A Three-Dimensional Numerical Model of the Sea Breeze for the Plymouth Region

Ian W. Clark

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A Three-Dimensional Numerical Model of the Sea Breeze for the Plymouth Region

by

Ian W. Clark
B.Sc. M.Sc.

Submitted to the Council for National Academic Awards
in partial fulfilment of the requirements for the
Degree of Doctor of Philosophy

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In collaboration with the
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R.A.F. Mount Batten,
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ABSTRACT

A three-dimensional, hydrostatic, primitive equation model is developed to simulate the Plymouth sea breeze. The equations are integrated forward in time on a staggered mesh with a domain of 64km X 64km X 3km, using a combination of central and upstream differencing. The ground surface is assumed to be smooth and the heat input to the atmosphere transferred vertically without the explicit use of diffusion coefficients. This results in a less stringent stability condition on the timestep, thus reducing the computational cost of the simulations.

Sensitivity tests for a two-dimensional version are presented, examining the influence of atmospheric stability, the synoptic scale flow and the magnitude of the surface heat flux. The major features of the mesoscale circulation are well represented including the strong overland updraughts associated with the sea breeze front. Frontal propagation rates are estimated in each simulation and are found to be in general agreement with available data for Southern England.

The preliminary three-dimensional results concern the sensitivity of the model to variations in the synoptic scale flow and the coastal configuration. The former tests show a more vigorous system developing with an offshore synoptic flow and a much weaker circulation for the onshore. The second test illustrates the development of a bay-induced landward bulge in the temperature gradient resulting in an asymmetric distribution of onshore convergence zones.

The final simulations represent two case studies of Plymouth sea breeze events during August 1983 and May 1984. The major features of the system are again well simulated, however several key problem areas are identified. These involve the influence of topographic variations, the numerical grid resolution, the heat flux parameterisation and the role of turbulent transfer.

Recommendations for further research are proposed and include the application of a terrain-following coordinate scheme and a new observational initiative. In addition, the need for an improved heat flux parameterisation and turbulence closure are identified.
ACKNOWLEDGEMENTS

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Finally, I am indebted to my wife Elizabeth for her love and support when I most needed it.
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CHAPTER 1 - INTRODUCTION

1.1 Mesoscale Motions and the Sea Breeze

Atmospheric fluid motions exhibit a wide spectrum of horizontal scales, ranging from earth size dimensions down to scales comparable with the mean free path of the individual molecules. Within this range, the sea breeze circulation can be classed as a mesoscale motion. These are defined as waves, eddies or jet-like features with horizontal dimensions ranging from a few tens of kilometres up to a few hundred (Wallace and Hobbs (1977)).

Mesoscale meteorology has received a great deal of attention in recent years as evidenced by the proliferation of not only academic papers but also conferences and symposia (Nowcasting (1981, 1984)) and general textbooks (Atkinson (1981), Pielke (1984)). The reason for this expansion lies in the knowledge that to achieve an understanding of atmospheric motion as a whole the various scales which contribute must be assessed individually and in their interactions with each other. The FRONTIERS program under development by the Meteorological Office is attempting to provide increased forecasting accuracy on the local variations of weather in the period up to a day ahead (Browning and Carpenter (1984) and Carpenter and Browning (1984)). This complex system involves an interactive network of radar and satellite observations coupled with a mesoscale numerical model allowing better resolution of the mesoscale features impinging on the UK. The development of FRONTIERS is a reflection of the increased awareness of the necessity to improve the knowledge and understanding of mesoscale disturbances.
The sea breeze is a thermally-induced circulation which develops in the following manner. In a sufficiently calm, clear atmosphere solar radiation heats up the land surface more efficiently than the adjacent water surface creating a horizontal temperature gradient. As a consequence, the overland air heats up and expands more rapidly than over the corresponding water surface and, hence, a large vertical gradient of pressure prevails over the colder air above the water. This results in a slight flow of air aloft from land to sea as illustrated in Figure 1. Convergence occurs at the point C due to an increase in pressure and a significant departure from hydrostatic equilibrium at that level. With this subsidence, a return flow develops from D to A due to the formation of a hydrostatic pressure gradient; this is the sea breeze component of the system. The circulation cell is completed with divergence at B causing air to rise from A in response to the change in hydrostatic equilibrium. The optimum conditions for the development of this type of thermal breeze system is anticyclonic summer weather when almost cloudless skies, high insolation and weak gradient winds combine to allow the maximum differentiation between surface climates.

The above description is of the 'pure' sea breeze circulation which will inevitably be modified to some degree by a range of environmental controls. The prevailing ambient atmospheric state is a major factor, with the magnitude and direction of the synoptic scale flow and atmospheric stability being the dominant influences. Other interactions involve surface friction, Coriolis effects and the influence of topography and coastal configuration. The degree by which these factors undermine the 'pure' sea breeze will be discussed in subsequent chapters.
1.2 Plymouth and its Environs

Plymouth is situated on the south coast of Devon between the uplands of Dartmoor and the English Channel, with the city itself lying at the confluence of the southward draining rivers Tamar and Plym (see Figure 2). Plymouth Sound forms a major part of the extensive ria physiography found in South West England. The dimensions of the bay are approximately 5km at the entrance narrowing to about 3km at the northward end, with a north-south extent of 5km. The coastline itself is a particularly rugged combination of steep cliffs rising to over 100m and long, curvilinear beaches.

As a consequence of the complicated coastal shape and the extensive granite moors, the sea breeze regime in this region is extremely complex. It is likely that the influences of topographic steering and anabatic effects will exert a strong influence on the systems development. These features will be discussed at greater length in subsequent chapters. The sea breeze itself and the climatology of the region will be examined in Chapter 6.

1.3 Literature Survey

The major reference works in the numerical modelling of the sea breeze system are the extensive bibliographies of Baralt and Brown (1965) and Jahn (1973). Unfortunately, both these reports are somewhat out of date due to the proliferation of sea breeze modelling in the late 1970's. Perhaps more appropriate reference sources are the recent textbooks by Atkinson (1981) and Pielke (1984).
Early theoretical work on sea breezes prior to the 1950's involved the development of linearised models of the fundamental atmospheric equations (Haurwitz (1947), Schmidt (1947), Defant (1950) and Pierson (1950)). These models helped to elucidate the major factors affecting the development and formation of the sea breeze circulation but were of limited applicability because of their inherent inability to cope with the advection terms in the governing equations. Since the system under consideration is thermally forced, the feedback between the temperature and velocity fields is of primary importance.

With the advent of high-power computers in the 1950's, a great deal of effort was directed towards the development of mathematical models of all atmospheric systems at all scales. The first sea breeze model was developed by Pearce (1955); the framework of this model provided a basis for all future numerical analysis, i.e. the pressure, velocity and temperature fields were integrated forward in time from an initial isothermal, static state. The forcing involved the differential heating of the surface across a long, straight coastline with vertical distribution by convection currents. Pearce noted two motions resulting from this procedure; firstly, a tidal motion of large horizontal extent and low velocity and, secondly, a smaller, more vigorous component with higher velocities and restricted dimensions. The former feature he considered to be responsible for the development of continental lows and the latter he identified as the pure breeze.

The next major development was the work of Fisher (1960, 1961) who introduced a parameterisation for surface friction and a more realistic heating function. This was the forerunner of the K-type models, where diffusion coefficients are employed in the prognostic equations. Although severe numerical instability was encountered, this was the earliest attempt to compare the results
with observations. Contemporaneously, Estoque (1961, 1962) introduced a
dual-layer formulation involving a constant flux layer of 50m depth characterised
by constant heat and momentum diffusion coefficients and a transition layer of
around 2km in depth, within which the governing equations were solved. His
major concern was with the relationship between the prevailing synoptic state
and the heat flux from the land surface and the intensity of the ensuing
circulation cell. Although the results were plausible, he was forced to employ
an incompressible form for the continuity equation to maintain model stability.
The implications of this will be discussed in the following chapter.

The model of Estoque was used and modified extensively in the 1960's and early
1970's. For example, Magata (1965) introduced a parameterisation for the effects
of non-adiabatic processes involving latent heat exchange and introduced vertical
shear in the general synoptic flow. Moroz (1967) adapted the dual-layer scheme
to the Lake Michigan breeze problem. In this study, reasonable agreement with
observations was found; however, the author did note difficulties in resolving all
the transient features of the turbulent transport of heat and momentum. The
first three-dimensional model was introduced by McPherson (1970), in which he
examined the role of coastal shape on the sea breeze circulation. He noted, in
particular, an asymmetric distribution of convergence zones around a simulated
bay relating these to preferred locations for convective shower development.
Nordlund (1971) introduced an adaptation of Estoque's scheme involving the use
of upstream differencing in the solution of the advection terms. The results of
his study for Southern Finland were encouraging; however, he did note stability
problems which restricted the prognosis to a maximum of 12 hours. Tingle
(1971) simulated the Lake Michigan breeze in order to assess the movement of
pollutants around the lake. More details of this analysis were reported in
Dieterle and Tingle (1976). The first non-hydrostatic model was introduced by
Neumann and Mahrer (1971); their simulations examined the land and sea breeze cycle over a three day period. The authors also used the original continuity equation, thus maintaining mass continuity in the simulations.

Other models to appear in the early 1970's included Clarke (1973), who examined the energetics of the sea breeze. Recognising that the major component in the energy budget was that due to vertical motions, he concluded that the sea breeze 'engine' was not widely different from larger scale processes. Pearson (1973, 1974) attempted to elucidate certain properties of the sea breeze front; his major conclusion was that frontal propagation was mainly dependant on the magnitude of the heat input into the atmosphere on a particular sea breeze day. Lambert (1974) also looked in more detail at the frontal structure of the sea breeze and noted that only very high resolution simulations will correctly predict the strong updraught velocities found in this region.

Perhaps the most influential sea breeze modelling research was that reported by Pielke (1973, 1974a); in an analysis of the sea breezes over Florida, the author developed a new three-dimensional model involving the decomposition of the dependent variables into three components, namely the synoptic scale, mesoscale and microscale. By analogy with Reynolds averaging, the averages of the latter components over a grid volume were considered to equal zero. Furthermore, by recognising that the synoptic scale changes relatively slowly with time, the synoptic scale dynamic terms were substracted out of the governing equations leaving a set solving only for the grid averaged mesoscale perturbations. The Florida simulations showed a fair agreement with observations especially with regard to the preferred locations of convective shower development over the peninsula and the influence of Lake Okeechobee.
The structure of the Pielke model proved to be the basis for a series of studies of particular sea breeze problems and detailed examination of the assumptions and solution methods employed. Due to the increased complexity of the model, computations were somewhat expensive and a two-dimensional version was developed (Pielke (1974b)). The main conclusion from the analysis was that full three-dimensional calculations were necessary for most practical sea breeze problems. Despite this, Mahrer and Pielke (1975) developed a two-dimensional sigma coordinate model for the simulation of flow over irregular terrain. In their study of flow over a bell-shaped mountain, a drawback was noted concerning the inability of the scheme to simulate non-hydrostatic features such as lee-waves. Pielke and Mahrer (1975) outlined a further parametric study involving the representation of the heated planetary boundary layer. In this study a prognostic equation for the planetary boundary layer depth proposed by Deardorff (1974) was found to be superior to the original diagnostic equation. Furthermore, the growth of the planetary boundary layer into a region with significant horizontal wind shear was shown to exert an influence on the locations of sea breeze convergence zones. The results were also compared with data from the Wangara experiment (Clarke, Dyer, Brook, Reid and Troup (1971)) and good agreement was found.

The problem of preferred locations for convective shower development was further analysed in Cotton, Pielke and Gannon (1976), where the one-dimensional transient cumulus model of Cotton (1975) was coupled with the sea breeze model. The first three-dimensional simulation using sigma coordinates was described in Mahrer and Pielke (1976) which gave an analysis of the airflow over Barbados. The results clearly identified the importance of topography in modifying the low-level airflow. A more general discussion of this numerical scheme can be found in Mahrer and Pielke (1977a). The latter paper also
discussed several other modifications involving a surface heat budget parameterisation and shortwave and longwave radiative fluxes. More verification of the three-dimensional sigma coordinate model was presented in Mahrer and Pielke (1977b), where the simulations were correlated with data from the White Sands Missile Range in New Mexico. The updated model was also compared with the original data for the Florida problem and presented in Pielke and Mahrer (1978). In the same context, Gannon (1978) examined the influence of the earth’s surface and cloud properties on the South Florida sea breeze using a two-dimensional version. His main conclusions were that soil moisture was the dominant controlling surface property and that the presence of cirrus clouds could prevent the evolution of the sea breeze. He also noted the importance of cumulus clouds in shielding the surface from solar radiation.

On the more theoretical side, Pielke’s model has been used to assess the application of cubic spline interpolation to the solution of the advection terms in the primitive equations (Mahrer and Pielke (1978)). In addition, Pielke and Martin (1981) gave a general derivation and discussion of the terrain-following coordinate system. The same two authors also looked at the adequacy of the hydrostatic assumption in the modelling of thermally induced systems (Martin and Pielke (1983)). In general terms they concluded that the assumption becomes less valid as the intensity of the surface heating increases and the temperature lapse rate becomes less stable.

In a study related to Gannon (1978), McCumber and Pielke (1981) examined the relationship between the sea breeze and the underlying ground surface, concluding again that the soil moisture is a vital factor as it regulates the strength of the heat fluxes between the atmosphere and the ground. More details on this subject are given in Ookouchi, Segal, Kessler and Pielke (1984)
and Segal, Pielke and Mahrer (1984). In general, they concluded that circulations induced by large contrasts in soil moisture over flat terrain can be equivalent in magnitude to that of a sea breeze.

In recent years, the detailed numerical model based on Pielke's original work has been applied to a range of mesoscale flows around the world. For example, Alpert, Cohen, Neumann and Doron (1982) carried out simulations of the summer circulation in the Lake Kinneret region of Israel, with increased resolution in the planetary boundary layer, a modified formulation for friction and heating and changes in the numerical solution. Several important flow features were well simulated, including the development of strong winds on the western side of the lake; however, the authors did note problems in predicting the correct times of onset. Segal, Mahrer and Pielke (1982) described an application of the three-dimensional sigma coordinate model to the heterogeneous terrain of Central Israel in an attempt to assess the wind energy characteristics of three typical synoptic situations. Alpert and Neumann (1983) extended their simulations to the Lake Michigan winter land breeze. The available observations show the simulated results to be reasonably accurate, with the updraught core features especially good. This also appears to be the first documented example of a numerically predicted land-breeze front. Two other recent applications of the model are the simulations of the meteorological patterns associated with the Dead Sea (Segal, Mahrer and Pielke (1983)) and the effect of an ocean surface temperature gradient on the mesoscale atmospheric circulation (Mizzi and Pielke (1984)).

Although the aforementioned papers formed the backbone of sea breeze research in the 1970's and early 1980's, several other important papers appeared in related areas. Glatt (1975), for example, developed a numerical sea breeze simulation
under inversion conditions by introducing an additional term into the continuity equation to account for large scale subsidence. Sheih and Moroz (1971, 1975) continued the research on the Lake Michigan problem, with improvements to the Moroz (1967) model, including the eddy diffusivity formulation, the boundary conditions and the numerical solution scheme. A very detailed non-hydrostatic model was introduced by Tapp and White (1976), with the numerical results presented in comparison with those of Pielke (1974). Good qualitative agreement was found and the authors proposed that full non-hydrostatic equations can be used with very little increase in computational cost. Physick (1976) introduced a two-dimensional sigma coordinate scheme for looking at the particular problem of lakes and gulfs. His major contribution concerned the detailed heat balance approach he adopted for the forcing of the model. More recently, Physick (1980) also examined the role of varying Bowen ratios on the inland penetration of the sea breeze; in common with several other presented results he found the most significant factor to be the soil moisture.

In addition to these more sophisticated models, Danard (1977) adopted a different approach by introducing a simple one-level primitive equation model for computing mesoscale influences of orography, friction and heating on surface winds, using synoptic scale data for the initialisation procedure. This scheme proved to be most effective in predicting processes involving orographic channeling, but was less so for the sea breeze situation where cause and effect are less clearly defined. Further work on the Lake Michigan breeze problem was reported by Patrinos and Kistler (1977), who used a two-dimensional non-hydrostatic model similar to Neumann and Mahrer (1971). In this modified version, so called Alternating Direction Implicit methods were used in the integration of the governing equations. The use of these methods removed the necessity to apply upstream differencing and spatial filtering, however, it is
difficult to assess whether the extra effort involved produced improved results. In addition to these general analyses of the sea breeze circulation, Maddukuri, Slawson and Danard (1978) examined the boundary layer structure. The model used predicted reasonably consistent profiles for the ambient temperature and horizontal wind fields for the north shore of Lake Erie.

Thus far the majority of the outlined research has originated in the U.S.A., however, over the same period several important papers appeared in Japan. Yoshikado and Asai (1972) carried out a series of numerical experiments to assess the merits of four schemes accounting for the turbulent transport of heat and momentum in a sea breeze model. The authors noted that the primary uncertainty lay with the vertical profile of $K$, with differing schemes producing widely varying results. They also suggested that so called 'penetrative convection' may be an important process in the sea breeze turbulence structure. Saito (1976) used a numerical model after Pielke (1974a) to examine the interactions between an urban heat island and the sea breeze. By specifying a differential temperature change between the urban and suburban areas, the model produced two circulation cells at the coast and at the urban/suburban interface. Saito also documented a simulation of the land and sea breeze regime for the Kanto district. The three-dimensional analysis clearly illustrated the importance of steering by complex coastal configurations. Other Japanese research included an analysis by Kimura and Takeuchi (1978) who examined the problem of sea breeze induced fumigation of pollutants by applying mixing length theory and a simple advection-diffusion model. Ookouchi, Uryu and Sawada (1978) examined the effects of a mountain on the land and sea breeze circulation using a two-dimensional formulation. The authors found that the sea breeze penetrated faster and deeper beyond the simulated isolated mountain than in the flat case and that for the heated mountain, inland penetration was restricted. Asai and
Mitsumoto (1978) looked at the influence of an inclined land surface on the land and sea breeze circulation. The two-dimensional results showed a distinct interaction between the simulated plain, slope and plateau depending on the specified surface heating function. The pollution studies were further enhanced by Kondo and Gambo (1979) who examined numerically the influence of the developing overland mixing layer during a sea breeze event. In general terms, the authors concluded that the sea breeze penetration was enhanced by a well developed mixing layer. They also suggested that frontal updraughts prevented pollutants emitted at the coastline attaining high concentrations within the mixed layer. Kikuchi, Arakawa, Kimura, Shirasaki and Nagano (1981) continued to elucidate the complex structure of the sea and land breeze regime in the Kanto district. Using a three-dimensional sigma coordinate model, the authors concluded that the mountainous nature of the region severely deforms the mesoscale circulations. An interesting feature they also noted was the prediction of an earlier occurrence of mountain and valley winds compared to both the land and sea breezes. Further work on pollution was documented by Ozoe, Shibata, Sayama and Ueda (1983) who used a two-dimensional scheme to simulate the transport of effluent within the land and sea breeze cycle over three day periods. The authors noted in particular the trapping of the pollutants in the closed circulation cells of both systems and distinct diurnal variations in near-field concentration levels accordingly. They also noted longshore drift of pollutants due to the influence of Coriolis forces. In a recent application of the model of Asai and Mitsumoto (1978), Kitada, Carmichael and Peters (1984) examined the transport of chemically reactive species within the land and sea breeze cycle. The transport/chemistry model was highly complex and involved 84 gas-phase and 10 heterogeneous chemical reactions. The predicted results compared favourably with the available profile data of pollutants such as O\textsubscript{3}. In addition, the development of clouds at the sea breeze front was shown to be
important in affecting the complex chemical processes taking place.

A further group of researchers were also looking at mesoscale modelling during the 1970's; Anthes and Warner (1978) first described a two-dimensional scheme as applied by Anthes (1978) to the development of the planetary boundary layer in the sea breeze regime. In terms of pollutant transport, the author noted Coriolis forces as being significant in causing longshore drift. Anthes also noted the model sensitivity to the rate of overland heating and the initial temperature stratification. Warner, Anthes and McNab (1978) reported on further progress with a three-dimensional model as applied to mesoscale systems ranging from the sea breeze and mountain and valley winds to lee troughs. Anthes, Warner and Seaman (1979), using the two-dimensional version, examined the role of diabatic heating for a range of disturbances. In accordance with several other workers, they found this to be the major controlling factor on the strength and dimensions of the sea breeze circulation.

During the late 1970s and early 1980s several important contributions were published independently from the aforementioned research groups. Dalu (1978) and Dalu and Green (1980) developed a two-dimensional sea breeze model without the explicit use of diffusion coefficients or the characteristic dual-layer structure. As a result, the model was computationally less expensive and the results compared favourably with more rigorous schemes. The same authors also commented on the energetics of the sea breeze system; their major conclusion was that for a twelve hour prognosis the kinetic energy was about 25% of the diabatically produced mesoscale available potential energy. The authors were also critical of the upstream differencing scheme employed in terms of numerical efficiency; they estimated that the numerical scheme lost 20-30% of the available potential energy. Clancy, Thompson, Hurlbert and Lee (1979) introduced a
complex two-dimensional model which coupled an atmospheric and oceanic simulation in order to assess the oceanic response to sea breeze forcing, changes induced in the sea breeze by coastal upwelling and any associated air-sea feedback. The authors noted the enhancement of the sea breeze by the colder coastal water introduced by the upwelling and the weakening of the land breeze by the same process. In addition, they also noted that the sea breeze contributed significantly to the longshore wind stress thus enhancing the coastal upwelling circulation. Augustynowicz and Flatau (1981) reported on some preliminary results of a two-dimensional non-hydrostatic model; the results represented the main features of the sea breeze and further work was proposed to study the turbulence flux parameterisations generally employed in the analysis of mesoscale flows.

Work on lake breezes continued apace during this period; for example, Estoque and Gross (1981a, 1981b) examined the three-dimensional structure and behaviour of the Lake Ontario breeze. They found that the major influences were the direction of the synoptic scale flow and the variation of the surrounding terrain. Yamada (1979, 1980) discussed the application of an industrial cooling pond model to the mesoscale atmospheric circulation over Lake Michigan. The presented sensitivity studies showed the relationship between the lake-induced disturbances and the synoptic scale. Kozo (1982a, 1982b), using the two-dimensional scheme after Estoque (1961), examined the sea breezes of the Alaskan Beaufort Sea coast. The model results were compared with a detailed set of experimental data and reasonable agreement was found in selected case studies for the boundary layer wind profiles, the transient change of wind direction at the surface and the variation of inversion height. Dunst and Lagrange (1983) used a three-dimensional hydrostatic model to study the effect of humidity, roughness length and heat capacity of land and sea on mesoscale
coastal disturbances. The roughness studies revealed realistic horizontal wind distributions, however, the wind angle results at the coast were at variance with other research and the authors emphasised a need for a comprehensive observational study. The thermally induced simulations were in accordance with expectations, reproducing satisfactory wind angles and wind speed variations. The humidity results were used in an attempt to simulate advection fog. The results appeared plausible, however the authors did note difficulties in specifying appropriate boundary and initial conditions.

In a recent study of sea breeze energetics, Richiardone and Pearson (1983) compared the availability of potential energy in inland convection and in a sea breeze circulation. One of the more interesting findings was that entrainment at the top of the mixed layer increased the available potential energy at the mesoscale while decreasing it at the convective scale, thus enhancing the sea breeze circulation. Pearson, Carboni and Brusasca (1983) studied in detail the interaction of the sea breeze with the synoptic scale flow. Their numerical analysis showed that the speed of the sea breeze front on a calm day and the speed on a day with overall mean flow onshore differed as a linear function of the speed of the mean flow. They also suggested that the shape and magnitude of the disturbance are unaffected by a constant mean flow. Fett and Tagg (1984) discussed an interesting comparison between a sea breeze induced calm zone over S. Florida as identified by satellite and the results from a two-dimensional planetary boundary layer model. The calm zone was predicted to appear initially at the coastline and, as the overland heating progressed, to move seaward. The speed of this movement and the total distance attained was shown to be highly sensitive to the land-sea temperature gradient and the speed of the synoptic scale component.
A very recent study of the sea breeze front has been presented by Van de Berg and Oerlemans (1985) who applied a high resolution non-hydrostatic model with cloud physics. Their main conclusions were that the formation of clouds tended to strengthen the sea breeze and that even with a very light offshore wind the sea breeze tended to develop into a classic gravity current with a well developed head. The authors did note, however, the relative simplicity of the model and the need for reliable observational verification. Alestalo and Savijarvi (1985) have also recently reported an application of the model of Alpert et al (1982) to the development of mesoscale disturbances normal to or along an 80km wide sea gulf. For an onshore geostrophic wind the model induced rising motion at the coastline due to differential roughness, with the opposite effect for offshore flow. For a longshore geostrophic component with low pressure over the sea, the model predicted three regions; rising motion 25km offshore, sinking at the coast and topographically induced descent at greater distances inland. The final reported application at the time of writing is that of Moussiopoulos (1985) who presented a two-dimensional simulation of the sea breeze disturbance for Athens. The results compared favourably with the available data, however the author did note the need for a more complete three-dimensional simulation in order to resolve the intricacies of the sea breeze interaction with pollutant dispersal in the Athens locale.

1.4 The Aims of the Research Programme

The aims of the presented research are as follows:

* To develop a two-dimensional model of the sea breeze applicable to the Plymouth locale.
To extend the above to a three-dimensional model, allowing for the introduction of coastal complexities and onshore topographic variations.

To correlate the models with data from the on-going observational study and relate the simulations to the experimental work of Hope-Hislop (1974).

To make an assessment of the complex development of the Plymouth sea breeze and its influence on coastal pollutant dispersal and thunderstorm activity over Dartmoor.

To aid the increased research effort in understanding and forecasting mesoscale disturbances.

The thesis will take the following format: Chapter 2 will discuss the development of the models and will present a derivation of the fundamental equations. Chapter 3 discusses the finite difference analogue and solution procedure, detailing the finite difference approximations to the governing equations. Chapter 4 presents the results from a series of sensitivity tests on the two-dimensional model followed by a similar set of tests on the three-dimensional model in Chapter 5. Chapter 6 looks at the application of the three-dimensional model to two sea breeze events in August 1983 and May 1984. Chapter 7 is the concluding chapter, summarising the results and making a range of recommendations. The thesis is completed with the nomenclature and references, followed by a series of appendices incorporating summaries of the equations, an outline of the computational sequence, listings of the model programmes and tables of sea breeze data for the 1982 and 1983 seasons.
CHAPTER 2 - THE MODEL EQUATIONS

2.1 Introduction

The foundation for any model of a mesoscale system such as the sea breeze circulation is a set of principles controlling the conservation of mass, the conservation of motion and the conservation of heat.

The equation for the latter principle determines the amplitude of the forcing function in the sea breeze mechanism, i.e. the diabatic heat flux over the land surface. Furthermore, the equation also controls how the heat overland is advected spatially and hence determines the magnitude of the horizontal temperature gradient across the coastline at any particular point in time.

This temperature gradient directly controls the magnitude of the pressure gradient across the shoreline. The principle for the conservation of motion determines how this feature generates a particular horizontal wind field. In the sea breeze situation, the two flows of primary interest are the onshore component and the offshore return flow aloft.

The principle for conservation of mass states that mass is neither created or destroyed in the earth's atmosphere. Using this concept, the horizontal mass flux can be used at any particular point in time to determine the vertical mass flux. In other words, knowledge of the horizontal wind fields allows determination of the vertical velocity field.

These three components form the basis of both the two- and three-dimensional
sets of equations. In the following sections, the three-dimensional form of the equations will be derived with a two-dimensional summary being given in Appendix A1. The development of the equation sets is based on the original work of Dalu (1978) and Pielke (1974a).

The dependent variables can be decomposed as follows:

\[ \varphi = \bar{\varphi} + \varphi' \]  

(2.1);

where:

- \( \varphi \) = total field (i.e. u, v, w, \( \theta \) or \( \pi \));
- \( \bar{\varphi} \) = domain averaged synoptic component; and
- \( \varphi' \) = mesoscale perturbation component.

The mesoscale perturbations are of primary importance in the context of this project.

2.2 Conservation of Mass

The mass continuity equation for a compressible fluid can be expressed as:

\[ \frac{\partial \rho v}{\partial t} + \nabla \cdot (\rho v \vec{v}) = 0 ; \]

where \( \rho \) is density; \( \vec{V} \) is the velocity vector; and \( D/Dt \) is the substantial derivative defined by:
\[ \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z} \] ; and

\( \nabla \) is the del operator defined by:

\[ (i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y} + k \frac{\partial}{\partial z}) . \]

Assuming incompressibility and application over a depth less than the scale depth of the atmosphere, this can be simplified to:

\[ \nabla \cdot V = 0 . \]

Expressed in component form, the mesoscale perturbation equation can be written as:

\[ \frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0 \]

(2.2);

where \( u' \), \( v' \) and \( w' \) are the velocity components in the west-east, north-south and vertical directions respectively.

2.3 Conservation of Motion

From Newton's second law, the equation of motion in a rotating frame of reference can be expressed as:

\[ \frac{D}{Dt} + f_k V = - \frac{1}{\rho} \nabla P + F ; \]
where $f$ is the Coriolis parameter; $k$ is the unit vector in the vertical; $P$ is atmospheric pressure; and $F$ incorporates the frictional terms.

The perturbation decomposition and the following assumptions can be applied to modify this equation.

1. The ground surface is assumed to be smooth, allowing the complex frictional terms, $F$, to be neglected.

2. The synoptic variables, $\tilde{\phi}$, are assumed to be functions of only synoptic components. A set of equations derived for these variables is subtracted from the total field resulting in equations representing the mesoscale perturbations.

3. The atmosphere is assumed to be barotropic with no initial vertical motion, i.e.:

\[
\begin{align*}
\frac{\partial \tilde{u}}{\partial x} &= \frac{\partial \tilde{v}}{\partial x} = \frac{\partial \tilde{\omega}}{\partial x} = 0 \\
\frac{\partial \tilde{u}}{\partial y} &= \frac{\partial \tilde{v}}{\partial y} = \frac{\partial \tilde{\omega}}{\partial y} = 0 \\
\tilde{\omega} &= 0.
\end{align*}
\]

4. The atmosphere is treated as an ideal gas allowing the horizontal pressure gradient term to be modified as follows:

\[
\frac{1}{\rho} \frac{\partial P}{\partial x} = 0 \quad \text{or} \quad \frac{\partial P}{\partial x} = 0.
\]
where \( \theta \) is the potential temperature and \( \pi \) is the Exner function.

5. The domain-averaged velocities are assumed to be geostrophic, therefore:

\[
\begin{align*}
\mathbf{u} &= \mathbf{U}_g + u' \\
\mathbf{v} &= \mathbf{V}_g + v'
\end{align*}
\]

\[
\begin{align*}
\theta' \frac{\partial \pi}{\partial y} &= f \mathbf{V}_g \frac{\partial \theta'}{\partial \theta} \\
\theta' \frac{\partial \pi}{\partial x} &= -f \mathbf{U}_g \frac{\partial \theta'}{\partial \theta} ;
\end{align*}
\]

where \( \mathbf{U}_g \) and \( \mathbf{V}_g \) are the geostrophic components in the east-west and north-south directions respectively. In the present model constructs, a constant value is assigned to these components throughout the model depth, due to the inherent assumption of a smooth ground surface. If the friction terms had been incorporated then it would have been appropriate to include a logarithmic wind profile for the terms through the depth of the planetary boundary layer.

The resulting longitudinal and latitudinal equations of motion are:

\[
\begin{align*}
\frac{\partial u'}{\partial t} &= - (\mathbf{U}_g + u') \frac{\partial u'}{\partial x} - (\mathbf{V}_g + v') \frac{\partial u'}{\partial y} - w' \frac{\partial}{\partial z} (\mathbf{U}_g + u') \\
&- f \frac{\partial}{\partial \theta} \mathbf{V}_g - \frac{\partial}{\partial x} \frac{\partial \pi'}{\partial x} + f v' - \mathbf{F}_w' 
\end{align*}
\]

(2.3)
The vertical equation of motion is reduced by assuming the atmosphere to be hydrostatic, i.e.:

\[ \frac{\partial P}{\partial z} = - \rho g \]

where \( g \) is the acceleration due to gravity. The hydrostatic assumption has been examined at some length by several mesoscale modellers (Patrinos and Kistler (1977), Physick (1976), Kozo (1982b) and Martin and Pielke (1983)). The criterion for validity is that the ratio of the horizontal to vertical grid lengths must be greater than or equal to three. This condition is never exceeded in either the two- or three-dimensional simulations.

