Gravity-wave observations using an array of microbarographs in the
Balearic Islands

By S. MONSERRAT1 and A. J. THORPE
Dept. of Meteorology, University of Reading, 2 Earley Gate, Whiteknights, Reading RG6 2AU

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SUMMARY

Time series of surface pressure from an array of microbarographs in the Balearic Islands (western Mediterranean) are used to obtain wavelength, phase speed and direction of propagation of gravity waves on the Balearic Islands. These pressure waves are commonly found during the summer and they force the large sea-level oscillations observed in the inlet of Ciutadella; a phenomenon locally known as 'rissaga'. Four events are described here, each of which has a duration of between 30 and 48 hours. Radiosonde data from Palma de Mallorca are used to relate the waves to the vertical structure of the atmosphere. The various source mechanisms of these waves and also the possibility of the existence of a duct layer trapping the wave energy near the surface are discussed. The waves are found to have no significant variation of phase speed with frequency and so they are basically non-dispersive. This is an indicator pointing either to a local generation of gravity waves by wind-shear instability or to the existence of neutral trapped waves, in the lower troposphere.

1. INTRODUCTION

Between May and September each year, large-amplitude gravity-wave trains are usually detected on the Balearic Islands (Fig. 1). These waves have been found to have an unusually large amplitude during some events (more than 3 mb trough-to-crest amplitude) with a relatively short period (less than one hour) (Ramis and Jansà 1983; Monserrat et al. 1991a). The source mechanism for these atmospheric waves is not yet well understood owing to the lack of data. They are not clearly associated with convective activity, and topographic origin is unlikely because the only significant mountains are far remote in north Africa (some mountains exist in the Balearic Islands but during the events the wind usually blows parallel to the main axis of the range).

The atmospheric gravity waves observed on the Balearic Islands during July 1989 were successfully associated with a phenomenon observed in the inlet of Ciutadella (Minorca) and other bays and inlets of the Balearic Islands. The phenomenon, locally known as 'rissaga', involves large sea-level oscillations in the inlet (up to 3 metres trough-to-crest amplitude) with a period of some 10 minutes (Ramis and Jansà 1983; Tintore et al. 1988; Monserrat et al. 1991b). The normal sea-level oscillations in the port without rissaga are of 5–10 cm amplitude. In the previously referenced work, it is shown that these sea-level oscillations could be produced by resonance between the normal mode of the inlet basin and an atmospheric gravity wave; the inlet would then behave as a damped harmonic oscillator externally forced by the atmospheric gravity wave. Knowledge of the atmospheric gravity wave, specially concerning the source mechanism, will allow a better understanding of the rissaga and, therefore, will help to predict the phenomenon.

In order to give further insight into possible generation mechanisms for the atmospheric waves, this paper considers time series of surface pressure from three microbarographs situated on Mallorca (Balearic Islands) and deployed as a triangle. The pressure records are used to obtain wavelength, phase speed and direction of propagation of the waves by means of the cross-correlation function. Pressure time series from a

1 Permanent address: Dept. de Fisica, Universitat Illes Balears, E-07071 Palma de Mallorca, Spain.
fourth instrument placed at Ciutadella are also available. This instrument is used for relating the atmospheric gravity waves with the rissaga phenomenon.

Other authors have previously calculated wave properties using time series from three or more instruments. Keliker (1975) analysed pressure time series from microbarograph arrays at Boulder (Colorado) and Washington D.C. Gedzelman and Rilling (1978) used an array of four microbarographs to determine speeds, directions and wavelengths of gravity waves at Palisades (N.Y.). Cross-correlation techniques were also applied to several National Weather Service barograph traces in the north central United States by Stobie et al. (1983). Rees and Mobbs (1988) used time series of wind speed and direction for studying gravity waves observed at Halley Base (Antarctica) while Samah (1990) studied gravity waves associated with atmospheric fronts from an array of three microbarographs in the United Kingdom.

Four periods of wave activity in the Balearic Islands during the summer of 1990 are selected, and the wavelength, phase speed and direction of propagation of the waves for every event are calculated. The results are used for discerning the possible generation mechanisms for the atmospheric waves. The possible existence of a wave duct, trapping the wave energy near the surface (Lindzen and Tung 1976), is also discussed.

