Numerical modelling of a sea-breeze circulation over Cleveland Bay

Jianmin Ma
Atmospheric Environment Service, Downsview, Ontario, Canada
(Manuscript received April 1995; revised August 1996)

A tropical sea-breeze circulation is simulated by using a two-dimensional mesoscale primitive equations model. The numerical experiments are carried out for a special case during winter under the typical prevailing trade wind regime over northern Queensland. Particular attention is paid to the behaviour of the onshore flow component, which occurs over the nearshore sea surface at night, and the corresponding sea-breeze circulation offshore. The interaction between a land breeze and the onshore wind component is discussed in the case where the land breeze lies at the coast rather than penetrates out to sea. The effects of local terrain (a hill), an island and large-scale wind on the sea-land breeze circulation are also investigated.

Introduction

Sea-breeze circulations in northern Australia have been the subject of many studies due to their characteristics under certain synoptic conditions and their apparent effects on local weather phenomena. These studies, including field observations and numerical treatments, were mainly focused on Cape York Peninsula and the Gulf of Carpentaria region, where the so-called ‘morning glory’ occurs frequently (e.g., Clarke et al. 1981; Clarke 1983a, b, 1984; Physick and Smith 1985; Smith et al. 1982; Smith and Morton 1984; Noonan and Smith 1986). These investigations explore weather characteristics and outline possible explanations in terms of dynamics and physics, which provide a fundamental understanding and insight into the interaction of the thermally induced sea-breeze circulation with the local terrain and the large-scale wind regime.

The sea-breeze circulation in Cleveland Bay near Townsville occurs, in general, under a similar large-scale wind regime as Cape York Peninsula during winter. An investigation of the sea-breeze circulation offshore has been conducted by Dexter et al. (1985) using sea-surface backscatter from an HF radar. An important offshore feature obtained by their investigation was the persistence of an onshore surface wind component almost throughout the night, while surface winds almost ceased near the coast after the land-sea temperature difference became negative (around sunset). A similar phenomenon was reported by Hsu (1969) on the Texas coast, where an onshore wind with wind speed up to 1 m s\(^{-1}\) persisted through 0400-0500 local time at a station some 20 km offshore when a land-breeze of about 4 m s\(^{-1}\) prevailed on the coast, with a corresponding cumulus line observed shoreward from that station indicating low-level convergence. In their study of the tropical sea-breezes in the Gulf of Carpentaria region of northern Australia, Physick and Smith (1985) reported that under suitable conditions the deep (1100 m) sea-breezes observed at the Gulf were shown to possess a stronger circulation at night than during the day, while the sea-breeze flow was cut off inland during the night (an important factor leading to the ‘morning glory’), but they did not report the persistence of the onshore flow offshore. It has been found that the nocturnal persistent onshore component in Cleveland Bay is still evident up to some 50 km offshore, according to the routine meteorological observation at Rib Reef (Dexter et al. 1985).
A numerical simulation was also carried out by Clarke (upon request from Dexter) for this case in order to compare with radar observations (Dexter et al. 1985). The results of Clarke showed a persistent onshore component from the sea surface near the shore up to 5 km inland until about 0300 local time.

Much less is known about the vertical profiles of the sea-breeze circulation and possible physical factors attributed to this weather phenomenon. Further evidence and possibilities are worth exploring. Firstly, we would like to ascertain whether this type of sea breeze can be regarded as a fluctuation of a large-scale wind regime, as is the usual definition of a sea breeze in the geostrophic wind field. Secondly, while the onshore wind component persisted over the sea surface during the night, the land breeze persisted over land only, rather than penetrating over the sea surface, leading to a low-level region of convergence near the coast. This will certainly affect the structure of the nocturnal boundary layer and pollutant transport, and it is therefore important to make a detailed analysis of this process so that we will be able to gain more knowledge about the problems of pollutants trapped in the circulation. Finally, the scale and intensity of the onshore wind circulation need to be further investigated, because there is still uncertainty about them due to the limitations of the radar observations.

In this paper, we perform a series of numerical experiments by using a mesoscale primitive equations model (Ma 1995), with particular emphasis on the onshore wind component near the coast during the night, in order to investigate the structure and evolution of the sea-breeze circulation.