In accordance with the pressure gradient terms in Equations (2.3) and (2.4), the following form involving \( \pi \) is used in the simulations, i.e.:

\[ \frac{\partial \pi'}{\partial z} = g \frac{\partial'}{\partial^2} \]  

(2.5)

There is a major advantage to employing this equation in that there is no need to calculate a density perturbation as the Boussinesq approximation has been applied in the derivation. The validity of this approximation has been discussed by Spiegel and Veronis (1959) and more recently by Wipperman (1981) and Pielke (1984). Essentially, it allows the equations for a compressible fluid to be simplified by neglecting density variations except when coupled with
gravity to produce buoyancy forces. This is only valid when the vertical dimension of the system under consideration is less than the scale depth of the atmosphere, a condition which, as stated earlier, is never exceeded.

2.4 Conservation of Heat

The first law of thermodynamics for an ideal gas can be written as:

\[ \frac{\partial \theta}{\partial t} = - \nabla \cdot v \theta + \delta \theta \]

where \( \delta \theta \) is the diabatically perturbed potential temperature. In component form and after the decomposition given in Equation (2.1), this can be written as:

\[ \frac{\partial \theta'}{\partial t} = - (U_g + u') \frac{\partial \theta'}{\partial x} - (V_g + v') \frac{\partial \theta'}{\partial y} \]

\[ - w' \frac{\partial}{\partial z} (\theta' + \bar{\theta}) + \delta \theta \]

(2.6).

2.5 Parameterisation of Diabatic Heat Flux and Diffusion

The term \( \delta \theta \) in Equation (2.6), strictly, represents all the sources and sinks of heat as incorporated in the changes of potential temperature. This would involve the latent heat changes associated with freezing and melting, condensation and evaporation, and deposition and sublimation. Further changes would also be caused by chemical reactions in the atmosphere, by radiative flux convergence or divergence and by molecular motion in the process of dissipating kinetic energy. Obviously, the exact solution of this term would be extremely complex.
and time consuming. In the models presented here, $\delta \theta$ is calculated using one of two parameterisations.

The first scheme to be used was developed by Dalu (1978) and is based on the assumption that the surface heat flux, $Q^*$, in the model is simply a fraction of the solar energy reaching the top of the atmosphere, i.e.:

$$Q^* = CC \ S \ cos \lambda \ sin \Omega t$$  \hspace{1cm} (2.7);

where $S$ is the solar constant, $\lambda$ is latitude, $\Omega$ is the earth's angular velocity and $t$ is time. The constant, $CC$, is chosen to match the observed windspeed values and, in the current simulations, will normally have a value of 0.1. The sinusoidal form of the heat flux produced by this equation is illustrated in Figure 3.

The second and somewhat more realistic scheme employs the surface energy balance model of Wood (1977a). With this method, $Q^*$ is assumed to equal the sensible heat flux at the surface, $Q_H$, given by the flux-gradient relationship:

$$Q^* = Q_H = - \rho \ C_p \ K \ \frac{\delta \theta}{\delta z}$$  \hspace{1cm} (2.8);

where $C_p$ is the specific heat of dry air at constant pressure. The model assumes the similarity hypothesis for conditions of neutral stability, such that a bulk turbulent transfer coefficient can be used for the coefficient of eddy transfer of heat, $K$, i.e.:

$$K = \frac{\rho \ k^2 \ U_2}{[\ln(Z_2/Z_0)]^2};$$
where \( k \) is Von Karman's constant, \( Z_0 \) is the roughness length, \( Z_2 \) is the logarithmic profile depth and \( U_2 \) is the windspeed at this height. It is not, however, always feasible to assume neutral stability in the surface layers of the atmosphere and it is, therefore, necessary to correct for this. A useful measure of stability is the Richardson number, given by:

\[
R_i = \frac{g}{\theta} \left[ \frac{\partial \theta}{\partial z} / \ln \left( \frac{Z_2}{Z_0} \right) \right] \left[ \frac{U_2}{\ln \left( \frac{Z_2}{Z_0} \right)} \right]^{-2}.
\]

Using this, a dimensionless stability function, \( \psi \), can be calculated from:

\[
\psi = (1 - a R_i)^n.
\]

There is a certain amount of debate in the literature as to the appropriate values for \( a \) and \( n \). Wood's model uses \( a=32 \) and \( n=0.5 \) for both stable and unstable conditions, however, Oke (1978) suggested \( a=5, n=2 \) for the stable case and \( a=16, n=0.75 \) for the unstable case. The former values roughly represent a real situation in that surface temperatures are reduced during unstable, daytime conditions and raised during the more stable, night situation.

Using either of these methods, a suitable parameterisation for the diffusion of the heat input must also be applied due to the simplified form of the governing equations. The scheme outlined by Dalu (1978) results in a realistic mixed convective layer, being totally adiabatic throughout its depth. The method is developed as follows: initially, a perturbation potential temperature as calculated using Equation (2.9) is added to the lowest overland grid points. However, if this results in a super-adiabatic layer then Equation (2.9) is recalculated and the resulting value added to the first two levels. The criterion of a non-super-adiabatic layer is continued vertically upwards until the diabatically
perturbed potential temperature field at a level \( z-1 \) is less than the value at a level \( z+1 \).

Expressed algebraically, a perturbation potential temperature is calculated using the formula:

\[
\delta \theta = \Delta t \sum_{i=1}^{m} \frac{Q}{C_p} \rho \Delta z
\]  

Starting at the first level above the ground (i.e. \( z=m\Delta z \), where \( m=1 \)), \( m \) is increased until the following condition is satisfied:

\[
\theta(x,y,z - \Delta z) + \delta \theta \leq \theta(x,y,z + \Delta z)
\]

The result of this procedure is a convective layer of depth equal to \( m\Delta z \). The temperature values in the layer can then be corrected as follows:

\[
\theta(x,y,z) = \theta(x,y,z) + \delta \theta \text{ for } 0 < z < m\Delta z
\]

2.6 Solution of the Diagnostic Equations

One of the problems associated with mesoscale modelling is the appropriate solution of the diagnostic equations, (2.2) and (2.5). Because of their nature it is necessary to specify a boundary condition at each timestep; for the continuity equation, (2.2), this will be that the vertical velocity at \( z=0 \) will always equal zero. This obviously allows integration from the ground upwards. Problems, however, ensue at the top grid point level unless the integration of the
hydrostatic equation, (2.5), is suitably coupled to maintain mass continuity. To prevent the resulting encroachment of gravity waves, early two-dimensional models used the so-called rigid lid approximation (Potter (1973)). This method, first employed by Estoque (1960), uses the differentiated form of the continuity equation, i.e.:

\[
\frac{\partial^2 w}{\partial z^2} = - \frac{\partial}{\partial z} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)
\] (2.10).

However, as noted by Neumann and Mahrer (1971) and Peterson (1971), this form of the equation does not satisfy the law of conservation of mass and the solution, although stable, can only be regarded as spurious. The reason for this can be explained as follows. If Equation (2.10) is integrated the result is:

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

Only if \(f(x,y)\) is identically equal to zero can the derivative form of the equation be valid. Since the models based on the work of Estoque do not consider this requirement, the validity of the resultant wind fields must be questioned. Estoque and Bhumralker (1970) attempted to improve on this by introducing a variable pressure at the model top. However, as Pielke (1974) pointed out, this is inconsistent in a hydrostatic model.

The technique employed in the present model simulations was first developed by Pielke (1974) and is applied as follows. The continuity equation is integrated upwards using a lower boundary condition, \(w'=0\), at the first grid point level. The hydrostatic equation is also integrated upwards after the application of a surface pressure tendency equation, i.e.:
This equation implies that the addition of heat overland in a column is equivalent to the removal of mass from that column. The parameter, \( H \), represents the height of a so-called material surface which allows for the vertical spreading of mass by hydrostatic adjustment. The prognostic tendency equation for \( H \) is given by:

\[
\frac{\partial H}{\partial t} = - (Ug + u') \frac{\partial H}{\partial x} - (Vg + v') \frac{\partial H}{\partial y} + w_H'.
\]

Several other sea breeze modellers (Dalu (1978), Mahrer & Pielke (1977a), Kozo (1982b)) have found it more convenient to adopt a system where the pressure is calculated by integration from the top downwards. The scheme developed by Dalu (1978) was adopted in the present model construct, however, the maintenance of computational stability throughout the simulations proved impossible.

In the following chapter a finite difference analogue will be described for the three-dimensional primary equations (2.3) to (2.6), (2.11) and (2.12).
3.1 **Description of Model Domains and Grid Meshes**

Figure 4 illustrates the model domains for both the two- and three-dimensional simulations. The former consists of a rectangle of horizontal dimension 128Km and depth 3Km and the three-dimensional system is a box of dimensions 64Km x 64Km x 3Km. The change in the horizontal lengths between the two models is to allow a clearer view of the situation when a rectangular bay is incorporated into the three-dimensional simulations. In both cases, the dimensions are considered to be sufficiently large for the system under study to be free of any undue influence from the lateral and upper boundaries. In an extensive study of the sea breezes over the Lasham-Reading area of Southern England, Simpson, Mansfield and Milford (1977) concluded that deep penetration by the sea breeze is not a frequent occurrence but may, on occasion, extend up to 100Km inland. For the Plymouth region, Hope-Hislop (1974) confirmed that the sea breeze generally travels no further than 15Km inland which is well within the dimension range chosen.

The parameter position on a finite-difference grid is vitally important with regard to the ease with which the governing equations can be solved; this is especially true at the lateral boundaries. The grid mesh set up for both the two- and three-dimensional models is given in Figure 5. For the former system, the mesh will therefore consist of 17x11 points for the horizontal velocity, pressure and potential temperature and 16x11 points for the vertical velocity. The three-dimensional grid mesh consists of 17x17x11 and 16x16x11, respectively, for the aforementioned parameters. The formulation of a staggered
mesh in this manner also results in a very convenient solution for the velocity divergence as defined at each cell centre.

The prognostic equations are integrated forward in time, with centred differencing employed in the spatial terms. The advection terms in Equations (2.3), (2.4), (2.6) and (2.12) are solved using an upstream differencing technique outlined as follows:

For any variable, ϕ,

\[
\begin{align*}
\frac{\partial \varphi}{\partial x} &= u_m \frac{\varphi_x - \varphi_{x-m} \Delta x}{\Delta x} \\
\frac{\partial \varphi}{\partial y} &= v_m \frac{\varphi_y - \varphi_{y-m} \Delta y}{\Delta y} \\
\frac{\partial \varphi}{\partial z} &= w_m \frac{\varphi_z - \varphi_{z-m} \Delta z}{\Delta z}
\end{align*}
\]

where, \( m \) is assigned the value 1 when \( u, v \) or \( w \) is positive, and -1 when \( u, v \) or \( w \) is negative. The applicability of this scheme in mesoscale modelling has been questioned in the past by Molenkamp (1968) who criticised the computational damping which the scheme incorporates into the model. He suggested that the numerical diffusion is almost as large as turbulent diffusion and the latter cannot therefore be modelled accurately using this scheme. As Pielke (1974a) points out, computational damping is desirable in controlling non-linear computational instabilities and upstream differencing has proved especially useful in describing the sea breeze (Nordlund (1971), Lambert (1974), Pielke (1974a), Kozo (1982b)). It is likely that Molenkamp's criticism is valid when considering more transient features such as clouds, where the turbulence terms are of vital importance, but less so in a system where the circulations
develop rather slowly with time. In recent years, Pielke's model has been modified to use cubic spline interpolation techniques in the advective terms (Mahrer & Pielke (1978)) and, although accurate, Purnell (1976) pointed out that despite the increase in complexity, the difference between the results using the two methods would seem to be slight in most situations.

3.2 Finite-Difference Form of the Governing Equations

If \( \phi \) is any variable and \( \Delta t, \Delta x, \Delta y \) and \( \Delta z \) the appropriate increments such that \( t=n\Delta t, x=i\Delta x, y=j\Delta y \) and \( z=k\Delta z \), then the following shorthand notation can be employed:

\[
\phi(n\Delta t, i\Delta x, j\Delta y, k\Delta z) = \phi^n_{ijk}
\]

For the following description all velocities in the advective terms are considered positive, with \( u'=u, v'=v, w'=w, u'g=\tilde{u}, v'+Vg=\tilde{v}, \theta'+\theta=\tilde{\theta} \) and \( \theta'=\theta \). The grid numbers and positions are as illustrated in Figure 6.

The longitudinal equation of motion (2.3) is approximated by:

\[
\begin{align*}
\frac{u^n_{i+1,j,k} - u^n_{i,j,k}}{\Delta t} &+ \frac{u^n_{i,j+1,k} - u^n_{i,j-1,k}}{\Delta y} + \frac{u^n_{i,j,k+1} - u^n_{i,j,k-1}}{\Delta z} \\
&+ f \left( \frac{\partial^n_{i,j,k}}{\theta_{i,j,k}} \right) Vg + \tilde{\theta}_{i,j,k} \left( \frac{u^n_{i+2,j,k} - 2u^n_{i,j,k} + u^n_{i-2,j,k}}{2\Delta x} \right) - f v^n_{i,j,k} + \tilde{w}^n_{i,j,k}
\end{align*}
\]

(3.1)
The latitudinal equation of motion (2.4) is approximated by:

\[
\begin{align*}
  v_{ij}^{n+1} &= v_{ij}^n - \Delta t \left[ u_{ijk}^n \left( \frac{v_{ijk}^n - v_{i-2,j,k}^n}{\Delta x} \right) 
  
  + \tilde{v}_{ijk}^n \left( \frac{v_{ijk}^n - v_{ij,i-2,k}^n}{\Delta y} \right) + w_{ijk}^n \left( \frac{v_{ijk}^n - v_{ij,i,k-2}^n}{\Delta z} \right) 
  
  - f \left( \frac{\theta_{ijk}^n}{\alpha_{ijk}} \right) u_g + \theta_{ijk}^n \left( \frac{\chi_{ij,i+2,k}^n - \chi_{ij,i-2,k}^n}{2\Delta y} \right) + f w_{ijk}^{n+1} \right] 
  
  + V. (\cdot) + W. (\cdot) 
\end{align*}
\]  

(3.2).

The new vertical velocity field is calculated using Equation (2.5) as approximated by:

\[
\begin{align*}
  w_{i+1,j+1+k+2}^{n+1} &= w_{i+1,j+1+k+1}^{n+1} - \Delta z \left[ u_{i+2,j+1+k+1}^n - u_{i+1,j+1+k+1}^n \right] 
  
  + \tilde{v}_{i+1,j+2+k+1}^{n+1} - v_{i+1,j+1+k+1}^{n+1} 
\end{align*}
\]  

(3.3).

The material surface at the new time step is calculated by the approximation to Equation (2.12).

\[
\begin{align*}
  H_{ij}^{n+1} &= H_{ij}^n - \Delta t \left[ u_{ij}^n \left( \frac{H_{ij,i}^n - H_{ij,i-2}^n}{\Delta x} \right) 
  
  - \tilde{v}_{ij}^n \left( \frac{H_{ij,j} - H_{ij,j-2}}{\Delta y} \right) + w_{ij}^{n+1} \right] 
\end{align*}
\]  

(3.4).

The new potential temperature field is calculated using Equation (2.6) as approximated by:
The surface pressure tendency Equation (2.11) is approximated by:

\[ \phi_{ijk}^{n+1} = \phi_{ijk}^n + \Delta t \left[ \nabla_{ijk}^{n} \left( \frac{\phi_{ijk}^{n} - \phi_{ijk-2}^{n}}{\Delta x} \right) + \nabla_{ijk}^{n} \left( \frac{\phi_{ijk}^{n} - \phi_{ijk-2}^{n}}{\Delta y} \right) \\
\right. \\
\left. + w_{ijk}^{n} \left( \frac{\phi_{ijk-1}^{n} - \phi_{ijk-2}^{n}}{\Delta z} \right) + \delta \phi_{ijk}^{n} \right] \quad (3.5). \]

The pressure distribution can then be obtained using the hydrostatic Equation (2.5) as approximated by:

\[ \pi_{ij}^{n+1} = \pi_{ij}^n - g \sum_{k}^{H} \text{SUM} \quad (3.6); \]

where \( \text{SUM} = \frac{\phi_{ijk+1}^{n} - \phi_{ijk}^{n} - (\bar{\phi}_{ijk})^2}{\Delta z} \) for \( k=0,2,\ldots,H-2 \).

The two-dimensional finite-difference approximations are readily obtained from Equations (3.1) to (3.7) and are given in Appendix A2.

### 3.3 Additional Numerical Techniques

In any attempt to solve a set of non-linear differential equations by finite-difference techniques, one of the most serious problems is the spurious growth of shortwaves which interfere with the solution. Several workers, including Haltiner (1971), Pepper, Kern and Long (1977) and Pielke (1984), have
discussed this aspect at considerable length. Its occurrence is a result of the models inability to adequately simulate certain energy cascades. In any mesoscale system such as the sea breeze, kinetic energy is continually being produced and then subsequently dissipated by molecular interactions into heat. The scale of the former motions can range from 100m up to 100Km, whereas the latter will have a range of less than 1cm. It is, therefore, obvious that modelling of this energy cascade is virtually impossible.

To illustrate the problem mathematically, consider the following situation:

\[ \varphi_1 = \varphi_0 \cos k_1 \Delta x \quad \text{and} \quad \varphi_2 = \varphi_0 \cos k_2 \Delta x \]

represent two waves with the same amplitude, \( \varphi_0 \).

A non-linear interaction between these waves can be expressed as:

\[ \varphi_1 \varphi_2 = \varphi_0^2 \cos k_1 \Delta x \cos k_2 \Delta x \]

\[ = \frac{1}{2} \varphi_0^2 \left[ \cos(k_1+k_2) \Delta x + \cos(k_1-k_2) \Delta x \right] . \]

This implies that the interaction of the two waves of number \( k_1 \) and \( k_2 \) results in two new waves of number \( (k_1+k_2) \) and \( (k_1-k_2) \). Assuming the interaction takes place between two waves of length \( 2\Delta x \) and \( 4\Delta x \) respectively, the resulting interaction can be expressed as:

\[ \varphi_1 \varphi_2 = \frac{1}{2} \varphi_0^2 \left[ \cos 2\pi \left( \frac{6}{8} \right) \Delta x + \cos 2\pi \left( \frac{1}{4} \right) \Delta x \right] . \]

The new waves are therefore of length 1.33\( \Delta x \) and 4\( \Delta x \). Since the resolvable
limit to waves in a finite-difference model is $2\Delta x$, the former cannot be modelled. Short waves which grow in this manner are described as being aliased or folded. Figure 7 illustrates how the $1.33\Delta x$ wave will be seen as a $4\Delta x$ wave on a finite-difference grid.

Two methods have been employed in the past to try to remove these erroneous waves. Firstly, by applying an appropriate parameterisation for the sub-grid scale correlation terms, i.e.:

$$K_H \frac{\partial^2 u}{\partial x^2}, \quad K_H \frac{\partial^2 u}{\partial y^2}$$ etc.

Pielke (1974a) used the following formulation for the eddy coefficient:

$$K_H = \alpha (\Delta x)^2 \left[ \frac{1}{2} \left( \frac{\partial v'}{\partial x} + \frac{\partial u'}{\partial y} \right)^2 + \left( \frac{\partial u'}{\partial x} \right)^2 + \left( \frac{\partial v'}{\partial y} \right)^2 \right]^{1/2},$$

where the coefficient $\alpha$ can be adjusted until the $2\Delta x$ wavelengths do not appear to be influencing the solution to any significant degree.

The second method is to apply a spatial filter which will remove the smaller wavelengths and leave the larger ones relatively unaffected. Several types of filter have been employed in mesoscale models; Sheih and Moroz (1975), for example, used an explicit, 5-point smoother of the form:

$$\varphi_{i,j}^* = \frac{1}{4} \varphi_{i,j} + \frac{1}{8} \left( \varphi_{i+1,j} + \varphi_{i-1,j} + \varphi_{i,j+1} + \varphi_{i,j-1} \right);$$

where $\varphi$ and $\varphi^*$ are the dependent variables before and after being smoothed.

More recently, low-pass, highly selective filters of the following form have been
used (Mahrer & Pielke (1978), Pepper, Kern and Long (1979), Alpert (1981),
Alpert, Cohen, Neumann & Doron (1982)):

\[(1-\delta) \varphi_{i+1}^* + 2\varphi_i^* + (1-\delta)\varphi_{i-1}^* = \varphi_{i+1} + 2\varphi_{i} + \varphi_{i-1} \quad (3.8)\]

The filter completely eliminates the $2\Delta x$ waves, while its effect on larger
wavelengths is controlled by the value assigned to $\delta$ (Pielke (1984)).
Experiments with both types of filter led to the incorporation of Equation (3.8)
with $\delta=0.01$ which is applied to the $u$, $v$ and $\theta$ fields at each time step in the
$x$- and $y$-directions.

Computational stability analysis of the finite-difference sets of equations is
important with regard to determination of the appropriate value of the timestep.
Traditionally, the restrictions placed on the stability of a particular model have
been a function of the grid steps and of the values assigned to the diffusion
coefficients. The latter takes the form:

$\Delta t \leq \frac{\Delta z^2}{4k}$.

This diffusive criterion is always the most stringent, for example, Pielke (1974a)
had to use timesteps varying from 20s to 60s to maintain stability, while Kozo
(1982b) used values of less than 120s. The former restriction concerning the
gridsteps is the so-called Courant-Friedrichs-Lewy (CFL) condition and takes the
form:

$\Delta t \leq \min \left( \frac{\Delta x}{u_{\text{max}}}, \frac{\Delta y}{v_{\text{max}}}, \frac{\Delta z}{w_{\text{max}}} \right)$. 

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In the models under discussion this leads to a criterion:

\[ \Delta t \leq \min (1600s, 1600s, 3000s) . \]

Normally, \( \Delta t \) will be in the range 120s to 600s. The use of equations which do not include diffusion coefficients has, therefore, resulted in a model set up which is far more efficient with regard to CPU time.

3.4 Computational Sequence

The integration sequence for the governing equations is illustrated in the flow chart given in Appendix B. The first few stages involve the set up of physical and grid constants and the initial state. After this the equations are integrated in a set of subroutines contained within a temporal loop, the first step being the calculation of the diabatic heat flux using either Equation (2.7) or (2.8). The computational order was developed due to considerations of stability. Rosenthal (1970) pointed out that the momentum equations should be evaluated before the thermodynamic equation, i.e. \( \theta' \) is computed using the updated values of \( u', v' \) and \( w' \). Tests on the present developments have verified this conclusion.

The programmes BREEZE.2D.F77 and BREEZE.3D.F77 are reproduced in Appendix B and represent the standard two- and three-dimensional formulations as outlined in this chapter.
3.5 **Boundary and Initial Conditions**

In the models presented here, the initial conditions are similar to those chosen in other mesoscale models (Pieike (1974a), Physick (1976), Dalu (1978), Kozo (1982b)). The model is integrated from a state of rest with no horizontal accelerations and no horizontal variations in potential temperature and pressure. The initial potential temperature at the surface and the vertical profile are taken from local analysis of tephigrams and suitable values chosen. Normally, the surface potential temperature will be set at 286K with a range of profiles from 1.0 to 4.0K km⁻¹. The initial synoptic scale wind field is set up by assigning values to Ug and Vg as determined from synoptic charts. The use of the material surface concept at the top grid point level means an initial value must be assigned to H.

In summary, at the initial time, t=0, the following conditions are set up:

\[
\begin{align*}
    u' &= v' = w' = \theta' = \pi' = 0 \\
    \hat{\theta} &= 286K \text{ for } Z=0 \\
    \hat{\theta} &= (286-BZ) K \text{ for } Z>0 \\
    H &= 3000m.
\end{align*}
\]

Pielke (1984) summarised four types of lateral boundary conditions, i.e.:

(i) constant inflow, gradient outflow conditions;
(ii) radiative boundary conditions;
(iii) sponge boundary conditions; and
(iv) periodic boundary conditions.

Periodic boundary conditions can be set up by assuming that variables at one lateral boundary are identically equal to those at the other boundary. Sponge boundary conditions are set up by using enhanced filtering to dampen any advection or wave disturbances as they approach the edges of the model domain. The radiative boundary conditions are set up by allowing the values at the lateral edges to change in a manner which minimises any reflection of outward developing systems back into the model domain. This latter scheme has been successfully adopted by Physick (1976). The most commonly used scheme and the one employed in the models under discussion is the constant inflow, gradient outflow condition which is set up by assuming that variables at the lateral boundaries at all timesteps are equal to those found one grid point upstream. If the lateral boundaries are kept sufficiently far away from the region of interest this will be a satisfactory boundary condition procedure.

In summary, the lateral boundary conditions at all timesteps are:

\[ \frac{\partial}{\partial x} (\theta', \pi', H) = \frac{\partial}{\partial y} (\theta', \pi', H) = 0 ; \]
\[ \frac{\partial}{\partial x} (u', v') = \frac{\partial}{\partial y} (u', v') = 0 . \]

At the lower boundary, the specified conditions are of a characteristic type, i.e.:

\[ u' = v' = w' = 0 . \]
On the material surface, \( z = H \), the boundary conditions are after Pielke (1974a), i.e.:

\[
\begin{align*}
  u &= U_g; \\
  v &= V_g; \\
  \tau' &= 0.
\end{align*}
\]

The sensitivity of the two-dimensional finite difference scheme, equivalent to equations (3.1) to (3.7), is tested in the following chapter with variations in the initial stratification, the direction of the synoptic scale flow and the amplitude of the forcing function.
4.1 Introduction

The model programmes are coded using ANSI x.39 (1978) standard FORTRAN 77 and processed on a system developed by the University of Salford for the PRIME 50 series. The Plymouth Polytechnic Computer Centre operates a network of PRIME super-mini computers of which the PRIME 850 was used in the simulations presented in this thesis.

The resulting model data is displayed using the FORTRAN subroutine libraries NAG (NAG (1981)) and GINO-F (CAD Centre (1985)). Both systems consist of a general purpose set of drawing and administration subroutines which allow the development of sophisticated, user-controlled graphics programs. The output can be displayed on graphics terminals; either a TEKTRONIX 4010 or the high resolution, colour AED767. Hardcopy is readily available using a CALCOMP 1039 drum plotter or a CALCOMP 81 flatbed. Sample programmes for vector plotting, contour charts and general graph drawing are reproduced in Appendix C.

Before any discussion of sensitivity can take place, the major features of the system under scrutiny must be identified:

(1) Time of onset.
(2) Depth of the sea breeze current.
(3) Maximum velocity in the sea breeze current.
(4) Depth of the return current.
Maximum velocity in the return current.

Horizontal extent inland.

Maximum velocity in overland updraught.

Horizontal extent seaward.

Maximum velocity in offshore downdraught.

These features will be examined in the following series of tests:

(a) **Test 1**: Variation of initial stratification.
   i.e.: $\frac{\partial\theta}{\partial z} = 1.0, 2.0, 3.0, 4.0 \text{ K km}^{-1}$

(b) **Test 2**: Variation in the value of the heat flux constant.
   i.e.: $CC = 0.025, 0.05, 0.1$

(c) **Test 3**: Variation in the geostrophic component.
   i.e.: $V_g = 0, 2.0$ (onshore), $-2.0$ (offshore) ms$^{-1}$

(d) **Test 4**: Variation in sensible heat flux values in surface climate model of Wood (1977a).
   i.e.: $Q^* = Q_H =$ values typical for March and June

The test criteria for the nine sea breeze features are as follows:

(1) Time of onset: time when the horizontal windspeed, $v'$, at the surface first exceeds 1.0ms$^{-1}$.

(2) Sea breeze depth: mean height of the level of neutrality, where $v'$ equals 0.0ms$^{-1}$.
(3) Maximum velocity in sea breeze current: maximum onshore velocity attained.

(4) Return current depth: difference between (2) and the upper level where \( v' \) exceeds \(-0.5\text{ms}^{-1}\).

(5) Maximum velocity in return current: maximum offshore velocity attained.

(6) Horizontal extent inland: maximum distance where the maritime air advection suppresses the temperature rise in the lowest model layer by more than \(0.1^\circ\text{C}\) (Warner, Anthes and Seaman (1979)).

(7) Maximum velocity in overland updraught: maximum value attained by \( w' \) overland.

(8) Horizontal extent seaward: maximum offshore distance where \( v' \) exceeds \(1.0\text{ms}^{-1}\).

(9) Maximum velocity in offshore downdraught: maximum value attained by \( |w'| \) offshore.

4.2 Sensitivity Test for the Initial Stratification

Table 4.1 shows the results from the first tests involving the stability of the atmosphere. The values ranging from 1.0 to 4.0K \(\text{km}^{-1}\) for the potential stratification imply a gradual increase in stability. Several features are immediately apparent; firstly, as the stability increases the depth of the sea
breeze circulation decreases. This development is well-documented in observational studies of the sea breeze system; Pearce (1968) noted the dependence of the strength of the sea breeze on atmospheric stability, with greater circulation depths for less stable atmospheres and vice-versa. Although this conclusion is generally agreed upon, there seems to be a certain amount of debate as to the degree of instability which most favours or inhibits sea breeze development. It is likely that local effects of topography and other impinging synoptic conditions combine to require differing stability criteria for different regions.

The second feature to note regards the variation in the times of onset, where an increase in stability results in a later onset. Wexler (1946) suggested that stability was the main control on the time of onset of the sea breeze and Watts (1955) related the stability to the manner in which it arrived at Thorney Island in Southern England. He concluded that under stable conditions the sea breeze tends to arrive gradually, whereas in unstable conditions, the tendency is for a sudden, sharp change in windspeed and direction.

A further development is the variation in the windspeed values. The horizontal components in both the sea breeze and return currents are reduced as the stability increases and similarly with the vertical velocities. Wallington (1977) related atmospheric stability to the strength and vigour of the sea breeze circulation. He noted that where inland convection was limited to less than 500m, the sea breeze component will be weak with little frontal development. For greater depths up to 2Km, he noted increased levels of windspeed and more defined sea breeze fronts.
<table>
<thead>
<tr>
<th>STRATIFICATION (K km⁻¹)</th>
<th>1.0</th>
<th>2.0</th>
<th>3.0</th>
<th>4.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEPTH OF SEA BREEZE (m)</td>
<td>1350</td>
<td>1050</td>
<td>750</td>
<td>750</td>
</tr>
<tr>
<td>DEPTH OF RETURN CURRENT (m)</td>
<td>1500</td>
<td>1200</td>
<td>900</td>
<td>750</td>
</tr>
<tr>
<td>DEPTH OF CIRCULATION (m)</td>
<td>2850</td>
<td>2250</td>
<td>1650</td>
<td>1500</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN SEA BREEZE (ms⁻¹)</td>
<td>4.09</td>
<td>3.74</td>
<td>3.45</td>
<td>3.25</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN RETURN CURRENT (ms⁻¹)</td>
<td>1.44</td>
<td>1.39</td>
<td>1.31</td>
<td>1.26</td>
</tr>
<tr>
<td>HORIZONTAL EXTENT INLAND (Km)</td>
<td>&gt;56</td>
<td>&gt;56</td>
<td>&gt;56</td>
<td>&gt;56</td>
</tr>
<tr>
<td>TIME OF ONSET (Hrs, simulated time)</td>
<td>02:50</td>
<td>03:00</td>
<td>03:20</td>
<td>03:30</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN UPDRAUGHT (cm s⁻¹)</td>
<td>5.65</td>
<td>3.73</td>
<td>2.77</td>
<td>2.41</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN DOWNDRAUGHT (cm s⁻¹)</td>
<td>4.70</td>
<td>3.05</td>
<td>2.46</td>
<td>2.09</td>
</tr>
</tbody>
</table>

**TABLE 4.1:** SENSITIVITY TEST FOR THE INITIAL POTENTIAL TEMPERATURE STRATIFICATION
To illustrate these developments, Figures 8 to 11 show the velocity fields for the initial stratification 1.0K km\(^{-1}\) (A) and 3.0K km\(^{-1}\) (B). In the horizontal wind field diagrams, the positive isotachs represent regions where the flow is onshore and the negative isotachs indicate regions where the flow is offshore. For the vertical velocity fields, the positive isotachs delineate regions of updraught and negative isotachs regions of downdraught. As well as clearly illustrating the aforementioned features, the diagrams also display some general facets of the sea breeze circulation. Immediately apparent is the characteristic development of the system with time. From 03:00hrs to 09:00hrs there is a gradual growth followed by dissipation to 12:00hrs as the heat flux from the land surface lessens. The onshore components are well simulated, with a depth of 1350m and a maximum velocity of 4.09ms\(^{-1}\) for (A) and 750m and 3.45ms\(^{-1}\) for (B). The return flows, as expected, are of a greater depth and are more diffuse, with values of 1.44ms\(^{-1}\) for (A) and 1.31ms\(^{-1}\) for (B).

