A simple model of airflow and cloud conditions over Baw Baw Plateau

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(Manuscript received May 1991; revised November 1991)

A simple two-dimensional airflow model has been adopted to calculate cloud conditions over Baw Baw Plateau. The model is intended to assist the Melbourne and Metropolitan Board of Works (MMBW) cloud seeding personnel at Essendon Airport by providing a rough estimate of cloud conditions over the mountain which may help in the planning of seeding flights. The northwest–southeast orientation of the Baw Baw Plateau limits the model's applicability to wind directions from southwest to west-southwest. This wind direction commonly occurs during post-frontal conditions.

In the model, Baw Baw Plateau is represented by a bell-shaped ridge, and the airflow is assumed adiabatic and hydrostatic. The only model input data is an upstream sounding from which the cloud parameters are calculated assuming conservation of entropy and total water along streamlines. The model predicts cloud extent, liquid water content and cloud-top temperature over the mountain. Calculations are presented to decide if the cloud field is convectively unstable.

Introduction

The Melbourne and Metropolitan Board of Works (MMBW) is currently undertaking a cloud seeding experiment aimed at augmenting the precipitation amounts falling into the Thomson Catchment near Baw Baw Plateau. The seeding is performed from an instrumented aircraft, which seeds with silver iodide (AgI) smoke for temperatures below −7°C, and with dry ice (CO₂) for temperatures between −2°C and −7°C. The dry ice seeding is a recent addition to the five-year study, which started in 1987. A more detailed description of the experiment is given by Long and Shaw (1988).

Long and Huggins (1991) have described the synoptic variability of cold fronts which move across eastern Victoria. The wind directions are usually northwesterly or westerly during the pre-frontal and frontal stages; after the frontal passage the wind moves to southwesterly, and a cool marine airflow moves over Baw Baw Plateau. Pre-frontal, frontal and post-frontal activities accounted for 61 per cent of rain days in the Thomson Catchment during 1983–1984 (Shaw and King 1986). Stratocumulus clouds during the post-frontal stage frequently form a persistent cap-cloud as the air moves over Baw Baw Plateau. The cap-cloud is a prime candidate for cloud seeding, provided the meteorological conditions are suitable. Suitable conditions are determined by the cloud-top temperature, liquid water content, wind direction and speed, horizontal extent and continuity of the cloud. Approximately half of the suitable conditions occur during shallow post-frontal cap-cloud situations.

An operations office manned by MMBW staff is located at the Essendon Airport. Here the staff has real-time access to Japanese Geostationary Meteorological Satellite imagery, S-band radar imagery, Laverton rawinsonde data, data from a vertically pointing liquid water radiometer located at Baw Baw Plateau, as well as special forecasts issued by the Bureau of Meteorology. However, even this large amount of information is rarely sufficient to determine if suitable cloud conditions exist near Baw Baw Plateau. Often it is necessary to perform a surveillance flight during which temperature, wind direction and speed, and liquid water content are measured from the instrumented aircraft. It is desirable to have as much information as possible regarding the cloud conditions at the Essendon office in order to determine the necessity of such a surveillance flight.

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The main aim of this study is to provide pre-flight information on cloud conditions over Baw Baw Plateau using omegasonde data. In this paper we describe a simple, analytical two-dimensional model which can be used to calculate airflow trajectories as the air moves up over Baw Baw Plateau. A model which executes rapidly on a small workstation is necessary if the model calculations are to have operational benefits. For this reason it is not practical to run more complex two-dimensional and three-dimensional models which require a supercomputer for performing the necessary calculations. We have chosen to use Queney’s (1947) analytical model for the calculation of air trajectories. The simplicity of this model imposes obvious limitations on the physics which can be represented; this is a necessary trade-off. The model is two-dimensional, hydrostatic, and employs a simplified geometry represented by a bell-shaped mountain. The model does not allow for convection, but we have used it to predict when convection will happen, at which point the model assumptions break down. The model is mainly applicable for the extended periods of southwesterly airflow for which a steady cap-cloud exists over the mountain.

We have coupled the trajectory model with entropy calculations which allow the calculation of temperature and liquid water content along the trajectories. By contouring the liquid water content values, it is possible to create a picture of the outline of the cloud, as well as the liquid water content inside the cloud. It is also possible to calculate the cloud-top temperature. We envisage that the model may provide a rough indication of the cloud conditions near Baw Baw Plateau based on sounding data from Laverton.

