Summer thermal lows in the Iberian peninsula: A three-dimensional simulation

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SUMMARY

A summary of some of the most relevant climatological characteristics of summer thermal lows in the Iberian peninsula is given followed by a description of a high-resolution numerical model which has been used to simulate the mesoscale system of a typical summer's day. The results from this model show the appearance of small low-pressure centres before noon, one of which prevails over the others and intensifies during the afternoon. Other small-size structures related to this system, and not evident in the analysis from the ECMWF (European Centre for Medium-Range Weather Forecasts) forecast model, appear in the simulation, such as the existence of areas with strong convergence near the surface, and the deviation towards the interior of the peninsula of the surface winds as the depression intensifies. Also, the structures associated with certain variables, such as potential temperature, vertical velocity and potential vorticity, become more noticeable in the model's simulation compared with the ECMWF analysis. Finally, a detailed description of the dynamical development of the simulated surface low-pressure system is presented, taking all these findings into consideration.

KEYWORDS: Iberian peninsula Mesoscale modelling Regional climate Thermal lows

1. INTRODUCTION

The formation of summer thermal depressions over continental areas is a phenomenon well-known to climatologists (Barry and Chorley 1985). The origin of these systems is related to the intense heating of the land surface during the day, thus causing the formation there of a thermal low-pressure cell that tends to weaken or disappear during the night when solar heating, the cause of its origin, ceases (Hufty 1984). Among the few studies done on the subject of this kind of meteorological system in different areas of the globe both in the sub tropics and in the extratropics, mention might be made of the following.

Blake et al. (1983) and Smith (1986a, b) analysed the characteristics of the Arabian heat low and the role it has in controlling moisture transport in the south-west regions of the Arabian peninsula affected by monsoon rains. Ramage (1971) and Chang (1972) studied the thermal lows that dominate the region of western Pakistan and north-western India during summer—thermal lows that are responsible for the summer Indian monsoon. Junning et al. (1984) analysed the thermal low-pressure systems formed over the Qinghai-Xizang plateau (China), which produce most of the summer rains in that area. They examined the climatological factors that favour their development, with a view to finding a frequency parameter that would make it possible to determine the location of maximum probability for the low centre to be situated at. Sellers and Hill (1974), Gilliland (1980) and Rowson and Collucci (1992) studied the thermal lows that form over the desert regions of southern Arizona, the south-east of California and the north-west of Mexico during the summer

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months. They determined the frequency and the position of the centre of the thermal lows, their vertical extent and the influence that the flow in the upper levels has on development of the system at the surface.

The summer thermal lows over Australia, related to the monsoon responsible for the main part of the annual rainfall in the northern tropical area, have been analysed in depth by Leslie (1980). His study centres mainly on the development of a limited-area model capable of simulating this phenomenon, which is not well predicted by the synoptic operational model used by the Australian Meteorological Service. The small vertical thickness and the weak horizontal gradients that are typical of these depressions, and also the inadequacy of the scheme for surface heat balance in the forecast model, are the reasons why the Australian thermal low has not been well predicted. Finally, Ramage (1971) and Pedgley (1972) have drawn attention to the summer thermal lows over north Africa, which have no rainfall associated with them owing to the great dryness of the air over the Sahara Desert.

The frequent formation of thermal lows over the Iberian peninsula (IP hereafter) in summer is a well-known fact (Linés 1977; Soler 1977; Capel 1981; Font 1983; Barry and Chorley 1985). However, interest in this phenomenon so far has been limited to its inclusion in classifications of the weather types in south-western Europe, with a mere reference to some of its characteristics. An initial approach to establishing a specific climatology of the thermal lows in the IP is to be found in the paper by Portela and Castro (1991), where an objective method of selection and classification of these meteorological systems is described. The mean-sea-level pressures as measured every three hours at 82 synoptic stations in Spain and Portugal were used. The results show that the presence of the thermal low over the IP acquires a quasi-permanent character in summer, constituting the most frequent surface meteorological situation.