The vertical velocity fields show the characteristic overland updraughts and offshore downdraughts, with maximum speeds in the former of 5.65cm s\(^{-1}\) for (A) and 2.77cm s\(^{-1}\) for (B). The magnitude of these values is somewhat less than those observed; for example, Wallington (1959) reported updraught speeds of more than 2ms\(^{-1}\) in a narrow core of 100-250m in width. Obviously resolution of a feature of this scale is impossible on such a coarse grid and use of the updraught core as an indicator of sea breeze frontal propagation is somewhat unreliable. With regard to model results, values are generally in the range of 5cm s\(^{-1}\) (Patrinos and Kistler (1977) and Sheih and Moroz (1975)) to 15cm s\(^{-1}\) (Neumann and Mahrer (1971) and Pearson (1973)). Lambert (1974) in a high resolution study of the sea breeze came closest to simulating strong core velocities, with values of more than 50cm s\(^{-1}\).
Because of this problem with the updraught core, the movement inland of the sea breeze was determined using the criterion proposed by Warner, Anthes and Seaman (1979) as defined in Section 4.1. Figure 12 illustrates frontal propagation rates for a range of stratifications from 1.0 to 4.0 K km$^{-1}$. The relationship in all four cases is approximately linear. Although the difference between the curves is minimal there is a tendency for slightly quicker propagation rates for lower stabilities. Pearson (1973) examined the role of stability in determining frontal propagation and concluded that the frontal speed was independent of it, which is in fair agreement with the results presented. The average propagation rates range from 12.44 K m hr$^{-1}$ for 1.0 K km$^{-1}$ and 11.2 K m hr$^{-1}$ for 4.0 K km$^{-1}$. Comparison between modelled results and observational data is somewhat misleading due to the large range of values to be found; for example, Clarke (1955) gave an observed range of 14.4 to 30.6 K m hr$^{-1}$ for South Australia compared to 7.2 to 12.6 K m hr$^{-1}$ for Poland given by Koschmeider (1936) and 7.2 to 10.8 K m hr$^{-1}$ for Southern England as quoted by Simpson (1964). As Pearson (1973) noted, the variation is likely due to differences in the heat flux input to the atmosphere in different regions of the earth. Modelled values also display a distinct range; for example, Neumann and Mahrer (1971) found a value of 8 K m hr$^{-1}$ for their Israeli simulations and Physick (1976) found values ranging from 10.7 K m hr$^{-1}$ to 17.6 K m hr$^{-1}$ for Australia.

As a further illustration of the models ability to simulate a typical sea breeze circulation, Figure 13, gives the horizontal wind profiles for 03:00 to 12:00 hrs for the situation with an initial stratification of 3.0 K m hr$^{-1}$. The development of the strong onshore flow near the surface and the broad, relatively weaker offshore flow aloft are well simulated. Figure 14 shows potential temperature profiles 24 K m inland for the same situation. This clearly illustrates the
development of the mixed convective layer with time; at 03:00hrs the depth is about 600m, increasing to about 1200m at 09:00hrs. The profiles also show a sharp increase in the stability of the land air at low levels due to the efflux of the cooler maritime air as the sea breeze grows in intensity.

4.3 Sensitivity Test for the Heat Flux Constant

Using the parameterisation given in Equation (2.7) a value for the constant CC must be chosen. Table 4.2 outlines the variations resulting from a range of values from 0.025 to 0.1. The first feature to note is the increased vigour of the system as the heat flux is increased; this is reflected in the dimensions of the various circulation components. The total depth of the system increases from 750m for CC=0.025 to 1650m for CC=0.1, which is a reasonable result caused by the increase in the heat input. The velocities also change in a similar manner; the horizontal windfields show velocities of 1.19 and 3.62ms$^{-1}$ for the sea breeze components at the same values for CC and 0.50ms$^{-1}$ and 1.38ms$^{-1}$ respectively, for the return flows. The vertical velocities also exhibit this trend, with maximum updraught speeds of 0.87cm s$^{-1}$ increasing to 4.29cm s$^{-1}$. Another expected result of the variation is the effect on the time of onset of the sea breeze; with a value of 0.025 the sea breeze starts at 06:40hrs compared with 03:10hrs for a value of 0.1. This result is understandable as the pressure gradient driving the sea breeze will reach appreciable levels at earlier times as the amplitude of the heat flux is increased.
<table>
<thead>
<tr>
<th>Heat Flux Constant, CC</th>
<th>0.025</th>
<th>0.05</th>
<th>0.1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth of Sea Breeze (m)</td>
<td>450</td>
<td>450</td>
<td>750</td>
</tr>
<tr>
<td>Depth of Return Current (m)</td>
<td>300</td>
<td>900</td>
<td>900</td>
</tr>
<tr>
<td>Depth of Circulation (m)</td>
<td>750</td>
<td>1350</td>
<td>1650</td>
</tr>
<tr>
<td>Maximum Velocity in Sea Breeze (ms⁻¹)</td>
<td>1.19</td>
<td>2.22</td>
<td>3.62</td>
</tr>
<tr>
<td>Maximum Velocity in Return Current (ms⁻¹)</td>
<td>0.50</td>
<td>0.85</td>
<td>1.38</td>
</tr>
<tr>
<td>Horizontal Extent Inland (Km)</td>
<td>32</td>
<td>&gt;56</td>
<td>&gt;56</td>
</tr>
<tr>
<td></td>
<td>(11:20)</td>
<td>(11:00)</td>
<td>(08:50)</td>
</tr>
<tr>
<td>Horizontal Extent Seaward (Km)</td>
<td>8</td>
<td>32</td>
<td>&gt;60</td>
</tr>
<tr>
<td>Time of Onset (Hrs, simulated time)</td>
<td>06:40</td>
<td>04:20</td>
<td>03:10</td>
</tr>
<tr>
<td>Maximum Velocity in Updraught (cm s⁻¹)</td>
<td>0.87</td>
<td>2.28</td>
<td>4.29</td>
</tr>
<tr>
<td>Maximum Velocity in Downdraught (cm s⁻¹)</td>
<td>0.70</td>
<td>1.47</td>
<td>2.75</td>
</tr>
</tbody>
</table>

**Table 4.2: Sensitivity Test for the Heat Flux Constant**

* Initial stratification held constant at 2.5 K Km⁻¹.
Perhaps the most interesting feature of this test is the effect of an increase in the heat flux on the frontal propagation rate. Figures 15 to 18 give the horizontal wind fields and potential temperature fields for the heat flux constant values of 0.025 and 0.1. The aforementioned points concerning the vigour of the system are readily discerned, as are the more general features of the sea breeze circulation as outlined in the previous section. Looking at the potential temperature fields in more detail, it is interesting to note the development of the convectively mixed layer with time and the gradual encroachment of the cooler maritime air as it is advected inland behind the sea breeze front. A comparison of the two diagrams shows greater depths for the convective layer as the heat flux is increased and similarly for the inland extent of the sea air.

As in the previous section it is possible to estimate frontal propagation rates; these are given in Figure 19. The difference between the three values for the heat flux is quite evident; for CC=0.025, the average propagation rate is about 6.6Km hr⁻¹, but only manages to extend inland a distance of 32Km by 11:20 hrs. With CC=0.05, the sea breeze extends 56Km inland by 11:00hrs, with an average frontal propagation of about 8.8Km hr⁻¹. For CC=0.1, the front again reaches 56Km inland but at the much earlier time of 08:50hrs and hence a higher propagation rate of about 10.8Km hr⁻¹ results. The control on the rate of advection of maritime air is the pressure gradient across the coastline which has resulted from the contrast in surface sensible heat flux. It is, therefore, reasonable to expect a relationship between the frontal propagation and the amplitude of the heating function.
Following the sensitivity tests assuming a calm synoptic situation, it is a logical step to attempt some assessment of the model behaviour to variations in the impinging synoptic state. Observational studies have confirmed that the majority of sea breeze circulations develop with some sort of large scale wind field; Hope-Hislop (1974) in his study of the sea breezes for the Plymouth region found that for the 1972 season (March to September), 60% of them occurred with a synoptic scale component and for the 1973 season he found a value of 82%. In this model development, a geostrophic windfield is incorporated by setting the value of $V_g$ at the initial timestep. In the results presented in Table 4.3, the geostrophic component is set at 0.0, 2.0 and $-2.0\text{ms}^{-1}$, where a positive value implies an onshore flow and a negative value implies an offshore flow.

For the onshore situation, the horizontal temperature gradient driving the sea breeze circulation is expected to be hindered in its development and a relatively weaker system should result. An examination of the horizontal and vertical velocity maxima bears this out; for the sea breeze component, the velocity maximum decreases from 3.62 to $2.93\text{ms}^{-1}$ and in the return current from $1.38$ to $1.20\text{ms}^{-1}$. Figure 20 illustrates the horizontal wind field and a comparison with Figure 16 shows the tangible change to a weaker, more diffuse system.

The reduction in the return flow aloft may explain why the return flow is less readily observed. Frizzola and Fisher (1963) in a study of sea breezes at Long Island noted that the return current was difficult to detect and in many other studies the existence of the flow is noted but is not given a depth or speed eg. Craig, Katz and Harney (1945) and Johnson and O'Brien (1973).
<table>
<thead>
<tr>
<th>GEOSTROPHIC COMPONENT Vg (ms⁻¹)</th>
<th>-2.0</th>
<th>0.0</th>
<th>2.0</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEPTH OF SEA BREEZE (m)</td>
<td>450</td>
<td>750</td>
<td>750</td>
</tr>
<tr>
<td>DEPTH OF RETURN CURRENT (m)</td>
<td>1500</td>
<td>900</td>
<td>900</td>
</tr>
<tr>
<td>DEPTH OF CIRCULATION (m)</td>
<td>1950</td>
<td>1650</td>
<td>1650</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN SEA BREEZE (ms⁻¹)</td>
<td>4.23</td>
<td>3.62</td>
<td>2.93</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN RETURN CURRENT (ms⁻¹)</td>
<td>1.55</td>
<td>1.38</td>
<td>1.20</td>
</tr>
<tr>
<td>HORIZONTAL EXTENT INLAND (Km)</td>
<td>32</td>
<td>&gt;56</td>
<td>&gt;56</td>
</tr>
<tr>
<td></td>
<td>(11:40)</td>
<td>(08:50)</td>
<td>(05:40)</td>
</tr>
<tr>
<td>HORIZONTAL EXTENT SEAWARD (Km)</td>
<td>40</td>
<td>&gt;60</td>
<td>40</td>
</tr>
<tr>
<td>TIME OF ONSET (Hrs, simulated time)</td>
<td>04:20</td>
<td>03:10</td>
<td>03:40</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN UPDRAUGHT (cm s⁻¹)</td>
<td>6.90</td>
<td>4.29</td>
<td>1.23</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN DOWNDRAUGHT (cm s⁻¹)</td>
<td>2.88</td>
<td>2.75</td>
<td>2.37</td>
</tr>
</tbody>
</table>

**TABLE 4.3:** SENSITIVITY TEST FOR THE GEOSTROPHIC COMPONENT
The vertical velocity components in the sea breeze circulation are given in Figure 21 and immediately apparent is the somewhat uncharacteristic development of a stronger offshore downdraught region, with a maximum of 2.37 cm s\(^{-1}\) compared to the updraught maximum of 1.23 cm s\(^{-1}\). This weaker updraught development has been noted by several of the glider meteorologists (Findlater (1963), Wallington (1977)). The latter author in particular noted that the steady flow from the sea bringing in cool stable air, will dampen the potential for frontal development and hence strong updraught cores.

To further illustrate the enhanced advection of the cooler maritime air, Figure 22 gives the potential temperature fields for the onshore geostrophic component. A comparison with Figure 18 shows clearly the more rapid efflux of the sea air. The effect of this increased rate of advection with regard to the propagation of the system inland will be discussed later in this section.

The offshore gradient wind situation has received a great deal more attention in observational studies as it is generally accepted that the most pronounced sea breezes occur on days when there has been a fairly high offshore wind component during the morning, which will be replaced by the sea breeze during the late morning or early afternoon hours. Koschmeider (1936, 1941) noted that offshore winds were likely to force the temperature gradient and hence the pressure gradient behind the sea breeze circulation out to sea. The result of this is a more accentuated system with a shorter horizontal extent, larger velocity values and a shorter life cycle. Table 4.3 shows that the onset of the sea breeze does not occur until 04:20hrs compared to 03:10hrs for the calm situation. Previously, the sea breeze has started at the coastline, however, in this case the system is first seen to develop some 8Km offshore which is in fair agreement with Koschmeider's work. The maximum velocity in the sea breeze component
increases to 4.23ms\(^{-1}\) from 3.62ms\(^{-1}\) and in the return flow to 1.55ms\(^{-1}\) from 1.38ms\(^{-1}\). These features are illustrated in Figure 23 which gives the sea breeze horizontal wind fields.

A further striking feature of the sea breeze development under offshore geostrophic conditions is the growth of a much stronger overland updraught core as illustrated in Figure 24. There have been a great number of observational studies carried out to elucidate sea breeze frontal structures (Mackenzie (1956), Saunders (1958), Findlater (1963, 1964) and Simpson (1964, 1967)). The characteristic updraught core with large windspeeds of around 1 to 2 ms\(^{-1}\) has already been discussed in Section 4.2, as have the problems in adequately simulating the feature on a coarse grid. Despite this, the strong updraught is remarkably clear in the model simulations, with a maximum velocity of 6.90cm s\(^{-1}\) compared to 4.29cm s\(^{-1}\) in the calm situation. As mentioned earlier, the horizontal extent of the sea breeze under these conditions is also inhibited and this explains the restricted inland displacement of the updraught core. To further elucidate this, the potential temperature fields are given in Figure 25. Comparison with the earlier Figures 18 and 22 show the much slower encroachment of the maritime air. Using the same criteria as before, the frontal propagation rates can be determined. For the onshore situation, the sea breeze only reaches a distance of 32Km inland by 11:40hrs compared to 56Km at 08:50hrs for the calm situation and 56Km at 05:40hrs for the offshore geostrophic component. The frontal propagation rates are illustrated in Figure 26 and show generally linear displacements for all three situations. The rates on average are 19.8Km hr\(^{-1}\) for the onshore component, 5.5Km hr\(^{-1}\) for the offshore component and 11.2Km hr\(^{-1}\) for the calm situation. Clearly, the rate of encroachment is highly dependent on the general synoptic state.
4.5 Sensitivity Test for the Parameterisation used in Equation (2.8)

In Section 2.5 two types of surface heat flux parameterisation were introduced. In this section the sensitivity test concerns variations of the sensible heat flux as determined by the surface climate model of Wood (1977a). This model is run for any particular day by setting certain astronomical parameters; in particular, the apparent declination (DEC) and the true geocentric distance (R). Reasonable values must also be assigned to the dry bulb (TA) and wet bulb (TW) temperatures, with all other input parameters remaining unchanged, i.e., cloudless skies and a constant value of 0.5 for atmospheric diffusivity.

To assess the ability of the surface climate model in producing reasonable values for the heat fluxes, comparisons between measured and modelled Bowen ratios were made, where the Bowen ratio is defined as:

$$\beta = \frac{Q_H}{\lambda E}$$

where, $Q_H$ is the sensible heat flux density, and $\lambda E$ the latent heat flux density. Wood (1977b) studied heat fluxes at a range of sites in Southern England during the summers of 1974 and 1975. The Bowen ratios determined are given below in Table 4.4.
<table>
<thead>
<tr>
<th>SITE</th>
<th>DATE</th>
<th>TIME PERIOD (GMT)</th>
<th>BOWEN RATIO</th>
</tr>
</thead>
<tbody>
<tr>
<td>WINDSOR</td>
<td>11/9/74</td>
<td>1000 - 1500</td>
<td>0.36</td>
</tr>
<tr>
<td>ASCOT</td>
<td>11/9/74</td>
<td>1000 - 1500</td>
<td>0.38</td>
</tr>
<tr>
<td>WINDSOR</td>
<td>10/6/75</td>
<td>0900 - 1600</td>
<td>0.47</td>
</tr>
<tr>
<td>SILWOOD</td>
<td>10/6/75</td>
<td>0900 - 1600</td>
<td>0.51</td>
</tr>
<tr>
<td>WINDSOR</td>
<td>6/8/75</td>
<td>0900 - 1600</td>
<td>0.66</td>
</tr>
<tr>
<td>CARDINGTON</td>
<td>6/8/75</td>
<td>0900 - 1600</td>
<td>0.81</td>
</tr>
<tr>
<td>WINDSOR</td>
<td>7/10/75</td>
<td>1000 - 1500</td>
<td>0.56</td>
</tr>
<tr>
<td>HARLINGTON</td>
<td>7/10/75</td>
<td>1000 - 1500</td>
<td>0.50</td>
</tr>
<tr>
<td>CARDINGTON</td>
<td>7/10/75</td>
<td>1000 - 1500</td>
<td>0.73</td>
</tr>
</tbody>
</table>


The above table shows a range of values from 0.36 to 0.81, with a mean of 0.55. The surface climate model was run several times to make sure the simulated Bowen ratios fell within this range. The major control on the magnitude of $\beta$ proved to be the value set for the surface water availability factor (SWAF), as used in the calculation of the latent heat flux. For the sensitivity tests, the model was set up to run for a typical day in March and in June as follows:

<table>
<thead>
<tr>
<th>MARCH</th>
<th>JUNE</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEC = -4° 53' 48.7&quot;</td>
<td>DEC = 23° 20' 31.0&quot;</td>
</tr>
<tr>
<td>R = 0.9927</td>
<td>R = 1.0163</td>
</tr>
<tr>
<td>TA = 10°C</td>
<td>TA = 15°C</td>
</tr>
<tr>
<td>TW = 7°C</td>
<td>TW = 10°C</td>
</tr>
<tr>
<td>SWAF = 0.6</td>
<td>SWAF = 0.6</td>
</tr>
</tbody>
</table>
The Bowen ratio ranges between the hours of 1000 to 1500 GMT were 0.42 to 0.72 for March and 0.50 to 0.56 for June, with averages of 0.65 to 0.54 respectively for the same two months. Since these values are in reasonable agreement with those observed, the above input data sets were maintained.

The surface climate model is run at set intervals for the period 0000 to 2400 GMT. The resulting sensible heat flux profiles are illustrated in Figure 27. These type of profiles, with ranges of 134 to -88 Wm\(^{-2}\) for March and 234 to -128 Wm\(^{-2}\) for June, are fairly typical. Oke (1978) looked at several vegetated and non-vegetated surfaces, and produced curves similar to those presented. The heat flux values from 0600 to 1800 GMT correspond to the simulated time period 00:00hrs to 12:00hrs in the sea breeze model. With this assumption, it is clear that part of the simulated time period will have negative values for the heat flux implying, therefore, that heat will be lost from the surface layers of air to the ground surface, causing a subsequent temperature fall overland.

Table 4.5 gives the data for the sensitivity test. The major features are similar to those found in Section 4.2, where an increase in the amplitude of the heating function results in a stronger, more intense circulation. With regard to the horizontal velocities, in the sea breeze component, the maximum windspeed increases from 4.52 m\(s^{-1}\) in March to 6.70 m\(s^{-1}\) in June. Similarly for the return current, with values of 1.34 and 2.39 m\(s^{-1}\) respectively. The circulation depth also shows a marked change, with 1650m in March increasing to 2550m in June.
<table>
<thead>
<tr>
<th>HEAT FLUX FROM SURFACE CLIMATE MODEL</th>
<th>MARCH</th>
<th>JUNE</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEPTH OF SEA BREEZE (m)</td>
<td>750</td>
<td>1050</td>
</tr>
<tr>
<td>DEPTH OF RETURN CURRENT (m)</td>
<td>900</td>
<td>1500</td>
</tr>
<tr>
<td>DEPTH OF CIRCULATION (m)</td>
<td>1650</td>
<td>2550</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN SEA BREEZE (ms$^{-1}$)</td>
<td>4.52</td>
<td>6.70</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN RETURN CURRENT (ms$^{-1}$)</td>
<td>1.34</td>
<td>2.39</td>
</tr>
<tr>
<td>HORIZONTAL EXTENT INLAND (Km)</td>
<td>&gt;56</td>
<td>&gt;56</td>
</tr>
<tr>
<td></td>
<td>(1550)</td>
<td>(1310)</td>
</tr>
<tr>
<td>HORIZONTAL EXTENT SEAWARD (Km)</td>
<td>&gt;64</td>
<td>&gt;64</td>
</tr>
<tr>
<td>TIME OF ONSET (GMT)</td>
<td>1150</td>
<td>0920</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN UPDRAUGHT (cm s$^{-1}$)</td>
<td>5.18</td>
<td>7.90</td>
</tr>
<tr>
<td>MAXIMUM VELOCITY IN DOWNDRAUGHT (cm s$^{-1}$)</td>
<td>3.05</td>
<td>5.95</td>
</tr>
</tbody>
</table>

**TABLE 4.5:** SENSITIVITY TEST FOR HEAT FLUX FROM SURFACE CLIMATE MODEL
The times in the table are all given in GMT in accordance with the sensible heat flux data and the onset times as expected are different, with 1150 GMT for March and 0920 GMT for June. The vertical velocity fields also reflect the change in the circulation intensity, with values of 5.18 and 7.90 cm s\(^{-1}\) for the overland updraughts and 3.05 and 5.95 cm s\(^{-1}\) for the downdraughts.

The horizontal wind fields for the two situations are given in Figures 28 and 29. As well as the aforementioned features, it is interesting to note the very light land breeze present at 0900 GMT for March. This is a consequence of the negative values for the heat flux from 0600 to 0840 GMT; Atkinson (1981) noted that at this time the land breeze may still be prevalent as the air temperature overland has not yet risen above that over the sea. By 1200 GMT and 1500 GMT the sea breeze circulation has rapidly developed and spread inland in both March and June. By 1800 GMT the heat flux has again become negative and the circulation rapidly dissipates.

Figures 30 and 31 show the potential temperature fields for the same situations and once again the varying advection rates of the maritime air are quite apparent. Also worth noting is the increased convective layer depth achieved in the June situation. The average frontal propagation rates are determined as before and give values of 15.3 km hr\(^{-1}\) for March and 15.4 km hr\(^{-1}\) for June. Figure 32 illustrates these rates and shows that although they are more or less equal, the times of onset and the times which they extend furthest inland are approximately 2.5 hrs apart.

The sensitivity test analyses will be extended in the next chapter to include the influence of variations in the synoptic scale flow and coastal indentation in the full three-dimensional model.
CHAPTER 5 - THREE-DIMENSIONAL MODEL: SENSITIVITY TESTS

5.1 Introduction

The three-dimensional model tests are concerned with two specific areas; firstly, the response of the model to variations in the synoptic scale windfield, and, secondly, the influence of coastal configuration. As pointed out in the previous chapter the latter factor is one of the major controls on how the sea breeze circulation develops with time. The three-dimensional set up gives increased scope to examine this interaction. The first four sections will therefore look at the sea breeze system developing in response to a geostrophic component of 2.0 ms\(^{-1}\) from each of the quadrants.

Up to this stage the coastline has, by necessity, been regarded as straight, which is, of course, an unrealistic assumption for the majority of coastal regions. Since the model is now fully three-dimensional, coastal indentation can be readily incorporated. For these simulations a square bay of dimensions 8 x 8 x 8 Km is included as a simple representation of Plymouth Sound (see Figure 2). The last section will, therefore, examine the wind fields resulting from this added complication to the model geometry.

With regard to the run parameters, several things have changed in the three-dimensional simulations. The horizontal grid step is reduced from 8 Km to 4 Km in order to allow increased focus around the Plymouth Sound region. In addition, the time-step is also reduced from 600 to 120 s. This last alteration was necessary to allow a completely stable twelve hour simulation. The sea breeze windfields in the x–y plane are represented as vector plots, with
the section \( z=300 \text{m} \) used throughout.

5.2 Calm Geostrophic Situation

Figure 33 shows the \( u', v' \) vector-fields for 03:00hrs to 12:00hrs simulated time. The maximum depth attained by the system is approximately 1650m, with a maximum sea breeze velocity of 2.6ms\(^{-1}\). The depth of the return current is about 900m, with velocities reaching a maximum of 0.95ms\(^{-1}\). Using the same criterion as in the previous chapter, the time of onset is approximately 03:40hrs. The vertical velocity structure shows an updraught overland, with a maximum of 3.2cm s\(^{-1}\) compared to the offshore downdraught, with a maximum velocity of 2.6cm s\(^{-1}\).

Examining the vector fields in more detail, after 03:00hrs there is almost no perturbation to the calm state. By 06:00hrs, however, the sea breeze has developed both offshore and inland, with a slight deflection to the right. By 09:00hrs the perturbation is much more intense, with increased wind speeds and a more pronounced eastward deflection. At 12:00hrs the temperature gradient across the coastline has equalised and as a result the \( v' \)-component is greatly reduced. The resultant motion is therefore solely from the west.

The gradual development of the sea breeze system with increased turning to the right has been recorded at several coastal sites. Gill (1968) examined the diurnal cycle of sea breezes at Kinloss; he calculated hourly vector mean winds based on a treatment by Haurwitz (1947) and plotted the results on hodographs. For the months of April to September, a characteristic elliptic formation was found illustrating the gradual veering of the wind during the day. Pearce
(1968) gave an explanation of the turning by examining the role of Coriolis deflection; with zero-gradient wind conditions, the onshore sea breeze component is not in equilibrium since there is no related pressure gradient parallel to the coast. The air will, therefore, have a tendency to accelerate in a direction parallel to the coast, towards the side where high pressure would be required to maintain a steady wind of the same strength i.e. to the right in the northern hemisphere.

With regard to model results, Neumann and Mahrer (1971) noted the elliptic formation for wind direction in their numerical integration for the land and sea breeze systems. Saito (1976) developed a two- and three-dimensional simulation of the sea breeze over the Kanto district in Japan. He noted that the major control on the sea breeze direction was a combination of Coriolis effects and coastal configuration. This latter aspect will be examined in Section 5.5.

As a check on the models ability to simulate adequately flow fields throughout the model depth, a check is made using cross-sections in the y-z plane. The potential temperature fields in Figure 34 show clearly the gradual encroachment of the cooler, maritime air. As before, a frontal propagation rate is calculated giving a value of 8.8Km hr⁻¹, which compares favourably with both observed and modelled results as outlined in the previous chapter. The horizontal wind fields are also encouraging as shown in Figure 35, with the characteristic onshore and offshore components of the sea breeze circulation.

5.3 Northerly Geostrophic Situation

In this section, Ug is set to zero and Vg to -2.00 ms⁻¹, resulting in an offshore
geostrophic component impinging upon the sea breeze perturbation. Figure 36 shows vectors of $U$ and $V$, where $U = U_g + u'$ and $V = V_g + v'$, respectively, for 03:00hrs to 12:00hrs simulated time. The depth of the circulation reaches a maximum of around 1950m, with a maximum velocity in the sea breeze component of about 2.9ms$^{-1}$. In the return current, the maximum velocity perturbation reaches about 1.12ms$^{-1}$. As expected, these values are all larger than in the calm situation, emphasising the more intense circulation system which develops under these conditions. The time of onset is estimated as 04:00hrs and in the updraught overland a maximum speed of 6.69cm s$^{-1}$ is found, with 2.84cm s$^{-1}$ in the offshore downdraught region.

Looking at the vector fields, at 03:00hrs the general wind field is still relatively unperturbed and the northerly breeze is maintained. By 06:00hrs, however, there is a marked change around the coastal region, with a distinct onshore component at about 8Km offshore. This deflection out to sea is an observed feature and occurs as a result of the seaward advection of the temperature gradient, normally found at the coast for a calm situation. The sea breeze increases in range and intensity, encroaching about 8Km inland by 09:00hrs. By 12:00hrs the flow direction has been reversed around the coastal region, with the remnants of the perturbation resulting in a large u-component acting on the offshore velocities.

Observations of the classic development of a sea breeze system are arguably more consistently made during offshore gradient wind conditions when the system is at its most intense. This feature has been noted in several other sea breeze models. Estoque (1962), found that the strongest vertical motions occurred and that the landward penetration of the sea breeze was greatly reduced under offshore conditions. With regard to the latter aspect, the frontal
propagation rate is estimated as only 2.57 km hr\(^{-1}\), reaching a distance of only 12 km compared to >32 km for the calm synoptic state. Magata (1965), in his numerical study, noted that the shear as well as the velocity and the direction of the synoptic scale wind field was a major influence on the development of the sea breeze circulation. Pielke (1974a), in his definitive numerical study of the sea breezes over Florida, noted the role of sea breeze convergence zones on the locations of thunderstorm complexes over the peninsula. The exact position of these regions proved highly dependent on the synoptic scale wind field assigned to the model. In a study of the sea breezes developing over the Beaufort Sea in Alaska, Kozo (1982a, 1982b) noted stronger overland updraughts and mentioned the stalling of the sea breeze advance with offshore large scale flow. More recent work by Pearson, Carboni and Brusasca (1983) suggested that when a constant mean flow is introduced, the general structure of the sea breeze remains unchanged. The model used in this analysis is somewhat more complex than that described here, incorporating a detailed boundary layer parameterisation. Furthermore, the intensity criterion employed are related to the buoyancy gradient for point measurements and the total kinetic energy as a measure of the system as a whole. These aspects cannot be readily estimated in the present model due to the inherent assumptions in its construction and, therefore, comparison is difficult. It is likely that the true role of the mean flow can only be fully determined in a model which as well as including buoyancy and boundary layer calculations also includes adequate simulation of the turbulence momentum fluxes in the lower atmosphere.

As an additional check on the models ability to simulate adequate flow fields, several y-z plane cross-sections are reproduced. The potential temperature fields, given in Figure 37, show again the distinct growth of the overland convective boundary layer and its subsequent erosion by the sea breeze. The
movement of the system offshore at the time of onset is also apparent at 03:00hrs. The horizontal wind fields, shown in Figure 38, emphasise the more intense sea breeze component and the restricted spread inland of the system. Figure 39 gives the vertical velocity fields and shows clearly the increased intensity of the overland updraught.

5.4 Southerly Geostrophic Situation

For the southerly geostrophic component, $U_g$ is set to zero and $V_g$ to 2.00ms⁻¹ resulting in the addition of a component in the same direction as the sea breeze. The result of this should be a much weaker sea breeze system with the salient features less clearly defined. As outlined in the previous chapter this has been noted in several observational studies, including Frizzola and Fisher (1963) and Wallington (1977). The obvious reason for this decrease in intensity is that the continuous advection of cooler sea air inhibits the growth of the heated convective layer overland resulting in a much weaker horizontal pressure gradient.

Figure 40 gives the vector fields for the onshore geostrophic situation. The depth of the system is approximately 1650m, with maximum windspeeds of 2.09 and 0.79ms⁻¹ in the sea breeze and return flow respectively. The onset time is somewhat later than before at 04:40hrs and the vertical velocities are reduced, with a maximum updraught speed of only 1.09cm s⁻¹. The vector fields show the increase in the velocity of the onshore flow due to the sea breeze superimposing itself on the mean flow. In addition to the increased velocity, the flow field also shows a distinct clockwise rotation between 03:00hrs and 12:00hrs simulated time. This turning is not as great as in the calm situation.
due to the overriding influence of the geostrophic component and the pressure gradient it implies.