Model and numerical experiments design

The basic model equations are the primitive equations of atmospheric motion as simplified by the incompressible and hydrostatic approximations with a topography coordinate transformation in the vertical coordinate (Pielke 1984). The model is two-dimensional and calculates y-z cross-sections perpendicular to the mean coastline orientation in the vicinity of Townsville (i.e., Cleveland Bay). Hence, the x-axis is defined to be aligned with the mean coastline orientation, that is 113°-293°. Thus, in the topography coordinate system the governing equations are:

\[
\frac{\partial \theta}{\partial t} + \frac{\partial \theta}{\partial y} + w_a \frac{\partial \theta}{\partial z_a} = \left( \frac{H}{H-Z_o} \right)^2 \frac{\partial}{\partial y} \left( \frac{K_H}{\theta} \frac{\partial \theta}{\partial y} \right) + \frac{\partial}{\partial y} \left( K_m \frac{\partial \theta}{\partial y} \right) + F \quad \text{...3}
\]

\[
\frac{\partial q}{\partial t} + \frac{\partial q}{\partial y} + w_a \frac{\partial q}{\partial z_a} = \left( \frac{H}{H-Z_o} \right)^2 \frac{\partial}{\partial y} \left( K_m \frac{\partial q}{\partial y} \right) + \frac{\partial}{\partial y} \left( K_H \frac{\partial q}{\partial y} \right) \quad \text{...4}
\]

\[
\frac{\partial \bar{v}}{\partial y} + \frac{\partial \bar{v}}{\partial z_a} - \frac{1}{H-Z_o} \frac{\partial Z_o}{\partial y} = 0 \quad \text{...5}
\]

\[
\frac{\partial \Pi}{\partial z} = \frac{H-Z_o}{H} \frac{\partial \theta}{\partial y} \quad \text{...6}
\]

where \( F \) in Eqn 3 is the radiation heating or cooling, \( \Pi = (p/p_0)^{\gamma/(\gamma-1)} \), \( K_m \), \( K_H \), and \( K_q \) are eddy coefficients for momentum, potential temperature and water vapour, respectively (\( K_Q = K_q \) is assumed in the model), \( K_H \) is the horizontal exchange coefficient, \( q \) is specific humidity and other symbols have their usual meaning.

In the model we adopt a turbulent kinetic energy (TKE) prognostic equation and diagnostic length-scale as the turbulent closure, that is, the so-called E-model, which uses only the turbulent kinetic energy equation and a simple relation for the ratio between eddy diffusivities for momentum and heat. The TKE equation in the terrain-following coordinate is written as (Pielke 1984):

\[
\frac{\partial E}{\partial t} + \frac{\partial E}{\partial y} + w_a \frac{\partial E}{\partial z_a} = K_m \left( \frac{H}{H-Z_o} \right)^2 \left( \frac{\partial \theta}{\partial z} \right) + \left( \frac{\partial \varphi}{\partial z} \right)^2 - \frac{1}{H-Z_o} K_q \left( \frac{\partial \theta}{\partial z} \right)^2 - \frac{1}{H-Z_o} K_H \left( \frac{\partial \theta}{\partial z} \right)^2 - \frac{1}{H-Z_o} \left( \frac{\partial \theta}{\partial z} \right) \left( \frac{\partial \varphi}{\partial z} \right)^2 - \frac{1}{H-Z_o} \left( \frac{\partial \theta}{\partial z} \right) \left( \frac{\partial \varphi}{\partial z} \right) \quad \text{...7}
\]

where respective terms in the right-hand side of Eqn 7 denote shear production, buoyancy, turbulent dissipation and turbulent transport, \( c_T \) is a constant and \( E \) is the grid-volume average turbulent kinetic energy. The eddy diffusivities for momentum and heat are defined by (Huang and Raman 1991):

\[
K_m = c_2 S_m / (2E)^{\frac{1}{2}} \quad K_q = P \frac{c_2 S_m / (2E)^{\frac{1}{2}}}{(2E)^{\frac{1}{2}}} \quad \text{...8(a) and (b)}
\]

where \( c_2 = 0.2 \), \( S_m \) is a stability function, \( P \) is the inverse turbulent Prandtl number and \( t \) is the eddy mixing length. For detailed descriptions of these functions and parameters, refer to Huang and Raman (1991).

