, in the south. The interesting difference of the semi-sinusoidal nature in the south (Fig.49) is mainly due to steric effect that is caused by upwelling in the southern part of the region, while such a phenomenon is not present in the northern region.

According to Prasada Rao and LaFond (1953) the annual variation of sea level along the east coast is maximum at Kidderpore amounting to about 1.7 m, and agrees well with the present investigation. Also, the occurrence of the highest sea level in July coincides with the observation of Prasada Rao and La Fond (1953). However, the lowest sea level is noticed to occur in February in the present study, while they observed it in January.

3.4. Secular variation of mean sea level along the east coast of India

The secular variation of mean sea level is worked out for Madras, Visakhapatnam, Dublet, Sagar Island, Diamond Harbour, Kidderpore and Calcutta from the available mean sea level data, using the five year moving average technique. The results are presented in the Figs.53 to 55.

3.4.1. Madras and Visakhapatnam

The secular variation of mean sea level, for the period 1937-1964 at Visakhapatnam and for the period 1953-1964 for Madras, is shown in Figs.53 a and b. Secular variation for Madras for the earlier period could not be worked out due to the non-availability of data.

The sea level at Madras (Fig.53b) decreases very slowly between 1955-1957 and then increases with the same order of magnitude. At Visakhapatnam (Fig.53a) the sea level in general shows an increasing tendency from 1937-1957 and from thereafter it decreases. The sudden decrease of sea level during 1961-1962 may not be of much consequence, as it may be due to the boundary effect in the moving average technique employed.

3.4.2. Dublet and Kidderpore

Figs.54 a and b depict the secular variation of mean sea level at Dublet and Kidderpore. Unfortunately,

Fig. 53. LONGTERM ANNUAL VARIATIONS ALONG THE EAST COAST OF INDIA.

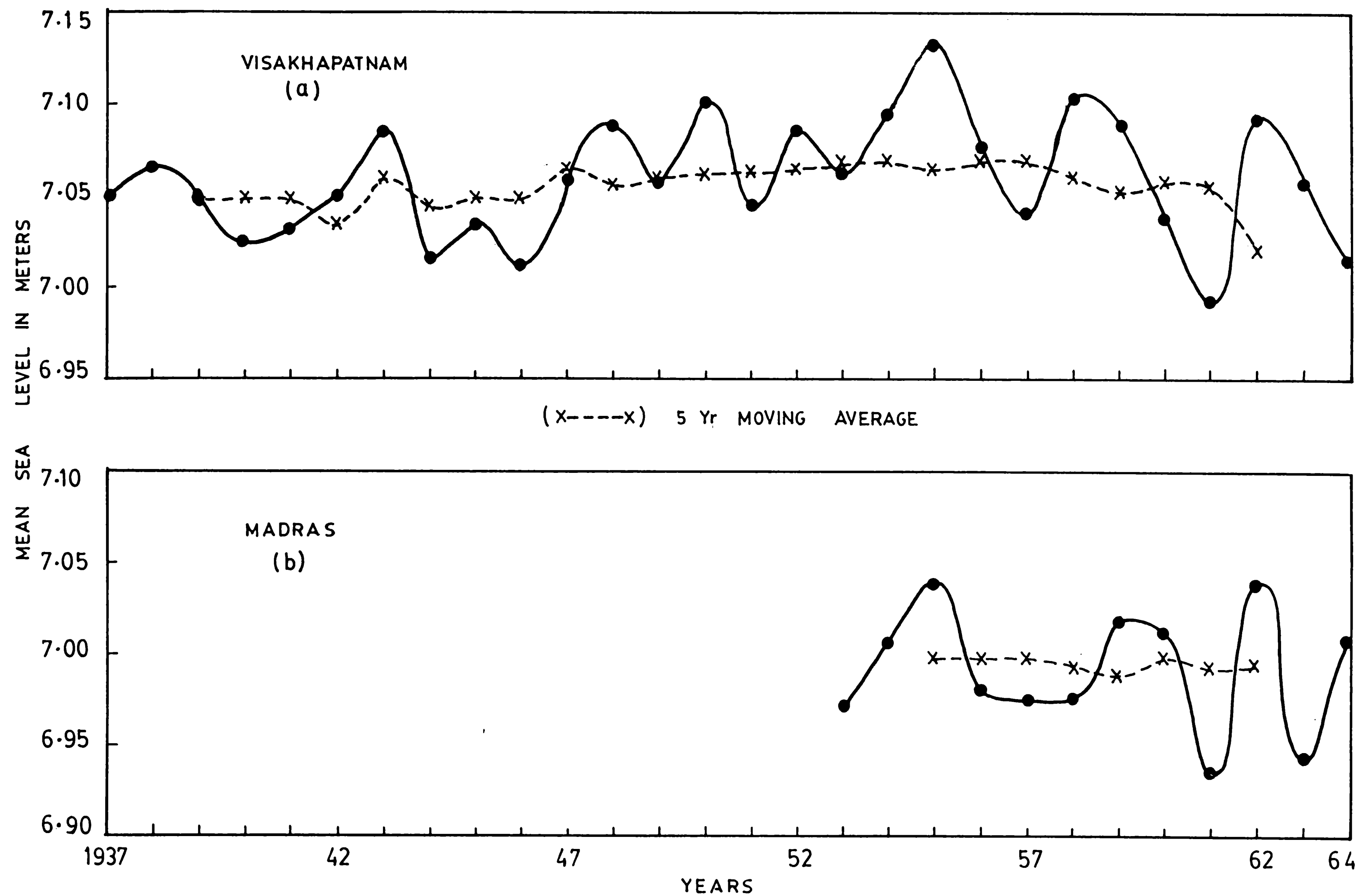


Fig. 54. LONGTERM ANNUAL VARIATIONS OF SEA LEVEL ALONG THE EAST COAST OF INDIA.

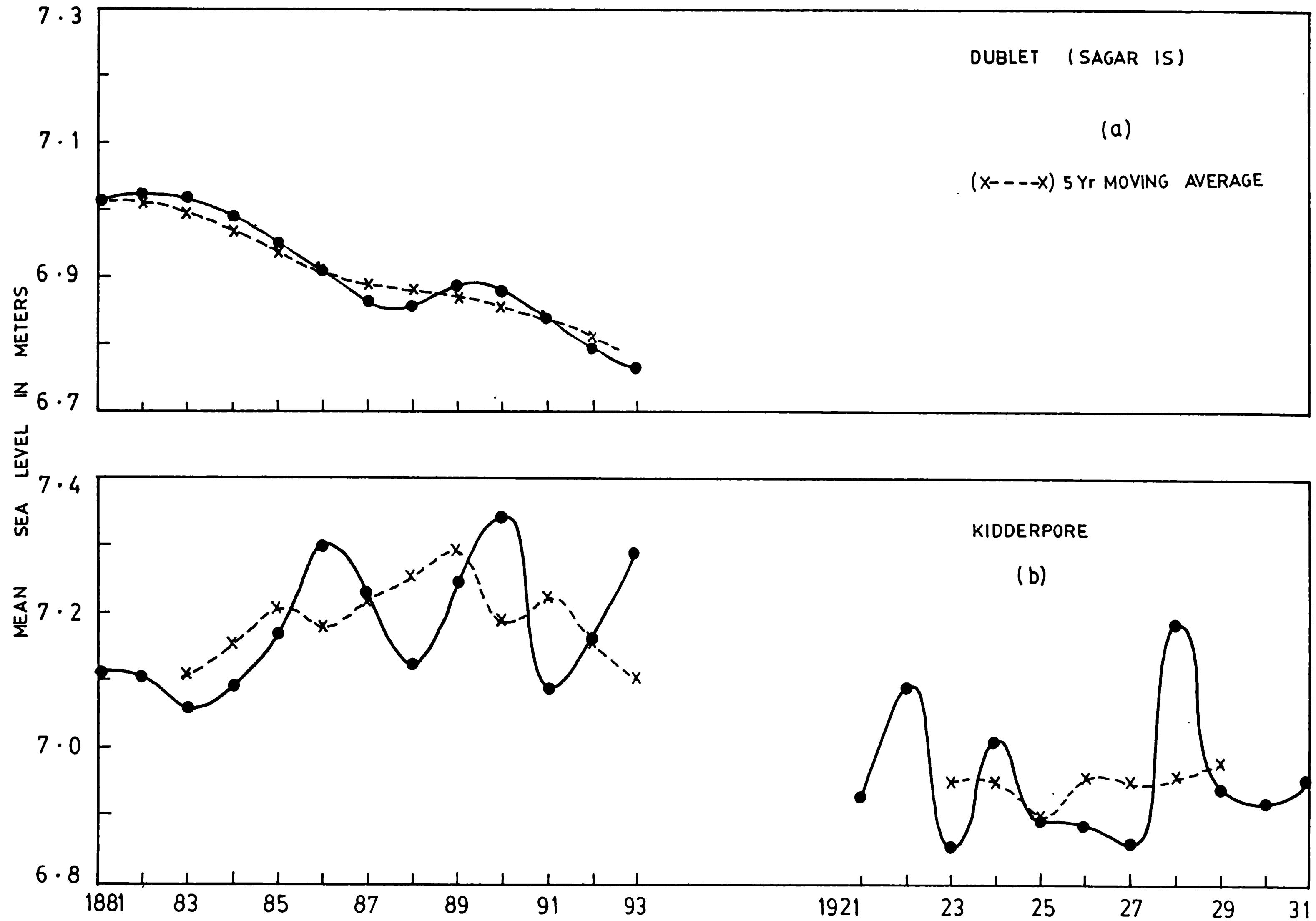
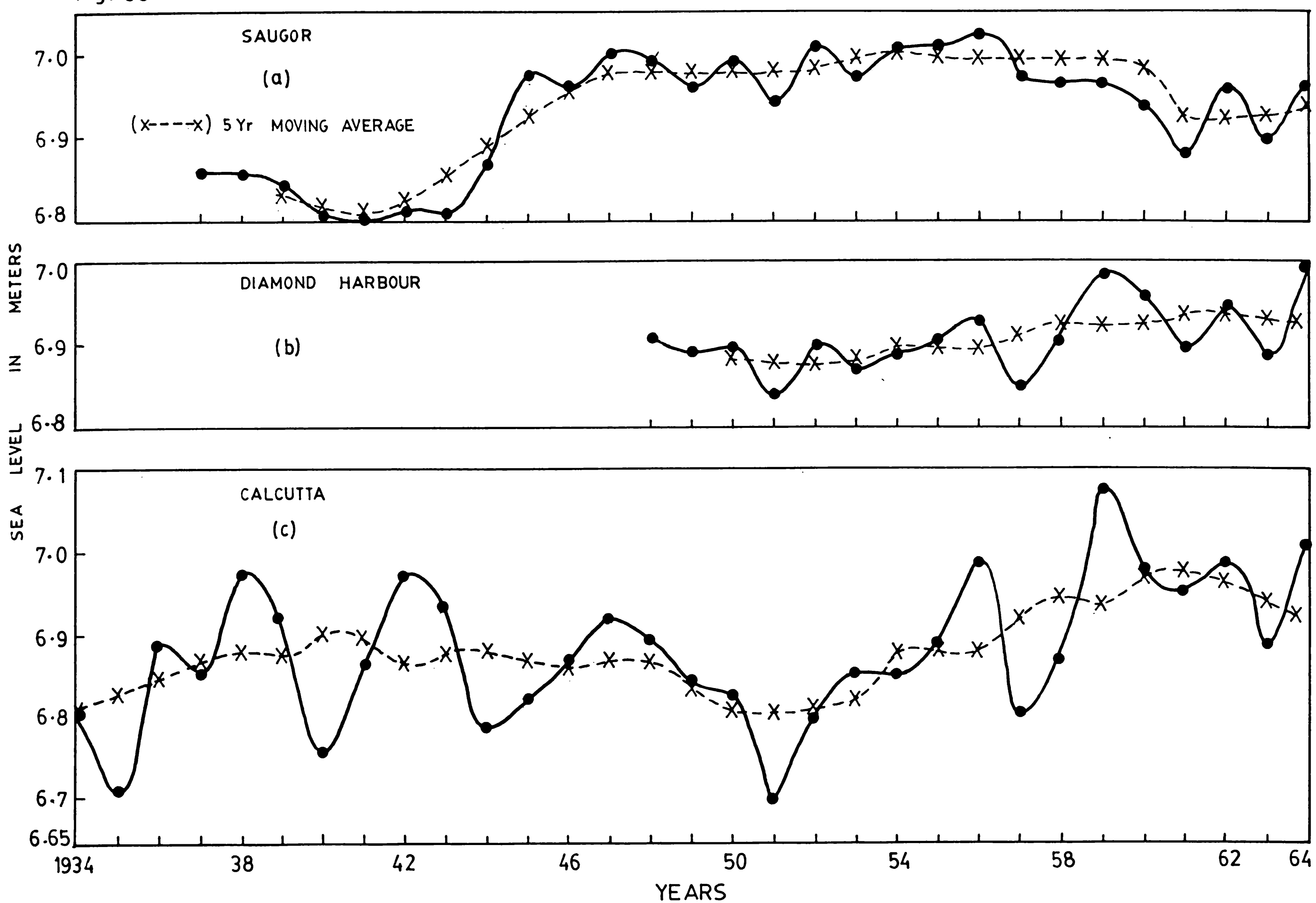


Fig. 55. LONGTERM ANNUAL VARIATIONS OF SEA LEVEL ALONG EAST COAST OF INDIA.



as there was no data available for the period between 1893 and 1921 at Kidderpore and for the period beyond 1893 at Dublet, it was not possible to present the secular variation at these places continuously for the whole period. However, it is interesting to note that the variation of mean sea level at Kidderpore is increasing from 1883-1889, whereas, there is a continuous decrease of mean sea level at Dublet from 1883-1891. It could be inferred, from the tendency of the secular variation curve at Kidderpore, that from 1889 onwards, the sea level is decreasing (Fig.54b). There is an increasing trend of sea level from 1925 onwards.

3.4.3. Sagar, Diamond Harbour and Calcutta

The secular variation of mean sea level at Sagar, Diamond Harbour and Calcutta is presented in Fig.55. While continuous data is available for Sagar and Calcutta, they are limited only for the period 1948 to 1964 for Diamond Harbour.

At Sagar, the sea level increases sharply from 1941-1947 and thereafter, there is not much variation. On the other hand, there is an abrupt fall of sea level during the period 1960-1961 at this station. A tendency of gradual increase in sea level is observed at Diamond Harbour during the period 1950-1962. The variation of sea level at Calcutta indicates alternate increase and decrease with a period of about 20 years which is superimposed by minor alternate variations of a smaller period.