As mentioned above, the weakening of the circulation has been observed in several sea breeze studies; in addition, the weakening has also been numerically simulated. Estoque (1962) and Kozo (1982b) both noted reduced vertical motion and the rapid decay of the sea breeze system under conditions with an onshore gradient wind component. Figure 41 gives the potential temperature fields in the y-z plane and immediately clear is that the initial atmosphere is virtually undisturbed due to the continuous efflux of the cooler maritime air which does not give the convective boundary layer overland a chance to develop. This feature is borne out in the horizontal wind fields given in Figure 42. The most obvious feature is the very rapid advance of the perturbation inland; estimation of the frontal propagation rate as before resulted in a value of about 12Km hr⁻¹ reaching 28Km inland after only 05:00hrs simulated time. Despite, this much more rapid advance, all the velocity values are considerably reduced.

To summarise the results in the previous sections, Table 5.1 lists the various components of the sea breeze simulation for the three sensitivity tests.
### Table 5.1: Sensitivity Tests for the Geostrophic Component in the Three-Dimensional Simulations

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Offshore</th>
<th>Onshore</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Gradient Wind (ms⁻¹)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calm</td>
<td>2.0</td>
<td>2.0</td>
</tr>
<tr>
<td><strong>Depth of Sea Breeze (m)</strong></td>
<td>750</td>
<td>750</td>
</tr>
<tr>
<td><strong>Depth of Return Current (m)</strong></td>
<td>900</td>
<td>1200</td>
</tr>
<tr>
<td><strong>Depth of Circulation (m)</strong></td>
<td>1650</td>
<td>1950</td>
</tr>
<tr>
<td><strong>Maximum Velocity in Sea Breeze (ms⁻¹)</strong></td>
<td>2.6</td>
<td>2.9</td>
</tr>
<tr>
<td><strong>Maximum Velocity in Return Current (ms⁻¹)</strong></td>
<td>0.95</td>
<td>1.12</td>
</tr>
<tr>
<td><strong>Horizontal Extent Inland (Km)</strong></td>
<td>&gt;32</td>
<td>12</td>
</tr>
<tr>
<td><strong>Horizontal Extent Seaward (Km)</strong></td>
<td>&gt;32</td>
<td>&gt;32</td>
</tr>
<tr>
<td><strong>Time of Onset (hrs, simulated time)</strong></td>
<td>03:40</td>
<td>04:00</td>
</tr>
<tr>
<td><strong>Maximum Velocity in Updraught (cm s⁻¹)</strong></td>
<td>3.2</td>
<td>6.69</td>
</tr>
<tr>
<td><strong>Maximum Velocity in Downdraught (cm s⁻¹)</strong></td>
<td>2.6</td>
<td>2.84</td>
</tr>
</tbody>
</table>

**5.5. Westerly/Easterly Geostrophic Situations**

It is an idiosyncracy of sea breeze analysis that comparatively little work has been carried out in examining the system under conditions where the general wind field blows parallel to the coast. Both Koschmeider (1936) and Wexler
(1946) considered that sea breeze development was unlikely under these conditions; Wallington (1977), however, suggested that airstreams parallel to the coast provided the optimum situation for sea breeze frontal development, especially when coupled with deep inland convection. A similar lack of attention is also to be found with numerical studies of the sea breeze system. The underlying cause of this neglect is that it is much easier to determine the major features of the sea breeze when the general wind field is blowing offshore. Furthermore, the incorporation of an alongshore component can only be simulated in a complete three-dimensional construct.

Figures 43 and 44 give the vector fields for the situations with an easterly and westerly geostrophic component respectively. For the latter, there is a complete reversal in the wind direction between 03:00hrs and 12:00hrs overland. At the intermediate time of 06:00hrs the easterly flow is perturbed in a band around the coast some 20Km inland and 16Km offshore, within which the flow is predominantly in a northerly direction. At 09:00hrs the flow over the whole domain is predominantly latitudinal, with a distinct turning to the right.

For the easterly geostrophic situation the influence of the sea breeze has resulted in a slight onshore turning of the wind at 06:00hrs and 09:00hrs respectively. At all times the wind speeds are considerably greater than the ambient geostrophic value reaching a maximum at 12:00hrs. This westerly enhancement is due to the rightwards deflection of the sea breeze by the Coriolis force.

5.6 Coastal Indentation

As outlined in Section 5.1, the development of a three-dimensional model, as
well as allowing increased scope in the direction of the large-scale wind, also allows the incorporation of coastal orography. In this section, two sets of results are presented; the first is for a calm synoptic state with the inclusion of a square shaped bay of 8 x 8 x 8Km and with the use of the simpler forcing function as described in Section 2.5. The second set of results uses the same model geometry, however, the diabatic heat flux is calculated using the surface climate model of Wood (1977a) for a typical day in March.

Several sea breeze models have been developed in recent years which have incorporated some sort of coastal variation. The definitive paper by McPherson (1970) provided a simulation for the Galveston Bay region in Texas. He noted that the presence of the bay led to a landward distortion of the sea breeze convergence zone which diminished with time as the system moved inland. Furthermore, the complex coastline led to two zones of ascent in opposite corners of the bay region. Since this early paper, several projects have incorporated increasingly complex coastal configurations. For example, Pielke (1974) simulated the complex coastal outline of the Florida peninsula, while Mahrer and Pielke (1976) provided a numerical simulation of the flow over Barbados using an improved sigma coordinate version. Similar computations have also been carried out in Japan; Saito (1976) noted the development of particular zones of convergence and divergence around the Boso peninsula in the Kanto district. More recently, Kikuchi, Arakawa, Kumura, Shirasaki and Nagano (1981) have undertaken a more complex analysis of the same region by incorporating topography in a complete terrain-following coordinate simulation.

Several more general mesoscale models have included coastal variation; for example, Golding and Machin (1984) reported on the mesoscale model developed at the Meteorological Office in which the coastline of the UK is simulated on a grid consisting of 15Km squares. Several one-level mesoscale models also
incorporate coastal variation, including Danard (1977) and Mass and Dempsey (1984).

Figure 45 shows the horizontal wind vector fields for 3, 6, 9 and 12 hrs simulated time. Although at first glance the bay does not appear to have significantly distorted the sea breeze flow field, a closer look at the 03:00hrs and 06:00hrs situations reveals some interesting features. At 03:00hrs the sea breeze is still very light, however, on magnification (Figure 46), the divergent flow field centred on the bay region is quite apparent. By 06:00hrs the inland extent of the sea breeze as well as the magnitude of the velocities has greatly increased. As a consequence the disturbance caused by the bay has been advected inland; the region 20-30Km inland immediately north of the bay has much stronger velocities than corresponding areas to either side. Another interesting feature is the increased veering of the wind to the east of the bay region compared to the west. The reason for this asymmetry is that the pressure gradient force and the Coriolis force are acting in concert to the east, whereas they are in opposition to the west. By 09:00hrs the sea breeze system has reached maximum intensity and the influence of the bay on the flow field has disappeared.

As a further illustration of the more complex situation resulting from the bay, the potential temperature fields at 300m are shown in Figure 47. At 03:00hrs the temperature gradient is centred over the coastline, with a distinct landward distortion around the bay region. By 06:00hrs the amplitude of the distortion has decreased considerably while the wavelength has increased. The zone of influence, as a result, has extended further inland and alongshore. By 09:00hrs the bay disturbance has disappeared, with a much weaker temperature gradient prevailing normal to the coastline.
The inclusion of the bay indentation also leads to a distorted vertical velocity field as illustrated in Figure 48. At 03:00hrs the bay has resulted in a cellular fragmentation of the vertical velocity field, with a downdraught core centred over the bay region and two updraught cores to the west and east. By 06:00hrs the downdraught core has expanded and increased in intensity, with an associated band of descending air about 8Km offshore. The updraught cores are still prominent but have now been advected about 20Km inland. The picture at 09:00hrs is less clearly defined, with the disappearance of the updraught cores and the breakdown of the structured downdraught belt offshore. By 12:00hrs the vertical velocity field has now lost its cellular structure completely, with the dissipation of the temperature gradient across the coastline.

The information shown in Figure 48 is somewhat misleading with regard to the detailed structure of the updraught cores. A closer examination of the data files revealed that the core to the northeast of the bay had relatively higher velocities than that to the northwest. Although the difference between the two maxima was quite small, it is likely that a finer mesh model would highlight this feature to an even greater extent. McPherson (1970) noted an asymmetric distribution of the updraught cores in his model and, as outlined earlier, the cause is likely due to the opposing actions of the pressure gradient and Coriolis forces on either side of the bay.

In the following chapter, two case study runs will be presented which will include the surface climate model of Wood (1977a) in the determination of the diabatic heat flux. As a further test on the influence of the bay indentation, a simulation for a typical day in March, as in Section 4.5, is described.

The horizontal wind field vectors are given in Figure 49; at 03:00hrs the
remnants of an overnight land breeze are still in evidence, with a light offshore flow. By 06:00hrs, however, a sea breeze has developed and the influence of the bay is clearly in evidence. At 09:00hrs and 12:00hrs the sea breeze has expanded and increased in intensity, with the consequence that the initial perturbation of the field caused by the bay has disappeared.

The potential temperature fields given in Figure 50 show a similar growth pattern. At 03:00hrs the temperature gradient across the coastline is commensurate with a land breeze, however, by 06:00hrs the situation has been reversed, with a distortion evident landward of the bay region. With the growth of the sea breeze and the increased advection of the cooler sea air, the temperature gradient becomes less well-defined and by 12:00hrs has reverted back to the land breeze situation.

The vertical velocity situation is given in Figure 51; at 03:00hrs there is very little perturbation to the general pattern, however, by 06:00hrs the characteristic asymmetric distribution has been initiated. The dichotomy between the two updraught cores is more pronounced than the previous simulation, with the strongest development to the east of the bay region and the downdraught area focussed over the bay. At 09:00hrs these features are still maintained although to a much lesser degree due to the rapid extension of the sea breeze system horizontally.

The next chapter presents two case studies of particular sea breeze events for Plymouth during August 1983 and May 1984 and compares available observations with results from a three-dimensional simulation.
6.1 Introduction

This chapter will examine two particular sea breeze events on the 11th August 1983 and the 9th May 1984. As a precursor to this discussion, an analysis of sea breeze data for the summers of 1982 and 1983 is presented using data from the M.O. station at R.A.F Mountbatten, Plymouth and the Faculty of Maritime Studies weather station at Plymouth Polytechnic. In addition, pilot balloon ascents and upper air analyses from M.O. Camborne are used for details of the vertical structure on particular case study days.

6.2 Climatology of the Plymouth Region

A major observational study of the sea breezes indigenous to the Plymouth locale was carried out by Hope-Hislop (1974). As a component of this research he also presented a concise analysis of the climatology of the area, the major results of which are summarised in Table 6.1. The important features to note are first of all the high amount of insolation during the summer months in accordance with the preponderance of high pressure systems over the U.K. and the resulting high temperatures overland. This feature is also borne out in the precipitation figures. The maximum temperature differential between land and sea occurs in June, which corresponds well with the wind data. The maximum frequency for a wind direction in the 200-220° sector in the speed range 2-5ms⁻¹ also occurs in the same month.
<table>
<thead>
<tr>
<th></th>
<th>MEAN ANNUAL (mm)</th>
<th>MEAN MONTHLY (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>PRECIPITATION</strong></td>
<td>1000</td>
<td>50 (JUNE)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>112 (NOV/DEC)</td>
</tr>
<tr>
<td><strong>MEAN ANNUAL (hrs)</strong></td>
<td></td>
<td>1677</td>
</tr>
<tr>
<td><strong>SUNSHINE</strong></td>
<td>1677</td>
<td>219 (MAY)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>222 (JUNE)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>198 (JUL/AUG)</td>
</tr>
<tr>
<td><strong>MEAN ANNUAL FREQUENCY OF</strong></td>
<td></td>
<td>83</td>
</tr>
<tr>
<td><strong>THICK FOG (&lt;200m) (days)</strong></td>
<td></td>
<td>1.1 (MAR)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.5 (APR/JUN)</td>
</tr>
<tr>
<td><strong>MEAN MONTHLY MINIMUM °C</strong></td>
<td>6.2 (JANUARY)</td>
<td>16.2 (AUGUST)</td>
</tr>
<tr>
<td><strong>MEAN MONTHLY MAXIMUM °C</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>AIR TEMP (T_A)</strong></td>
<td>6.2 (JANUARY)</td>
<td>16.2 (AUGUST)</td>
</tr>
<tr>
<td><strong>SEA TEMP (T_S)</strong></td>
<td>8.0 (FEB/MAR)</td>
<td>15.8 (AUGUST)</td>
</tr>
<tr>
<td></td>
<td>4.0 (JUNE)</td>
<td></td>
</tr>
<tr>
<td><strong>MAXIMUM MEAN FREQUENCY (%)</strong></td>
<td>55 (180-360°)</td>
<td>2.0 (DECEMBER)</td>
</tr>
<tr>
<td></td>
<td>36.4 (0-180°)</td>
<td>21.5 (JUNE)</td>
</tr>
<tr>
<td></td>
<td>8.4 (CALM)</td>
<td></td>
</tr>
</tbody>
</table>

**TABLE 6.1:** SELECTED CLIMATOLOGICAL DATA FOR PLYMOUTH, 1916-1950. [SOURCE: HOPE-HISLOP (1974)]
Hope-Hislop's study provided some excellent background material on the structure and development of the sea breezes in the Plymouth region as presented below for the 1972 and 1973 seasons.

(1) Predominantly a summer phenomenon but possible from March to September, with a maximum intensity in July.

(2) Sea breeze only developed when the synoptic scale wind field was calm or blowing offshore, with a speed not greater than $8.0\text{ms}^{-1}$.

(3) Sea breeze only occurred when the land-sea temperature differential was greater than $2^\circ\text{C}$.

(4) The sea breeze mean wind speed and direction was recorded as $3.5\text{ms}^{-1}$ blowing at $190^\circ$.

(5) Mean duration of breeze calculated at 5hrs 42mins, with a mean time of onset at 1045 GMT and 1535 GMT for the time of abatement.

(6) The horizontal extent was on average 13Km inland and 7Km offshore.

(7) Sea breeze depth varied temporally between 200 and 700m, with a mean of about 500m.

(8) The majority of sea breezes occurred under stable atmospheric conditions.

(9) Detailed observations of wind speed and direction showed the sea breeze to be topographically constrained by Plymouth Sound, with corresponding
divergent and convergent wind fields.

As an adjunct to these conclusions, sea breeze data for the 1982 and 1983 seasons from April to September are presented in Appendix D. In total, over the two summer periods, 52 sea breeze events were recorded with 23 in 1982 and 29 in 1983. Table 6.2 outlines the mean values ascertained from the recorded data.

<table>
<thead>
<tr>
<th></th>
<th>R.A.F MOUNTBATTEN</th>
<th>PLYMOUTH POLYTECHNIC</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>OBSERVED SEA BREEZES (1982)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TIME OF ONSET (GMT)</td>
<td>1100</td>
<td>1145</td>
</tr>
<tr>
<td>TIME OF ABATEMENT (GMT)</td>
<td>1815</td>
<td>1750</td>
</tr>
<tr>
<td>DURATION (hrs)</td>
<td>7.25</td>
<td>6.08</td>
</tr>
<tr>
<td>MEAN WIND SPEED (ms⁻¹)</td>
<td>3.67</td>
<td>3.84</td>
</tr>
<tr>
<td>MEAN WIND DIRECTION (º)</td>
<td>200</td>
<td>190</td>
</tr>
<tr>
<td><strong>OBSERVED SEA BREEZES (1983)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TIME OF ONSET (GMT)</td>
<td>1100</td>
<td>1115</td>
</tr>
<tr>
<td>TIME OF ABATEMENT (GMT)</td>
<td>1800</td>
<td>1750</td>
</tr>
<tr>
<td>DURATION (hrs)</td>
<td>7.00</td>
<td>6.58</td>
</tr>
<tr>
<td>MEAN WIND SPEED (ms⁻¹)</td>
<td>3.14</td>
<td>3.91</td>
</tr>
<tr>
<td>MEAN WIND DIRECTION (º)</td>
<td>206</td>
<td>193</td>
</tr>
</tbody>
</table>

**TABLE 6.2:** ANALYSIS OF SEA BREEZE DATA APRIL TO SEPTEMBER 1982 AND APRIL TO SEPTEMBER 1983.

The above averaged data compares favourably with the results of Hope-Hislop, although there are slight differences in the sea breeze duration. An additional feature of interest is the difference between the data from the two stations; for R.A.F. Mountbatten, the sea breeze tends to have a slightly longer duration with an earlier onset and a later abatement. These differences are commensurate with
the more exposed situation of the Meteorological Office station. Another interesting feature is the slight difference in the wind speeds, with larger values recorded at Plymouth Polytechnic; this may be a feature of the convergent/divergent windfield resulting from the coastal configuration. The statistical significance of the above analysis is not very high and a more detailed study of the local structure both spatially and temporally is required.

6.3 Case Study 1: 11th August 1983

6.3.1 The synoptic situation

The synoptic situation over the U.K. was dominated by an anticyclone to the west of Ireland, with an associated ridge of high pressure extending west to east across the country into Western Europe. To the north, a relatively weak low pressure system was centred over Iceland, with frontal troughs trailing across Scandinavia. The synoptic chart for 1200 GMT on the 11th August 1983 is presented in Figure 52.

More locally, an upper analysis at 1200 GMT for M.O. Camborne is given in Figure 53. The temperature environment curve shows two distinct layers between the surface and 770mb; the first layer between 1000 and 900mb is conditionally unstable topped by a second absolutely stable layer from 900 to 770mb. This relatively complex profile in the lower atmosphere is averaged to give a figure for the initial stratification,

\[ \frac{\Delta \theta}{\Delta z} = 2.7 \text{ K km}^{-1}. \]
The midday ascent for Camborne also yielded data on the geostrophic wind field with a windspeed of 6.5 ms$^{-1}$ at 850mb. Clearly a value of $V_g$ at a height of around 1.5Km is not a representative figure for the influence of the synoptic scale on the sea breeze circulation. To try to produce an average figure for the depth of the sea breeze system, a power-law profile is assumed, i.e.

$$V_g = V_{go} \left( \frac{Z}{Z_o} \right)^\alpha$$

where: $V_{go} = 6.5$ ms$^{-1}$; $Z_o = 1500$ m; $\alpha = 0.6$.

A more realistic reference level is taken as 300m; from the pilot balloon ascents for a range of sea breeze events this appears to be an average figure for the level of neutrality. In addition, mass continuity suggests a total circulation depth of 900m for a return current windspeed of 50% of the sea breeze value; for a column of this height, the approximate level of mean density is also 300m. At the 300m level the resulting value for the geostrophic windspeed is,

$$V_g = 2.47$$ ms$^{-1}$.

This component is considered to blow at the same angle as that found for the 850mb level.
6.3.2 The observed sea breeze

On the 11th August 1983 the anemograph at the Plymouth Polytechnic weather station showed a light northerly breeze at the surface backing, to a southerly around 0815 GMT. By 0900 GMT the breeze had risen to a speed of around 1.0 ms\(^{-1}\); this gradually increased in intensity with a maximum gust of 8.0 ms\(^{-1}\) at 1300 GMT and an average of 4.0 ms\(^{-1}\). The sea breeze blew steadily throughout the afternoon and dissipated very sharply at 1700 GMT.

During the day several pilot balloon ascents were made from the Plymouth Polytechnic weather station; the southerly components calculated are presented in Figure 54. These show clearly the gradual increase in intensity of the sea breeze with onshore maxima of 1.84, 3.78, 4.82 and 4.98 ms\(^{-1}\), respectively, for the four runs. The return flow aloft is less easily defined due to the swamping effect of the synoptic field.

Using the above details, a set of sea breeze simulation criteria can be defined as follows:

- Time of onset.
- Mean maximum velocity in sea breeze.
- Time of mean maximum velocity.
- Mean sea breeze depth.
- Time of offset.

Additional features to note are:

- Evidence of a temporal increase in sea breeze depth.
Limited horizontal extent, both landward and seaward.

Evidence of any topographical constraints on the sea breeze.

6.3.3 The simulated sea breeze

The results are presented as previously with 3, 6, 9 and 12 hours simulated time being the equivalent of the period 0600 GMT to 1800 GMT. This corresponds to a 12 hour period from the surface climate simulation of Wood (1977a).

Figure 55 shows the horizontal wind fields in a cross-section through the centre of the simulated bay; at 0900 GMT the sea breeze is not yet evident, with a very slight offshore breeze prevailing. By 1200 GMT a sea breeze has started, with the time of onset estimated as 1030 GMT. The maximum velocity in the sea breeze component at 1200 GMT is about 1.1ms\(^{-1}\), with about 0.2ms\(^{-1}\) in the return flow aloft. By 1500 GMT the sea breeze is more intense, with a maximum of about 3.6ms\(^{-1}\) in the onshore component and 0.9ms\(^{-1}\) for the offshore. Figure 56 shows the horizontal wind profiles at the coast in the same cross-section through the bay region and shows clearly the gradual increase in intensity of the sea breeze with time. The profiles also show the sea breeze depth to be, on average, around 1Km, with a much weaker and less well-defined flow aloft.

With regard to the potential temperature structure, the cross-sections in Figure 57 describe the evolving situation between 0900 and 1800 GMT. At 0900 GMT the temperature gradient is aligned with slightly higher values over the sea at the lowest levels leading to a slight landward drift of restricted depth. By 1200 GMT a characteristic adiabatic convective profile has developed overland which is undergoing erosion by the cooler maritime air. By 1500 GMT the depth of
this layer has increased slightly and the maritime inflow is being restricted. This inhibition of the inland efflux is probably due to two factors; firstly, the increasing eastward Coriolis component forcing the sea breeze to blow at a more significant angle alongshore and, secondly, the damping effect of the offshore synoptic scale wind. The final cross-section at 1800 GMT depicts a depth-restricted inflow region, with the shoreline temperature gradient beginning to equalise.

As a further illustration of the three-dimensional potential temperature structure, Figure 58 shows the patterns at the 300m level over the Plymouth Sound region. At 0900 GMT the temperature gradient is commensurate with a land breeze regime. In addition, an interesting feature is the seaward 'kink' caused by the presence of the bay. By 1200 GMT the land breeze situation has been superseded by a landward temperature gradient typical of a sea breeze; the bay simulation has resulted in a landward distortion of the temperature field. By 1500 GMT the temperature gradient along the coast is less sharply defined, with the distortion now displaced slightly to the right. As mentioned previously this deflection to the east is a function of the increasing importance of the Coriolis component. At 1800 GMT there is little perturbation of the surface layers, with a return to an equilibrium state between land and sea.

Further details of the three-dimensional nature of the flow patterns can be ascertained in Figure 59 which shows the vertical velocity situation at the 300m level. At 0900 GMT there is no visible sign of any mesoscale disturbance, however, by 1200 GMT a bimodal updraught distribution has developed around the bay, with two cores to the east and west of the indentation and maximum velocities of 3.4cm s⁻¹ and 4.6cm s⁻¹, respectively. The downdraught is focussed over the bay region, with a maximum of 2.8cm s⁻¹. At 1500 GMT the system
appears more intense, with maxima of 7.0 and 7.9 cm s$^{-1}$. Although still
displaying the two updraught cores, there is some evidence that the
topographically forced perturbations are dissipating with the merging of the
updraughts and the significant spread of the downdraught region. The final
diagram reveals a less well defined cellular pattern for 1800 GMT although the
salient features still remain.

6.3.4 Discussion

Using the criterion as proposed in section 6.3.2, the table given overleaf
summarises the general comparison between the observed and simulated sea
breezes.

The difference between the time of onsets is about 2.5 hrs and suggests a
significant failure of the simulation. However, the differences are a direct result
of the coarse grid resolution and the form of heating function employed. The
former problem is a far-reaching one in all numerical simulations in that the
grid employed may have a bearing on the results obtained. In this instance, the
horizontal domain is 64 x 64 Km consisting of 256 cells each 4 x 4 Km. In
terms of the Plymouth locale, the distance between the Polytechnic and the north
edge of Plymouth Sound is about 1.5 Km, with another 6 Km of complex
coastline to the seaward limit of the bay. To do justice to a feature such as
the onset time a much finer grid is required to adequately resolve all the
extraneous factors influencing the development of the sea breeze. With regard
to the form of the heating function, the sinusoidal variation around a maximum
at midday accordingly reproduces a characteristic smooth onset for the simulated
onshore flow. This is somewhat unlike the real situation where the sea breeze
onset can be very sudden indeed.
<table>
<thead>
<tr>
<th><strong>OBSERVED</strong></th>
<th><strong>SIMULATED</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>TIME OF ONSET (GMT)</strong></td>
<td>08:15</td>
</tr>
<tr>
<td><strong>MEAN MAXIMUM VELOCITY (ms(^{-1}))</strong></td>
<td>4.0</td>
</tr>
<tr>
<td><strong>TIME OF MAXIMUM VELOCITY (GMT)</strong></td>
<td>1330</td>
</tr>
<tr>
<td><strong>MEAN SEA BREEZE DEPTH (m)</strong></td>
<td>300</td>
</tr>
<tr>
<td><strong>TIME OF OFFSET (GMT)</strong></td>
<td>17:00</td>
</tr>
<tr>
<td><strong>AVERAGE FRONTAL PROPAGATION RATE (Km hr(^{-1}))</strong></td>
<td>UNKNOWN</td>
</tr>
<tr>
<td><strong>TOTAL HORIZONTAL EXTENT (Km)</strong></td>
<td>UNKNOWN</td>
</tr>
</tbody>
</table>

**EVIDENCE OF TEMPORAL CHANGE IN DEPTH OF SEA BREEZE**
- From 225m at 09:15 GMT to 325m at 14:15 GMT
- From 450m at 1200 GMT to 750m at 1500 GMT

**EVIDENCE OF TOPOGRAPHICAL CONSTRAINTS**
- Bimodal updraught distribution and curvilinear temperature distribution
- UNKNOWN

**TABLE 6.3: COMPARISON DATA FOR SIMULATED AND OBSERVED SEA BREEZES: CASE STUDY 1 - 11th AUGUST 1983**

Comparison of the mean maximum velocities is also a difficult problem due to the 'gusty' nature of the sea breeze. However, the numerical results should be regarded as time-averaged ensembles of the highly turbulent onshore flow more often observed. Only with a more realistic turbulent closure will more precise comparisons with the numerical data be appropriate.

The simulated time of the maximum velocity is also severely compromised due to the aforementioned problems with the time of onset. Clearly, improvements...
in the initial onset time will enhance the predictive potential of all other temporal data.

The numerical results for the mean sea breeze depth and the transient change in depth appear initially to be greatly at odds with the observations, however, the simulations are critically influenced by the vertical grid resolution. The 300m cells are inhibiting in that this only allows an accuracy of ±150m. Despite this, perhaps the most encouraging feature is the prediction of a definite change in the depth of the system with time; accurate values, however, will only be attained with improved vertical resolution.

The predicted offset times also need care in interpretation due again to the heating function discussed previously. For the observed sea breeze of 11th August 1983, the onshore component dissipated suddenly around 1700 GMT, however, commensurate with the sinusoidal forcing in the model, a more smooth transition is predicted. Hope-Hislop (1974) reported some concern over the extreme variability of sea breeze duration. During his analysis he noted nine sea breeze events which persisted beyond sunset; however, using the formulation of Zambakas (1973) where the time of abatement is calculated as a function of the rate of heating and latitude, the mean offset time for the Plymouth locale was determined as falling before sunset. Clearly, the offset time is not a simple function of two variables; it is likely that other external factors such as topography will play an important role.

The frontal propagation rate is estimated to be about 7.11Km hr\(^{-1}\). Although no local observational data is available this does appear to be a consistent value; Simpson (1964), for example, considered 7.2 to 10.8Km hr\(^{-1}\) to be fairly typical for Southern England. As with the other results, consideration must be given to
the assumptions inherent in the model framework. It is likely that the role of topography is crucial in determining the inland propagation of the sea breeze; this will either prevent or enhance the inland efflux, depending on the orientation of the topographic feature in question to the sun.

The simulated horizontal extent is somewhat greater than the average figure quoted by Hope-Hislop; this lack of agreement may be related to the topographic question, although the rate of heating and the influence of the synoptic scale are also critical.

One of the more striking features of the simulation concerns the evidence of topographical constraints. Reference has already been made to the prominent features of the vertical velocity and temperature distributions; the reasons for the development of perturbations normal to the axis of the sea breeze is directly related to the convergent/divergent windfield which occurs as a consequence of the complex coastal configuration in and around Plymouth. Hope-Hislop did find some evidence of a divergent flow field at the northward end of Plymouth Sound and associated eddies in the smaller bays in the immediate locale. Although more corroborating evidence is required, it seems likely that the complex coastal structure does play an important role in the surface wind fields.

6.4 Case Study 2: 9th May 1984

6.4.1 The synoptic situation

The synoptic situation over the U.K. was dominated by an elongated ridge of high pressure, running north to south off the west coast of Ireland. To the
north of the country a weak trailing frontal trough resulted in somewhat cloudier weather in Scotland. In more southerly parts, the weather was clearer and the synoptic flow was dominated by a north easterly airstream. The synoptic chart for 1200 GMT on the 9th May 1984 is presented in Figure 60.

The upper-air analysis, given in Figure 61, shows the vertical temperature and dew-point structure at 1200 GMT for M.O. Camborne. The former profile displays a layered structure, with a conditionally unstable layer up to approximately 800mb capped by a small inversion of 2 degK. In the 1000-800mb layer the mean potential temperature stratification is estimated as,

\[
\frac{\Delta \theta}{\Delta z} = 3.7 \text{ K km}^{-1}
\]

The synoptic scale wind is calculated as before using a power-law; with \( V_{go} = 11 \text{ ms}^{-1} \) at 850mb, this results in a characteristic geostrophic wind at 300m of:

\[
V_g = 4.19 \text{ ms}^{-1}
\]

6.4.2 The observed sea breeze

Examination of the anemograph at the Plymouth Polytechnic weather station showed a northerly breeze overnight on the 9th May which backed over a period of about 1.5hrs to a southerly onset around 0900 GMT. The intensity of the sea breeze gradually increased to an average of about 5.0ms\(^{-1}\) at 1100 GMT which continued unabated until about 1645 GMT when a very rapid veer to the
north occurred. The maximum gust of about $9.0 \text{ms}^{-1}$ was recorded at 1400 GMT.

Two pilot balloon ascents were made from the Plymouth Polytechnic weather station as presented in Figure 62. The calculated southerly components show a sea breeze depth at 1030 GMT to 1050 GMT of around 250 to 400m, with a wind speed at the surface of 4.1 to $3.2 \text{ms}^{-1}$. As with all sea breeze situations developing in an offshore flow, details of the return component are impossible to ascertain.

The sea breeze criteria specified in section 6.3.2 will again be used to compare the observed and simulated data.

6.4.3 The simulated sea breeze

Figure 63 shows the horizontal wind fields in a cross section through the centre of the bay region; at 0900 GMT the flow regime shows a light land breeze, with a maximum offshore windspeed of $0.4 \text{ms}^{-1}$. A sea breeze circulation has developed by 1200 GMT, with a time of onset of approximately 1015 GMT. Midway through the simulation, the wind velocity in the sea breeze is about $0.9 \text{ms}^{-1}$, with a maximum of $0.2 \text{ms}^{-1}$ in the return flow. By 1500 GMT the sea breeze is at its peak, with a maximum horizontal wind speed of $2.56 \text{ms}^{-1}$ and $0.9 \text{ms}^{-1}$ in the return current. At 1800 GMT the sea breeze current has dissipated, with the return flow aloft somewhat enhanced to about $1.1 \text{ms}^{-1}$. A feature which is remarkable to this simulation is the apparent displacement of the whole sea breeze system offshore; this problem is related to the magnitude of the mean flow component and will be discussed in a later section.
The horizontal profiles commensurate with the above are given in Figure 64 and show the developing situation at the head of the simulated bay. The aforementioned seaward displacement is clearly illustrated, with the profile for 1800 GMT showing no sign of any mesoscale perturbation. Despite this, there is a distinct increase in the intensity and depth of the sea breeze between 1200 and 1500 GMT; 0.75ms\(^{-1}\) to 1.6ms\(^{-1}\), with a depth change from 575m to 800m. The return flow aloft also becomes more prominent over this time period, with an increase from 0.1 to 0.6ms\(^{-1}\).

