THE EFFECT OF LAND-SEA TEMPERATURE CONTRAST ON SHORT-TERM NUMERICAL FORECASTS

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ABSTRACT

A simple scheme for the calculation of land surface temperature by means of the solution of the heat balance equation is outlined. Features include diurnal variation of solar insolation and a stability dependent drag coefficient. Experiments are described in which the scheme is used to gauge the relevance of land-sea surface temperature contrasts to short-term numerical predictions in the Australian region. In the experiments land surface temperature is specified using the scheme while sea surface temperatures are set to climatological values. Prognoses for three summer situations are compared with prognoses performed with the scheme omitted. In each case the prognosis performed by the model including the surface heat balance gives improvements in the mean sea level pressure pattern which would be of direct consequence for weather forecasts in the southern Australian States.

INTRODUCTION

Due mainly to operationally imposed running-time restrictions, numerical models used to date for routine forecasting purposes in the southern hemisphere, and in particular in the Australian region, have included no radiation schemes. This is justified, as far as radiative processes in the free atmosphere are concerned, by the general agreement in the scientific literature that these processes have a minimal effect on short-term predictions (see for example GARP Joint Organising Committee (1970)). Radiation, however, is one component of the heat balance of the atmosphere's underlying surface. Its absence from the models has consequently been accompanied by an unrealistically small and time invariant land-sea temperature contrast. This contrast exerts a pronounced synoptic effect in the Australian summer period as is demonstrated by the persistent appearance on daily synoptic charts of thermally produced lows over the Australian continent.

This study describes experiments conducted with a very simple surface heat balance scheme. The scheme was designed to test the hypothesis that positive short-term effect can be gained in large-scale hemispheric numerical forecasting models from the use of empirical radiation relations in combination with bulk aerodynamic eddy flux parameterisations for the specification of surface temperature. Surface temperature over the land is the resultant of the balance of the energy fluxes at each grid point, and is controlled by a diurnally varying downward flux of solar insolation. Sea surface temperatures are set to climatological values and the temperatures over the antarctic land mass are set to the values of temperature at the lowest model level in the starting analysis.
Such an approach has the advantage of economy in computer storage space and calculation time. The joint objectives of this paper are to show that the approach can be used successfully in a forecasting model and to test the sensitivity of model forecasts to its implementation.

DESCRIPTION OF THE MODEL AND THE RADIATION SCHEME

The heat balance scheme is here described in the context of the Australian Numerical Meteorology Research Centre's six level restructured form of the GFDL primitive equation model. The model is run on a polar stereographic projection of the southern hemisphere with 30 grid points from pole to equator and a smoothed topography. A schematic representation of the distribution of vertical levels is shown in Fig 1 where the vertical coordinate is the normalised pressure \( \sigma = p/p_\ast \). \( p \) is pressure, and \( p_\ast \) is the local surface pressure.

<table>
<thead>
<tr>
<th>( \sigma )</th>
<th>VALUE</th>
<th>LEVEL</th>
</tr>
</thead>
<tbody>
<tr>
<td>.000</td>
<td>—</td>
<td>( K = \frac{1}{2} )</td>
</tr>
<tr>
<td>.020</td>
<td>—</td>
<td>( K = 1 )</td>
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<tr>
<td>.074</td>
<td>—</td>
<td>( K = 2 )</td>
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<tr>
<td>.156</td>
<td>—</td>
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<tr>
<td>.259</td>
<td>—</td>
<td>( K = 4 )</td>
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<tr>
<td>.376</td>
<td>—</td>
<td>( K = 5 )</td>
</tr>
<tr>
<td>.500</td>
<td>—</td>
<td>( K = 6 )</td>
</tr>
<tr>
<td>.624</td>
<td>—</td>
<td>( K = 6\frac{1}{2} ) (i.e. the surface)</td>
</tr>
<tr>
<td>.741</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>.844</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>.926</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>.980</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>1.000</td>
<td>—</td>
<td></td>
</tr>
</tbody>
</table>

Fig 1 Distribution of levels in the model.

The basic form of the model used in this study is that described by Gauntlett and Hincksman (1971). A fundamental difference between this model and the 1971 version is that in the current form vertical diffusion is extended into the higher levels of the troposphere per medium of the partially implicit algorithm of Smagorinsky et al. (1965). This eddy exchange in the free atmosphere was introduced into the model by Kininmonth (1974) as part of an overall viscosity and convection parameterisation scheme.
Kininmonth's convective parameterisation represents an extension of the 'integrated moisture convergence' ideas of Kuo (1965) and is a modification of the scheme described by Barker and Kininmonth (1973). The modifications have been designed to minimise numerical noise by treating the convection as complementary to the local dynamics. Other added features are the incorporation of momentum transports and the inclusion of internal cloud downdrafts to modify the boundary layer.