3.4.4. Comparison of secular sea level variation at different stations along the east coast of India

Similar to the variation noticed in the seasonal mean sea level, there is a distinct variation in secular sea level also between the southernmost (Madras and Visakhapatnam) and the northernmost stations under investigation. While the variation of mean sea level is relatively less in the southern region, it is much higher in the north. The change of sign is also more abrupt in the north compared to the south except at Diamond Harbour, where the variation is relatively more gradual. The common feature at almost all the places with the exception of Calcutta is, that the sea level increases from about 1937 till 1957. Similarly, a decreasing tendency in sea level beyond 1959 is conspicuous at most of the stations.

CHAPTER IV

NOMOGRAMS

CHAPTER - IV

NOMOGRAMS

The primary goal of the storm surge study is the safeguarding of life and property with the minimum of inconvenience to the public by forecasting the surge in advance of the cyclone crossing the coast. In carrying out the forecasting function, one considers the available information on the present and recent state of the atmosphere to determine the future state which he thinks has the likelihood of occurrence. The forecast may suggest alternative possibilities.

Three steps are mainly considered in storm surge forecasting: (i) forecasting the cyclone motion and development, (ii) determining the storm surge which is implied by the cyclone forecast and (iii) determining the practical importance of the storm surge in terms of specific flood depths at a particular place.

For maximum usefulness, the technique of estimation of surge should be simple, and it should be based on cyclone parameters that are obtainable before the cyclone crosses the coast.

Since the present thesis is primarily concerned with the objective of developing a method by which the storm surge generated by an impinging cyclone of known parameters could be predicted before it approaches the coast, it is proposed to discuss the different techniques and finally present the nomograms that are worked out to give a forecast of probable surges that can emerge out of impinging cyclones of various intensities and angles of orientation of crossing the east coast of India.

The first step in formulating a storm surge is forecasting the future position and intensity of the cyclone. According to usual practice, the location of low pressure centre is first determined and then the wind field and the other features of the cyclone are related to the position of the centre.

4.1. Surge forecasting

The techniques of forecasting storm surges in partly enclosed sea areas have been classified by Groen (1963) as being either purely empirical, purely theoretical and partly empirical-partly theoretical.

4.1.1. Purely empirical

In purely empirical methods, the surge at a particular place and time is related directly to the values of barometric pressure or wind at fixed points of the sea at the same or earlier times. The relation is derived on the basis of statistical analysis of sea level and meteorological data corresponding to the past surges. Account may be taken of the surge propagation to a given place by including a contribution to the surge at that place involving the disturbance of sea level at another place at an earlier time. An example of this formula is derived by Corkan (1948).

$$R_S = R_D + 0.033\,N|N| - 0.055\,E|E| - 0.075\,n|n| - 0.095\,e|e|$$

where R_S is the disturbance in feet where the surge is to be forecasted after eliminating the effect of local barometric pressure assuming a statical law at time t hours, R_D is the observed disturbance in feet at some place away from the place of forecast at time t - 9 hours after elimination of the effect of barometric pressure assuming the statical law.

N and E are the north and east pressure gradients in the sea at time t hours. n and e are the north and east pressure gradients in the sea at time t - 9 hours. Clearly, in using this formula to forecast the surge at time t hours, a meteorological forecast is required of the pressure gradients in the sea at time t hours, t - 6 hours together with the observed surge and pressure at t - 9 hours.

The empirical formula used for prediction of storm surge in Japan is

$$H = -a(p-p_o) + b\,W^2 \cos\theta$$

where p is the atmospheric pressure (mb), p_o the monthly mean atmospheric pressure (mb), W the wind velocity (m/s), θ the angle between the wind direction and the direction in which the wind setup is most effective (usually, the direction normal to the coast or the direction of the bay axis), H the surge height (cm), and a and b the numerical constants. The first and the second term of the right-hand side represent the sea level variation due to the atmospheric pressure fall and the wind effect respectively. The coefficients a and b for each station are calculated by means of the least square method from hourly or three hourly observations, and the effect of wind on the storm surge was compared with that of atmospheric pressure fall. Further, similar empirical techniques have been developed by Cline (1920,1926), Tomczak (1954), Connar et al. (1957), Hoover (1957), Harris (1957,1962), Donn (1958), Bretschneider (1959a,1959b), Harris and Angelo (1963) and Keers (1968), Chan and Walker (1979).

The empirical methods rely on statistical analysis of past observations and much depends on the degree to which the results of such analysis are relevant to future situations. The tendency is for the statistical averaging processes to give more weight to the greater number of small surges and, as a consequence, an empirical formula might well give unsatisfactory forecasts for the vitally important large surges which occur infrequently but which often produce disastrous floods. The empirical method uses values of barometric pressure or wind at one or more discrete points over the sea and is, therefore, liable to omit the effects of strong inhomogeneities in such producing wind and pressure fields.

4.1.2. Purely theoretical

In purely theoretical approach surge forecasts are obtained by integrating the dynamical differential equations which govern the generation and propagation of storm surges. For this purpose a numerical sea model is postulated. The differential equations are replaced by finite difference equations which then form the basis of step-by-step procedure yielding surge values at intersection points of a grid covering the sea. Uncertainties in the determination of surface wind stress from meteorological data and lack of proper knowledge of values to be assigned to the coefficients of bottom friction are some important weaknesses of this method. Harris (1962) and Groen (1963) discussed the advantages and disadvantages of purely empirical and purely theoretical lines of approach. In the theoretical method the accuracy of the surge forecast depends upon the assumptions employed in translating the physical description of nature into mathematical terms.

4.1.3. Partly empirical - Partly theoretical

Partly empirical - partly theoretical methods has been developed by Schalkwijk (1947), Weenink (1958), Weenink and Groen (1958). The method first determines the wind induced elevation of sea level at a particular time assuming the changing wind conditions as quasi-stationary. In these determinations the elevation at any particular time (known as the equilibrium elevation) is found by summing up the conditions from wind fields at that time acting respectively over each from different points. The contribution to the equilibrium elevation from each area is deduced from the gradient wind over the area using tables and graphs established partly on an empirical basis and partly on the basis of theoretical computation. The real elevations are subsequently obtained from the equilibrium elevation by applying simple corrections to take account of certain dynamic effects of a time-lag between the application of wind to the sea surface and the corresponding response of the surface level heights associated with a surge peak, and damped oscillations of the sea level following a surge.

Allied to the problem of surge forecasting as discussed above is the problem of determining the frequency with which abnormally high sea levels might be expected to occur in the future at coastal locations. Knowledge of such frequencies of occurrence helps in making the design of sea defences safe and economical.

4.2. Present technique

Of the three techniques discussed, partly theoretical - partly empirical technique appears to have less limitations compared to either purely empirical or purely theoretical technique. Therefore, it is proposed to apply such a technique for the prediction of surges that are produced by cyclones originated in the Bay of Bengal and impinge the east coast of India.

The various factors that contribute to the generation of a surge and their relative significance are discussed in the introductory chapter. From that it is realised that the wind setup plays a major role and it is mainly controlled by the bathymetry and wind stress. The winds are computed for various intensities of pressure drop between the centre and the outer periphery of the cyclone identified by the outer most closed isobar around a low pressure. The wind setup developed at different places along the east coast of India by the cyclones approaching the coast at angles of 90° and 135° from the north are estimated, taking into the account, the actual bathymetry at these places. Graphs are drawn between the wind setup for varying pressure drop and the radius of curvature of the outermost closed isobar of the low pressure system. In the computation, the pressure gradient is considered as the pressure drop between the lowest reported pressure and the pressure of the outermost closed isobar of the cyclone. The pressure gradient in the cyclone varies from a highest value near the eye of the cyclone to the lowest value at the outer periphery of the cyclone. Therefore, the pressure gradient evaluated gives an overall pressure gradient instead of the *in situ* pressure at different points. Hence, the computed surge may have a slight deviation from the real which is subsequently obtained by applying a simple correction to take account of certain dynamic effects caused by the wind fields not being stationary as has been assumed. The correction is obtained by drawing a graph between the surge values obtained from the nomograms and from the case studies.

4.3. Winds against the pressure drop

The winds generated by varying pressure drop for different radii of curvature of the cyclones are computed using the gradient wind equation. They are presented as the ordinate against the pressure drop on the abscissa. In fact, the pressure drop is used as a substitute to the pressure gradient, which is the variation of the pressures over the radius of the outer most closed isobar. According to Bretschneider (1967) the recorded pressure drop in a tropical cyclone is 45 mb and the maximum radius of curvature is 500 km. Hence, in the present study, the winds are computed for pressure drops varying from 0 to 50 mb and for the radii of curvature of 100,200,300,400 and 500 km. Figs.56 to 60 show the graphs representing winds generated by the varying pressure drop at Nagapatnam, Madras, Masulipatnam, Visakhapatnam and Paradip, respectively.

As expected, for the gradient wind equation, the magnitude of the wind decreases with increasing radius of curvature and Coriolis force, for the same pressure drop. Accordingly, the wind speed is found decreasing from Nagapatnam to Paradip (Figs.56 to 60) for any pressure drop over a constant radius of curvature. For a maximum pressure drop of 50 mb and a radius of curvature of 100 km, the winds

Fig.56. RELATION OF WIND AND PRESSURE DROP AT NAGAPATNAM

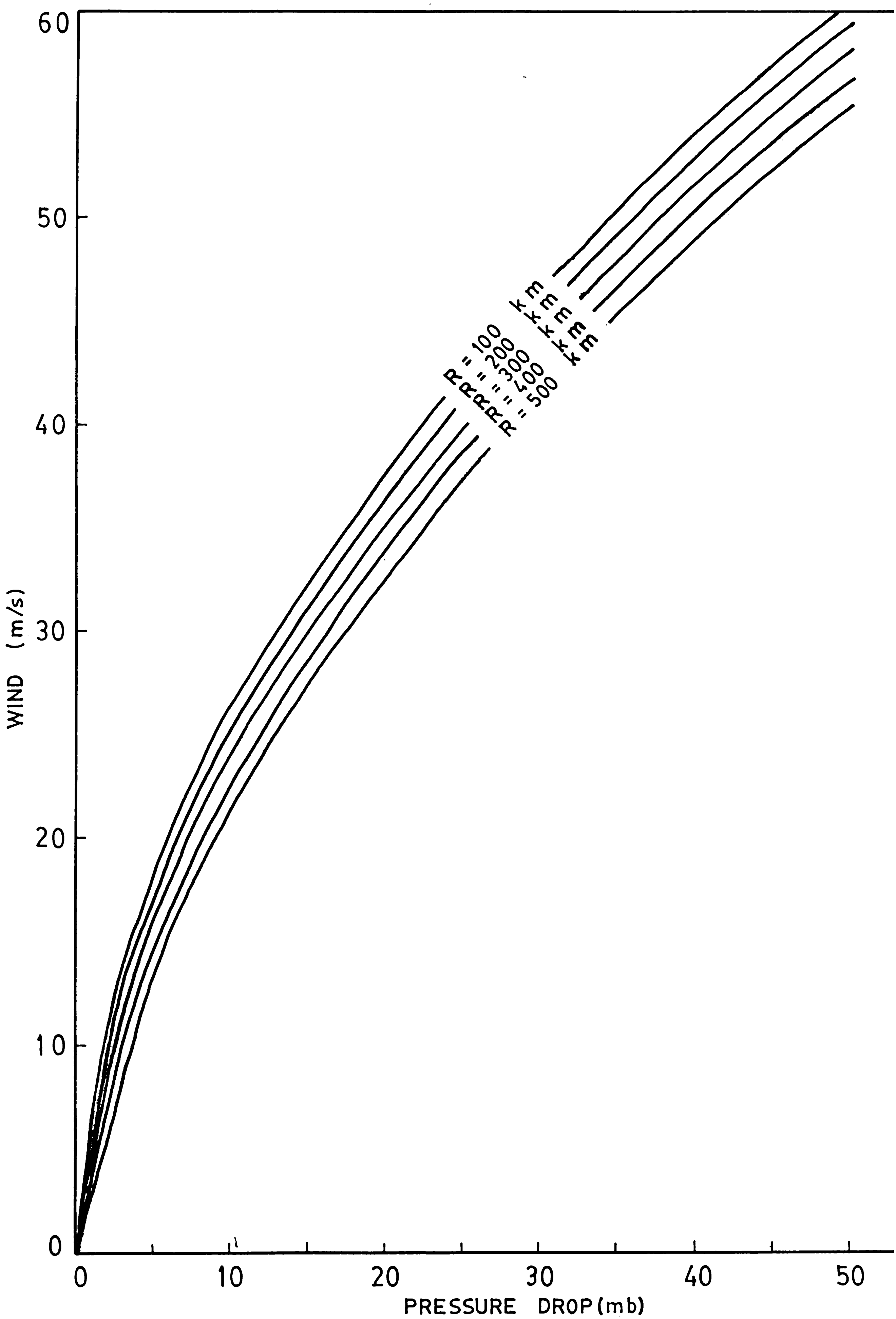


Fig.57. RELATION OF WIND AND PRESSURE DROP AT MADRAS.

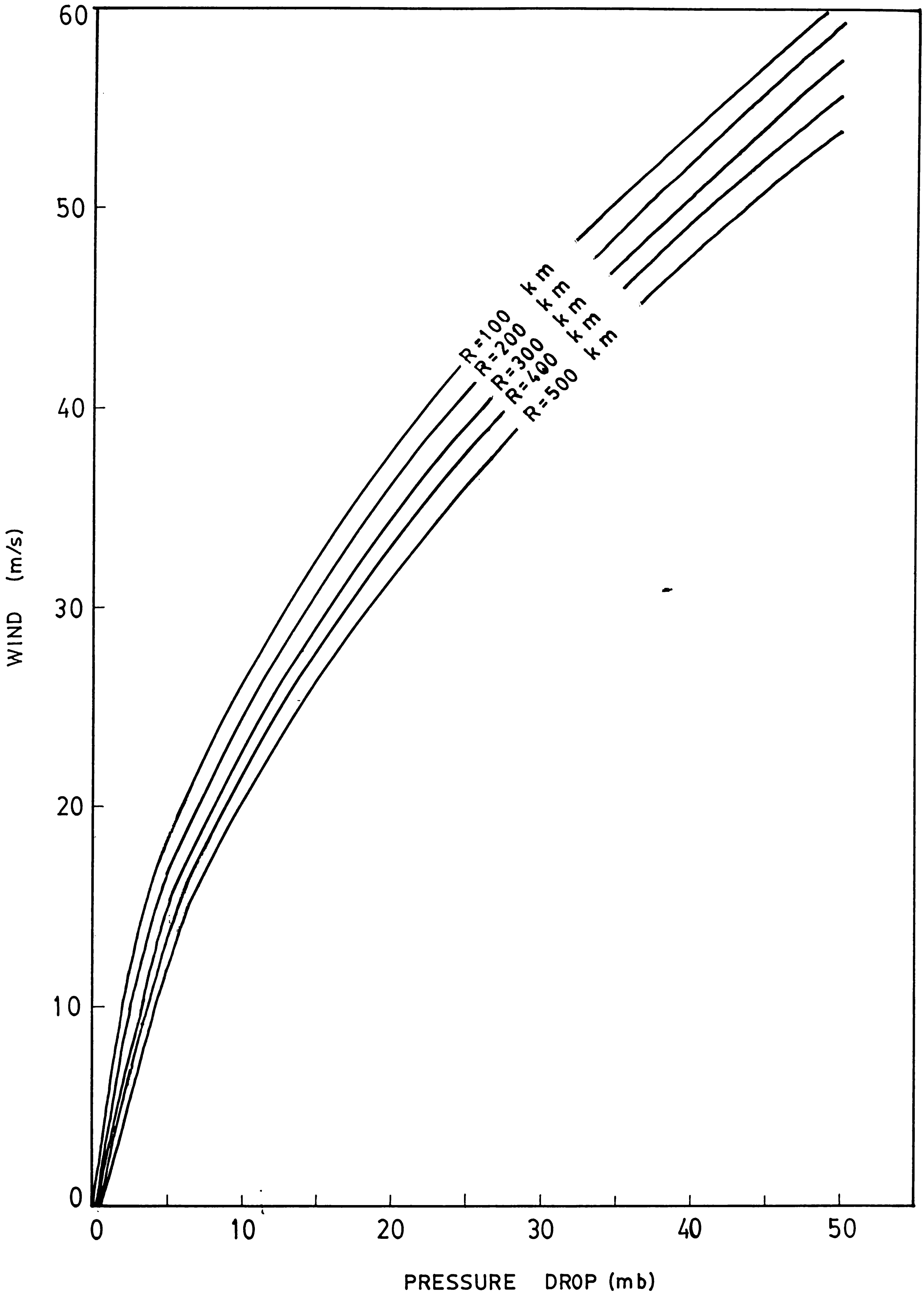


Fig.58. RELATION OF WIND AND PRESSURE DROP AT MASULIPATNAM.

