The detection and analysis of a gravity wave observed over Casey in East Antarctica using radiosonde data

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Casey station in East Antarctica is prone to storms of hurricane-force easterly winds in situations where deep cyclones track eastward just off the coast of Casey. However, the strong wind experienced at Casey is more than simply synoptically driven and appears to be sensitive to atmospheric stability and strongly influenced by the flow around Law Dome, situated to the southeast. A strong wind event on 12 January 1993 is analysed in some detail using radiosonde data and the flow at Casey argued to be in response to a transient orographic gravity wave generated by the flow over Law Dome.

Introduction

Weather forecasters based at Casey Station over the summer months have long recognised the significant effect Law Dome plays in modifying the wind regime in the Casey locale. Casey is situated on the Antarctic coast at approximately 66°S, 110°E (Fig. 1) and lies in the lee of Law Dome, a large regularly shaped ice dome rising to a centre with a height of 1395 m some 120 km to the southeast of Casey. Strong wind events at Casey Station occur infrequently over the summer months but may be extreme in their magnitude with mean wind speeds of 30 to 40 m s⁻¹ not uncommon, and peak gusts measured in excess of 65 m s⁻¹ in one storm during March of 1992. The summertime strong wind events are associated with cyclones in the polar trough moving southeastward towards the Antarctic coast and sweeping eastward around the coast to the north of Casey. The more extreme events have occurred when cyclones deepen to below 960 hPa and track south of 60°S at Casey longitudes. The unusual aspect of the Casey area becomes apparent when analysing wind profiles prior to the onset of gale-force surface wind at the Casey Antarctic Meteorological Centre (AMC). On many occasions the gradient (1000 m) wind from the Casey radiosonde flight may be strong, consistent with the pressure gradient generated by the approaching cyclone, yet at the surface (10 m) be less than 5 m s⁻¹ and on some occasions westerly in direction and oppos-

Fig. 1 Map of the Casey and Law Dome area in East Antarctica.

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ing the gradient. On occasions when the strong wind does reach the surface at Casey, the wind speed is typically greater than would be expected from the gradient associated with the cyclone and significantly stronger than the wind measured at Law Dome, some 1395 m above mean sea level and inland of Casey.

The ability to forecast the onset and strength of strong surface wind events is of prime importance over the summer months where research and construction programs rely on helicopter and small boat operations to successfully carry out summer programs. The severity of the wind events coupled with the isolation of the base necessitate reliable and timely warnings to station personnel.

Wilson (1992), in a study of strong wind events in the Casey area, provided an analysis of wind data from around Casey, analysed output from several different numerical models and reviewed towing tank experiments on flow over a regularly shaped hill. The conclusions drawn from the study were that Casey is in an area little affected by katabatic flow, due to the influence of Law Dome, and that although the Casey strong wind events are synoptically driven the wind speed and direction were not representative of the synoptic situation. The most probable cause of the onset of very strong wind was the formation of a hydraulic jump upstream of Casey in the strengthening gradient wind as the cyclone approached. As the gradient wind increased further, the hydraulic jump moved downstream over the Casey area with the resulting significant increase in local wind speed. The analysis provided by Wilson (1992), and the description of the onset of strong wind at Casey, appears to follow closely the observations made by the author over the two years spent in the area, but the conclusion that a hydraulic jump is acting in the area appears not to be well founded. With the approaching strong wind, drift tails of snow begin to appear on the moraine line, inland and upslope from Casey, prior to the onset of gales at the station. These plumes of drifting snow have been described by Wilson (1992) and other meteorologists stationed at Casey and are a very good indicator of impending bad weather. However, the drifting snow is in the form of very distinct plumes that move across the moraine line as single entities. Such phenomena appear to be associated with strong wind separating from the surface on the steep slopes of the moraine rather than with a hydraulic jump where a distinct ‘wall’ of blowing snow ought to be observed moving up and down the moraine. The author has observed only one such jump event at Casey over a two-year period, whereas the

Fig. 2 Surface analysis over East Antarctica at 0000 UTC on 12 January 1993.
plumes of snow associated with the onset of very strong wind have occurred on a very frequent basis. It is the contention of this paper that the onset of strong wind at Casey is associated with a topographically induced internal atmospheric gravity wave generated in the flow over Law Dome. An analysis is presented of the 12-hourly high resolution radiosonde data following closely the work of Shutt et al. (1988) in their study of a large amplitude gravity wave detected using radiosonde data from Shanwell in Scotland. The strong wind event presented here was not particularly severe, by Antarctic standards, but the analysis provides a useful model of the flow characteristics around Law Dome and demonstrates the effectiveness of routine operational radiosonde data in the diagnosis of atmospheric stability and turbulence.