The brief survey of the best-known thermal lows everywhere showed a common characteristic: they all form over arid soils. Although this is not the case in the IP, nevertheless the thermal lows that form there are, most frequently, located over the driest areas and show a daily cyclical behaviour with maximum intensity after midday. The relatively strong pressure gradient between the periphery and the central area of the IP has a tendency to cause low-level winds flowing from the coastal areas towards the inner regions; this has been confirmed by the study carried out by Millán et al. (1991) on the trajectories of polluting trails (SO₂) issuing from tall chimney stacks situated in the coastal areas of northern and eastern Spain. The intensity which these depressions can reach does not seem to correspond with the relatively small continental extent over which they form, nor with the non-arid character of its soils. This leads us to consider the possible existence of other factors responsible for their formation. For instance, the geographic location of the IP, almost surrounded by colder marine surfaces, and its complex orography, with two high sub-plateaux in the centre and important mountain ranges (see Fig. 1) that would induce anabatic circulation during the day, reinforcing the divergence in the upper levels and the surface pressure contrast between the interior and peripheral areas of the IP.

The first results obtained from a study of the dynamic structure and the processes of formation and development of the Iberian thermal low, through analysing the behaviour of the potential vorticity field, were published by Alonso et al. (1994). The most relevant characteristic is the existence of a 'dome' of negative potential vorticity over the centre of the depression extending up to nearly 700 hPa. This seems to be connected with what is the typical structure of the field of potential temperature and with the great importance of the effects that the friction and diabatic heating have near the surface in this situation, as detailed in the aforementioned paper.

The scarcity of studies on the thermal low in the IP appearing in the literature might be attributed to the little influence that this mesoscale system has on the predictions made
by global forecast models. The relatively small size of the Iberian thermal low makes some aspects of it, though relevant, pass unnoticed in the meteorological fields obtained from the low-spatial-resolution models. It is for this reason that a mesoscale model has been used in this study to simulate a typical day in a thermal-low situation. In fact, a two-dimensional version of this model was used in a study carried out by Gaertner et al. (1993) to analyse the sensitivity of the Iberian thermal low to several factors such as soil moisture, plateau elevation, thermal stratification, synoptic mean flow and proximity of the ocean. The results of the study showed that the low is less intense when the soil dryness on the central plateau is reduced or when forested land replaces the sea; it becomes deeper when the plateau is more elevated, and it is not significantly affected by an increase in the thermal stratification stability. However, when there is a larger-scale flow, the surface depression has a shorter life.

The 3-D version of the model which we used is capable of representing more accurately phenomena of a size of the order of the system under study, thus making it possible to reveal the dynamic and diabatic processes operating in this meteorological system. Before analysing the most significant results, we shall give a brief description of the main characteristics of the summer Iberian thermal low, the model used, and the simulation conditions.

2. BRIEF DESCRIPTION OF THE MAIN CHARACTERISTICS OF THE IBERIAN THERMAL LOW

(a) Frequency of occurrence

In this preliminary study an analysis has been done using the mean-sea-level pressures as measured at 82 stations of the synoptic network in Spain and Portugal during the periods 1973 to 1977 and 1985 to 1988. The criteria which were used when selecting the days on which thermal lows occurred have been prepared considering the typical features characteristic of this type of low and the macroscale atmospheric conditions that affect that part of Europe (Portela and Castro 1991).
Figure 2 shows the monthly frequency distribution of thermal-low days in the IP throughout the aforementioned nine-year period. The monthly variation in the distribution is similar in each year. Maximum frequencies above 50% can be observed in July and August, but in March and October this phenomenon occurs on average on only 10% of the days. Because February and November showed only negligible percentages, they have not been included in the figure, neither have December and January, in which months the thermal low has no incidence.

Although in all the years considered the monthly frequency distribution is quite similar, some differences can be detected which seem to be related to whether the winters and
springs of the corresponding years were wet or dry, as was noted by Portela and Castro (1991). Thus, for example, the driest month of those studied was August 1985; this also was the month showing the highest number of thermal-low days (81%).