Surface temperature directly influences the model dynamics through the fluxes of latent and sensible heat. Surface eddy fluxes of heat and momentum follow the bulk aerodynamic formulae given by Manabe et al. (1965).

At each grid point the level of the cloud base is calculated according to a correlation between the mean mass-weighted 1000 to 500 mb relative humidity and the type of cloud, as shown in Table 1. If there is convection at a grid point (the criterion for which is that there is both synoptic scale boundary layer convergence and a lapse rate unstable for pseudo-adiabatic ascent from the boundary layer), the convective cloud is assumed to occupy only a small fraction of the grid area, so that radiative processes are allowed to proceed as if for clear skies.

Table 1  
Relation between the level of cloud base and the model generated  
1000 to 500 mb mass-weighted average relative humidity. The  
figures shown are based on regression relations determined from  
satellite cloud photographs (see, for example, Thompson and West  
(1967)).

<table>
<thead>
<tr>
<th>Relative Humidity</th>
<th>Model Level Considered to Represent the Cloud Base</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.618 ≤ RH &lt; 0.475</td>
<td>Level 5 (low cloud)</td>
</tr>
<tr>
<td>0.475 ≤ RH &lt; 0.325</td>
<td>Level 4 (middle cloud)</td>
</tr>
<tr>
<td>0.325 ≤ RH &lt; 0.475</td>
<td>Level 3 (high cloud)</td>
</tr>
<tr>
<td>0 &lt; RH &lt; 0.325</td>
<td>Clear skies</td>
</tr>
</tbody>
</table>

Long-wave radiative cooling at each level is a function of the level itself and of the level of cloud base, and is determined from a table compiled by Paltridge (1973a) from the results of an application of the radiation model of Rodgers and Walshaw (1966).

To obtain the rate of heating due to short-wave radiative flux, we first determine the short-wave flux at each half level. The short-wave radiation varies diurnally and is a function of zenith angle, which is an easily calculated function of latitude, longitude, solar declination and model timestep. (See, for example, Haltiner (1971), page 171.) At level K+½ (corresponding to the top of the atmosphere) the short-wave radiative flux equals $S_0 \cos \alpha$ where $S_0$ is the solar constant and $\alpha$ is the local zenith angle. At lower half levels the flux is obtained according to the assumptions that a solar beam which has passed through a layer of cloud is depleted by 18% and that the atmosphere above level K+½ has an absorptivity equal to 0.099 w(K)\(^{-1}\) (Paltridge 1973b, Yamamoto 1962). w(K) here is the precipitable water above model level K+½. The short-wave radiative heating at a grid point for the level K is then calculated as the flux divergence of levels K-½ and K+½. The depletion due to cloud is unrealistically small, but the results presented later justify it in the present context.

The surface temperature at land grid points (excluding those over Antarctica) is determined from the heat balance at the surface. This is a statement that the sum of the fluxes directed towards the surface equals the sum of the fluxes directed away from the surface, and is solved iteratively by Newton's method (Sokolnikoff and Redheffer 1958, page 684). The component energy fluxes of the heat balance equation are as follows:
(i) The downward flux of long-wave radiation. The values used are empirical and are functions only of the level of cloud base. They were derived from representative temperature and moisture profiles for the Australian region using the radiation scheme of Manabe and Moller (1961).

(ii) The downward flux of short-wave radiation. The incoming solar radiation at the surface is determined as described above. The flux into the surface is equal to this incident radiation less that part reflected by the surface.

(iii) The upward transport of sensible heat by eddy exchanges. Here the bulk aerodynamic formula is used:

\[ \text{FLUX} = C_p \cdot \rho \cdot C_D \cdot |V|_6 \cdot (\theta_s - \theta_6) \]

\[ \approx C_p \cdot \rho \cdot C_D \cdot |V|_6 \cdot (T_s - \theta_6). \]

\[ \theta_s \] is the potential temperature at the surface. \( T_s \) is the surface temperature being determined. \( \theta_6 \) and \( |V|_6 \) are respectively the potential temperature and the magnitude of the horizontal wind velocity at the model's lowest dynamic level (sigma = 0.98). \( \rho \) is the air density and \( C_D \) is the specific heat at constant pressure. The drag coefficient \( C_D \) is equal to 0.004 at each land grid point under normal day time circumstances. In regions of very light wind, for avoidance of excessive heating of the surface, convective eddies must be parameterised. For heat balance purposes this is done by allowing in equation 1 a minimum value of 2 m s\(^{-1}\) for \( |V|_6 \). When this minimum wind is used, the drag coefficient is set equal to 0.006.