Cross-sections of the potential temperature structure are presented in Figure 65. At 0900 GMT a positive temperature gradient exists between sea and land resulting in the light land breeze regime mentioned earlier. By 1200 GMT a well mixed convective boundary layer has developed overland to a depth of 1Km with a strong temperature gradient across the coastline. By 1500 GMT the depth of this layer has increased to about 1.5Km and the sharp temperature discontinuity has encroached slightly further northwards. The extent of the inland efflux is somewhat restricted compared to the previous case study which is a direct result of the lower heating rate and the stronger offshore gradient flow. The situation at 1800 GMT is more static and reflects a return to a state of equilibrium between land and sea.

Figure 66 presents plan views of the simulated potential temperature fields at the 300m level. These diagrams show clearly the significant offshore displacement of the temperature gradient. At 0900 GMT a typical land breeze situation is present, with higher temperatures over the sea. As in the previous case study, a seaward distortion in the temperature field has developed in response to the simulated bay region. By 1200 GMT the temperature gradient has reversed to that typical for a sea breeze regime; the bay region has now caused a landward
distortion which is reflected in all isotherms over the seaward part of the domain. At 1500 GMT the strong temperature gradient remains largely offshore and the bay-induced wave has reduced in amplitude. This lack of any landward efflux of cooler air has serious repercussions on the intensity and range of the sea breeze front. The final diagram for 1800 GMT shows a relatively unperturbed situation and a return to an isothermal state at the surface.

As mentioned above, the lack of inland penetration is of great importance in terms of the sea breeze frontal structure. Figure 67 presents the vertical velocity distribution and in general terms shows the restricted, but intense frontal zone which develops under these conditions. At 1200 GMT the typical bimodal updraught distribution has been initiated at either side of the simulated bay. This situation increases in intensity, with maximum updraughts around 1500 GMT of 5.6 and 6.4 cm s\(^{-1}\) to the west and east of the bay respectively. The downdraught velocity at this stage is approximately 2.4 cm s\(^{-1}\) at some 24Km out to sea. The situation at 1800 GMT is less clearly defined, with a reduction in the velocities, however, the remnants of the cellular distribution remain. Hope-Hislop (1974) noted that sea breezes will develop in the Plymouth region with offshore gradient winds of up to 8.0 ms\(^{-1}\), however, the numerical results show a lower critical level of around 4.0 ms\(^{-1}\) as being sufficient to seriously suppress any inland penetration. This limitation in the model may be due to the simplistic approach adopted to calculate a synoptic scale flow value typical for a depth of 3Km.

6.4.4 Discussion

The same criteria as set out for Case Study 1 are employed in Table 6.4 which summarises the comparative data for the sea breeze of 9th May 1984.
### Table 6.4: Comparison Data for Simulated and Observed Sea Breezes: Case Study 2 - 9th May

<table>
<thead>
<tr>
<th></th>
<th>Observed</th>
<th>Simulated</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Time of Onset (GMT)</strong></td>
<td>09:00</td>
<td>10:15</td>
</tr>
<tr>
<td><strong>Mean Maximum Velocity (m/s)</strong></td>
<td>5.0</td>
<td>2.6</td>
</tr>
<tr>
<td><strong>Time of Maximum Velocity (GMT)</strong></td>
<td>1400</td>
<td>1500</td>
</tr>
<tr>
<td><strong>Mean Sea Breeze Depth (m)</strong></td>
<td>300</td>
<td>600</td>
</tr>
<tr>
<td><strong>Time of Offset (GMT)</strong></td>
<td>16:45</td>
<td>17:00</td>
</tr>
<tr>
<td><strong>Average Frontal Propagation Rate (Km/hr^1)</strong></td>
<td>UNKNOWN</td>
<td>-</td>
</tr>
<tr>
<td><strong>Total Horizontal Extent (Km)</strong></td>
<td>UNKNOWN</td>
<td>10 Km inland &gt;32 Km offshore</td>
</tr>
</tbody>
</table>

**Evidence of Temporal Change in Depth of Sea Breeze**
- From 200m at 10:30 GMT to 400m at 11:00 GMT
- From 500m at 1200 GMT to 800m at 1500 GMT

**Evidence of Topographical Constraints**
- Unknown
- Bimodal updraught distribution and curvilinear temperature distribution

The onset times are in slightly better agreement than in the previous case study; however, the reservations involving the forcing function of the model and the grid cell structure are still valid. Although the offset times also compare favourably, the same difficulties with this and other temporal features make definitive conclusions difficult.
The strength and dimensions of the system in this simulation are strongly
controlled by the offshore gradient wind of 4.19 ms\(^{-1}\). The resulting sea breeze
is a much weaker affair with a very restricted inland spread. The mean
horizontal velocity of 2.6 ms\(^{-1}\) is approximately 50% of that observed. These
features clearly illustrate the severe constraints exerted by the synoptic scale
flow; it would seem that an offshore flow exceeding 4.0 ms\(^{-1}\) is a critical level
in damping the realistic development of the disturbance. Another aspect which
is probably of vital importance is the role of topography; several workers,
including Sumner (1977) and Atkinson (1981) have discussed the observational
aspects and the former gave a detailed assessment of the steering effect of
coastal hills and valleys in the Cardigan Bay area of North Wales. It is certain
that the influence of Dartmoor on the sea breeze is a vital controlling factor
and a suitable representation of this should be sought. Recommendations on
how this may best be achieved will be given in the concluding chapter.

Because of the aforementioned restriction in the inland extent of the system,
details on frontal propagation are difficult to ascertain. Despite this, the severe
disruption of the flow patterns by the complex coastal configuration is still a
prominent feature. Clearly a more detailed observational study of these
perturbations is necessary to confirm their significance.

With regard to the vertical variation of the system, the grid resolution is again a
problem. It appears that only a more careful consideration of the boundary
layer physics, possibly involving some sort of turbulence modelling is required.
This aspect will also be discussed in the concluding chapter.
CHAPTER 7 - CONCLUSIONS AND RECOMMENDATIONS

7.1 Conclusions

The research outlined concerned the development of a three-dimensional primitive equation model for the purposes of simulating the sea breeze for the Plymouth locale. The results presented have taken the format of general sensitivity studies of the three-dimensional scheme and a related two-dimensional version, followed by the application of the former to two particular sea breeze events during August 1983 and May 1984.

The details of the two-dimensional tests are summarised in Table 7.1. This is intended to describe the general trends associated with selected changes in the major factors determining the strength and dimensions of the sea breeze system. The increase in atmospheric stability results in a sea breeze circulation which is less vigorous and of reduced size and duration. This type of development has been noted in several observational studies (Pearce (1968); Wallington (1977)).

The influence of an increase in the surface heat flux shows an opposite effect, with a more vigorous system of greater horizontal and vertical extent and with higher velocities. The importance of the magnitude of the surface heat flux has been noted in several numerical analyses (Estoque (1961); Anthes, Warner and Seaman (1979); McCumber and Pielke (1981)). The geostrophic wind test is summarised in the table as a comparison with an equivalent calm situation. For the onshore flow, the sea breeze system is somewhat weaker in terms of velocity levels; the dimensions, however, are relatively unchanged, with a slightly later onset and an increased frontal propagation rate. In contrast, for the offshore flow the circulation is deeper and more vigorous; in addition, the onset time is
considerably later and the frontal propagation rate reduced. The final two-dimensional test shows results in accordance with the second test on the heat flux parameterisation, with a more vigorous circulation developing for the June simulation.

<table>
<thead>
<tr>
<th>STABILITY</th>
<th>HEAT FLUX</th>
<th>( V_g )</th>
<th>( V_g )</th>
<th>HEAT</th>
<th>HEAT</th>
</tr>
</thead>
<tbody>
<tr>
<td>INCREASE</td>
<td>INCREASE</td>
<td>ONSHORE</td>
<td>OFFSHORE</td>
<td>FLUX</td>
<td>FLUX</td>
</tr>
<tr>
<td>DEPTH OF CIRCULATION</td>
<td>-</td>
<td>+</td>
<td>0</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>TIME OF ONSET</td>
<td>+</td>
<td>-</td>
<td>+</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>HORIZONTAL WIND SPEED</td>
<td>-</td>
<td>+</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>UPDRAUGHT VELOCITY</td>
<td>-</td>
<td>+</td>
<td>-</td>
<td>+</td>
<td>-</td>
</tr>
<tr>
<td>FRONTAL PROPAGATION RATE</td>
<td>-</td>
<td>+</td>
<td>+</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

**KEY**

+ : REFERS TO INCREASING  
- : REFERS TO DECREASING

**TABLE 7.1: SUMMARY OF THE TWO-DIMENSIONAL SENSITIVITY TESTS**

The three-dimensional tests were concerned with two particular aspects; firstly, the influence of the synoptic scale flow and, secondly, the influence of coastal indentation. The results from the first test are summarised in Table 7.2. As in the two-dimensional analysis, the offshore synoptic flow produces the most vigorous circulation, with a reduced frontal propagation rate. The test on the coastal indentation showed the development of an asymmetric vertical velocity
distribution around the simulated bay. This is argued as being a consequence of the landward distortion in the temperature gradient and the combination of the Coriolis force acting in concert to the east of the bay and in opposition to the west. These conclusions are in accordance with other research involving complex coastlines (McPherson (1971); Saito (1976)).

<table>
<thead>
<tr>
<th></th>
<th>( V_g ) ONSHORE</th>
<th>( V_g ) OFFSHORE</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEPTH OF CIRCULATION</td>
<td>0</td>
<td>+</td>
</tr>
<tr>
<td>TIME OF ONSET</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>HORIZONTAL WIND SPEED</td>
<td>-</td>
<td>+</td>
</tr>
<tr>
<td>UPDRAUGHT VELOCITY</td>
<td>-</td>
<td>+</td>
</tr>
<tr>
<td>FRONTAL PROPOGATION RATE</td>
<td>+</td>
<td>-</td>
</tr>
</tbody>
</table>

**KEY**

+ : REFERS TO INCREASING  
- : REFERS TO DECREASING

**TABLE 7.2: SUMMARY OF THE THREE-DIMENSIONAL SENSITIVITY TEST FOR THE SYNOPTIC SCALE FLOW**

In Chapter 6 two case studies employing the three-dimensional model were presented, helping to identify the key problem areas. Firstly, the numerical model predicts times which are later than those observed in both case studies. This is argued as being related to a possible grid-dependency in the model and the rather smooth heating function employed. The predicted velocities in both case studies are underestimates, which is probably a reflection of the
time-averaged nature of the results as compared to the highly transient and turbulent nature of the observed sea breeze. Comparison of depth data in the two case studies is also considered problematic due to the resolution of the vertical grid which will only allow an accuracy of ±150m. Despite this, the model does demonstrate a characteristic temporal change in the depth of the circulation. The frontal propagation rate simulated for Case Study 1 is in reasonable agreement with the available general data (e.g. Simpson et al (1976)), however, Case Study 2 illustrates a severe, and probably unrealistic, influence on the mesoscale disturbance by the stronger geostrophic component. The restriction in the inland spread to less than 10Km suggests that the critical value for the synoptic scale component is approximately 4m s⁻¹. This is at variance with the general conclusion of Hope-Hislop (1974), who quoted a figure of 8m s⁻¹. It is clear that further analysis, both theoretical and experimental, of this aspect is required. More observations are also required to elucidate the horizontal extent of the system, not only inland but also seaward. On the more positive side the model does predict an asymmetric updraught distribution in response to the divergent/convergent wind field created by the simulated bay. This type of disturbance was noted by Hope-Hislop (1974), however, clearer observational verification should be sought.

7.2 Discussion of Conclusions

The two-dimensional model has successfully reproduced the major features of the sea breeze for a theoretical straight coastline. The sensitivity tests have demonstrated a model response in accordance with observations and other numerical analyses of the sea breeze.
The three-dimensional model has successfully reproduced the major features of the sea breeze for a straight and an indented coastline with no topographic variations. The model sensitivity is shown to be in accordance with available observations and other numerical analyses.

The two presented case studies demonstrate several key areas where the model construction requires improvement and illustrates the difficulties involved in the numerical prediction of mesoscale phenomenon. The major areas of concern are:

- The lack of topographic variations overland.
- The poor grid resolution in the vertical in the lowest layers.
- The relatively simple heat flux parameterisation and transfer mechanism employed.
- The neglect of the turbulent interactions in the boundary layer.

With the drawbacks outlined above it is difficult to make a proper assessment of the influence of the Plymouth sea breeze on thunderstorms over Dartmoor and the dispersal of pollutants. However, it is clear that the presence of a bay like Plymouth Sound results in a double convergence zone with larger updraught velocities than would otherwise be the case. The efflux of the cooler, moist air into these zones would suggest preferential locations for convective shower development. This, in conjunction with anabatic effects over the moors, may result in enhancement of pre-existing showers. In terms of pollutant dispersal, the classic fumigation problem is a possibility, however, the complex coastal configuration and orography make a definitive conclusion impossible at this stage.
7.3 Recommendations

* The most urgent requirement is the incorporation of topographic variations in a full sigma coordinate derivation in both the two-dimensional and three-dimensional models. Some preliminary tests on the two-dimensional scheme using a blockage procedure to simulate a plain, slope and plateau proved difficult to stabilise, however, it is clear that this remains one of the major controls on the surface wind fields and associated convergence zones.

* The heat flux parameterisation of Wood (1977a) should be coupled more closely with the models to allow easier initialisation. The surface heat balance approach appears to be the most appropriate, but a study is required to improve the assumptions employed in regions where the terrain is complex.

* The heat and momentum transfer in the boundary layer is currently carried out without the explicit use of diffusion coefficients. Although this has important consequences on computational efficiency, a more realistic closure of the turbulent transfer processes should be sought. The most common procedure involves K-theory; however, other possibilities which may be worth pursuing involve one- or two-equation turbulence modelling (Launder and Spalding (1974)) and algebraic-stress modelling (Donaldson (1973)).

* More numerical studies are also necessary to improve the grid resolution in the boundary layer. At present it is difficult to predict the finer details of the sea breeze circulation with any great accuracy.
Commensurate with the above numerical details, a new observational initiative is also urgently required. This should focus mainly on the three-dimensional structure of the Plymouth sea breeze and in particular the temperature and moisture fields. Additional data on other features such as the seaward extent of the sea breeze, the possible location of bay induced convergence zones and anabatic effects of Dartmoor would also be of considerable use.

With these data available, more effort should be directed towards the effects of the sea breeze on pollutant dispersal and thunderstorm enhancement. Numerically, this could take the form of a joint sea breeze/cloud model with complete moist thermodynamics to analyse the cloud development over the moor. The pollutant problem, although not serious for Plymouth, should be quantitatively assessed by the application of trajectory or particle tracer techniques in the numerical models.

A further benefit with more available data will be improved verification of the numerical scheme. Only with an increased effort in this direction can the local forecasting potential of a numerical model of the Plymouth sea breeze attain a suitable level of credibility.
NOMENCLATURE

B = Initial potential temperature stratification.
CC = Heat flux constant.
Cp = Specific heat of air at STP.
f = $2 \Omega \sin \lambda$.
$\tilde{f}$ = $2 \Omega \cos \lambda$.
F~ = Frictional forces.
g = Acceleration due to gravity.
h = Convective layer height.
H = Height of material surface.
i, j, k = Grid number positions.
j, j, k = Unit vectors.
k = von Karman's constant.
K = Transfer coefficient.
P = Pressure.
$P_0$ = Reference pressure.
$Q_H$ = Sensible heat flux.
$Q^*$ = Surface heat flux.
R = Gas constant for dry air.
Ri = Richardson number.
S = Solar constant.
t = Time.
u, v, w = Velocity components in Cartesian space.
Ug, Vg = Geostrophic components in x-y directions.
$U_2$ = Windspeed at height $Z_2$.
V~ = Velocity vector.
\[ x, y, z = \text{Cartesian coordinate directions.} \]

\[ Z_0 = \text{Roughness length.} \]

\[ Z_2 = \text{Logarithmic profile depth.} \]

\[ \rho = \text{Density.} \]

\[ \nabla = \text{Del operator} \quad = i \frac{d}{dx} + j \frac{d}{dy} + k \frac{d}{dz} \]

\[ \Delta x, \Delta y, \Delta z = \text{Grid steps.} \]

\[ \Delta t = \text{Timestep.} \]

\[ \theta = \text{Potential temperature.} \]

\[ \pi = \text{Exner function} \quad = \frac{C_P(P/P_0)^\kappa}. \]

\[ \kappa = \frac{R}{C_p}. \]

\[ \lambda = \text{Latitude.} \]

\[ \Omega = \text{Angular velocity of earth.} \]

\[ \delta \theta = \text{Diabatically perturbed potential temperature.} \]

\[ \psi = \text{Stability function.} \]

\[ \beta = \text{Bowen ratio.} \]

\[ \lambda E = \text{Latent heat flux.} \]
9. REFERENCES


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Zambakas, J. D. 1973: The diurnal variation and duration of the sea-breeze at the National Observatory of Athens, Greece. Met. Mag., 102, 224-228.
FIGURE 1: SCHEMATIC REPRESENTATION OF A SEA BREEZE WHEN THE SYNOPTIC SCALE WIND IS LIGHT

(Source: Munn (1971) and Oke (1978))
FIGURE 3: DIABATIC HEAT FLUX PROFILES USED IN THE FORCING FUNCTION OF THE MODEL
FIGURE 4: MODEL DOMAINS IN (A) TWO-DIMENSIONAL MODEL; AND (B) THREE-DIMENSIONAL MODEL
FIGURE 5: THE STAGGERED GRID MESHES USED IN THE (A) TWO-DIMENSIONAL NUMERICAL SCHEME; AND (B) THREE-DIMENSIONAL NUMERICAL SCHEME
FIGURE 6: GRID NUMBERS AND POSITIONS IN THE (A) TWO-DIMENSIONAL MESH; AND (B) THREE-DIMENSIONAL MESH
Figure 7: Schematic representation of how the physical solution of the interaction between two waves decomposes into an erroneous aliased solution.

(Source: Pielke (1984))
INITIAL STRATIFICATION = 1.00 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 8: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS
SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 3.00 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 9: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 3.00 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 10: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
FIGURE 11: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FRONTAL PROPOGATION RATES

FIGURE 12: FRONTAL DISPLACEMENT INLAND AS A FUNCTION OF TIME AND INITIAL STRATIFICATION
INITIAL STRATIFICATION = 3.00 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 M/S

HORIZONTAL WIND PROFILES ON THE COAST

--- 3 HRS
--- 6 HRS
--- 9 HRS
--- 12 HRS

FIGURE 13: HORIZONTAL WIND PROFILES FOR 3 TO 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 3.00 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 M/S

FIGURE 14: POTENTIAL TEMPERATURE PROFILES FOR 0 TO 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.025
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 15: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
FIGURE 16: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.025
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 17: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 18: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 M/S

FRONTAL PROPAGATION RATES

\[ \text{FRONTAL DISPLACEMENT INLAND AS A FUNCTION OF TIME AND AMPLITUDE OF HEATING FUNCTION} \]

FIGURE 19
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 2.00 m/s

FIGURE 20: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = 2.00 m/s

FIGURE 21: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
FIGURE 22: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
OT = 600 S
GEOSTROPHIC COMPONENT = -2.00 m/s

FIGURE 23: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
**INITIAL STRATIFICATION** = 2.50 \( \text{K/Km} \)

\( CC = 0.1 \)

\( DT = 600 \text{ S} \)

**GEOSTROPHIC COMPONENT** = -2.00 \( \text{m/s} \)

**FIGURE 24:** VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 600 S
GEOSTROPHIC COMPONENT = -2.00 m/s

FIGURE 25: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 600 S

FRONTAL PROPAGATION RATES

FIGURE 26: FRONTAL DISPLACEMENT INLAND AS A FUNCTION OF TIME AND SPEED AND DIRECTION OF Vg
FIGURE 27: SENSIBLE HEAT FLUX PROFILES AS CALCULATED USING THE SURFACE CLIMATE MODEL OF WOOD (1977a)
FIGURE 28: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/km
SENSIBLE HEAT FLUX FOR A TYPICAL DAY IN JUNE
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 29: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
SENSIBLE HEAT FLUX FOR A TYPICAL DAY IN MARCH
DT = 600 S
GEOSTROPHIC COMPONENT = 0.80 M/S

FIGURE 30: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
SENSIBLE HEAT FLUX FOR A TYPICAL DAY IN JUNE
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 31: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE TWO-DIMENSIONAL MODEL
INITIAL STRATIFICATION = 2.50 K/Km
DT = 600 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FRONTAL PROPAGATION RATES

FIGURE 32: FRONTAL DISPLACEMENT INLAND AS A FUNCTION OF TIME AND THE AMPLITUDE OF THE HEATING FUNCTION FROM SURFACE CLIMATE MODEL OF WOOD (1977a)
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 M
GEOSTROPHIC COMPONENT = 0.00 M/S

FIGURE 33: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 34: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9, AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE
FIGURE 35: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 m
GEOSTROPHIC COMPONENT = 2.00 m/s NORTHERLY

FIGURE 36: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN OFFSHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120S
GEOSTROPHIC COMPONENT = 2.00 m/s NORTHERLY

FIGURE 37: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN OFFSHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
GEOSTROPHIC COMPONENT = 2.00 m/s NORTHERLY

FIGURE 38: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN OFFSHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
GEOSTROPHIC COMPONENT = 2.00 m/s NORTHERLY

FIGURE 39: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN OFFSHORE GEOSTROPHIC COMPONENT
**Figure 40**: Horizontal wind fields for 3, 6, 9 and 12 hours simulated time in the three-dimensional model with a straight coastline and an onshore geostrophic component.
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
GEOSTROPHIC COMPONENT = 2.00 m/s SOUTHERLY

FIGURE 41: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN ONSHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
GEOSTROPHIC COMPONENT = 2.00 m/s SOUTHERLY

FIGURE 42: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN ONSHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 M
GEOSTROPHIC COMPONENT = 2.00 M/S EASTERLY

FIGURE 43: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN ALONG SHORE GEOSTROPHIC COMPONENT

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**INITIAL STRATIFICATION = 2.50 K/Km**

CC = 0.1

DT = 120 S

Z = 300 m

**GEOSTROPHIC COMPONENT = 2.00 M/S WESTERLY**

**HORIZONTAL WIND FIELD VECTORS (M/S)**

**FIGURE 44:** HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH A STRAIGHT COASTLINE AND AN ALONG SHORE GEOSTROPHIC COMPONENT
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 M
GEOSTROPHIC COMPONENT = 0.00 M/S

FIGURE 45: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 M
GEOSTROPHIC COMPONENT = 0.00 M/S

HORIZONTAL WIND FIELD VECTORS (M/S)

FIGURE 46: HORIZONTAL WIND FIELDS FOR 3 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 300 m
GEOSTROPHIC COMPONENT = 0.00 M/S

FIGURE 47: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
CC = 0.1
DT = 120 S
Z = 600 M
GEOSTROPHIC COMPONENT = 0.00 M/S

FIGURE 48: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE
INITIAL STRATIFICATION = 2.50 K/Km
HEAT FLUX FOR A TYPICAL DAY IN MARCH
\( \Delta T = 90 \) S
\( Z = 300 \) M
GEOSTROPHIC COMPONENT = 0.00 M/S

**FIGURE 49:** HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR HEAT FLUX FROM SURFACE CLIMATE MODEL OF WOOD (1977a)
INITIAL STRATIFICATION = 2.50 K/Km
HEAT FLUX FOR A TYPICAL DAY IN MARCH
\( DT = 90 \text{ s} \)
\( Z = 300 \text{ m} \)
GEOSTROPHIC COMPONENT = 0.00 m/s

FIGURE 50: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR HEAT FLUX FROM SURFACE CLIMATE MODEL OF WOOD (1977a)
**INITIAL STRATIFICATION = 2.50 K/Km**

**HEAT FLUX FOR A TYPICAL DAY IN MARCH**

**DT = 90 S**

**Z = 600 M**

**GEOSTROPHIC COMPONENT = 0.00 m/s**

**FIGURE 51:** VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR HEAT FLUX FROM SURFACE CLIMATE MODEL OF WOOD (1977a)
FIGURE 52: CASE STUDY 1: GENERAL SYNOPTIC SITUATION AT 1200 GMT ON 11th AUGUST 1983
FIGURE 53: CASE STUDY 1: TEPHIGRAM FOR M.O. CAMBORNE AT 1200 GMT ON THE 11TH AUGUST 1983
FIGURE 54: HORIZONTAL WIND PROFILES FROM PILOT BALLOON ASCENTS AT PLYMOUTH POLYTECHNIC ON 11TH AUGUST 1983
INITIAL STRATIFICATION = 2.70 K/km
CASE STUDY 1: 11th AUGUST 1983
DT = 90 S
GEOSTROPHIC COMPONENT = 2.47 m/s - 313°

FIGURE 55: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 1: 11th AUGUST 1983
INITIAL STRATIFICATION = 2.70 K/Km

CASE STUDY 1: 11th AUGUST 1983
DT = 90 S
GEOSTROPHIC COMPONENT = 2.47 m/s - 313°

SOUTHERLY COMPONENTS IN SEA BREEZE AT THE COAST

(A) 03:00 HRS
(B) 06:00 HRS
(C) 09:00 HRS
(D) 12:00 HRS

FIGURE 56: HORIZONTAL WIND PROFILES FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 1: 11th AUGUST 1983
**INITIAL STRATIFICATION = 2.70 K/Km**
**CASE STUDY 1: 11th AUGUST 1983**
**DT = 90 S**
**GEOSTROPHIC COMPONENT = 2.47 m/s - 343°**

**FIGURE 57:** POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 1: 11th AUGUST 1983
INITIAL STRATIFICATION = 2.70 K/Km
CASE STUDY 1: 11th AUGUST 1983
DT = 90.5
Z = 300 M
GEOSTROPHIC COMPONENT = 2.17 m/s - 343°

FIGURE 58: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 1: 11 AUGUST 1983
FIGURE 59: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 1: 11th AUGUST 1983
FIGURE 60: CASE STUDY 2: GENERAL SYNOPTIC SITUATION AT 1200 GMT ON 9th MAY 1984
FIGURE 61: CASE STUDY 2: TEPHIGRAM FOR M.O. CAMBORNE AT 1200 GMT ON 9TH MAY 1984
FIGURE 62: HORIZONTAL WIND PROFILES FROM PILOT BALLOON ASCENTS AT PLYMOUTH POLYTECHNIC ON 9th MAY 1984
INITIAL STRATIFICATION = 3.70 K/Km
CASE STUDY 2: 9th MAY 1984
DT = 90 s
Z = 300 m
GEOSTROPHIC COMPONENT = 1.19 m/s - 025°

FIGURE 63: HORIZONTAL WIND FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 2: 9th MAY 1984
INITIAL STRATIFICATION = 3.70 K/Km
CASE STUDY 2: 9TH MAY 1984
DT = 90 S
GEOSTROPHIC COMPONENT = 4.19 m/s - 025°

SOUTHERLY COMPONENTS IN SEA BREEZE AT THE COAST

FIGURE 64: HORIZONTAL WIND PROFILES FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 2: 9TH MAY 1984
INITIAL STRATIFICATION = 3.70 K/Km
CASE STUDY 2: 9th MAY 1984
OT = 90 S
Z = 300 m
GEOSTROPHIC COMPONENT = 1.19 m/s - 025°

FIGURE 65: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 2: 9th MAY 1984
INITIAL STRATIFICATION = 3.70 K/km
CASE STUDY 2: 9th MAY 1984
OT = 90S
Z = 300 m
GEOSTROPHIC COMPONENT = 1.19 m/s - 025°

FIGURE 66: POTENTIAL TEMPERATURE FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 2: 9th MAY 1984
INITIAL STRATIFICATION = 3.78 K/Km
CASE STUDY 2: 9th MAY 1984
DT = 90 S
Z = 300 M
GEOSTROPHIC COMPONENT = 4.19 M/S - 025°

FIGURE 67: VERTICAL VELOCITY FIELDS FOR 3, 6, 9 AND 12 HOURS SIMULATED TIME IN THE THREE-DIMENSIONAL MODEL WITH AN INDENTED COASTLINE AND FOR CASE STUDY 2: 9th MAY 1984
APPENDIX A1

THE MODEL EQUATIONS
APPENDIX A1

Three-Dimensional Model Equations

Conservation of Mass

\[
\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0 \quad (A1-1).
\]

Conservation of Motion

\[
\frac{\partial u'}{\partial t} = - (U_g + u') \frac{\partial u'}{\partial x} - (V_g + v') \frac{\partial u'}{\partial y} - w' \frac{\partial}{\partial z} (U_g + u')
\]

\[- f \frac{\partial}{\partial \theta} \frac{\partial}{\partial x} + \nu' \frac{\partial}{\partial x} + f v' - f w' \quad (A1-2).
\]

\[
\frac{\partial v'}{\partial t} = - (U_g + u') \frac{\partial v'}{\partial x} - (V_g + v') \frac{\partial v'}{\partial y} - w' \frac{\partial}{\partial z} (V_g + v')
\]

\[+ f \frac{\partial}{\partial \theta} \frac{\partial}{\partial y} + \nu' \frac{\partial}{\partial y} - f u' \quad (A1-3).
\]

\[
\frac{\partial w'}{\partial z} = \frac{g}{\theta^2} \quad (A1-4).
\]
Conservation of Heat

\[
\frac{\partial \theta'}{\partial t} = - (U_g + u') \frac{\partial \theta'}{\partial x} - (V_g + v') \frac{\partial \theta'}{\partial y} - w' \frac{\partial}{\partial z} (\theta' + \bar{\theta}) + \delta \theta
\]  
(A1-5).

Material Surface Equation

\[
\frac{\partial \bar{H}}{\partial t} = - (U_g + u') \frac{\partial \bar{H}}{\partial x} - (V_g + v') \frac{\partial \bar{H}}{\partial y} + W_H'
\]  
(A1-6).

Two-Dimensional Model Equations

Conservation of Mass

\[
\frac{\partial \bar{v}'}{\partial y} + \frac{\partial \bar{w}'}{\partial z} = 0
\]  
(A1-7)

Conservation of Motion

\[
\frac{\partial \bar{u}'}{\partial t} = - (V_g + v') \frac{\partial \bar{u}'}{\partial y} - w' \frac{\partial}{\partial z} (U_g + u')
\]
\[-f \frac{\partial}{\partial \theta} v_g - \tilde{\theta} \frac{\partial \tilde{x}'}{\partial x} + f v' - \tilde{f} w' \quad (A1-8).\]

\[\frac{\partial v'}{\partial t} = - (V_g + v') \frac{\partial v'}{\partial y} - w' \frac{\partial}{\partial z} (V_g + v') - fu' - \tilde{\theta} \frac{\partial \tilde{x}'}{\partial y} \quad (A1-9).\]

\[\frac{\partial \tilde{x}'}{\partial z} = \frac{g}{\tilde{\theta}^2} \quad (A1-10).\]

**Conservation of Heat**

\[\frac{\partial \theta'}{\partial t} = - (V_g + v') \frac{\partial \theta'}{\partial y} - w' \frac{\partial}{\partial z} (\theta' + \tilde{\theta}) + \delta \theta \quad (A1-11).\]

**Material Surface Equation**

\[\frac{\partial H}{\partial t} = - (V_g + v') \frac{\partial H}{\partial y} + W_H \quad (A1-12).\]
APPENDIX A2

THE FINITE-DIFFERENCE APPROXIMATIONS
APPENDIX A2

Three-Dimensional Finite-Difference Equations

Conservation of Mass

\[ w_{i+1, j+1, k+2}^{n+1} - w_{i+1, j+1, k}^{n+1} = \Delta z \left[ \frac{u_{i+2, j+1, k+1}^{n+1} - u_{i+1, j+1, k+1}^{n+1}}{\Delta x} \right] \]

\[ + \frac{v_{i+1, j+2, k+1}^{n+1} - v_{i+1, j+1, k+1}^{n+1}}{\Delta y} \]

(A2-1).