It is also the intention of this paper to suggest that through a regular analysis of radiosonde profiles of temperature, stability and wind a more thorough understanding of the three-dimensional flow may be achieved and a useful source of data compiled from which to gauge the effectiveness of numerical modelling studies, such as those undertaken by Wilson (1992). Regular analysis of the radiosonde data may also be useful in building a climatology of the intensity and frequency of gravity wave activity in the Casey area. Studies dating back to Corby (1957), Lalas and Einaudi (1980) and, more recently, Shutt et al. (1988) have emphasised the necessity for a detailed analysis of the broadscale frequency, intensity and geographical distribution of forced gravity wave events. Lilly (1972) provided observational data to contend that orographically induced gravity waves in the upper troposphere and lower stratosphere may be of sufficient magnitude and extent to have a significant effect on the larger scale flow. McFarlane (1987) further emphasised that the redistribution of momentum through breaking gravity waves has an influence on the zonal momentum budget, and hemispheric circulation that needs to be incorporated into gravity wave drag parametrisation schemes and remains a major field of research in numerical modelling.

Fig. 3 Wind speed and direction from the Casey 0000 UTC radiosonde on 12 January 1993.
Meteorological observations

On 12 January 1993, a gale-force east to northeasterly wind dominated the Casey weather. The surface synoptic pattern at 0000 UTC (Fig. 2), consisting of a mean sea-level pressure analysis north of 70°S and a streamline analysis south of 70°S, was dominated by a deep low pressure system centred near 60°S 109°E and with a central pressure of 956 hPa. The east-northeasterly wind generated to the south of the low pressure system was close to 30 m s⁻¹ in the lowest few hundred metres above Casey (Fig. 3) with large fluctuations in speed throughout the troposphere. The cyclonic flow was still evident at 500 hPa (Fig. 4), although the strength of the system was weaker, with the easterly wind having abated to near 18 m s⁻¹ (Fig. 3). The surface synoptic situation at 1200 UTC had changed very little with the low pressure system quasistationary near 60°S 110°E but with the central pressure having risen to 962 hPa and the 500 hPa (5135 m) flow strengthened to around 25 m s⁻¹ (Fig. 5). It should be noted that due to the sparsity of data in the Casey region, the surface and 500 hPa synoptic flow has been inferred from the Casey radiosonde data and any changes will be a response to both the synoptic evolution of the systems and changes resulting from any gravity waves.

Coincident with the storm-force easterly wind was a spectacular formation of altostratus lenticularis (cloud base around 2000 m), with underlying rotor cloud around 1000 m (Fig. 6), visible over and to the west of the Casey Station. Accompanying the cloud formation were significant tropospheric fluctuations in the ascent rate profile of the radiosonde (Fig. 7), released at 0000 UTC from the Casey AMC. The fluctuations had a maximum amplitude of approximately 3 m s⁻¹, measured in the upper troposphere but attenuated sharply at the tropopause (10127 m or 238 hPa), with no significant signal evident in the stratosphere. The error in radiosonde-derived vertical velocity was estimated at near five per cent at the height of the maximum fluctuation (~9000 m) which compares favourably with studies by Corby (1957) and the error analysis of Lalas and Einaudi (1980) in their discussion of the ability of rawinsonde data to adequately distinguish velocity fluctuations attributable to internal atmospheric waves. The mean ascent rate of the radiosonde was 5.38 m s⁻¹ resulting in a measurement error of approximately 0.27 m s⁻¹, establishing the peak fluctuation of near 3 m s⁻¹ as a significant feature of the velocity profile.