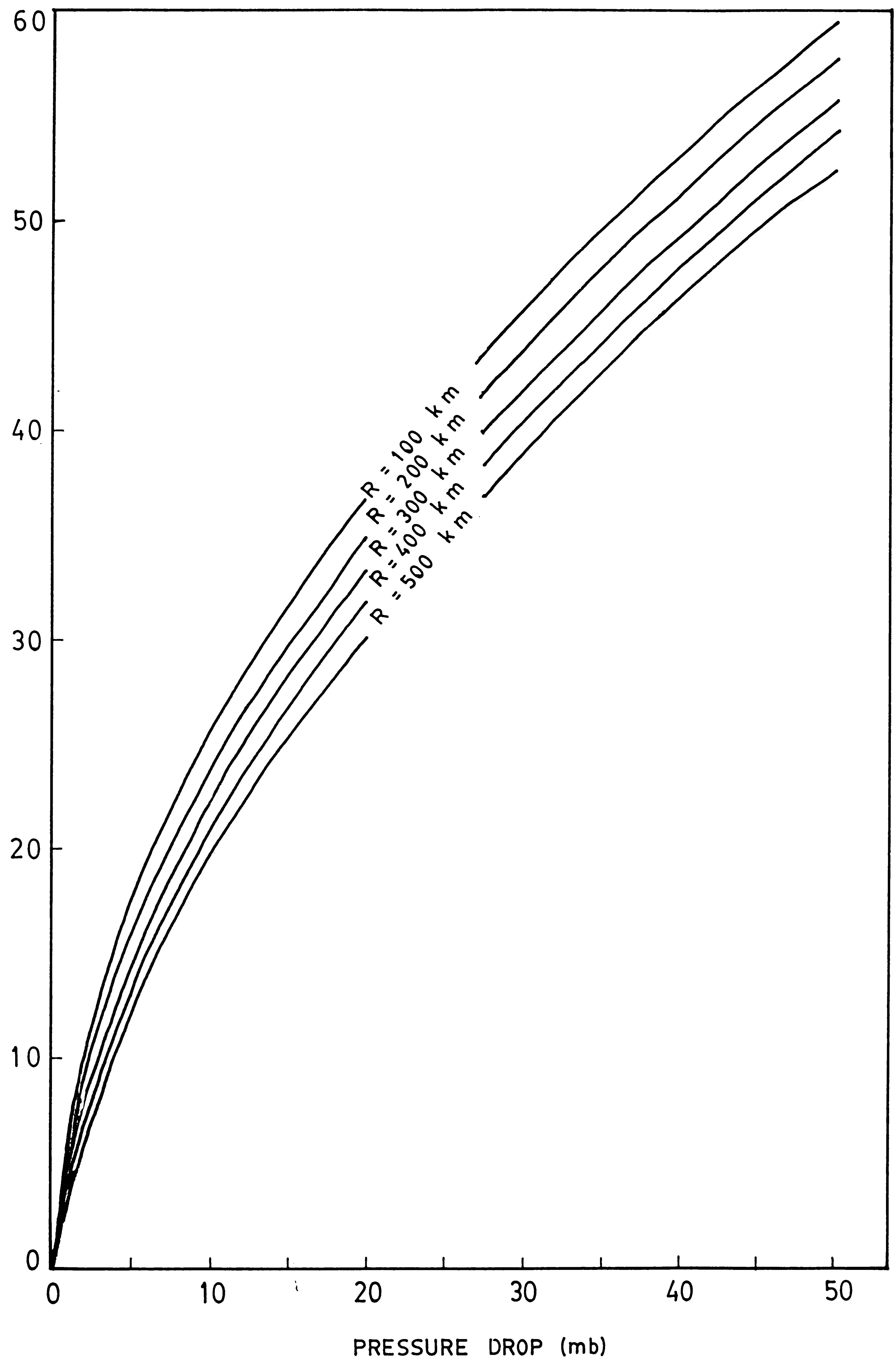


Fig 59. RELATION OF WIND AND PRESSURE DROP AT VISAKHAPATNA

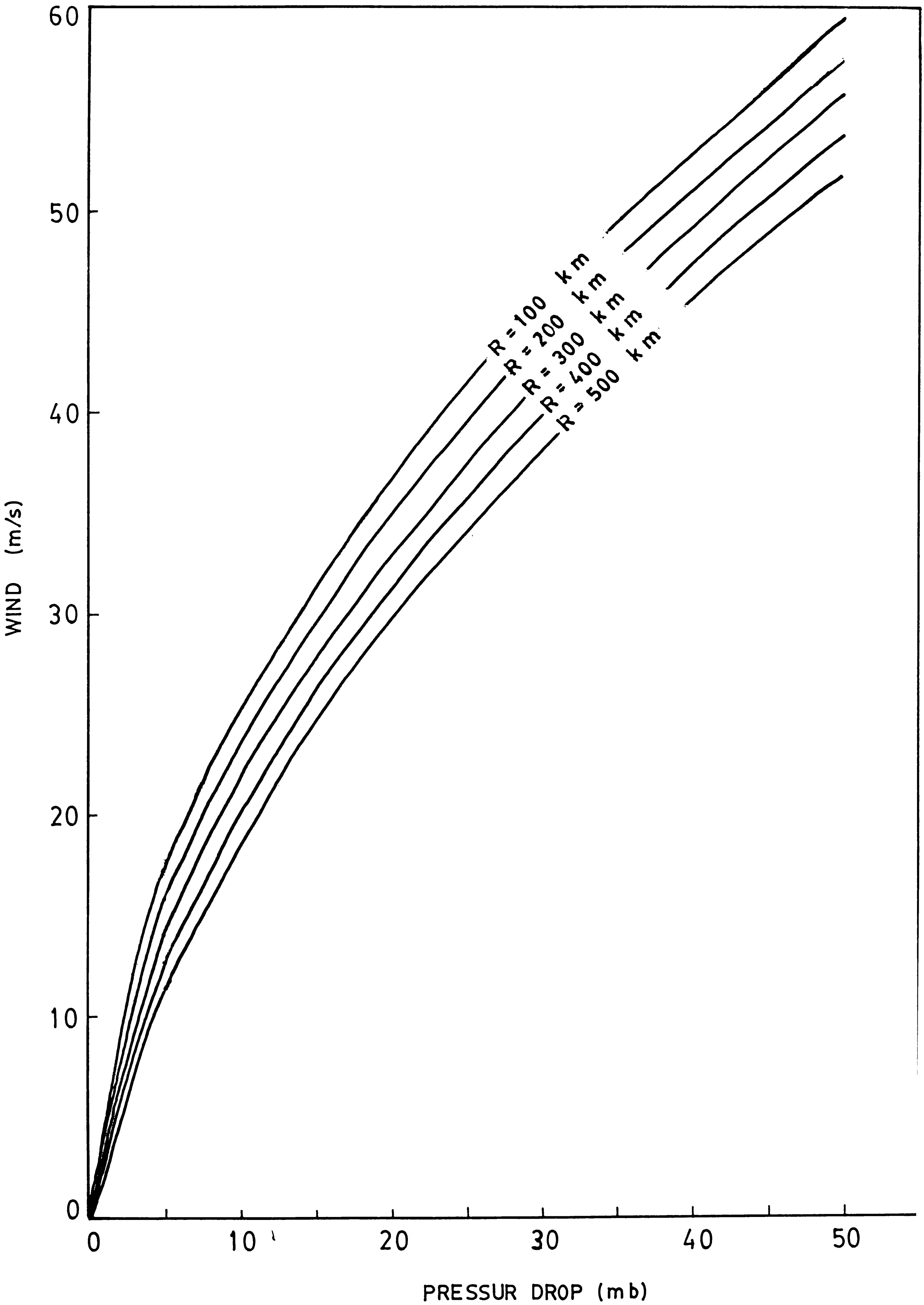
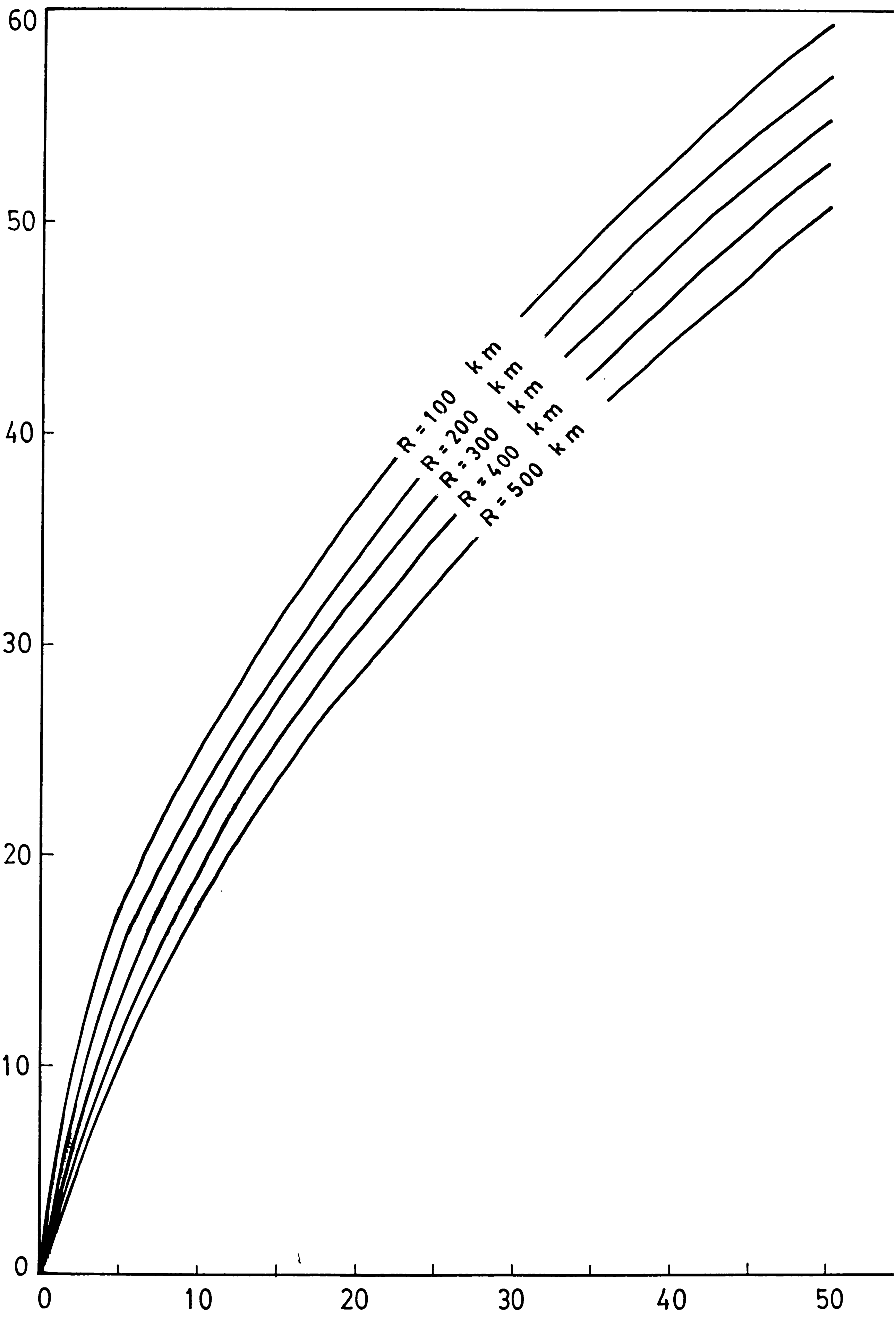


Fig.60. RELATION OF WIND AND PRESSURE AT PARADIP.



at Nagapatnam and Paradip are found to be 61 m sec^{-1} and 59 m sec^{-1}, respectively. For the same pressure drop and a radius of curvature of 500 km the corresponding values, respectively, are 55 and 51 m sec^{-1}. Thus it is evident that Coriolis force is too insignificant and for that matter even the effect of the radius of curvature also is much less compared to that of the pressure drop in the generation of wind.

4.4. Surge prediction from nomograms

After evaluating the winds for varying pressure drop and different radii of curvature, the surge generated by the winds is computed using the actual bathymetry at Nagapatnam, Madras, Masulipatnam, Visakhapatnam and Paradip, for an impinging cyclone with angles of incidence of 90° and 135° from the north. Graphs are drawn with the surge on the ordinate and pressure drop along the abscissa. These graphs serve as nomograms to predict the surge when the pressure drop, radius of curvature and the angle of incidence of the cyclone are known. Normally, the maximum surge that can be expected does not exceed 10 m and so the computations are limited to a maximum surge height of 10 m.

4.4.1. Nagapatnam

Figs.61 and 62 represent the nomograms of the surge

Fig.61. SURGE AT NAGAPATNAM FOR ANGLE OF INCIDENCE - 90°.

