THE SEA SURFACE HEAT BALANCE
IN THE BENGUELA UPEWELLING REGION

BY

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1987
The surface heat balance of the Benguela upwelling area on the west coast of southern Africa is analysed. Measurements of the components of the heat balance were made in the St Helena Bay area from 14-21 October 1986. Additional long-term data was obtained from Alexander Bay and Cape Town.

An average net heat gain of 227 W.m\(^{-2}\) was received over the eight days of the field study. The presence of cold water determined that latent heat loss by the sea surface was small, while the sensible heat flux represented a small gain by the sea. These two turbulent heat fluxes are roughly equal and opposite and therefore approximately cancel each other. Use of a model, assuming idealised conditions, indicated that most turbulent heat exchange between the air and takes place in the nearshore region where air-sea contrasts are greatest. The net radiation was found to provide a good estimate of the total heat balance, thus the major contributing term to a high heat balance over the Benguela area is the input solar radiation.

Minimal synoptic variation in the heat balance during the eight-day field programme was observed, but additional global radiation data analysed revealed that synoptic variations over the 3-6 day period are in fact more significant than the longer term seasonal variations. Both synoptic and seasonal variations in the heat balance are greater in the south than in the north.

The high heat flux into the sea surface is capable of increasing the temperature of the upwelled water at a fairly rapid rate. During summer the heat exchange is capable of increasing the temperature of the upper 10 m mixed layer by as much as 0.65°C over one day. This input heat is used to realise the high biological potential of the upwelled waters.
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The direction of the heat exchange between the ocean and the atmosphere has important implications for global climatology. A vital contrast in this heat exchange is evident between the eastern and western boundary currents of the world. The eastern boundary currents which are dominated by coastal upwelling have been found to receive greater amounts of heat from solar radiation and the atmosphere than their more studied western counterparts. This study provides a detailed investigation of the heat budget over the Benguela upwelling system.

A knowledge of the heat budgets over the oceans is essential for a complete understanding of the oceanographical and meteorological interactions that take place. The atmosphere and oceans are intricately linked and mutually interact over a variety of scales. Benton, et al. (1963) stated that "the atmosphere and oceans constitute a single mechanical and thermodynamical system of two coupled fluids and they interact in a manner which is so complex that cause and effect cannot always be distinguished". The ocean-atmosphere system thus essentially regulates its own behaviour through the application of internal feedback mechanisms.

The heat balance at the sea surface indicates the net heat gain or loss by the surface from the sum of incident solar radiation, back radiation, sensible heat conduction and evaporative heat loss. A
The major aim of this thesis is to examine and characterise the observed heat transfer processes, relevant to the overall heat balance, between the ocean and atmosphere in the coastal upwelling environment of the Benguela system. A significant aspect of the study concerns the relative importance and magnitude of each of the various terms contributing to a high positive heat balance over the upwelled waters. It is necessary to study the exchange of heat between the atmosphere and ocean over the cold subsurface upwelled water in order to investigate the degree of change in the temperature of the cold water brought about by the contributions of radiation and air-sea interactions, and to examine the degree of heat transfer to the atmosphere.

All of the important processes by which the oceans gain or lose heat operate only at the surface of the sea. Even the small scale processes, such as wind stress, evaporation and heat conduction, are important for the operation and maintenance of all atmospheric and oceanic circulations. At the sea surface the flux of heat is influenced by environmental conditions in the sea, such as the sea surface temperature (SST), and by conditions in the atmosphere, such as...
as air temperature, humidity, wind speed and cloud cover. In addition, advective processes in the sea, along with the effect of the high thermal inertia of water, indicate that the long-term balance can depend on the regional patterns of warm and cold currents.

Prevailing climatic conditions generally determine the main characteristics of the local energy flux, but synoptic events can cause variations on this general theme. Over the west coast of southern Africa environmental parameters on the synoptic time scale of a few days exhibit larger ranges than those experienced over the longer seasonal time scale. It was noted by Perry and Walker (1979) that short-term fluctuations in individual terms of the heat balance result from changes in air and sea surface temperatures over various time scales, so that the energy transfer processes do not proceed continuously but in an interrupted, pulsating form. Certain environmental conditions during the synoptic cycle can result in unusually high heat gains at the sea surface of the west coast. It is therefore of interest to measure and investigate the variability of the major heat transfer processes over the Benguela region on the shorter synoptic time scale of a few days and to explore the consequences thereof. The investigation also focuses on the possibility that an unusually high heat flux into the surface waters of the Benguela Current may have a profound effect on the physical, dynamic and biological characteristics of the surface waters.

In subsequent chapters an historical background to relevant published works consistent with the scope of this study is given. A discussion of the heat transfer processes pertinent to the heat
budget study is then given, in which aspects such as the usefulness and validity of the applied equations are dealt with. The study area and influence of climatological and oceanographical processes is introduced on various temporal and spatial scales. In addition, the heat balance and ongoing physical processes over the Benguela area and other parts of the South Atlantic Ocean are considered. Attention is also given to the inherent similarities between the Benguela and other upwelling sites. On a slightly smaller spatial scale, the longshore and offshore variability in the sea and air along the west coast is dealt with.

On the synoptic scale, attention is given to the weather systems which control the atmospheric and oceanographic climate of the Benguela area, as this is instrumental in determining the heat budget of the area. The Benguela upwelling process, a result of periodic atmospheric forcing, is also given due consideration. On a smaller scale the modelling of inherent offshore and onshore variations in the heat transfer processes for a study area typical of the Benguela under idealised conditions is considered.

A suitable area typifying an upwelling environment was chosen for an eight-day field measurement programme. Details of the localised study site at St Helena Bay (32°45'S 18°05'E) are given, as well as the instruments used and methods employed for data extraction. The observed results from the study area are dealt with in a fair amount of detail.

The discussion chapter begins with the interpretation and usefulness of the field programme results. An investigation into the synoptic
variability in the heat balance over the Benguela upwelling area is made. Because longshore variations in the air and sea are found to be weak in comparison to the offshore variations, the results of the study are extended to characterise the heat transfer processes relevant to the entire Benguela system. This is followed by an investigation into the physical and biological consequences of a large amount of heat being available for heating of the surface waters of the Benguela system as a whole. Relevance of the results gained from the field programme and from additional long-term data to the Benguela climatology as a whole is considered.

The concluding chapter summarises the findings of the study. Similarities between the Benguela system and other upwelling systems of the world are used to justify an extension of the results obtained to typify the heat balance over other upwelling areas. Finally, recommendations for future research are made.
In comparison to other aspects of oceanography, the field of energy exchange at the sea surface appears to have received scant attention relative to the important role it has in maintaining the world's energy balance. The first part of this historical review considers the research specific to the study of heat budgets that has been accomplished. Thereafter, the study of heat budgets at more localised upwelling areas is considered. This is followed by a synopsis of the research which is applicable to the study of the Benguela upwelling system.

Most of the studies involving the heat budget of the surface layer of the ocean, and the exchange of energy between the ocean surface and the atmosphere, have concentrated on events occurring over fairly long time scales, such as the seasonal, and have considered annual averages occurring over large spatial scales, such as the global, oceanic and latitudinal. One of the first detailed accounts of the heat budget of the oceans was that given by Sverdrup, et al. (1942). This included a good introduction to the various physical processes which contribute to heating and cooling of the sea surface. Comparisons were made between the heat budgets of the different oceans, and seasonal, as well as general diurnal, fluctuations of the constituents of the heat budget were dealt with.

In the article entitled "On the Energy Exchange between the Sea and Atmosphere", Jacobs (1942) introduced a more accurate bulk aerodynamic method for the evaluation of evaporation and sensible heat exchange, in order to compute the annual and seasonal energy
fluxes over the North Atlantic and North Pacific Oceans. Comparisons were made of the terms of the heat budget over large areas (latitudinal and oceanic) and long time scales (seasonal) and attention was given to the fact that low evaporation is associated with the northerly flowing cold currents from the southern hemisphere. Recognition was also given to the suggestion that for every major meteorological change over the oceans, a corresponding change in the energy relationships between the sea and atmosphere must result.

A further notable attempt to determine the climatological distribution of various aspects of the energy budget of the oceans was made by Budyko (1963) in his world atlas of the heat balance. Included in this long-term and large scale study was a more localised study of the region affected by the Benguela Current, to illustrate the annual variation of the heat budget components typical of an eastern boundary current area. A large overall heat gain was observed due to a high radiation balance, a small sensible heat gain and only a small latent heat loss. Particularly large heat expenditures on heating the cold water masses transported by the current were noted in summer. Pronounced annual variations in the radiation balance were also observed and were ascribed to the annual march of the sun. The accuracy of the monthly values calculated by Budyko for the oceans of the world is, however, uncertain as the data was obtained from a variety of sources, differing in time and method of observation.

Privett (1960) was the first to concentrate on the seasonal and regional variations of energy transfer at the ocean surface in the
southern hemisphere. On a broad scale general agreement with the results gained by Jacobs (1942) and Budyko (1963) was shown, even though computational methods differed. A similar study of the radiation budget of the southern hemisphere performed by Sasanori, et al. (1972) and Newton (1972) indicated a high average radiation over the west coast of southern Africa.

In recent years the study of heat budgets, in particular the air-sea energy exchange processes, has received increased attention. A greater accuracy in the assessment of the turbulent heat transfer has evolved with the development of more accurate estimates of the bulk transfer coefficients, which apply to the bulk aerodynamic formulae. Parameterization of the air-sea heat fluxes through use of the bulk aerodynamic formulae allows for the opportunity to calculate heat fluxes wherever and whenever complete weather and sea surface temperature observations are made.

The first to practically use bulk exchange coefficients which varied with wind speed and stability to calculate the turbulent heat fluxes was Bunker (1976), in his study of energy fluxes through the surface of the North Atlantic Ocean. Some attention was given to local regions within the North Atlantic, such as the Spanish Sahara upwelling region. It was found that the effect of the cold water was to increase the heat gain through decreasing the evaporation. The eastern and south-eastern parts of the Atlantic were also noted to absorb more heat from the sun than they lost through the sum of evaporation, conduction and infrared radiational exchange.

Using a constant drag coefficient for the bulk method, Hastenrath
(1977a & b) calculated all the components of the heat budget derived from conventional ship observations from the Tropical Atlantic and Eastern Pacific Oceans during 1911-1970. More localised areas such as the cold Canary, Benguela and Humboldt Current domains, and the warm Gulf Stream and Brazil Current, were studied in order to gain a better understanding of the zonal asymmetries between the eastern and western parts of the world’s oceans. The cold ocean currents off the west coasts were found to stand out as regions of downward directed sensible heat flux and minimal evaporative heat loss, contrasting with the larger exchange from ocean to atmosphere in the warm current domain.

The most recent calculations of long-term monthly and annual means of the energy budget constituents for the global ocean were those made by Haung (1986), through the use of a comprehensive data set. Although the heat flux distributions are more accurate than the earlier estimates, the results still show a great deal of agreement with previous studies.

Only a general indication of local conditions can be derived from the long temporal and spatial scales adopted in the literature reviewed above. Lately there has, however, been an increased interest generated in the more localised treatment of heat budgets in space and time. It is necessary to investigate short-term changes, as the processes which operate on the hourly or daily periods, in fact, determine the seasonal and long-term trends.

Attempts at studying the heat budget on shorter time scales at more localised upwelling areas have been made mainly at the coastal areas
Off California and North West Africa. Off the coast of Oregon
(California) insolation and heat budgets have been examined by,
and Tont (1975 and 1981). Tont (1975) found that increased
upwelling during summer off the Oregon coast resulted in increased
cloudiness and hence decreased solar irradiance. Upwelling of cold
water acts to decrease the temperature of the air above (Tont,
1981). After examining data from a time series station in the
Oregon area, Reed & Halpern (1975) concluded that, during intense
upwelling, heat content changes of the mixed layer were influenced
more by horizontal advection and diffusion than by surface heat
exchange. However, off the Spanish Sahara, Halpern & Reed (1976)
found that the contributions of horizontal advection and diffusion
to the heat balance were negligible.

Investigations into the heating of the surface waters of the North
West African upwelling area by Richman & Badan-Dangon (1983)
indicated that solar heating is an obvious source of heat to replace
the offshore heat transport of the previously upwelled waters. A
study of the same area by Bowden (1977) showed that a high net heat
gain by the ocean surface was sufficient to cause an appreciable
rise in temperature and corresponding decrease in density of the
upwelled water. It was also suggested that the change in
temperature of the water in relation to the rate of heating through
the surface could be used to estimate the upwelling velocity.

Off Cabo Nazca, Peru, Stevenson, et al. (1981) provided an insight
into the short-term variations in the circulation, heat content and
surface mixed layer of upwelling plumes. The presence of warmer
water surrounding a coastal upwelling area is attributed to cumulative solar heating of the coastal upwelled water as it moves away from the area of the plume.

The Benguela system remains relatively unexplored from the purely physical point of view, most of the literature having a physical component which is merely in support of fisheries science (Nelson & Hutchings, 1983). Physical features of the Benguela system, including the macroscale meteorology and mesoscale coastal upwelling processes have been considered by Nelson & Hutchings (1983). A comprehensive study of the physical features and processes influencing the Benguela system was more recently compiled by Shannon (1985). Few researchers have specifically concentrated on the heat balance of the area.

Systematic measurements of total and diffuse radiation over southern Africa have been made by the Weather Bureau since 1951 and data has been published in the following forms: Quarterly Radiation Bulletin (hourly data), Annual Radiation Report (hourly data) and Solar Radiation and Sunshine (summaries 1951-1962). An analysis of the incoming short wave radiation over South Africa by Maaren (1976) revealed that the annual income of radiation over South Africa is high compared to the rest of the world.

Hoflich (1984) dealt in depth with the surface heat balance in a study of the "Climate of the South Atlantic Ocean". Seasonal comparisons (January and July) were made between all the components of the heat budget for the entire South Atlantic Ocean. The more localised treatment of offshore areas, such as off South West
Africa, was dealt with in a greater amount of detail. Map distributions produced by Hoflich indicate that the largest heat gain by the South Atlantic Ocean occurs in the region off the west coast of southern Africa. A detailed discussion of Hoflich's results is considered in a subsequent chapter of this thesis.

To date, the only relatively detailed study of the heat budget of the Benguela upwelling region appears to be that done by Hughes (1984, unpubl.). Physical data for three typical days and one night, acquired from a cruise in St Helena Bay, was used in order to calculate the local heat budget. An attempt was subsequently made to relate the heat content of the water column to the amount of input heat energy available, in order to determine the residence time of the water in the bay. Limited reference was made to the implications of this by Waldron (1985). Although the Benguela upwelling system is a dominant feature of the west coast of South Africa, aspects of the heat budget of the area thus far appear to have received minimal attention. A comprehensive heat budget study of the area is thus long overdue.
HEAT TRANSFER PROCESSES BETWEEN THE OCEAN AND ATMOSPHERE

The equation used to determine the net heat gain (or loss) by the water through a unit area of sea surface can be represented by:

\[ Q_t = Q_r - Q_e - Q_h \]  

(1)

Where \( Q_t \) represents the total heat flux into/out of the sea
\( Q_r \) " net radiation into the sea
\( Q_e \) " latent heat flux from the sea
\( Q_h \) " sensible heat flux into/out of the sea

This equation essentially represents the exchange of energy between the ocean and the atmosphere. A positive value of \( Q_t \) indicates a heat gain by the sea, while a negative value indicates a heat loss by the sea and a gain by the atmosphere. The values of the terms in equation (1) above can be computed according to standard equations.

Incoming Radiation

The radiation balance is represented by:

\[ Q_r = (Q_I + q)(1 - \alpha) - \rho_b \]  

(2)

(Sellers, 1963).

Where \( Q_I \) = Incident solar beam
\[ q = \text{Diffuse solar radiation} \]
\[ \alpha = \text{Sea surface albedo} \]
\[ Q_o = \text{Long wave back radiation from the sea surface} \]

The term \((Q_i + q)\) represents the global radiation, which can be defined as the total amount of radiation received from all parts of the horizon on a flat surface. "Solar radiation is the main source of heat energy for almost all the natural processes developing in the atmosphere, hydrosphere and in the upper layers of the lithosphere" (Budyko, 1985), therefore the amount of solar radiation received at the surface is extremely significant to the heat budget of an area. Other sources of heat available to the ocean, such as the heat influx from the interior of the earth, heat gained from radio-active decay, frictional heat and heat given off by chemical-biological processes, are of minor importance in comparison with the input from solar radiation.

The amount of incoming solar radiation received at the earth's surface differs locally, depending on:

(1) the sun's altitude. This depends on the hour of day, the season and the geographical latitude of the area concerned. When the sun is set lower in the sky, solar radiation needs to pass through a greater thickness of atmosphere in order to reach the earth's surface, and is therefore subjected to a greater degree of attenuation in the atmosphere. This, coupled with the fact that when the sun angle is more acute, the oblique sun's rays are distributed over a larger area, means that less solar radiation is available for heating at the surface than when the sun is situated
closer to the zenith. The hours of sunshine received each day is also a significant factor and is dependent on season and geographical latitude.

(ii) attenuation in the atmosphere. The electromagnetic spectrum of solar radiation which traverses the atmosphere consists of a wide range of wavelengths. However, each of the numerous gases present in the atmosphere is capable of absorbing or reflecting radiation of distinctive wavelengths. Fortunately the atmosphere is relatively transparent to the short wave radiation (wavelengths < 4 µm), which comprises 99% of the sun's energy (see Fig. 1). Part of the radiant energy passing through the atmosphere is, however, absorbed, scattered and reflected by gas and cloud particles, dust, haze, smoke, etc. Thus the amount of radiant energy reaching the earth's surface is dependent on the nature and amount of absorbing, reflecting and scattering matter present in the atmosphere. Those gases present in the atmosphere in the smallest amounts, such as water vapour, carbon dioxide, oxygen and ozone are, in fact, responsible for absorbing the greatest portion of radiation.

Water vapour, which absorbs strongly in the 1-6.5 µm band and at wavelengths greater than 18 µm, usually comprises less than 3% of the gases in the atmosphere, but is capable of absorbing almost six times as much solar energy as all the other gases combined (Barry & Chorley, 1976). Of the incident radiation absorbed in the atmosphere, an average of

55% is absorbed by water vapour,
17% " " clouds,
Figure 1. Atmospheric absorption in clear air for solar radiation with a zenith angle of 50° and for diffuse terrestrial radiation at (a) the portion of the atmosphere lying above the 11-km level (mid-latitude tropopause), and (b) at ground level (after Wallace & Hobbs, 1977).

Figure 2. Absorption of solar radiation by different components of the atmosphere (data obtained from Newton (ed), 1972).
These estimates are depicted in Figure 2. An estimate of the average percentage of incoming radiation that reaches the earth's surface is given in Figure 3.

Unlike the other gases, which are more constant in quantity in the atmosphere, water vapour varies greatly from place to place, from percentages as low as 0.02% in desert regions to as high as 1.8% in humid regions. The absorption of radiation by water vapour thus varies greatly according to climatic conditions. The amount and type of cloud cover present in the atmosphere is another important factor in determining the amount of solar energy available at the surface because radiation can be reflected, absorbed and scattered by cloud particles. Under heavy overcast conditions the combined absorption and reflection of solar energy by clouds can prevent 35-80% of the incoming radiation from reaching the earth's surface, while on a clear day absorption and reflection by gas particles may only prevent 20% of the incident radiation from reaching the surface.
Figure 3. The annual mean radiation and heat balance of the earth-atmosphere system, relative to 100 units of incoming solar radiation (adapted from "Understanding Climatic Change", National Academy of Sciences, 1975).
Reflectance of solar radiation from the sea surface

The solar radiation that reaches the sea surface can either be absorbed by the sea or reflected from its surface. The albedo of the sea surface, which represents that portion of the incoming radiant energy that is reflected, is low in comparison to land surfaces and depends largely on the sun's altitude and the sea state. Values for the albedo of the sea surface in calm conditions are given in Table 1(a). For rough seas, defined by Danard, et al. (1983) as "those seas where the roughness length exceeds 10^{-3} \text{ cm}, corresponding to a 10 \text{ m wind speed of } 2 \text{ m.s}^{-1} ", the albedo changes and a correction factor needs to be added to the calm sea values.

When the sun is high above the horizon, the incident radiation is more likely to strike a sloping wave face, resulting in greater reflectance from the surface. However, when the sun is closer to the horizon, less reflectance of the incident rays striking the sloping waves will take place. These changes in the albedo are not marked and do not affect the absorption values to any significant degree. Table 1(b) indicates the corrections to be added to the values given in Table 1(a) to account for the wind disturbance of the surface.
Table 1(a). Albedo of the sea surface for calm seas

<table>
<thead>
<tr>
<th>ALTITUDE OF SUN</th>
<th>ALBEDO</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°</td>
<td>1.000</td>
</tr>
<tr>
<td>10°</td>
<td>0.350</td>
</tr>
<tr>
<td>20°</td>
<td>0.135</td>
</tr>
<tr>
<td>30°</td>
<td>0.061</td>
</tr>
<tr>
<td>40°</td>
<td>0.035</td>
</tr>
<tr>
<td>50°</td>
<td>0.025</td>
</tr>
<tr>
<td>60°</td>
<td>0.022</td>
</tr>
<tr>
<td>70°</td>
<td>0.021</td>
</tr>
<tr>
<td>80°</td>
<td>0.021</td>
</tr>
<tr>
<td>90°</td>
<td>0.021</td>
</tr>
</tbody>
</table>

Table 1(b). Correction $\Delta a$ to be added to albedo to account for sea surface roughness (winds $> 2 \text{ m.s}^{-1}$)

<table>
<thead>
<tr>
<th>ALTITUDE OF SUN</th>
<th>$\Delta a$</th>
</tr>
</thead>
<tbody>
<tr>
<td>10°</td>
<td>-1.000</td>
</tr>
<tr>
<td>20°</td>
<td>-0.015</td>
</tr>
<tr>
<td>30°</td>
<td>0.010</td>
</tr>
<tr>
<td>40°</td>
<td>0.018</td>
</tr>
<tr>
<td>50°</td>
<td>0.025</td>
</tr>
<tr>
<td>60°</td>
<td>0.029</td>
</tr>
</tbody>
</table>
Back Radiation

The long wave (3-80,\mu m) thermal radiation emitted from the sea surface \((Q_b)\) is radiated out at an intensity proportional to the absolute temperature of the sea surface. Owing to the fact that long wave radiation measurements are not easily or often accomplished, the \(Q_b\) term is usually computed with the aid of empirical formulae, which use primarily sea surface temperature (SST) and vapour pressure as input variables. A variety of formulae to compute the back radiation from the sea surface are used in the available literature. Difficulties in the choice of the most accurate formula have been referred to by Weare, et al. (1981). The most widely used appear to be those formulated by Brunt (given by Budyko, 1958), Berliand (given by Budyko, 1963), Krauss and Rooth (1961), Efimova (given by Reed, 1976) and Lonngvist (given by Pickard & Emery, 1982).