In the following sections we describe the foundations for the numerical model, the entropy scheme used for calculating thermodynamic parameters along the trajectories, and the results for some selected input conditions. In the final section, we discuss the performance of the model and its limitations.

**Airflow model and topography**

**Topography**

Baw Baw Plateau is an elongated ridge oriented roughly in a northwest-southeast direction as shown in Fig. 1. A southwesterly wind, perpendicular to the ridge, is more likely to result in a two-dimensional airflow field than any other wind directions. A cross-section of Baw Baw Plateau along the wind direction of 230° is shown in Fig. 2. Also shown is an idealised, bell-shaped mountain with a half-width a = 10 km, and with a height \( h_m = 1300 \) m. The height of 1300 m is roughly the mean height along the length direction of the ridge; the absolute maximum height is approximately 1500 m. The bell-shaped mountain is given by the so-called ‘witch of agnesi’:

\[
h(x) = h_m \frac{a^2}{(x^2 + a^2)}
\]

The bell shape is a fair approximation to the mountain on most of the upstream side, but much more complex terrain exists on the downstream side. There will undoubtedly be some upstream effects due to the complex downstream topography, but the extent of these is unknown; we will assume that they are small for the airflow in the lowermost few kilometres upstream of the mountain where the cap-cloud forms.

In this study we will only be considering two-dimensional flow over the mountain. Even if the mountain is an elongated ridge of about 40 km length, there will undoubtedly be some flow around the mountain. The degree to which this will happen depends on wind speed and stability. Weak wind and very stable conditions will favour airflow around the mountain; strong wind and unstable conditions will favour airflow more predominantly over the mountain than around, but exact calculations cannot be made due to the complex topography.
Fig. 2 The profile of Baw Baw Plateau along the wind direction of 230° and a superimposed profile of a bell-shaped ridge with a height of 1.3 km and a half-width of 10 km.

A model of steady two-dimensional inviscid airflow

Calculation of the airflow over a mountain range is a complicated task even if we assume a model of two-dimensional steady flow. Ideally the model should consider atmospheric stability, wind shear, the mountain's topography, the effect of adiabatic and diabatic heating and cooling on airflow, and the earth's rotation for a large-scale flow, etc. Many studies have attempted to tackle these various aspects. Since there are a good number of reviews, such as Queney et al. (1960) and Eliassen (1974), we will not discuss the various aspects of available models.

For operational reasons mentioned earlier, we have chosen to use a highly simplified model. We have used Queney's (1947) approach to calculate the streamlines of airflow over a bell-shaped mountain. The following discussion closely follows Smith (1979). We assume that the flow is adiabatic and inviscid. Under these conditions, the perturbation of an airstream caused by a mountain can be written as follows:

$$\frac{\partial u'}{\partial x} + \frac{\partial w'}{\partial z} = \frac{g}{c^2}$$  \hspace{1cm} (2.2)

where $u'$ stands for perturbation of the horizontal velocity component, $u(x, z)$, which is a sum of $U(z)$ and $u'(x, z)$. The perturbation of the vertical component, $w(x, z)$, is given by $w'$, $g$ is the gravitational constant, and $c^2 = \gamma RT$. A full symbol list is given in the Appendix.

Equation 2 shows the relationship between divergence in the velocity field and the adiabatic ascent of an air parcel. This equation is drawn from the following sets of equations: the horizontal and vertical momentum equations, the equation of continuity, the first law of thermodynamics and the perfect gas law. For details, see Smith (1979).

Equation 2 can be expressed in terms of the vertical velocity $w'(x, z)$. The subscripts $xx$ and $zz$ stand for $\partial^2/\partial x^2$ and $\partial^2/\partial z^2$, respectively:

$$\hat{w}_{xx} + \hat{w}_{zz} + l^2(z)\hat{w} = 0$$  \hspace{1cm} (3)

where

$$\hat{w} = \frac{\bar{p}(z)}{\bar{p}(0)} w'$$  \hspace{1cm} (4)

The square of $\hat{w}$ is proportional to the energy of the wave disturbance (Eliassen and Palm 1960).

$$l^2(z) = \frac{\beta g}{U^2} + \frac{SU_z}{U} - \frac{1}{4} \xi^2 + \frac{1}{2} \xi_z - \frac{U_{zz}}{U}$$  \hspace{1cm} (5)

This length scale $l^2(z)$ is the Scorer parameter. In this equation $S$ is called the heterogeneity (Queney 1947):

$$\bar{S} = \frac{d}{dz} \ln \bar{p}(z)$$  \hspace{1cm} (6)

$S$ is not related to the generation of buoyancy forces since it does not include the effect of adiabatic expansion. It describes the effect of density $\rho$ variation in the divergence of the velocity field (Smith 1979).