Conservation of Motion

\[ u_{i, j, k}^{n+1} = u_{i, j, k}^{n} - \Delta t \left[ u_{i, j, k}^{n} \left( \frac{u_{i, j, k}^{n} - u_{i, j, -2}^{n}}{\Delta x} \right) \right] \]

\[ + \frac{v_{i, j, k}^{n} (u_{i, j, k}^{n} - u_{i, j, -2}^{n})}{\Delta y} + \frac{w_{i, j, k}^{n} (u_{i, j, k}^{n} - u_{i, j, -2}^{n})}{\Delta z} \]

\[ - f \left( \frac{\theta_{i, j, k}^{n}}{\theta_{i, j, k}^{n}} \right) v_{g} - \frac{v_{i, j, k}^{n} (\pi_{i+2, j, k}^{n} - \pi_{i, j, -2}^{n})}{2\Delta x} + f v_{i, j, k}^{n} - f w_{i, j, k}^{n} \]

(A2-2).

\[ v_{i, j, k}^{n+1} = v_{i, j, k}^{n} - \Delta t \left[ v_{i, j, k}^{n} \left( \frac{v_{i, j, k}^{n} - v_{i, j, -2}^{n}}{\Delta x} \right) \right] \]

\[ + \frac{v_{i, j, k}^{n} (v_{i, j, k}^{n} - v_{i, j, -2}^{n})}{\Delta y} + \frac{w_{i, j, k}^{n} (v_{i, j, k}^{n} - v_{i, j, -2}^{n})}{\Delta z} \]

A2-1
\[- f \left( \frac{\phi_{n, ik}}{\phi_{i, jk}} \right) u_g + \partial_{i, jk} \left( \frac{\pi_{i+2, ik} - \pi_{i-2, ik}}{2\Delta y} \right) + f u_{n+1} \right] \quad (A2-3). \]

\[\pi_{i, jk+2}^{n+1} = \pi_{i, jk}^{n+1} + \Delta z \cdot g \left( \frac{\phi_{i, jk+1}}{\phi_{i, jk+1}} \right)^2 \quad (A2-4). \]

\[\Delta \text{Conservation of Heat} \]

\[\theta_{i, jk}^{n+1} = \theta_{i, jk}^n - \Delta t \left[ u_{i, jk} \left( \theta_{i, i, ik}^{n+1} - \theta_{i, i, 2ik}^{n} \right) + v_{i, jk} \left( \frac{\theta_{i, ik}^{n} - \theta_{i, i, 2ik}^{n}}{\Delta y} \right) \right. \]
\[\left. + w_{i, jk} \left( \frac{\theta_{i, ik}^{n} - \theta_{i, ik-2}^{n}}{\Delta z} \right) \right] + \delta_{i, jk} \theta_{i, jk}^n \quad (A2-5). \]

\[\Delta \text{Material Surface Equation} \]

\[\Delta \text{Two-Dimensional Finite-Difference Equations} \]

\[\Delta \text{Conservation of Mass} \]

A2-2
\[ w_{j+1k+2}^{n+1} = w_{j+1}^{n+1} - \Delta z \left[ \frac{v_{j+1k+1}^{n+1} - v_{jk+1}^{n+1}}{\Delta y} \right] \] (A2-7).

**Conservation of Motion**

\[ u_{jk}^{n+1} = u_{jk}^{n} - \Delta t \left[ v_{jk}^{n} \left( \frac{u_{jk}^{n} - u_{i-2k}^{n}}{\Delta y} \right) \right. \]
\[ \left. + w_{jk}^{n} \left( \frac{u_{jk}^{n} - u_{i-2k}^{n}}{\Delta z} \right) + f \left( \frac{\theta_{jk}^{n}}{\theta_{jk}} \right) \right] \] (A2-8).

\[ v_{jk}^{n+1} = v_{jk}^{n} - \Delta t \left[ v_{jk}^{n} \left( \frac{v_{jk}^{n} - v_{i-2k}^{n}}{\Delta y} \right) + w_{jk}^{n} \left( \frac{v_{jk}^{n} - v_{i-2k}^{n}}{\Delta z} \right) \right. \]
\[ \left. + \theta_{jk}^{n} \left( \frac{z_{i+2k}^{n} - z_{i-2k}^{n}}{\Delta y} \right) + f \left( \frac{u_{jk}^{n+1}}{u_{jk}} \right) \right] \] (A2-9).

\[ z_{jk+2}^{n+1} = z_{jk}^{n+1} + \Delta z g \frac{\theta_{jk+1}^{n+1}}{\left( \theta_{jk+1} \right)^{2}} \] (A2-10).

**Conservation of Heat**

\[ \theta_{jk}^{n+1} = \theta_{jk}^{n} - \Delta t \left[ v_{jk}^{n} \left( \frac{\theta_{jk}^{n} - \theta_{i-2k}^{n}}{\Delta y} \right) \right] \]
\[ + w_{jk}^n \left( \frac{\theta_{ik}^n - \theta_{ik}^{n-2}}{\Delta z} \right) + \delta \theta_{jk}^n \] (A2-11).

**Material Surface Equation**

\[ H_j^{n+1} = H_j^n - \Delta t \left[ \nabla_{ij}^n \left( \frac{H_j^n - H_j^{n-2}}{\Delta y} \right) + w_j^{n+1} \right] \] (A2-12).
APPENDIX B

COMPUTATIONAL SEQUENCE AND MODEL PROGRAMS
START
OPEN DATA FILES
INPUT PHYSICAL CONSTANTS
INPUT GRID CONSTANTS
SET UP INITIAL STATE
CALL ECONST
TIME = 0
TIME = TIME + DT
CALCULATE DIABATIC HEAT FLUX
CALL FORCE
CALL ADVHW → CALL XGEN → CALL YGEN → CALL ZGEN
CALL FILTH
CALL VWIND
CALL MATGEN → CALL XGEN → CALL YGEN
CALL ADVPFL → CALL XGEN → CALL YGEN → CALL ZGEN
CALL FILTPT
CALL PGEN
TIME = 12 HRS ?
Y OUTPUT DATA CLOSE FILES
N
TIME = 3, 6 OR 9 HRS ?
Y OUTPUT DATA
STOP

COMPUTATIONAL SEQUENCE IN THE THREE-DIMENSIONAL MODEL
MODEL 2

DIMENSION V(0:62,0:32)
DIMENSION W(0:62,0:32)
DIMENSION Z(0:62,0:32)

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ PTE
COMMON /CB5/ PT,PTIN
COMMON /CB6/ OLDPTE
COMMON /CB7/ P
COMMON /CB8/ Z
COMMON /CB9/ S

REAL BB,CC,C,DELTA,DENS,DY,F,S,LAT,SOL,R,RA,STA,TMP,VG
REAL HOUR
INTEGER DT,JK,KW,YY,ZZ
INTEGER T,TIML

OPEN(5,FILE=•••>DATA>6.DAT•••
OPEN(6,FILE=•••>DATA>PT.20.DAT•••
OPEN(7,FILE=•••>DATA>Y.20.DAT•••
OPEN(8,FILE=•••>DATA>2U.DAT•••
C OPEN(11,FILE=•••>DATA>SURF.2D.DAT•••
C OPEN(12,FILE=•••>DATA>PRESS.2D.DAT•••

C— PHYSICAL CONSTANTS

BE=0.0025
STR=0.0043
CC=0.1
CP=1009.0

DENS=1.21
DELT=0.1
G=5.81
LAT=50.375
SOL=1370.3
R=7.29E-5
F=2.0*R*SIN(LAT+2.0*3.14159/360.0)
FEAP=2.0*K*COS(LAT+2.0*3.14159/360.0)
RA=287.0
TMP=286.0
VG=1.3

C— GRID SIZE CONSTANTS

DY=800.0
DT=600
TIME=43200
YY=32
Z2=20

DO 15 J=0,YY
  Z(J,0)=0.0
DO 12 K=1,Z2
  Z(J,K)=Z(J,K-1)*150.0
12 CONTINUE
S(K)=Z(J,Z2)

B-3
C____MODEL 2

15 CONTINUE

C____INITIAL AND BOUNDARY CONDITIONS
    DO 20 J=0,YY
    PT(J,0)=TEMP
    DO 10 K=1,ZZ
    PT(J,K)=PT(J,0)+20+Z(J,K)
10 CONTINUE
    DO 21 K=0,ZZ
    U(J,K)=0.0
    V(J,K)=0.0
    PTIN(K)=PT(J,K)
    PTET(J,K)=0.0
    OLDPTET(J,K)=0.0
    P(J,K)=0.0
    W(J,K)=0.0
21 CONTINUE
20 CONTINUE

C WRITE(5,503) (PT(22*K),K=1,150,0<=22*K<=2)

C____SET UP OF CONSTANTS FOR USE IN IMPLICIT FILTER
CALL FC0NST(YY,ZZ,DELTA)

C____TIME INTEGRATION
KNT=0
    DO 70 T=DT,TIML,DT
    HOURL=T/3600.0
    KNT=KNT+1
    IF(KNT.EQ.1)KNT=0

C____CALCULATION OF DIABATIC HEAT FLUX, C
30 READ(5,606)C
    C=CC*30*COS(LAT*2.0+3.14159/550.0)*SIN(R*T)

C____FORCING OF POTENTIAL TEMPS AND HORIZONTAL WIND FIELDS
CALL FORCE(CP,DEV,DT,DY,FPAR,0,VG,YY,ZZ)

C____CALCULATION OF ADEPTION TERMS FOR HORIZONTAL WIND FIELDS
CALL ADVHV(DT,DY,VG,YY,ZZ)

C____APPLICATION OF IMPLICIT FILTER TO HORIZONTAL WIND FIELD
IF(KNT.EQ.6)THLA
    CALL FILTH(YY,ZZ)
END IF

C____CALCULATION OF VERTICAL VELOCITY FIELD
CALL WIND(DY*YY*ZZ)

C_____CALCULATION OF MATERIAL SURFACE
CALL MAT6E(VI,DY,KG,YY*ZZ)

C_____CALCULATION OF JOINT CTMS FOR POT. TEMPR. FIELD
CALL AUVPT(DY,KG,YY*ZZ)

C_____APPLICATION OF IMPLICIT FILTER TO POTENTIAL TEMPRATURE FLD
IF(INIT. EQ. 0) THEN
  CALL FILTPT(YY,ZZ)
END IF

C_____CALCULATION OF PRESSURE FIELD
CALL PSN(SIR,YY,YY,ZZ)

C_____PRINT OUT TIMES
IF(T,1080.0 .LT. T .LE. 21 & T .GT. 0.0 .OR. T .GT. 43200) THEN

C_____OUTPUT DATA
C
WRITE(6,1111) HOUR
WRITE(6,500) ((DPIV(K)+PTE(J,K)),J=0,YY,2),K=0,ZZ,2) C
WRITE(7,2222) HOUR
WRITE(7,501) (1V(J,K)),J=0,YY,2),K=0,ZZ,2) C
WRITE(3,3333) HOUR
WRITE(3,502) ((X(J,K)+100.0),J=1,YY-1,2),K=0,ZZ,2) C
WRITE(11,4444) HOUR
WRITE(11,5555) HOUR
WRITE(12,5555) HOUR
WRITE(12,5555) HOUR
WRITE(5,503) (P1N(K)+P1L(Z2,K),K=156,0),K=0,ZZ,2) C
END IF

70 CONTINUE
CLOSE(5)
CLOSE(3)
CLOSE(7)
CLOSE(8)
C
CLOSE(11)
C
CLOSE(12)

STOP

C_____FORMAT STATEMENTS
500 FORMAT(1X,1/F7.2)
501 FORMAT(1X,1/F7.2)
SUBROUTINE CONST(YY,ZZ,DELTA)

DIMENSION APK(C:32),BK0(D:32),APKZ(L:32),EZ2Z(0:32)

DIMENSION Z(0:32)

INTEGER YY,ZZ,J,K
REAL SM,SP,DELTA

COMMON /C3S/?
COMMON /CB12/ SM,APK,BQK,APKZ,BQKZ

SM = 1.0-DELTA
SP=2.0*(1.0+DELTA)

C______Y - IMPLICIT FILTER CONSTANTS

APK(0)=1.0
BKX(1)=1.0

DO 700 J=2,YY
    APK(J)=SM*BK0(J-2)+SP
    BK0(J)=-3H/K/J
    CONTINUE

APK(YY)=BK0(YY-2)+1.0
BKX(YY)=0.0

C______Z - IMPLICIT FILTER CONSTANTS

APKZ(0)=1.0
BKXZ(0)=1.0

DO 710 K=2,ZZ
    APKZ(K)=SM*BKXZ(K-2)+SP
    BKXZ(K)=SM/APKZ(K)

CONTINUE

APKZ(ZZ)=BKXZ(ZZ-2)+1.0
BKXZ(ZZ)=0.0

RETURN

END
C____MODEL 2

END

SUBROUTINE FORCING(CP,DENS,DT,DY,F,FBAR,OG,VG,YY,ZZ)

DIMENSION P(0:62,0:32),PT(0:62,0:32)
DIMENSION PTE(0:62,0:32)
DIMENSION PTIN(:32)
DIMENSION U(0:62,0:32)
DIMENSION V(0:62,0:32)
DIMENSION Z(0:62,0:32),W(0:62,0:32)

COMMON /CB1/U
COMMON /CB2/V
COMMON /CB3/W
COMMON /CB4/PT
COMMON /CB5/PTE,PTIN
COMMON /CB7/P
COMMON /CB8/Z

INTEGER DT,J,K,YY,ZZ
REAL CP,DENS,DP,DY,F,FBAR,OG,VG

C____FORCING OF HORIZONTAL WIND FIELDS

DO 300 J=2,YY-2
DO 301 K=2,ZZ-2
DHDY=(P(J,K)-P(J-2,K))/DY
W(J,K)=(W(J+1,K)+W(J-1,K))/2.
U(J,K)=U(J-1,K)-0.5*V(J,K)*F*VG*W(J,K)
V(J,K)=V(J+1,K)-0.5*U(J,K)*F*VG*W(J,K)

301 CONTINUE
U(J,ZZ)=0.0
V(J,ZZ)=0.0

300 CONTINUE

DO 362 K=0,ZZ+2
U(0,K)=U(2,K)
V(0,K)=V(2,K)
U(YY,K)=U(YY-2,K)
V(YY,K)=V(YY-2,K)

362 CONTINUE

C____PARAMETERISATION OF HEAT CONVECTION IN ISENTROPIC C.L. OVERLAND

DO 52 J=2,YY-2
IF(J.LE.16)THEN
MM=1
IF(J.GE.MM+16)THEN
MM=MM+2
END IF

DPT=DTR/C*(C*DLS+Z(J,MM))
DIF(P(J,MM)*DPT)*G*PT(J,MM+2))GO TO 100

100 K=0,MM+2
PT(J,K)=PT(J,K)+DPT
PTE(J,K)=PTE(J,K)+DPT

110 CONTINUE

B-7
C__MODEL 2

END IF

RETURN

SUBROUTINE ADVH(UT, DY, VG, YY, ZZ)

DIMENSION U(3:62, 0:32)
DIMENSION V(0:62, 0:32)
DIMENSION Z(0:62, 0:32), W(0:62, 0:32)

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB5/ Z

INTEGER DT, J, K, YY, ZZ
REAL DY, VG
REAL U1, U2, U3, U4, U5, V1, V2, V3, V4, V5, W1, Z1, Z2, Z3
REAL & YADV, ZADV

C___CALCULATION OF ADVECTION TERMS FOR HORIZONTAL WIND FIELDS

DO 310 J=2*YY-2, 2
DO 320 K=2*ZZ-2, 2

U1 = U(J+K)
U4 = U(J+K+2)
U5 = U(J+K-2)
V1 = V(J+K)
V4 = V(J+K+2)
V5 = V(J+K-2)
W1 = W(J+K)
Z1 = Z(J+K)
Z2 = Z(J+K-2)
Z3 = Z(J+K+2)

CALL 7GLFJ(U1, U4, U5, V1, V4, V5, W1, Z1, Z2, Z3, YADV)
U(J, K) = U(J, K) - LT*SN(U1, U4, U5, V1, V4, V5, W1, Z1, Z2, Z3, YADV)

CALL 2GLFJ(V1, V4, V5, W1, VG, V6, V5, Z1, Z2, Z3, ZADV)
V(J, K) = V(J, K) - LT*SN(V1, V4, V5, W1, VG, V6, V5, Z1, Z2, Z3, ZADV)

310 CONTINUE

U(J, K) = U(J, K)
V(J, K) = V(J, K)

320 CONTINUE

DO 311 K=0, ZZ+2
U(J, K) = U(J, K)
V(J, K) = V(J, K)

311 CONTINUE

B-8
C____MODEL 2

U(YY, K) = U(YY-2, K)
V(YY, K) = V(YY-2, K)

311 CONTINUE

DO 330 K = 0, ZZ + 2
DO 340 J = 2*YY - 2 + 2

U1 = U(J, K)
U2 = U(J + 2, K)
U3 = U(J - 2, K)
V1 = V(J, K)
V2 = V(J + 2, K)
V3 = V(J - 2, K)

CALL YGEN(DY, VG, U1, U2, U3, V1, YADV)
U(J, K) = U(J, K) - DT*SGL(YADV)

CALL YGEN(DY, VG, V1, V2, V3, V1, YADV)
V(J, K) = V(J, K) - DT*SGL(YADV)

340 CONTINUE

V(0, K) = V(2, K)
V(YY, K) = V(YY-2, K)
U(0, K) = U(2, K)
U(YY, K) = U(YY-2, K)

330 CONTINUE

DO 370 J = 0, YY + 2
DO 380 K = 1, ZZ - 1 + 2

U(J, K) = (U(J, K + 1) + U(J, K - 1))/2
V(J, K) = (V(J, K + 1) + V(J, K - 1))/2

380 CONTINUE
370 CONTINUE

RETURN
END

SUBROUTINE FILTH(YY, ZZ)

DIMENSION U(0:KY+1:12)
DIMENSION V(0:KZ+1:12)
DIMENSION UUX(0:KZ+1:12), VXK(0:KZ+1:12)
DIMENSION APX(0:KZ+1:12), B(KZ+1:12)

INTEGER J, JJ, K, YY, ZZ
REAL SM

COMMON /CR1U/ U
COMMON /CR2V/ V
COMMON /CR3Z/ SM, APX, B(KZ), APKZ, B(KZ)

C____ APPLY IMPLICIT FILTER IN Y - DIRECTION
SUBROUTINE VWND(DY,YY,ZZ)

DIMENSION V(0,62,0,32)
DIMENSION W(0,62,0,32),Z(0,62,0,32)

COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ Z

INTEGER J,K,YY,ZZ
REAL DY,YY

DO 550 J=1,YY-1,2
DO 551 K=2,ZZ+2

DY = (V(J+1,K-1) - V(J-1,K-1))/DY

W(J,K) = W(J-2) - (Z(J,K) - Z(J,K-2))*DY

W(J,K) = (W(J-1,K) + W(J+1,K))/2.

551 CONTINUE
550 CONTINUE

DO 552 J=2*YY-2,2
DO 553 K=1,ZZ

U(J,K) = (U(J-1,K)*U(J+1,K))/2.

553 CONTINUE
552 CONTINUE

RETURN
END

SUBROUTINE YATLEN(DT, DY, VG, YY, ZZ)

DIMENSION SI(0,62),S2(0,62),S3(0,62),V1(0,62)

COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ Z

INTEGER DT, J, YY, ZZ
REAL DY, VG
REAL SI, S2, S3, V1
REAL*8 YADV

C____CALCULATION OF ADVECTION TERM

DO 110 J=2,YY-2

S1 = SI(J)
S2 = SI(J+2)
S3 = SI(J-2)
V1 = V(J,ZZ-2)

CALL YCHN(DY, VG, S1, S2, S3, V1, YADV)

B-11
C___MODEL 2

S(J) = S(J) - DT * SGL(YADV)

CONTINUE

C____INTEGRATION OF MATERIAL SURFACE PROGNOSTIC EQUATION

DO 413 J = 2, YY - 2
S(J) = S(J) + DT * 2(J, Z2)
CONTINUE

S(0) = S(2)
S(YY) = S(YY - 2)

DO 414 J = 1, YY - 1, 2
S(J) = (S(J+1) + S(J-1)) / 2.0
CONTINUE

DO 415 J = 0, YY
Z(J, Z2) = S(J)
CONTINUE

RETURN
END

SUBROUTINE ADV: T(DT, DT, VG, YY, Z2)

DIMENSION PTE(0:3, 0:32)
DIMENSION PT1(0:32, 0:32), PTIN(0:32)
DIMENSION V(0:32, 0:32)
DIMENSION W(0:32, 0:32), Z(0:32, 0:32)

COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ PTE
COMMON /CB5/ PTIN
COMMON /CB6/ 7

INTEGER DT, J, K, YY, Z2
REAL DT, VG
REAL 2 YADV, ZADV

C____CALCULATION OF ADVECTION TERM FOR POTENTIAL TEMPERATURE FIELD

DO 92 J = 2, YY - 2, 2
DO 93 K = 2, Z2 - 2, 2

PTIN1 = PTIN(K)
PTIN2 = PTIN(K + 2)
PTIN3 = PTIN(K - 2)
PTE1 = PTE(J, K)
PTE2 = PTE(J, K + 2)
PTE3 = PTE(J, K - 2)
W1 = V(J, K)
W2 = W(J, K)
Z1 = Z(J, K)
Z2 = Z(J, K + 2)

B-12
C____MODEL 2

Z3 = Z(J, K-2)

CALL ZGEN(PTE1, PTE4, PTE5, W1, PTIME, PTIME2, PTIN5, Z1, Z2, Z3, ZADV)
PTI(J, K) = PTE(J, K) - DT*SNGL(ZADV)

53 CONTINUE

PTI(J, 0) = 2*3 + PTE(J, 2) - PTE(J, 4)
PTI(J, 2) = PTE(J, 2)

92 CONTINUE

DO 85 K = 0, Z2 + 2
PTI(0, K) = PTE(2, K)
PTI(YY, K) = PTE(YY - 2, K)
83 CONTINUE

DO 94 K = 0, Z2 + 2
DO 95 J = 2, YY - 2
PTI1 = PTE(J, K)
PTI2 = PTE(J + 2, K)
PTI3 = PTE(J - 2, K)

CALL YSEN(YV, VGL, PTE1, PTE2, PTE3, V1, YADV)
PTI(J, K) = PTE(J, K) - DT*SNGL(YADV)

93 CONTINUE

PTI(0, K) = PTE(2, K)
PTI(YY, K) = PTE(YY - 2, K)
94 CONTINUE

RETURN
END

SUBROUTINE FILIFT(YY, Z2)

DIMENSION PTE(0:32, 0:32)
DIMENSION PT(0:62, 0:32), PTIME(0:32)
DIMENSION WU(0:32), APK(0:12), BOK(0:32)
DIMENSION APKZ(0:32), BKZ(0:32)

INTEGER J, JJ, K, YY, ZZ
REAL SM

COMMON /CE54/ PTE
COMMON /CE85/ PT, PTIME
COMMON /CE27/ SM, APK, BOK, APKZ, BKZ

C____APPLY IMPLICIT FILTER IN Y - DIRECTION

UU1(Y) = 0.4
C MODEL 2

DO 2260 K = 1, ZZ-2, 2
DO 2250 J = 2, YY-2, 2
UU1(J) = (PT(j, K-2) + PTE(J, K) + 2.0 * PTE(J, K) - S4 * UU1(J-2)) / APK(J)
2250 CONTINUE

UU1(YY) = UU1(YY-2) / APK(YY)
PTE(YY, K) = UU1(YY)
DO 2251 J = 0, YY-2, 2
J = YY-(J+2)
PTE(J, K) = PT(E(J, K) + PTE(J, K) + UU1(J)
2251 CONTINUE
2260 CONTINUE

C APPLY IMPLICIT FILTER IN Z - DIRECTION

C UU1(1) = 0.0
C DO 2270 J = 2, YY-2, 2
C DO 2280 K = 2, ZZ-2, 2
C UU1(K) = (PT(E(J, K-2) * PTE(J, K) * 2.0 * PTE(J, K) - S4 * UU1(K-2)) / APK(K)
C CONTINUE
C UU1(ZZ) = UU1(ZZ-2) / APK(ZZ)
PTE(J, ZZ) = UU1(ZZ)
C DO 2281 KK = 3, ZZ-2, 2
C K = ZZ-(K+2)
PTE(J, K) = 3 * Z(K) * PTE(J, K) * UU1(K)
C CONTINUE
C 2281 CONTINUE
C 2270 CONTINUE

DO 98 K = 0, ZZ, 2
PTE(0, K) = PTE(0, K)
PTE(YY, K) = PTE(YY-2, K)
98 CONTINUE

RETURN
END

SUBROUTINE GETR(STR, DT, G, YY, ZZ)

DIMENSION G, UPIF(J:5, 0:32), P(J:5, 0:32)
DIMENSION PT(J:5, 0:32), PTIN(J:32)
DIMENSION PTE(J:52, 0:32)
DIMENSION S(1, 2, 2, 2, 0:32)

COMMON /C04/ PT
COMMON /C33/ PTE, PTIN
COMMON /CGG/ UPIF
COMMON /CS7/ G
COMMON /C8R/ Z
COMMON /CS9/ S

INTEGER DT, J, K, YY, ZZ
REAL G, PERT, SUM, STR, TCPPT

DO 423 J = 2, YY-2, 2
SUM = C.0
MODEL 2

DO 424 K=Q,22-2*2
PTE(J,K+1)=PTE(J,K)+PTE(J,K+2)/2.0
SUM=SUM+(PTE(J,K+1)-OLOPT(J,K))/?IN(K+1)**2)*
(Z(J,K+2)-Z(J,K))

CONTINUE

C TOPPT=PTIN(ZZ)+ST+(S(J)-300.0)*
PCT(J,ZZ)=PTIN(ZZ)-TOPPT
C PERT=PTE(J,ZZ-2)-OLOPT(J,ZZ)+PTE(J,ZZ)+OLOPT(J,ZZ))/*
C *(PTIN(ZZ-2)+TOPPT/2.0)**2)
C PRT=0.5*PERT*(S(J)-2*Z(J,ZZ-2))
C SUM=SUM+PERT

UO = SUM/K = ZZ,K

OLDPTE(J,K)=PTE(J,K)

CONTINUE

OLDPTE(J,22)=PTE(J,22)

CONTINUE

DO 427 K=0,22-2,2
P(J,K)=P(2,K)

CONTINUE

RETURN

END

SUBROUTINE YGEN(Y1,VG,A1,A2,A3,V1,YADV)

REAL DY,VG,A1,A2,A3
REAL YADV

IF(V1*VG).GT.0.0.AND.YADV.0.0.
ELSE IF(V1*VG).LT.0.0.AND.YADV.0.0.
YADV=V1*VG*(A1-A3)/DY
ELSE
YADV=(V1*VG)*(A2-A1)/DY
END IF

RETURN

END

SUBROUTINE ZGEN(A1,A4,A5,A6,A7,A8,A9,A10,A11,A12,A13,A14,A15,A16,A17)

REAL A1,A4,A5,A6,A7,A8,A9,A10,A11,A12,A13,A14,A15,A16,A17
REAL YADV

IF(V1*VG).LT.0.0.

B-15
C.MODEL 2

ZADV=0.0
ELSE IF (W1.GT.0.0) THEN
   ZADV=W1*(((A1+G1)-(A5+G3))/Z1-Z3)
ELSE
   ZADV=W1*(((A4+G2)-(A1+G1))/Z2-Z1)
END IF

RETURN
END
C____MODEL 2

C____MODEL 2 - DIMENSIONAL SEA BREEZE MODEL WITH FLAT LAND
C SURFACE AND STRAIGHT COASTLINE.
C
C____NAME LIST FOR BREEZE.2D.F77
C
C BB = INITIAL SYNOPTIC POTENTIAL TEMPERATURE
C SC = STRATIFICATION (K/M)
C CP = SPECIFIC HEAT CAPACITY (J/Kg*K)
C DELTA = SMOOTHING CONSTANT
C DENS = DENSITY (Kg/M**3)
C DPDY = HORIZONTAL MESOSCALE PRESSURE GRADIENT
C DPT = DIABATIC POTENTIAL TEMPERATURE PERTURBATION (K)
C DT = TIMESTEP (SEC)
C DXY = HORIZONTAL GRADIENT OF WINDSPEED IN THE
C D Y = Y - DIRECTION
C D = Y GRIDSTEP (Km)
C F = CORIO.1ST PARAMETER=2*R*SIN(LAT)
C FDIS = CORIO.1ST PARAMETER=2*R*COS(LAT)
C G = ACCELERATION DUE TO GRAVITY (M/SEC**2)
C J = Y GRID NUMBER
C K = Z GRID NUMBER
C LAT = LATITUDE
C OLDOPT = MESOSCALE POTENTIAL TEMPERATURE PERTURBATION
C AT PREVIOUS TIMESTEP (K)
C P = EXNER FUNCTION=C*(PRESS/1000)**(R/A/C)
C PRESS = MESOSCALE PRESSURE PERTURBATION (N/M**2)
C PT = SYNOPTIC POTENTIAL TEMPERATURE (K)
C PTF = MESOSCALE POTENTIAL TEMPERATURE PERTURBATION (K)
C PTIN = INITIAL SYNOPTIC POTENTIAL TEMPERATURE (K)
C Q = DIABATIC HEAT FLUX (W/M**2)
C R = EARTHS ANGULAR VELOCITY (R.AO/SEC)
C RA = SPECIFIC GAS CONSTANT FOR DRY AIR (J/Kg*K)
C S = HEIGHT OF MATERIAL SURFACE (K)
C SOL = SOLAR CONSTANT (W/SEC)
C T = TIME (SEC)
C TIME = MAXIMUM OF SIMULATED TIME (SEC)
C U = MESOSCALE VELOCITY COMPONENT IN X - DIRECTION (M/SEC)
C V = MESOSCALE VELOCITY COMPONENT IN Y - DIRECTION (M/SEC)
C VG = GEOSTROPHIC COMPONENT IN THE Y - DIRECTION (M/SEC)
C W = MESOSCALE VERTICAL VELOCITY (M/SEC)
C YADV = ADVECTION TERM IN Y - DIRECTION
C YY = MAXIMUM GRID POINT NUMBER IN Y - DIRECTION
C ZADV = ADVECTION TERM IN Z - DIRECTION
C Z = ARRAY OF VERTICAL GRID LEVELS (M)
C ZZ = MAXIMUM GRID POINT NUMBER IN THE VERTICAL

DIMENSION P(U:32+D:32)*PT(C:H2+D:32)
DIMENSION OLDFIL(0:2+D:32)*PTIN(U:32)
DIMENSION PT1(L:32+G:32)
DIMENSION S(J:E0)
DIMENSION U(0:H2+J:32)