Fig. 4  500 hPa analysis over East Antarctica at 0000 UTC on 12 January 1993.
Fig. 5  Wind speed and direction from the Casey 1200 UTC radiosonde on 12 January 1993.

The high resolution image from the NOAA-12 satellite at 2120 UTC on 11 January 1993 (Fig. 8) showed banding within the altostratus evident on the southwest edge of Casey, aligned orthogonally to the prevailing east-northeastly wind. The image, from the band 2 near-infrared (0.725 - 1.10 μm) sensor, had a subsatellite resolution of 1 km with measurements from the image giving an estimated horizontal wavelength for the cloud banding varying between 8 and 12 km. Inspection of the band 4 thermal infrared image from the same satellite pass gave estimates for the cloud-top temperatures of between -47°C and -53°C, placing the cloud tops between 8500 m and 9150 m (307 to 277 hPa) with a cloud depth of approximately 7000 m.

The profiles of temperature and dew-point taken at 0000 UTC (Fig. 9) indicated a very moist atmosphere with an isothermal stable layer near the surface, coincident with the very strong surface wind shear evident from Fig. 3. Values of the Richardson number (Eqn 1) are also displayed in Fig. 9 where the long marks indicate layers in which the value is less than one and the short marks indicate layers in which the value is less than one-quarter. In the layer around 850 hPa (1100 m) the values were less than one quarter and coincident with the rotor cloud under the lenticularis in Fig. 6.

\[ \text{Ri} = \frac{N^2}{(\frac{dU}{dz})^2} \]  

...1

where \( U \) is the horizontal wind speed and \( z \) is height. Also displayed on the far right-hand side of Fig. 9 is a graph of the Brünt Väisälä frequency, \( N^2 \) (Eqn 2).

\[ N^2 = \frac{g}{\theta_0} \frac{d\theta}{dz} \]  

...2

where \( g \) is the gravitational acceleration, \( \theta \) is potential temperature and \( \theta_0 \) is the layer mean potential temperature. The scale of the graph is indicated by the maximum value printed along the side of the graph with the zero line running down to the right of the Richardson number data. The atmospheric layer just below 250 hPa (9810 m) was notable for values of the square of the Brünt Väisälä frequency, \( N^2 \) being less than zero and absolutely unstable. In comparison, the layer above the
tropopause (233 hPa or 10243 m) was very stable with a strong temperature inversion from 233 hPa to 220 hPa and large values of the Brünt Väisälä frequency, $N$, peaking at near 0.039 s$^{-1}$. By 1200 UTC the atmosphere had changed substantially with the stability of the layer below 250 hPa (9780 m), originally absolutely unstable, increasing markedly (Fig. 10) with the Brünt Väisälä frequency rising to a maximum of 0.033 s$^{-1}$ at 249 hPa (9810 m). Fluctuations in temperature and dew-point, of the order of 5 K, had formed in the stratosphere and coincided with an oscillatory signal in the radiosonde ascent rate profile (Fig. 11) between 6000 m and 13000 m, (442 to 153 hPa) where the velocity fluctuations had reached a peak amplitude of 1.3 m s$^{-1}$ at 10000m (241 hPa). The wave signal in the troposphere was markedly different and dominated by a large amplitude (10.9 m s$^{-1}$) peak in vertical motion near 2000 m (760 hPa).

The theoretical discussion

The wave-like signal apparent in both the vertical velocity profile (Fig. 7) and the horizontal wind profile (Fig. 3) at 0000 UTC, coupled with the formation of altostratus lenticularis, suggested an internal forced gravity wave existed on the day of interest; with cloud banding in the high resolution satellite image further supporting the conjecture. To further test the existence of a gravity wave a two-dimensional, non-linear model was compared with the data from the radiosonde released from Casey.