By comparing measured back radiation values to those calculated from the Berliand formula, Reed (1976) discovered that the calculated values overestimated the backward irradiance. Efimova's formula was found to provide acceptable estimates of the oceanic net long wave radiation, but the formula is not always valid (Reed, 1976). After careful consideration the following formula given by Krauss & Rooth (1961) and used by Bowden (1977), was selected for the purposes of this study:

\[
Q_b = \varepsilon \sigma T_s^4 (0.39 - 0.05 \sqrt{e_a})(1 - 0.6c^2)
\]  

(3)

Where \(\sigma = \text{Stefan-Boltzmann constant} = 5.6696 \times 10^{-8} \text{ W.m}^{-2}.\text{deg}^{-4}\)
\[ \varepsilon = \text{emissivity of sea surface} = 0.985 \]

\[ T_s = \text{SST in } ^\circ\text{C} \]

\[ e_a = \text{vapour pressure of air} \]

\[ C = \text{cloud cover as a fraction} \]

From the formula above it is evident that the SST determines the rate of outward emission of the thermal energy, but the vapour pressure present in the overlying air acts to reduce the radiation by itself radiating by virtue of its temperature. An increase in SST is accompanied by an exponential increase in the humidity of the overlying air. Because the rate of increase in the humidity of the air exceeds the rate of increase of the saturation specific humidity of the sea surface, the long wave radiation of the water vapour in the atmosphere to the sea increases at a rate greater than the long wave radiation from the sea surface to the atmosphere. Thus, contrary to what might be expected, an increase in SST can actually result in a decrease in \( Q_b \). Clouds act to decrease the effective back radiation from the sea surface because the water droplets present in the clouds block the outward passage of long wave radiation and the cloud bases radiate back to sea (see Fig. 1 & Fig. 3).
The turbulent heat fluxes

(i) Latent Heat Flux \((Q_a)\)
The process of latent heat transfer from the sea surface to the atmosphere above is an important aspect of the heat budget, but is difficult to accurately determine. Heat that is lost by the ocean during the process of evaporation is transferred to the atmosphere in the form of latent heat, which is then released into the atmosphere during the process of condensation. Evaporation can occur whenever the air immediately above the sea surface is unsaturated with water vapour, and therefore takes place whenever there is an upward vertical gradient of specific humidity (or vapour pressure).

In order for evaporation to take place, heat must be supplied either from an outside source, or from the liquid itself. In the case of the ocean, the energy for evaporation is supplied from the ocean water. According to Mc Lellan (1965), on a global average, approximately half of the radiant energy absorbed by the waters of the ocean is returned to the atmosphere by evaporation.

(ii) Sensible Heat Flux \((Q_h)\)
The temperature of the sea surface usually differs from that of the overlying air so that a temperature gradient exists between the two. Depending on the direction of the temperature gradient, heat can either be lost or gained by the sea due to conduction and turbulent heat exchange. This exchange of energy is the sensible heat flux term, \(Q_h\), which is usually of a much smaller magnitude than the \(Q_a\) term (i.e. \(\pm 10\%\) thereof). Where the air temperature gradient
decreases upward from the sea surface, heat is conducted out of the sea and $Q_h$ is a loss term. If the air temperature increases upward from the sea surface heat is conducted from the air to the sea and $Q_h$ is a gain term in the heat budget.

Both of these vertical energy transfers, referred to by some authors as the turbulent heat fluxes, are important components of the heat budget of the ocean, but their direct measurement requires complicated and expensive technical equipment. The latent and sensible heat fluxes are therefore computed from equations called the bulk aerodynamic formulae, which are based on the largely empirical results of research on atmospheric turbulence (Privett, 1960). At present the bulk aerodynamic method is the only practical means of calculating the turbulent heat fluxes from measurable variables. Meteorological parameters which can be measured from shipboard with simple instruments can be used for the determination of the vertical transports and the decisive processes are parameterized (Dietrich, et al, 1972). Fortunately the calculation of these fluxes over the sea is more simplified than over the land due to the fact that the sea surface is relatively uniform compared with land surfaces and standard SST's can be observed. The fluxes are determined with the use of transfer coefficients which relate the fluxes to the variables measured (Liu, et al. 1979).

The accepted bulk aerodynamic formulae for estimating the latent and sensible heat fluxes are respectively

$$Q_e = L f C_e u (q_e - q_a) \quad (4)$$
and

\[ Q_h = \rho c_p c_h u (T_s - T_a) \]  \hspace{1cm} (5)

where

\[ L = \text{latent heat of vaporisation} = 2494 - 2.2 (T_s) \text{ kJ.kg}^{-1} \]

(Pickard & Emery)

\[ \rho = \text{air density} = 1.2 \text{ kg.m}^{-3} \]

\[ c_p = \text{Specific heat of air at constant pressure} = 1004 \text{ J.deg}^{-1} \]

\[ c_e = \text{latent heat transfer coefficient} \]

\[ c_h = \text{sensible heat transfer coefficient} \]

\[ u = \text{wind speed (m.s}^{-1}) \text{ at height h} \]

\[ q_s = \text{specific humidity (g/kg) at sea surface} \]

\[ q_a = \text{specific humidity (g/kg) of air at height h} \]

\[ T_s = \text{SST (°C)} \]

\[ T_a = \text{air temperature at height h (°C)} \]

These formulae are valid only for log vertical wind profiles and a reference height, h, of 10 m above sea level. The transfer coefficients \( c_e \) and \( c_h \) are usually assumed to be approximately equivalent and equal to the drag coefficient for momentum transfer, \( C_d \) (Friehe & Schmidt, 1976; Fissel, et al. 1977; Gadd & Keers 1970; Grachev and Panin, 1984; Liu, et al. 1979; Bunker, 1976; etc.). Blanc (1985) discovered that over 20 bulk transfer coefficient schemes (or combinations of schemes) are encompassed within the bulk method.

Many authors, among them Friehe & Schmidt (1976), Budyko (1963), Hastenrath & Lamb (1978) and Fissel et al. (1977), have assumed that the transfer coefficients are constant. However, the values of these
constants tend to vary from author to author, from as low as 0.66 \times 10^{-3} for a stable environment (e.g. Large & Pond, 1982) to as high as 2.1 \times 10^{-3} (e.g. Budyko, 1963). The reason for the discrepancies between the values probably lies in the fact that the experimental data used for verifying the coefficients is taken from different oceanic regions. Measurement techniques may also differ. There is no single, universally accepted bulk transfer coefficient scheme (Blanc, 1985).

Bunker (1976) and Kondo (1975) developed bulk coefficients which take the wind speed and variable roughness of the sea surface, as well as the atmospheric stability into account. The values given by Bunker appear to underestimate the fluxes, while those given by Kondo were found to have a satisfactory agreement with direct observations (Kondo, 1977). The values given by Kondo were therefore selected for the purposes of the study and are illustrated in Table 2. Both turbulent heat fluxes decrease with decreasing wind speed as a result of diminished mixing or turbulence and weaker advection. Under calm conditions, with no wind, zero values for the fluxes result. Uncertainty still exists as to the degree of accuracy of the empirically based bulk aerodynamic formulae and further verification of the calculated fluxes is required. There is at present no way of knowing which of the bulk schemes, if any, is correct, but there is currently no alternative but to use the bulk method (Blanc, 1985). However, the errors associated with the fluxes are fixed and do not affect comparison studies among different time periods or areas.

<table>
<thead>
<tr>
<th>$u \text{ (m.s}^{-1}\text{)}$</th>
<th>$c_d \times 10^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.3 - 2.2</td>
<td>$1.08u^{-0.15}$</td>
</tr>
<tr>
<td>2.2 - 5.0</td>
<td>$0.771 + 0.0858u$</td>
</tr>
<tr>
<td>5.0 - 8.0</td>
<td>$0.967 + 0.0667u$</td>
</tr>
<tr>
<td>8.0 - 25.0</td>
<td>$1.2 + 0.025u$</td>
</tr>
<tr>
<td>25.0 - 50.0</td>
<td>$0.073u$</td>
</tr>
</tbody>
</table>

The magnitude and direction of the air-sea fluxes is dependent on the temperature difference between the air and ocean, which in turn determines the stability and amount of vertical convection in the air above. Where the water is colder than the overlying air, as in the case of upwelling areas, the sensible heat flux is directed downwards from the air to the sea surface. The lowest air layers, which are in contact with the sea surface, become denser and are cooled from below, thereby inhibiting convection. This has the effect of stabilizing the air and reducing the degree of turbulence and wind stress and therefore also has the effect of further decreasing the sensible and latent heat fluxes. Evaporated water contained in the lowest air layers is trapped by the stable air layers, resulting in an increase in the relative humidity of the
air. The air is thus unable to acquire more water vapour and the latent heat flux is consequently reduced.

Air above warmer ocean water is heated from below, expands and rises and becomes convectively unstable, carrying heat away from the sea surface in the process. Due to the direction of the temperature gradient, sensible heat is transferred away from the sea surface, assisted by upward convection. Water vapour is transported along with the upward turbulent motions of the air and the relative humidity is not allowed to increase above a threshold value of approximately 78% (Flohn, 1982). A gradient in specific humidity (or vapour pressure) from the sea surface to the air above is set up, allowing for continued evaporation to occur.

Over the oceanic areas the temperature of the ocean surface is greater than that of the air and the specific humidity (or vapour pressure) at the surface is similarly greater than that of the air. Thus the sea usually loses heat to the atmosphere through sensible heat conduction and evaporation. Areas with low evaporation are those with relatively small sea-air temperature differences, or, in exceptional cases, where the SST is less than the air temperature, such as over upwelling cold water regions.
Summary of the Heat Balance

Generally speaking, the heat budget is negative over warm currents due to the loss of heat from the sea surface to the atmosphere by chiefly turbulent latent heat transfer. This is a consequence of both positive vertical temperature and humidity gradients in the overlying air. Most of the net radiation available at the surface is therefore used for the process of evaporation, leaving little or no radiant energy to heat the surface waters.

The negative vertical temperature gradient and resulting stable stratification of the air layers above cold water results in diminished evaporation rates over areas with cold ocean currents. The heat budget therefore tends to be positive in those regions and the radiant heat is available to be stored in the ocean.
B. PHYSICAL PROCESSES INFLUENCING THE BENGUELA SYSTEM

Section Overview

This part of the thesis gives a physical background to the study area in terms of orographic, bathymetric, climatological, and oceanographic influences on various scales. The large scale processes deal with the long-term physical effects over the west coast area as a whole. Aspects such as the general climatology and macroscale (+1000 km) semi-permanent pressure controls over the area are referred to, as well as interannual and seasonal variations of the physical controls on upwelling.

Continuing on this large scale theme the general oceanographic and climatological conditions and their influence on the heat balance of the Benguela system as compared to other regions of the South Atlantic Ocean are considered. In a similar context long-term similarities between the Benguela system and other upwelling sites of the world receive mention.

Breaking down the scales and superimposed on the macroscale influences are physical processes on the synoptic scale (100 - 1000 km), which occur over a few days. This section includes a synopsis of weather systems affecting the area and the short-term influences on upwelling. Longshore variability and, on a smaller scale, offshore variability in the ocean and overlying air are also considered. This leads to an even smaller scale (+100 km) approach in terms of the inherent offshore and onshore variations in the heat transfer processes. A model of these smaller scale variations is
used as an illustration.

The heat balance and its components are considered at a specific site on the west coast in order to typify the heat balance of the Benguela area. Because the physical influences need to be taken into account, local aspects such as the surrounding topography and orography and their controls on the microclimate and upwelling are discussed. The features of this small scale, typical west coast site are then extended longshore to categorize the heat transfer processes typical of the Benguela system as a whole.
1. **LARGE SCALE PROCESSES**

Although some variability is evident between the northern and southern parts of the Benguela system, the upwelling-dominated west coast of southern Africa is uniform in nature on a broad scale. The typical topography of the western coastal area consists of a relatively narrow coastal plain which ascends 50-200 km inland to the continental escarpment. Much of the coastal region north of about 32°S is dominated by sand dunes and experiences an arid climate, classified by Hoflich (1984), according to the Koppen system, as a dry trade climate, type BW. Topographically the regular coastline to the north of approximately 32°S differs from that south of 32°S, the latter being more irregular, punctuated by several headlands and northward sweeping log-spiral shaped bays. The climate in the south, classified as temperate Mediterranean, type C, is characterised by warm, dry summers and cool, rainy winters.

**Climatic Controls**

The pressure systems which are regarded as exerting the strongest influence on the west coast area are the South Atlantic Anticyclone (SAA), eastward moving cyclones to the south associated with perturbations in the subtropical jet-stream, and the inland pressure over the subcontinent. The large, semi-permanent South Atlantic high pressure system to the west of the sub-continent exerts the most significant control on the climate of southern Africa. Centred at approximately 30°S 05°W during summer, the SAA
migrates some 6° north-west during winter, allowing the free zonal passage of eastward-migrating sub-tropical cyclones over the country. During summer these sub-tropical cyclones are usually forced south of the country by the SAA.

The South Atlantic High, in conjunction with a contrasting heat low which forms over the interior during summer, is responsible for the steep zonal pressure gradients along the west coast, which atmospherically force strong southerly upwelling-favourable winds. At the coastal belt both the orographical barrier and the thermal barrier to cross-flow caused by the desert-like conditions inland, contribute to forcing the curved anticyclonic flow associated with the SAA to be southerly in direction and therefore upwelling-favourable. These southerly winds are seen to be relatively consistent along the west coast during summer (see Fig. 4a).

During winter, however, the picture is somewhat different as the northward shift of the pressure systems allows for the passage of sub-tropical cyclones over the southern Benguela area. These are accompanied by north-westerly winds which effectively suppress upwelling (see Fig. 4b). The effect of the westerlies in winter can be felt as far north as the Orange River. North of about 31°S to just north of Luderitz less seasonal variation in the macroscale wind field is evident and the southerly winds blow consistently. Although the northward shift of the SAA causes a strengthening of equatorward winds over the northern Benguela during winter, seasonal variations are not as pronounced over this area as in the south.

An additional significant climatic feature of the west coast is that
Figure 4. Mean wind speed vectors for (a) summer, and (b) winter (after Kamstra, 1985).
of the strong presence of temperature inversions. Subsiding air associated with the strong SAA, in conjunction with the presence of the cold Benguela water, is largely responsible for the high frequency and strength of non-surface and surface inversions at low levels over the west coast. The dry, subsided air on the eastern side of the SAA is separated from the cool, moist, maritime air near the sea surface by a marked semi-permanent inversion of an average 5°C intensity and 500 m depth at a height of approximately 400 m to 1200 m above the sea surface (Preston-Whyte, et al. 1973). An increase in frequency and intensity and a lowering of the inversion base is evident during the summer months when the SAA is situated closer to the sub-continent. The subsidence inversion, which always tends to be strongest over the west coast, has been found to extend a considerable distance along the coast and out to sea (Diab, 1977).

At the surface the cold upwelled waters of the Benguela system create a stable, stratified surface layer in the atmosphere. According to Tyson, et al. (1976), only on the cold-water Benguela coast is surface cooling sufficient to ensure the occasional occurrence of some midday surface inversions. These inversions persist due to the low-level advection of cool air in the sea-breeze system (Diab, 1977).

A feature of many temperature inversions present in the atmosphere is a sharp decrease in water vapour content as one progresses upward into the warmer air layers. It is a known fact that the least water vapour in the upper air is found where the strongest temperature inversions occur. Thus the water vapour content of the upper air above the Benguela upwelling area is low, as is evident from the 850
Figure 5. Dew-point depression at 850 mb-level for (a) January, and (b) July (after Schulze, 1965).
mb dew-point depression distribution (Fig 5). The dew-point depression represents the difference between the dry-bulb air temperature and dew-point temperature and is an indication of the dryness of the air.

Owing to the stable stratification of the air layers over the Benguela area, turbulence in the air is suppressed and the convective activity required for cloud formation is inhibited. In the region north of 32°S, the cloud cover, which frequently occurs in the form of low stratus or stratocumulus below the trade inversion, is variable (Hoflich, 1984). Fog, which usually forms when maritime air moves over the cold upwelled water at night, is common along the northern coastal areas, occurring approximately 100 days a year on average in the Walvis Bay area. Offshore the frequency of fog decreases to about 1%, but a corresponding increase in cloud cover is evident (Hoflich, 1984). A zone of minimum cloud cover exists between 25°S and 34°S in summer. This zone migrates north in winter to between 22°S and 32°S.

Off the South West Cape coast the duration of sunshine reaches an average of 70% of the possible (Hoflich, 1984). The greatest amount of cloud cover occurs during winter, accompanying the cold frontal systems. Fog only occurs on an average of 1-2 days a month in the southern Benguela region, increasing in autumn to an average of 5 days a month.

The solar radiation received obviously depends on the season and latitudinal location. The northern Benguela accordingly receives more insolation than the southern part, but this meridional effect
is offset by the increased cloud cover in the north, in particular inshore where the frequency of fog is high (Shannon & Pillar, 1986). The effects of solar radiation over the Benguela were noted by Pomeroy (undated) to be severe. In January the maximum average amount of solar radiation received at the earth's surface has been found to occur at 2 regions, viz: 30°S 17°E and 30°S 116°E (Pomeroy, undated). The first region corresponds to an area between Hondsklip Bay and Port Nolloth on the west coast of South Africa.
Large Scale Upwelling

The Benguela current forms the eastern boundary of the South Atlantic Ocean gyre, and is typically shallow (100-200 m), broad (100-200 km), and cold, flowing sluggishly (10-20 cm.s⁻¹) equatorward parallel to the west coast of southern Africa. Eastern boundary currents flowing from high latitudes, together with upwelling, are known to result in cold water masses along the western edges of continents (Tont, 1981). Upwelling in response to favourable winds (i.e. equatorward) is a dominant feature of the inshore waters off the west coast of southern Africa. Coastal upwelling occurs whenever alongshore wind stress induces offshore Ekman transport and, therefore, divergence in the surface current at the coastal boundary. Central Atlantic water from depths of some 200-300 m ascends to replace the surface waters. On reaching the surface this freshly upwelled water is cold, with a temperature of some 10°C, and is characterised by low salinity and oxygen concentrations and a high nutrient content.

The intensity of upwelling differs seasonally and can obviously be correlated with the surface wind stress. Calculated Ekman transport off the west coast of South Africa during summer (Fig. 6a) indicates that, over a broad scale, the offshore surface transport is relatively consistent. A spring-summer maximum in upwelling occurs in the southern Benguela area. The upwelling is, however, suppressed in the southern region during winter as a result of the predominant north-westerly winds (Fig. 6b). In the central part of the Benguela system upwelling occurs throughout the year, while north of about 25°S a late winter-spring maximum occurs. The
Figure 6. Ekman transport during (a) summer, and (b) winter (after Parrish, et al. 1984).
Figure 7. Average seasonal SST (°C) for (a) summer, (b) autumn, (c) winter, and (d) spring (after Boyd & Agenbag, 1984).
Figure 8. Seasonal SST cycles at Swakopmund (1967-1985) and Cape Town (1956-1985). Error bars are standard deviations of the monthly means (after Shannon, et al. 1986).
upwelling season in the southern Benguela is thus out of phase with that in the north. The significance of this phase lag is that the upwelling in the northern Benguela is in phase with the seasonal insolation cycle, whereas in the southern Benguela it is not (Shannon, et al. 1986).

The offshore and longshore extent of cold water is also seen to be less in summer and autumn than during winter and spring off the west coast (see Fig. 7). Decreased solar heating of the ocean surface during winter implies that the upwelled water in the northern part is not substantially heated and therefore remains fairly cool (Fig. 7c & d). During summer decreased upwelling in the north, in conjunction with increased solar heating, results in higher SST's in the northern Benguela (see Fig. 7a & b). Although upwelling is suppressed in the southern Benguela during winter, the surface water remains cool because less solar heat is available to warm the water. During the maximum upwelling season in spring-summer the cold upwelled waters are rapidly heated as they move offshore. As a result of these phase differences between north and south, there is no clear seasonal SST cycle in the southern Benguela, but a pronounced seasonal SST cycle exists in the northern part (see Fig. 8). The large-scale seasonal surface temperatures broadly reflect changes in insolation, upwelling, vertical mixing and horizontal advection (Shannon, 1985).

On the interannual scale it has been found that several distinctive warm and cool periods have occurred in the northern Benguela. Only 3 anomalous warm events synonymous with the Peru "El Nino" have been documented off South West Africa since 1934. These events are
thought to recur consistently over a large, ocean-wide scale, and are initially driven by variations in the trade wind belt of the Equatorial Atlantic Ocean, which are caused by a meridional shift of the Inter Tropical Convergence Zone. During the warm events equatorward wind stress over the northern Benguela is found to be stronger than usual, but upwelling is suppressed by the southward progression of warm, highly saline water from the Angolan region (Shannon, et al. 1986). The response is channelled in an eastward direction along the equator and then southward down the Benguela. The proximity of the 2-3°C warmer water near the west coast can result in dramatically increased rainfall over Namibia.

The 3 major recorded warm perturbations (1934, 1963 and 1984) in the northern Benguela exhibit similar features to the Pacific "El Niño", but are less intense and less frequent than their eastern Pacific counterparts (Shannon, et al. 1986). The effect on the heat balance is not known, but it could be anticipated that the presence of warmer water would lead to increased evaporation and therefore a slightly greater heat loss than usual. These major events have a negligible effect on the southern Benguela.