In section 2, a short description of the instruments and their location, the available data and the selected four periods of wave activity are given. The method of analysis used in this study for obtaining wavelength, phase speed and direction of propagation of the waves is explained in section 3. The results for every event are given in section 4. In section 5 the instrument at Ciutadella is used in order to relate the atmospheric wave with the rissaga phenomenon. An interpretation of the results, analysing the likely source mechanisms and the possibility of a wave duct are given in section 6. Summary and conclusions are presented in section 7.
2. INSTRUMENTS AND OBSERVATIONS

The instrumentation used in this study consists of four microbarographs that are an improved version of the instruments described in Monserrat et al. (1991b). Each barometer consists of an outer insulated box containing a thermostatically-controlled oven, itself insulated with polystyrene, operating at 50°C. The thermostatic control is effective until the outside air temperature exceeds about 40°C. This oven houses an absolute pressure sensor (commercially available) which has been calibrated in the laboratory.

Three of these instruments, which will be referred to as SGR, CLO and AUL, were situated on the main island (Mallorca) to form a triangular array (Fig. 2) and the fourth one (CIU) was situated at Ciutadella (Minorca). Figure 1 shows the location of the instruments in the zone. The instrument SGR was chosen to be at the origin of the coordinate system and the CLO and AUL coordinates are (0.64, 3.60) km and (3.40, 2.60) km respectively; the CIU instrument has the coordinates, in the same reference axis, (92.42, 58.80) km. Time series from the instrument at Ciutadella are not used for doing calculations of wavelength or phase speed with the other three because it is placed relatively far from them, however, it will be used for testing some results when the atmospheric wave maintains its coherence travelling such a large distance, and for relating the gravity waves to the rissgaga phenomenon. Sea-level records from a tide gauge in the harbour of Ciutadella are also available. The vertical structure of the atmosphere simultaneous to the events can be obtained from the radiosonde ascent made every 12 hours from a location near the microbarograph array at Palma de Mallorca.

The data analysed in this study consist of pressure time series, with a sampling interval of 30 seconds, obtained from July 1990 to September 1990. Time series from the tide gauge are only available for August and September. During this period some gravity-wave events with pressure variations greater than 1 mb have been observed. In this work four events have been selected with durations ranging between 30 and 48 hours (Table 1).

![Figure 2. The relative positions of the three microbarographs (CLO, AUL, and SGR) on Mallorca.](image)

3. METHOD OF DATA ANALYSIS

We are interested in calculating wavelengths, phase speed and direction of propagation of the wave for every event. In general, every frequency can have a different
TABLE 1. List of selected gravity-wave events from the measurements during the summer of 1990

<table>
<thead>
<tr>
<th>Event number</th>
<th>Period</th>
<th>Max. pressure variation</th>
<th>Max. sea-level variation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>From</td>
<td>To</td>
<td>CLO (mb)</td>
</tr>
<tr>
<td>1</td>
<td>00 GMT 28 July</td>
<td>06 GMT 29 July</td>
<td>2.50</td>
</tr>
<tr>
<td>2</td>
<td>21 GMT 7 Sept.</td>
<td>18 GMT 9 Sept.</td>
<td>1.50</td>
</tr>
<tr>
<td>3</td>
<td>06 GMT 24 Sept.</td>
<td>06 GMT 26 Sept.</td>
<td>2.50</td>
</tr>
<tr>
<td>4</td>
<td>12 GMT 6 Aug.</td>
<td>06 GMT 8 Aug.</td>
<td>1.75</td>
</tr>
</tbody>
</table>

The question mark means that no sea-level record is available for this event. However, witnesses at the port observed the maximum oscillations during this case of around 30-40 cm.

Phase speed and so we used a band-pass filter to select a particular frequency band in order to obtain the dispersion relation for the waves. Moreover, when a long time series is analysed, the disturbances may be non-stationary and different modes may be involved; an analysis of the whole event would give an average of all the modes involved. In order to take this fact into account we divided the total interval of an event into subintervals of 6 hours (720 points) every 3 hours (the subintervals are then overlapped). A mean and standard deviation for every event and frequency band was found. The standard deviation gives information about the spread of the results but, owing to the use of a relatively small number of data points, is only useful for providing an estimate of the uncertainty of the mean value.

The frequency bands were selected for the range of periods between 7.5 min and 75 min. The range of frequencies we can study is limited by the distance between the instruments. The phase difference between the perturbation registered at two sites, which depends on this distance, cannot be too large because it could lead to erroneous results, but neither too small because the phase difference must be significant to be distinguishable in both time series. For the above frequency range, assuming a phase speed of around 20 m s⁻¹ and taking into account that the instruments were about 4 km apart, the method will give phase differences of between π/12 and π.