When the flux as given by equation 1 is negative (ie, a sensible heat flux towards the surface), the value of \( C_D \) used is 0.0005. This value was obtained from numerical experimentation to give a realistic diurnal variation of sensible heat flux.

The same drag coefficient value is used for all calculations at a particular grid point at a given timestep. These are the sensible heat flux and the latent heat flux calculations used in the heat balance equation for the determination of surface temperature, and the surface sensible heat, latent heat and momentum flux used in the model's vertical diffusion parameterisation. Thus in stable conditions \((C_D = 0.0005)\) very little energy and momentum is exchanged between the model atmosphere and the ground.

(iv) The upward transport of latent heat by eddy exchanges:

\[ \text{FLUX} = DW \cdot L \cdot \rho \cdot C_D \cdot |V|_6 \cdot (q_s(T_s, p_s) - q_6). \]

\( DW \) is the availability of soil moisture \((0 < DW \leq 1)\). \( L \) is the latent heat of evaporation \( q_s(T_s, p_s) \) is the saturation mixing ratio at the surface, and \( q_6 \) is the mixing ratio at level 6.

(v) The upward long-wave radiation flux. The value used is \( \sigma T_s^4 \) where \( \sigma \) here is the Stefan-Boltzmann constant.

(vi) The ground flux parameterisation consists of a vertical finite difference formulation (Brook 1975) of the equations for heat conduction into the ground,

\[ \frac{\partial T}{\partial t} = K_1 \frac{\partial^2 T}{\partial z^2}, \]

\[ G = K_2 \frac{\partial T}{\partial z}, \]
where $Z$ is the vertical coordinate; $T$ is temperature; $t$ is time; $G$ is the ground flux; $K_t$ is the thermal diffusivity; and $K_s$ is the thermal conductivity. For this parameterisation an extra model variable of temperature at a depth in the ground is carried. A third climatological temperature must be specified corresponding to a temperature at such a depth that its value does not change over the period of a short-term forecast. In practice, for this temperature the value at 30 cm depth is used, taken from the maps of world patterns of monthly soil temperature prepared by Chang (1958).

As stated in the introduction, the objectives of this paper are to show that very simple parameterisations can be used to represent the energy transformations at the earth's surface and to examine the model atmosphere's reaction to such a scheme's inclusion. The third stage of refining the scheme to incorporate the most recent and exact empirical radiative relations should logically be carried out after the completion of these first two steps. Following this philosophy the description given above of the present flux parameterisations has been brief. The scheme as described has limitations such as the conceptual inconsistency of a 'constant' downward long-wave radiation flux and the use of bulk transfer coefficients determined traditionally by numerical experimentation.

The variation with time of fluxes (i) to (vi) over the period of a two day model integration at a grid point on the Australian continent is shown in Fig 2. The fluxes compare favourably with the experimentally determined fluxes in Fig 3. Discontinuities in the model flux curves correspond to discontinuities in the values of the level of cloud base and of the drag coefficient. In Fig 4 we show the resultant surface temperatures at four different Australian grid points. This figure and Fig 2 are based on the two day prognosis beginning at 2300 GMT on 17 December 1973. This prognosis is discussed in more detail in a later section of this paper.

Comparing Fig 2 with the curve for grid point A (latitude 31°S, longitude 118°E) in Fig 4, the lag in time of the surface temperature maximum behind the short-wave radiative flux maximum has the realistic value of two hours. The amplitude and median value of the diurnal range, however, seems to be too small. For comparison, at Perth (32°S, 116°E) the mean December one inch depth 6 am soil temperature was 18.8°C over the years 1957 to 1961 and the corresponding 3 pm temperature was 49.9°C (Kininmonth 1962).