On a smaller scale, minor warm water perturbations have been recorded in the extreme southern part of the Benguela, e.g. 1976/77 and 1982/83. These have been found to be localised features, due to the partial failure of the upwelling-favourable southerly winds in the extreme south as a result of an unusual position of the semi-permanent SAA. The atmospherically caused warming of the surface waters during the austral summer of 1983 was related to the period of greatest anomalies in the Pacific Ocean and indicates a more
rapid connection at the higher southern latitudes than in the north (Shannon, et al. 1986). The effects of the warm and cool periods on the heat balance are not accurately known, but are not thought to be significant.
Heat Balance of the South Atlantic Ocean

The heat balance of the world's oceans has been dealt with over large time and spatial scales by many authors (e.g. Sverdup, 1942; Budyko, 1963; Privett, 1960; Weare, et al. 1981; Hsiung 1986; Hoflich, 1984). This treatment serves to give a general indication of the heat received or lost by the sea surface in particular climatic areas. In the book entitled "Climates of the Oceans" (van Loon (ed.), 1984), Hoflich devoted a large part of the section "Climate of the South Atlantic Ocean" to a heat budget study of the area. An insight into the regional long-term situation over the South Atlantic Ocean can be obtained through a comparison of the monthly mean heat balance and the variables which control it in January and July.

The amount of incident radiation received by the sea surface varies mainly according to cloudiness, latitude and season. During January (Fig. 9a) the distribution of insolation over the South Atlantic Ocean depicts an approximate meridional gradient with values decreasing to the north and south of the maximum (250 W.m⁻²) centred between 20° and 30°S. Another maximum of over 275 W.m⁻² exists over the west coast of southern Africa, reflecting a departure from the meridional distribution. In July (Fig. 9b) the meridional gradient which, as a result of a lower sun angle and reduced daylight, decreases from north (200 W.m⁻² at 0°) to south, is again slightly upset along the west coast where an above average 125 W.m⁻² is received. It is evident that a significant seasonal variation in the radiation received occurs, with summer insolational values approximating twice those of winter.
Figure 9. Insolation at the sea surface (W.m\(^{-2}\)) during (a) January, and (b) July (after Hoflich, 1984).
Figure 9. Insolation at the sea surface (Wm$^{-2}$) during (a) January, and (b) July (after Hoflich, 1984).
Figure 10. Heat balance at the sea surface (W.m\(^{-2}\)) during (a) January, and (b) July (after Hoflich, 1984).
Figure 11. Evaporation (mm) from the sea surface during (a) January, and (b) July (after Hoflich, 1984).
Figure 12. Sensible heat flux (W.m\(^{-2}\)) at the sea surface during (a) January, and (b) July (after Hoflich, 1984).
The monthly mean heat balance at the sea surface of the South Atlantic Ocean for a summer month (January) and a winter month (July) are shown in Figure 10a & b respectively. The average heat gain over much of the South Atlantic Ocean during January is of the order of 75 W.m\(^{-2}\), while a heat loss of some 100 W.m\(^{-2}\) is evident during July. This indicates that a compensation contrast exists in the heat balance over the South Atlantic Ocean between summer and winter.

On the other hand, in the region of the retroflection of the warm Agulhas current to the south of Africa, there is always a heat loss from the ocean, varying from an average of 50 W.m\(^{-2}\) in January to more than 225 W.m\(^{-2}\) in July. This heat loss is the largest in the South Atlantic Ocean and is mainly due to a large amount of the available insolation heat being expended during the evaporation process. More than 210 mm is evaporated, on average, during both seasons (Fig. 11a & b). The sensible heat loss (Fig. 12) is small, with little variation between summer (20 W.m\(^{-2}\)) and winter (30 W.m\(^{-2}\)).

In contrast to the Agulhas current retroflection area, the Benguela region always experiences a heat gain, which exceeds 200 W.m\(^{-2}\) near the coast in January and is of the order of 50 W.m\(^{-2}\) in July, decreasing from the coast to open sea. The heat gain by the cold upwelled Benguela waters is the highest in the South Atlantic Ocean and can be ascribed to the facts that clear skies predominate over this area and, throughout the year, a minimum of the high insolation heat is expended for evaporation and a small gain in
sensible heat occurs. The Benguela area is thus a heat source for
the ocean and contrasts strongly with the Agulhas retrolection area
to the south, where large heat loss acts as a source of heat for
the atmosphere.

It should be noted, however, that even in the anomalous regions of
the Benguela current and Agulhas retroreflection, similar differences
between summer and winter exist in the heat balance. Reasons for
the compensation contrast in the heat balance between seasons must
lie in the fact that insolation varies between summer and winter.
The unusual heat fluxes may exist as a result of unusual values for
the constituents and in the environmental conditions which control
them.
It has been indicated by Hsiung (1986) that the upwelling regions in both the Atlantic and the Pacific Oceans receive the largest amount of heat. The cold ocean currents off the west coasts of Peru and Chile, California, South West Africa and North West Africa, are dominated by coastal upwelling and stand out as regions of downward directed heat flux, contrasting with the largest exchange from ocean to atmosphere in the domain of the warm currents (Hastenrath, 1977).

These eastern boundary currents, which flow from high to low latitudes, are generally colder than the water in the centres of the oceans and, in conjunction with the effect of equatorward winds and a suitable bathymetry for upwelling, even colder water results at the surface. Inland of the coast, dry typically desert-like conditions are experienced. Diurnal amplification of the equatorward winds off all the four major upwelling regions has been reported by Hawkins & Stuart (1980), Jury (1985), Elliot & O'Brien (1976), and Uhart (1976). The longshore winds result from the Coriolis deflection of sea-breezes and tend to be at a maximum during summer afternoons when solar radiation is maximised. The SST is usually less than the air temperature at all the upwelling sites so that the overlying air layers are stable and sensible heat is directed downwards to the sea surface. A minimal amount of evaporation takes place. The cold upwelling regions therefore gain large amounts of heat and are strong exporters of heat.

Direct comparisons between the heat balances at the different
upwelling sites are made difficult for the following reasons:

(i) variations in latitudinal location of the different upwelling sites;
(ii) methods of measurement of the atmospheric variables vary from site to site and from author to author, and a variety of formulas are used to calculate the same heat budget terms;
(iii) few comprehensive published works are available on the heat budgets of more localised areas.

The SST's at the different upwelling regions tend to differ according to the latitudinal position of the upwelling sites and the magnitude of the Coriolis force, which increases polewards. The coldest water is generally found close to the shore where the SST of the upwelled waters of the Benguela ($2^\circ S$ - $34^\circ S$) can be as low as $9-10^\circ C$, and that off California ($23^\circ N$ - $48^\circ N$) can reach $8-10^\circ C$. Off Chile and Peru ($3^\circ S$ - $33^\circ S$) the actual surface temperature is commonly about $20^\circ C$ and off North West Africa ($5^\circ N$ - $27^\circ N$) SST's of approximately $13-20^\circ C$ are the norm.

The maximum upwelling seasons tend to vary according to the seasonal variation in the wind at the different sites. Off the South African and Californian coasts the upwelling index tends, generally, to be highest in spring and summer, while off the North West African coast the north-east trades are upwelling-favourable throughout the year, although a spring-summer maximum is evident. However, off Peru north of $30^\circ S$ and off South West Africa north of $25^\circ S$ a winter maximum in upwelling occurs.
During the upwelling season off Oregon the wind fluctuates in both speed and direction with periods of 3-7 days (Huyer, 1978), while the Benguela system has been shown to be influenced by synoptic cycles of periods ranging from 3-6 days (Preston-Whyte & Tyson, 1973). The fact that the African continent ends at 35°S makes for an important distinction between the Benguela upwelling system and upwelling systems of other parts of the world as this allows easterly moving cyclones a free zonal passage across the subcontinent, unimpeded by steep topography, such as the Andes, or extensive land masses (Nelsen & Hutchings, 1983). A corresponding synoptic variation in the heat balance is therefore to be expected over the Benguela area. The Benguela and Californian systems tend to be weather dominated, while the Peru and North West African systems are more climatically controlled.

The Benguela system is fairly compact, as its offshore extent is only some 200 km compared to the 900 km and 700 km extents of the Peru and Californian systems respectively (King, 1975). Similarly, the frontal temperature gradients of the Benguela have been reported to exceed considerably those found in the upwelling regions of the American and North African west coasts (Bang, 1973). The irregularity of the southern Benguela coastline is said to be unusual in its dimensions among upwelling coastlines of the world, and the exceptionally low SST can be attributed to the subsurface topography and orographic features of rocky outcrops (Nelson & Hutchings, 1983).

In general, the upwelling regions of the world can be regarded as similar in that the forcing mechanism required for upwelling is the
same and the effects of the lowered SST on the environment and on the heat balance are universal, even though they may vary in intensity from one site to another. Localised differences do occur as a result of latitudinal position, seasonality, topography and availability of cold water in proximity to the coast. Of paramount importance, however, is the fact that all upwelling regions of the world are remarkable as regions of heat gain at the sea surface.
Winds in a longshore direction along the west coast of southern Africa are fairly consistent in speed and direction, with southerly winds predominating (see Fig. 4). A greater variation is, however, evident in the southern Benguela area, particularly during the winter season. The offshore Berg winds, which occur more commonly during winter, have been found to be directionally coherent over a distance of nearly 1500 km along the west coast (Shannon, 1985). The offshore extent of these winds has been indicated through satellite imagery of aeolian transport to reach some 150 km from the coast.

The mean air temperature along the west coast is 19°C in summer and 16°C in winter, with mean daily ranges exceeding 10°C. The minimum diurnal range is 4°C (Jury, 1985). Longshore temperature gradients are weak, but cross-shore temperature gradients in the coastal region are usually intense due to the desert-like interior being in close proximity to the cold waters of the Benguela. The air inland is heated intensely by insolation and often has a temperature 6-10°C higher than that at the coast, where the air above the water is cooled from below. The resulting cross-shore thermal gradients are manifested as thermal fronts along the coast. This heat imbalance in a cross-shore direction contributes to the formation of an isobaric layer aloft, which slopes downwards from the land to the sea. The air subsequently moves onshore at the surface and offshore above in response to the pressure gradient force. The cross-shore sea-breeze intensifies during the day and is deflected to the south (longshore) during the afternoon by the Coriolis force. These
diurnal period winds have been found to rotate in a predominant anticlockwise sense (Jury & Guastella, 1987). In the northern hemisphere a clockwise turning of the diurnal winds was observed off Oregon by Hawkins & Stuart (1980).

The diurnal cycle of the sea-breeze, and the corresponding land-breeze system which occurs at night, is superimposed on the synoptic cycle and is known to be common along the entire west coast of southern Africa. The effects of the sea-breeze are, however, only experienced up to a distance of approximately 10 km inland owing to the fact that the thermal front acts as a restriction to cross-flow. On the other hand, the coastal winds were found to extend approximately 20-35 km seawards (Jury, 1985 and Keen, 1979).

Seasonal variations in the sea-breeze are evident, with the maximum strength occurring in summer when solar radiation is maximised, and the temperature contrast between land and sea is therefore greatest. Land-breezes are more intense during winter. Obviously, the diurnal pulsing of winds and the seasonal variation exhibited therein must be important for the coastal upwelling dynamics of much of the Benguela region (Shannon, 1985).

An examination of the SST's along the west coast (see Fig. 7) revealed that longshore temperature gradients are weak in comparison with the offshore gradients. In addition, the sigma-t distribution along the southern African coast indicates minimal longshore variability in the density of the coastal water from approximately Cape Agulhas to Cape Frio in the north (see Fig. 13). Offshore SST gradients are generally constant in a longshore direction, but these
Figure 13. Sigma-t compared with depth approximately 200 km from the coast around southern Africa (after Parrish, et al. 1983).
Figure 14. Smoothed air-sea temperature differences (°C) over the southern Benguela for (a) summer, and (b) winter (after Kamstra, 1985). Isolines were omitted further offshore where less than 50 ship observations were made as the accuracy is unknown.
gradients may be more intense at the local upwelling sites such as the Cape Peninsula, Cape Columbine, Hondeklip Bay and Walvis Bay.

The distribution of air-sea temperature differences for summer and winter (Fig. 14) reveals that longshore and offshore variations are similar to that of the SST, with the inshore regions experiencing the largest air-sea contrast. The zero line of equal sea and air temperatures was noted to be closer inshore during winter (Kamstra, 1985). To summarise, it can be assumed that atmospheric and oceanographic variations along the west coast of southern Africa are minimal on a broad climatic scale.
2. SYNOPSIS SCALE PROCESSES

Meteorology

The Benguela upwelling area is weather dominated. Environmental parameters tend to have large ranges over the synoptic time scale of a few days. It is these environmental variables which determine the synoptic variation in the heat balance over the Benguela area. Weather changes over the Benguela are dominated largely by perturbations in the southern hemisphere’s westerly circulation, which appear on the surface as a succession of cyclones or anticyclones moving around or across the country from west to east (Schulze, 1965). Thus the coastal winds which induce upwelling are determined by synoptic weather systems of scales varying between 100 and 1000 km.

During summer the semi-permanent South Atlantic Anticyclone (SAA) is commonly situated west of the country (see Fig. 15a) and, in conjunction with the interior heat low, produces strong upwelling-favourable south-easterly winds along the west coast. The dominant south-easterly conditions are, however, disturbed by eastward-migrating low pressure cells accompanied by frontal systems, which originate in the south. During summer these low pressure systems usually pass south of the country, but this movement is accompanied by an eastward shift of the SAA. During its eastward passage the SAA is forced to ridge south of the country (Fig. 15b). As the South West Cape coast is approached the isobars are compressed parallel to the coast by both a pressure trough inland and the low pressure
Figure 15. Weather systems which affect the Benguela area: (a) SAA established west of country, southerly winds along west coast; (b) SAA ridging, gale-force southerly winds along west coast; (c) SAA south of country, coastal low over west coast; (d) Berg wind situation over west coast; (e) cold front and associated low pressure cell approaching west coast, north-west winds are precursor; (f) cold front passed Benguela area, winds back to south-westerly.
systems to the south. This has the effect of producing gale force south-easterly winds over the west coast, resulting in intense atmospheric forcing for upwelling. These gale force south-easterly winds can persist for some days.

In summer the low pressure systems may occasionally penetrate the south of the country, but their effect is slight, normally manifested as a weakening of the south-easterly winds. The migrating anticyclone moves to the Indian Ocean and is replaced by another high pressure system over the South Atlantic Ocean.

A slightly smaller scale synoptic disturbance is that of the coastal low pressure system, which migrates along the coast from west to east at periodic intervals throughout the year. First described by Taljaard and Van Loon (1963) as "leader cells" which rhythmically propagate anti-clockwise around the continent preceding eastward moving pressure perturbations and cold fronts, coastal lows are now defined by South African Weather Bureau forecasters as "small areas of relatively lower pressure which appear in the lower levels of the atmosphere (below 700 mb) along the coast." The coastal low is associated with subsidence off the interior escarpment and is accompanied by an inversion and wind shear in the lower levels, but the essential features are the pressure minimum and the shallowness of the system (Coastal Low Workshop, 1984).

The occurrence of a coastal low is determined by the synoptic scale circulations. In summer they first appear in the Luderitz area on the west coast during the ridging process of the SAA. During winter these systems usually form ahead of an approaching cold
front to the south-west of the country (see Fig. 15c). The coastal lows migrate from Luderitz south along the coast to Cape Town at an average speed of 7 m s\(^{-1}\) after which they reappear in the Cape Agulhas area and move eastwards along the coast as "leader fronts".

The inner core of the system (wind speeds less than 4 m s\(^{-1}\)) has an average diameter of 150-200 km. The cyclonically rotating wind is offshore ahead of the low and the inversion level is strengthened and lowered. Upwelling is locally suppressed as the wave travels along the coast as a result of the cyclonic rotation of air about the low pressure cells and light wind conditions at the centre (Nelson and Hutchings, 1983). After the passage of the low pressure cell the inversion is raised and the wind switches to south-west (i.e. onshore), often accompanied by fog on the west coast.

Should a strong, stationary high pressure system exist over the continent, as is sometimes the case during winter, a strong offshore (east to north-east) airflow, known as a "Berg wind", may develop along the west coast ahead of a coastal low or a frontal system approaching the country from the south-west (see Fig. 15d). The anticyclonically rotating air from the interior high descends from the plateau down to the coast ahead of the coastal low and is adiabatically heated, causing high air temperatures along the west coast. Upwelling is then suppressed and rapid changes in the interior thermohaline structure of the Benguela waters have been observed by Nelson and Hutchings (1983).

As the warm air moves offshore, a stark air-sea temperature difference may be set up over the cold Benguela waters. In the
extreme case at the coast this difference may be as much as 15°C, but the deviation diminishes offshore as the cold Benguela waters act to modify the air - sea temperature contrast. This effect is dealt with in more detail in a subsequent chapter.

During winter the westerly wind belt migrates northwards, allowing the unimpeded passage of frontal systems, associated with eastward moving lows over the continent (see Fig. 15e). These systems are often accompanied by rain. Inversion characteristics along the coast are altered with the passage of a cold front in that preceding the frontal system an intensification and lowering of the inversion base is evident, while post-frontal conditions eliminate the inversion. Upwelling is suppressed by the north to north-westerly winds, which prevail ahead of the cold front. After the passage of the frontal system a rise in air pressure is evident and the winds back to south-west and can again become upwelling-favourable (see Fig. 15f).

Progressive wind vectors from Koeberg from a one-year record (1983) (see Fig. 16) indicate the dominance of the summer southerly winds and the anti-clockwise synoptic scale cycling of winds with the clear, calm, stable conditions associated with anticyclones and the unsettled, cloudy, windy conditions with the possible precipitation that accompanies coastal low pressure systems and frontal depressions (Tyson, et al. 1976).
Figure 16. Progressive wind vectors recorded at Koeberg, 1983 (from Guastella, 1985).
Synoptic scale upwelling

The phenomenon of upwelling is wind dependent and therefore determined by mesoscale atmospheric perturbations. Coastal upwelling tends to vary according to local wind conditions over both short time and spatial scales. Changes in climatic variables on the synoptic scale suggest that an upwelling season is characterised by a series of upwelling events. A "coastal upwelling event" may be defined as the period from the onset of winds favourable for the initiation of upwelling to the arrival of the coldest, densest water on the surface (Hawkins & Stuart, 1980). The upwelled water appears on the surface within a few hours following the onset of upwelling-favourable southerly winds.

The west coast of southern Africa is typically dominated by southerly winds, which are favourable for the upwelling process, but is disturbed by the periodic passage of coastal low pressure systems and eastward moving cyclones from the south, which act to relax and/or modify the winds, suppressing upwelling in the process. Bang (1971) stated that the oceanic processes in the Benguela Current are dominated by short-term atmospheric interactions. Every few days the characteristic southerly wind increases in magnitude, leading to a renewed burst of coastal upwelling. During this 'active' phase the surface waters respond directly to the wind stress, resulting in a time scale of 3-5 days variability, while during the 'passive' phase upwelling ceases and the system is permitted to equilibrate physically and develop biologically (Waldron, 1985).

Although on a broad scale, the west coast can be regarded as
uniform, it has been found that locally upwelling is variable along the coast, being more pronounced at particular sites. This spatial variability can be ascribed to smaller scale bathymetric and topographic features. Thus far 4 major upwelling sites have been identified off the west coast of southern Africa, viz. off the Cape Peninsula, Cape Columbine, Hondeklip Bay, and further north at Luderitz (see Fig. 17). All of the sites where intense upwelling occurs are adjacent to prominent orographic features, while areas of lesser upwelling tend to lie to the west of lower lying areas (Tauntor-Clark, 1985).

The main centres of upwelling are characterised by plumes or tongues of cold water, which extend in a north-westerly direction from the coast. During the 'active' phase rapid growth in the upwelling plume occurs, with the coldest water (10-12°C) indicating the core of the plume. With the cessation of upwelling-favourable winds (coastal low or Berg winds) the plume starts to decay. The cold, dense upwelled water in the surface mixed layer tends to sink both at the oceanic front and across the upwelling plume (Jury, 1986). The decay during the 'passive' phase is indicated by a shrinkage of the plume and warming of the water (to 12-16°C) by insolation.
Figure 17. The 4 major upwelling sites of the Benguela system (after Shannon, et al. 1984).
3. HEAT TRANSFER MODELLING

Air-Sea Interactions Near the Coast

Mention of the offshore variation in the average air-sea temperature differences over the west coast area has been made with reference to Figure 14. Evident from the isolines is that the air-sea temperature difference is at a maximum near the coast and north of the upwelling centres of the Cape Peninsula, Cape Columbine and Hondeklip Bay, but the anomaly diminishes offshore, where the sea water is also warmer than that experienced at the coast (see also SST distributions in Fig. 7). A corresponding decrease in the magnitude of the turbulent heat fluxes in an offshore direction would therefore be expected due to the decrease in temperature contrast between the air and the sea surface. This decrease in the air-sea temperature contrast offshore must be due to the cooling of the air as it passes over the colder upwelled waters and/or due to the warming of the surface water as it moves offshore.

Kamstra (1985) ascribed the possible inconsistencies in his data and the implied higher air temperatures near the coast at Alexander Bay to the formation of warm core (shallow) coastal lows in this region. Ahead of these low pressure systems the inversion base is lowered and fine weather with dry, offshore airflow prevails. In the more extreme case hot, dry Berg winds may occur as a precursor to the low. The temperature contrast between the air and the sea surface is usually greatest at the coast, when the warm, dry air (sometimes with a temperature as high as 30°C) ahead of the low pressure system...
passes over the cold upwelled waters. As the air near the sea surface moves offshore it is progressively cooled by the water with which it is in contact in this stable environment, until the air-sea temperature difference offshore reaches zero and no further alteration of the air properties can take place.

Small scale modelling of an idealised study area under these idealised conditions can provide useful information on the variations in the air properties offshore. In addition, the inherent offshore and onshore variation in the heat transfer processes can be estimated. The progressive decrease in the temperature anomaly offshore can be explained through the application of a model which considers the ideal conditions of a coastal low as representing a stable closed system, with only redistribution of heat from air to sea taking place. Due to the thermal inertia of sea water, conditions in the sea can be assumed to be steady. Thus theoretically we have a volume of air, $V_{air} = hA$, moving at a constant rate over the sea surface, where $h$ = height of inversion and $A$ = oceanic area.
For simplicity the height of the inversion lid is assumed constant for the period when airflow is offshore.

The air temperature is always assumed to be greater than the SST so that the sensible heat flux represents a loss of heat from the air to the sea. It is convenient to introduce the air temperature deviation, $T$, from the SST which is assumed to be invariant.

$$\text{Air temperature} = \text{SST} + T$$

and $Q_h = C_h C_p \int_{A} \rho_{air} T u$ from Chapter A(3).