For this analysis we chose nine frequency bands, each of width 0.8 h⁻¹, within the 75–7.5 min range periods, i.e. from 0.8 to 8 h⁻¹, limited by the frequencies

\[ f_n = 0.8n \text{ h}^{-1} \quad n = 1, 2, \ldots, 10. \]

For each subinterval and frequency band we obtained the value of the lag, τᵢⱼ, giving the maximum correlation (i, j = SGR, CLO, AUL). A quality control decision was made at this stage: if \( \tau_i + \tau_j - \tau_k \geq 2\Delta t \), where \( \Delta t \) is the sampling interval, or if the maximum cross-correlation coefficient was less than 0.6 then this estimate was neglected. For these cases we assumed that this particular frequency band was very short lived and/or weakly coherent for this subinterval. This may occur when a frequency band does not have significant energy and then we cannot assign a phase speed or wavelength to this wave. Table 2 shows, for each frequency band and event, how many subintervals satisfy both quality controls.

For the remaining estimates we followed the method used by Rees and Mobbs (1988) for obtaining wavelength, phase speed and direction of propagation (azimuth). The method, briefly, consists of the following. Given the instruments 1, 2 and 3, with coordinates \((x_1, y_1), (x_2, y_2)\) and \((x_3, y_3)\) respectively, and assuming a wave-like
TABLE 2. NUMBER OF SUBINTERVALS SATISFYING THE QUALITY CONTROLS FOR EACH FREQUENCY BAND AND EVENT

<table>
<thead>
<tr>
<th>Frequency band (Median period) (min)</th>
<th>Event 1</th>
<th>Event 2</th>
<th>Event 3</th>
<th>Event 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td>7</td>
<td>10</td>
<td>2</td>
<td>7</td>
</tr>
<tr>
<td>30</td>
<td>6</td>
<td>9</td>
<td>8</td>
<td>6</td>
</tr>
<tr>
<td>21.4</td>
<td>4</td>
<td>5</td>
<td>9</td>
<td>9</td>
</tr>
<tr>
<td>16.7</td>
<td>2</td>
<td>3</td>
<td>10</td>
<td>4</td>
</tr>
<tr>
<td>13.6</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>11.5</td>
<td>0</td>
<td>0</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>10</td>
<td>0</td>
<td>0</td>
<td>5</td>
<td>0</td>
</tr>
<tr>
<td>8.8</td>
<td>3</td>
<td>0</td>
<td>3</td>
<td>0</td>
</tr>
<tr>
<td>7.9</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

disturbance with frequency \( f \) passing over the three instruments, the following relations for the inverse horizontal wavelengths \((k, l)\) hold (Rees and Mobbs 1988):

\[
k x_1 + l y_1 = k x_2 + l y_2 - f \tau_{12}
\]

\[
k x_1 + l y_1 = k x_3 + l y_3 - f \tau_{13}.
\]

Equations (1) and (2) can be solved for \( k \) and \( l \) and it is possible then to obtain the values of wavelength \((\lambda)\), phase speed \((c)\), and azimuth of the wave \((\alpha)\). Therefore

\[
k = \frac{f \tau_{12} (y_1 - y_3) - \tau_{13} (y_1 - y_2)}{(x_1 - x_3) (y_1 - y_2) - (x_2 - x_3) (y_1 - y_3)}
\]

\[
l = \frac{f \tau_{13} (x_1 - x_2) - \tau_{12} (x_1 - x_3)}{(x_1 - x_3) (y_1 - y_2) - (x_2 - x_3) (y_1 - y_3)}
\]

\[
\lambda = \frac{1}{\sqrt{k^2 + l^2}}
\]

\[
c = \frac{f}{\sqrt{k^2 + l^2}}
\]

and by using the meteorological convention for the direction of propagation, increasing clockwise from north, the azimuth is given by:

\[
\alpha = 180^0 + \tan^{-1} \frac{k}{l} \quad \text{if} \quad k > 0, \quad l > 0 \quad (7a)
\]

\[
\alpha = 180^0 - \tan^{-1} \frac{k}{l} \quad \text{if} \quad k < 0, \quad l > 0 \quad (7b)
\]

\[
\alpha = 0^0 + \tan^{-1} \frac{k}{l} \quad \text{if} \quad k < 0, \quad l < 0 \quad (7c)
\]

\[
\alpha = 360^0 - \tan^{-1} \frac{k}{l} \quad \text{if} \quad k > 0, \quad l < 0 \quad (7d)
\]