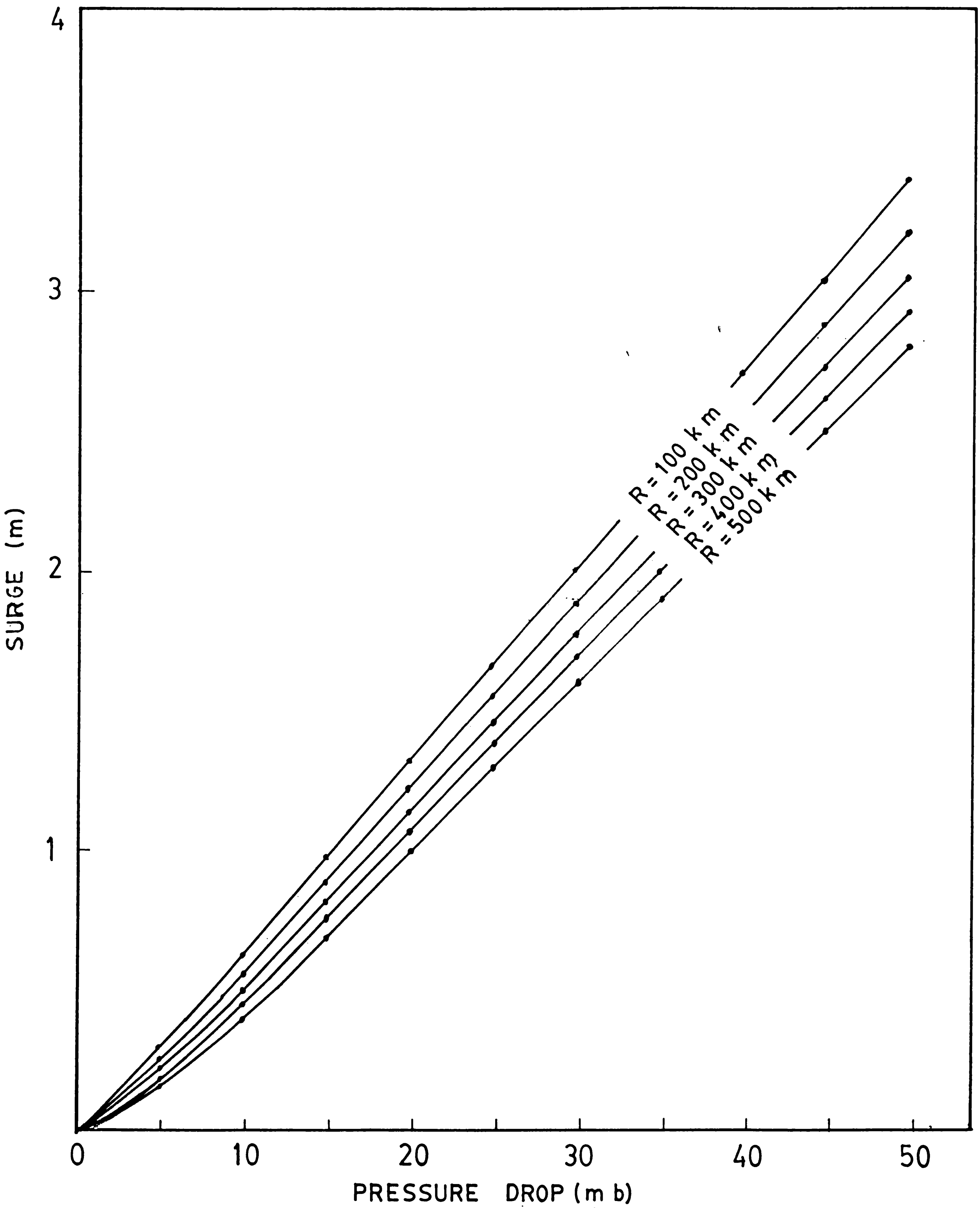


Fig. 61a. BATHYMETRY BETWEEN NAGAPATNAM & MADRAS.

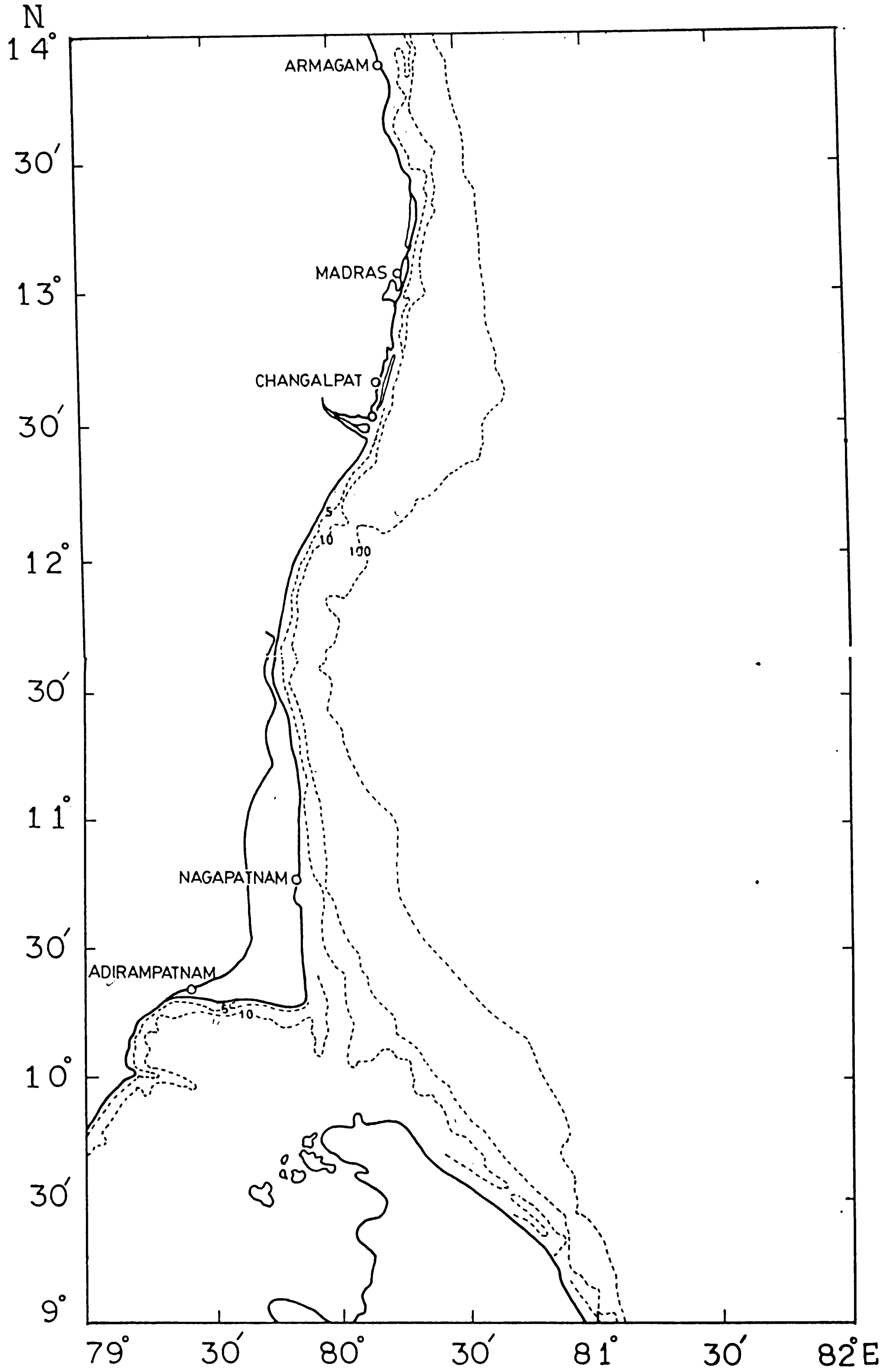
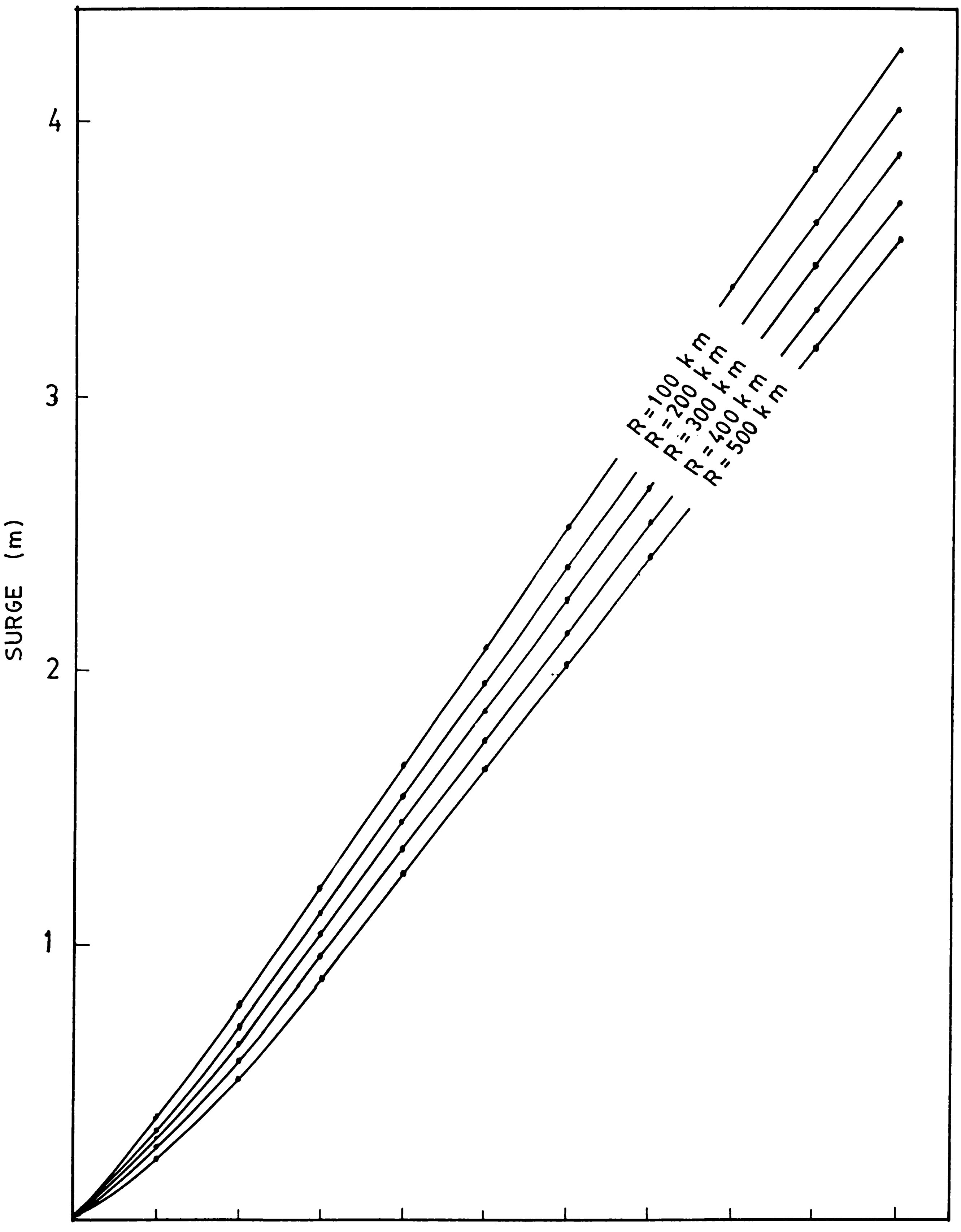


Fig.62. SURGE AT NAGAPATNAM FOR ANGLE OF INCIDENCE - 135°.



produced at Nagapatnam for the angles of incidence of 90° and 135°, respectively. At Nagapatnam a surge of about 3.4 m occurs for a pressure drop of 50 mb and a radius of curvature of 100 km when the angle of incidence is 90°, whereas it is about 4.2 m for the same radius of curvature and pressure drop but an angle of incidence is 135°. Therefore, the surge height increases with the increase in angle of incidence. This is mainly due to the difference in bathymetry with respect to the two orientations (Fig.61a). The inclination of the curves with the abscissa indicates that the surge height increases rapidly with the pressure drop and more so when the angle of incidence is 135°. The variations of surge height for a pressure drop of 50 mb and for the radii of curvature of 100 km and 500 km are from 3.4 to 2.8 m and 4.2 to 3.6 m for the angles of incidence of 90° and 135° respectively. Thus the maximum contribution due to the change in radius of curvature from 100 to 500 km is only 0.6 m for a pressure drop of 50 mb.

4.4.2. Madras

The surge generated at Madras for varying pressure drop and different radii of curvature is shown in Figs.63 and 64 for the angles of incidence of 90° and 135° respectively. For a pressure drop of 50 mb and a radius of curvature of 100 km, a surge of about 1.9 m is induced

Fig .63. SURGE AT MADRAS FOR ANGLE OF INCIDENCE - 90°.