The loss of heat from the air to the sea over area $A$ during time, $t$ to $t + \Delta t$ will result in a fall in the temperature deviation from $T$ to $T - \Delta T$. This loss can be represented in two ways:

(a) the heat flux out of the air into the sea

$$Q_h A \Delta t \text{ (Joules)}$$

(b) the change in heat capacity of the air

$$- \Delta T C_p \rho_{air} A h \text{ (Joules)}$$

Thus

$$Q_h A \Delta t = - \Delta T C_p \rho_{air} A h$$

or using the expression for the heat flux in terms of the environmental parameters and rearranging to a derivative for the change

$$\frac{dT}{dt} = - \frac{C_h u}{h} T$$
Introducing the initial temperature deviation $T_0$ at time $t = 0$, the complete form for the temperature deviation is then

$$T = T_0 e^{-\frac{C_h}{h} u t}$$

(1)

An alternative form, which uses the offshore distance $x = ut$ rather than time as the independent variable, is

$$T = T_0 e^{-\frac{C_h}{h} x}$$

(2)

The air moving off the land is usually warm and dry, i.e. unsaturated, therefore similar to the air-sea temperature contrast which exists near the coast, so a vapour pressure or specific humidity contrast is evident between the sea surface and the overlying air. This difference in specific humidity determines, along with the wind speed, the magnitude of the latent heat flux from the sea. The specific humidity of the sea surface is dependent on the SST, while the specific humidity of the air, in addition to being controlled by the moisture content, is influenced by the temperature of the air. Hence the variables that determine the latent heat flux are similar to those which determine the sensible heat flux.

For evaporation to take place the saturation specific humidity of the sea ($q_{s(sst)}$) must exceed the specific humidity of the air above ($q_a$) so that $Q_e = L \rho_{air} C_e (q_{s(sst)} - q_a) > 0$. Heat is therefore lost by the sea and gained by the air, although in a latent form, and the specific humidity of the air is allowed to increase until
As in the case of the air-sea temperature deviation, the deviation in specific humidity of the air from the saturated value of the assumed invariant sea surface can be represented as $Q$.

Thus

$$q_a = q_a(sst) + Q$$

and

$$Q_a = L C_a C_p \gamma_{air} Q u$$

Through application of the same principles used in the derivation of the equation for the temperature anomaly decrease, the estimate of the decrease in the $Q$ anomaly (or increase in the specific humidity of the air) can be given by

$$Q = Q_0 e^{-\frac{C_a}{h} \text{ut}}$$  \quad (3)

or

$$Q = Q_0 e^{-\frac{C_a}{h} x}$$  \quad (4)

Thus, within the constraints of the listed assumptions of the model, along with the assumption that $C_h$ and $C_e$ are both constant, it can be concluded that the anomalies of temperature and specific humidity between the air and sea surface both decrease at an exponential rate with time or distance moved offshore. Because a constant SST was assumed this implies that the rate of decrease in temperature and increase in specific humidity of the air as it moves offshore is exponential.
Thus the greatest portion of the loss of heat by the air to the sea (sensible heat exchange) and the gaining of moisture by the air (latent heat loss from the sea) takes place in the nearshore region.

At first sight it would appear that a high wind speed should extend the influence of the continental air further offshore. This is not the case, however. The width of the land influence can be given by the e-folding distance, which is derived from equations (2) and (4) as:

\[
x = \frac{h}{C_h} \quad \text{and} \quad x = \frac{h}{C_e}
\]

respectively.
The area of maximum exchange is dependent on the thickness, \( h \), of the turbulent layer below the inversion, and on the turbulent exchange coefficients \( C_e \) and \( C_h \), which in turn depend on the wind speed. An increase in wind speed leads to increased values of \( C_h \) and \( C_e \) and leads to a greater turbulent exchange, which means that modification of the air by the sea is more rapid and the width of land influence, in fact, becomes smaller.

The thickness of the turbulent layer below the inversion is important in determining the rate of decrease of the deviations \( T \) and \( q \). The model assumed an idealised pre-coastal low condition, which is obviously characterised by a lowered inversion. If the height of this inversion was, say, 100 m the resulting e.folding distance of 100 km then corresponds to the dimensions of a coastal low. The actual diameter of a coastal low is not accurately known, but 100-300 km appears to be a reasonable approximation.

With the passage of a coastal low pressure system winds at the coast are known to typically change from warm and offshore ahead of the system to an onshore influx of cool, maritime air, often accompanied by fog, once the system has passed (see Fig. 18). It is therefore feasible that the air advected onshore behind the coastal low could in fact be the same as that which previously blew offshore ahead of the low pressure system, only the properties have been altered offshore to become maritime in nature.

It is important to remember that the assumptions of the model are simplified, and applications to the real situation are limited to the extreme synoptic case at the leading edge of a coastal low,
Figure 18. A typical coastal low migrating down the west coast of South Africa.
where airflow is warm, dry and offshore, sea conditions are fairly calm and invariant, the inversion is low and maximum heat gain of the surface waters is permissible. This idealised situation indicates that the greatest sensible and latent heat exchange between the air and sea takes place near the west coast, while further offshore this exchange is exponentially less and the air - sea temperature and specific humidity deviations are reduced until the air and sea eventually reach an equilibrium state where no further turbulent heat exchange can take place.

In reality, the SST is colder inshore and warmer in an offshore direction, thus the air temperature is more likely to equalise the SST sooner and nearer to the shore than actually predicted by the model. Should warmer water be present offshore, as is usually the case, the latent heat loss by the sea surface will obviously increase in an offshore direction. In addition, the temperature and humidity deviations in the model are treated as separate processes, whereas they can, in fact, be interlinked. The relative humidity of the air increases in sympathy with a decrease in air temperature. This increase in relative humidity is realised as an increase in the specific humidity of the air, so that saturation of the air is, in fact, achieved sooner than what might be predicted from the model. The model can therefore only be used to indicate the general trends in the coastal zone. It is still, however, useful in that it indicates that the coast itself is the extreme site of heat exchange.
Relative Magnitudes of Heat Transfer Components Near the Coast

Small scale modelling of the Benguela area serves to indicate that the gain of heat by the sea is greater near the coast than at the oceanic region further offshore. The relative magnitudes of the heat transfer processes need to be considered in order to illustrate the importance of this to the heat balance of the area. A review of the assumptions applied are that airflow is offshore and the SST is lowest near the coast, but higher in the oceanic region offshore. The air - sea temperature contrast is greatest near the coast, decreasing in an offshore direction, and most turbulent heat exchange takes place in the nearshore area. The boundary between the coastal and oceanic water lies at least 100 km in an offshore direction, thereby allowing sufficient distance and time for air property changes to occur.

From the idealised picture presented in Figure 19, the overall heat gain by the ocean at the coastal region can be regarded as being of a greater magnitude than at the oceanic region further offshore due to the following:

1. Solar radiation, $Q_1$, can be treated as approximately equivalent at both areas and represents a large heat gain by the sea (+).
2. Although the loss due to back radiation, $Q_2$, is probably slightly less at the coastal area than further offshore, for the purposes of the comparison study the $Q_2$ values can be regarded as equivalent losses (-).
3. Owing to the fact that the air temperature exceeds the SST at the coastal area, there is a gain (+) of sensible heat ($Q_h$).
Figure 19. (a) Model of the heat budget components near the west coast and more than 100 km offshore at the oceanic region; (b) similar to (a) but in cross-section, indicating relative magnitudes of the components.
The converse is true in the oceanic area because the water is warmer than the now cooled air, causing $Q_h$ to be a heat loss term ($-$).

(4) Loss of heat from the sea in latent form ($Q_e$) is small near the coast because the SST is less than the air temperature and the stable air layers prevent any upward mixing. Over the oceanic area, however, the latent heat loss is greater because the SST is now warmer than that of the overlying air in this region. Upward mixing and transport of the evaporated moisture is sustained by the convectively less stable air layers, allowing the evaporative heat loss, $Q_e$, to be maintained at the surface.

Collectively the factors (1) - (4) contribute to a higher heat gain in the coastal area than in the oceanic area more than 100 km offshore because $Q_t = Q_1 - Q_0 + Q_h - Q_e$ at the coast is greater than $Q_t = Q_1 - Q_0 - Q_h - Q_e$ at the oceanic area. This represents the typical situation along the west coast of southern Africa and variations in a longshore direction are not anticipated as being significant. One may therefore be justified in measuring the heat budget components at one particular site along the west coast to broadly typify the situation along the whole coast. The only variations expected along the coast can be listed as:

(i) seasonal wind variations between the northern and southern Benguela, and therefore seasonal variations in upwelling;
(ii) the existence of some upwelling sites along the coast, where the effects may be more pronounced;
(iii) variations in the amount of fog and cloud cover between the
north and south; and

(iv) slight variations in radiation intensity along the coast.

Over a broad scale, however, longshore variations are negligible in terms of the factors contributing to the overall heat balance. It is for this particular reason that a field measurement programme was conducted at one localised site, St Helena Bay, to characterise the heat balance of the Benguela system as a whole.
In order to accurately categorize the relevant terms of the heat balance of the Benguela system, it was necessary to obtain data representative of the Benguela upwelling area. The need for a time series of heat budget-related data for the study of the synoptic and diurnal scale heat budget over a typical Benguela upwelling area was fulfilled through the undertaking of an eight-day field measurement programme to St Helena Bay from 14-21 October 1986. The location of the study site is indicated in Figure 20. Centered at 32°45'S 18°05'E and approximately 130 km north of Cape Town, St Helena Bay lies in the southern Benguela region, adjacent to Cape Columbine, which is one of the four most important active upwelling sites on the west coast of southern Africa. Climatic conditions experienced in the area are roughly representative of the southern Benguela, although one must be aware of local climatic controls.

To make the study more complete, additional meteorological data was obtained from the meteorological station at Langebaanweg, situated approximately 15 km south of St Helena Bay at 32°59'S 18°09'E. Data from this station provides an insight into the cross-shore variations near the coast. Meteorological data to verify the measured data and to extrapolate the heat budget components when no measurements were made at St Helena Bay was obtained from the Cape Columbine lighthouse at 32°50'S 17°51'E.

Some more extensive meteorological data, including global radiation data, recorded at the D.F. Malan Airport Weather Office near Cape
Figure 20. Location of St Helena Bay and meteorological stations from which additional data was obtained.
Town (33°58'S 18°36'E) was also obtained. D.F. Malan is situated at an altitude of 44 m and is approximately 15 km inland from the Cape Peninsula upwelling site. Data gained from the site is broadly representative of the southern extremity of the Benguela system.

Additional long-term global radiation data was obtained from Alexander Bay (28°34'S 16°32'E). Located only approximately 8 km inland and at an altitude of 21 m, the global radiation data obtained from the Alexander Bay weather station can generally be taken as representative of the northern extremity of the southern Benguela region. The D.F. Malan and Alexander Bay data can be used to interpret the heat balance in a longshore direction along the west coast. Unfortunately global radiation data is not measured at any other site along the west coast of southern Africa.

An investigation into the variability in both atmospheric and oceanographic environmental conditions over the diurnal cycle and the synoptic cycle of a few days was necessary in order to determine and therefore characterise the variation of the heat flux from the atmosphere into the surface waters of the Benguela over short time scales. Obviously the relatively short time span and the fact that the components of the heat balance are measured at a localised site will contribute to some biasing in the results, but little variation in the overall trend is anticipated.
The Study Area

Due to its location, the St Helena Bay area is influenced by the dry trade climate of the coast to the north and the temperate Mediterranean climate of warm, dry summers and colder, wet winters to the south. St Helena Bay itself receives little rainfall (≤ 300 mm p.a.), with the maximum occurring during winter. Because this area lies in the path of migrating mid-latitude weather systems, a synoptic 6-day cycle has been found (Jury, 1986).

On an annual basis fog is not extremely common in the area, although higher frequencies are found during autumn, when the coastal low pressure systems occur more frequently. Fog was, in fact, found to be present 30-40% of the time spent at St Helena Bay by Sea Fisheries personnel during 19 March - 18 April 1987 (P. Chapman, pers. comm.). The Berg River flows out into the southern extremity of St Helena Bay. The outflow from this river is, however, not large enough to significantly alter the composition of the water within the bay.

In the Cape Columbine area the bathymetry and the orographic influence on the wind field are important controls on upwelling. Rising to a height of some 250 m above sea level and extending approximately 40 km seawards in a north-westerly direction, the Cape Columbine headland partly shelters the wide, log-spiral shaped St Helena Bay from the dominant southerly winds in summer. Diurnal land-sea breezes are commonly superimposed on the 6-day synoptic cycle, the strongest equatorward winds being recorded during summer afternoons.
With the Cape Canyon situated only 30 km offshore, west of Cape Columbine, cold water from depths of 200-300 m is readily available for upwelling as soon as the strong southerly winds blow. The canyon acts to channel the subsurface water so that on reaching the surface, it appears as a large scale surface tongue of cold water (minimum temperature 10-12°C) extending northward from the headland. Horizontal crowding of the isotherms and isotherlines is evident off Cape Columbine and indicates the intense perennial upwelling there (Nelson & Hutchings, 1983).

The water in St Helena Bay for the most part throughout the year originates as upwelled water which, to a greater or lesser extent, has been heated by the sun (Buys, 1959). Water within St Helena Bay is usually warmer (13-15°C) than newly upwelled water and is often separated from the coast by a narrow strip of cold upwelled water with temperatures ranging from 11-12°C (Taunton-Clark, 1985). According to Waldron (1985) it is inevitable that the warm surface layer in St Helena Bay is the result of insolation.

Jury (1985) indicated two different cases of southerly wind flow which influence upwelling in St Helena Bay (see Fig. 21). During shallow southerly wind events a distinct leeward wake and cyclonic curvature is evident in St Helena Bay. Lowest SST's are found directly west of Cape Columbine and minimal upwelling occurs along the coast north of St Helena Bay. However, when deep southerly winds blow, the cyclonic curvature of the wind in the lee of Cape Columbine is reduced and Ekman upwelling can take place along the coast to the north and east of the headland. A cool upwelling tongue is evident off Cape Columbine.
Figure 21. Comparison of (a) shallow (1 November 1980) and (b) deep (23 November 1980) cases of southerly wind flow over the St Helena Bay region. (c) and (d) are SST ($^\circ$C) distributions on the respective days (from Jury, 1985).
Figure 22. (a)-(e) Changes in the Cape Columbine upwelling tongue in response to local winds (SST in °C). (f) Progressive winds during the same period, 12-16 February 1980 (after Shannon, et al. 1984).
The formation and decay of the upwelling tongue in response to local winds has been depicted by Shannon, et al. (1984) and is illustrated in Figure 22. Under the influence of strong southerly winds a tongue of cool water (minimum temperature 12°C) extends NNE of Cape Columbine (Fig. 22a & b) and upwelling is induced further north along the coast. A tendency for the cool plume to curve around the headland towards the St Helena Bay coast in response to cyclonic curvature of the wind is indicated in Figure 22c. Note also the narrowing of the plume and the intense thermal front. During the decay stage (Fig. 22d & e) a warming and apparent shrinkage of the plume occurs and warmer water is present in the bay.

A dominant feature of the currents in the St Helena Bay area is the permanent presence of a strong north-eastward flowing baroclinic jet between the 200 m and 300 m isobaths west of Cape Columbine. The speed is typically greater than 50 cm.s\(^{-1}\) and its width has been estimated at between 20 and 30 km (Nelson & Hutchings, 1983). Holden (1985) has shown evidence of a sluggish southward flow along the St Helena Bay coast. A schematic representation of the currents around St Helena Bay in Figure 23 indicates a general northward flow and the presence of a cyclonic gyre, with associated eddies, in the coastal area south of Elands Bay. The residence time of the water in St Helena Bay has been found to be approximately 25 days, but periodic upwelling over the 3-5 day cycle has a pulsing effect on the water (Waldron, 1985).
Figure 23. Schematic diagram of near-surface currents in St Helena Bay during February 1979 (after Holden, 1985).
Instruments and Methods

Global radiation and meteorological data, relevant to the components of the heat balance, was measured at a beach site and at various locations in St Helena Bay (see Fig. 24). At sea daytime measurements were taken on the hour every hour (local solar time (LST)), weather permitting on 14/10 - 17/10 inclusive and 20/10 - 21/10 inclusive. Initially an anchoring station NNW of the town of Lasiplak and at the 30 m depth contour in St Helena Bay was selected, but thereafter stations situated slightly closer inshore were chosen as little variation between the offshore sites was noted.

Measurements at the beach station were co-ordinated to take place simultaneously with the sea station readings. In order to gain a more continuous time series, measurements were also taken at the beach station when the research vessel was not in use. Additional data was collected at night in order to quantify the nocturnal outward heat exchange.

Due to the nature of the field work involved, the use of small, transportable instruments, which could be set up with minimal complication or time delay, were selected. Solar radiation data was measured using two different instruments, a pyranometer and a solarimeter. Both instruments measure the total short wave radiation incident from the sun and sky on a horizontal surface at the ground. The pyranometer was a calibrated Li-cor Model LI-170 Quantum/Radiometer/Photometer which, due to its compactness, was used mainly at sea. The direct readout in micro-einstein$\cdot$m$^{-2}\cdot$s$^{-1}$
Figure 24. Location of observation stations at St Helena Bay.
was multiplied by a factor of 0.1 in order to obtain the equivalent values in W.m\(^{-2}\). The solarimeter, which was used mainly for the beach station readings, was a standard Kipp & Zonen type instrument connected to a millivoltmeter. The equivalent reading in W.m\(^{-2}\) was obtained by multiplying the output by a factor of 93.43. In field use instrument error has been estimated to be approximately 5% (Drummond, 1964). The solarimeters are reputed to be stable for use over a long period of time in the field, with minimal maintenance being required.

An estimation of the diffuse (scattered) component of incident radiation was obtained by shading the sensors. These values were taken in order that the direct beam of solar radiation (total incident - scattered component) could be compared to theoretically calculated values. In addition, measurements of net radiation over the sea surface were taken from a net pyranometer, but the values obtained were found to disagree with the calculated net radiation. This inaccuracy can probably be attributed to increased reflection from the research vessel and personnel in the vicinity.

A Thies hand-held anemometer giving a 10-second average digital readout was used for wind speed measurements. A one-minute average wind speed was then calculated from 6 consecutive 10-second average readings. In order to apply the bulk aerodynamic formulae, the wind speed measurements were transformed from the 2-m level of measurement above the sea surface to the 10-m level by means of the wind stress formula applying to a log wind profile. In this case the 2-m winds were multiplied by a constant factor of 1.38. Wind direction was estimated through the use of a compass.
Wet- and dry-bulb air temperatures were measured by an Assman psychrometer. The relative humidity (%) of the air, as required for the vapour pressure and specific humidity calculations, is read from a moisture chart (see Appendix 1). Measurements of the wet- and dry-bulb temperatures were taken at the approximate 2-m level above the surface at both the beach and sea stations. However, the bulk aerodynamic formulae for which the measurements are required apply to the 10-m level. This difference in height of measurement is not foreseen as representing a problem as regards the accuracy of the bulk aerodynamic calculations as only very small differences were observed between the 2-m and 10-m levels when tested by Hughes in 1984. It was also found that the resulting relative humidity (%) was always the same at the two different levels.

The vapour pressure of the air (mb) required for the back radiation calculations is derived from the formula:

$$e_a = \frac{e_s \times RH}{100}$$

where $e_s$ = saturation vapour pressure of the air (mb), read from the table in Appendix 2, and

$RH = \text{relative humidity (}%)$.

The specific humidity of the air, required for latent heat flux calculations, is derived from:

$$q_a = \frac{0.622 \cdot e_a}{p}$$

where $p = \text{air pressure (mb)}$. 
Saturation specific humidity at the SST is calculated according to the formula:

$$q_a(sst) = \frac{0.662 e_s(sst)}{p - 0.378 e_s(sst)}$$

Air pressure was simply measured from a standard aneroid barometer.

In addition to the SST measurements, profiles of temperature and salinity were planned for each sea station. Use of a Salinity Temperature Bridge Type M.C.5 was made for the above-mentioned measurements, but a malfunction of the salinity sensor unfortunately limited the data set to temperature only. Ocean current speed and direction were estimated through the use of drift cards, a stop watch and compass. A Secchi disc was used to estimate the transparency of the water. Lastly, the cloud cover required for back radiation calculations was estimated in tenths. All the physical data was manually recorded on data sheets at both beach and sea stations for processing at U.C.T. (see Appendix 3 & 4).
Results of the Field Study

(1) Synoptic weather cycle during the study

Variations in atmospheric conditions over the short synoptic scale determine variations in the heat balance over the west coast. It is therefore necessary to investigate the prevailing weather conditions experienced over the west coast during the field study period. An indication of the synoptic controls on the weather pattern from 14-21 October 1986 is provided in Figure 25. Further reference is made to Figure 25, which includes wind stick vectors recorded at D.F. Malan Airport and at St Helena Bay, air pressure recorded at Langebaanweg and 08h00, 14h00 and 20h00 air temperatures from Langebaanweg and Cape Columbine.

Synoptic conditions on the 14th were characterised by a strong South Atlantic high pressure system beginning to ridge south of the country, advecting cold air inland behind a cold front, which passed on the 13th. Note the relatively low 14h00 air temperatures at Cape Town, Robertson and Vredendal (Fig. 25) and at Langebaanweg and Cape Columbine (Fig. 26d). Compression of the isobars and a strengthening of the pressure gradient along the west coast was responsible for the fresh to strong (> 6 m.s\(^{-1}\)) upwelling-favoured south-easterly winds during the afternoon of the 14th. The superimposed diurnal cycle contributed to a strengthening of this equatorward wind.

The South Atlantic Anticyclone (SAA) continued to ridge south of the country during the 15th and dominated the west coast area, still
Figure 25. Synoptic weather cycle during the field programme, 14-21 October 1986. Air temperatures are at Vredendal, Robertson and Cape Town.

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Figure 26. (a) 2-hourly wind vectors recorded at D.F. Malan Airport, 14-21 October 1986; (b) Winds recorded at St Helena Bay. Gaps in the series are where no measurements were taken; (c) 2-hourly air pressure recorded at Langebaanweg; (d) Smoothed air temperatures from Cape Columbine and Langebaanweg.
advecting cold air inland, as is evident from the inland 14h00
temperature at Robertson of only 17°C. The ridging process of the
SAA resulted in deep upwelling-favourable gale force south-easterly
winds along the west coast. At St Helena Bay a high average wind
speed of 13 m.s⁻¹ was recorded at noon. Later in the day the wind
dropped to an average 9 m.s⁻¹. The wind abated during the night as
the SAA migrated eastwards, thereby allowing a migrating coastal low
to proceed on its southward passage.