4. Results

In this section the analysis of every event in Table 1 is made by obtaining wavelength, phase speed and azimuth of the waves. For every case, the data time series from the instrument CLO, after removing periods less than 5 min by means of a low-pass filter, is shown as an example (AUL and SGR time series are always quite similar to the CLO time series). It should be noted that every 12 hours the mean value of the measurements
over that period has been removed and only the pressure variations are actually plotted; therefore the time series during an event does not match at the end of one 12-hour plot and the beginning of the following one. The logarithmic spectrum for the whole event indicates the dominant wave frequencies during every event. The calculation of the number of degrees of freedom follows Bloomfield (1976). Values of wavelength, phase speed and direction of propagation of the wave are plotted against frequency, following the method explained in section 3. The standard deviation for every estimate is also indicated as a measurement of the uncertainty in this estimate. Finally, the vertical profiles of potential temperature, wind speed, and direction are shown using data from the routine Palma de Mallorca sounding.

(a) Case 1: 28–29 July 1990

Event 1 consists of 3600 data-points (1800 min) starting at 0000 GMT 28 July 1990. Figure 3 shows the 12 hours with the largest pressure oscillations at CLO. During this event two aspects are evident: the very-short-period (around 8 min) wave train soon after 1400 GMT 28 July, with a relatively large amplitude (up to 0.5 mb); and the sudden pressure decrease of some 2.5 mb over a few minutes around 1930 GMT.

![Pressure time-series from CLO for the 12 hours starting at 1200 GMT 28 July 1990. Periods less than 5 min have been removed by means of a low-pass filter. Pressure records from AUL and SGR are quite similar.](image)

Figure 4 shows the logarithmic spectrum for the whole event. Relatively large values of wave energy are found at periods of around 7–8 min, reflecting the wave train commented upon in the previous paragraph. On the other hand, the event presents a very broad spectrum with no other significant peaks.

The results for wavelength ($\lambda$), phase speed ($c$) and azimuth ($\alpha$) for every frequency band are shown in Fig. 5. It can be seen that $\lambda$ clearly decreases with frequency. For this particular event, and for shorter periods, few subintervals fulfill the quality controls specified in the previous section; the only exception being the frequency band centered at 8.8 min, which shows an unusual good correlation. In this case, more energy than usual is present at these frequencies (see Fig. 4). Figures 5(b) and 5(c) show that no strong variation of phase speed or direction with frequency exists; the mean values are 20 ms$^{-1}$ and 218° respectively.

Figure 6 shows the vertical profiles and the mean wind speed and direction. Also the height at which the Richardson number is a minimum is indicated. It is notable that the wind direction is almost uniform and is the same as the direction of the wave propagation. However, the wave is propagating somewhat faster than the wind speed at
any level. The potential-temperature profile shows that there is weaker static stability at the level of minimum Richardson number.

(b) Case 2: 7–9 September 1990

Event 2 consists of 5400 data-points (2700 min) starting at 2100 GMT 7 September 1990. Pressure records from the instrument CLO for the period from 0000 to 1200 GMT 8 September 1990 are shown in Fig. 7 after filtering periods less than 5 min.

This event is characterized by the presence of relatively weak pressure oscillations but over a rather long interval of time. A pressure decrease of some 1.5 mb around 0330 on 8 September followed by some smaller fluctuations is the most notable aspect during this event. The spectrum for this case (Fig. 8) presents no significant peaks and it is less energetic than the spectrum for case 1 at shorter periods.

Figure 9 shows the results for wavelength, phase speed and direction of propagation of the wave. No result is given for periods less than 13 min because the cross-correlation coefficients did not fulfill the quality control. This is consistent with the lack of significant energy for these periods in the spectrum (Fig. 8). Comparing the energy around the 8 min period between this event and the previous one (see Fig. 4) we observe a difference of almost an order of magnitude. Wavelength clearly decreases with frequency as in case 1 and the phase speed and direction are again essentially independent of frequency with mean values of $17 \text{ m s}^{-1}$ and $234^\circ$.

Figure 10 shows that there is only a small wind shear and that the minimum Richardson number is near 300 mb. Also shown is the level of minimum static stability as this will be of some relevance when we discuss ideas of wave trapping in section 6. The wave propagates faster than the wind at any level whilst the wave direction is similar to the wind direction in the mid and upper troposphere.
Figure 5. (a) Wavelength, (b) phase speed and (c) direction of propagation of the wave plotted against frequency for case 1 (28–29 July 1990). The standard deviations are indicated by error bars.
The period of measurements selected as event 3 consists of 5760 data-points (2880 min) starting at 0600 GMT 24 September 1990. This event is the longest in duration during the summer of 1990 and also the event with greatest pressure oscillations. Figure 11 shows the pressure disturbances from the instrument CLO from 0000 GMT 25 September to 0000 GMT 26 September. The greatest disturbances occur just before 0800 GMT with a 1.5 mb pressure jump and around 1800 GMT with a sudden rise of some 2.5 mb.