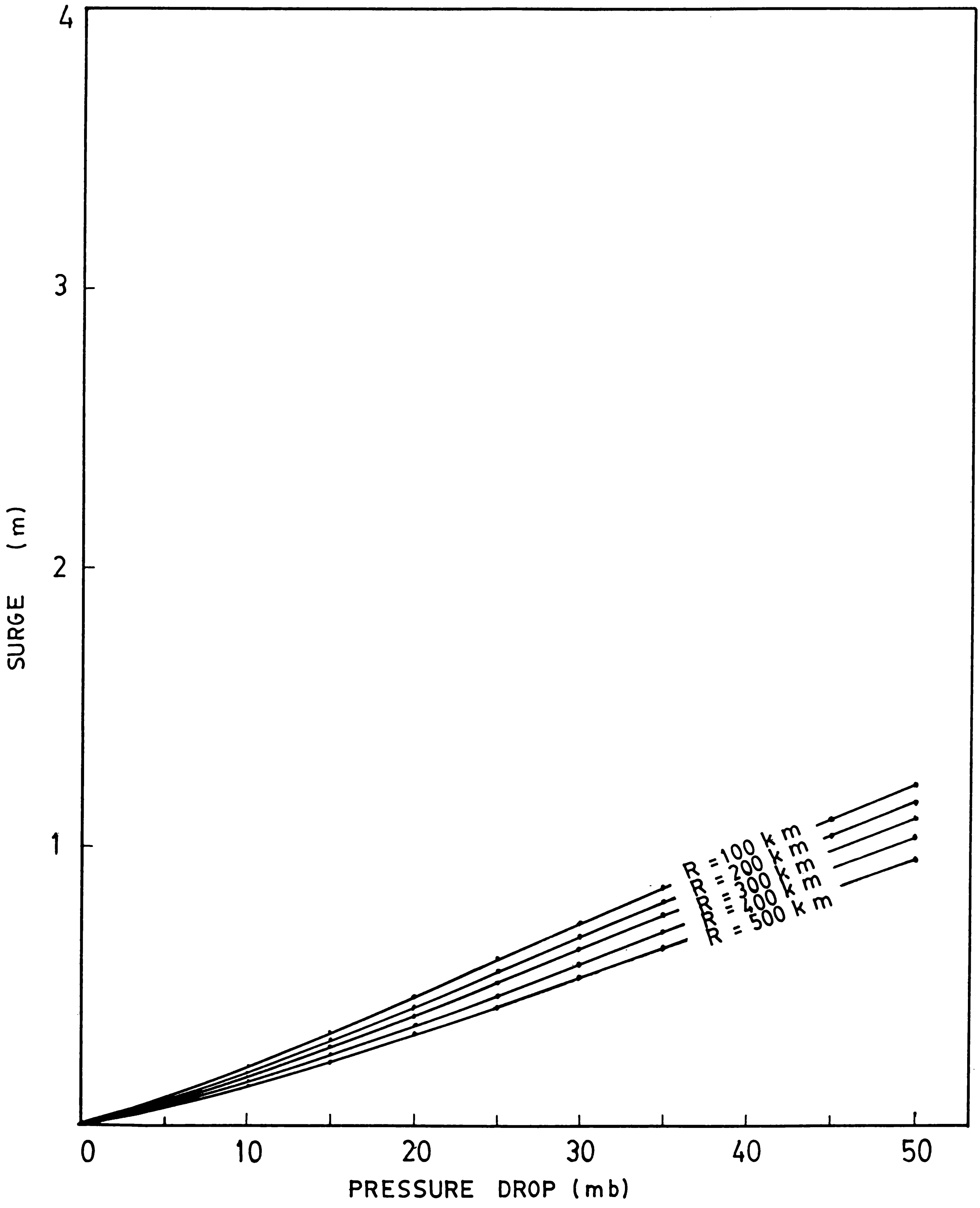
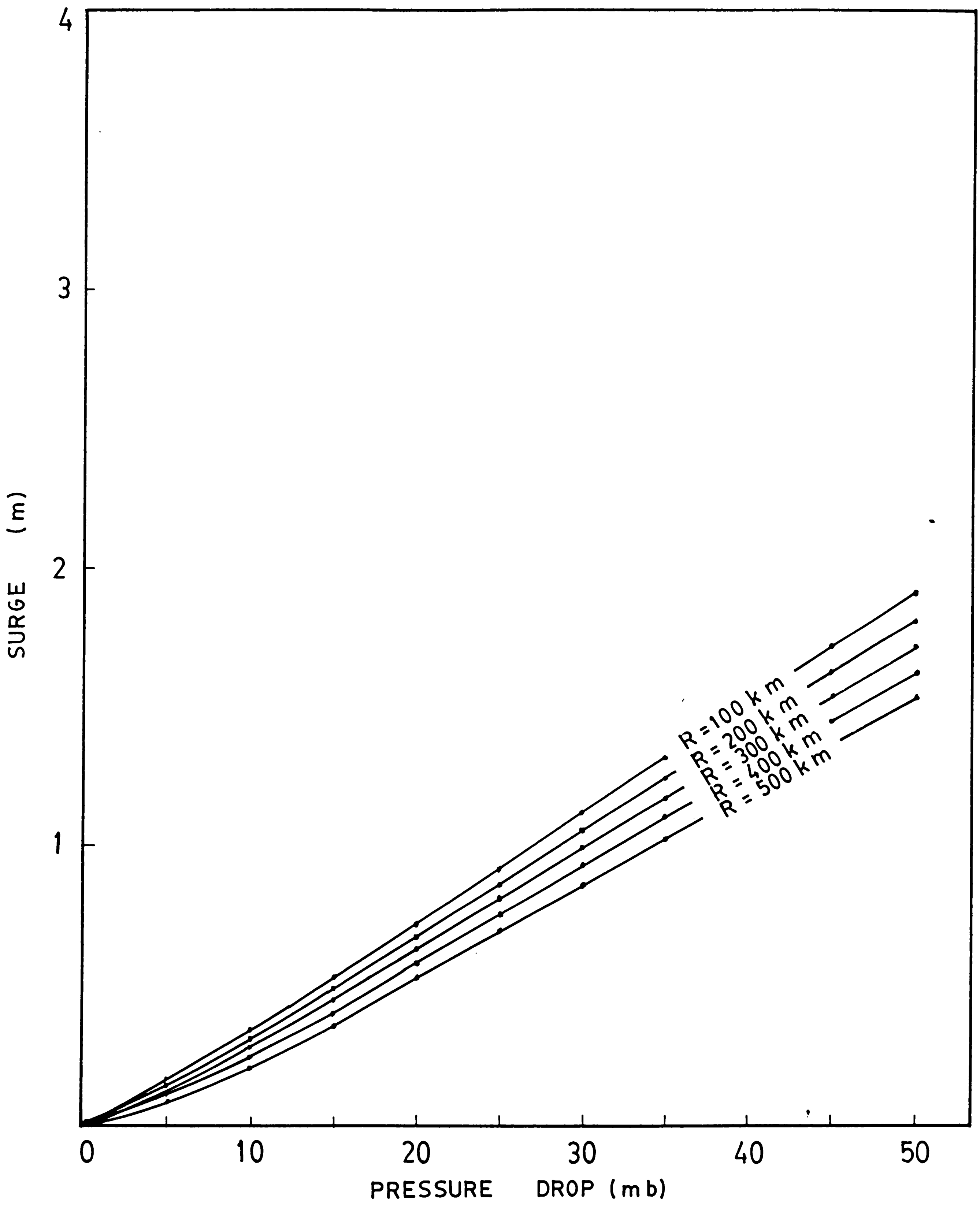


Fig. 64. SURGE AT MADRAS FOR ANGLE OF INCIDENCE - 135°.



when the angle of incidence is 135° and about 1.2 m for the angle of incidence is 90°. Hence, it can be inferred that the change in the orientation of the impinging cyclone can change the surge height by about 0.7 m. Although this number appears to be small, its relative contribution to the total surge height is considerable as the maximum surge is only 1.9 m. It can also be seen that the surge height increases with the increase of angle of incidence from 90° to 135°. From the closeness of the curves for different radii of curvature, it is obvious that the variation of the radius of curvature has negligible effect on the surge height, which increases with the increasing angle of incidence.

4.4.3. Masulipatnam

Graphs presenting the surge height for varying pressure drop and different radii of curvature are presented in Figs.65 and 66 for the angles of incidence of 90° and 135° respectively. It may be noted that the maximum surge of over 10 m occurs at Masulipatnam for an angle of incidence of 90° whereas, it is not even 5 m for the same pressure drop of 50 mb and radius of curvature of 100 km when the angle of incidence is 135°. Because of such a high surge for the angle of incidence of 90° at Masulipatnam a relatively higher scale is taken for

Fig. 65. SURGE AT MASULIPATNAM FOR ANGLE OF INCIDENCE - 90°.

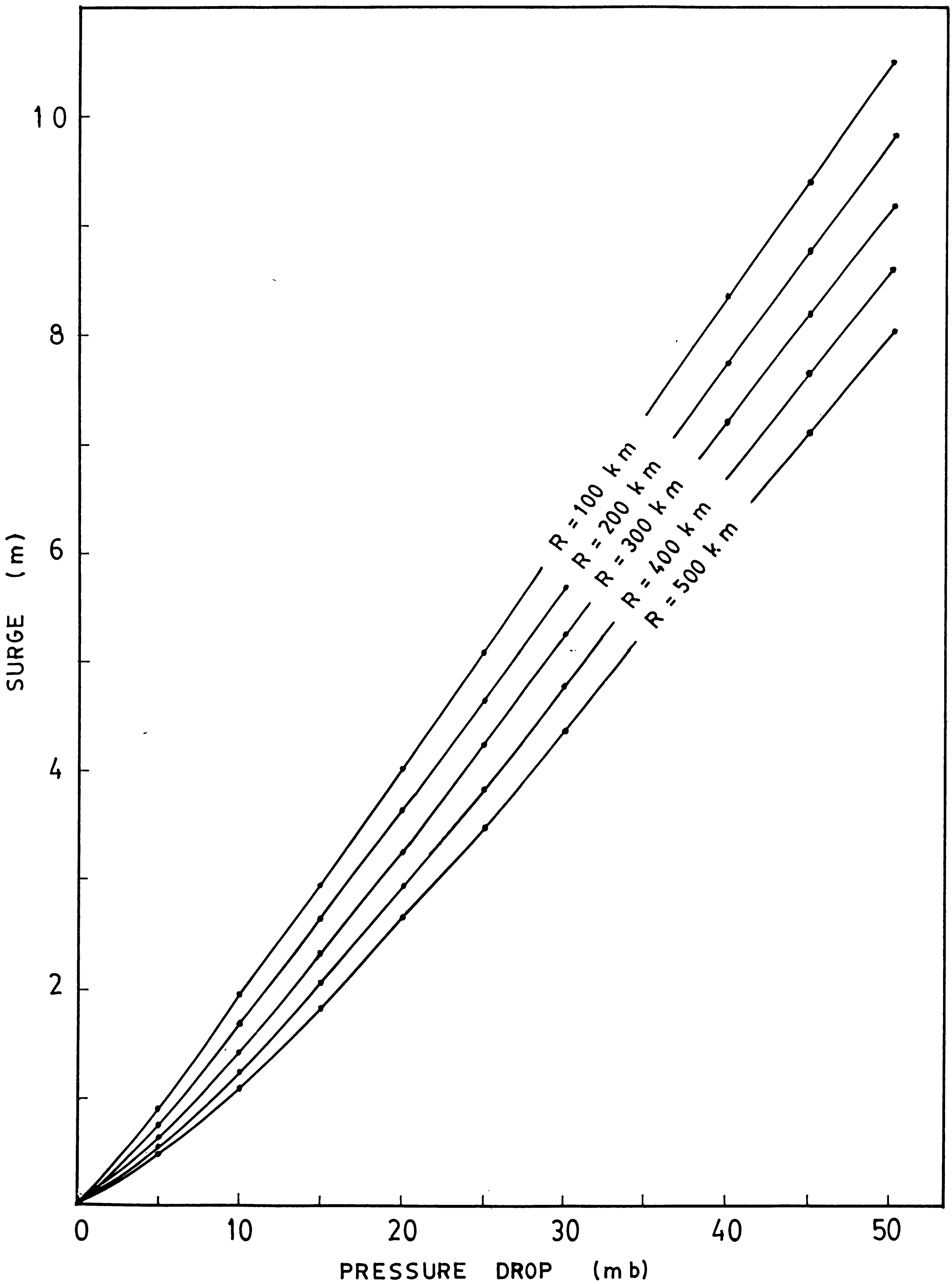
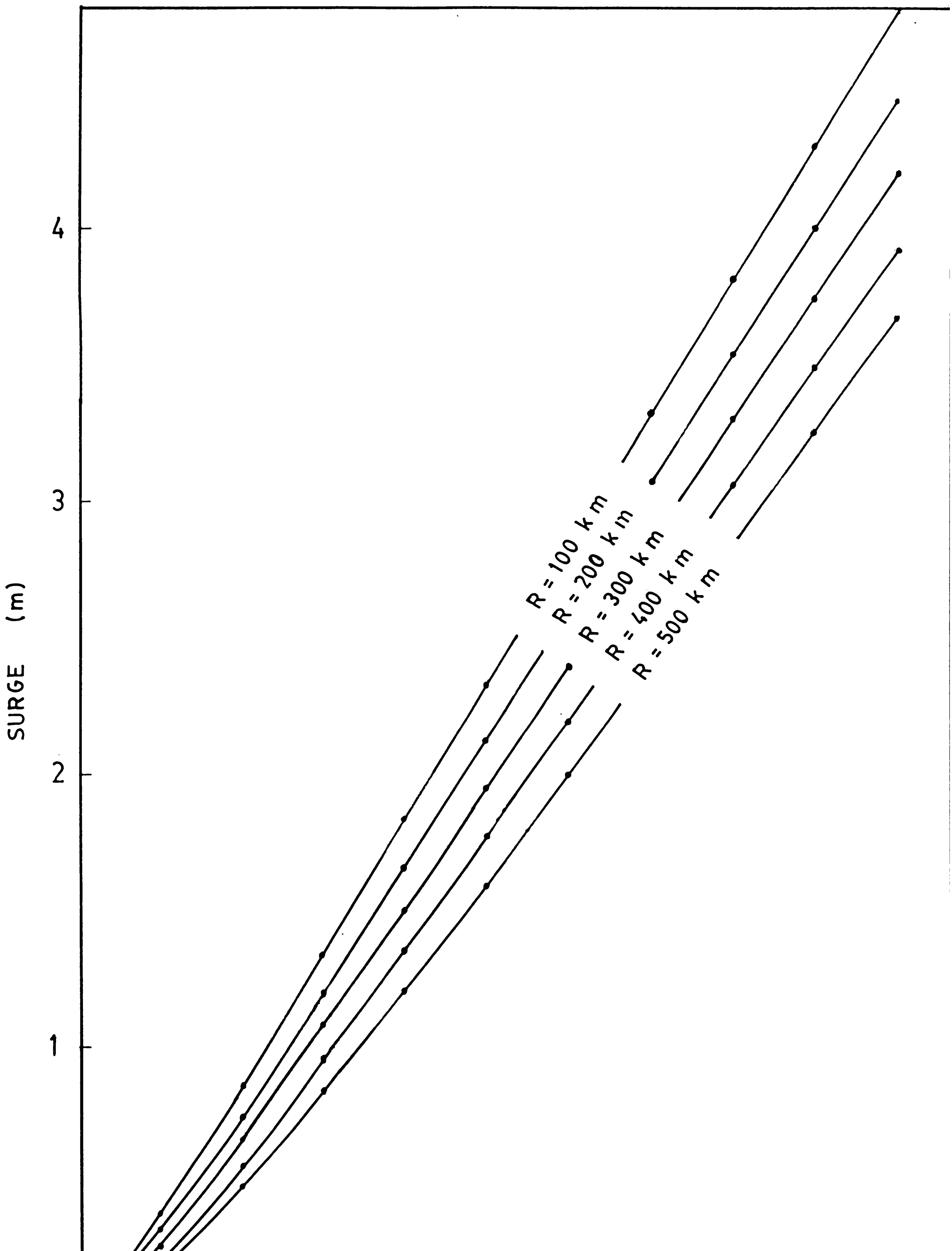


Fig.66. SURGE AT MASULIPATNAM FOR ANGLE OF INCIDENCE -135°.



presenting the surge height in this particular diagram (Fig.65), while a smaller and uniform scale on the ordinate is used for the rest of the diagrams. Obviously, the angle of incidence has a major influence on the induced surge at Masulipatnam. For both the angles of incidence the surge height increases sharply with the increase of pressure drop. The effect of the radius of curvature is relatively insignificant compared to other factors.

4.4.4. Visakhapatnam

The surge generated at Visakhapatnam for varying pressure drop and radii of curvature for the angles of incidence of 90° and 135° is represented in Figs.67 and 68 respectively. For a maximum pressure drop of 50 mb and for an angle of incidence of 90° and radii of curvature of 100 km the surge produced is about 1.5 m. Therefore, it can be construed that the surge developed at Visakhapatnam is of much smaller magnitude, may be mainly due to the narrow continental shelf. The surge height increases with the decrease of angle of incidence from 135° to 90°. It is interesting to note that the surge height does not depend much on the radius of curvature.