On the 16th a major shift from deep, cold south-easterly advection
to warm, shallow south-east advection occurred over the western part
of the country. Calm to light south-easterly conditions present
during the early morning changed to light north-westerly winds,
varying in speed from 3-5 m.s⁻¹, preceding the coastal low during
the latter part of the morning and early afternoon at St Helena Bay
(see Fig. 26b). High temperature offshore airflow was evident at
the beach station (maximum of 30°C) and both Langebaanweg and Cape
Columbine recorded higher air temperatures on this day (see Fig.
26d). The passage of the coastal low at Langebaanweg is indicated
by the distinct drop in air pressure during the afternoon of the
16th (Fig. 26c). During the latter part of the afternoon the wind
swung back to onshore at St Helena Bay. At D.F. Malan the passage
of the coastal low is indicated by a wind calm during the early hours
of the 17th (Fig. 26a).

Further calm weather prevailed on the 17th, with the low pressure
cell along the west coast forming part of a trough system extending
from approximately Luderitz in the north to the South Coast. At St
Helena Bay north-west winds, varying from a calm 0.5 m.s⁻¹ to light
3 m s\(^{-1}\), preceded the arrival of the low pressure system. A drop in air pressure is indicated during the afternoon of the 17th at Langebaanweg (Fig. 26c). High air temperatures were again experienced inland (e.g. Vredendal, Robertson, Langebaanweg) and at the beach station (maximum of 29.5°C), but at the exposed Cape Columbine lower air temperatures, reaching a maximum of only 16°C, were recorded. With the advancement of the coastal low during the latter part of the afternoon the airflow backed to the south-west. This onshore advection was accompanied by overnight fog. Near-perfect weather and sea conditions on the 16th and 17th allowed for the collection of a good data set at sea during the hours 07h00 - 16h00 on both days.

The fog dissipated during the morning of the 18th and another coastal low passed over the area during the afternoon. Winds ranged from calm to light north-west in the early morning to moderate south-west later in the day. High air temperatures were again experienced inland, with Vredendal recording a 14h00 temperature of 35°C. The passage of the coastal low on the 18th is indicated on the pressure record from Langebaanweg as a distinct pressure minimum of 1011.8 mb. Fog formed again overnight and a weak cold front approached the coast from the south-west.

Low stratus clouds on the morning of the 19th cleared by midday and a weak SAA, with a central pressure of only 1020 mb, was present in a more northerly position off the west coast. Air pressure rose in sympathy with the high pressure system to the west, but dropped again with the progression of yet another weakly formed coastal low. Decreased air temperatures were evident both inland and at Cape.
Weak synoptic conditions were found to exist on the 20th, with no marked change in conditions from the previous day. Winds experienced at St Helena Bay were mainly moderate (2-4.5 m s\(^{-1}\)) north-westerly, but swung to the west later in the day. The calm conditions which resulted from the weak pressure gradient allowed for the formation of overnight fog, which dissipated between 09h00 and 10h00 (LST) on the morning of the 21st. With the approach of the cold front to the south-west of the country, west to north-west winds, reaching a high of 9 m s\(^{-1}\) at 14h00, prevailed over the west coast on the final day of the field programme.

To summarise, the synoptic conditions experienced on the west coast after the progression of a cold front on the 13th, consisted of an initial strongly developed SAA, which dominated the area, causing an influx of cold air and deep, strong upwelling-favourable south-easterly winds. Ridging of the high pressure system south of the country further strengthened the south-easterly winds and was the forcing mechanism for the first coastal low which migrated down the coast on the 16th. A semi-permanent low pressure trough dominated the interior and cold fronts passed well to the south of the country during the following four days. The weak pressure gradients and unusually frequent progression of coastal lows allowed for calm and warmer weather conditions with only weak periodic atmospheric forcing for upwelling from 16-21 October.

It is of interest to note the marked difference in air temperature between Langebaanweg and Cape Columbine indicated in Figure 26d.
The air temperature at the upwelling centre of Cape Columbine depicts a much weaker diurnal variation than that at Langebaanweg only 15 km inland. This is due to the direct influence and moderation of the air temperature by the cold seawater at the upwelling site. During offshore wind conditions and calms, as experienced on the 16-18th, heating of the air at Langebaanweg is maximised. The converse is, however, true at night when cooling of the air inland is more pronounced than at the site near the sea.
Heat budget results

Incident solar radiation
An initial aspect of the data analysis involved an assessment of whether the incident solar radiation received at the coast differed from that received ± 10 n.mi offshore at the sea stations. Because the distances between the beach and sea stations were fairly small, no significant variation was in fact anticipated. Only data measured at both stations and at the same time (LST) could be utilised for comparison purposes. Initial impressions gained from the two data sets suggested that differences between the sites were small and that the amount of variation usually fell within the percentage of instrument error (± 5%). A statistical t-test performed on the data pairs revealed that no significant difference between incident radiation recorded at the beach and that recorded a relatively short distance (± 10 n.mi) offshore was evident. A description of the statistical test is given in Appendix 5.

It therefore became possible to combine and average the incident radiation received at the two stations in St Helena Bay in order to represent the area as a whole. Hourly values of incident solar radiation, representative of St Helena Bay, for the days 14-21 October 1986 are plotted in time series in Figure 27a. Where data gaps occurred extrapolations of the values were made through a knowledge of sun angles (appendix 6), cloud cover and a direct measurement of solar radiation at some hour during the day when skies were clear. The diurnal cycle obviously dominates the time series and was found to be fairly regular during the entire study period. Peak midday radiation values were high, varying over the
Figure 27. (a) Incident radiation measured at St Helena Bay, 14-21 October 1986. Dashed lines are the extrapolated values.
(b) Incident radiation for St Helena Bay predicted by Lumb's formula.
(c) Hourly air temperature recorded at Langebaanweg, 14-21 October 1986.
Figure 28. (a) Daily averages of insolation, $Q_l$; back radiation, $Q_b$; sensible heat flux, $Q_h$; and latent heat flux, $Q_e$: 14-21 October 1986 (b) Daily averages of net radiation, $R_n$; and total heat flux, $Q_t$. 
eight days from 980-1030 W.m\(^{-2}\). Little significant day to day or synoptic variation in the incident radiation received occurred, but one must bear in mind that the synoptic cycle was weak during the eight-day period.

High daily averages of incident solar radiation were recorded over the eight days with values ranging from 290-336 W.m\(^{-2}\) (see Fig. 28a). The daily averages are tabulated in Table 1. The minimum of 290 W.m\(^{-2}\) on the 19th was due to fog and low stratus clouds preventing a portion of the incident solar radiation from reaching the sea surface during the early part of the morning. The presence of fog on the early morning of the 18th and the mornings of the 19th and 21st is reflected as a negative skewness of the radiation curves (Fig. 27a).
TABLE 1. Daily average values of surface energy budget terms (Wm$^{-2}$) recorded at St Helena Bay, 14-21 October 1986.

<table>
<thead>
<tr>
<th>Days of Oct '86</th>
<th>$Q_i$</th>
<th>$Q_o$</th>
<th>$m$</th>
<th>$Q_e$</th>
<th>$Q_n$</th>
<th>$Q_t$</th>
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</thead>
<tbody>
<tr>
<td>14</td>
<td>311</td>
<td>80</td>
<td>215</td>
<td>40</td>
<td>-17</td>
<td>192</td>
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<td>323</td>
<td>74</td>
<td>233</td>
<td>4</td>
<td>-24</td>
<td>253</td>
</tr>
<tr>
<td>18</td>
<td>322</td>
<td>75</td>
<td>248</td>
<td>14</td>
<td>-10</td>
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</tr>
<tr>
<td>19</td>
<td>290</td>
<td>60</td>
<td>208</td>
<td>27</td>
<td>-12</td>
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<tr>
<td>20</td>
<td>336</td>
<td>65</td>
<td>242</td>
<td>2</td>
<td>-11</td>
<td>251</td>
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<td>58</td>
<td>242</td>
<td>23</td>
<td>-16</td>
<td>235</td>
</tr>
<tr>
<td>Ave. over 8 days</td>
<td>319</td>
<td>71</td>
<td>231</td>
<td>26</td>
<td>-22</td>
<td>227</td>
</tr>
</tbody>
</table>
The incident radiation recorded at St Helena Bay was found to exhibit a good agreement with values predicted over the eight-day period according to the formula given by Lumb (1964). Lumb's formula for clear-sky conditions is:

\[ Q_o = 1353 \times (0.61 + 0.20 \times s) \]

where \( Q_o = \) clear-sky insolation (W.m\(^{-2}\))
\( s = \) sine of solar altitude.

This formula could be adjusted for cloud cover according to the following:

\[ Q_s = Q_o \times (1 - 0.62 \times C + 0.0019 \times \alpha) \]

where \( Q_s = \) insolation under cloudy conditions
\( Q_o = \) insolation under clear skies
\( C = \) cloud cover (tenths)
\( \alpha = \) noon sun altitude

The predicted values shown in Figure 27b only slightly exceeded those recorded at St Helena Bay during the same period.

Hourly air temperatures recorded at the inland (\(\sim 15 \text{ km}\)) station at Langebaanweg (Fig. 27c) indicate the warming effect during the diurnal and synoptic cycles. The air temperature generally follows the diurnal variation of solar radiation. The higher daytime air temperatures evident on the 16-18 can be ascribed to the influence of low velocity warm offshore winds preceding the coastal low.
pressure systems which passed over the area.

(ii) Back radiation
On the whole, it was found that back radiation values at St Helena Bay were relatively low, averaging 71 Wm\(^{-2}\) during the field study. Factors responsible for the relatively low values include low SST's (average of 14.9°C in St Helena Bay during the field study) and a relatively high value of vapour pressure (average 15.7 mb over the eight days) due to the trapping of moisture near the surface by stable air layers. A factor which significantly affects the amount of back radiation is the proportion of sky covered by cloud. The presence of fog on the nights of the 17th, 18th and 20th had the effect of reducing back radiation to a mere 30 Wm\(^{-2}\) (see Fig. 29a). Extrapolations of values to a reasonable degree of accuracy were permissible through a knowledge of the cloud cover, averaged SST and vapour pressure. On a daily average (see Fig. 28) little variation was evident, only a slight decrease on the days when fog was present.

(iii) Net radiation
The net radiation into the sea surface was calculated through use of equation (2) in Chapter A(3). The incident radiation, albedo of the sea surface, and back radiation are taken into account. The loss terms of the equation, i.e. the sea surface albedo, which averages 6% of incident radiation, and the back radiation, which averages approximately 22% of incident radiation, are fairly constant and do not significantly alter the trend of incident radiation. The net radiation received at St Helena Bay during the field programme is illustrated in Fig. 29b. Where data gaps occurred in the time
Figure 29. (a) Back radiation measured at St Helena Bay, 14-21 October 1986. Dashed lines are the extrapolated values.
(b) Net radiation at St Helena Bay during the same time span.
(c) Heat balance (total heat flux) at St Helena Bay.
series, values were extrapolated through use of the extrapolated incident radiation and back radiation values and through a knowledge of cloud cover and wind conditions. The net radiation received obviously follows the trend of incident radiation and therefore compares favourably with Figure 27a. Peak noon values over the eight days of the study reached 860-890 W.m\(^{-2}\) and daily averages varied from 208-248 W.m\(^{-2}\) (see Fig. 28b and Table 1).

A higher daily averaged net radiation was evident where fog prevailed overnight followed by a dissipation in the early morning, as occurred on the 18th and 21st. This has the effect of minimising the back radiation overnight, while clear skies during the day permit maximum solar radiation. Where fog and cloud cover persisted to a slightly greater extent during the day, as occurred on the 19th, both incident radiation and back radiation are diminished, but because the magnitude of incident radiation lost is greater, the net effect is an overall decrease in net radiation for the day. Where atmospheric conditions are clear both at night and during the day, both the back radiation and the incident radiation are maximised, but the greater magnitude of incident radiation determines that an increased amount of net radiation is received for the day.

(iv) **The turbulent heat fluxes**

As stated in Chapter A(3), the main factors which determine turbulent heat transfer between the air and sea are wind speed and the temperature contrast between the sea surface and the overlying air. During the day the air temperature is almost always greater than the SST in this area, while at night the reverse might occur,
although contrasts are small.

Examination of the latent and sensible heat fluxes, estimated from the bulk aerodynamic formulae, at the beach and sea stations at St Helena Bay revealed that, as expected, magnitudes were small in comparison with the solar radiation received. The latent heat flux amounted to a small heat loss from the sea surface and the sensible heat flux represented a small heat gain by the sea surface. The fluxes generally operate in opposite directions over the cold waters and the calculated values often, in fact, lie within the region of error of the insolation measurements.

These energy transfer processes do not proceed continuously, but in an interrupted, pulsating form (Perry & Walker, 1977) as a result of short-term fluctuations in the wind speed and air and sea surface temperatures over various time scales. Due to this short-term variability it is easier to deal with average values taken over a longer time span to create a general idea of conditions. The fluxes are therefore averaged for the purposes of the study and data from the beach and sea stations was grouped together to represent the St Helena Bay area as a whole.

Daily averaged values of sensible and latent heat transfer during the study period are indicated in Figure 28a. The relatively small values are shown as approximate converses of each other. Variations in the wind speed and its associated turbulence appears to be the main cause of variations in the fluxes. Both fluxes show maximum heat gain and heat loss respectively by the sea on the 15th, when high velocity winds prevailed, generating increased turbulence.
Averaged over the eight days, the small sensible heat gain of 22 W.m\(^{-2}\) amounts to 7% of the average amount of incident radiation received (see Table 1), while the small latent heat loss of average 26 W.m\(^{-2}\) represents 8% of the incident radiation amount. Due to the fact that the turbulent fluxes are approximately equal and opposite, their effect on the overall heat balance is minimal.

(v) Heat balance at St Helena Bay

Calculated values of the heat balance at the beach station and at the sea stations located approximately 10 n.mi offshore were compared to assess whether any differences existed between the coastal and offshore sites. Because both the incident and back radiation exhibited little variation between the sites and the turbulent heat fluxes approximately cancel each other, no variation in the heat balance between the coastal and offshore sites was, in fact, anticipated. A statistical t-test (see Appendix 7) confirmed our expectations.

Variations in the heat balance over the diurnal period were obviously dominated by the diurnal trend of incident radiation (see Fig. 29c & cf. Fig. 29b). Peak hourly values of the total amount of heat available to the sea surface varied from 808 W.m\(^{-2}\) on the slightly more overcast day of the 19th, to a high of 950 W.m\(^{-2}\) on the 17th. The trend over the synoptic period was, likewise, also dominated by the radiation received, the other components of the heat budget making negligible contributions in comparison. Daily averages of the energy available for heating of the surface Benguela waters over the eight days in October varied from 192 W.m\(^{-2}\) on the 14th to a high of 253 W.m\(^{-2}\) on the 17th (see Fig. 28b and Table 1).
A large surplus of energy (average of 227 W.m\(^{-2}\) over the eight days) was, therefore, available to heat the oceans in the St Helena Bay area during the field study period of 14-21 October 1986. The essential result of the field study is that the major contributing term to a high heat balance in the St Helena Bay area is the input solar radiation.
Overview

The overall aim of this thesis is to examine and characterise the observed heat transfer processes within the overall heat balance between the ocean and atmosphere in the coastal upwelling environment of the Benguela region.

In realising this aim it became clear that synoptic scale variability could play an unusually important role in characterising the longer term heat transfer processes. A further striking feature of the Benguela region is the weakness of change in the coastal environment longshore throughout the entire region, as compared with the strong change from the coastal to the oceanic environment offshore.

The field study in St Helena Bay provided vital information on the magnitude and relative importance of the various heat transfer processes within the heat balance at the sea surface. It was intended to highlight the synoptic scale variability, and with the inclusion of further observations from neighbouring sites, this became possible. The longshore similarity of conditions within the Benguela coastal environment means that the characterisations arising from the field study are representative of the Benguela region as a whole.

In Section B, the topics were introduced in the order of decreasing
spatial and temporal scales. In order to recognise the central place of the field study at St Helena Bay, the discussion reverses the scale order. It begins with the interpretation of the results of the field study and establishes the nature of the heat budget contributions. This is followed by an interpretation of the observed offshore trends in terms of the model application. The scale is then broadened to consider the synoptic variability of the heat budget and its components. Results gained from the eight-day field trip are used, but in order to utilise a more complete data set, global radiation values recorded at D.F. Malan are also applied. The field results are then extended longshore to characterise the heat budget processes relevant to the entire Benguela area. Additional data from Alexander Bay and D.F. Malan is used to confirm the expected trends.

A further objective of the study was to examine the effects of a high heat input to the cold surface waters of the Benguela. Temperature variations recorded at sea during the study are examined with the aid of satellite-derived SST maps of the area. An investigation is made into the degree of change of the cold upwelled water as a result of a high positive heat balance.

Over a larger scale the results gained from the study are applied to the general climatology of the Benguela area. The discussion ends with a summary of the essential features of the heat balance typical of an upwelling area.
1. **INTERPRETATION OF FIELD RESULTS**

Results obtained from the field study indicate that solar radiation is the most important term in the heat balance of the Benguela upwelling area. Data obtained from the field study revealed that both synoptic and diurnal fluctuations in the heat budget are governed by the incident radiation term, $Q_1$. A large amount of incident heat is available to heat the ocean water in this area. The incident solar radiation is only modified by the back radiation, which approximated an almost constant loss of 71 W.m$^{-2}$ during the field study. Variations in $Q_b$ only occurred with increasing cloud cover, 10/10 cloud cover diminishing the loss to approximately 30 W.m$^{-2}$.

The latent and sensible heat fluxes at the typical Benguela site of St Helena Bay were calculated to be approximate converses of each other. Thus from the heat balance equation $Q_t = R_n - Q_e - Q_h$ it is apparent that the $Q_e$ and $Q_h$ terms cancel out, leaving the heat balance term, $Q_t$, to be approximated by the net radiation, $R_n$. Because the loss terms of $R_n = (Q_1 + q)(1 - \alpha) - Q_b$ are roughly constant, any fluctuations in $R_n$ will be governed by fluctuations in the incident radiation term, $Q_1$.

The calculated $Q_t$ and $R_n$ values derived from the measurements made at St Helena Bay were statistically compared to investigate as to whether any significant differences existed between them. The t-test performed on the data revealed that it is 99% certain that no significant difference between the net radiation ($R_n$) and the total heat flux ($Q_t$) exists (see Appendix 8). The net radiation is
therefore the main contributor to a high heat input to the waters of St Helena Bay. Daily averages of $R_n$ and $Q_c$ are depicted together in graphical form in Figure 28b for comparison purposes. The distribution reveals that departure of the average daily total heat balance from the average daily net radiation is minimal. The average net radiation measured over the eight days amounted to 231 W.m\(^{-2}\), slightly more than the average total heat input of 227 W.m\(^{-2}\). This insignificant deviation can be accounted for by the fact that the latent heat flux slightly exceeded the sensible heat flux by an average 4 W.m\(^{-2}\) (see Table 1).

The relationship between the 2 variables, $Q_c$ and $R_n$, was found to be significant, with a correlation coefficient of 0.992 indicating a very strong association (see Appendix 9). This linear relationship is graphically depicted in a scatter diagram (Fig. 30). The goodness of fit linear regression line represented by the equation $y = -9 + 1.04x$ is superimposed in order to indicate the strong association between the 2 variables.

The coastal and offshore data measured over the eight days was grouped together according to hour of day and averaged, in order to provide an average diurnal trend of the heat budget components at St Helena Bay for the period of the field study. The resultant trend is illustrated in Figure 31a. The curve of average daily insolation, $Q_i$, is typically sinusoidal in shape. Daytime insolation values are noted to be high, with the average peak at noon almost attaining 1000 W.m\(^{-2}\). Little variation in back radiation is evident, only a slight decrease, roughly from the hours 23h00 to 08-09h00, due to the effect of water vapour in fog acting to minimise the back
Figure 30. Scattergram of the relationship between net radiation and the heat balance over the Benguela area, 14-21 October 1986. Solid line is the best fit regression line.
Figure 31. Average diurnal trend of (a) the heat budget components, $Q_i$, $Q_b$, $Q_h$, and $Q_e$; and (b) net radiation ($R_n$) and the total heat flux ($Q_t$); 14-21 October 1986.
radiation from the sea surface.

The turbulent heat fluxes are almost complete converses of each other, with both reaching maxima during the late afternoon when the sea-breeze causes an increase in wind speed and turbulence. Heating of the air by solar insolation during the day results in a sensible heat gain by the sea which, in conjunction with the effect of the strengthening sea-breeze during the afternoon, increases from about 12h00-17h00. The latent heat flux is always negative. There is a general drop-off in the fluxes at night, with both reaching a minimum near the time of sunrise. A decrease in the air temperature at night implies that the SST may then exceed the air temperature, causing a reversal in the sensible heat flux from a heat gain to a heat loss between the hours of approximately 20h00-05h30. Magnitudes are, however, small (1-2 W.m⁻²) due to the absence of strong winds at these hours. The latent heat flux tends to slightly exceed the sensible heat loss.

The average diurnal trends of \( R_n \) and \( Q_l \) show a good agreement with each other (see Fig. 31b). The only visible differences occur around noon when higher air temperatures contribute to an increased sensible heat flux into the ocean, resulting in an increased heat balance (898 W.m⁻²), and at night when both turbulent heat fluxes represent losses from the ocean surface and, therefore, effectively contribute to a slightly greater total heat loss.

This situation, where the surface exchange of heat is dominated by the net radiation was indicated by Reed (1978) to be a rare occurrence. Reasons for this rare situation were listed as winds
being light to moderate, air vapour pressure being greater than water vapour pressure and sea - air temperatures being small. Over the Benguela upwelling area the main reason for the surface heat exchange being dominated by net radiation lies in the fact that the turbulent heat fluxes of latent and sensible heat are usually equal and act in opposite directions, thus cancelling each other. The facts that the air vapour pressure is usually greater than the water vapour pressure and that sea - air temperatures are small and usually negative implies that both turbulent heat fluxes are normally small in magnitude in comparison to the net radiation term. Winds can reach gale force over the Benguela area, thereby increasing the turbulent heat fluxes, but effects on the total heat balance are, nevertheless, still negligible as the increased turbulent heat fluxes will continue to act in opposite directions and therefore still approximately cancel each other. Thus Reed's pre-requisite that wind speeds be light to moderate is not fundamental to the application over the cold Benguela waters.
2. APPLICATIONS TO THE MODEL

The results from the field study, particularly in respect of comparisons between the beach station and the sea stations in St Helena Bay, can be used to assess the usefulness of the conclusions of the idealised model. In its most extreme case the model assumed a Berg wind situation preceding a coastal low, with a steady warm, dry offshore airflow, and a low air temperature inversion level. On the whole, the system was assumed to be closed with only a redistribution of heat from air to sea taking place. Sea conditions were assumed to be steady. An important conclusion of the model was that the actual fluxes of sensible and latent heat are confined to the coastal region. These fluxes decrease exponentially in an offshore direction.