Figure 12 shows the logarithmic spectrum for case 3. It is a broad spectrum but with some significant peaks at 15, 18 and 25 min.

Figure 13 shows that again wavelength decreases with frequency and that the phase speed is remarkably constant with frequency. The mean values for wave speed and
direction are 29 m s\(^{-1}\) and 230° respectively. In this case every frequency band except two satisfy both quality controls and this is perhaps consistent with the presence of large wave energy particularly at lower wave periods (see also Fig. 12).

Figure 14 shows a remarkably large wind shear below 700 mb and almost none above this level. The waves move at the speed of the wind in this upper layer whilst the direction of the wave propagation is some 20° different from the wind direction. Of particular note is the layer with almost zero static stability and minimum Richardson number at the top of this shear layer.

(d) Case 4: 6–8 August 1990

Event 4 consists of 5040 data-points (2520 min) starting at 1200 GMT 6 August 1990. The period of measurements from 0000 to 1200 GMT 7 August is shown in Fig. 15 (again the time series has been pre-filtered to retain only periods higher than 5 min). Significant pressure oscillations are observed. In the spectrum for the whole event (Fig. 16) a clear peak appears at 12–13 min with a sharp wave energy decrease for shorter periods. In consequence, no subinterval, for periods shorter than 12 min, fulfills the quality controls owing to the lack of energy at these frequencies.

The results for this case are similar to the other events concerning the wavelength and phase speed behaviour with frequency (Fig. 17). In this case the average direction of propagation of the wave is more from the west, 280°, than in the other cases. The mean wave speed is 18 m s\(^{-1}\).

Figure 18 shows that there are two levels at which the Richardson number is a minimum. The wave speed and direction are similar to the wind speed and direction in the unsheared layer above the upper level of minimum Richardson number.
Figure 9. As Fig. 5 but for case 2 (7–9 September 1990).
Figure 10. As Fig. 6 but for the sounding made at Palma de Mallorca at 1200 GMT 8 September 1990 and case 2. The level of minimum static stability is shown in (a).

Figure 11. As Fig. 3 but for case 3: (a) From 0000 GMT 25 September 1990 and (b) from 1200 GMT 25 September 1990.
5. **Atmospheric Gravity Waves and the Rissaga Phenomenon**

In the previous section, an analysis of gravity waves passing over the triangular array of microbarographs in Mallorca has been made without reference to the instrument at Ciutadella which is 105 km distant. In case 3, the most energetic event, it is possible to use the data from Ciutadella, for the largest wavelengths, to assess their horizontal coherence. Usually gravity waves lose their identity after travelling perhaps less than about half a wavelength (DeMaria et al. 1989) unless some mechanism allows the energy to remain trapped near the surface (Lindzen and Tung 1976). If we try to follow the method indicated in section 3 for the shorter wavelengths, but using CIU, we find that many subintervals do not fulfill the required quality controls. It is to be expected that the wave does not maintain its coherence over so large a distance, especially for shorter periods (which also have shorter wavelengths). As an example, we show the results obtained for the first 12 hours of 25 September 1990. This interval has been selected because of the good correlations found between the time series from the triangular array in Mallorca and from the instrument at Ciutadella. Perhaps better results could be expected for the next 12 hours, when the largest pressure oscillations were registered, but unfortunately there exists a gap of several hours in the Ciutadella pressure record after 1700 GMT 25 September.

If we obtain the cross-correlation functions for the three instruments in Mallorca during the period from 0000 GMT to 1200 GMT 25 September 1990, we obtain the following lags of maximum cross-correlation coefficients:

\[
\text{AUL} - \text{SGR} = 5 \Delta t
\]

\[
\text{SGR} - \text{CLO} = -4 \Delta t
\]

\[
\text{AUL} - \text{CLO} = 1 \Delta t.
\]
Figure 13. As Fig. 5 but for case 3 (24-26 September 1990).
In this case the condition $\tau_{\text{AUL}-\text{SGR}} + \tau_{\text{SGR}-\text{CLO}} - \tau_{\text{AUL}-\text{CLO}} = 0$ is fulfilled.