Long and short wave radiation heating are taken into account under the assumption of a clear-sky condition. The long wave radiation flux was evaluated using the scheme developed by Chen and Cotton (1983). The computation of short wave radiation uses the scheme of Katayama (1974). Surface fluxes of heat, momentum and moisture are computed using the technique described by Louis (1979).
The initial wind profile in the vertical is obtained by solving Ekman's equation and the initial temperature distribution is specified by:

\[ T(p) = T(p_0) \left( \frac{p}{p_0} \right)^{\frac{\gamma}{R_s}} \]

where \( p_0 \) is 1000 hPa and the lapse rate is taken to be \( \gamma = 6.1 \) K km\(^{-1}\). The initial temperature at the surface is taken to be the sea-surface temperature \( (T(p_0) = 294.2 \) K). The land-surface temperatures are obtained by a prognostic surface energy balance equation which considers the surface sensible and latent heat fluxes, short and long wave radiation, and thermal diffusion in the soil (Blackadar 1979).

Grid-points in the model domain are evenly spaced in the horizontal, the distance between grid-points being 5 km. There are 80 grid-points in all and the total horizontal domain is 400 km, of which 250 km is land and 150 km sea. In the vertical, grid-points increase from \( \Delta z = 10 \) m at the first layer above the surface to \( \Delta z = 500 \) m at layers 1500 m to 8000 m with a total of 25 levels.

The model equations are solved numerically by a finite difference method. The horizontal wind components are defined at grids using the Arakawa-C stagger. Time differencing is forward for all terms in which the advection terms are solved with the forward-in-time, spline-interpolation upstream scheme (Pielke 1984). To allow a large time-step, vertical diffusion is solved implicitly with the Crank-Nicholson scheme, while horizontal diffusion and pressure gradient as well as other spatial derivatives are centred. The time-step is taken to be 30s. The model is run from 4 am (i.e., 0400 AEST) and then integrated over the next 24 hours.

The horizontal wind components normal to the coastline (i.e., y-direction in the model) will be regarded as the sea or land breeze. To compare with Dexter et al.'s measurements (1985), we chose the same weather conditions as those during the period of Dexter's observations, which roughly reflect the typical situation at Townsville in that period. The analyses of the general synoptic situation for the eight-day period covering the corresponding experiments were given by Dexter et al. (1985). Essentially, the flow pattern over northern Queensland displayed a broad southeast trade wind regime. The coastline orientation and local topography contribute to dry, relatively stable and slightly offshore winds. We will regard a total surface wind directional variation between 110°-290° as an index of the sea-breeze circulation. Considering this background situation, we shall use a prescribed wind direction of 110° as the input geostrophic wind direction. (Clarke used the same value of geostrophic wind direction in his numerical experiment in order to compare with the radar observations, see Dexter et al. (1985)).

The results presented in the following section are derived from four main numerical experiments as listed in Table 1. In these four experiments the geostrophic wind direction remains unchanged while the geostrophic wind speed is chosen to be 7 m s\(^{-1}\) (the same value used by Clarke in his numerical experiment, Dexter et al. (1985)) except for one of the experiments where it is chosen to be 12 m s\(^{-1}\). The topography included in the experiments is a smoothed, idealised hill roughly representing the dominant topography of Townsville, with a peak height of 300 m, centred at 30 km inland. In one of the experiments we shall consider the approximate impact of Magnetic Island (see Fig. 1) upon the sea-breeze circulation. We would like to point out, however, that by including an island in a 2D simulation, effectively an infinitely long peninsula is modelled. Therefore, we shall assume that there is no island within the model water surface domain in the direction normal to the coastline of Cleveland Bay (see Fig. 1).

<table>
<thead>
<tr>
<th>Run</th>
<th>Topography</th>
<th>Island</th>
<th>Geostrophic wind speed</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Excluded</td>
<td>Excluded</td>
<td>7 m s(^{-1})</td>
</tr>
<tr>
<td>2</td>
<td>Included</td>
<td>Excluded</td>
<td>7 m s(^{-1})</td>
</tr>
<tr>
<td>3</td>
<td>Excluded</td>
<td>Included</td>
<td>7 m s(^{-1})</td>
</tr>
<tr>
<td>4</td>
<td>Included and excluded</td>
<td>Excluded</td>
<td>12 m s(^{-1})</td>
</tr>
</tbody>
</table>

Fig. 1 Location diagram showing the general experiment area. The topographic contours at 200 and 500 m are also indicated.
Results