Fig. 67. SURGE AT VISAKHAPATNAM FOR ANGLE OF INCIDENCE - 90°.

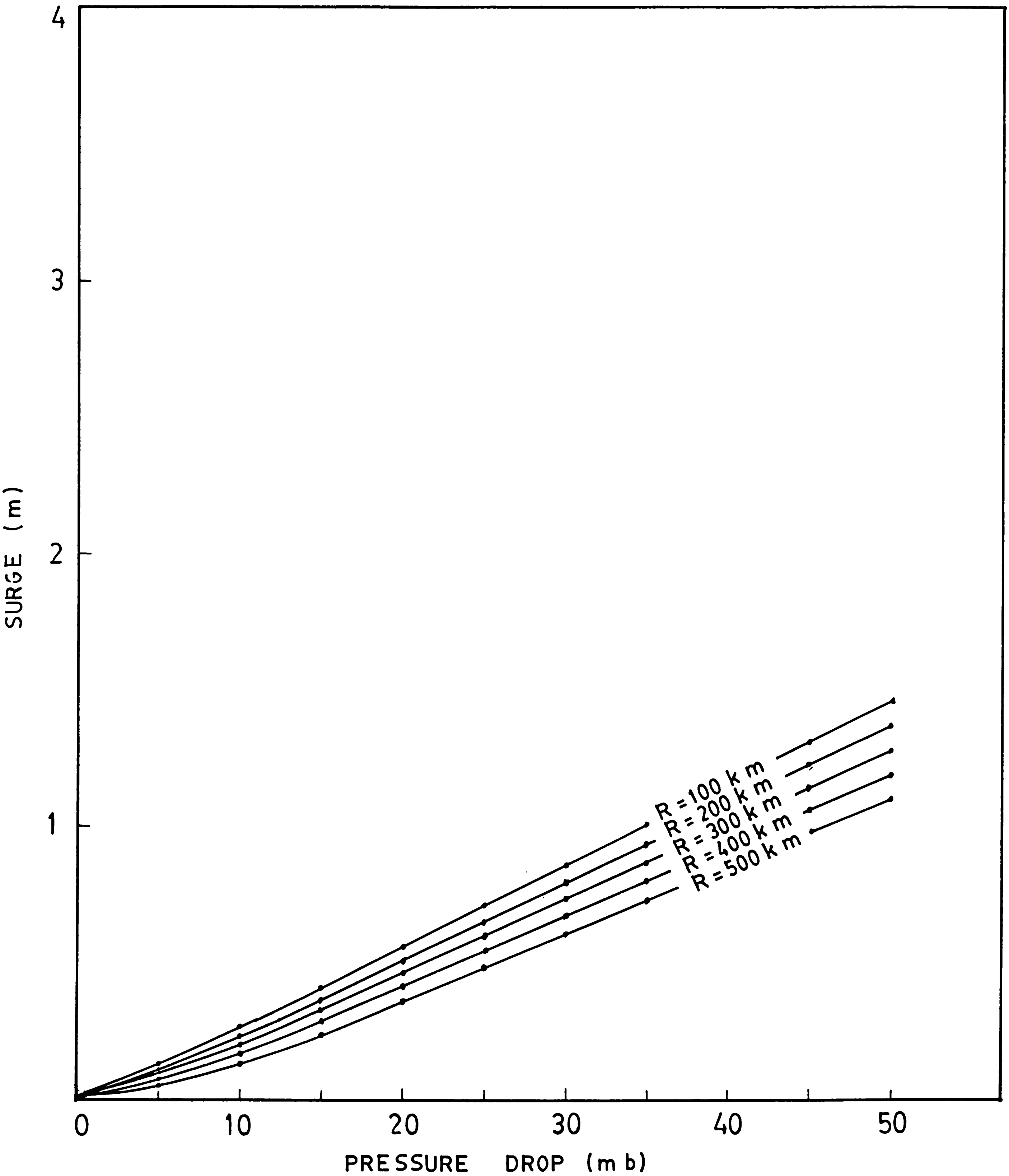
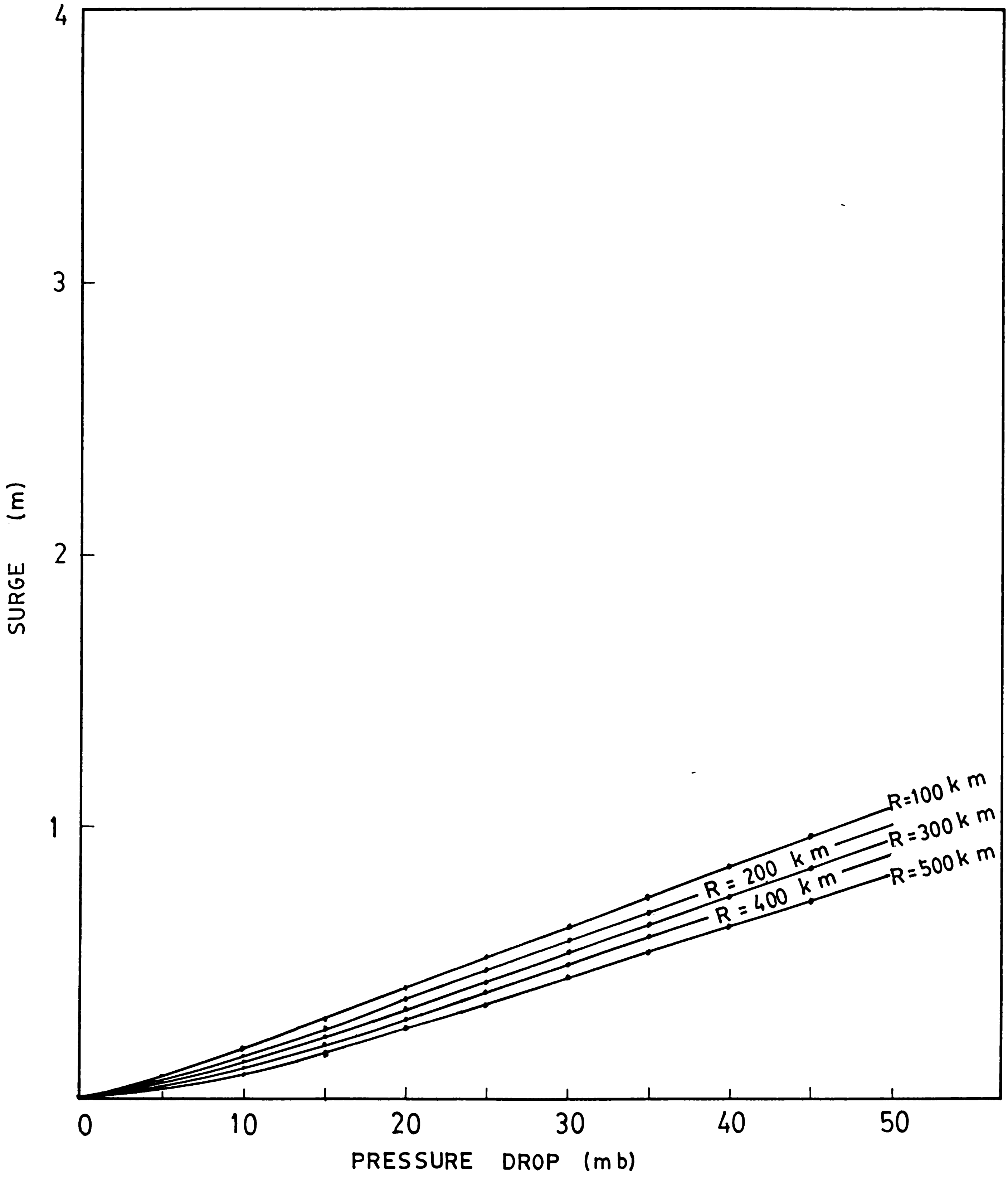


Fig.68. SURGE AT VISAKHAPATNAM FOR ANGLE OF INCIDENCE - 135°.



4.4.5. Paradip

Figs.69 and 70 show the surge evolved at Paradip for the angles of incidence of 90° and 135° respectively. When the angle of incidence is 90°, a pressure drop of 50 mb generates a surge of over 5 m for a radius of curvature of 100 km. For the same pressure drop and radius of curvature a surge of less than 2.5 m is developed when the angle of incidence is 135°. Thus, it is obvious that the angle of incidence is a predominant factor controlling the surge height. A change in the radius of curvature from 100 to 500 km results in a drop of about 1.4 and 0.7 m in the surge height for the angles of incidence of 90° and 135° respectively when a pressure drop of 50 mb occurs. Therefore, the effect of the radius of curvature also decreases with the increase of angle of incidence from 90° to 135°.

4.5. Storm surges along the east coast

In this section, it is proposed to discuss the relative effects of impinging cyclones along the east coast of India with varying intensity and angle of incidence. From the preceeding section it is evident that along the east coast the most vulnerable areas of storm surges are off Masulipatnam and Paradip; particularly, when the angle

Fig.69: SURGE AT PARADIP FOR ANGLE OF INCIDENCE - 90°.

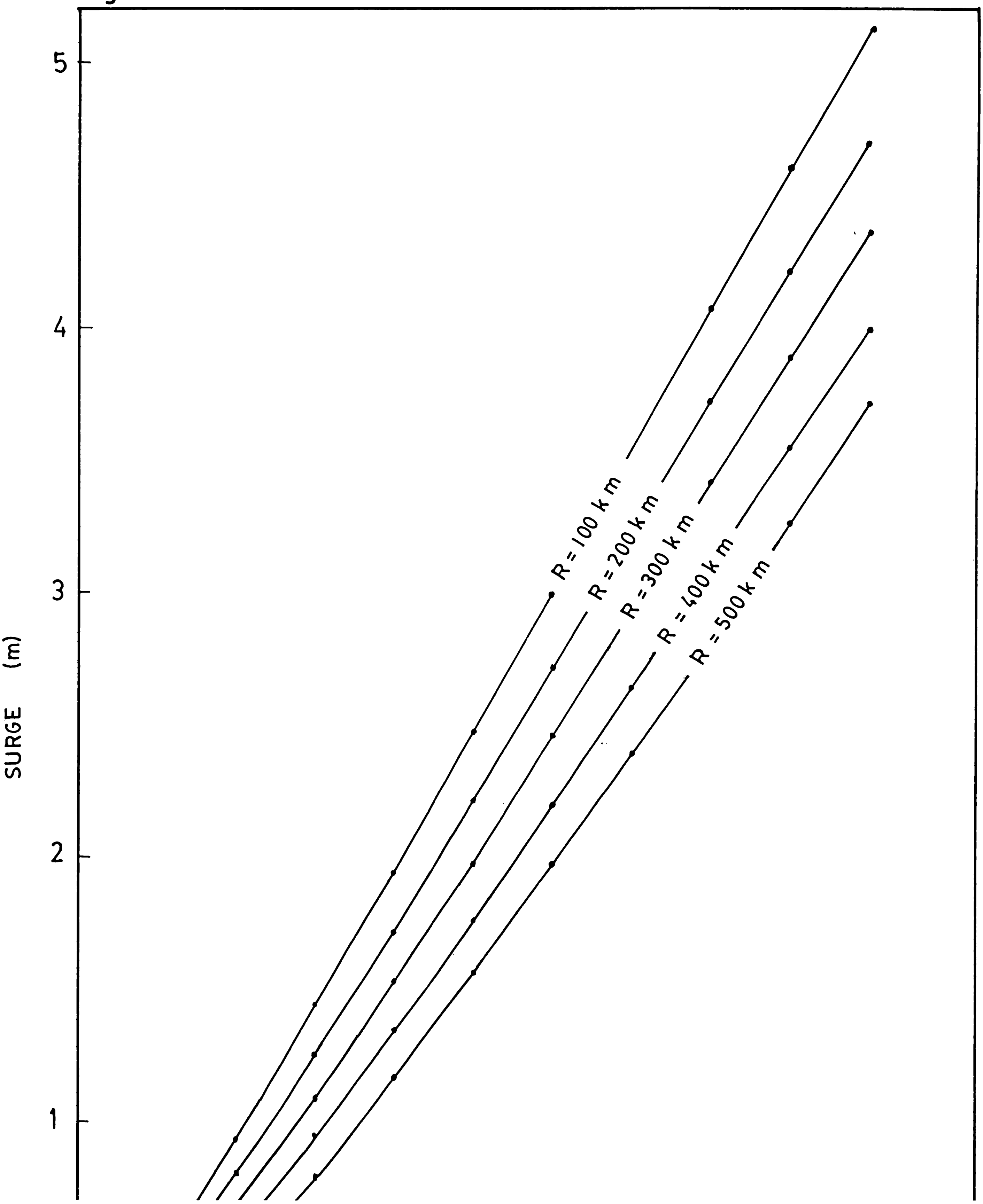


Fig.69: SURGE AT PARADIP FOR ANGLE OF INCIDENCE - 90°.