The following analysis is based on the work of Shutt et al. (1988) and Lilly and Klemp (1979) in the analysis of two-dimensional flow over a sinusoidal lower boundary. If steady two-dimensional flow is considered then the two-dimensional inviscid anelastic equations, in perturbation form, may be reduced to

$$\nabla^2 \psi' + \frac{N^2}{\alpha^2} \psi' = 0 \quad \ldots 3$$

where the stream function $\psi$ is defined by $(u, w) = (\partial \psi/\partial z, -\partial \psi/\partial x)$ and $\psi = \psi_0 + \psi'$ with the upstream flow defined by $\psi_0(z) = Uz$. Equation 3 is a special case of the Taylor-Goldstein equation and known as Long's Equation (Shutt et al. 1988). A solution to Eqn 3, without reference to boundary conditions is...
\[
\psi' = -aU \sin(kx + mz) \quad \ldots 4
\]

where
\[
m^2 = \frac{N^2}{U^2} - k^2 \quad \ldots 5
\]

Fig. 7 Ascent rate profile of the radiosonde released from Casey at 0000 UTC on 12 January 1993.

The solution represents an upward propagating wave with horizontal wave number \(k\) and vertical wave number \(m\). The boundary condition for the problem has been applied at \(z = \delta\) and not \(z = 0\), so despite Eqn 4 being linear the solution is non-linear and applicable to finite amplitude waves. It follows from Eqn 4 and the definition of the stream function that the flow is defined by
\[
u = U[1 - am \cos(kx + mz)] \quad \ldots 6
\]
and
\[
w = Uak \cos(kx + mz) \quad \ldots 7
\]

Lilly and Klemp (1979) have developed an iterative method for obtaining the non-linear solutions starting from the linear Eqns 6 and 7. Figure 12 shows the streamline pattern for uniform flow over a corrugated lower boundary and highlights the track followed by a balloon released into the model atmosphere from the surface. Over-turning or stagnation in the flow occurs where \(u = 0\) and \(dwdz = 0\), where \(b\) is the buoyancy. The static stability of the atmosphere is governed by Eqn 8 and from the analysis of Shutt et al. (1988) the static stability measured by the radiosonde can be calculated
\[
\frac{db}{dz} = N^2[1 - am \cos(kx + mz)] \quad \ldots 8
\]
using the parametric equations derived from Eqns 6 and 7 and given by
\[
x = Ut - Uam \left[ \frac{\sin(\chi t)}{\chi} \right] \quad \ldots 9
\]
and
\[
z = Wt - Uak \left[ \frac{\sin(\chi t)}{\chi} \right] \quad \ldots 10
\]
where \(\chi = Uk + Wm\) giving a measured static stability of
\[
\left. \frac{db}{dz} \right|_* = (\frac{w}{w' + W}) \frac{db}{dz} \quad \ldots 11
\]

As may be seen, the actual stability, measured in the vertical (a Eulerian profile), is modified by a factor \(W/(w' + W)\) as the radiosonde moves through the forced gravity wave. For regions where the wave-induced motion is upward the apparent stability will be less than the real stability and in regions where the motion is downward the static stability will be greater, provided the magnitude of the downmotion does not exceed the radiosonde ascent rate, in which case the apparent stability becomes negative and absolutely unstable. Profiles of the square of the Brunt Väisälä frequency, \(N^2\) and the Richardson number, \(Ri\), in Figs 9 and 10 have been corrected for radiosonde vertical motions using Eqn 11. The \(w' + W\) term was the measured radiosonde velocity at a given height, calculated using second order, centred, finite differencing of the radiosonde height data.
Fig. 9  Aerological diagram from the Casey 0000 UTC radiosonde on 12 January 1993.

Fig. 10  Aerological diagram from the Casey 1200 UTC radiosonde on 12 January 1993.
From an analysis of the radiosonde ascent rate profile (Fig. 7) the apparent vertical wavelength of the disturbance was (2403 ± 119) m in the upper troposphere, with an amplitude maximum of (3.05 ± 0.15) m s\(^{-1}\) at 8550 m (304 hPa). The horizontal wind structure (Fig. 3) gave a somewhat larger apparent vertical wavelength of (3142 ± 156) m with a maximum amplitude of (8.2 ± 0.41) m s\(^{-1}\) at 8550 m. The depth-averaged value of \( N \) through the troposphere was near 0.013 s\(^{-1}\), and \( U \) (18.9 ± 0.95) m s\(^{-1}\), with the average ascent rate of the radiosonde over the entire flight measured at 5.38 m s\(^{-1}\).