(b) Preferred location of the centre of the Iberian thermal low

The centre of a low-pressure system is taken as being located over the area where the minimum in the local surface pressure field is reached. Some authors, such as Rowson and Colucci (1992) in their synoptic climatology of the thermal lows in south-western North America, defined it as the approximate geometric centre of the innermost closed isobar, and applied this process while carrying out an inspection of surface pressure maps in which the isobars had been drawn at 4-hPa intervals. A less subjective method has been used for our study, because, since the surface pressure readings from the analysis done by the ECMWF (European Centre for Medium-Range Weather Forecasts) model are available in a regular grid with a horizontal resolution of 0.5 deg both in latitude and longitude, the determination of the centre of the low has been accomplished with reference to that geographic location (in longitude–latitude coordinates) on the grid where the pressure showed its minimum value. As the gridpoints are quite close together, it was not considered necessary to perform interpolations with a higher resolution.

Figure 3 shows the number of days on which a thermal low was centred at various points of the domain over the four years of the analysis. The existence of a preferred area in the south-west of the IP can be seen where this system tended to form and remain. Specifically, on 52% of the days analysed, the centre of the low was located in the region of the peninsula within the domain $7^\circ W \leq \lambda \leq 5.5^\circ W$ and $38^\circ N \leq \varphi \leq 39.5^\circ N$, where $\lambda$ and $\varphi$ are longitude and latitude, respectively. A possible explanation for this preferred location might be that this particular area has certain characteristics in its favour, such as the dryness of the soil and its proximity to the Saharian thermal low.
Figure 4. Monthly frequency distribution of (a) intensity and (b) horizontal extent of the Iberian thermal low for the period 1985–1988.

(c) Intensity and horizontal extent of the Iberian thermal low

The typical mean-sea-level pressure field accompanying the Iberian thermal depression is characterized by having its strongest gradients, not in proximity to the centre of the system—that is to say in the south of the peninsula—but in areas close to the north and east coastal regions, where the surface pressures are higher. So as to underline this fact, the intensity of the thermal low, I, has been quantified in terms of the difference between the average pressure at 1800 UTC in the central points of the ECMWF analysis grid situated over land and the minimum pressure recorded at the same hour. The ECMWF analysis for the days of thermal low in the period considered above has been used.

Figure 4(a) shows the monthly frequency of intensity of the Iberian thermal low calculated in the manner explained above, the values having been obtained at regular pressure intervals of 0.5 hPa. The month of July shows a higher frequency of high intensities. For example, the maximum value (27%) occurs within a range of intensities from 3.75 to 4.25 hPa in that month; meanwhile in August the intensities which have the maximum incidence (36%) have lower values (between 2.25 and 2.75 hPa). Although the higher intensity of
the thermal low does not necessarily imply that it has a lower minimum pressure, nevertheless very intense lows usually have relatively low pressures of this magnitude. Thus, it is noticed that July not only has the highest frequency of high intensities, but also it is in that month that the minimum pressures generally have their lowest values.

On the other hand, the horizontal extent of the thermal low was determined by taking the difference between the average pressure at 1800 UTC within the 24 gridpoints surrounding the centre of the thermal low, and the pressure at the centre. The value so obtained is normalized by the area corresponding to those same gridpoints, using the same data-base that had been used when calculating the intensity of the thermal low.

Figure 4(b) shows the monthly frequency, as observed, of the amounts of the horizontal extents grouped at intervals of 0.1 hPa/3.9 × 10^4 km^-2. (Note that the inverse of these values would represent the area around the low centre, where the pressure would differ by less than 1 hPa from the minimum pressure). In July the most frequent horizontal extent of the thermal low (39%) has a value of 0.5 hPa/3.9 × 10^4 km^-2; in August and September of 0.4 hPa/3.9 × 10^4 km^-2, with a monthly frequency of 33% and 43%, respectively, and of 0.3 hPa/3.9 × 10^4 km^-2 in June, with a frequency of 35%. It can be seen that the maximum extent occurs in July, followed by August.