Little work has been done at this stage on the specification of surface fluxes in areas of high topography; for the sake of the integrations in this study fluxes suitable for the virtually non-topographical Australian region were specified: the albedo was set equal to 0.3; and the availability of moisture was set to 0.8 over the sea and 0.05 over the land. It was found necessary, however, to reduce the temperature which resulted from the heat balance equation by the standard lapse rate of 0.65°C for every 100 m by which the surface height exceeded 500 m. This procedure seems to be sufficient for hemispheric prognoses used specifically for Australian regional forecasting.

Using an IBM 360/65 computer, the addition of the heat balance scheme to the model increases the time required to perform a forecast by less than one per cent.

**SYNOPTIC CASE STUDIES**

Two prognoses, subsequently denoted as the 'control' and 'heat balance' prognoses, were carried out for each of the three randomly chosen situations discussed in this section. In the control prognosis the sea surface temperatures were climatologically specified (Bureau of Meteorology, Australia, 1973) and the temperatures over all land surface grid points were set equal to the values of temperature found at the lowest model level as specified by the starting analysis. The land-sea temperature difference in the region of interest in this prognosis was less than 4°C and is illustrated for a December situation in Fig 5.
Fig 3 The daily course of the heat balance components over a crop of Sherpa wheat at Rutherglen (36°S, 145°E), averaged over 15 days in December 1971 (from data of Patridge et al. (1972)). The eddy flux of sensible heat was not measured. The curve here represents the residual of the other fluxes.
Fig 4 Variation of surface temperature at four Australian grid points during the 2-day prognosis beginning at 2300 GMT on 17 December 1973.
Fig 5  Climatological sea surface temperatures, in degrees celsius, used in the 2-day prognoses beginning 2300 GMT 17 December 1973. The values in brackets are the temperatures used at grid points A, B, C and D during the control prognosis. The variation of temperature at these grid points during the heat balance prognosis is given in Fig 4.

Fig 6  Mean sea level pressure analysis for 2300 GMT 17 December 1973.
Fig 7 A sequence of charts showing the evolution of the real and numerically simulated atmospheres during the 48 hours from 2300 GMT 17 December 1973. In the top two charts are MSL pressure analyses at 12-hour intervals. In the centre charts are heat balance prognoses for the same times. In the bottom two charts are the corresponding control prognoses.
Fig 8 A sequence of charts showing the evolution of the real and numerically simulated atmospheres during the 48 hours from 2300 GMT 17 December 1973. In the top two charts are MSL pressure analyses at 12-hour intervals. In the centre charts are heat balance prognoses for the same times. In the bottom two charts are the corresponding control prognoses.
The drag coefficient over the land was in the control prognosis equal to 0.004. In the heat balance prognosis, non-antarctic land surface temperatures were specified at each timestep at each grid point by the solution of the heat balance equation. The drag coefficient was allowed to vary as described above. In both prognoses the drag coefficient over the sea was set to 0.002.

2300 GMT 17 December 1973

A section of the hemispheric mean sea level pressure analysis for this situation is shown in Fig 6. A significant feature of the analysis is the easterly trough development near the west coast of Australia. Troughs of this type directly control the weather over Western Australia during summer and the prognosis of their development and movement is of importance for regional forecasting and the prediction of maximum temperature.

The analysed and forecast pressure patterns for the subsequent 48 hours are shown in Figs 7 and 8. In the real atmosphere the trough deepens in the first 12 hours of the sequence and over the two days moves approximately 15 degrees of longitude towards the east. The heat balance prognosis simulates the sequence very well up to 36 hours, whereas in the control run there is no longer any trace of an easterly trough after 24 hours. Differences between the two prognoses which would be of particular consequence are the simulation of the wind direction and pressure gradient over the southwest corner of the continent and the position of the area of light winds in the Great Australian Bight region. A diurnal variation is discernible in the intensity of the trough and can be seen in both the analyses and the heat balance prognoses. The prognosis trough, however, is at all times less intense than the trough in the real atmosphere.

2300 GMT 10 January 1975

This sequence (Fig 9) represents the type of situation which current southern hemisphere forecasting models have difficulty simulating. Over 36 hours the high pressure system centred in the Great Australian Bight is steady in both intensity and position. The synoptic consequences of this are persistent warm winds to the southwest corner of the continent and cold southerly winds to the densely populated southeast corner. Both prognoses move the central position of the high north by more than 5 latitude degrees and a similar distance to the east, so that both of these synoptic effects are removed. The heat balance prognosis, however, still gives an improvement in spot pressure values over the Australian continent, primarily over the 'heat trough' areas over Western Australia, the Northern Territory and Queensland. Another improvement is the maintenance, through the better delineation of the heat trough, of a southerly gradient along the northwest coast.