The upstream stability is characterised by the parameter $\beta$:

$$\beta = \frac{d\ln \theta}{dz}$$  \hspace{1cm} (7)

where $\theta$ is the potential temperature.

The Scorer parameter $l^2(z)$ is usually dominated by the first term, the buoyancy force term $\beta g/U^2$. Three terms with $S$ can be neglected under the Boussinesq approximation that density variations are not important unless they affect the buoyancy. The last term, which may be important if there is a strong shear, is included in our calculation.

Airstream over an isolated bell-shaped ridge

The perturbation equation, Eqn 3, in terms of the vertical velocity $w(x, z)$, is used for the calculation of airflow streamlines over the mountain. The solution is best obtained by taking Fourier transforms. Following Queney et al. (1960), a one-sided Fourier integral is used. Then Eqn 3 becomes:

$$\hat{w}_{zz} + [l^2(z) - k^2]\hat{w} = 0$$  \hspace{1cm} (8)

where

$$\hat{w}(x, z) = \Re \int_{0}^{\infty} \hat{w}(k, z) \exp(+ikx) dk$$  \hspace{1cm} (9)

$\Re$ stands for the real part of Fourier integration.

The lower boundary condition is that the airstream on the ground level follows the terrain profile; the upper boundary condition is that the solution does not tend to $\infty$ as $z$ approaches $\infty$. 
We can then obtain the solution of Eqn 8 in terms of vertical displacement, \( \eta(x, z) \). The vertical displacement of a streamline from its upstream height above ground, \( \eta(x, z) \), is given by

\[
\eta(x, z) = \frac{\partial}{\partial x} \left[ \int_0^z h(k) \exp[-i(k^2 - \ell^2)^{1/2}z] \exp(ikx) dk \right] + \frac{\partial}{\partial z} \left[ \int_0^x h(k) \exp[-(k - \ell)^2z] \exp(ikx) dk \right] \]

Here \( h(x) \) is a function representing a bell-shaped ridge and \( h(k) \) is a Fourier transform of \( h(x) \):

\[
h(k) = h_m a \exp(-ka) \]

where \( h_m \) is the height of the mountain and \( a \) is its half-width.

For a narrow mountain, weak stability and strong winds, \( a/l \) is small and the first integral in Eqn 10 becomes small. For the case of a wider mountain (such as Baw Baw Plateau), moderate stability and moderate wind speed, we get \( a/l \gg 1 \).

A case study to be presented in subsequent sections has \( a/l \approx 6.2 \). In this case the flow is nearly hydrostatic and the second part of Eqn 10 is insignificant. The vertical displacement of the trajectory over the mountain, \( \eta(x, z) \), is then as follows:

\[
\eta(x, z) = \frac{\partial}{\partial z} \left[ \frac{h_m a}{a^2 + x^2} \right] \]

In our case, we have only included the buoyancy force term and the wind shear term (the first and last terms of Eqn 5) in the calculation of \( l \). This is because the buoyancy force is dominant for \( a/l \gg 1 \) where flow is hydrostatic.

### Entropy calculation

#### Definitions of total water and entropy

We take the airflow calculated in the preceding section to be a given kinematic flow field, regardless of the fact that in reality the air may be cloudy or clear as it moves up over the mountain. It has been shown by Fraser et al. (1973), who used a somewhat more complex model, that the presence of cloud had only a small effect on the wind field. The streamlines are calculated for 64 levels using Eqn 12 described in the preceding section. Labeling each of these trajectories with the subscript \( j \), we can calculate the upstream conditions, \( p_{0,j}, T_{0,j} \), and \( T_{d0,j} \), by interpolation of the upstream sounding.