Applied to the field study results, the model predicts that the turbulent heat fluxes recorded at the beach station should exceed those recorded at the sea stations approximately 10 n.mi offshore. Corresponding values of both latent and sensible heat fluxes recorded at the beach and sea stations were compared in order to investigate as to whether any significant differences existed over this +10 n.mi distance. From the matching values which were compared, it was found that both the sensible and the latent heat fluxes recorded at the beach approximately doubled those recorded +10 n.mi offshore. Thus, on average, it would appear that a greater exchange of turbulent heat occurred at the coast than at the offshore sites, but because these exchanges were small (7-9% of incident radiation received) and operated in opposite directions, any effect on the overall heat balance between the different sites
Statistical tests on the different data sets revealed that it was 95% certain that both turbulent heat fluxes were greater at the beach than at the sites in St Helena Bay (see Appendixes 9 and 10). This hypothesis was, however, rejected at the 99% level of significance. Reasons for the rejection at the higher level of significance probably lie in the facts that the turbulent heat fluxes are highly variable and a large standard deviation has the effect of reducing the calculated 't' values required for the significance test. It is clear that further research involving a greater number of sea stations positioned further offshore would be required to more accurately assess the trend predicted by the model.

Analysis of the variables which affect the turbulent heat fluxes revealed that both the temperature and specific humidity deviations between the air and sea showed little variation between the beach and the offshore positions. The heat flux differences between the different sites must therefore be due, in part, to a greater wind speed at the beach than slightly offshore. Reference to the wind speeds recorded at the different sites confirmed this trend. This obviously dampens the conclusions gained from the model, as the idealised situation assumed a constant wind speed from the coast offshore. The distance between the beach and sea stations of approximately 10 n.mi is also not really sufficiently large to indicate any substantial differences in the air-sea temperature and humidity deviations.

The question as to the usefulness of the model must then rely on
more widely spaced data. Turbulent heat fluxes recorded from a
greater number of stations by Bowden (1977) off Cabo Bojoador, North
West Africa, can be applied to the model to confirm the expected
features. Bowden found a clear systematic variation of $Q_h$ and $Q_e$
with distance from the coast (see Fig. 32). The turbulent heat flux
(sensible and latent heat) was found to decrease on approaching the
coastal area and became negative (i.e. a heat gain) within
approximately 20 km from the coast. A corresponding increase in the
net gain of the sea surface was also observed. The increased
negative values of the turbulent heat flux near the coast are the
result of a greater sensible heat exchange to the sea surface in the
near-shore region, where the air - sea temperature contrast is
greatest. The turbulent heat fluxes approximate zero at the
transition area approximately 20 km offshore from the coast because
the air properties have been altered so that the temperature and
humidity of the air equals that of the sea surface. This trend
supports the model theory. Further offshore the turbulent heat flux
is positive (heat loss) due to the presence of warmer water.

The results gained by Bowden (1979) from North West Africa are
typical of an upwelling area and confirm the conditions expected
over the Benguela upwelling area. In a longshore direction little
variation of these typical features is expected. Thus the situation
can be assumed to be representative of the entire Benguela upwelling
region. The use of coastal measurements from one site along the
west coast can therefore justifiably be used to typify conditions
along the entire west coast.
Figure 32. Heat flux, $Q_h$ and $Q_a$ in cal.cm$^{-2}$.hr$^{-1}$ at Cabo Bojador stations, north-west Africa (after Bowden, 1977). Negative values are heat gains by the sea ($1 \text{ cal.cm}^{-2} \text{.hr}^{-1} = 11.63 \text{ W.m}^{-2}$).
3. SYNOPTIC VARIABILITY

One of the aims of the study was to examine the variation in the heat budget (and its components) over the synoptic time scale of a few days. A synoptic 6-day cycle in the weather systems along the west coast has been indicated by Preston-Whyte & Tyson (1973). Large ranges in the environmental parameters over a few days should therefore result in similarly large variations in the heat balance over the same time scale.

Results gained from the field programme, however, indicated that minimal variation in the heat balance occurred during the weak synoptic cycle over the eight days. The consistently high values of solar radiation received, and therefore the high heat balance throughout the eight days, can be explained by the fact that skies were clear throughout almost the entire study period. Fog and low stratus cloud was present on only 3 of the mornings. The fact that the field programme coincided with an anomalously weak synoptic cycle therefore limits the interpretation of any synoptic variability in the heat balance from the field data alone. Had the field programme been a few days longer, more pronounced synoptic events would have revealed greater synoptic variability in the heat balance and in the environmental factors which control it.

Due to the limited synoptic application of the data recorded at St Helena Bay, it was deemed necessary to make use of additional data measured on a continual basis by the Weather Bureau to provide a more complete assessment of the synoptic variation. The only continuous measurement of global radiation at a site representative
of the west coast of southern Africa is made at D.F. Malan Airport, Cape Town, which represents the southern extremity of the Benguela system. Only long-term data is available from Alexander Bay, 450 km to the north, as the continuous measurement of solar radiation has been discontinued at that site.

Hourly global radiation received at D.F. Malan during the coincident eight-day period, 14-21 October 1986, is illustrated in Figure 33a in order to make comparison studies with the St Helena Bay data over the diurnal and synoptic scales. Compared to Figure 27a of solar radiation measured at St Helena Bay, little variation between the two locations is evident. Peak midday values at D.F. Malan varied from 960-1040 W.m\(^{-2}\) during the study period, as compared to the 980-1030 W.m\(^{-2}\) range at St Helena Bay. Air temperatures recorded at D.F. Malan (Fig. 33b) also showed a similar diurnal and synoptic trend to those recorded at Langebaanweg (Fig. 27c). Daily average global radiation received at D.F. Malan during the eight-day period ranged from 290-315 W.m\(^{-2}\), while at St Helena Bay the values ranged from 290-336 W.m\(^{-2}\) (see Table 2).
Figure 33. (a) Incident radiation measured at D.F. Malan Airport, 14-21 October 1986.
(b) Hourly air temperature recorded at D.F. Malan Airport during the same time period.
TABLE 2. Daily average global radiation recorded at Cape Town (D.F. Malan airport) and at St Helena Bay: 14-21 October 1986.

<table>
<thead>
<tr>
<th>Days of October</th>
<th>D.F. Malan</th>
<th>St Helena Bay</th>
</tr>
</thead>
<tbody>
<tr>
<td>14</td>
<td>290</td>
<td>311</td>
</tr>
<tr>
<td>15</td>
<td>303</td>
<td>327</td>
</tr>
<tr>
<td>16</td>
<td>311</td>
<td>329</td>
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<td>17</td>
<td>313</td>
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<td>18</td>
<td>305</td>
<td>322</td>
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<td>19</td>
<td>300</td>
<td>290</td>
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<tr>
<td>20</td>
<td>315</td>
<td>336</td>
</tr>
<tr>
<td>21</td>
<td>299</td>
<td>313</td>
</tr>
</tbody>
</table>

| Ave over 8 days | 305 | 319 |

The daily averages of global radiation received at St Helena Bay superimposed on the graph of daily averages recorded at D.F. Malan for the entire month of October (Fig. 34) indicate that, on the whole, the radiation received during the field programme at St Helena Bay exceeded that received during the same period at D.F. Malan. Global radiation at St Helena Bay over the eight days was calculated to average 4.7% more than the radiation recorded at D.F. Malan. This variation in incident radiation received is realistic if one takes into account the fact that D.F. Malan is situated 130 km to the south of St Helena Bay and thus potentially receives less radiation than those areas further north. The incident radiation
Figure 34. Daily average incident radiation, including the scattered component, measured at D.F. Malan Airport for the month of October 1986. Daily average incident radiation measured at St Helena Bay from 14-21 October is superimposed.
recorded at D.F. Malan, although on average slightly less than that recorded at St Helena Bay, can be extended some distance alongshore and used to approximate conditions experienced in the southern Benguela.

The daily average global radiation recorded at D.F. Malan for October (Fig. 34) serves to illustrate the fact that the eight-day period of the field study coincided with a synoptic period characterised by clear skies and maximised incident radiation. From the graph it is evident that prior to the field study daily fluctuations in the radiation received occurred. With the passage of a cold front on the 12th, average global radiation dropped to a distinct minimum of 62.5 W.m\(^{-2}\). Thereafter the somewhat extended period of increased solar heating for the duration of the field study occurred, with an average incident radiation of 304.5 W.m\(^{-2}\) per day being received at D.F. Malan and a higher 318.9 W.m\(^{-2}\) average being received at St Helena Bay over the eight days. This event was punctuated by the passage of a cold front on the 23rd, when only 99 W.m\(^{-2}\) was received for the day, after which daily fluctuations were again the norm.

The average diurnal trend of incident radiation received at St Helena Bay during the eight days was also compared to the October average diurnal trends at D.F. Malan and Alexander Bay. From the graphs represented in Figure 35 it is clear that the radiation received at St Helena Bay from 14-21 October 1986 exceeds the long-term diurnal October averages at both D.F. Malan and Alexander Bay. The eight-day daily average global radiation of 318 W.m\(^{-2}\) also far exceeds the October averages of 287.8 W.m\(^{-2}\) and 253.6 W.m\(^{-2}\) for
Figure 35. Average diurnal trend of global radiation derived from long-term data for the month of October at D.F. Malan and Alexander Bay compared to the 8-day diurnal average from St Helena Bay (14-21 October 1986).
Alexander Bay and D.F. Malan respectively. An anomalously high amount of heat was thus available to the Benguela area during the period of the field study.

Over the synoptic time scale, the results of the field study and the more complete D.F. Malan data set can be used to characterise the heat budget and its components over the Benguela system for 3 main synoptic events: (i) deep southerly wind advection; (ii) coastal lows, and (iii) cold fronts. Little variation in the heat balance was evident between the first two cases, only slight variations in the components thereof, while the third case exhibits a pronounced variation from the first two cases.

(i) Deep southerly wind advection

Typical conditions consist of the South Atlantic Anticyclone (SAA) ridging south of the country, forcing strong equatorward southerly winds along the west coast (see Fig. 15b and Fig. 25b). Skies are generally clear, thus maximum solar radiation is allowed to penetrate the sea surface throughout the day. The back radiation term exhibits minimal variation, but the turbulent heat fluxes increase as a result of the high velocity winds (cf. 15 Oct 1986). Due to the fact that these turbulent fluxes transfer heat in opposite directions (sensible heat to the ocean, latent heat to the air) their effect on the heat balance is largely damped out and the heat balance is determined by the net radiation. The increased evaporation may cause a greater amount of moisture to be present in the air and to higher levels. This may have the effect of absorbing a slightly greater proportion of the incident radiation before it reaches the surface.
Convection due to wind mixing of the surface mixed layer immediately redistributes any heat gained by the surface ocean layer downwards, so that the temperature profile of the mixed layer does not exhibit a strong thermocline. All the available heat is used to heat the ascending upwelled water. When the temperature of the surface water is decreased as a result of upwelling, the sensible heat flux into the water increases and the overlying air layer stabilises, ultimately resulting in a decrease in the wind stress. Both these effects act to increase the SST, thus acting against the initial temperature decrease.

(ii) Coastal lows

This situation consists of a coastal low migrating from north to south down the coast, with warm, dry, offshore winds ahead of the low followed by cool, moist, onshore advection behind it (see Figs. 15c & 25c,d,e). Preceding the low, skies are clear, permitting maximum insolation at the surface. Weather conditions are relatively calm, thus the turbulent heat fluxes are both small and continue to transfer heat in opposite directions. These fluxes do not affect the heat balance prior to the passage of the low, therefore the heat balance is determined by the net radiation. There is thus little variation in the overall heat balance from case (i).

The effect of a lowered inversion preceding the coastal low is to limit the depth of convection in the air. According to the model theory the rate of change of the air properties from warm and dry at the coast to cool and moist offshore will increase with a decreasing inversion level. Calm, often glassy, sea conditions prevail and
heating of the surface layer during the day may be intense as little wind mixing takes place. Surface warming due to solar radiation therefore produces shallow thermoclines within the upwelled area.

Once the low has passed cool, moist air is advected onshore, sometimes accompanied by low stratus clouds or fog. The stratus cloud or fog occurs more commonly at night, when its net effect is to cause an increase in the heat balance due to the fact that the heat loss from the back radiation is reduced. When fog is present the air properties are so similar to the sea that the turbulent fluxes approximate zero. Should fog persist throughout the day, however, both incident radiation and back radiation are diminished, resulting in a decreased overall heat balance. More commonly, the fog dissipates during the morning as a result of solar heating. The overall heat balance will then exhibit little variation from case (i).

(iii) Cold fronts
Low pressure cells accompanied by cold fronts, which originate as perturbations on the circumpolar jetstream, are known to periodically influence the southern Benguela area during winter. Higher velocity north-westerly winds ahead of the front will have the effect of increasing the turbulent heat fluxes (still opposite in sign), but as skies are usually clear prior to the passage of the front, the heat balance will remain relatively high and exhibit little variation from cases (i) or (ii). Cirrus clouds just preceding the front may have the effect of reducing the insolation and therefore the heat balance.
With the onset of cloudy, cooler weather, often accompanied by rain, the incident radiation received at the surface is drastically reduced. Back radiation losses are subsequently diminished and the colder air may contribute to a sensible heat loss from the ocean. Because the latent heat flux is also a loss term, the turbulent heat fluxes both represent a loss of heat from the ocean surface. This, together with the fact that a large proportion of incident radiation is reduced by the cloud cover, results in a sometimes dramatically decreased heat balance over the Benguela region during the passage of a cold frontal system.

From the above-mentioned synoptic cases it is evident that variations in the total energy flux into the ocean on the west coast of southern Africa are dominated by variations in the solar radiation, which is moderated by the cloud cover. The extent of cloud cover is in turn determined by the phase of the synoptic cycle. A somewhat surprising observation from the west coast is that variability is not at its maximum on the seasonal time scale, but on the synoptic time scale, where large ranges in the environmental parameters may occur over a few days. Because solar radiation is the main contributor to the high heat balance over the west coast, and longshore variations are insignificant, the global radiation data recorded from D.F. Malan Airport can be used to illustrate this synoptic variability over the southern Benguela area.

As an example, note the variability in daily average global radiation recorded at D.F. Malan between the dates 12th and 13th October 1986 (see Fig. 34). With the passage of a cold front on the
12th, a low average daily global radiation of 62.5 W.m\(^{-2}\) was recorded, while on the following day after the passage of the front, a high daily average of 305.4 W.m\(^{-2}\) was measured. This variation of 243 W.m\(^{-2}\) exceeds the large seasonal variation of 226.5 W.m\(^{-2}\) at D.F. Malan. During December and January, when the sun is at its most southern extent over the southern hemisphere, more heat is obviously available and even greater day to day variations are possible, e.g. during January 1986 the range in average daily global radiation between 2 consecutive days amounted to 274 W.m\(^{-2}\). Thus, if it is unusual environmental conditions which lead to the high heat gain in January (summer) as a whole, there will be certain periods during January when the environmental conditions will be even more unusual and the heat gain by the sea surface will be even higher.
Because minimal longshore variability is evident along the west coast of southern Africa, results gained from the localised site at St Helena Bay may be regarded as broadly representative of the Benguela system as a whole. The heat balance and the controls thereof exhibit a consistency in a longshore direction. The presence of cold surface waters along the entire west coast upwelling area causes the turbulent heat fluxes to generally be small in comparison to the total insolation received. These turbulent fluxes were found to transfer heat in opposite directions over the cold water. The latent heat flux may be expected to be greater in the more extreme northern areas during late summer as a result of the advection of warmer Angolan waters southward. Heat gains will consequently be slightly less. Relatively little variation in back radiation along the coast is anticipated, only a slight decrease to the north as a consequence of increased cloud cover. The average sum of heat loss to the atmosphere from the surface of the Benguela is fairly consistent along the coast, approximating the back radiation amount of some 71 W.m\(^{-2}\).

Results gained from the field programme indicated that the net radiation received was the main contributor to the high heat balance over the St Helena Bay area. This application may be extended in a longshore direction to represent the entire Benguela upwelling area. The net radiation is obviously determined by the amount of solar radiation received. Thus an indication of the magnitude of the overall heat gain by the Benguela upwelling area can be obtained by referring to values of global radiation recorded along the coast.
Figure 36. Annual variation of global radiation at Cape Town (D.F. Malan) and Alexander Bay (after Tegen, 1987).
Long-term global radiation representative of the west coast is only available from two meteorological stations, viz, D.F. Malan Airport and Alexander Bay. Both stations are situated slightly inland from the coast, but because these inland distances are small little variation in global radiation between the coast and the slightly inland positions is anticipated. The global radiation recorded at the stations can therefore be treated as representative of the surrounding coastal areas.

Annual patterns of global radiation received at D.F. Malan and Alexander Bay are shown in Figure 36. Hourly global radiation data from a 13-year record processed by Tegen (1987) was used to obtain the illustrations. Situated further north, Alexander Bay receives a greater amount of solar radiation than D.F. Malan throughout the year, with annual averages reaching 248.7 W.m\(^{-2}\) and 216.1 W.m\(^{-2}\) respectively. Both stations record highest values during December (342.7 W.m\(^{-2}\) and 326.2 W.m\(^{-2}\) respectively) and minimum values during June (148.5 W.m\(^{-2}\) and 99.7 W.m\(^{-2}\) respectively), but the annual range is greater in the south at D.F. Malan (226.5 W.m\(^{-2}\)) than at the more northerly positioned Alexander Bay (194.2 W.m\(^{-2}\)).

The northern Benguela receives more insolation than the southern part, but because the cloud cover generally increases to the north and in an offshore direction, the meridional differences are slightly offset. The highest yearly mean net radiation for the South Atlantic Ocean occurs at the coast near 25\(^\circ\)S, but the trade clouds cause a decrease in net radiation farther out to sea (Hoflich, 1984). The probability of fog or low stratus clouds over the cold upwelling water and below the trade inversion is high in the area.
north of about 32°S, with the Walvis Bay region experiencing the highest frequency of about 100 days per year on average (Hoflich, 1984). The frequency of fog decreases out to sea. The fog occurs more commonly during winter in the north, coincident with the season of maximum upwelling and occurs less frequently in the southern areas during summer.

Seasonal variations are greater in the south than in the north due to the increase in latitude and the passage of cold fronts over the southern area during the synoptic cycle in winter. During winter the solar radiation is further reduced over the Benguela region by the cloud cover which accompanies cold fronts in the south, and in the north due to an increase in the frequency of fog. Although fog might result in some decreased solar irradiance before it is burned off during the day, the cooling effect during daylight may be compensated or over-ridden by warming during the night by what may be termed the "lid effect" (Tont, 1981), i.e. the cloud cover acts as a lid by keeping in part of the outgoing radiation.

The diurnal variation in incident radiation obviously exhibits a significant control on the heat balance of the Benguela region and is also the forcing mechanism for the diurnal land-sea breeze cycle, which is common along the entire coast. Land-breezes may occur during the night, particularly in winter, when the air temperature drops below that of the sea. The sensible heat flux then becomes a loss from the sea, as does the latent heat flux, but the generally low wind speeds at night prevent any large heat losses from the ocean. During the day the air temperature usually rises above the SST and the sea-breeze intensifies, resulting in increased turbulent
heat conduction to the sea surface and an increased latent heat loss from the surface. The two turbulent heat fluxes roughly cancel out, leaving the heat balance to be dominated by the net radiation.

The reason for the high values of insolation received over this area can probably be explained by the fact that the atmosphere is dry over the west coast of southern Africa. The low water vapour content of the atmosphere above the Benguela upwelling area (see Section B1 and Fig. 5) and limited cloud cover allows for the penetration of a greater amount of incident radiation than in the surrounding oceanic areas. This large amount of radiation is then responsible for the high energy balance of the Benguela area. The consequence of the high positive heat balance over the Benguela is that warming of the cold upwelling water at the surface balances the cooling caused by further upwelling and the horizontal heat transfer of the Benguela Current as it moves north-westwards.
5. OCEANOGRAPHIC IMPLICATIONS

The effect of a high input of heat to the ocean surface obviously has important physical and biological implications to the Benguela system. In order to gain a thorough knowledge of the ongoing physical processes in the ocean which result from strong surface heating, it is important to fully understand the heating process of the ocean water. Unlike the land masses, the oceans, with a high specific heat of $4.2 \times 10^3 \text{ J.kg}^{-1}\text{.deg}^{-1}$, have an enormous capacity to store heat. The rate of heating of the ocean water is, however, slow. The temperature of the sea depends on various physical processes such as the radiation intensity, ability of the sea to absorb radiation, heat exchange between the sea surface and the overlying atmosphere, horizontal and vertical advection, mixing by convective stirring and turbulence, and wave action.

The vertical distribution of solar heating is important as it determines the thermal structure of the upper ocean. Incident radiation that penetrates the sea consists of energy distributed over a wide range of wavelengths. The incident light is scattered by the water and suspended particles contained therein. A small part radiates back as diffuse underlight, which is responsible for the colour sensed by an observer. Selective absorption of the radiation at short wavelengths toward the ultra-violet and at longer wavelengths (red and infra-red) allows only radiation in the blue-green part of the spectrum to penetrate into deeper water. The exponential law which determines the progressive decrease with depth of the solar radiation intensity due to absorption and scattering is as follows:
\[ I_z = I_0 \exp(-kz) \]  

where 

- \( I_0 \) = radiation intensity penetrating at the sea surface
- \( I_z \) = remaining radiation intensity at depth \( z \) metres below surface
- \( k \) = absorption coefficient of water.