From Eqn 10 the apparent vertical wavelength is \( 2\pi W/\chi \) giving an apparent vertical wave number of

\[
m_\ast = \frac{Uk}{W} + m \quad \ldots 12
\]

If the wave in question was stationary, and satisfied the hydrostatic long wave approximation, then \( k \) could be considered sufficiently small in Eqn 12 such that \( m_\ast = m \) and, from Eqn 5, \( m_\ast = N/U \). Substituting the above estimates for \( N \) and \( U \), from the Casey data, gives \( m_\ast = 0.000688 \) m\(^{-1}\) and an apparent vertical wavelength of 9134 m which is substantially larger than either of the measured vertical wavelengths. So if the wave is assumed to be stationary, then \( k \) cannot be considered small and horizontal phase variations need to be included. With the dispersion relation, \( m_\ast^2 = N^2/U^2 - k^2 \) the transformation \( k = (N/U) \sin(\phi) \) and \( m = (N/U) \cos(\phi) \) can be made where \( \phi \) is the zenith angle of the wave number vector, reducing Eqn 12 to

\[
\phi = \arcsin \left( \frac{m_\ast W U}{N\Delta} \right) - \arctan \left( \frac{W}{U} \right) \quad \ldots 13
\]

where \( \Delta = \sqrt{U^2 + W^2} \). Equation 13, in conjunction with observed atmospheric values of \( m_\ast, N, U \) and \( W \), can be used to calculate \( \phi \) from which \( k \) and \( m \) may be inferred using Eqn 14.

\[
k = \frac{N}{U} \sin(\phi) \quad \text{and} \quad m = \frac{N}{U} \cos(\phi) \quad \ldots 14
\]

Table 1 details the calculated wavelengths for differing flow speeds using the depth-averaged values of \( W \) and \( N \) cited above and an apparent vertical wavelength of (2773 ± 194) m, the mean of the wavelengths calculated from the horizontal and vertical velocity profiles.

### Table 1. Horizontal and vertical wavelengths for different wind speeds, derived from Eqn 14.

<table>
<thead>
<tr>
<th>( U ) (m s(^{-1}))</th>
<th>( \lambda_x ) (m)</th>
<th>( \lambda_z ) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8</td>
<td>13099 ± 916</td>
<td>4047 ± 283</td>
</tr>
<tr>
<td>9</td>
<td>11259 ± 788</td>
<td>4716 ± 330</td>
</tr>
<tr>
<td>10</td>
<td>10507 ± 735</td>
<td>5443 ± 381</td>
</tr>
<tr>
<td>15</td>
<td>10786 ± 754</td>
<td>9792 ± 685</td>
</tr>
<tr>
<td>20</td>
<td>12649 ± 885</td>
<td>14988 ± 1048</td>
</tr>
</tbody>
</table>
From Eqns 6 and 7 the wave-induced oscillations in horizontal and vertical velocity are given by

\[ u' = Ua\cos(kx + mz) \]

\[ w' = Uak\cos(kx + mz) \]

giving the ratio \( lw'/lu' = \lambda_z/\lambda_x \). So the magnitude of the fluctuations in both the horizontal and vertical velocity (8 m s\(^{-1}\) and 3 m s\(^{-1}\) respectively) would imply a horizontal wavelength of between two and three times the vertical wavelength suggesting, from Table 1, a flow speed of near 9 m s\(^{-1}\) and a wavelength of near 12.5 km. The measured mean tropospheric speed near Casey of 18.9 m s\(^{-1}\) is substantially larger than this figure, intimating the possibility of a transient wave with a phase speed near 10 m s\(^{-1}\). The image taken at 1149 UTC on 12 January 1993 by the NOAA-11 polar-orbiting satellite showed distinct banding some 150 km to the west-southwest of Casey (Fig. 13) where the wave had been evident some twelve hours earlier. If both cloud signatures (Figs 8 and 13) were the result of a propagating wave, then the horizontal phase speed would have been at least 4 m s\(^{-1}\), supporting the notion of transience although with some doubt on the magnitude of the phase speed and spatial and temporal variation. The surface synoptic situation appeared to be static over the twelve-hour period between 0000 UTC and 1200 UTC but the atmospheric stability profiles were dynamic with marked changes evident in both the profiles of temperature and the square of the Brunt-Väisälä frequency, \( N^2 \). The dynamic nature of the stability profiles may account for the ephemeral nature of the wave. The unstable layer near 250 hPa (9810 m) was characterised by complex values of the vertical wave number intimating the possibility of trapping within the troposphere. By 1200 UTC the upper troposphere had stabilised considerably allowing vertical propagation of the wave into the stratosphere, with increasing instability in the lower troposphere suppressing wave generation.