Sea breeze with and without terrain

At first we shall simply review the earlier stages of the sea-breeze flow. After a ten-hour model run to 1400 AEST, we can see from Fig. 2 that the sea breeze has become well established on the coast in both Runs 1 and 2. The maximum wind speeds of the onshore flow at 10 m height are 1.6 m s$^{-1}$ for Run 1 and 2.0 m s$^{-1}$ for Run 2 at the coast, increasing to 2.9 m s$^{-1}$ and 3.0 m s$^{-1}$ over the sea (20 km offshore). Compared with the wind speed of 3.5 m s$^{-1}$ from Dexter et al.'s HF radar observation in pixel 7 (21 km offshore), the simulated sea breezes in both cases are weaker over the sea. However, after sunset the predicted onshore velocities of 3.0 m s$^{-1}$ for Run 1 and 2.3 m s$^{-1}$ for Run 2 at 50 km offshore are in fair agreement with the observed value of 2.6 m s$^{-1}$.

The convectively well-mixed layer has developed to a height of 1500 m or so. However, at this stage, there is no remarkable difference in the intensity of the sea-breeze circulation between the two cases. The vertical extent of the sea breeze is about 200 m in both cases. The seaward expansion of the sea breeze is much greater than landward. We shall show later that if the large-scale wind speed with the same direction increases to a certain value (12 m s$^{-1}$ in Run 4), the sea-breeze cell will reduce significantly. Mahrer and Pielke (1977) have carried out numerical experiments to investigate the effects of a mountain on sea-land circulation and found that the combined mountain and sea-land simulations produced a more intense circulation than they did separately, and that by early afternoon the sea-breeze had reached the lee side of the mountain. For a small-scale hill as in our case, we have not found that the sea-land circulation is much stronger after introducing the hill than in the flat situation (Run 1). However, in agreement with Mahrer and Pielke's results, the sea breeze in Run 2 penetrates faster than in Run 1, which is a result of the combined effects of upslope wind and onshore wind inflow.

Following Arritt (1989, 1993), we define the inland penetration of the sea breeze as the westernmost location within the lowest 1000 m that had a negative $v$-component of at least 1 m s$^{-1}$. From our simulation we find that when surface heating has disappeared, the sea breeze in Run 1 reaches about 30 km inland but in Run 2 extends to 45 km inland. The inland penetration between 1400 and 2000 AEST has a rate of 2.0 km h$^{-1}$ for Run 1 and 3.6 km h$^{-1}$ for Run 2. These figures and the results in Fig. 3 for Run 1 and 1800 AEST for Run 2 until 0100 AEST on the next day (see Fig. 4). This accelerating cool-air movement has been explained using gravity current theory by Physick and Smith (1985) and Sha et al. (1991). In the late afternoon the solar radiative heating decreases but the temperature of the land-air still increases gradually. At the same time, the temperature of the sea-air reaching the front decreases continuously as it has had less heat added to it since crossing the coast. Thus, the temperature difference across the front at low levels increases and the sea-breeze front begins to accelerate. Although it is difficult to identify the sea-breeze front from our model results, the gravity current theory can explain the inland penetration of the sea breeze.

The further evolution of the sea-land circulation in Run 1 and Run 2 during the night is illustrated in Fig. 4 in which we only plot up to 3000 m in the vertical so that the sea-breeze effects near the surface become more apparent. The plots show the vertical cross-section of the horizontal wind speed normal to the coastline every two hours from 2200 AEST to 0200 AEST on the next day. For Run 1 the onshore flow has penetrated 50 km inland and about 100 km offshore at 2200 AEST with a maximum wind speed of 2.7 m s$^{-1}$ within the feeder flow at 50 m above the sea surface, but the maximum wind speed (centred at about 50 km offshore) and the vertical extent are reduced significantly. Accordingly, the height of the return flow aloft above the sea-breeze cell also reduces but the intensity remains unchanged, the maximum wind velocity being 4.8 m s$^{-1}$. After two hours, at 0200 AEST, we observe that the area of the onshore flow has decreased. The sea-breeze inflow begins to be cut off at about 50 km inland but can still be found at 40 km offshore and near the coast. By 0200 AEST, the sea breeze reaching deep inland is cut off completely and a vortex about 50 km in length centred at 70 km inland is seen clearly, in good agreement with the numerical experimental results obtained by Sha et al. (1993) by using a two-dimensional nonhydrostatic numerical model with high spatial resolution. They found that at 0200 LST a horizontal vortex was completely detached from the feeder flow and was established and identified as a sea breeze cut-off vortex.