As an example, Fig. 19(a) shows the cross-correlation function between stations AUL–SGR. A significant maximum cross-correlation coefficient is evident at 2.5 min (lag = 5$\Delta t$). Furthermore, a periodic behaviour of around 65 min period is clearly visible in this figure, indicating that 65 min is the dominant period during this interval of measurements (the calculation has been made by filtering only periods less than 5 min).

Using Eqs. (1) to (7) we obtain for this dominant period the following values: $\lambda = 105$ km, $c = 27.4$ m s$^{-1}$ and $\alpha = 216^\circ$. These values are consistent with the results shown in Fig. 13.

Figure 19(b) shows the cross-correlation function between stations CIU–AUL (cross-correlation functions between CIU–CLO and CIU–SGR are quite similar). This figure has a similar periodic behaviour to Fig. 19(a), with a dominant period of some 64 min and with a significant maximum value (greater than 0.6) at 54 min (lag = 108 $\Delta t$). We conclude that, in this case, a dominant wave of about 65 min period maintains its identity.
travelling from Mallorca to Ciutadella. Taking into account that the distance between CIU-AUL is approximately 105 km in the direction 215°, we can obtain, by using the CIU-AUL maximum correlation lag, a value for the phase speed of 32.4 ms⁻¹, in reasonable agreement with the results obtained by using the triangular array of microbarographs in Mallorca.

These results suggest that, at least on some occasions, gravity waves maintain their identity over at least one wavelength, which could indicate that some mechanism exists which is able to trap the energy near the surface.

In three of the four cases indicated in Table 1 (cases 1, 2 and 3) similar pressure oscillations were also registered at Ciutadella at approximately the same time. The direction between Ciutadella and the triangular array in Mallorca is approximately 215°, close to the direction of propagation of the waves for cases 1, 2 and 3. During the fourth case no pressure oscillations were registered at Ciutadella. The explanation is that for this case the direction of propagation of the wave was found to be more from the west (around 280°). Only when the horizontal scale of the perturbation is very large would a disturbance propagating with the direction 280° be observed at both Mallorca and Ciutadella.

During event 3 some relatively large sea-level oscillations (of around 1 metre trough-to-crest amplitude) were registered in the harbour of Ciutadella (Fig. 20). Similar sea-level oscillations with smaller amplitude were also observed during events 1 and 2 but, during case 4, when no pressure oscillations were registered at Ciutadella, no significant sea-level variations were observed. These results confirm the relation, observed in previous studies, between the presence of atmospheric gravity waves at Ciutadella and the large sea-level oscillations in the harbour (Monserrat et al. 1991b). However, large sea-level oscillations have been associated with the presence of great energy in the atmospheric gravity wave at around 10 min period (Monserrat et al. 1991b). These short-
Figure 17. As Fig. 5 but for case 4 (7–8 August 1990).
period waves have wavelengths in the range 10–20 km and do not maintain their phase coherence travelling from Mallorca to Ciutadella. Therefore it is difficult to extract conclusive results about the intensity of the rissaga just from pressure records on Mallorca.

6. THEORETICAL IDEAS ON THE POSSIBLE MECHANISMS INVOLVED

The synoptic situation and vertical structure of the atmosphere during periods of wave activity on the Balearic Islands have been identified in previous studies (Ramis and Jansà 1983; Monserrat et al. 1991a, 1991b). The main aspects of the vertical structure of the atmosphere can be summarized as follows:

(i) Two distinct air masses in the mid and lower troposphere, one cold and humid below and the other warm and dry above, separated by a shallow and usually strong temperature inversion near the surface.

(ii) Very weak winds normally from the east or south-east below the inversion veering toward south-west above it.

(iii) Strong wind shear at middle levels with a layer of small Richardson number near 500 mb.

In the cases presented here two are similar to the above description (cases 3 and 4) whilst cases 1 and 2 have the mid-level minimum Richardson number but exhibit relatively weak wind shear (in fact, in case 2 the lack of wind shear in the middle troposphere is the reason why the Richardson number is not small in the layer of minimum static stability). These three aspects of the meteorological situation, therefore, cannot fully describe the necessary conditions for gravity-wave activity. They take a synoptic viewpoint but perhaps are not the most dynamically significant factors.

The vertical structure of the atmosphere will now be analysed to determine the potential for wave generation by shearing instability and whether it fulfills the conditions suggested by Lindzen and Tung (1976) for the existence of a wave duct which could
account for the maintenance of the wave coherence. Case 3, which is the one with the largest pressure oscillations, is examined in more detail.