Total water mixing ratio and the entropy can be calculated for each of the \( j \) streamlines. The entropy is calculated separately for dry air conditions (air plus vapour) and for cloud conditions (air, vapour and liquid water). We will assume that total water mixing ratio and the entropies are constant along each streamline. Constancy of entropy and total water implies the exclusion of diabatic processes such as mixing, radiation, formation of ice and sedimentation of precipitation.

From the known altitude of the streamlines, we can deduce the pressure by comparison with the upstream sounding. Given a known pressure, total water mixing ratio and entropy, we can use iterative techniques to find the temperature, vapour mixing ratio and possible cloud liquid water content along all the trajectories. This procedure will be described in subsequent sections.

The definition of total water mixing ratio, \( Q \), is:

\[
Q = q_v + q_l \]

where \( q_v \) and \( q_l \) are the vapour mixing ratio and the liquid water mixing ratios, respectively.

The definition of specific entropy for a system of dry air and water vapour is (Paluch 1979; Iri-barne and Godson 1981):

\[
S_{sub} = c_{pd} \ln T - R_d \ln p_d + q_v (c_{pv} \ln T - R_v \ln e) \]

and for cloudy, saturated air:

\[
S_{sat} = (c_{pd} + Qc_v) \ln T - R_d \ln p_d + Q \frac{L_v}{T} \]

where \( S_{sub} \) and \( S_{sat} \) are the entropies of sub-saturated and saturated air. The heat capacities for liquid water and for dry air and vapour are given by \( c_w, c_{pd} \) and \( c_{pv} \), respectively. \( T \) is the temperature, \( p_d \) is the dry air partial pressure, and \( e \) is the vapour pressure. \( L_v \) is the latent heat of vaporisation, and \( R_d \) and \( R_v \) are the gas constants for dry air and water vapour, respectively.

Although the upstream sounding is assumed to be taken in cloud-free conditions, it is necessary to calculate both \( S_{sub} \) and \( S_{sat} \) for the sounding. For each trajectory we can thus calculate \( Q, S_{sub,j} \) and \( S_{sat,j} \) from the upstream sounding values \( p_{0,j}, T_{0,j} \) and \( T_{d0,j} \).

#### Iteration to find thermodynamic conditions along trajectories

Along each of the \( j \) trajectories we define \( i \) grid-points. The altitude, \( z_{i,j} \), is calculated from the trajectory model (see Eqn 12). The corresponding pressure, \( p_{i,j} \), is found from interpolation of the upstream sounding data. Along each of the trajectories \( Q, S_{sub,j} \) and \( S_{sat,j} \) are constant.

Given this pressure, \( p_{i,j} \), and the trajectory values of total water and entropy, we calculate the temperature, vapour mixing ratio and liquid water mixing ratio in the following way. In order to determine if the air at grid-points \((i, j)\) is saturated or not, it is necessary to follow three steps.

(a) Assume saturated air at pressure \( p_{i,j} \). Calculate the saturated air entropy. (b) Compare the thus calculated entropy to the trajectory entropy, \( S_{sat,j} \). If the entropy exceeds the trajectory entropy, then the parcel is saturated, otherwise the parcel is sub-
saturated. (c) Choose the appropriate entropy definition \( S_{\text{sat},j} \) or \( S_{\text{sub},j} \), use the total water mixing ratio \( Q_j \) and pressure \( p_{i,j} \) to iterate to find \( T_{i,j} \) and \( q_{i,j} \). These three steps will now be elaborated.

(a) Assuming exactly saturated air and no liquid water, we use the definition of vapour mixing ratio to calculate the saturation vapour pressure:

\[
\varepsilon_s = \frac{p_{i,j} Q_j}{Q_j + \varepsilon}
\]

We can then calculate the dry air partial pressure:

\[
p_d = p_{i,j} - \varepsilon_s
\]

Using the Clausius-Clapeyron equation we can calculate the dew-point temperature, \( T_d \), corresponding to \( \varepsilon_s \). In practice we will use Bolton’s (1980) empirical relationship between dew-point temperature and saturated vapour mixing ratio:

\[
T_d = \frac{243.5 \ln \varepsilon_s - 440.8}{19.48 - \ln \varepsilon_s}
\]

At this stage the saturated entropy can be calculated from:

\[
S_{\text{sat}} = (c_{pd} + Q_j c_{wp}) \ln T - R_d \ln p_d + Q_j \frac{L_v}{T_d}
\]

(b) We now evaluate whether the air parcel, \((i, j)\), is saturated. If \( S_{\text{sat}} \) from Eqn 19 is equal to or larger than \( S_{\text{Sat},j} \), then the parcel is saturated. Otherwise the parcel is clear.