Scorer (1949) demonstrated that if the parameter \( \beta^2 \) (Eqn 15)

\[ \beta^2 = \frac{N^2}{U^2} - \frac{1}{U} \frac{d^2U}{dz^2} \]

...15

decreases sufficiently with height, then trapped lee waves may occur. In a two-layer atmosphere in which the lower layer has a vertical wave number \( m_2 \), such that \( m_2^2 > 0 \) and the upper layer a wave number \( m_1 \) such that \( m_1^2 < m_2^2 \) then an upward-propagating wave from the lower layer will be reflected at the boundary, with a transmitted wave propagating through the upper layer. If travelling wave solutions of the form

\[ \Psi = \hat{\Psi}(z) e^{i(kx-\omega t)} \]

...16

are sought, then the stream function \( \hat{\Psi} \) satisfies the Taylor-Goldstein equation (Eqn 17) which is analogous to Long’s equation with \( \beta^2 = N^2/U^2 \).

\[ \frac{d^2\hat{\Psi}}{dz^2} + (\beta^2 - k^2)\hat{\Psi} = 0 \]

...17

Solutions to Eqn 17 take the form

\[ \hat{\Psi} = A e^{ikx-m_2z} , \ m_2<0 \]

...18

\[ \hat{\Psi} = A e^{ikx+m_2z} , \ m_2>0 \]

...19

where Eqn 18 represents a perturbation decaying exponentially with height and Eqn 19 a wave solution. In a two-layer model, where \( m_1^2<0 \) in the upper layer, the transmitted wave decays exponentially with height, as described by Eqn 20, and the lower layer has a solution made up of the sum of an upward-propagating wave and a downward one (Eqn 21).

\[ \psi_1 = ce^{ikx-m_1z} \]

...20

\[ \psi_2 = \beta e^{ikx+m_2z} + \gamma e^{ikx-m_2z} \]

...21

Assuming \( \alpha \) is known and using the requirement that both \( \hat{\Psi} \) and \( \frac{d\hat{\Psi}}{dz} \) must be continuous at the interface between the two layers to conserve density and pressure, then in the lower layer

\[ \psi_2 = \alpha [\cos(m_2z) - \frac{m_1}{m_2} \sin(m_2z)] e^{ikz} \]

...22

The \( x \) and \( z \) dependence in Eqn 22 are decoupled giving a horizontally propagating wave which is trapped in the lower layer and decays exponentially in the upper layer on a scale of \( 1/|m_1| \).

The requirement that \( m_1 \) be less than \( m_2 \) is analogous to the Scorer parameter decreasing with height and negative values of the Scorer parameter imply regions in which \( m_1^2 < 0 \). Figure 14 is the profile of the Scorer
parameter (Eqn 15) from the 0000 UTC flight. A running mean over approximately 1000 m has been applied to the profile in an attempt to reduce the significance of the errors introduced in calculating the second-order derivative of the wind profile. A sharp decrease in the Scorcer parameter existed in the layer around 8000 m (331 hPa), falling to less than zero between 8500 m and 9000 m (307 to 284 hPa) resulting in a complex vertical wave number in the upper layer. The form of the profile was consistent with Scorcer's analysis of trapping, with a lee wave likely to have existed in the troposphere, trapped below the layer near 9000 m (284 hPa), a height consistent with the observed cloud tops between 8500 m and 9150 m (307 to 277 hPa).