An important distinction is whether the sounding being used from Palma de Mallorca is representative of the region of wave generation or of the region of ducted wave propagation. This sounding is in a location adjacent to the microbarograph observations. It is plausible to suppose that the wave generation region is of considerably smaller horizontal and vertical extent than the region of ducting which inherently involves horizontal wave propagation. Therefore we expect to find evidence in this sounding for wave ducting but perhaps not for shearing instability.
(a) Wave source mechanisms

Shearing instability has been considered by many authors as a suitable mechanism for the generation of gravity waves and it has been found to be the actual generation mechanism on numerous occasions (Kelihor 1975; Gedzelman and Rilling 1978; Stobie et al. 1983). In these papers evidence of the waves being generated by the dynamic instability of the wind profile is based on the observation that the wave speeds and directions tend to match the wind speeds and directions at the level of minimum Richardson number which, following the classical criterion for shearing instability, should be less than $1/4$. This evidence is also supported by the observation that every frequency band has the same phase speed, i.e. the waves are not dispersive.

Rees and Mobbs (1988) in their analysis of Antarctic gravity waves conclude that they are generated by unsteady flow associated with mountains. Clearly mountain waves must be stationary, so that propagating gravity waves can only be forced by mountains from unsteadiness in the flow. This mechanism is extremely difficult to verify with routine observations, and remains speculative. There are significant mountains in the Balearic islands but they are oriented in a range parallel to the observed flow and our microbarograph observations are located at the upstream edge of the range. The only other significant mountains are remote in Africa (800 km distant) and so can probably be discounted.

Convective activity is another possible source for gravity waves. However, in our four cases this is not present. As described in Uccellini and Koch (1987) another mechanism is that associated with geostrophic adjustment downstream of a strong upper-level jet streak. Although Monserrat et al. (1991a) found evidence of such a jet streak at 300 mb in another similar gravity-wave event over the Balearic Islands, the presence of such a jet is not evident in the cases studied here.

Of these mechanisms the main point supporting the idea that the observed waves on the Balearic Islands are generated by shearing instability is the fact that they are not noticeably dispersive. For our observations the relationship between wavelength and frequency is given by $\lambda = f^{-1/3.2\pm0.1}$ if all the observations are taken into account. This is significantly more non-dispersive than the waves observed in the Antarctic by Rees and Mobbs (1988). Also the profiles exhibit a mid-tropospheric level of minimum Richardson number. In case 3 this minimum value is zero although it is greater in the other cases. For a dynamically unstable wave the phase speed and direction must equal the wind speed and direction at the level of minimum Richardson number. This is possible in cases 3 and 4 but not in cases 1 and 2; however, even in the latter cases the wave direction is similar to the wind direction.

In case 3, the depth of the layer with minimum Richardson number is only about 500 m, implying that the vertical extent of the waves will be small. Given the elevation of the layer at 4 km, we might, therefore, not expect to find strong wave activity at the ground. This perhaps argues against dynamical instability.

Therefore, although some evidence exists that the waves are generated by dynamical instability, we cannot conclusively show that any of the previously mentioned wave source mechanisms are the unambiguous cause for the observed gravity waves. There exists the possibility that the Palma de Mallorca sounding is representative of the region of wave propagation rather than the region of wave generation.

(b) Maintenance of wave coherence

Whatever the source mechanism, the waves can only propagate far from their source region and maintain their coherence if they become neutral and are vertically trapped modes in a duct. Lindzen and Tung (1976) show that an effective duct, able to trap the
wave energy near the surface, must be a statically stable layer deep enough to accommodate at least a quarter of the vertical wavelength corresponding to the observed phase speed. The duct must be capped by a reflecting layer with the Richardson number less than 0.25. This is also the condition for dynamic instability but here is being used in the context of neutral modes. Such a duct can, therefore, produce wave over-reflection. It is interesting to note that the Richardson number has to be small for both dynamic instability and for neutral mode ducting.

This reflecting layer can be produced by either of the following possibilities:

(i) if the Brunt–Väisälä frequency is almost zero in this layer, and
(ii) if the layer is approximately saturated and conditionally unstable.

In case 3 we see that condition (i) is in fact satisfied although over a shallow layer. This layer was about 500 m deep at a height of about 4 km. Although Lindzen and Tung (1976) show that such a shallow layer will be a poor reflector we imagine that the layer from the ground to 4 km may act as a duct.