B-2
**C** MODEL 5

**C** 3 - DIMENSIONAL SEA BREEZE MODEL WITH FLAT LAND
C SURFACE AND STRAIGHT COASTLINE.

**C** NAME LIST FOR BREEZE_3D.F77

**C**

<table>
<thead>
<tr>
<th>Variable</th>
<th>Description</th>
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<tr>
<td><strong>38</strong></td>
<td>INITIAL SYNOPTIC POTENTIAL TEMPERATURE</td>
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<tr>
<td><strong>CC</strong></td>
<td>HEAT FLUX CONSTANT</td>
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<tr>
<td><strong>CP</strong></td>
<td>SPECIFIC HEAT CAPACITY (J/Kg. K)</td>
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<td><strong>DELTA</strong></td>
<td>SMOOTHING CONSTANT</td>
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<td><strong>DENS</strong></td>
<td>DENSITY(Kg/M<strong>3</strong>)</td>
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<tr>
<td><strong>DPDX</strong></td>
<td>HORIZONTAL MESOSCALE PRESSURE GRADIENT</td>
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<tr>
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<tr>
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<tr>
<td><strong>DT</strong></td>
<td>TIMESTEP(SEC)</td>
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<tr>
<td><strong>DUDX</strong></td>
<td>HORIZONTAL GRADIENT OF WINDSPEED IN THE X - DIRECTION</td>
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<tr>
<td><strong>DVDY</strong></td>
<td>HORIZONTAL GRADIENT OF WINDSPEED IN THE Y - DIRECTION</td>
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<tr>
<td><strong>DX</strong></td>
<td>X GRIDSTEP (Km)</td>
</tr>
<tr>
<td><strong>DY</strong></td>
<td>Y GRIDSTEP (Km)</td>
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<tr>
<td><strong>F</strong></td>
<td>CORIOLIS PARAMETER=2. R. SIN(LAT)</td>
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<tr>
<td><strong>FBAR</strong></td>
<td>CORIOLIS PARAMETER=2. R. COS(LAT)</td>
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<tr>
<td><strong>G</strong></td>
<td>ACCELERATION DUE TO GRAVITY (M/SEC<strong>2</strong>)</td>
</tr>
<tr>
<td><strong>I</strong></td>
<td>X GRID NUMBER</td>
</tr>
<tr>
<td><strong>J</strong></td>
<td>Y GRID NUMBER</td>
</tr>
<tr>
<td><strong>K</strong></td>
<td>Z GRID NUMBER</td>
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<td><strong>LAT</strong></td>
<td>LATITUDE</td>
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<td><strong>OLDPTE</strong></td>
<td>M CSOSCALE POTENTIAL TEMPERATURE PERTURBATION AT PREVIOUS TIMESTEP (K)</td>
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<tr>
<td><strong>P</strong></td>
<td>EXNER FUNCTION=CP*(PRESS/1000)**(RA/CP)</td>
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<td><strong>PRESS</strong></td>
<td>M ESOSCALE PRESSURE PERTURBATION (N/M<strong>2</strong>)</td>
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<td><strong>PT</strong></td>
<td>SYNOPTIC POTENTIAL TEMPERATURE (K)</td>
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<td><strong>PTIN</strong></td>
<td>INITIAL SYNOPTIC POTENTIAL TEMPERATURE (K)</td>
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<tr>
<td><strong>Q</strong></td>
<td>DIABATIC HEAT FLUX (W/M<strong>2</strong>)</td>
</tr>
<tr>
<td><strong>R</strong></td>
<td>EARTHS ANGULAR VELOCITY (RAD/SEC)</td>
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<tr>
<td><strong>RA</strong></td>
<td>SPECIFIC GAS CONSTANT FOR DRY AIR (J/Kg.K)</td>
</tr>
<tr>
<td><strong>S</strong></td>
<td>HEIGHT OF MATERIAL SURFACE (M)</td>
</tr>
<tr>
<td><strong>SOL</strong></td>
<td>SOLAR CONSTANT (W/M<strong>2</strong>)</td>
</tr>
<tr>
<td><strong>T</strong></td>
<td>TIME (SEC)</td>
</tr>
<tr>
<td><strong>TIME</strong></td>
<td>MAXIMUM OF SIMULATED TIME (SEC)</td>
</tr>
<tr>
<td><strong>U</strong></td>
<td>M ESOSCALE VELOCITY COMPONENT IN X - DIRECTION (M/SEC)</td>
</tr>
<tr>
<td><strong>UC</strong></td>
<td>GEOSTROPHIC COMPONENT IN THE X - DIRECTION (M/SEC)</td>
</tr>
<tr>
<td><strong>V</strong></td>
<td>M ESOSCALE VELOCITY COMPONENT IN Y - DIRECTION (M/SEC)</td>
</tr>
<tr>
<td><strong>VG</strong></td>
<td>GEOSTROPHIC COMPONENT IN THE Y - DIRECTION (M/SEC)</td>
</tr>
<tr>
<td><strong>W</strong></td>
<td>M ESOSCALE VERTICAL VELOCITY (M/SEC)</td>
</tr>
<tr>
<td><strong>XADV</strong></td>
<td>ADVECTION TERM IN X - DIRECTION</td>
</tr>
<tr>
<td><strong>XX</strong></td>
<td>MAXIMUM GRID POINT NUMBER IN X - DIRECTION</td>
</tr>
<tr>
<td><strong>YADV</strong></td>
<td>ADVECTION TERM IN Y - DIRECTION</td>
</tr>
<tr>
<td><strong>YY</strong></td>
<td>MAXIMUM GRID POINT NUMBER IN Y - DIRECTION</td>
</tr>
<tr>
<td><strong>ZADV</strong></td>
<td>ADVECTION TERM IN Z - DIRECTION</td>
</tr>
<tr>
<td><strong>Z</strong></td>
<td>ARRAY OF VERTICAL GRID LEVELS(M)</td>
</tr>
<tr>
<td><strong>ZZ</strong></td>
<td>MAXIMUM GRID POINT NUMBER IN THE VERTICAL</td>
</tr>
</tbody>
</table>
DIMENSION P(0:54,0:54,0:22), PT(0:54,0:54,0:22)
DIMENSION OLDPTE(0:54,0:54,0:22), PTIN(0:22)
DIMENSION PTE(0:54,0:54,0:22)
DIMENSION S(0:54,0:54)
DIMENSION U(0:54,0:54,0:22)
DIMENSION V(0:54,0:54,0:22)
DIMENSION W(0:54,0:54,0:22)
DIMENSION Z(0:54,0:54,0:22)

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ PTE
COMMON /CB5/ PT, PTIN
COMMON /CB6/ OLDPTE
COMMON /CB7/ P
COMMON /CB8/ Z
COMMON /CB9/ S

REAL BB, CC, CP, DELTA, DENS, DX, DY, F, G, LAT, SOL, R, RA, TEMP, UG, VG
REAL HOUR
INTEGER DT, I, J, K, KNT, KRAP, XX, YY, ZZ
INTEGER*4 T, TIME

OPEN(5,FILE='*>DATA>PT.3D.DAT')
OPEN(6,FILE='*>DATA>V.3D.DAT')
OPEN(7,FILE='*>DATA>W.3D.DAT')
OPEN(8,FILE='*>DATA>UP.3D.DAT')
OPEN(9,FILE='*>DATA>VECT.3D.DAT')
OPEN(10,FILE='*>DATA>TEMP.3D.DAT')

C PHYSICAL CONSTANTS
BB=0.0025
CC=0.1
CP=1004.0
DENS=1.21
DELTA=0.1
G=9.81
LAT=50.375
SOL=1370.0
R=7.29E-5
F=2.0*R*SIN(LAT*2.0*3.14159/360.0)
F BAR=2.0*R*CGS(LAT*2.0*3.14159/360.0)
RA=287.0
TEMP=286.0
UG=2.0
VG=0.0

C GRID SIZE CONSTANTS
DX=4000.0
DY=4000.0
DT=120
TIME=43200
C____MODEL 5

XX=32
YY=32
ZZ=20

DO 15 J=0,YY
DO 14 I=0,XX
Z(I, J, 0)=0.0
DO 12 K=1,ZZ
Z(I, J, K)=Z(I, J, K-1)+150.0
12 CONTINUE
S(I, J)=Z(I, J, ZZ)
14 CONTINUE
15 CONTINUE

C____INITIAL AND BOUNDARY CONDITIONS

DO 19 I=0,XX
DO 20 J=0,YY
PT(I, J, 0)=TEMP
DO 10 K=1,ZZ
PT(I, J, K)=PT(I, J, 0)+BB*Z(I, J, K)
10 CONTINUE
DO 21 K=0,ZZ
U(I, J, K)=0.0
V(I, J, K)=0.0
PTIN(K)=PT(I, J, K)
PTE(I, J, K)=0.0
OLDPTE(I, J, K)=0.0
P(I, J, K)=0.0
W(I, J, K)=0.0
21 CONTINUE
20 CONTINUE
19 CONTINUE

C____SET UP CONSTANTS FOR USE IN FILTER SUBROUTINES

CALL FCONST(XX, YY, DELTA)

C____TIME INTEGRATION

KNT=0
KRAP=0

DO 70 T=DT, TIME, DT
HOUR=T/3600.0
KNT=KNT+1
KRAP=KRAP+1
 IF(KNT.EQ.1)KNT=0
 IF(KRAP.EQ.5)KRAP=0

C____CALCULATION OF DIABATIC HEAT FLUX, Q

C READ(5, 500)Q
Q=CC*SGL*COS(LAT*2.0*3.14159/360.0)*SIN(R*T)
C____MODEL 5

C____FORCING OF POTENTIAL TEMP. AND HORIZONTAL WIND FIELDS
   CALL FORCE(CP, DENS, DT, DX, DY, F, FBAR, Q, UG, VG, XX, YY, ZZ)

C____CALCULATION OF ADVETION TERMS FOR HORIZONTAL WIND FIELDS
   CALL ADVHW(DT, DX, DY, UG, VG, XX, YY, ZZ)

C____APPLICATION OF FILTER TO HORIZONTAL WIND FIELD
   IF(KNT.EQ.0)THEN
      CALL FILTXX(XX, YY, ZZ)
   END IF

C____CALCULATION OF VERTICAL VELOCITY FIELD
   CALL VWIND(DX, DY, XX, YY, ZZ)

C____CALCULATION OF NEW MATERIAL SURFACE HEIGHT
   CALL MATGEN(DT, DX, DY, UG, VG, XX, YY, ZZ)

C____CALCULATION OF ADVETION TERMS FOR POT. TEMP. FIELD
   CALL ADVPT(DT, DX, DY, UG, VG, XX, YY, ZZ)

C____APPLICATION OF FILTER TO POTENTIAL TEMPERATURE FIELD
   IF(KNT.EQ.0)THEN
      CALL FILTPT(XX, YY, ZZ)
   END IF

C____CALCULATION OF PRESSURE FIELD
   CALL PGEN(DT, Q, XX, YY, ZZ)

C____PRINT OUT TIMES
   IF(T.EQ.10800. OR. T.EQ.21600. OR. T.EQ.32400. OR. T.EQ.43200)THEN
      IF(KRAP.EQ.0)THEN
         C OUTPUT DATA
         WRITE(5, 600) HOUR
         WRITE(5, 551)(PTIN(K)+PTE(B, J, K), J=0, YY, 2), K=0, ZZ, 2)
         WRITE(6, 601) HOUR
         WRITE(6, 551)(V(B, J, K), J=0, YY, 2), K=0, ZZ, 2)
         WRITE(7, 602) HOUR
         WRITE(7, 552)(W(B, J, K)*100.0, J=1, YY-1, 2), K=0, ZZ, 2)
         WRITE(8, 603) HOUR
         WRITE(8, 552)(W(I, J, K)*100.0, I=1, XX-1, 2), J=1, YY-1, 2)
         WRITE(9, 604) HOUR
         WRITE(9, 553)(VG+V(I, J, 2), I=2, XX-2, 2), J=2, YY-2, 2)
   END IF

B-20
C____MODEL 5

C    WRITE(9,605)HOUR
WRITE(9,553)((UG+U(I,J,2),I=2,XX-2,2),J=2,YY-2,2)
C    WRITE(10,606)HOUR
C    WRITE(10,553)(((PTIN(2)+PTE(I,J,2)),I=2,XX-2,2),J=2,YY-2,2)
END IF

70 CONTINUE

C    CLOSE(5)
C    CLOSE(6)
C    CLOSE(7)
C    CLOSE(8)
C    CLOSE(9)
C    CLOSE(10)

STOP

C____FORMAT STATEMENTS

551 FORMAT(1X,17F7.2)
552 FORMAT(1X,16F7.2)
553 FORMAT(1X,15F7.2)
554 FORMAT(1X,15F9.2)
600 FORMAT(1X,'TIME='F7.2,,'POT. TEMP. FIELD NORMAL TO COAST')
601 FORMAT(1X,'TIME='F7.2,,'HORIZONTAL WIND FIELD NORMAL TO COAST')
602 FORMAT(1X,'TIME='F7.2,,'VERTICAL VELOCITY FIELD NORMAL
* TO COAST')
603 FORMAT(1X,'TIME='F7.2,,'X/Y VERTICAL VELOCITY FIELD')
604 FORMAT(1X,'TIME='F7.2,,'X/Y HORIZONTAL WIND (V) FIELD')
605 FORMAT(1X,'TIME='F7.2,,'X/Y HORIZONTAL WIND (U) FIELD')
606 FORMAT(1X,'TIME='F7.2,,'X/Y POT. TEMP. FIELD')

END

SUBROUTINE FCONST(XX, YY, DELTA)

DIMENSION APK(0:54), BQK(0:54)
DIMENSION APKX(0:54), BQXX(0:54)

INTEGER XX, YY, I, J
REAL SM, SP, DELTA

COMMON /CB12/ SM, APK, BQK, APKX, BQXX

SM=1.0-DELTA
SP=2.0*(1.0+DELTA)

C____X - FILTER CONSTANTS

APKX(0)=1.0
BQXX(0)=1.0

DO 699 I=2,XX-2,2
  APKX(I)=SM*BQXX(I-2)+SP
  BQXX(I)=-SM/APKX(I)

END
C MODEL 3

CONTINUE

APKX(XX)=BQKX(XX-2)-1.0
BQKX(XX)=0.0

C Y - FILTER CONSTANTS

APK(0)=1.0
BQK(0)=1.0

DO 700 J=2,YY-2,2
APK(J)=SM*BQK(J-2)+SP
BQK(J)=-SM/APK(J)
700 CONTINUE

APK(YY)=BQK(YY-2)-1.0
BQK(YY)=0.0

RETURN
END

SUBROUTINE FORCE(CP, DENS, DT, DX, DY, F, FBAR, Q, UG, VG, XX, YY, ZZ)
DIMENSION P(0:54,0:54,0:22), PT(0:54,0:54,0:22)
DIMENSION PTE(0:54,0:54,0:22)
DIMENSION PTIN(0:22)
DIMENSION U(0:54,0:54,0:22)
DIMENSION V(0:54,0:54,0:22)
DIMENSION Z(0:54,0:54,0:22), W(0:54,0:54,0:22)
COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ PTE
COMMON /CB5/ PT,PTIN
COMMON /CB7/ P
COMMON /CBB/ Z
INTEGER DT, I, J, K, XX, YY, ZZ
REAL CP, DENS, DPDX, DPDY, DPT, DX, DY, F, FBAR, Q, UG, VG

C FORCING OF HORIZONTAL WIND FIELDS

C EQUATIONS (3.1) AND (3.2)

DO 315 I=2,XX-2,2
DO 300 J=2,YY-2,2
DO 301 K=2,ZZ-2,2
DPDX=(P(I+2,J,K)-P(I-2,J,K))/(DX+DX)
DPDY=(P(I,J+2,K)-P(I,J-2,K))/(DY+DY)
W(I,J,K)=(W(I+1,J+1,K)+W(I+1,J-1,K)+
W(I-1,J+1,K)+W(I-1,J-1,K))/4.0
U(I,J,K)=U(I,J,K)-DT*(-F*V(I,J,K)+PTIN(K)*DPDX+F*VG+
*(PTE(I,J,K)/PTIN(K))+FBAR*W(I,J,K))
V(I,J,K)=V(I,J,K)-DT*(F*U(I,J,K)+PTIN(K)*DPDY-F*UG+
*(PTE(I,J,K)/PTIN(K)))

B-22
C____MODEL 5

301 CONTINUE
U(I, J, ZZ) = 0.0
V(I, J, ZZ) = 0.0

300 CONTINUE
DO 302 K = 0, ZZ, 2
U(I, 0, K) = U(I, 2, K)
V(I, 0, K) = V(I, 2, K)
U(I, YY, K) = U(I, YY - 2, K)
V(I, YY, K) = V(I, YY - 2, K)

302 CONTINUE
315 CONTINUE
DO 303 J = 0, YY, 2
DO 304 K = 0, ZZ, 2
U(0, J, K) = U(2, J, K)
V(0, J, K) = V(2, J, K)
U(XX, J, K) = U(XX - 2, J, K)
V(XX, J, K) = V(XX - 2, J, K)

304 CONTINUE
303 CONTINUE

C____PARAMETERISATION OF HEAT CONVECTION IN ISENTROPIC B. L. OVERLAND

C____EQUATION (3.5)

DO 79 I = 2, XX - 2, 2
DO 80 J = 2, YY - 2, 2
IF(J .GE. 16) THEN
MM = 0
100 MM = MM + 2
DPT = DT * Q / (CP * DENS * Z(I, J, MM))
IF((PT(I, J, MM) + DPT) .GT. PT(I, J, MM + 2)) GO TO 100
DO 110 K = 0, MM, 2
PT(I, J, K) = PT(I, J, K) + DPT
PTE(I, J, K) = PTE(I, J, K) + DPT
110 CONTINUE
END IF
80 CONTINUE
79 CONTINUE

RETURN
END

SUBROUTINE ADVH(DT, DX, DY, UG, VG, XX, YY, ZZ)

DIMENSION U(0:54, 0:54, 0:22)
DIMENSION V(0:54, 0:54, 0:22)
DIMENSION Z(0:54, 0:54, 0:22), W(0:54, 0:54, 0:22)

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB8/ Z

INTEGER DT, I, J, K, XX, YY, ZZ
REAL DX, DY, UG, VG

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C MODEL 5

REAL U1, U2, U3, V1, V2, V3, W1, Z1, Z2, Z3
REAL*8 XADV, YADV, ZADV

C CALCULATION OF ADVECTION TERMS FOR HORIZONTAL WIND FIELDS

C EQUATIONS (3.1) AND (3.2)

C IN THE Z - DIRECTION

DO 305 I=2, XX-2, 2
DO 310 J=2, YY-2, 2
DO 320 K=2, ZZ-2, 2

U1=U(I, J, K)
U2=U(I, J, K+2)
U3=U(I, J, K-2)
V1=V(I, J, K)
V2=V(I, J, K+2)
V3=V(I, J, K-2)
W1=W(I, J, K)
Z1=Z(I, J, K)
Z2=Z(I, J, K+2)
Z3=Z(I, J, K-2)

CALL ZGEN(U1, U2, U3, W1, UG, UG, UG, Z1, Z2, Z3, ZADV)
U(I, J, K)=U(I, J, K)-DT*SNGL(ZADV)

CALL ZGEN(V1, V2, V3, W1, VG, VG, VG, Z1, Z2, Z3, ZADV)
V(I, J, K)=V(I, J, K)-DT*SNGL(ZADV)

320 CONTINUE

U(I, J, 0)=U(I, J, 0)
V(I, J, 0)=V(I, J, 0)
U(I, J, ZZ)=U(I, J, ZZ)
V(I, J, ZZ)=V(I, J, ZZ)

310 CONTINUE

DO 311 K=0, ZZ, 2
U(I, 0, K)=U(I, 2, K)
V(I, 0, K)=V(I, 2, K)
U(I, YY, K)=U(I, YY-2, K)
V(I, YY, K)=V(I, YY-2, K)

311 CONTINUE

305 CONTINUE

DO 306 K=0, ZZ, 2
DO 307 J=0, YY, 2
U(0, J, K)=U(2, J, K)
V(0, J, K)=V(2, J, K)
U(XX, J, K)=U(XX-2, J, K)
V(XX, J, K)=V(XX-2, J, K)

307 CONTINUE

C IN THE Y - DIRECTION
MODEL 5

DO 330 K=0, ZZ, 2
DO 335 I=2, XX-2, 2
DO 340 J=2, YY-2, 2

U1=U(I, J, K)
U2=U(I, J+2, K)
U3=U(I, J-2, K)
V1=V(I, J, K)
V2=V(I, J+2, K)
V3=V(I, J-2, K)

CALL Y GEN(DY, VG, U1, U2, U3, V1, YADV)
U(I, J, K)=U(I, J, K)-DT*SNGL(YADV)

CALL Y GEN(DY, VG, V1, V2, V3, V1, YADV)
V(I, J, K)=V(I, J, K)-DT*SNGL(YADV)

340 CONTINUE

V(I, 0, K)=V(I, 2, K)
V(I, YY, K)=V(I, YY-2, K)
U(I, 0, K)=U(I, 2, K)
U(I, YY, K)=U(I, YY-2, K)

335 CONTINUE

DO 336 J=0, YY, 2
U(0, J, K)=U(2, J, K)
V(0, J, K)=V(2, J, K)
U(XX, J, K)=U(XX-2, J, K)
V(XX, J, K)=V(XX-2, J, K)

336 CONTINUE

330 CONTINUE

C IN THE X - DIRECTION

DO 405 J=2, YY-2, 2
DO 406 K=0, ZZ, 2
DO 407 I=2, XX-2, 2

U1=U(I, J, K)
V1=V(I, J, K)
U2=U(I+2, J, K)
V2=V(I+2, J, K)
U3=U(I-2, J, K)
V3=V(I-2, J, K)

CALL X GEN(DX, UG, U1, U2, U3, U1, XADV)
U(I, J, K)=U(I, J, K)-DT*SNGL(XADV)

CALL X GEN(DX, UG, V1, V2, V3, V1, XADV)
V(I, J, K)=V(I, J, K)-DT*SNGL(XADV)

407 CONTINUE

U(0, J, K)=U(2, J, K)
SUBROUTINE FILTH(XX, YY, ZZ)

DIMENSION U(0:54, 0:54, 0:22)
DIMENSION V(0:54, 0:54, 0:22)
DIMENSION UU1(0:54), UU2(0:54), APK(0:54), BQK(0:54)
DIMENSION APKX(0:54), BQKX(0:54)

INTEGER I, II, J, JJ, XX, YY, ZZ
REAL SM

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB12/ SM, APK, BQK, APKX, BQKX

C_____SMOOTHING OF HORIZONTAL WIND FIELDS
C_____EQUATION (3.10)
C_____APPLY FILTER IN Y - DIRECTION

UU1(0)=0.0
UU2(0)=0.0

DO 1160 K=0, ZZ-2, 2
DO 1155 I=2, XX-2, 2
DO 1150 J=2, YY-2, 2
UU2(J)=((V(I, J-2, K)+V(I, J+2, K)+2.0*V(I, J, K)-SM*UU2(J-2))/APK(J)
UU1(J)=(U(I, J-2, K)+U(I, J+2, K)+2.0*U(I, J, K)-SM*UU1(J-2))/APK(J)

1150 CONTINUE
UU1(YY)=-UU1(YY-2)/APK(YY)
UU2(YY)=-UU2(YY-2)/APK(YY)
U(I, YY, K)=UU1(YY)
V(I, YY, K)=UU2(YY)

DO 1151 JJ=0, YY-2, 2
J=YY-(JJ+2)

B-26
MODEL 5

U(I, J, K) = BQ(J) * U(I, J + 2, K) + UU1(J)
V(I, J, K) = BQK(J) * V(I, J + 2, K) + UU2(J)

1151 CONTINUE
1155 CONTINUE
DO 1156 J = 0, YY, 2
U(0, J, K) = U(2, J, K)
V(0, J, K) = V(2, J, K)
U(XX, J, K) = U(XX-2, J, K)
V(XX, J, K) = V(XX-2, J, K)
1156 CONTINUE
1160 CONTINUE

APPLY-FILTER IN X DIRECTION

UU1(0) = 0.0
UU2(0) = 0.0

DO 4460 K = 0, ZZ-2, 2
DO 4455 J = 2, YY-2, 2
DO 4450 I = 2, XX-2, 2
UU2(I) = (V(I-2, J, K) + V(I+2, J, K) + 2.0 * V(I, J, K) - SM * UU2(I-2)) / APKX(I)
UU1(I) = (U(I-2, J, K) + U(I+2, J, K) + 2.0 * U(I, J, K) - SM * UU1(I-2)) / APKX(I)

4450 CONTINUE

UU1(XX) = -UU1(XX-2) / APKX(XX)
UU2(XX) = -UU2(XX-2) / APKX(XX)
U(XX, J, K) = UU1(XX)
V(XX, J, K) = UU2(XX)
DO 4451 I = 0, XX-2, 2
I = XX-(II+2)
U(I, J, K) = BQKX(I) * U(I+2, J, K) + UU1(I)
V(I, J, K) = BQKX(I) * V(I+2, J, K) + UU2(I)

4451 CONTINUE
4455 CONTINUE

RETURN
END

SUBROUTINE VWIND(DX, DY, XX, YY, ZZ)

DIMENSION U(0:54, 0:54, 0:22), V(0:54, 0:54, 0:22)
DIMENSION W(0:54, 0:54, 0:22), Z(0:54, 0:54, 0:22)
COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ Z

INTEGER I, J, K, XX, YY, ZZ
REAL DUDX, DUDY, DX, DY

C____CALCULATE NEW VERTICAL VELOCITY FIELD
C THROUGH MASS CONTINUITY
C____EQUATION (3.3)

DO 349 I=1, XX-1, 2
DO 350 J=1, YY-1, 2
DO 351 K=1, ZZ-1, 2
V(I, J+1, K) = (V(I+1, J+1, K+1) + V(I+1, J+1, K-1) + V(I-1, J+1, K+1) + V(I-1, J+1, K-1))/4.0
V(I, J-1, K) = (V(I+1, J-1, K+1) + V(I+1, J-1, K-1) + V(I-1, J-1, K+1) + V(I-1, J-1, K-1))/4.0
U(I+1, J, K) = (U(I+1, J+1, K+1) + U(I+1, J+1, K-1) + U(I-1, J+1, K+1) + U(I-1, J+1, K-1))/4.0
U(I-1, J, K) = (U(I-1, J+1, K+1) + U(I-1, J+1, K-1) + U(I-1, J-1, K+1) + U(I-1, J-1, K-1))/4.0
DUDY = (V(I, J+1, K) - V(I, J-1, K))/DY
DUDX = (U(I+1, J, K) - U(I-1, J, K))/DX
W(I, J, K) = 0.5*(W(I, J, K+1) + W(I, J, K-1))/0.0
351 CONTINUE
350 CONTINUE
349 CONTINUE

DO 354 I=2, XX-2, 2
DO 352 J=2, YY-2, 2
W(I, J, ZZ) = 0.5*(W(I+1, J+1, ZZ) + W(I+1, J-1, ZZ) + W(I-1, J+1, ZZ) + W(I-1, J-1, ZZ))/4.0
352 CONTINUE
354 CONTINUE

RETURN
END

SUBROUTINE MATGEN(DT, DX, DY, UG, VG, XX, YY, ZZ)

DIMENSION U(0: 54, 0: 54, 0: 22)
DIMENSION V(0: 54, 0: 54, 0: 22)
DIMENSION U(0: 54, 0: 54)
DIMENSION W(0: 54, 0: 54, 0: 22), Z(0: 54, 0: 54, 0: 22)

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB9/ Z
COMMON /CB9/ S

INTEGER DT, I, J, XX, YY, ZZ
REAL DX, DY, UG, VG
REAL SI, S2, S3, S4, S5, U1, V1
REAL*8 XADV, YADV

C____CALCULATION OF ADVECTION TERM
C____MODEL 5

C____EQUATION (3. 4)

DO 409 I=2, XX-2, 2
DO 410 J=2, YY-2, 2

S1=S(I, J)
S2=S(I, J+2)
S3=S(I, J-2)
S4=S(I+2, J)
S5=S(I-2, J)
U1=U(I, J, ZZ-2)
V1=V(I, J, ZZ-2)

CALL YGEN(DY, VG, S1, S2, S3, V1, YADV)
S(I, J)=S(I, J)-DT*SNGL(YADV)

CALL XGEN(DX, UG, S1, S4, S5, U1, XADV)
S(I, J)=S(I, J)-DT*SNGL(XADV)

410 CONTINUE
409 CONTINUE

C____CALCULATION OF NEW MATERIAL SURFACE HEIGHT

C____EQUATION (3. 4)

DO 412 I=2, XX-2, 2
DO 413 J=2, YY-2, 2

S(I, J)=S(I, J)+DT*W(I, J, ZZ)

413 CONTINUE

S(I, 0)=S(I, 2)
S(I, YY)=S(I, YY-2)

412 CONTINUE

DO 411 J=0, YY, 2
S(0, J)=S(2, J)
S(XX, J)=S(XX-2, J)

411 CONTINUE

DO 417 I=0, XX, 2
DO 415 J=0, YY, 2
Z(I, J, ZZ)=S(I, J)

415 CONTINUE
417 CONTINUE

RETURN
END

SUBROUTINE ADVPT(DT, DX, DY, UG, VG, XX, YY, ZZ)

DIMENSION PTE(0:54, 0:54, 0:22)
DIMENSION PT(0:54, 0:54, 0:22), PTIN(0:22)
DIMENSION U(0:54, 0:54, 0:22), V(0:54, 0:54, 0:22)
DIMENSION W(0:54, 0:54, 0:22), Z(0:54, 0:54, 0:22)
C____MODEL 5

COMMON /CB1/ U
COMMON /CB2/ V
COMMON /CB3/ W
COMMON /CB4/ PTE
COMMON /CB5/ PT, PTIN
COMMON /CB8/ Z

INTEGER DT, I, J, K, XX, YY, ZZ
REAL DX, DY, UG, VG
REAL*8 XADV, YADV, ZADV
REAL PTIN1, PTIN2, PTIN3, PTE1, PTE2, PTE3

C____CALCULATION OF ADVECTION TERMS FOR POTENTIAL TEMPERATURE FIELD

C____EQUATION (3.5)

C____IN THE Z - DIRECTION

DO 91 I=2, XX-2, 2
DO 92 J=2, YY-2, 2
DO 93 K=2, ZZ-2, 2

PTIN1=PTIN(K)
PTIN2=PTIN(K+2)
PTIN3=PTIN(K-2)
PTE1=PTE(I, J, K)
PTE2=PTE(I, J, K+2)
PTE3=PTE(I, J, K-2)
W1=W(I, J, K)
Z1=Z(I, J, K)
Z2=Z(I, J, K+2)
Z3=Z(I, J, K-2)

CALL ZGEN(PTE1, PTE2, PTE3, W1, PTIN1, PTIN2, PTIN3, Z1, Z2, Z3, ZADV)
PTE(I, J, K)=PTE(I, J, K)-DT*SNGL(ZADV)

93 CONTINUE

PTE(I, J, 0)=2.0*PTE(I, J, 2)-PTE(I, J, 4)
PTE(I, J, ZZ)=PTE(I, J, ZZ)

92 CONTINUE

DO 83 K=0, ZZ, 2
PTE(I, 0, K)=PTE(I, 2, K)
PTE(I, YY, K)=PTE(I, YY-2, K)

83 CONTINUE

91 CONTINUE

DO 94 K=0, ZZ, 2
DO 95 J=0, YY, 2
PTE(0, J, K)=PTE(2, J, K)
PTE(XX, J, K)=PTE(XX-2, J, K)

95 CONTINUE

94 CONTINUE

C____IN THE Y - DIRECTION
C_____MODEL 5

DO 96 K=0, ZZ, 2
DO 97 I=2, XX-2, 2
DO 98 J=2, YY-2, 2

PTE1=PTE(I, J, K)
PTE2=PTE(I, J+2, K)
PTE3=PTE(I, J-2, K)
V1=V(I, J, K)

CALL YGEN(DY, VG, PTE1, PTE2, PTE3, V1, YADV)
PTE(I, J, K)=PTE(I, J, K)-DT*SNGL(YADV)

98 CONTINUE

PTE(I, 0, K)=PTE(I, 2, K)
PTE(I, YY, K)=PTE(I, YY-2, K)

97 CONTINUE

DO 99 J=0, YY, 2
PTE(0, J, K)=PTE(2, J, K)
PTE(XX, J, K)=PTE(XX-2, J, K)

99 CONTINUE

96 CONTINUE

C_____IN THE X - DIRECTION

DO 304 J=2, YY-2, 2
DO 306 K=0, ZZ, 2
DO 307 I=2, XX-2, 2

PTE1=PTE(I, J, K)
PTE2=PTE(I+2, J, K)
PTE3=PTE(I-2, J, K)
U1=U(I, J, K)

CALL XGEN(DX, UG, PTE1, PTE2, PTE3, U1, XADV)
PTE(I, J, K)=PTE(I, J, K)-DT*SNGL(XADV)

307 CONTINUE

PTE(0, J, K)=PTE(2, J, K)
PTE(XX, J, K)=PTE(XX-2, J, K)

306 CONTINUE

304 CONTINUE

DO 308 I=0, XX, 2
DO 309 K=0, ZZ, 2

PTE(I, 0, K)=PTE(I, 2, K)
PTE(I, YY, K)=PTE(I, YY-2, K)

309 CONTINUE

308 CONTINUE

RETURN
SUBROUTINE FILTPT(XX, YY, ZZ)

DIMENSION PTE(0:54, 0:54, 0:22)
DIMENSION UUl(0:54), APK(0:54), BQK(0:54)
DIMENSION APKX(0:54), BQKX(0:54)
INTEGER I, II, J, JJ, K, XX, YY, ZZ
REAL SM

COMMON /CB4/ PTE
COMMON /CB12/ SM, APK, BQK, APKX, BQKX

C____SMOOTHING OF POTENTIAL TEMPERATURE FIELD

C____EQUATION (3.10)

C____APPLY FILTER IN Y - DIRECTION

UU1(0)=0.0

DO 2250 K=0, ZZ-2, 2
DO 2255 I=2, XX-2, 2
DO 2250 J=2, YY-2, 2
UU1(J)=PUi(I, J-2, K)+PTE(I, J+2, K)+2.0*PTE(I, J, K)
-*SM*UU1(J-2))/APK(J)
2250 CONTINUE