Equation 22 may be used to model the flow over Law Dome through tracking an imaginary radiosonde released into the model atmosphere. Figure 15 is the vertical velocity profile a radiosonde would experience if released into a two-layer atmosphere with a boundary at 9500 m. Figure 14 was used to estimate depth-averaged values of $\beta$ with $3.58 \times 10^{-7} \text{m}^{-2}$ employed in the lower atmosphere and $-2.01 \times 10^{-6} \text{m}^2$ in the top layer. The horizontal velocity $U$, was set to 19 m s$^{-1}$, the mean wind speed in the lower layer, and a mean ascent rate of 6.0 m s$^{-1}$ used for the radiosonde, and also measured from the lower layer. A horizontal wavelength of 12 500 m was assumed as a compromise of the estimated values from Table 2 and the measured wavelengths from the NOAA-12 image (Fig. 8). The value of $\alpha = 1000 \text{ m}^2 \text{s}^{-1}$ in Eqn 22, was chosen to provide a signal amplitude similar to that measured in the real ascent. The ascent rate profile from the model is remarkably similar to the profile measured at 0000 UTC on 12 January (Fig. 7), however, it should be stressed that the solution described by Eqn 22 is linear and the amplitude of the bottom topography assumed small in the derivation of the solution to the Taylor Goldstein equation. The value chosen for $\alpha$ in the calculation of Fig. 15 compromises the linearity of the solution but nonetheless, the solution does model the form of Fig. 7 with a reasonable degree of accuracy furthering the premise that a trapped lee wave was the cause of the signal visible in the ascent rate profile at 0000 UTC.

Discussion

The evidence presented in the previous section supports the notion that the disturbance observed on 12 January from the Casey radiosonde ascent was caused by a transient orographic gravity wave with a horizontal wavelength near 12.5 km. By 1200 UTC stability conditions had altered sufficiently to allow propagation of
the wave into the stratosphere (Fig. 11) although a low-level wave train was still visible some distance downstream from Casey (Fig. 13). The deviations of $u' = 8 \text{ m s}^{-1}$ and $w' = 3 \text{ m s}^{-1}$, measured at 8550 m (304 hPa), were quite large and indicated the likely existence of large fluxes in momentum associated with the wave. If $u'$ and $w'$ were perfectly correlated then an estimate of the mean momentum flux would be given by $\overline{u'w'} / 2$, resulting in a flux of 5.9 Nm$^{-2}$ at 0000 UTC on the day under study. If the vertical profiles are split into mean and wave components by applying a running mean over an averaging length large in comparison to the apparent vertical wavelength, then profiles of vertical flux of horizontal velocity may be estimated from Eqn 23.

$$\overline{u'w'} = \overline{uw} - \overline{u^2} \overline{w}$$

$$\overline{w'w'} = \overline{ww} - \overline{w^2} \overline{w}$$

Figure 16 shows the profiles of zonal and meridional fluxes of horizontal momentum with a maximum downward momentum flux of 2.15 Nm$^{-2}$ at 9800 m. The averaging length was set to 5973 m, just over twice the apparent vertical wavelength. The magnitude of the momentum flux was somewhat smaller than when perfect correlation was assumed. Aircraft measurements of momentum flux associated with orographic waves over Scotland (Brown 1983) found values of 0.4 Nm$^{-1}$, somewhat smaller than measured here. However, the magnitude of the flux is similar to that measured by Shutts et al. (1988) from the Shanwell wave event; described as an 'extreme' event by Shutts. Given the height and extent of Law Dome and the high gradient wind speed in the Casey area on 12 January, the magnitude of the momentum flux was probably genuine.

It would appear that as cyclones approach the Casey area causing the easterly flow to strengthen, the high static stability forces the flow to separate over Law Dome. If a sufficient decrease in stability occurs, the flow regime switches to a state similar to that shown in Fig. 12 where the air accelerates down the lee slope of the obstacle with an orographic gravity wave generated within the flow. Such strong downslope wind is often observed over Casey station in comparison with the gradient flow and the wind speed observed at the drill site on Law Dome (Fig. 1). On 12 January at 1200 UTC the easterly wind measured at Casey was 30 m s$^{-1}$ yet the flow at the drill site was measured at only 20 m s$^{-1}$.