(c) If the parcel is sub-saturated, then we can use Eqn 14 to solve for \( T \):

\[
T_{i,j} = \exp \left\{ \frac{S_{\text{sub},j} + R_d \ln p_d + Q_j R_v \ln \varepsilon_s}{c_{pd} + Q_j c_{pv}} \right\}
\]

Given that the parcel is sub-saturated, it follows that it does not contain any liquid water.

If the parcel is saturated, then it is necessary to iterate to find the values of \( T_{i,j} \) and \( q_{i,j} \). First a guess of temperature, \( T_i \), is made. From this temperature the saturated vapour pressure, \( \varepsilon_s \), is found. This allows for the dry air partial pressure, \( p_d \), and for the vapour mixing ratio, \( q_s \), to be calculated. Using these values the entropy, \( S_{\text{sat}} \), can be calculated from Eqn 15. If this entropy is nearly identical to the trajectory entropy, \( S_{\text{Sat},j} \), then the temperature \( T = T_{i,j} \). The liquid water mixing ratio is given by \( q_{i,j} = Q_j - q_v \).

Once the entire domain of 101 \( i \)-values and 64 \( j \)-values are calculated, then the fields of \( T_{i,j} \) and \( q_{i,j} \) can be contoured.

**Example of a post-frontal airflow**

For the present study we have used sounding data from Blue Rock, which is located 30 km southwest of Baw Baw Plateau. We chose the sounding data of 6 August 1988 for the presentation.

Figure 3 shows the Skew T-Log p diagram and wind vectors of the sounding data. The sounding is fairly stable, and the air is very moist between the surface and 850 hPa. There is some backing of the wind with altitude from west-southwest to south-southwest from the surface to high levels. The wind components perpendicular to the mountain were calculated at all levels up to 5 km and from these the average wind speed was calculated. The perpendicular wind direction is 230° and the average wind speed is 18.7 m s\(^{-1}\). The lowest 5 km sounding data were also used for calculating an average Scorer parameter.

**Fig. 3** A Skew T-Log p diagram from the Blue Rock omegasonde released at 12:00 noon, 6 August 1988. Pressure is in hPa and temperature is in °C. The right box shows observed wind vectors, predominantly a southwestely wind.

The altitude and pressure for the 64 streamline levels were calculated based on the sounding data. For each trajectory, the vertical displacement \( \eta \) was first calculated for each horizontal grid spaced by 1 km using Eqn 12. Adding to its initial upstream height from the ground, we get the trajectory height at each grid-point. Liquid water content was calculated from the liquid water mixing ratio described in the previous section.

Figure 4 shows the airflow trajectories and distribution of liquid water content. Near the surface the streamlines show the familiar gentle slope up over the mountain, followed by a rapid descent immediately on the lee side of the mountain. Phase lines of the wave crests slope upstream, as found by Queney (1947). At an altitude of about 5.5 km (half the vertical wavelength) the airflow is symmetric up and downstream, but inverted relative to the surface topography.
The cloud location is determined by the airflow trajectories and by the high moisture content low in the atmosphere. The cloud is asymmetric with respect to the mountain; it is much more extensive on the upstream side of the mountain than on the downstream side. The maximum cloud top is likewise displaced upstream from the mountain top. However, the maximum liquid water content occurs right over the top and is associated with the low-level air. The maximum cloud top reaches about 3 km, but the cloud top (as indicated by the 0.0 liquid water contour) decreases rapidly upstream.

Figure 5 shows the distribution of temperature and liquid water content. The main feature to focus on is the cloud-top temperature upwind of the mountain. As explained in the introduction, cloud seeding would be performed with dry ice if the cloud temperature is between $-2^\circ$ and $-7^\circ$C. The $-8^\circ$C contour intersects the cloud top about 30 km upwind of the mountain top. Given a growth time for an ice particle from nucleation to fallout of about half an hour (1800 s) and a wind speed of about 19 m s$^{-1}$, it can be calculated that seeding should take place about 35 km upstream of the target area. The target area is the Thomson Catchment, which extends from the mountain top to 5–15 km downwind of the mountain. Figure 5 shows cloud top with $-8^\circ$C at 2 km height and 30 km upstream of the mountain top. In summary, the calculated conditions for this sounding (wind, temperature and moisture) suggest that the conditions for cloud seeding may be fulfilled for this particular case.