Also, according to the Lindzen and Tung (1976) trapping model, it is necessary for the wave to have a steering level above or in the reflecting layer. Looking at case 3, and assuming a phase speed of 29 m s⁻¹, a steering level exists below the reflecting layer. However, this is a singularity which would cease to be a steering level if the wave speed was slightly faster. Therefore, a more plausible steering level is indeed above the reflecting layer (see Fig. 14).

If we examine the details of the vertical profile of potential temperature for case 3 we find that the duct is disturbed, whereas the region above the reflecting layer is not (Fig. 21(a)). The different potential-temperature behaviour below and above the reflecting layer is more apparent in Figs. 21(b) and (c) where potential temperature is plotted after removing a linear trend.

If these variations in potential temperature were caused by the waves themselves then they are not the gravest mode of the duct. The gravest mode for waves with long horizontal wavelength can be approximated by the expression

\[ L = \frac{2\pi c}{N} \]

where \( L \) is the vertical wavelength, \( c \) the phase speed relative to mean flow and \( N \) the Brunt–Väisälä frequency in the duct. For our case taking \( c = 29 \text{ m s}^{-1} \) we obtain that \( L = 12 \text{ km} \). The trapped wave is such that a quarter vertical wavelength occupies the full depth of the duct; in this case, therefore, the duct must be of depth 3 km or greater. Thus the observed 4 km deep duct could support the gravest mode. However, the potential-temperature variations in Fig. 21 represent a much smaller vertical scale. Such short vertical-scale waves might be expected to have a smaller phase speed than the observed speed of about 29 m s⁻¹.

Therefore, it seems likely that the step-like structures in the potential-temperature profile within the duct are not trapped waves. Although the exact causes of the oscillations cannot be established from the radiosonde data alone, a possible explanation for the disturbances is that they are caused by wave-induced turbulence. In that sense, it is interesting to note that the oscillations are absent from the region above the reflecting layer, thus giving indirect evidence of the existence of the waves in the duct.

Lindzen and Tung (1976) refer to the phase speed relative to a constant mean flow, but in our case, because of the wind speed shear in the duct, it is difficult to know how to interpret this prediction. The simplified and idealized vertical profiles of wind and temperature used by Lindzen and Tung (1976) perhaps do not adequately describe those from our case 3. A further study is needed in order to describe the characteristics of
Figure 21. (a) High-resolution vertical structure of the potential temperature in the lower 7 km for case 3 (at 0000 GMT 25 September 1990). Oscillations are clearly visible in the lower 4 km but there is no evidence of such disturbances in the upper 3 km. (b) Potential temperature, after removing the linear trend, for the lower 3.5 km. (c) As (b) but for the layer above the reflecting layer (between 4 km and 7.5 km). No evidence of the oscillations remains in this layer.
trapped modes in such a wave-duct. However, the observational evidence presented here is at least suggestive that neutral trapped modes may exist in this situation.

7. Conclusions

We have used surface pressure time series from a triangular array of microbarographs in Mallorca (Balearic Islands) for obtaining wavelength, phase speed and direction of propagation of gravity waves during four selected events of wave activity in the summer of 1990.

The results show that the observed waves were not noticeably dispersive, the phase speed remained approximately constant with frequency for every event and the relation between frequency and wavelength was \( \lambda = f^{-1.03 \pm 0.1} \). This result is consistent with a wave generation by shearing instability but also with neutral trapped modes with small vertical wavelengths.

The calculated direction of propagation for these waves was close to the direction between the triangular array and Ciutadella during three of the events, and for these three cases pressure disturbances were recorded at both sites. The fourth case had a different direction of propagation and no pressure disturbances were recorded at Ciutadella. We conclude that when the wave propagates in the appropriate direction, gravity waves passing over Mallorca are also observed at Ciutadella. Furthermore, when pressure oscillations are registered by CIU there exists some intensity of the rissaga in the harbour. However, since the amplitude of sea-level oscillations depends on the atmospheric wave energy at around a 10 min period and these periods do not maintain their coherence travelling from Mallorca to Ciutadella, no conclusive results can be obtained about the amplitude of the rissaga just from pressure measurements on Mallorca.

Various source mechanisms for the observed gravity waves have been considered. The vertical profiles of wind and temperature suggest that dynamical instability by wind shear is the most likely source mechanism but some uncertainty still remains, perhaps because the Palma de Mallorca sounding is not the most representative sounding for the region of wave generation.

Some deductions suggesting that the observed waves could be trapped modes in a duct have been given. This could be the reason why on some occasions the waves maintain their coherence for long distances involving at least a whole wavelength.

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