The disturbance caused by the hill terrain on the onshore flow field is quite evident in the subsequent development of the sea-breeze circulation during the night compared to the case without topography involved (Run 1). The wind speed is still large landward of the hill at 2200 AEST, with a velocity of 3.7 m s$^{-1}$ within the feeder flow at 70 km inland and 50 m height, as shown in Fig. 4(c), extending to 600 m in the vertical. Seaward of the hill and 50 km out to sea, the onshore flow is relatively weak with a maximum velocity of 2.5 m s$^{-1}$ near the coast, and of lower extension in the vertical. Compared with Run 1, the intensity of the sea breeze in Run 2 is weaker on the windward side of the hill (including the coast) and offshore, which also suggests the effect of the hill. At this time, we still observe the mountain wind on the lee of the hill but not on the windward slope.
Fig. 2  Vertical cross-section of $v$ and $w$-isotachs and potential temperature isentropes for Run 1 and Run 2 at 1400 AEST. Contour intervals are 1 m s$^{-1}$ for $v$, 0.04 m s$^{-1}$ for $w$ and 2 K for $\theta$. The negative $v$ (dashed line) represent onshore flow and negative $w$ (dashed line) represent downward vertical motion.
Fig. 3  Same as Fig. 2 but for 2000 AEST.
Fig. 4  The vertical cross-section of the meridional wind for Run 1 and Run 2 from 2200 AEST to 0200 AEST.
At 2400 AEST, the flow has become offshore on the seaward side of the hill, but at the sea near the coast we can still find onshore flow up to 50 m in height. It is noted that the v-component in Figs 4 (e) and (f) is positive over the sea, but we shall still regard the flow as onshore flow from Fig. 6 (recalling the coastline orientation being 110°-290°), agreeing with the radar observations conducted by Dexter et al. (1985). They found a persistent onshore surface wind component over the sea, from 12 km to about 50 km offshore, almost entirely throughout the night. The simulated onshore flow from our model can be found to persist in a similar offshore region at 0200 AEST as that observed by radar. As a result, we may expect that there would be convergence near the coast due to the colliding of the offshore flow from the land and the onshore flow from the sea, as reported by Hsu (1969). At 60 km inland behind the hill, the simulated onshore flow decreases gradually in both intensity and vertical extension (Figs 4(d-f)).

In Fig. 5 we plot the vertical profiles of potential temperature for Run 2 at the grid-points 5 km inland and 10 km offshore at 2000 AEST, and 0200 AEST on the second day, respectively. A convectively well-mixed layer was still evident, particularly over land near the coast at 2000 AEST at the lower levels of the atmosphere, which is actually a remnant of the well-mixed layer in the daytime. At 0200 AEST on the second day, the nocturnal inversion became established, the surface temperature at the land decreased rapidly and the corresponding inversion layer extended to about 500 m.

Figure 6(a) contains a plot of the predicted wind direction variation at the first level (10 m in the vertical) over the sea surface for Run 1 and Run 2, in order to compare with the radar-derived wind directions and observations (Fig. 6(b), from Dexter et al. (1985)). It shows clearly a 70° change in wind direction after the sea breeze develops. This wind direction variation compares well with the measurements. The wind direction variations at about 50 km offshore can be also observed, indicating the seaward advance of the sea-breeze circulation, commencing at around 1000 AEST and extending to 25 km by 1100 hours and to 50 km by 1200 hours (Run 2). Figures 6(a) and (b) show that the wind direction alterations exhibit almost no difference between Run 1 and Run 2, indicating that the effect of the hill on the wind direction variation of the sea breeze offshore is negligible.

**Effects of the island**

Since our model is two-dimensional, but the modelled region is characterised by an island (Magnetic Island, see Fig. 1), a problem of confidence may arise in the applicability of the foregoing model simulations as we have set the x-axis parallel to the coastline and there is no island in the direction normal to the coastline. To focus investigations upon the effects of the island only, in Run 3 we have excluded the topography and located an island 20 km offshore with a width of 15 km.

Figure 7 shows the vertical cross-sections of the wind speed, potential temperature and vertical velocity from 1400 to 2000 AEST. It can be seen that at 1400 AEST the sea breeze has become established across the island but extends only to less than 200 m in height. The water body at the right of the island and between the island and land can be identified clearly from the potential temperature profiles. Accordingly, the convectively well-mixed layer is developed over both land and the
Fig. 6  (a) Predicted wind direction at the first model level (10 m) for the range from 0 to 50 km offshore. (b) Radar-derived and measured wind direction.