Queney's model may also be used to calculate horizontal wind speeds as the air approaches and moves over the mountain. The starting point for this calculation is that the mass flux is constant between streamlines. Using average density and altitude of two adjacent airstreams at the initial location, 50 km west of the mountain's peak, the mass flux was calculated at the averaged altitude. Horizontal wind was calculated at each grid-point using the given mass flux, the local distance between streamlines and the local density. At altitudes lower than 5 km, as the trajectories suggest in Fig. 4, the wind decreases as the air moves towards the mountain, but the wind increases up to as much as 75 m s$^{-1}$ around the peak of the mountain where the streamlines are close. This velocity seems unreasonably large; it is most likely due to the linear nature of the model.

**Instability due to convection**

Queney's model cannot allow for convection due to the two-dimensional analytical formulation. However, it can be used to calculate a 'base' flow field, which may have regions of instability due to the differential cooling rates of dry and moist adiabatic lifting.

We will examine the calculated thermodynamic fields for two types of moist instability: (a) convective instability, in which the vertical virtual temperature gradient exceeds the moist adiabatic temperature gradient ($\Gamma > \Gamma_{wa}$), and (b) conditional instability, in which the virtual temperature of a cloudy parcel exceeds the environmental (far upstream) virtual temperature. For the calculation of buoyancies (as expressed by virtual temperature) we have included the liquid water loading in cloudy air parcels.

We evaluate these two instabilities for each point along all trajectories. If instability occurs, then we take this location to be the base of a convectively ascending element. We calculate its
properties \( (T, q_v, \text{ and } q_l) \) based on its entropy and total water mixing ratio at subsequent higher locations. If it is convectively unstable, then the convective element is assumed to continue ascending. Once this condition is no longer fulfilled, the parcel has reached its level of neutral buoyancy, and this is assumed to be the top of the convective region.

By equating the ‘positive’ and ‘negative’ area of the sounding, we find that convective overshoot may occur. The depth of the negative area is approximately 50 per cent of the depth of the positive area. The convective overshoot is more predominant immediately upstream of the mountain peak.

We find that (a) convective instability is very extensive for the example under consideration, whereas (b) conditional instability is virtually absent. The vertical bars in Fig. 6 show the base and top of the convectively unstable region. It can be seen that almost the entire cloud is convectively unstable.

**Fig. 6** Cloud liquid water content \((g \text{ m}^{-3})\) and regions of convective instability. The vertical bars represent the extent of convective instability.

The degree of the convective activity is examined by calculating the virtual temperature difference between air parcels ascending adiabatically from any grid-point and their surroundings at all higher levels. The average virtual temperature difference for the entire convectively unstable region is 0.2°C with a maximum value of 0.7°C 5 km upstream of the mountain top. Since this average virtual temperature difference is small, we assume that it is likely that the convective activity has only a minor influence on the general airflow.

**Discussion**

In the present study we have implemented a very simple model for calculating two-dimensional airflow over a mountain. We have extended this model to include the calculation of thermodynamic fields and convective instability in the cloudy air. The simplicity is necessary in order to be able to execute the model quickly on the MMBW field office computer system, and thereby be of value for the planning of surveillance flights. The simplicity of the present model is apparent in a number of different ways.

We have employed a simplified topography by assuming a bell-shaped mountain. The actual topography of Baw Baw Plateau is only approximately bell-shaped from the mountain top and upstream; downstream the topography is very complex with several irregular ridges and valleys. The choice of a bell shape for this situation may be appropriate given that the main interest is in the cloud formations upstream and right over Baw Baw Plateau; however, there are certainly some reservations regarding the possible upstream flow field effects caused by the irregular downstream topography. It is an area we intend to investigate in more detail in future studies.

Queney’s (1947) analytical model is strictly speaking only valid for an infinitely deep fluid in which no latent heat effects are present. In reality the atmosphere may be better described by a fluid of finite depth (Long 1953), and there are certainly latent heat effects associated with the cloud formation as the air moves up over Baw Baw Plateau. Fortunately, as found by Fraser et al. (1973), for stable conditions it appears that the latent heat effects are secondary to the barrier effect of the mountain.