UU1(YY)=-UU1(YY-2)/APK(YY)

PTE(I, YY, K)=UU1(YY)

DO 2251 JJ=0, YY-2, 2
J=YY-(JJ+2)
PTE(I, J, K)=BQK(J)*PTE(I, J+2, K)+UU1(J)
2251 CONTINUE

2255 CONTINUE

DO 2256 J=0, YY, 2
PTE(0, J, K)=PTE(2, J, K)
PTE(XX, J, K)=PTE(XX-2, J, K)
2256 CONTINUE

2260 CONTINUE

C____APPLY FILTER IN X - DIRECTION

UU1(0)=0.0

DO 5550 K=0, ZZ-2, 2
DO 5555 J=2, YY-2, 2
DO 5550 I=2, XX-2, 2
UU1(I)=(PTE(I-2, J, K)+PTE(I+2, J, K)+2.0*PTE(I, J, K)
-*SM*UU1(I-2))/APKX(I)
5550 CONTINUE

UU1(XX)=-UU1(XX-2)/APKX(XX)
PTE(XX, J, K)=UU1(XX)

DO 5551 II=0, XX-2, 2
I=XX-(II+2)
PTE(I, J, K)=BQKX(I)*PTE(I+2, J, K)+UU1(I)
C____MODEL 5

5551 CONTINUE
5555 CONTINUE
   DO 5556 I=0, XX, 2
      PTE(I,0,K)=PTE(I,2,K)
      PTE(I,YY,K)=PTE(I,YY-2,K)
   5556 CONTINUE
   5560 CONTINUE
   RETURN
END

SUBROUTINE PGEN(DT,G,XX,YY,ZZ)

DIMENSION OLDPTE(0:54,0:54,0:22), P(0:54,0:54,0:22)
DIMENSION PT(0:54,0:54,0:22), PTIN(0:22)
DIMENSION PTE(0:54,0:54,0:22)
DIMENSION S(0:54,0:54), Z(0:54,0:54,0:22)

COMMON /CB4/ PTE
COMMON /CB5/ PT, PTIN
COMMON /CB6/ OLDPTE
COMMON /CB7/ P
COMMON /CB8/ Z
COMMON /CB9/ S

INTEGER DT, I, J, K, XX, YY, ZZ
REAL G, SUM

C____CALCULATION_OF_NEW_SURFACE_PRESSURE

C____EQUATION (3.6)
   DO 422 I=2, XX-2, 2
   DO 423 J=2, YY-2, 2
      SUM=0.0
      DO 424 K=0, ZZ-2, 2
         PTE(I,J,K+1)=(PTE(I,J,K)+PTE(I,J,K+2))/2.0
         SUM=SUM+(PTE(I,J,K+1)-OLDPTE(I,J,K+1))/PTIN(K+1)**2*
            (Z(I,J,K+2)-Z(I,J,K))
      424 CONTINUE
   P(I,J,0)=P(I,J,0)-G*SUM

C____NEW_PRESSURE_FIELD_CALCULATED_USING_THE
C____HYDROSTATIC_ASSUMPTION

C____EQUATION (3.7)
   DO 425 K=0, ZZ-4, 2
      P(I,J,K+2)=P(I,J,K)+G*PTE(I,J,K+1)/(PTIN(K+1)**2)*
            (Z(I,J,K+2)-Z(I,J,K))
   425 CONTINUE
   DO 426 K=1, ZZ-1, 2
      OLDPTE(I,J,K)=PTE(I,J,K)
   426 CONTINUE

B-33
SUBROUTINE XGEN(DX, UG, A1, A2, A3, U1, XADV)
REAL DX, UG, A1, A2, A3, U1
REAL*8 XADV
C____CALCULATION OF ADVECTION TERMS \frac{\partial A}{\partial x}
IF((U1+UG). EQ. 0.0) THEN
  XADV = 0.0
ELSE IF((U1+UG). GT. 0.0) THEN
  XADV = (U1+UG) * (A1-A3) / DX
ELSE
  XADV = (U1+UG) * (A2-A1) / DX
END IF
RETURN
END

SUBROUTINE YGEN(DY, VG, A1, A2, A3, V1, YADV)
REAL DY, VG, A1, A2, A3, V1
REAL*8 YADV
C____CALCULATION OF ADVECTION TERMS \frac{\partial A}{\partial y}
IF((V1+VG). EQ. 0.0) THEN
  YADV = 0.0
ELSE IF((V1+VG). GT. 0.0) THEN
  YADV = (V1+VG) * (A1-A3) / DY
ELSE
  YADV = (V1+VG) * (A2-A1) / DY
END IF
RETURN
END
SUBROUTINE ZGEN(A1, A4, A5, W1, G1, G2, G3, Z1, Z2, Z3, ZADV)
REAL A1, A4, A5, G1, G2, G3, W1, Z1, Z2, Z3
      REAL ZADV
C______CALCULATION OF ADVECTION TERMS \partial dA/\partial Z
      IF(W1 .EQ. 0. 0) THEN
        ZADV=0. 0
      ELSE_IF(W1 .GT. 0. 0) THEN
        ZADV=W1*[(A1+G1)-(A5+G3)]/(Z1-Z3)
      ELSE
        ZADV=W1*[(A4+G2)-(A1+G1)]/(Z2-Z1)
      END_IF
RETURN
END
APPENDIX C

GRAPHICS PROGRAMS
VECTOR DIAGRAMS USING NAG GRAPHICAL SUPPLEMENT

AND GINO-F AND HERSEY LIBRARIES

NAG SUBROUTINES USED

J06AAF - draws a scaled border to fit the current data region.
AHF - draws a centred title at top of data region.
AJF - draws an axis title.
WAF - initialises NAG graphical system and establishes mapping of initial data region onto default viewport.
WBF - declares current data region and establishes mapping of this data region onto current data region onto current viewport with margin option.
WCF - sets current viewport on plotting surface and establishes mapping of current data region onto viewport.
WDF - selects a new frame.
WZF - terminates graphical output to currently selected device.
YAF - moves pen to position (X,Y) in user coords.
YBF - increments pen position by (DX,DY) in user coords.
YDF - draws a line, advancing by an increment (DX,DY) in user coords.
YHF - draws character string.
YJF - sets size of markers.
YKF - sets width and height of characters.
YLF - sets spacing between characters.
X04AAF - returns or sets the current error message unit number.

GINO-F SUBROUTINES USED

CHAANG - sets orientation of character, with angle given in degrees.
CHASIZ - sets width and height of characters.
MOVTO2 - moves pen to position (X,Y) in user coords.
POSSPA - gives current pen position.

HERSEY SUBROUTINES USED

HERCEN - outputs graphical character centred at current position.
HERCHA - outputs graphical character at current position.
HEREND - close character information file.

LOCAL SCALARS

INTEGER I, J, K, NCHARS, XX, YY

LOCAL ARRAYS

INTEGER ICHARS(80)
REAL*8 SBUDAT(20,20), SBVDAT(20,20)
VECTOR DIAGRAMS USING NAG GRAPHICAL SUPPLEMENT

```fortran
REAL*8 DIST, DX, DY, XINC, YINC
REAL RAD, SIZE, Theta
CHARACTER*20 A, B, C, D

FILE HANDLING
A='**DATA>NEW.DAT'
B='**DATA>ERRORS'
C='**DATA>LAB1'
D='**DATA>TIME2.DAT'
OPEN(5, FILE=A)
OPEN(6, FILE=B)
OPEN(7, FILE=C)
OPEN(8, FILE=D)

SELECT OUTPUT CHANNEL FOR ERROR MESSAGES
CALL X04AAF(1, 6)

INITIALISE PLOTTING DEVICE
CALL GINO
CALL CCB1
CALL JO6WAF
CALL JO6WDF

XX=16
YY=16
RAD=360.0/(2.0*3.14549)
DO 20 K=1, 4

READ DATA
READ(5, 10)((SBVDAT(I, J), I=1, XX-1), J=1, YY-1)
READ(5, 10)((SBUDAT(I, J), I=1, XX-1), J=1, YY-1)

MAP DATA REGION TO VIEWPORT
CALL JO6WBF(0.0DO, 64.0DO, 0.0DO, 64.0DO, 1)
IF(K.EQ.1) THEN
CALL JO6WCF(0.15DO, 0.5DO, 0.5DO, 0.85DO)
ELSE IF(K.EQ.2) THEN
CALL JO6WCF(0.5DO, 0.85DO, 0.5DO, 0.85DO)
ELSE IF(K.EQ.3) THEN
CALL JO6WCF(0.15DO, 0.5DO, 0.15DO, 0.5DO)
ELSE
CALL JO6WCF(0.5DO, 0.85DO, 0.15DO, 0.5DO)
END IF

DRAW COASTLINE
CALL JO6YAF(0.0DO, 32.0DO)
CALL JO6YDF(64.0DO, 0.0DO)
CALL JO6YDF(28.0DO, 0.0DO)
CALL JO6YDF(0.0DO, 8.0DO)
```

C-2
C VECTOR DIAGRAMS USING NAG GRAPHICAL SUPPLEMENT

CALL J06YDF(8. ODO, 0. ODO)
CALL J06YDF(0. ODO, -8. ODO)
CALL J06YDF(28. ODO, 0. ODO)

C DRAW AND LABEL X AND Y-AXES

CALL J06AFF(8. ODO, 8. ODO)

IF(K. EQ. 3. OR. K. EQ. 4) THEN
  CALL J06AJF(1, 'HORIZONTAL DISTANCE X Km', 24)
END IF

IF(K. EQ. 1. OR. K. EQ. 3) THEN
  CALL J06AJF(2, 'HORIZONTAL DISTANCE Y Km', 24)
END IF

CALL J06AHF('HORIZONTAL WIND FIELD VECTORS (M/S)', 35)

C DRAW TIMES

CALL J06YAF(52. ODO, 59. ODO)
READ(8,444)(ICHARS(I), I=1, 6)
CALL J06YHF(ICHARS, 6)

C DRAW VECTORS

CALL J06YAF(0. ODO, 0. ODO)
DO 11 I=1, XX-1
  DX=4. ODO
  CALL J06YBF(DX, 0. ODO)
DO 12 J=1, YY-1
  DY=4. ODO
  CALL J06YBF(0. ODO, DY)

XINC=SBUDAT(I, J)
YINC=SBVDAT(I, J)

IF(XINC. EQ. O. O. AND. YINC. EQ. O. O) THEN
  GO TO 12
ELSE IF(XINC. EQ. O. O. AND. YINC. GT. 0. 0) THEN
  Theta=90. 0
ELSE IF(XINC. EQ. O. O. AND. YINC. LT. 0. 0) THEN
 Theta=-90. 0
ELSE IF(XINC. GT. 0. 0. AND. YINC. EQ. O. O) THEN
  Theta=0. 0
ELSE IF(XINC. LT. 0. 0. AND. YINC. EQ. O. O) THEN
  Theta=180. 0
ELSE IF(XINC. LT. 0. 0. AND. YINC. LT. 0. 0) THEN
  Theta=ATAN(YINC/XINC)
  Theta=Theta*RAD+180. 0
ELSE IF(XINC. GT. 0. 0. AND. YINC. GT. 0. 0) THEN
  Theta=ATAN(YINC/XINC)
  Theta=Theta*RAD
ELSE IF(XINC. GT. 0. 0. AND. YINC. LT. 0. 0) THEN
  Theta=ATAN(YINC/XINC)
  Theta=Theta*RAD
ELSE IF(XINC. LT. 0. 0. AND. YINC. GT. 0. 0) THEN
  Theta=ATAN(XINC/YINC)
  Theta=Theta*RAD
ELSE IF(XINC. LT. 0. 0. AND. YINC. LT. 0. 0) THEN
  Theta=ATAN(XINC/YINC)

C-3
VECTOR DIAGRAMS USING NAG GRAPHICAL SUPPLEMENT

THETA = THETA * RAD + 180.0
END IF

SIZE = SQRT (XINC**2 + YINC**2)
SIZE = SIZE * 2.0
CALL POSSPA (XP, YP, ZP)
CALL CHAANG (THETA)
CALL CHASIZ (SIZE, SIZE)
CALL HERCEN (1261)
CALL MOVTO2 (XP, YP)
12 CONTINUE
CALL J06YBF (0.0D0, -60.0D0)
11 CONTINUE

CALL CHAANG (0.0)
20 CONTINUE

DRAWS CHARACTER STRINGS

CALL J06WBF (0.0D0, 100.0D0, 0.0D0, 100.0D0, 1.0D0)
CALL J06WCF (0.0D0, 1.0D0, 0.0D0, 1.0D0)
CALL J06YUF (1.5D0)
CALL J06YKF (0.75D0, 1.5D0)
CALL J06YLF (1.5D0, 0.0D0)

DIST = 102.5D0

DO 100 L = 1, 5
IF (L .EQ. 1) THEN
NCHARS = 34
ASSIGN 61 TO NU
ELSE IF (L .EQ. 2) THEN
NCHARS = 40
ASSIGN 62 TO NU
ELSE IF (L .EQ. 3) THEN
NCHARS = 10
ASSIGN 63 TO NU
ELSE IF (L .EQ. 4) THEN
NCHARS = 9
ASSIGN 64 TO NU
ELSE
NCHARS = 40
ASSIGN 65 TO NU
END IF
READ (5, NU) (ICHARS(I), I = 1, NCHARS)
DIST = DIST - 2.5D0
CALL J06YAF (10.0D0, DIST)
CALL J06YHF (ICHARS, NCHARS)
100 CONTINUE

DRAW DIAGRAM LABELS

DIST = 7.5D0

DO 200 L = 1, 3
IF (L .EQ. 1) THEN
NCHARS = 67
ASSIGN 71 TO NU

C-4
ELSE IF(L. EQ. 2) THEN
    NCHARS=70
    ASSIGN 72 TO NU
ELSE
    NCHARS=70
    ASSIGN 73 TO NU
END IF
READ(7,NU)(ICHARS(I),I=1,NCHARS)
DIST=2.5D0
CALL J06YAF(0.0D0,DIST)
CALL J06YHF(ICHARS,NCHARS)
CONTINUE
CALL J06YAF(50.0D0,95.0D0)
READ(8,445)(ICHARS(I),I=1,23)
CALL J06YHF(ICHARS,23)
C DRAW ARROW KEY
SIZE=4.0
CALL J06WBF(0.0D0,64.0D0,0.0D0,64.0D0,1)
CALL J06WCF(0.5D0,0.85D0,0.65D0,1.0D0)
CALL CHASIZ1(SIZE,SIZE)
CALL J06YAF(0.0D0,49.0D0)
CALL HERCHA(1262)
C END PLOT
CALL J06W2F
CALL HEREND
CALL GINEND
CLOSE(5)
CLOSE(6)
CLOSE(7)
CLOSE(8)
STOP
10 FORMAT(IX,15F7.2)
61 FORMAT(34A1)
62 FORMAT(40A1)
63 FORMAT(10A1)
64 FORMAT(9A1)
65 FORMAT(40A1)
71 FORMAT(67A1)
72 FORMAT(70A1)
73 FORMAT(70A1)
444 FORMAT(6A1)
445 FORMAT(23A1)
END
CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT

CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT

NAG SUBROUTINES USED

J06AFFF - draws a scaled border to fit the current data region.
AHF - draws a centred title at top of data region.
AJF - draws an axis title.
WAF - initialises NAG graphical system and establishes mapping of initial data region onto default viewport.
WBK - declares current data region and establishes mapping of this data region onto current data region onto current viewport with margin option.
WCF - sets current viewport on plotting surface and establishes mapping of current data region onto viewport.
WDF - selects a new frame.
WZQ - terminates graphical output to currently selected device.
XGF - sets character size scale factor.
YAF - moves pen to position (X,Y) in user coords.
YBF - increments pen position by (DX, DY) in user coords.
YDF - draws a line, advancing by an increment (DX, DY) in user coords.
YHF - draws character string.
YJF - sets size of markers.
YKF - sets width and height of characters.
YLF - sets spacing between characters.
X04AFF - returns or sets the current error message unit number.

EXTERNAL ROUTINES

J06GBV - connects contour points with smooth lines.
GBY - processes data to give contour points using inverse linear interpolation.

LOCAL Scalars

INTEGER I, ICH, IFAIL, L, NCHARS, NCHTS
INTEGER IGRID, IHIOH, ILAB, MA, MB, NA, NB
INTEGER K, MDIM, NU

LOCAL Arrays

INTEGER ICHARS(70)
REAL*8 SBDAT1(17, 11), SBDAT2(17, 11), SBDAT3(16, 11)
REAL*8 WSPCE1(17, 11), WSPCE2(17, 11), WSPCE3(16, 11)
REAL*8 CHTS(22), DIST, XMAX, XMIN, YMAX, YMIN
CHARACTER*39 A, B, C, D, E

EXTERNAL J06GBU, J06GBZ

FILE HANDLING

A=’<DATA>SB. 3D. DAT’
CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT

B='*>DATA>TDUR. DAT'
C='*>DATA>LAB2'
D='*>DATA>TIME. DAT'
E='*>DATA>ERRORS'
OPEN(5, FILE=A)
OPEN(6, FILE=B)
OPEN(7, FILE=C)
OPEN(8, FILE=D)
OPEN(9, FILE=E)

SELECT OUTPUT CHANNEL FOR ERROR MESSAGES
CALL X04AAF(1, 9)

INITIALISE PLOTTING DEVICE
CALL GIND
CALL CC81
CALL J06WAF
CALL J06WDF

MA=1
NA=1
XMIN=FLOAT(MA)
YMIN=FLOAT(NA)

DO 251 K=1, 3
READ IN CONTOUR HEIGHTS
IF(K. EQ. 1) THEN
   NCHTS=7
   ASSIGN 11 TO NU
ELSE IF(K. EQ. 2) THEN
   NCHTS=8
   ASSIGN 12 TO NU
ELSE
   NCHTS=10
   ASSIGN 13 TO NU
END IF
READ<6, NU)(CHTS(I), I=1, NCHTS)

101 L=L+1

READ IN DATA FROM SEA BREEZE MODEL
IF(K. EQ. 1) THEN
   MB=17
   NB=11
   XMAX=FLOAT(MB)
   YMAX=FLOAT(NB)
   READ(5, 21) SBDAT1
ELSE IF(K. EQ. 2) THEN
   MB=17
   NB=11
   XMAX=FLOAT(MB)
   YMAX=FLOAT(NB)

C-7
C周围的等值线图使用NAG图形补充

READ(5,22)SBDAT2
ELSE
MB=16
NB=11
XMAX=FLOAT(MB)
YMAX=FLOAT(NB)
READ(5,23)SBDAT3
END IF

C. 映射数据区域到视口
CALL J06WBF(XMIN, XMAX, YMIN, YMAX, 1)
IF(L.EQ.1)THEN
CALL J06WCF(0.05D0, 0.5D0, 0.5D0, 0.8D0)
ELSE IF(L.EQ.2)THEN
CALL J06WCF(0.5D0, 0.95D0, 0.5D0, 0.8D0)
ELSE IF(L.EQ.3)THEN
CALL J06WCF(0.05D0, 0.5D0, 0.2D0, 0.5D0)
ELSE
CALL J06WCF(0.5D0, 0.95D0, 0.2D0, 0.5D0)
END IF

C. 标记每个等值线，绘制边界，并突出显示没有等值线
ILA3=1
IGRID=1
IHIGH=0

C. 绘制标题轴和标题
IF(K.EQ.1)THEN
CALL J06WBF(0.0D0, 64.0D0, 0.0D0, 3000.0D0, 1)
ELSE IF(K.EQ.2)THEN
CALL J06WBF(0.0D0, 64.0D0, 0.0D0, 3000.0D0, 1)
ELSE
CALL J06WBF(2.0D0, 62.0D0, 0.0D0, 3000.0D0, 1)
END IF
CALL J06AFF(4.0D0, 300.0D0)

IF(L.EQ.3.0D0.L.EQ.4)THEN
CALL J06AJF(1, 'SEA(Km) -> ', 30)
CALL J06AJF(1, 'LAND(Km) -> ', 30)
END IF

IF(L.EQ.1.0D0.L.EQ.3)THEN
CALL J06AJF(2, 'HEIGHT IN M', 11)
END IF

IF(K.EQ.1)THEN
CALL J06AHF('J06GBF POTENTIAL TEMPERATURE FIELD (K) ', 38)
ELSE IF(K.EQ.2)THEN
CALL J06AHF('J06GBF HORIZONTAL WIND FIELD (V) (M/S) ', 38)
ELSE
CALL J06AHF('J06GBF VERTICAL VELOCITY FIELD (W) (Cm/S) ', 41)
END IF
CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT

C_____DRAW TIMES

CALL J06YAF(4.0DO, 2850.0DO)
READ(8,444)((ICHARS(I), I=1, 6)
CALL J06YHF(ICHARS, 6)

C_____DRAW ISOPLETHS USING ALTERNATIVE CONTOURING METHOD (J06GBZ)
C AND JOINING POINTS ON CONTOUR WITH A SMOOTH CURVE (J06GBU)

IFAIL=1
ICH=1
MDIM=MB
CALL J06XGF(1.5DO, 1.5DO)
IF(K.EQ.1)THEN
CALL J06GBF(SBDAT1, MDIM, MA, MB, NA, NB, NCHTS, CHTS, ICH,
* J06GBZ, ILAB, IHIGH, J06GBU, IGRID, WSPCE1, IFAIL)
ELSE IF(K.EQ.2)THEN
CALL J06GBF(SBDAT2, MDIM, MA, MB, NA, NB, NCHTS, CHTS, ICH,
* J06GBZ, ILAB, IHIGH, J06GBU, IGRID, WSPCE2, IFAIL)
ELSE
CALL J06GBF(SBDAT3, MDIM, MA, MB, NA, NB, NCHTS, CHTS, ICH,
* J06GBZ, ILAB, IHIGH, J06GBU, IGRID, WSPCE3, IFAIL)
END IF
CALL J06XGF(1.0DO, 1.0DO)

C_____CHECK FAILURE EXIT

IF(IFAIL.EQ.0)GO TO 30
WRITE(4, 1&)IFAIL
STOP
30 CONTINUE

IF(L.LT.4)GO TO 101

C_____DRAWS CHARACTER STRINGS

CALL J06WBF(0.0DO, 100.0DO, 0.0DO, 100.0DO, 1)
CALL J06WCF(0.0DO, 1.0DO, 0.0DO, 1.0DO)
CALL J06YJF(1.5)
CALL J06YKF(0.75DO, 1.5DO)
CALL J06YLF(1.5DO, 0.0DO)

DIST=101.500
DO 100 L=1, 5
IF(L.EQ.1)THEN
NCHARS=34
ASSIGN 61 TO NU
ELSE IF(L.EQ.2)THEN
NCHARS=40
ASSIGN 62 TO NU
ELSE IF(L.EQ.3)THEN
NCHARS=10
ASSIGN 63 TO NU
ELSE IF(L.EQ.4)THEN
**CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT**

NCHARS=10
ASSIGN 64 TO NU
ELSE
NCHARS=43
ASSIGN 65 TO NU
END IF
READ(5,NU)(ICHARS(I),I=1,NCHARS)
DIST=DIST-2.5D0
CALL J06YAF(25.0D0,DIST)
CALL J06YHF(ICHARS,NCHARS)

CONTINUE

**DRAW DIAGRAM LABELS**

DIST=6.5D0

DO 300 L=1,3
IF( L.EQ.1 ) THEN
NCHARS=68
ASSIGN 71 TO NU
ELSE IF( L.EQ.2 ) THEN
NCHARS=70
ASSIGN 72 TO NU
ELSE
NCHARS=70
ASSIGN 73 TO NU
END IF
READ(7,NU)(ICHARSCI), 1 = 1, NCHARS)
DIST=DIST-2.5D0
CALL J06YAF(0.0D0,DIST)
CALL J06YHF(ICHARS,NCHARS)

CONTINUE

**DRAW ISOPLETH KEY**

CALL J06WCF(0.15D0,0.25D0,0.875D0,0.975D0)
CALL J06G2F(CHTS,NCHTS,2,7,1)
CALL J06WDF

CONTINUE

**END PLOT**

CALL J06WF
CLOSE(5)
CLOSE(6)
CLOSE(7)
CLOSE(8)
CLOSE(9)
STOP

11 FORMAT(1X,7F7.1)
12 FORMAT(1X,8F7.1)
13 FORMAT(1X,10F7.1)
21 FORMAT(1X,17F7.2)
22 FORMAT(1X,17F7.2)
23 FORMAT(1X,16F7.2)
CONTOUR PLOTS USING NAG GRAPHICAL SUPPLEMENT

16  FORMAT('J06GBF FAILS. IFAIL =',I2)
61  FORMAT(34A1)
62  FORMAT(40A1)
63  FORMAT(10A1)
64  FORMAT(10A1)
65  FORMAT(43A1)
71  FORMAT(68A1)
72  FORMAT(70A1)
73  FORMAT(70A1)
444 FORMAT(6A1)

END
VERTICAL PROFILES USING NAG GRAPHICAL SUPPLEMENT

NAG SUBROUTINES USED

JO6AAF - draws a scaled border to fit the current data region.
AHF - draws a centred title at top of data region.
AJF - draws an axis title.
CCF - draws a smooth, possibly multi-valued, curve through a set of data points.
WAF - initialises NAG graphical system and establishes mapping of initial data region onto default viewport.
WBF - declares current data region and establishes mapping of this data region onto current data region onto current viewport with margin option.
WCF - sets current viewport on plotting surface and establishes mapping of current data region onto viewport.
WDF - selects a new frame.
WZF - terminates graphical output to currently selected device.
YAF - moves pen to position (X,Y) in user coords.
YJF - draws character string.
YJF - sets size of markers.
YLF - sets width and height of characters.
X04AAF - returns or sets the current error message unit number.

GINO-F SUBROUTINES USED

BROKEN - selects broken line type.

LOCAL Scalars

DOUBLE PRECISION DIST, XMAX, XMIN, YMAX, YMIN
INTEGER I, IFAIL, K, METHOD, N, NU, NCHARS

LOCAL Arrays

INTEGER ICHARS(60)
DOUBLE PRECISION AX(100), AY(100)

FILE Handling

OPEN(5,FILE='*>DATA>SB.PROF.DAT')
OPEN(6,FILE='*>DATA>ERRORS')
OPEN(7,FILE='*>DATA>LAB3')

SELECT OUTPUT CHANNEL FOR ERROR MESSAGES

CALL X04AAF(1, 6)

INITIALISE PLOTTING DEVICE
C_____VERTICAL PROFILES USING NAG GRAPHICAL SUPPLEMENT

CALL GINO
CALL CC81
CALL J06WAF
CALL J06WDF

DO 251 K = 1, 3
KOUNT = 0
222 KOUNT = KOUNT + 1
READ(5, 25) N

C______READ IN DATA
READ(5, 30) (AX(I), AY(I), I = 1, N)
IF(K.EQ.1)THEN
   XMIN = 274.0
   XMAX = 289.0
ELSE
   XMIN = -2.0
   XMAX = 2.0
END IF
YMIN = 0.0
YMAX = 3200.0

C______MAP DATA REGION ONTO VIEWPORT
CALL J06W3F(XMIN, XMAX, YMIN, YMAX, 1)
CALL J06WCF(0.2D0, 0.7D0, 0.2D0, 0.7D0)

C______DRAW CURVE
METHOD = 1
IFAIL = 10
CALL J06CCF(AX, AY, N, METHOD, IFAIL)

C______CHECK FAILURE EXIT
IF(IFAIL.EQ.0)GO TO 50
WRITE(6, 55) IFAIL
STOP
50 CONTINUE
IF(KOUNT.LT.4)GO TO 222

C______DRAW AXES AND TITLE
IF(K.EQ.1)THEN
   CALL J06AHF(‘POTENTIAL TEMPERATURE PROFILES AT THE COAST’, 43)
ELSE IF(K.EQ.2)THEN
   CALL J06AHF(‘SOUTHERLY COMPONENTS IN SEA BREEZE AT THE COAST’, 47)
ELSE
   CALL J06AHF(‘WESTERLY COMPONENTS IN SEA BREEZE AT THE COAST’, 46)
END IF

CALL J06AAF
CALL J06AJF(2, ‘HEIGHT IN M’, 11)

C-13
C____ VERTICAL PROFILES USING NAG GRAPHICAL SUPPLEMENT

IF(K. EQ. 1) THEN
CALL J06AJF(1, 'POTENTIAL TEMPERATURE (K)', 25)
ELSE
CALL J06AJF(1, 'HORIZONTAL WINDSPEED (M/S)', 26)
END IF

C____ DRAWS CHARACTER STRINGS

CALL J06WBF(0.0DO, 100.0DO, 0.0DO, 100.0DO, 1)
CALL J06WCF(0.0DO, 1.0DO, 0.0DO, 1.0DO)
CALL J06YJF(1.5)
CALL J06YKF(0.75DO, 1.5DO)
CALL J06YLF(1.5DO, 0.0DO)

DIST=94.0DO

DO 100 L=1, 4
IF(L. EQ. 1) THEN
NCHARS=40
ASSIGN 61 TO NU
ELSE IF(L. EQ. 2) THEN
NCHARS=40
ASSIGN 62 TO NU
ELSE IF(L. EQ. 3) THEN
NCHARS=10
ASSIGN 63 TO NU
ELSE
NCHARS=40
ASSIGN 64 TO NU
END IF
READ(5, NU)(ICHARSCI), 1 = 1, NCHARS
DIST=DIST-4.0DO
CALL J06YAF(18.0DO, DIST)
CALL J06YHF(ICHARS, NCHARS)
100 CONTINUE

C____ DRAN DIAGRAM LABELS

DIST=6.5DO

DO 200 L=1, 3
IF(L. EQ. 1) THEN
NCHARS=70
ASSIGN 71 TO NU
ELSE IF(L. EQ. 2) THEN
NCHARS=70
ASSIGN 72 TO NU
ELSE
NCHARS=70
ASSIGN 73 TO NU
END IF
READ(7, NU)(ICHARSCI), I=1, NCHARS
DIST=DIST-2.5DO
CALL J06YAF(0.0DO, DIST)
CALL J06YHF(ICHARS, NCHARS)
200 CONTINUE

C-14
C____VERTICAL PROFILES USING NAG GRAPHICAL SUPPLEMENT

C____DRAW KEY
DIST=55.000
NCHARS=21

DO 300 L=1,4
READ(5,80)(ICHARS(I), I=1,NCHARS)
DIST=DIST-4.000
CALL J06YAF(60.000,DIST)
300 CONTINUE
CALL J06YHF(ICHARS,NCHARS)

251 CONTINUE

C____END PLOT
CALL J06WZF
CALL GINEND
CLOSE(5)
CLOSE(6)
STOP

25 FORMAT(IX, I3)
30 FORMAT(IX,F6.2,F7.1)
55 FORMAT(IX/I1X, 'J06CCF FAILS. IFAIL =',I2)
61 FORMAT(40A1)
62 FORMAT(40A1)
63 FORMAT(10A1)
64 FORMAT(40A1)
71 FORMAT(70A1)
72 FORMAT(70A1)
73 FORMAT(70A1)
80 FORMAT(21A1)

END
APPENDIX D

SEA BREEZE DATA
(1982-1983)
<table>
<thead>
<tr>
<th>DATE</th>
<th>WIND BEFORE SEA BREEZE</th>
<th>DIRECTION</th>
<th>SPEED (ms⁻¹)</th>
<th>TIME OF ONSET</th>
<th>OBSERVED SEA BREEZE</th>
<th>TIME OF ABATEMENT</th>
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**TABLE DI:** SEA BREEZES RECORDED AT RAF MOUNTBATTEN, PLYMOUTH APRIL TO SEPTEMBER 1982
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**TABLE D2**: SEA BREEZES RECORDED AT RAF MOUNTBATEN, PLYMOUTH APRIL TO SEPTEMBER 1983
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**TABLE D3:** SEA BREEZES RECORDED AT PLYMOUTH POLYTECHNIC METEOROLOGICAL STATION APRIL TO SEPTEMBER 1982
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**TABLE D4:**

SEA BREEZES RECORDED AT PLYMOUTH POLYTECHNIC METEOROLOGICAL STATION APRIL TO SEPTEMBER 1983