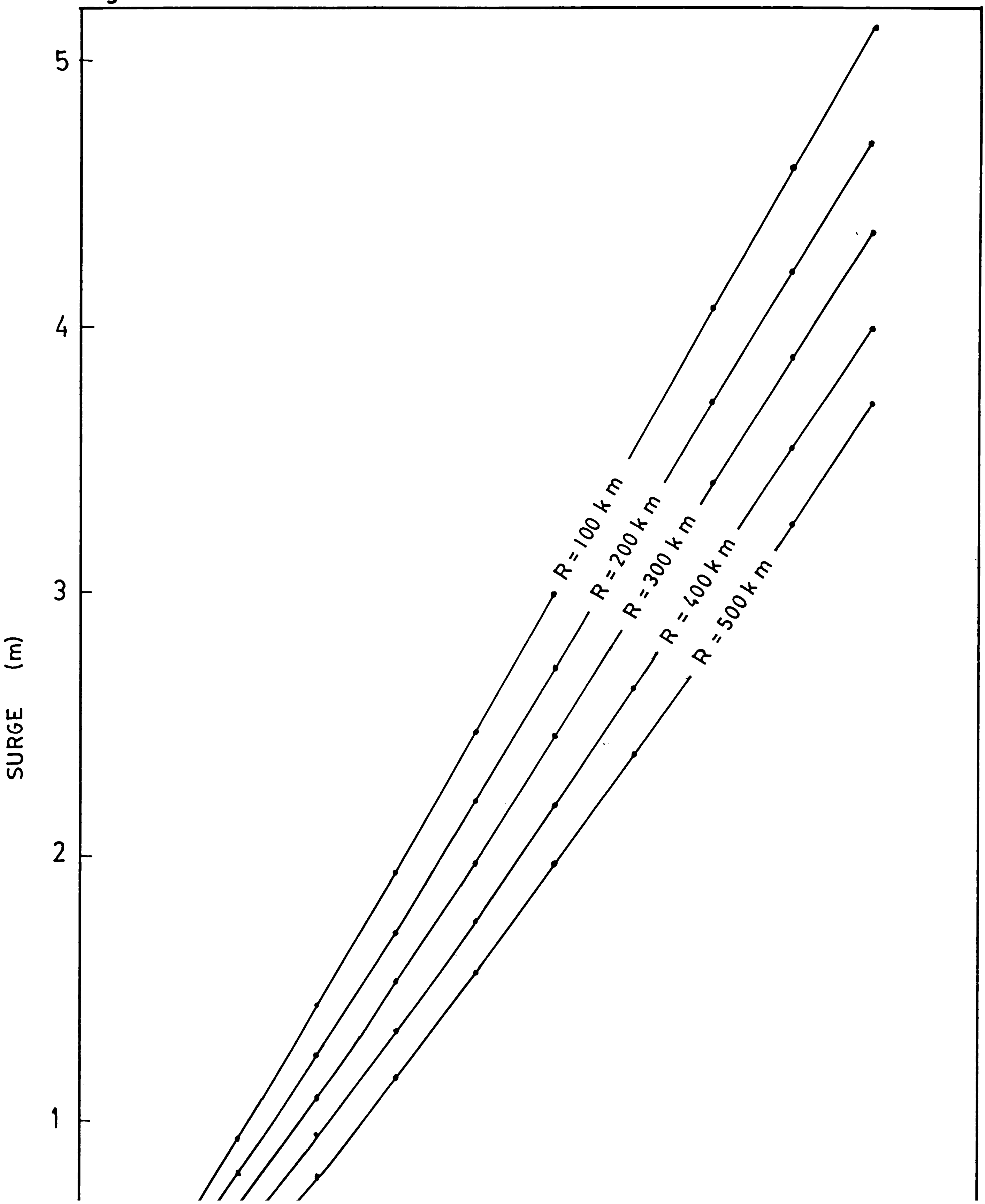
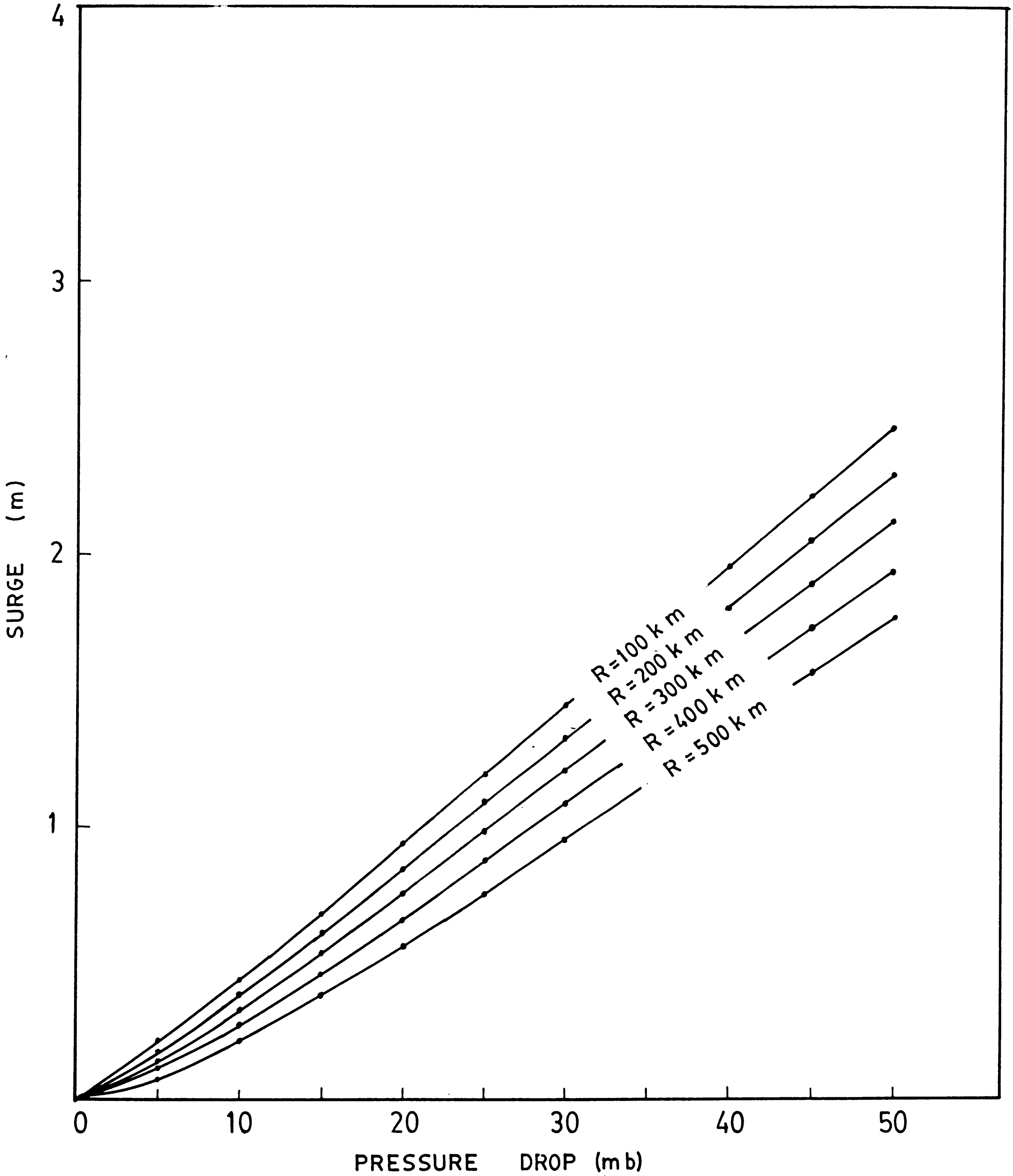


Fig. 70. SURGE AT PARADIP FOR ANGLE OF INCIDENCE - 135°.



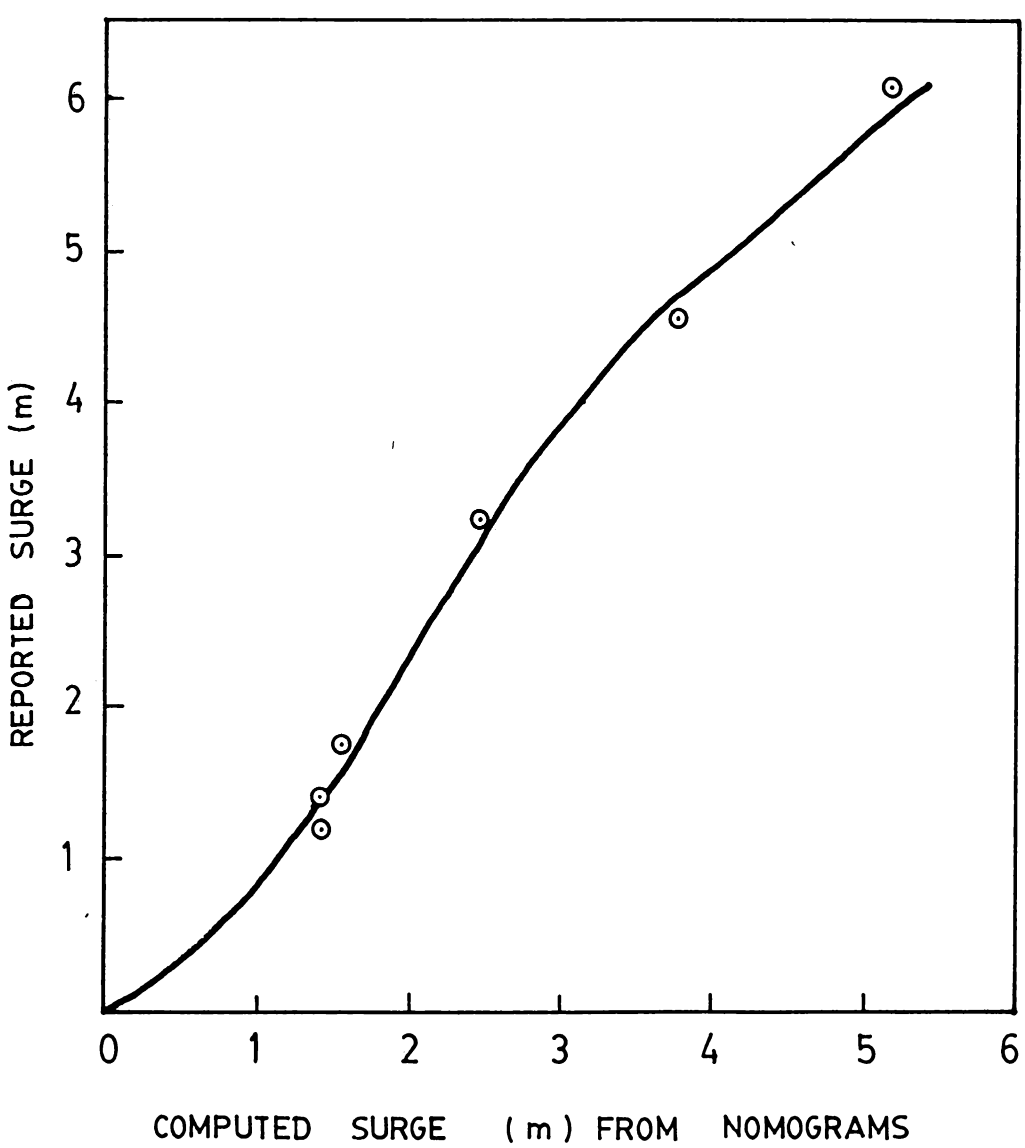
of incidence is less at both the places. The main factor that contributes to the surge height at these places is the bathymetry. An examination of the bathymetric charts (Figs.19 and 44) indicates a buildup in surge height when the cyclones impinge at angles equal and less than 90° as they travel over shallow regions for longer distances. At Masulipatnam the surge height is expected to be less for the angles of incidence between 100° and 170°, and again a strong buildup is expected when the angle of incidence is beyond 170°. At Paradip the surge is expected to increase with the decreasing angle of incidence and decreases with increasing angle of incidence from 90°. In the order of magnitude of surges generated at other places they can be ranked in the descending order as Nagapatnam, Madras and Visakhapatnam. Obviously, the least surge induced at Visakhapatnam is mainly because of a very narrow continental shelf (Figs.1b and 44). As a result, irrespective of the intensity of the cyclones the surge generated at Visakhapatnam is extremely small. It is also conspicuous that the variation in the angle of incidence does not change the surge height much, except when a cyclone impinges the coast at angles less than 90° or greater than 180°. Such a situation seldom occurs at this place. That is why there is not even a single incidence of storm surge reported at this place. The common feature between Madras and

Nagapatnam is that with increasing angle of incidence the surge height increases due to bathymetry. Perhaps, the effect of the storm surge at Nagapatnam may be the least when the angle of incidence is 90° and increases with decreasing and increasing angles of incidence from 90°. However, the increasing angle of incidence from 90° may be more effective than decreasing angle of incidence because the width of the continental shelf increases southward from Nagapatnam. At Madras the surge height is likely to increase further beyond the angle of incidence of 135°.

4.6. Correction for nomograms

As pointed out earlier, it is expected that the nomograms serve the purpose of predicting the surges that are produced by cyclones impinging the east coast of India at different angles of incidence and varying intensities in the neighbourhood of the stations for which the nomograms are worked out. In view of the limitations involved in the preparation of the nomograms, it becomes necessary to draw a correction graph to give a valid prediction. The correction graph is drawn between the surges that were observed in the case studies as well as those that were reported at other times and the computed surges (from the nomograms), produced by cyclones having the same pressure drops, radii of curvature and approximate angles of incidence (Fig.71). It is clear from this graph that at lower values of surge height less than 1.5 m, the computed values are an overestimation and it is an underestimation for the surge heights greater than 1.5 m. Therefore, to get a reasonably accurate prediction, negative and positive corrections have to be applied for the values of the nomograms less and greater than 1.5 m respectively, which can be identified from Fig.71. The overestimation of the surge at the lower values and underestimation at higher values is likely due to the fact that the average pressure gradient at lower pressure drops may be an overestimation and it is an underestimation with intense cyclones of higher pressure drops. The discrepancy at the lower values may sometimes be the result of the visual observation of the surge reported, which may be an overestimate.

Fig. 71. CORRECTION DIAGRAM FOR NOMOGRAMS.



CHAPTER V

D I S C U S S I O N

CHAPTER - V

DISCUSSION

As concerns the interested area of the east coast of India, the characteristics of the storm surge attribute to some meteorological factors, such as pressure drop of the storm, storm size, speed of the storm etc. The variation of surge at a station is not an isolated event, but closely associated with the variations of the surge in the cyclone affected area. Consequently, the surge of the interested area is the resultant of complex variations arising out of various factors. Of these, the most prominent are the wind, bathymetry of the place, orientation of the coast with reference to the track of the cyclone.

In chapter II surges are computed for the past incidents of the cyclones impinging on the east coast of India taking into account the winds computed from the pressure gradients as depicted in the synoptic maps, the actual bathymetry and the track of the cyclone. In all the cases, the surge generated is maximum on the right hand side of the track and minimum on the left hand side, while the surge height along the track has a magnitude in between these two extremes. As the coast under study runs southwest to northeast the tracks in all the case studies are oriented in southeast to northwest direction. The winds in cyclones having the orientation of the track specified above contain mostly onshore or parallel components with southerly direction on the right, and offshore and northerly ~~components~~ on the left of the track. Under the control of onshore components, sea water is driven and piledup toward the coast, and the winds parallel to the coast having a northerly component, with the influence of the Coriolis force, deflect the sea water towards the coast in the northern hemisphere. Similarly the offshore winds and those blowing parallel to the coast with a southerly component drive the water offshore. Thus, the surge buildsup on the right of the track and resurgence takes place on the left of the track in the northern hemisphere. Therefore, the peak surge and the resurgence are the result of a two fold effect.

The peak surge for 19 November 1977 cyclone is estimated to be 6.6 m and occurring at the time of landfall at a distance of 40 km to the right of the landfall, whereas the mean surge along AB, CD and EF (Fig.20) is 4.8 m in the present study. However, Pant et al. (1980) reported the surge of more than 5.0 m above the ground. The ground heights given in the Survey of India Charts in the locality indicate a gentle slope of about 0.5 m per km. As the reported surge (greater than 5 m) is above the ground at Hamsaladevi which is 1.5 km from the coast line, the total surge above the mean sea level after ground correction comes to more than 5.75 m. This figure still appears to be an underestimate because no visual reports are available exactly at the time of flooding but after a day or two when the survey teams could go to the region and supply this information. By the time the survey teams or the local people observed this, it appears some water had already receded to the sea. Under these circumstances the actual surge occurred would have been more than 6 m.