Price, et al. (1986) determined that the short wave radiation in fairly clear mid-ocean water is absorbed with an extinction scale of 20 m, and approximately half of the incident radiation is absorbed in the upper metre of the sea. The amount of absorption and scattering varies, depending mainly on the turbidity of the water. The transmission of solar radiation is more efficient in the open ocean, where organic production is low, than in coastal waters, which contain a larger number of impurities. Of the solar radiation absorbed by the sea a negligible amount of 0.02% is fixed by photosynthesis (Williams, et al. 1973), while the remainder is used to heat the upper ocean (Perry & Walker, 1977). The electromagnetic energy of radiation, which is absorbed by the upper layers of the ocean, is converted into internal molecular kinetic energy, which can be sensed as heat and results in an increase in temperature of the water.

Because solar heating is concentrated close to the surface it has a profound effect on the depth of convection (Woods, et al. 1986). In the absence of wind the solar energy flux can produce strong heating within a thin surface layer of the sea, which may turn out to be
Considerably warmer than the first metre of water. Strong surface heating induces stratification, which leads to stability in the surface layers. This has the effect of limiting the vertical mixing in the sea and results in rapid warming of the surface layers. Absorption of heat by the oceans from solar radiation, and in some cases from heat conduction, is most efficient at the surface, but is distributed to greater depths mainly by wind-generated turbulence. The thermal structure of the ocean is therefore usually determined more by wind-mixing than by radiation absorption alone (Price, et al. 1986). In the presence of wind, which is more frequently the case, the effects of solar heating are realised to a greater depth than that reached by solar radiation directly due to the downward vertical wind mixing, which produces a downward heat flux in the ocean. Below the mixed layer heating is weak, but over a longer time scale it can be of importance.

Although diurnal variations in solar heating are pronounced, the variation of solar heating over the synoptic and seasonal scales tends to be of greater importance. Over the west coast of southern Africa the variability, although significant on the seasonal scale, appears to be of greater importance over the synoptic time scale of a few days.

Sea temperature profiles obtained from the field programme were used to examine the effect of a high heat input to the surface mixed layer over the diurnal and synoptic time scales. The active upwelling phase of 3-5 day variability has a pulsing effect on the waters in St Helena Bay within the context of its 25-day residence time (Waldron, 1985), therefore synoptic variations are to be
expected in the surface mixed layer. The existence of a diurnal land-sea breeze cycle superimposed on the synoptic cycle, and strong solar heating during the day exceeding any heat losses, contributes to a diurnal thermal cycle in the mixed layer of the ocean.

A discontinuous time series of the thermal structure at sea stations A - F at St Helena Bay is indicated in Figure 37a and variations in the thermocline with time are shown in Figure 38. Winds recorded at the sea stations are illustrated in Figure 37b.

Although the bottom water at 30 m was cool (10.4-10.5°C) on the first day of the study, the upper mixed layer, which extended to a depth of approximately 20 m, was initially still relatively warm with temperatures of 12.1-12.5°C at 10 m and 12.6-13.6°C at the surface. Atmospheric forcing would have caused coastal upwelling off Cape Columbine the previous day, but this water had not yet advected into the bay itself by the 14th. The one sea station completed on the following day recorded a 10 m temperature of 11°C, indicating that colder upwelled water had advected into the bay. Deep south-easterly wind advection over the area would have forced strong upwelling off Cape Columbine and off the northern shores of St Helena Bay on the 15th.

The cool water present in the bay was well mixed by the strong south-easterly winds. Temperature profiles measured on the 15th and the morning of the 16th indicated an almost isothermal layer. Unfortunately no salinity measurements were able to be taken due to a malfunction in the salinity sensor. Salinity data would have provided useful information on the source of the water in St Helena.
Figure 37. (a) Sea temperature time series at the sea surface and at depths of 2 m, 10 m and 20 m. (b) Winds recorded at the corresponding times of the sea temperature measurements. (c) 10 m sea temperatures taken at 11h00 on 14-17, 20 & 21 October 1986.
Figure 38. Sea temperature profiles measured on 14-17, 20 & 21 October 1986. Notches on the x-axis represent 12°C for each profile made.
On the morning of the 16th conditions were calm and the water cold, with a surface temperature of only 11.2°C at 07h00 LST and temperatures of the order of 10.1°C below 12 m. Calm conditions throughout most of the day allowed for strong warming by solar heating within the surface layer. The calm conditions acted to restrict turbulent mixing during the day, therefore a shallow trapping depth of the surface heated layers was evident. The SST in the bay increased to a high of 13.8°C at 14h00 and the surface mixed layer extended to approximately 10 m. An increase in the wind speed after 15h00 resulted in a downward redistribution of the heat and cooling of the surface layer, even though the surface heat flux was still fairly strong and positive (~500 Wm⁻²).

Satellite-derived SST distributions were obtained in order to gain further insight into the temperature changes at the sea surface as a result of the strong solar heating. Images taken of the St Helena Bay area and surrounds for two suitable days, the 16th and 17th October 1986, are indicated in Figure 39. The images were taken at approximately 13h00 (LST) on both days. SST's recorded by the MC5 temperature bridge operated at the sea stations showed a good agreement with the satellite-derived SST's.

The image obtained from the afternoon of the 16th indicates a cold upwelling plume, with a central SST of less than 11°C, extending northwards from Cape Columbine. Deep southerly winds during the previous day forced cold water to be upwelled along the entire southern Benguela region. The only coastal area to record
Figure 39. Satellite-derived SST distributions over the southern Benguela for (a) 16/10/86, and (b) 17/10/86.
temperatures greater than 13°C was at the south and south-east shores of St Helena Bay itself. The heat gained by the sea surface during the day of the 16th was slowly distributed to a greater depth later during the day and convection deepened at night even though conditions were calm, mixing the water which was warmed during the day.

The cooling and mixing during the night erased the heat anomaly from the previous day so that by the morning of the 17th the SST was only 12°C and no distinct thermocline was evident (see Fig. 38). Below the surface the sea temperature was on the whole slightly greater than the previous day. Calm conditions, and therefore limited mixing, combined with a high positive surface heat flux lead to increased heating of the surface layer during the day. SST's rose to 14.2°C at noon and reached a high of 17.4°C by late afternoon. This increase in the SST may not, however, be the result of a high heat flux alone, but may be due, in part, to warm water advection in the bay. Salinity measurements would have assisted in determining the origin of this water. Solar heating stabilised the surface layer and stratification lead to a trapping of the heat at the surface within a layer approximately 2 m thick. The thermocline was at an average 12 m depth during the day. Temperatures in the mixed layer showed a gradual rise later in the day as the surface heat gain was slowly redistributed.

The satellite image taken from the afternoon of the 17th indicates a shrinkage of the cold plume north of Cape Columbine. This is a result of high solar heating and a decay of upwelling during the passage of the coastal low pressure systems. The core temperature
of the plume increased slightly to between 11° and 12°C. The water along the entire coastal region indicated in the image had warmed up sufficiently that the only areas with SST's colder than 13°C occurred north of the upwelling sites of the Cape Peninsula and Cape Columbine. Unfortunately the persistence of fog throughout the day in the region north of Lambert's Bay obscured any view of temperature changes in the northern area.

The satellite images and field results indicate that maximised solar heating on the 16th and 17th, along with calm conditions and a decay in the upwelling cycle, lead to widespread warming of the upwelled waters of the southern Benguela, with SST's on average increasing by 1-2°C. The greatest change appeared to take place in St Helena Bay itself where SST increases of 3-4°C occurred. The SST's derived from the satellite images only reflect conditions right at the surface and obviously depend on the time of observation and degree of wind mixing, therefore caution needs to be exerted in interpreting these results. The images are, however, still useful indicators of the degree of warming of the surface layers as a result of a high surface heat flux, even though the average conditions throughout the mixed layer are of greater significance.

No sea temperature profiles were made on the following two days and no further upwelling would have occurred off Cape Columbine as winds were predominantly light north-westerly. The final two days of the study were both characterised by relatively warmer water both at the surface (13.8-15.3°C on the 20th and 14.5-15.6°C on the 21st) and at a greater depth of 10m (11.5-11.8°C and 11.7-11.8°C respectively).
In order to summarise the average conditions over the synoptic scale, 10 m sea temperatures recorded at the same time (11h00) on each day of sea station occupation are indicated in Figure 37c. The 14th was characterised by warmish water not yet affected by upwelling. Strong winds on the afternoon of the 14th and the 15th caused deep upwelling off Cape Columbine and colder water was present in St Helena Bay on the 15th and 16th. A strong heat flux into the area and a cessation of upwelling from the 16th to the end of the field study allowed for rapid heating of the surface waters. Downward redistribution of some of this absorbed heat resulted in warmer water being present in the bay on the 17th, presumably the 18th and 19th, and on the 20th and 21st.

The high heat flux into the Benguela waters is capable of increasing the temperature of the upwelled water at a fairly rapid rate. The rate of temperature increase in the mixed layer due to solar heating depends on the surface energy budget, which in turn depends on the season and synoptic and diurnal cycles and the atmospheric variations exhibited therein. The time scale over which the upwelling events, and therefore major changes in the SST occurs, corresponds to the synoptic 5-6 day variability.

The net heat flux or rate of storage, \( H \), for a particular column of sea water is given by:

\[
H = \int C_p \frac{dT}{dt} dz \quad (7)
\]
where \( \rho' \) = density of sea water = \( 10^3 \) kg.m\(^{-3}\),
\( C_p \) = specific heat of sea water at constant pressure = 4180 J.kg\(^{-1}\).deg\(^{-1}\),
\( T \) = sea temperature,
\( t \) = time (secs), and
\( z \) = depth of mixed layer (m).

Assumptions applied to this formula are that vertical heat transport below the mixed layer is negligible and that the rate of temperature change in the mixed layer is approximately uniform.

The SST is known to increase in an offshore direction from the Benguela coast. This offshore gradient in the SST is fairly constant as far as the 18°C isotherm (Kamstra, 1985). As the cold upwelled water moves away from the most intense upwelling site at the coast it is able to absorb the available heat. Warmer water offshore or immediately surrounding an upwelling plume thus originates as previously upwelled water, which exists due to cumulative solar heating of the cool upwelled water as it moves away from the upwelling site. In fact the formation of upwelled water of sufficiently low density for it to move offshore as a surface layer is dependent on surface heating (Bowden, 1977). The heat flux which enters at the ocean surface is promptly distributed throughout the mixed layer and used to warm the ocean water. The previously upwelled water can be identified due to the fact that its salinity remains unchanged.

Use of equation (7) can be made in order to gain an insight into the temperature increase of the surface mixed layer as the water moves.
away from the upwelling centres along the west coast or during the
decay phase of the upwelling cycle. Thermohaline data from a
frontal zone cruise conducted along the west coast near Cape
Columbine suggests that the surface mixed layer, which effectively
gains the high heat flux and exhibits an overall increase in
temperature as it moves away from the upwelling site, extends to
some 10 m below the sea surface. In these coastal waters more than
half the short wave radiation is absorbed in the upper metre
and only an extremely negligible amount of energy is distributed
below a depth of 10 m. Little heat penetrates downward through
the thermocline at the upwelling sites because practically all of
the heat flux which enters at the sea surface is used to heat the
ascending water.

The heat input to the sea surface during the day exceeds the heat
loss at night, resulting in an overall positive heat flux into the
surface over the full diurnal period. Assuming an average total
heat flux of 227 W.m\(^{-2}\) (average for 8 days of field study) is
available to the ocean surface per day and that the heat gain is
mixed to a depth of 10 m, the change in temperature of the water
column as it moves away from the coast as a slab after one day has
elapsed would be:

\[
\frac{dT}{dT} = \frac{H \frac{dT}{dZ} + \int C_p \frac{dT}{dZ} \, dZ}{C_p \frac{dT}{dZ}} = \frac{227 \, \text{W.m}^{-2} \times 86400 \, \text{s}}{10^3 \, \text{kg.m}^{-3} \times 4.2 \times 10^3 \, \text{J.kg}^{-1}\text{deg.}^{-1} \times 10 \, \text{m}}
\]

\[
= 0.46^\circ\text{C per day}
\]

Although the rate of temperature increase might seem slow, it is
reasonable if one takes into account the fact that it is integrated over a 10 m column.

In summer during favourable phases of the synoptic cycle the global radiation recorded at D.F. Malan airport can reach 368 Wm$^{-2}$ over one day. The heat balance may be estimated through use of the extended results of the field study, i.e. that the sensible and latent heat fluxes on average approximately cancel each other, whilst back radiation averages some 70 Wm$^{-2}$. A heat balance of say 315 Wm$^{-2}$ during a favourable day in summer would then result in an averaged temperature increase over a 10 m column of 0.65°C per day. Because the heat input exhibits some variation over the synoptic scale, the increase in temperature of the surface mixed layer need not necessarily take place at a constant rate.

The averaged heat input over the extreme southern Benguela area during December amounts to an approximate 255 Wm$^{-2}$, which yields an integrated temperature increase of 0.52°C for the upper 10 m mixed layer. Further north near Alexander Bay a mid-summer heat balance of some 270 W.m$^{-2}$ would yield a slightly greater average integrated temperature increase of 0.55°C. Similar increases are to be expected along the South West African coast, but where further cloud cover and warmer water exists from the Walvis Bay region northwards in summer, the heat balance is expected to be less and temperature increases will not be quite as rapid. The surface heat flux, which is absorbed by the ocean, accumulates there until the surface temperature has risen sufficiently for enhanced evaporation, conduction and thermal radiation to occur (Woods, et al. 1984).
Information obtained by Olivieri (1983) from a Sea Fisheries monitoring programme during December 1979 is used to illustrate the reliability of the calculated temperature rise for the upper mixed layer as it moves offshore. A drogue released into cold upwelled water about 11 km off the Cape Peninsula coast was observed to follow the same body of water for 4 days. As the water progressed northwards away from the upwelling site gradual warming due to surface heating increased the temperature of the upper 10 m by 2°C over the 4 days in December. This strong heating acted to stabilize the water column by increasing the surface buoyancy in the upper mixed layer (Olivieri, 1983). The calculated temperature rise of an average 0.52°C per day during December for the southern Benguela therefore shows a good agreement with the average temperature increase of 0.5°C per day measured by Olivieri.

The drogue was observed to travel a distance of 28.2 km over 4 days, thus the horizontal SST indicates a rise of 2°C over 28.2 km or a gradient of 0.07°C.km⁻¹. The rate of current movement is, however, not fixed therefore variations in the horizontal gradient may occur. If an average surface current speed for the Benguela of 17 cm.s⁻¹ is assumed (Shannon, 1985), then a piece of seawater would be observed to move north-westwards at an averaged rate of 15 km.day⁻¹. This would then give rise to a horizontal SST gradient of 0.034°C.km⁻¹ over the southern Benguela in December.

The estimated rise in SST also shows a good agreement with the temperature increases derived by Stevenson, et al. (1981) for a surface mixed layer off the upwelling area of Peru. It was calculated that an influx of solar energy of 465 ly.day⁻¹ during
March (equinox = time of maximum insolation), which is equivalent to 255.5 W m\(^{-2}\), was sufficient to raise the temperature of a 6 m deep surface mixed layer by 0.81°C over 1 day. A similar and even better agreement is to be found from values derived by Bowden (1977). A net gain of heat by the surface of 506 cal. cm\(^{-2}\) day\(^{-1}\) (245 W m\(^{-2}\)) near the coast was calculated to result in a temperature increase of 0.51°C in the upper 10 m layer. Integrated over 20 m the resultant rise in temperature was calculated to be 0.25°C.

During summer, when maximum upwelling takes place over most of the southern and central Benguela, the high heat input to the surface water is capable of raising the temperature of the upwelled water as it moves away from its site of origin. Owing to fact that this increase in temperature of the surface mixed layer is fairly rapid, it cannot be used as a means of distinguishing upwelled water from offshore water for longer than a few days. Fortunately the salinity of the upwelled water remains high so that it can be used to trace the path of the upwelled water. The rapid warming of the surface water as it moves offshore gives rise to fairly sharp offshore SST gradients during the summer months. During winter less upwelling takes place in the southern Benguela and the input heat is diminished, therefore the surface waters are not heated as intensely and offshore SST gradients are weaker (see Fig. 7). Maximum upwelling takes place in the northern Benguela during winter, but because less insolational heat is available, the surface waters are not heated as intensely and offshore SST gradients are not as strong as the southern area experiences during its maximum upwelling season.
The cold, newly upwelled water is usually uniformly mixed down the water column and has a low salinity ($\pm 34.7\%$), relatively high density ($\sigma_t = 26.63$), high nutrient content and low 1% light level (i.e. the water is relatively clear). The water has a high biological potential, which requires warmth and light to generate into biological production. Once active upwelling ceases and/or as the upwelled water moves away from the coast, solar heating of the euphotic layer produces a thermocline and the absorbed heat and light is utilised for phytoplankton growth. The phytoplankton stock of the upwelled water observed by Olivieri (1983), in fact, showed an increase of as much as 18.8 mg.1$^{-1}$ of chlorophyll$_a$ after 4 days.

The increase in phytoplankton concentration during the warming cycle results in an increase in oxygen concentration above the thermocline, but is accompanied by a respective decrease in nutrient concentration. The phytoplankton acts to attenuate incident radiation, therefore the 1% light level is observed to decrease in depth and solar heating is thereby confined to the surface layers. The unusually high heat flux into the surface waters of the Benguela provides the heat and, if the input radiation is high during this phase of the synoptic cycle, also provides the light necessary to realise the biological potential of the upwelled waters and convert it into biological production. The high heat input into the surface waters of the Benguela is therefore of vital importance to the entire Benguela system.

In summary, the synoptic cycle of 5-6 days variability and the corresponding progressive upwelling cycle are important in maintaining the Benguela ecosystem as a whole. The cold, nutrient-
rich water, which is upwelled to the surface as a result of atmospheric forcing, is sun-warmed as it progresses offshore during the decay cycle, resulting in increased biological production. This short-term variability over a few days is of more significance to the Benguela system than the longer term seasonal variability.
Although the results of the field programme could be extended in a longshore direction, extensions over the climatic scale could prove to be more difficult as measurements were only taken over a limited period. The field results could, however, be used to typify springtime conditions over the southern Benguela during favourable phases of the synoptic cycle. Fortunately additional long-term global radiation data from Cape Town (D.F. Malan Airport) and Alexander Bay was available to provide more realistic long-term trends. Comparisons of the field results can, however, be made with the October data recorded by Hoflich (1984) over the Benguela area. Extensions are then possible over the climatic time scale. Even then, one must still bear in mind that only eight days data from a month will lead to some contribution biasing.

Hoflich’s climatic table for a $5^\circ \times 10^\circ$ area (Area C), which includes the southern Benguela is illustrated in Table 3. Data from a $5^\circ \times 5^\circ$ area (Area L), encompassing part of the central Benguela, is indicated in Table 4 and results from a $5^\circ \times 10^\circ$ area (Area G), including the extreme northern Benguela, are tabulated in Table 5. The actual areas sampled are indicated in Figure 40. Reference will be made to the eight-day averaged results obtained from the St Helena Bay field programme and tabulated in Table 1.

The high values of incident solar radiation recorded during the field programme (319 W.m$^{-2}$) are seen to far exceed the October average of 201 W.m$^{-2}$ recorded by Hoflich over the southern area.
Figure 40. Climatic areas sampled by Hoflich (1984).
<table>
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<th>Mean radiation (W.m(^{-2}))</th>
<th>Mean net radiation (W.m(^{-2}))</th>
<th>Mean latent sensib.</th>
<th>Mean Solar terrest.</th>
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### TABLE 5. CLIMATIC TABLE FOR AREA G

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<th>Mean heat flux (W.m⁻²)</th>
<th>Mean radiation (W.m⁻²)</th>
<th>Mean net radiation (W.m⁻²)</th>
<th>Mean heat bal. (W.m⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>latent sensib. solar terrest.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jan</td>
<td>78</td>
<td>-2</td>
<td>192</td>
<td>37</td>
</tr>
<tr>
<td>Feb</td>
<td>99</td>
<td>2</td>
<td>181</td>
<td>42</td>
</tr>
<tr>
<td>Mar</td>
<td>105</td>
<td>8</td>
<td>156</td>
<td>39</td>
</tr>
<tr>
<td>Apr</td>
<td>111</td>
<td>8</td>
<td>155</td>
<td>47</td>
</tr>
<tr>
<td>May</td>
<td>92</td>
<td>5</td>
<td>143</td>
<td>50</td>
</tr>
<tr>
<td>Jun</td>
<td>112</td>
<td>9</td>
<td>125</td>
<td>51</td>
</tr>
<tr>
<td>Jul</td>
<td>82</td>
<td>5</td>
<td>117</td>
<td>45</td>
</tr>
<tr>
<td>Aug</td>
<td>75</td>
<td>6</td>
<td>117</td>
<td>37</td>
</tr>
<tr>
<td>Sep</td>
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<td>3</td>
<td>144</td>
<td>37</td>
</tr>
<tr>
<td>Oct</td>
<td>64</td>
<td>-3</td>
<td>171</td>
<td>31</td>
</tr>
<tr>
<td>Nov</td>
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<td>30</td>
</tr>
<tr>
<td>Dec</td>
<td>77</td>
<td>-1</td>
<td>199</td>
<td>38</td>
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<tr>
<td>Annual</td>
<td>85</td>
<td>3</td>
<td>157</td>
<td>40</td>
</tr>
</tbody>
</table>
This discrepancy could, however, be ascribed to the fact that the southern Benguela area (including St Helena Bay), only represents a small portion of the larger space scale employed by Hoflich. The October average of 253.6 W.m\(^{-2}\) recorded at D.F. Malan is also noted to exceed the values recorded by Hoflich, as does the Alexander Bay October average of 287.8 W.m\(^{-2}\) exceed Hoflich's 208 W.m\(^{-2}\) for the area encompassing the central Benguela. The lower values recorded by Hoflich can probably be ascribed due to the fact that the sampling area extends further offshore to an area no longer influenced by the Benguela, but more representative of the oceanic area. An increase in cloud cover in an offshore direction would terminate a greater amount of solar radiation offshore, thereby causing the area averages to be lower than the values recorded at the coast.

The annual trend of global radiation recorded by Hoflich (1984) is exceeded at all times by the global radiation recorded at the coastal sites of D.F. Malan and Alexander Bay (see Fig. 35). The incident radiation received decreases to the north, where less annual variation and increased cloud cover is evident. The highest value of insolation calculated by Hoflich was quoted as 280 W.m\(^{-2}\) received at the South West African coast. This value is, however, still less than the December average of 342.7 W.m\(^{-2}\) measured by the Weather Bureau at Alexander Bay. It would thus appear that the incident solar radiation tends to be underestimated by Hoflich.