Some two-dimensional models predict the occurrence of blocking, i.e., closed circulations at low levels upstream of the mountain. This is not the case in the present model. Blocking is normally associated with infinitely long mountains, which the present model is intended to represent, but in practice it is likely that the finite length of Baw Baw Plateau allows for some limited flow around the mountain. This flow would tend to reduce the blocking effect.

Airflow around the mountain will tend to reduce the vertical displacement of the streamlines. Abbs and Jensen (personal communication) have used a three-dimensional numeric model to calculate the airflow around Mt Baw Baw for a single case study. They found only minor airflow around the mountain near the surface. Even so, the implication is that the present model probably overestimates the cloud-top height, liquid water content, etc., due to its two-dimensional basis.

We have used the present model to calculate temperature and cloud-water fields. This has been done using entropy and total water as the conserved parameters. Entropy and total water are
both affected by diabatic processes such as mixing and fallout of precipitation. Entropy is furthermore affected by radiative energy exchanges. None of these are modelled in the present study. The liquid water content calculated by the model is probably overestimated due to the omission of dry air entrainment across the cloud/clear air border and the subsequent evaporation during mixing of cloudy and clear air parcels. A reduction in liquid water content can also occur as a result of the growth of ice particles. It is therefore almost certain that the present model overestimates the liquid water content in the clouds, and to a lesser extent also the cloud amount and continuity.

Convection cannot be explicitly calculated in the present model. It is, however, possible to do buoyancy comparisons to see if convection would occur given the calculated dynamic and thermodynamic fields. Strictly speaking the model is not valid for the example shown in this paper; convection does occur and therefore the model assumptions break down. Since the virtual temperature difference was small, as mentioned in the previous section, it is likely that the convective activity has only a minor influence on the general airflow. In practice it may be possible to say that the model calculates a basic airflow, for which we can analyse the regions of convection and get a very rough idea of their extent. We note that the cloudy regions of convection are likely to have higher liquid water contents than their more quiescent surrounding regions.

There are many more sophisticated and complex two-dimensional models which describe the airflow over mountains. These may give more accurate answers, but operational considerations have led us to choose the present simple model. The model is intended to be used for calculating cloud fields over Baw Baw Plateau in a nowcasting mode. For this reason it is important to have quick access to the Laverton sounding data, but even so there may be differences between the air mass observed at Laverton and that at Baw Baw Plateau. The rough predictions of the cloud liquid water content, cloud extent, and cloud-top temperature model will be compared against observations taken from the MMBW cloud seeding aircraft in the future.

In a future study we plan to combine this airflow and cloud field model with trajectory calculations of growing precipitation particles.

### Appendix

#### Symbol list

- $\alpha$: Mountain half-width.
- $c^2$: The speed of sound for dry adiabatic motion.
- $c_{pd}$: Heat capacity of dry air, water vapour and liquid water.
- $e$, $e_s$: Vapour pressure and saturated vapour pressure.
- $g$: Gravity acceleration.
- $h$: Mountain height.
- $h_m$: Maximum mountain height.
- $i$, $j$: Horizontal and vertical grid-point number.
- $k$: Wave number.
- $l$: Scorer parameter.
- $p$, $p_d$: Pressure and dry air partial pressure.
- $q_h$, $q_v$: Liquid water and water vapour mixing ratio.
- $x$: Horizontal distance.
- $u$: Horizontal velocity.
- $w$: Vertical velocity.
- $z$: Altitude.
- $L_v$: Latent heat of vaporisation.
- $Q$: Total water mixing ratio.
- $R_d$, $R_v$: Gas constant for dry air and water vapour.
- $\bar{S}$: Heterogeneity.
- $S_{\text{sat}}$, $S_{\text{sub}}$: Specific entropy of saturated and subsaturated air.
- $T$, $T_d$: Temperature and dew-point temperature.
- $U$: Horizontal velocity upstream of the mountain.
- $\beta$: Stability parameter.
- $\gamma$: Ratio of specific heat capacities at constant pressure and volume.
- $\eta$: Vertical displacement of a streamline.
- $\rho$: Air density.
- $\theta$: Potential temperature.
- $\Gamma$, $\Gamma_{mv}$: Lapse rate and moist, virtual lapse rate.

### References


IND/N85/MMBW/1. Available from CSIRO Division of Radiophysics.