The trapped lee wave described here was of a wavelength sufficient for the radiosonde sampling rate (a measurement every 10 seconds) to adequately capture in detail, however, internal waves of a finer structure are also a feature of the flow in the Casey area. The wave cloud illustrated in Fig. 17 also occurred on 12 January and was observed at a height of several hundred metres to the north of Casey and believed to be due to Kelvin-Helmholtz instability. The wave appeared within an atmospheric layer that was isothermal and very stable (Fig. 9) with a wind profile (Fig. 4) that showed a strong increase in velocity from the surface to 400 m, through the stable layer. The observed conditions described are those under which Kelvin-Helmholtz instability may form. From Fig. 9 it is also evident that the Richardson number was less than one-quarter in the layer indicating that shear instability was a distinct probability. So despite the wave being too fine for the radiosonde data to 'measure', the stability and wind profiles obtained from the radiosonde flight appear capable of establishing atmospheric layers in which Kelvin-Helmholtz instability may arise. The instability may also be an important factor in the generation of clear-air turbulence (Scorer 1972) and such layers need to be identified. Figures 9 and 10 both have profiles of the Richardson number (Eqn 2) and indicate where values fall to less than one and also to less than one-quarter. The potential for instability and turbulence is found to occur in layers in which the Richardson number falls below one-quarter, although from geostrophic theory values of the parameter less than one lead to instability. A significant layer in which the parameter was less than one-quarter also existed around 850 hPa (1100 m) at 0000 UTC (Fig. 9) and was coincident with the rotor cloud visible under the altostratus lenticularis (Fig. 6), an area visibly turbulent.
Conclusions

The analysis of operational radiosonde data from Casey, in East Antarctica, has detected a significant wave-like disturbance attributable to an internal atmospheric gravity wave generated in the flow over Law Dome. Errors in radiosonde-derived profiles due to sensor and measurement limitations were not significant in comparison to the large-amplitude periodic fluctuations in profiles of horizontal and vertical velocity. Analysis techniques demonstrated by Shufts et al. (1988) were applied to the 12 January data in an attempt to describe the dynamics producing the wave-like disturbances. Some success was achieved in defining a trapped lee wave as the most likely disturbance producing the observed fluctuations, with the wave train generated as the flow over Law Dome switched from being separated flow to a down-slope wind. It is this switch from separated to down-slope flow that is believed to be the cause of the onset of very strong, super-synoptic wind at Casey.

The probable transient nature of the wave highlighted the inherent problems associated with balloon-based observations of the atmosphere. The radiosondes measure atmospheric parameters as they are advected both horizontally and vertically through the atmosphere making it difficult to differentiate between spatial variations in atmospheric parameters and temporal variations. Also, the Lagrangian profiles produced by the radiosonde may not realistically represent the true vertical structure of the atmosphere, with profiles being modified by fluctuations in vertical velocity as the radiosonde is advected through a wave disturbance. The former problem is an intractable one when dealing with balloon-based observations; other technologies, such as radar vertical profilers and microbarograph arrays, need to be employed. The latter problem of obtaining true vertical profiles from the radiosonde data may be addressed if the wave disturbances are assumed stationary. The technique described by Shufts et al. (1988) and reproduced, in part, in an earlier section, provides a means of modifying stability profiles from which profiles of temperature may be corrected for vertical velocity fluctuations experienced by the radiosonde. The modified profiles have the benefit of providing the meteorolo-
gist with a more genuine profile of the atmosphere from which turbulence and stability may more accurately be described. It is important that meteorologists be aware that typical radiosonde profiles sample the atmosphere over a long horizontal fetch and typically cover distances some four to five times greater than in the vertical.

The wave over Casey produced substantial downward fluxes of horizontal momentum, applying a significant influence on the tropospheric flow. The wave appeared nearly two hundred kilometres downstream some 12 hours after the initial disturbance was observed, highlighting the likely redistribution of momentum that may take place when forced orographic gravity waves are generated. Long-term analysis of radiosonde data may provide a climatology of gravity wave events, providing more realistic estimates of momentum fluxes and gravity wave drag which in turn may be incorporated into parametrisation schemes within numerical models.

Radiosondes continue to provide valuable data and careful analysis may reveal much about the state of the atmosphere and the dynamic processes in play.

References