0000 GMT 3 November 1969 (Figs 10 to 12)

The 36-hour heat balance prognosis gives improved gradients along the eastern Australian coast, resulting primarily from a better prediction of pressures over eastern Australia. In particular, the heat balance prognosis correctly predicts northerly gradients along the Queensland coast, while the control prognosis predicts southerly gradients.

However, the control prognosis is better in terms of gradients over southern Western Australia. It appears that the pressure gradients associated with the low in the Bight predominate over the thermal effect of the land mass in this situation. A similar situation arises in the 96-hour prognosis (Figs 11 and 12) where the heat balance model has produced a trough over the Northern Territory, where a ridge in fact developed.
Fig 9 Analyses and prognosis for the 36 hours from 2300 GMT 10 January 1975: (a) MSL pressure analysis 2300 GMT 10 January 1975; (b) MSL pressure analysis 1100 GMT 12 January 1975; (c) 36-hour heat balance prognosis valid at 1100 GMT 12 January 1975; (d) 36-hour control prognosis valid at 1100 GMT 12 January 1975.
Fig 10 Analyses and prognoses for the 36 hours from 0000 GMT 3 November 1969: (a) MSL pressure analysis 0000 GMT 3 November 1969; (b) MSL pressure analysis 1200 GMT 4 November 1969; (c) 36-hour heat balance prognosis valid at 1200 GMT 4 November 1969; (d) 36-hour control prognosis valid at 1200 GMT 4 November 1969
Fig 11 Initial and verifying mean sea level analyses for the 96-hour November sequence.
Fig 12 Mean sea level predictions based on the analysis of Fig 11(a).
Hemispherically after 96 hours we see little difference in the simulation of the major systems of the circumpolar westerly stream; however, there is evident an increase in the intensity of the subtropical high pressure belt due to the displacement of mass from the continents. An improvement in the Australian region is the forecast of the trough over southern Western Australia. The simulation of the cyclogenesis near the east coast is of interest. In the real atmosphere this cyclogenesis manifests itself as a low in the easterlies resulting from the interaction between the upper troposphere component of the low over the continent in Fig 10(a) and the residual heat low shown over the continent in Fig 10(b) and (c). In the heat balance prognosis (Fig 12(a)) cyclogenesis is evident in the area, but is seen as an extension of a high latitude westerly trough. A more fundamental difference between the real and the simulated systems can be seen on the 500 mb height analysis and prognosis (not shown). The real atmospheric system is a deep system quite evident at 500 mb, whereas the prognosticated system is very shallow. Reasons for this difference are associated with the model's jet stream simulation during the development stage of the system. This can be seen in Fig 13, which shows the 48-hour 500 mb heat balance prognosis field and the corresponding real atmosphere analysis. Superimposed are the 80 knot and 100 knot isotachs from the simultaneous 200 mb height prognosis and analysis fields. The feature of note with regard to cyclogenesis is the difference in the position of the jet core relative to the axis of the 500 mb trough. Another aspect of the model's performance which we associate with the poor forecasting of the east coast low is the characteristic lagging of mid-latitude systems as is evidenced by the relative positions of 500 mb troughs in Fig 13(a) and (b).

Considerable differences over the land masses of South America and South Africa are also indicated in Fig 12 (a) and (b), although inconsistencies in reduction to sea level techniques obstruct comparison with the analysed field.

Discussion

These three case studies affirm the importance of land-sea temperature contrast in the context of short-term numerical predictions in the Australian region. They establish also the feasibility of including this contrast through the use of a very simple heat balance scheme. From the experience gained in the implementation of these case studies, it seems apparent that an essential aspect of such a heat balance scheme is diurnally varying solar insolation accompanied by a stability dependent drag coefficient. One pleasing aspect of the results discussed is the improvement in the six level model's ability to simulate the behaviour of the Australian summer heat low.

These conclusions are quantitatively substantiated by the mean sea level verification statistics for the Australian region shown in Fig 14. The statistics, which are averaged over the three situations, demonstrate a marked improvement in the results of the heat balance prognoses over those of the control prognoses. The only statistic in which the control prognosis gives a better result is the bias in the 36-hour forecast. This result can be justified when we note from Fig 8 that at 1100 GMT on 19 December 1973 a ridge of high pressure extended across the Great Australian Bight. This ridge was forecast by neither form of the model so that they both exhibited a negative bias in the 36-hour forecast. In the case of the heat balance prognosis, however, the bias is increased by the improved (lower) pressure values over the continent. Similar remarks concerning bias can be made for the 36-hour prognoses valid at 1100 GMT on 12 January 1975 shown in Fig 9.