The average back radiation value of 71 W.m\(^{-2}\) recorded at St Helena Bay compares favourably with the October average of 68 W.m\(^{-2}\) calculated by Hoflich for Area O. Annual variations are not
pronounced and the back radiation is seen to decrease to the north, probably due to the effect of increased cloud cover.

The calculated turbulent heat fluxes from the field programme were, however, observed to differ from those derived by Hoflich (1984) for the month of October. Hoflich calculated the sensible and latent heat fluxes over the southern area to both be losses of 1 Wm\(^{-2}\) and 85 Wm\(^{-2}\) respectively. The fact that the area sampled by Hoflich included some offshore oceanic water obviously influenced the average values. A constant exchange coefficient of 1.3 x 10\(^{-3}\) could be the factor responsible for larger latent heat losses than what might be expected. The sensible heat flux measured by Hoflich was, in fact, only calculated to be a gain during the summer months. However, the isolines of sensible heat flux in Figure 12 appear to indicate a sensible heat gain by the ocean throughout the year. Hoflich calculated that both turbulent fluxes combined amounted to a loss of 20-25 Wm\(^{-2}\) off the west coast, with little annual variation evident.

The high heat balance of 227 Wm\(^{-2}\) calculated during the eight-day field study far exceeds the October average of 47 Wm\(^{-2}\) calculated by Hoflich (1984) for the southern area. Even the highest quoted value of 215 Wm\(^{-2}\) during January along the South West African coast is still 40 Wm\(^{-2}\) less than the summer average of 255 Wm\(^{-2}\) postulated from the D.F. Malan data (southern Benguela) and is 55 Wm\(^{-2}\) less than the projected Alexander Bay summer average of 270 Wm\(^{-2}\). Reasons for the underestimation of the heat balance by Hoflich lie mainly in the fact that offshore oceanic areas were included in the sample areas, while incident radiation appears to be
underestimated and the latent heat flux is slightly exaggerated.

A comparison of the heat balance (Fig. 10) and net radiation (Fig. 41) distribution maps provided by Hoflich (1984) supports the theory that the heat balance may be estimated by the net radiation. For example the high heat balance of 215 W.m\(^{-2}\) quoted along the South West African coast exhibits little variation from the average net radiation of 230 W.m\(^{-2}\) calculated by Hoflich (1984) for summer. The map distributions and climatic tables also serve to indicate that the variation of solar radiation according to season is the main cause of the seasonal variation in the heat balance.
Figure 41. Net radiation at the sea surface (W.m$^{-2}$) during (a) January, and (b) July (after Hoflich, 1984).
7. SUMMARY AND CONCLUSIONS

The surface heat budget of the Benguela upwelling system on the west coast of southern Africa was examined. An important feature of this coast is the occurrence of relatively low sea temperatures associated with the often intense upwelling. The lowering of the SST as a result of upwelling fluctuates in space, due to the coastal configuration, and in time due to the fluctuations in the wind fields.

Results gained from the field study were used to characterise the various terms contributing to a high heat balance over the Benguela upwelling area. Preliminary findings indicated that variations in the heat balance over a small distance of approximately 10 n.mi offshore were minimal. The illustrations of the components of the energy balance revealed that variations in solar radiation dominated variations in the energy balance. Because the upwelled water is cold, the sensible heat flux is directed downward to the sea surface and the latent heat loss from the surface is reduced. These turbulent heat fluxes are therefore observed to be approximate converses of each other and the stable stratification of the air layers above the cold water acts to limit the magnitude of the fluxes. An increase in wind speed was observed to increase both turbulent heat fluxes, but the sum of the fluxes was, nevertheless, still calculated to be nearly constant, averaging zero. Only a slight variation was exhibited by the back radiation.

Owing to the fact that the turbulent heat fluxes approximated zero, the net radiation term provided a good estimation of the total heat
balance. A similar trend was also evident from the long-term heat budget data calculated by Hoflich (1984). A strong diurnal cycle in the energy budget and in the winds was found to exist over the coastal upwelling area.

Modelling of the heat exchange between the nearshore and offshore regions under idealised conditions revealed that most energy exchange between the air and sea takes place in the nearshore region. The exchange of heat and mixing in the lower layers of the atmosphere were found to be governed largely by the strength and height of the atmospheric inversion and the wind speed. The actual gain of heat by the sea surface was calculated to be greater at the cold nearshore region than, for example, 100 km further offshore, where warmer water is present.

Variations in the upwelling and heat balance are controlled by fluctuations in meteorological variables. The main characteristics of the local energy budget are climatically determined, but synoptic events have been found to produce significant variations on this general theme. Although synoptic variation in the heat balance during the period of the field programme was observed to be negligible, additional global radiation data recorded over a longer time span revealed that the variation in environmental conditions, and therefore the heat balance, of the Benguela upwelling system over the synoptic time scale of a few days was, in fact, more significant than variations over the longer seasonal time scale.

Because longshore variations in the ocean and atmosphere are negligible in comparison with offshore variations, the results from
the localised field study could be extended to typify the situation along the entire Benguela coast. The heat balance of the entire west coast area is found to be dominated by the input solar radiation, which in turn depends on the astronomical cycle and atmospheric variables, particularly cloud cover. The high heat balance over the Benguela area is due to the facts that clear to fair skies are most frequently found off the west coast of southern Africa, and that the air is particularly dry above the subsidence inversion.

Seasonal variations in the heat balance are obviously dominated by the solar radiation, and are therefore dependent on the astronomical cycle. Winter insolation was, in fact, found to be approximately half that in summer. These variations are more pronounced in the southern area which is also influenced by the passage of cold fronts with a 3-6 day variability (Preston-Whyte & Tyson, 1973). It would be expected that the northern Benguela, owing to it position further north, would receive more insolation than the southern Benguela, but this meridional difference is damped out by the increased fog and cloud frequency in the north. Variations in average solar radiation in a longshore direction are therefore not pronounced.

The large amount of surplus heat available to the sea surface of the Benguela is used to increase the temperature of the cold upwelled waters. Most of the heating is caused by the absorption of solar radiation. The diurnal solar cycle causes strong heating in the surface layer of approximately 1 m. The absorbed heat is gradually distributed vertically downward into the cooler water below by wind-
generated turbulence and convection. This heat exchange has a substantial effect on the periodically upwelled waters, and is capable of increasing the temperature of the upper 10 m mixed layer by as much as 0.65°C per day in mid summer. The input of heat is vital to sustain the biological production in the upwelled water as it moves offshore.

Since the physical processes which operate in all upwelling areas are similar, the results gained from this study can generally be applied to the other upwelling sites of the world. Obviously, applications will differ according to location, latitude and synoptic influences, but the general characterisation of the heat budget terms over upwelling areas is similar.

It is known that equatorward winds along west coasts are the driving force for upwelling of cold subsurface waters. When cold water arrives at the sea surface the heat flux increases because the decreased water temperature acts to stabilise the overlying air layers and increases sensible heat conduction to the ocean surface, in the process reducing the evaporative heat loss. Both sensible and evaporative fluxes in upwelling regions are known to be generally less than 10% of the radiation heat gain (Bunker, 1976; Budyko, 1963). The fact that these fluxes operate in opposite directions, implies that the assumption that the net radiation approximates the total heat budget over cold water, may be applied to other upwelling areas of the world.

The effect of the upwelled water is to decrease the air temperature at the coastal regions and strengthen the inversion layer. These
low temperatures suppress convective cloud formation and therefore allow for the receipt of maximum solar radiation at the ocean surface. The presence of fog and low stratus clouds over the cold water can, however, reduce the insolation. The largest net energy surpluses have been found to occur at the upwelling regions of the world (Hsiung, 1986) where maximised heating of the cold waters can take place. The cold waters, therefore, become warmed mainly by insolation and ultimately become regions of strong heat export.

Seasonal variations in the heat budget and in its components are obviously expected over both upwelling and non-upwelling areas, these variations being greater at the higher latitude locations. Similarly, synoptic scale variations in the surface energy budget terms would be anticipated at all the upwelling sites, even though meteorological conditions may differ according to location. Over the 3-4 day time scale daily variations in the net energy flux at the surface, the heat content of the upper 15 m of the ocean and the ocean temperature were observed by Reed and Lewis (1980) over the eastern Tropical Atlantic Ocean. These variations are due to the passage of westward-travelling synoptic-scale disturbances in the atmosphere. Similar variations are to be anticipated off Peru, while off Oregon 3-7 day synoptic fluctuations occur (Huyer, 1978). The Benguela system is typically governed by synoptic disturbances over the 3-6 day period (Preston-Whyte & Tyson, 1973). It is therefore apparent that the environmental controls vary over the synoptic time period of a few days, subsequently causing synoptic variations in the heat budget at all the upwelling sites.

Over the diurnal period land-sea breezes have been found to occur at
all upwelling sites, therefore variations in the turbulent heat fluxes throughout the day are expected to be similar to those experienced over the Benguela area, with minimum values at night and a maximum during the afternoon when the sea-breeze cycle is at its most intense. Daily insolation values are expected to be high, contributing to a diurnal heating-cooling cycle in the surface mixed layer of the upwelled waters.

Heating of the surface waters is more intense over the coastal upwelling areas than over other oceanic areas, where a greater proportion of the available heat is used for the evaporation process. The latter areas ultimately receive their heat from the heat eventually distributed from the upwelling areas, thus maintaining the oceanic heat balance.

Findings from this study indicate that measurements taken at the coast can be used to characterise the heat budget of the nearshore areas along the west coast. It is apparent that further direct measurement of solar radiation and the meteorological variables over the west coast would lead to a more accurate assessment of the heat balance over the Benguela area. Future research should concentrate on the establishment of additional solarimeters at meteorological stations along the west coast of southern Africa in order to categorize the input solar radiation more accurately. In addition, the consistent measurement of solar radiation from research cruises in the Benguela should be investigated. This, together with the on-board measurement of meteorological variables, would provide a more accurate indication of the on-going heat transfer processes operating at the air-sea interface of the Benguela system and could be used to constrain the expected features of the model.
be used to confirm the expected features of the model.
## APPENDIX 1

### Moisture Chart for Psychrometer

<table>
<thead>
<tr>
<th>Difference between Dry and Wet Bulb, t - t' (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>t</td>
</tr>
<tr>
<td>-----</td>
</tr>
<tr>
<td>C</td>
</tr>
<tr>
<td>0.0</td>
</tr>
<tr>
<td>0.8</td>
</tr>
<tr>
<td>1.6</td>
</tr>
<tr>
<td>2.4</td>
</tr>
<tr>
<td>3.2</td>
</tr>
<tr>
<td>4.0</td>
</tr>
<tr>
<td>4.8</td>
</tr>
<tr>
<td>5.6</td>
</tr>
</tbody>
</table>

### OTA KEIKI

SEISAKUSHO.
### APPENDIX 2

#### SATURATION VAPOR PRESSURE OVER WATER

**Metric units**

<table>
<thead>
<tr>
<th>Temperature (^\circ)C</th>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
<td>mb</td>
</tr>
<tr>
<td>0</td>
<td>6.078</td>
<td>1.513</td>
<td>1.071</td>
<td>2.142</td>
<td>2.385</td>
<td>3.333</td>
<td>3.793</td>
<td>4.425</td>
<td>5.472</td>
<td>6.510</td>
</tr>
<tr>
<td>9</td>
<td>11.246</td>
<td>2.428</td>
<td>1.750</td>
<td>1.895</td>
<td>2.128</td>
<td>2.845</td>
<td>3.415</td>
<td>4.256</td>
<td>5.441</td>
<td>6.502</td>
</tr>
</tbody>
</table>

#### APPENDIX 3

The table above provides the saturation vapor pressure (in millibars) over water at different temperatures (in °C). The values are rounded to two decimal places. Each row represents a different temperature, with the pressure values increasing as the temperature increases.
# DATA SHEET - BEACH STATION

<table>
<thead>
<tr>
<th>DATE:</th>
<th>LOCATION:</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**WEATHER CONDITIONS & NOTES**

<table>
<thead>
<tr>
<th>Time</th>
<th>SOLAR RADIATION</th>
<th>TEMPERATURES</th>
<th>WIND</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Total</td>
<td>Dry B Wet B SST</td>
<td>10s Ave Speeds</td>
</tr>
<tr>
<td></td>
<td>Scattered</td>
<td></td>
<td></td>
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</tbody>
</table>
### APPENDIX 4

#### DATA SHEET

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</tr>
</thead>
<tbody>
<tr>
<td>TIME: SAST LOCAL</td>
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<tr>
<td>POSITION: LAT LONG</td>
<td></td>
</tr>
<tr>
<td>WATER DEPTH ATM. PRESS</td>
<td></td>
</tr>
<tr>
<td>SOLAR RADIATION: WIND SPEED (10s Ave):</td>
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<tr>
<td>TOT. INCIDENT SCATTERED</td>
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<tr>
<td>NET 1 MIN AVE</td>
<td></td>
</tr>
<tr>
<td>CLOUD COVER DIRECTION</td>
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</tr>
<tr>
<td>AIR TEMPERATURE: SURFACE CURRENT:</td>
<td></td>
</tr>
<tr>
<td>DRY BULB SPEED</td>
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</tr>
<tr>
<td>WET BULB DIRECTION</td>
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<tr>
<td>SECCHI DEPTH SEA STATE</td>
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</tbody>
</table>

#### WEATHER CONDITIONS & NOTES

#### PROFILES:

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<th>DEPTH</th>
<th>TEMP</th>
<th>SALINITY</th>
<th>O₂</th>
</tr>
</thead>
<tbody>
<tr>
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<td></td>
</tr>
<tr>
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<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td></td>
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<td></td>
</tr>
<tr>
<td>8</td>
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<td></td>
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<tr>
<td>10</td>
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</tr>
<tr>
<td>15</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<tr>
<td>30m</td>
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</table>

![Depth vs Temp](image1.png)

![Depth vs Salinity](image2.png)
Paired t-test to test whether there is a significant difference in incident solar radiation received at the beach and at positions ± 10 n.mile offshore.

Null hypothesis: There is no difference in the solar radiation values, i.e. \( \mu = 0 \).

Alternative hypothesis: There is a difference in the solar radiation values, i.e. \( \mu \neq 0 \).

Formula used:

\[
    t = \frac{\bar{X} - \mu}{s / \sqrt{n}}
\]


\( n = 29 \)

Degrees of freedom = \( n - 1 = 28 \)

\( \bar{X} = 6.55 \)

\( s = 25.74 \)

Calculation:

\[
    t = \frac{6.55 - 0}{25.74} = 1.37
\]

Significance level: 99% \( t_{(0.01)} = 2.76 \)

Conclusion: \( t_{calc} = 1.37 < 2.76 = t_{crit} \). accept null hypothesis, i.e. there is no significant difference (99% level) solar radiation recorded at the two different sites.
### APPENDIX 6

**ALTITUDE OF SUN IN DEGREES**

Latitude: 32° 40'S
Longitude: 18° 00'S

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<th>16</th>
<th>17</th>
<th>18</th>
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<td>5.4</td>
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<td>29.9</td>
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<td>30.2</td>
<td>30.4</td>
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<tr>
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<td>5.6</td>
<td>5.8</td>
<td>6.0</td>
<td>6.1</td>
</tr>
</tbody>
</table>
Paired t-test to test whether there is a significant difference in the heat balance between the beach and a position 10 n.mi offshore.

Null hypothesis: There is no difference in the heat balance at the two stations i.e. $\mu = 0$.

Alternative hypothesis: There is a significant difference in the heat balance at the two stations i.e. $\mu \neq 0$

Formula used:

$$ t = \frac{\bar{X} - \mu}{s / \sqrt{n}} $$

Difference in values (Beach-sea stations): 58, 10, -12, -9, -15, 27, -57, -70, -49, -41, -21, 59, 65, 40, 26, 35, 26, 39, 94, 109, 113, 49, -29, -19, 51, 52, 10, -26, -89.

$n = 29$

Degrees of freedom = 28

$\bar{X} = 14.72$

$s = 51.92$

Calculation:

$$ t = \frac{14.72 - 0}{51.92 / \sqrt{28}} = 1.53 $$

Significance level: 99% $t(0.01) = 2.76$

Conclusion: $t_{calc} = 1.53 < 2.76 = t_{crit}$. 'accept null hypothesis, i.e. there is no significant difference (99% level) in the heat balance at the two different sites.
Paired t-test to test whether any significant difference exists between the net radiation and the total heat balance at St Helena Bay.

Null hypothesis: There is no significant difference between the net radiation and total heat balance at both beach and sea stations in St Helena Bay, i.e. \( \mu = 0 \).

Alternative hypothesis: There is a significant difference between the net radiation and total heat balance recorded at St Helena Bay, i.e. \( \mu \neq 0 \).

Formula used: 
\[
\frac{\bar{x} - \mu}{s / \sqrt{n}}
\]

Difference in values \((R_n - Q_t)\): 7, 2, -13, -4, 2, -10, -13, 11, 11, 16, -18, 40, 2, -29, 5, -7, -24, -32, -31, -49, 1, 0, -3, -11, -4, -31, -30, -32, -10, -26, -82, 68, 28, 37, 74, 87, 130, 20, 75, 35, 100, -1, 0, -4, -13, -29, -13, -12, -14, 0, 6, -2, -8, -19, -19, -7, -101, -136, -77, -60, -5, 3, 9, 27, 3, 70, -11, 0, -7 -3, -13, -8, -25, 0, -3, 0, 6, -15, -37, -58, -1, -8, -14, -12, -11, 7, 40, -3, 18.

\( n = 89 \)

Degrees of freedom = \( n - 1 = 88 \)

\( \bar{x} = 3.17 \)
\( s = 38.18 \)

Calculation: 
\[
t = \frac{3.17 - 0}{38.18 / \sqrt{89}} = 0.78
\]

Significance level: 99\% \( t_{0.01} = 2.632 \)

Conclusion: \( t_{calc} = 0.78 < 2.632 = t_{crit} \). accept null hypothesis, i.e. there is no significant difference (99\% level) between the values of \( R_n \) and \( Q_t \) at St Helena Bay.
\[ \begin{align*}
\Sigma x &= 55278 \\
\bar{x} &= 581.87 \\
\Sigma x^2 &= 39920392 \\
\Sigma xy &= 41037481 \\
C_{xx} &= \frac{\Sigma x^2 - (\Sigma x)^2}{n} \\
&= \frac{39920392 - (55278)^2}{95} \\
&= 7755578.48 \\
C_{xy} &= \frac{\Sigma xy - \Sigma x \Sigma y}{n} \\
&= \frac{41037481 - 55278 \times 56657}{95} \\
&= 8070263.67 \\
C_{yy} &= \frac{\Sigma y^2 - (\Sigma y)^2}{n} \\
&= \frac{42322089 - (56657)^2}{95} \\
&= 8532450.59 \\
b &= \frac{C_{xy}}{C_{xx}} = 1.04 \\
a &= \bar{y} - b\bar{x} = 596.39 - 581.87 (1.04) = -9.09
\end{align*} \]

The linear regression line is represented by:
\[ y = -9.09 + 1.04x \]

The correlation coefficient is:
\[ r = \frac{C_{xy}}{\sqrt{C_{xx} C_{yy}}} = 0.992 \]
To test whether the latent heat flux recorded at the beach is greater than that recorded at the sea stations in St Helena Bay.

Null hypothesis: There is no difference in the latent heat flux at the two different sites, i.e. \( \mu = 0 \).

The model predicts that values at the beach should be greater than those offshore, therefore:

Alternatively hypothesis: The latent heat flux recorded at the beach station exceeds that recorded at the sea stations i.e. \( \mu > 0 \).

Formula used: \[ t = \frac{\bar{X} - \mu}{\frac{s}{\sqrt{n}}} \]

Difference in values (beach-sea): -8.7, -6.8, -5.5, 12.4, 45.6, 10.2, 18.6, 36.1, 17.3, 31.8, 20.0, 14.8, 8.4, -0.7, 0.0, 0.0, 0.0, -15.5, 0.0, -3.2, -16.8, -16.8, -8.4, 1.1, -0.9, -7.3, -5.7, 14.9, 28.6, 58.9.

\( n = 29 \)

Degrees of freedom = \( n - 1 = 28 \)

\( \bar{X} = 8.25 \)

\( s = 18.29 \)

Calculation: \[ t = \frac{8.25 - 0}{18.39} = 2.43 \]

Significance level: 99% \( t_{(0.01)} = 2.467 \)

95% \( t_{(0.05)} = 1.701 \)

Conclusion: \( t_{\text{calc}} = 2.43 > 1.701 = t_{(0.05)} \). reject null hypothesis and accept alternative at 95% significance level,

but \( t_{\text{calc}} = 2.43 < 2.467 = t_{(0.01)} \). accept null hypothesis at 99% level,

i.e. the latent heat flux at the beach (coast) is > \( \pm 10 \text{ n.m.i offshore} \) at the 95% level of significance.
To test whether the sensible heat flux recorded at the beach is greater than that recorded at the sea stations in St Helena Bay.

Null hypothesis: There is no difference in the sensible heat flux recorded at the two different sites, i.e. \( \mu = 0 \).

The model predicts that values at the beach should be greater than those recorded further offshore, therefore:

Alternative hypothesis: The sensible heat flux recorded at the beach station exceeds that recorded at the sea stations, i.e. \( \mu > 0 \).

Formula used: \( t = \frac{\bar{X} - \mu}{s / \sqrt{n}} \)

Difference in values (beach-sea): 3.6, 5.1, -5.3, -1.7, 0.3, 10.3, 11.6, 30.9, -13.6, -9.2, -14.5, 37.5, 43.3, -0.3, 8.2, 15.1, 5.1, -38.2, 88.2, 120.8, 56.4, 51.6, 2.2, 3.9, 6.9, 0.3, 10.4, -15.8, -39.2.

\( n = 29 \)

Degrees of freedom = \( n - 1 = 28 \)

\( \bar{X} = 12.85 \)

\( s = 33.92 \)

Calculation:

\[ t = \frac{12.85 - 0}{33.92 / \sqrt{29}} = 2.04 \]

Significance level: 99% \( t_{(0.01)} = 2.467 \)

95% \( t_{(0.05)} = 1.701 \)

Conclusion: \( t_{calc} = 2.04 > 1.701 = t_{(0.05)} \) :: reject null hypothesis and accept alternative at 95% significance level, but \( t_{calc} = 2.04 < 2.467 = t_{(0.01)} \) :: accept null hypothesis at 99% level, i.e. the sensible heat flux at the beach (coast) is > 10 n. mi offshore at the 95% level of significance.
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