Subbaramayya et al. (1979), who surveyed the affected area, worked out the surge using some empirical formulae developed by earlier workers. They came to the conclusion that all the predicted values were about 2/3 of the observed surge. They attributed that the discrepancy might be partly because the actual central pressure at the time of landfall could have been less than the estimated. The approximate value of the predicted surge, according to them, is 4.2 m. Therefore, the observed surge should have been around 6.3 m which agrees well with the present estimate.

Ghosh (1981) mentioned that while applying his nomograms to the surge of Andhra Cyclone, taking the radius of maximum winds, the angle of track of the cyclone relative to the coastline and speed of movement of the cyclone respectively, as 45 km, 90° and 15 km per hour, it works out to 5.2 m. This lower value may be due to the fact that the actual angle of incidence of the cyclone appears to be greater than 90°. As the effect of bathymetry increases, especially in the area of Divi beyond 90°, a value higher than 5.2 m is expected.

The numerical model of Johns et al. (1981) gave the surge height of 4.8 m at Divi at t = 55 hrs. that corresponds to the morning of 19 November 1977. While the surge height computed by this model agrees with the survey reports, this model appears to produce the peak elevation much in advance of the landfall and reported time of flooding. Also this model produced the maximum surge of 5.3 m at Kavali on 18th afternoon instead of at Divi on 19th afternoon where several thousands of people were killed. Therefore, the model of Johns et al. (1981) is unlikely to simulate the actual conditions, perhaps, because of the differences in the actual and theoretical cyclones. The assumption of boundary conditions in the numerical models are also artificial and are not at par with the actual circumstances. In a subsequent paper Johns *et al.* (1982) also arrived at a similar conclusion.

The present value also agrees with that of Dube *et al.* (1981), who computed this surge as 6 m using the formula of Jelesnianski (1972).

The occurrence of the peak for the 1976 Masulipatnam Cyclone (Fig.26), coinciding with the time of landfall may be the consequence of the fore-runner. Sharma and Satyanarayana Murty (1979) also drew a similar conclusion. Unfortunately, not much information on the surge induced by this cyclone is available nor any research work is reported on this cyclone. Therefore it could not be possible to make any comparative study with that of others.

The surge induced by Kavali Cyclone of 15-17 November 1976 maintains a higher sea level indicating the presence of the fore-runners and a relatively abrupt decrease appears to be the consequence of resurgence. According to Isozaki (1969,1970), the fore-runners play an important role when the movement of the approaching cyclone is slow.

The computed surge, in the case of the Contai Cyclone of 8-19 September 1976, is relatively less than the reported surge of more than 2 m (Pant *et al.*, 1978). Perhaps, the reported value of the sea level might not be the surge because subsequently a tidal bore was likely to be generated in the Hooghly river due to the all time record rainfall of 29 cm on 16 November 1976. The trend of the surge curves indicates a fore-runner before the landfall and a resurgence consequent to it. As explained earlier, the fore-runner may be attributed to the slow movement of the cyclone.

As the tide gauge records are not available at the exact place of landfall, a quantitative comparison of the magnitude of the mean residual sea level with the wind setup in case studies is not possible, although, qualitatively they can be compared with reference to the relative variations and times of their occurrence. The time of occurrence of the peak and the fall in the wind setup along the different parallels of the storm track coincides with the corresponding mean residual tidal structure in general for all the cyclones.

Relatively very high wind setup for the Andhra Cyclone of 1977 (6.6 m) compared to the Masulipatnam and Kavali Cyclones of 1976 (< 2.5 m) is not merely due to the difference in wind speeds but appears to be the consequence of the shallowness of the shelf between Ongole and Nizampatnam where the Andhra Cyclone crossed the coast.

While the 100 fathom depth contour is at a distance of about 80-85 km from the coast along the track of 1977 Cyclone, it is at a distance of only 40-50 km for the tracks of Masulipatnam and Kavali Cyclones in 1976 (Fig.1a). Apart from the shallowness of the shelf, the orientation of the coast with reference to the approach of the cyclone may also be responsible for the difference in wind setup between the cyclones. All these cyclones crossed the coast at an angle of 125° to 130° from north. While considering the numerical model for prediction of storm surges along the Gangetic Coast, Flierl and Robinson (1972) discussed that the coastal geometry and bottom topography are important in determining the surge response on the open coast.

The seasonal sea level variation along the east coast of India is primarily controlled by monsoons, cyclones, rainfall, and upwelling and sinking. It is already explained that the seasonal sea level variation along the east coast has a distinct variation between the regions in the south (Fig.49) and north (Figs.50 to 52) with a semi-sinusoidal and sinusoidal variation respectively. This difference is mainly because of the occurrence of upwelling and sinking in the south while no such phenomena takes place in the north. The two regions also react differently to the two monsoons. The southwest monsoon winds induce onshore transportation in the north while they produce offshore transport in the south. Similarly while the northeast winds generate offshore transport in the north they give rise to pilingup of water against the coast in the south. The present results in general agree with those of Prasada Rao and La Fond (1954) and Ramanadham and Varadarajulu (1969) for Visakhapatnam. However, Prasada Rao and La Fond (1954) attribute it mainly to the steric effect while Ramanadham and Varadarajulu (1969) considered it due to rainfall. In fact, the influence of rainfall is automatically incorporated in the steric effect. Nevertheless, it is doubtful if the influence of the steric effect or the rainfall would give rise to such a drastic variation of more than 50 cm in the sea level (Kesava Das, 1979). Not much information is available to compare the sea level variations at the northern stations except that of Prasada Rao and La Fond (1953). They pointed out that the amplitude of the annual variation is the largest in the north. As per their results the annual range in the sea level at Kidderpore is about 170 cm which agrees with the present results (Fig.51). The occurrence of the maximum and minimum in July-August and January-February respectively tally with the present results.

As far as the eustatic changes in sea level are concerned, there is a trend of rise in sea level during 1950's when compared to 1940's along the east coast of India (Figs.53 to 55). According to the present study, the maximum variation in mean sea level at Visakhapatnam from 1947-1955 is 13 cm and from 1955-1957 is 10 cm. The mean annual sea level change observed by Ramanadham and Varadarajulu (1964) at Visakhapatnam between the years 1955-1957 was 9.5 cm which agrees with the value of the present study (Fig.53a).

With the help of the present nomograms, the surges are computed, for the past cyclones that crossed the east coast, for their intensity and angle of incidence. The computed and the reported surges are presented in Table IX. The computed surge is agreeing extremely well with the reported.

Table IX. Comparison of reported and computed surges

Cyclone							Surge (m)	
Sl. No.	Year	Date	Maximum wind (kt)	Pressure drop (mb)	Approx. angle of incidence	Area affected	Reported	Computed from nomograms
1.	1679	Oct. 13	120	55	135^{o}	Masulipatnam	6.00	5.60
2.	1864	Nov. 1	115	45	135^{o}	Masulipatnam	4.00	4.40
3.	1949	Oct. 27	80-90	30	135^{o}	Masulipatnam	3-3.5	2.90
4.	1969	Nov. 4-9	95	32	135^{o}	Between Masulipatnam and Kakinada	3.00	3.02
5.	1972	Sept.7-14	110	43	135^{o}	Paradip	2.75	2.20
6.	1977	Nov. 19	120	66.3	135^{o}	Divi	>5.00	6.05

The cyclone information is taken from Angelis (1978) and Ananthakrishnan (1964)

CHAPTER VI

SUMMARY AND CONCLUSIONS

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SUMMARY AND CONCLUSIONS

The present thesis is an outcome of some of the investigations carried out by the author on storm surges and sea level variations along the east coast of India, as it is one of the regions in the world most vulnerable to storm surges. Its principal aim is to identify the areas that are maximum affected by the cyclones and suggest a method by which the surge of an impinging cyclone can be predicted with the knowledge of the pressure drop, the approximate size and the angle of incidence of a cyclone. The aim is achieved by estimating the wind induced surge for four cases of past cyclones that attacked the east coast of India and comparing the estimates with the reported surges to qualify the method of approach. Having realised that the method of approach gives reasonably satisfactory results, nomograms of the surge against the pressure drop are prepared for five representative stations for varying pressure drop and different radii of curvature of the cyclones impinging the east coast of India at angles of incidence of 90° and 135° from the north.

The wind induced surge is computed from winds, estimated using the gradient wind equation, and actual bathymetry along the path of the cyclone. The pressure gradient is obtained from synoptic maps. In order to assess qualitatively the time of occurrence of the peak surge and its order of magnitude, residual sea level is calculated by subtracting the hourly astronomical tide taken from the tide tables, from the recorded tide at the available tide gauge stations. As the data from the tide gauges is available only at selected stations, a qualitative study can alone be carried to qualify the validity of the results of the wind setup in the case studies.

The mean seasonal and secular sea levels are presented using the sea level data supplied by the permanent service to Mean Sea Level, I.O.S., Bidston observatory, U.K. The purpose of working out the mean sea level is to have an idea of the tendency of the change in mean sea level and to understand the combined height of the sea level by superimposing the wind induced surge on the mean sea level.

According to Wang and Liu (1982), the peak of the surge always appears on the right of the landfall at a distance of approximately equal to the radius of the maximum wind. It is a remarkable coincidence that the peak of the surge in all case studies is noticed on the right of the landfall at a distance of about 40-60 km which is approximately equal to the radius of maximum wind. In most of the instances the peak occurs at the time of landfall on the right of the track, but on the left and along the track the peak is sometimes noticed before the landfall. Normally, the resurgence commences earlier on the lefthand side of the landfall. Far left of the track, negative surge is occasionally observed. The surge progress is mainly dependent on the strength of the onshore component of the wind and bathymetry, rather than the tide.

Although, no clear-cut correlation is observed between residual sea level and surge at landfall, in general, the times of occurrences of higher sea levels almost agreed. It is also found that the effect of short period variations in the residual curves are very large during the period of cyclones when compared with that of the normal days. This information may sometimes be useful for correlating to the magnitude of the surge that occurs within 200 to 300 kms.

The mean sea level in the southern region (Fig.49) of the east coast of India depicts a semi-sinusoidal variation with the primary maximum and minimum in October-November and March-April respectively. The secondary maximum and minimum occur during May-June and July-August respectively. The sea level in the northern region shows a sinusoidal variation with maximum in August-September and minimum in January-February respectively (Figs. 50 to 52). As far as the eustatic changes are concerned they do not have much influence on the small scale time variation of sea level which is important in the surge prediction.

A close comparison of the nomograms (Figs.61 to 70) with that of the shelf width (Fig.1) clearly reveals that higher surges occur at places of wider continental shelf and least surges at narrow shelf. Thus, the maximum surge that can be generated at Masulipatnam and the least at Visakhapatnam are the results of such conditions. However, it is just not enough if we consider only the width of the continental shelf but the orientation of the cyclone track with reference to the coast is also equally important. It is evident that the surge induced at an angle of incidence of 90^{o} (Fig.65) is almost twice that of the one produced by a cyclone of the same intensity when the angle of incidence is 135^{o} (Fig.66) at Masulipatnam. Therefore, it is of particular significance to be cautious about the angle of incidence of the cyclene in the regions of broad continental shelf. On a similar analogy it can also be confirmed that the orientation of the cyclone track has a little effect at the places of narrow continental shelf. At Visakhapatnam the variation of the surge either with the intensity or the angle of incidence of the cyclone, is highly marginal. Therefore, it can be concluded that the angle of incidence is a significant factor when the shelf is broad. Under such circumstances, it is all the more important to consider the coastal configuration in the regions of wider continental shelf where it has a crescent-moon shape like that of Nizampatnam Coast. Obviously, the region near Nizampatnam (Fig.19) with a flat based shallow island is one of the most vulnerable areas to storm surges, particularly, when the angle of incidence of the cyclone is almost parallel to the coast. In such a place the mass transport of water by the wind swells up and superimposes on the peak surge because of funnelling effect. Similarly, Nagapatnam is another vulnerable area to the storm surge if the angle of incidence is about 180° and the effect is even worse if a cyclone impinges into the Palk Bay at an angle of incidence of less than 90°. The angle of incidence is equally important for the region around Paradip when it is either 90° or less than that. But, for higher angles of incidence the effect is not prominent.

The water elevation of a surge is decided primarily by the strength of the cyclone and related closely with the bathymetry. To simplify the problem, in practice the horizontal pressure gradients at sea level are, generally, used as the main index which represents the sea surface wind. Because each port has its individual geographic conditions such as geographic position, coastline configuration and bathymetry, the amplitude of the storm surge depends on the action of wind in certain direction. The topographic effect on the surge is remarkable. The values of the surges vary with the various tracks of the cyclone and landfall positions. For different types of cyclones, the time of occurrence of the peak surge is different. When the surge coincides with the high tide, the occurrence of the peak is slightly earlier than when it coincides with the low tide. Apart from the effect of storm surge, near the mouths of the rivers, the combined effect of storm surge and tidal bore are important in inland flooding. In the case of Contai Cyclone, while the computed surge is only 1.3 m, the reported surge is more than 2.0 m, which may be due to the tidal bore that might have been generated by an all time record rainfall of 29 cm in the Hooghly region in one day.

While predicting the surge from the nomograms, an element of caution is invariably essential as the limitations in the nomograms should not be underestimated.

The gradient wind equation does not give the direction of the wind; as a result of which it is not possible to identify the onshore component that gives the maximum surge and the offshore component that reduces the surge height. The available local bathymetric charts used are that of 1959, and the present bathymetry might have been changed substantially due to either siltation or erosion. Further, the pressure drop gives only an average pressure gradient over the whole radius of curvature of the outermost closed isobar. These limitations are to a certain extent eliminated by the correction graph.

The nomograms give only the wind induced surge, but it is the combined effect of the variation of sea level that is to be considered at the time of prediction. Therefore, the resultant of the sea level at the time of the cyclone impinging a coast should be the algebraic sum of the contributions to the sea level by steric effect, astronomical tide and inverted barometric effect besides the wind induced surge. Of these contributions, except the steric effect which does not contribute much to the sea level unless there is heavy rainfall, the rest can easily be obtained. The astronomical tide can be known from the tide tables and the inversed barometric effect is approximately one cm rise in sea level for every mb drop of atmospheric pressure which can be roughly estimated before the cyclone impinging the coast either from the radar or satellite pictures. Ofcourse, the wind induced surge can be estimated from the nomograms if the pressure drop, the approximate size of the cyclone and the angle of incidence of the cyclone are known. Besides, one should also consider the mean seasonal variation of the sea level over which the total contribution of the variation in the sea level is to be superimposed. For that matter, the annual range of the mean sea level along the east coast varies from about 0.3 m to 1.7 m. Unless this variation is taken into account while predicting the surge height, it may drastically mislead the prediction.

The author presumes that the present nomograms are the simplest of all the available nomograms for prediction of the storm surge, including that of Ghosh (1977). His attempts will be better rewarded with more accurate prediction if the nomograms are constructed for more angles of incidence at many other places with modified bathymetry.

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