(a)

(b)

Effects of geostrophic wind speed

The role of the large-scale wind regime in the generation of the sea-breeze circulation has been studied extensively in terms of the effects of the geostrophic or, more commonly, the gradient wind. In a recent investigation Arritt presented results of the effects of the large-scale flow on the sea breeze (Arritt 1993). However, many previous studies were focused on the onshore and offshore gradient winds (e.g., Atkinson 1981; Arritt 1993). For gradient winds parallel to the coastline it seems that no extensive investigations have been reported. In some cases such flow would cause sea breezes to rarely develop, while in other cases little hindrance would be placed on the development of the sea breezes (Atkinson 1981). As discussed above, the sea breeze becomes established in either the modelled or real situations under a moderate wind parallel to the coastline, and hence the large-scale wind at 110° (nearly parallel to shore) with a moderate speed will not directly retard the development of the sea-breeze circulation. In Run 4, however, when we impose a relatively strong wind (12 m s⁻¹) parallel to the coastline, the sea breeze is found to develop only over the sea, rather than to penetrate the coast. It is known that the critical factor determining whether sea breezes develop easily or are hindered is whether the frictional turning associated with the gradient wind parallel to the coastline in the boundary layer opposes the sea breeze or reinforces it. For our situation of winds from 110°, the frictional component is offshore and so opposes the sea.
Fig. 7  Vertical cross-section of $v$ and $w$-isotachs and potential temperature isentropes for Run 3. The contour intervals are the same as Fig. 2. (a), (b), (c), (d) are for 1400, 1600, 1800 and 2000 AEST.
breeze. For the imposed wind speed of 12 m s\(^{-1}\), the offshore component is obviously too strong for a sea breeze to overcome. This finding (for which no detailed results are shown) is obtained from both numerical calculations with and without the hill included.

**Conclusions**

The structure and detailed behaviour of the sea-breeze circulation over Cleveland Bay, Townsville, in northern Australia has been simulated by using a two-dimensional mesoscale numerical model. Under moderate conditions with a 7 m s\(^{-1}\) background wind and model topography, the major characteristics of the sea-land breeze circulation were reproduced by the numerical experiments. The deep inland penetration and subsequent cut-off vortex of the sea breeze in our case can be attributed to the same physical processes as the sea-breeze circulation in the Cape York region, investigated by other researchers, where similar conditions of a dominant background wind field prevailed.

The numerical results confirmed the persistence of the onshore flow offshore during the night. This onshore flow initially developed near the coast in the morning and then penetrated to deep inland in the afternoon. After sunset, the area covered by onshore flow over the water shrunk to an area extending from the shore to 50 km offshore.

The effects of a small-scale island located near the coast upon the sea-breeze circulation were also considered. It was found that the most significant influence of the island is to reduce significantly the deep inland penetration of the onshore flow and prevent the formation of a cut-off vortex inland during the night.

The large-scale wind parallel to the coastline with a moderate wind speed, in general, did not retard the development of the sea-breeze circulation. However, the variation of the wind speed must be considered as an important factor because the relatively strong gradient wind parallel to the coast could hinder the inland penetration of the onshore flow, even though the sea breeze can develop offshore near the coast.

Finally, we wish to point out that although the model reproduced the most important characteristics of the sea-breeze circulation observed at Cleveland Bay, in general, the intensity of the simulated circulation was weaker than the real situation. For example, the wind speed within the feeder flow was less than 5 m s\(^{-1}\) compared with about 6 m s\(^{-1}\) from measurement (from radar and boat, Dexter et al. (1985)). The maximum vertical extension of the sea breeze was only about 600 m rather than the somewhat higher values actually observed at Townsville Airport. Comparatively, Run 2, which includes the topography, gave more reasonable results in comparison with the observations. Only a full, three-dimensional model calculation can hope to produce better quantitative agreement with observation.

**Acknowledgments**

The benefit of fruitful discussions with Dr Robson, of James Cook University of North Queensland, is acknowledged. The author would also like to thank an anonymous reviewer for making valuable comments and suggestions on the original manuscript, which led to a significant improvement of this paper.

**References**