REPETITION USING THE 1971 MODEL

The results given above are related to the effect of land-sea temperature differences on the flux divergence of heat and moisture at the lowest model level. In the heat balance prognoses there is, entering the model atmosphere at level 6, an input of diurnal energy pulses.
Fig 13  500 mb geopotential fields for 0000 GMT 5 November 1969; (a) analysis; (b) heat balance prognosis. The superimposed isotachs are from the corresponding 200 mb height–wind analysis and prognosis. (The hatched areas correspond to 200 mb winds greater than 80 knots and the cross hatched to winds greater than 100 knots.)
Fig 14 Average time variation of various mean sea level verification statistics for the Australian region (110° to 160° E, 20° to 45° S).
In the model runs constituting this study, sub-grid scale processes were parameterised according to the scheme of Kininmonth (1974). This scheme involves vertical eddy exchange of momentum, moisture and heat right through the model troposphere and a convective parameterisation based on the CISK theory of penetrative convection forced by moisture convergence in the frictional boundary layer.

Over the three situations, however, the interaction between the diurnal energy input and the sub-grid scale physics led to, after 48 hours, a warming of the 500 mb heat balance temperature field of 0.8°C relative to the corresponding control prognosis field, and subsequently may be suspected of unduly influencing our results. Thus, for further substantiation, the experiments were repeated using the version of the primitive equation model described by Gauntlett and Hincksman (1971). In this model cumulus convection is included by means of a moist convective adjustment process (Manabe et al. 1965), whereby at any time that the lapse rate at a grid point exceeds the moist adiabatic value and the relative humidity exceeds 80 per cent, the lapse rate is adjusted back to the moist adiabatic value in such a way that the potential energy for the vertical column is conserved. The vertical eddy transfer of heat is included only next to the earth's surface.

This second series of prognoses yielded a uniform improvement in the performance of the model with the heat balance included (see Fig 15). There was no detectable difference between the upper tropospheric temperature fields of the heat balance and the control forecasts.

It is beyond the scope of this study to investigate the differences in the performances of the 1971 and the 1975 versions of the P.E. model or on the nature of the interaction between the diurnal land surface temperature variation and the two types of model physics. This is especially true when we consider that the Kininmonth convective parameterisation is still at a developmental stage. However, in the model using the convective adjustment the response to the land-sea temperature contrast is slower than in the penetrative convection form of the model. The two 96-hour hemispheric prognoses for the convection adjustment form of the model are shown in Fig 16. As before, the heat balance prognosis is better in the simulation of the trough over southern Western Australia. Comparing Figs 12 and 16, in the latter the overall thermal effect is not as pronounced. This is evidenced by the smaller increase in the intensity of the subtropical ridge and (with the qualification expressed at the end of the previous section) by the different relative intensities of the systems over Africa and South America.

CONCLUDING REMARKS

This paper has investigated the relevance of land-sea thermal contrast to short-term forecasts in the Australian region. The test was severe, as the magnitude and range of the surface temperature were underestimated. In each of the three situations studied, however, the synoptic fields were better represented than previously; and the skill scores and rms errors were significantly better for the heat balance prognosis than for the control prognosis.

In the experimental prognoses the land surface temperature was determined by balancing very simple energy flux parameterisations under the control of the model dynamics. The heat balance scheme has deficiencies which detract from its total effectiveness; but improvements to it can be made without significant loss of its essential simplicity. Examples of suitable radiative relationships are described by Corby et al. (1972) and by Paltridge (1974).

An alternative to the use of bulk transfer relationships is the simultaneous solution with the heat balance equation of either constant flux layer or outer layer similarity relations as described by Monin and Zilitinkevich (1967) and Clarke (1970). This method, though more physically sound, is far more expensive in computer storage space and calculation time, and has not yet proved justifiable in the context of short-term forecasts. A study of the actual values which should be used for the bulk transfer coefficients based on observational evidence has recently been performed by Garratt (1975).
Fig 15 Average time variation of various mean sea level verification statistics for the Australian region (110° to 160° E, 20° to 45° S), for the hemispheric primitive equation model described by Guenther and Hirschman (1971).
Fig 16. Mean sea level predictions based on the analysis of Fig 11(a) for the 1971 P.E. model.
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