(d) Vertical extent of the Iberian thermal low

The ECMWF analyses corresponding to thermal-low days show that the temperature distributions at the 1000 hPa level have an appearance similar to those for geopotential height. Isotherms are distributed approximately in a concentric pattern around the point where the maximum temperature is reached, which point often coincides with the minimum of geopotential height and with the centre of the low at the surface. A similar situation can be observed at the 850 hPa level. However, and as was expected, this similarity diminishes as the altitude increases, noticeably changing the curvature of the isotherms. A typical example of this behaviour was reported in the paper by Alonso et al. (1994).

Bearing this behaviour in mind, we could consider that the higher level influenced by the surface thermal low coincides with that level where a change in the curvature of the isentropes is observed. To detect this feature, the Laplacians of the potential temperature over the centre of the thermal low between 1000 and 300 hPa have been calculated at 50 hPa intervals. Since, at the surface, the potential temperature field shows a maximum at the thermal-low centre, the Laplacian will be negative there. The same will happen at the other levels where this structure is maintained—the sign of the Laplacian changing when the influence of the thermal low ceases. Thus, we conclude that the thermal low is limited vertically by that pressure level at which a change is detected in the Laplacian of the potential temperature.

The results obtained using this method, for the thermal-low days in 1985, show the relatively small thickness of this system compared with other thermal lows which had developed in mid latitudes. The most frequently occurring vertical extent of thermal lows in the IP is about 750 hPa (38%), followed by 800 hPa (18%). None of the lows analysed exceeded 500 hPa in height in any case.

3. Three-dimensional simulation of a thermal-low day

(a) Description of the numerical model

The model to be applied, the so-called PROMES (Spanish acronym for PRONóstico a MESoescala) described by Castro et al. (1993), is a primitive-equation, fully compressible,
hydrostatic model that uses Cartesian coordinates horizontally and, in the vertical, a terrain-following coordinate proposed by Phillips and modified by Shuman and Hovermale (1968), viz.

\[ \sigma = \frac{p - p_s}{p_s - p_i} \]

where \( p \) is the pressure, \( p_s \) is the surface pressure and \( p_i \) the pressure at the top of the domain. The parametrization of the surface energy balance is achieved by means of the force-restore method proposed by Bhurulkar (1975) and Blackadar (1976). In this method the forcing of radiative flux and latent and sensible heat fluxes at the surface is damped by the heat flux towards the soil layers, where the diurnal thermal wave has no influence. The turbulent fluxes in the surface layer are determined following the model developed by Blackadar (1976) and Zhang and Anthes (1982), in which four turbulent regimes are considered: stable, mechanical turbulence, forced convection and free convection. The vertical turbulent exchanges above the surface layer in the first three cases is parametrized by means of a first-order closure scheme (K-theory). The \( K \) coefficients are calculated locally and depend on atmospheric stability and vertical wind shear (Blackadar 1976; McNider and Pielke 1981). Under conditions of free convection, vertical mixing is not considered to be determined by local gradients but by the thermal structure throughout the entire mixing layer, following the method developed by Estoque (1968) and Blackadar (1978). In this way the turbulent exchanges take place between the lowest layer of the model and each level inside the mixing layer. Above the mixing layer the \( K \)-theory is applied once more. A fourth-order scheme is used for horizontal diffusion, with a coefficient proportional to total flow deformation (Smagorinsky et al. 1965; Doms 1990). The parametrization of radiative effects included both shortwave and longwave fluxes (Sasamori 1968, 1972) damped by the presence of clouds; values of the emission function were obtained from tabulations given by Mahrer and Pielke (1977). Finally, the water processes were modelled using an explicit method (Hsie et al. 1984), with prognostic equations for vapour, cloud and rainwater and phase-change coefficients proposed by Pielke (1984).

A Lorenz grid-mesh in the vertical and an Arakawa-C in the horizontal were used. The different terms of the primitive equations were solved using several explicit finite-difference schemes, except for vertical diffusion which is implicit. Although a more detailed description of the numerical schemes used may be found in the papers by Fernández (1992) and Gaertner (1994), we shall comment briefly on them here. For the gravity-wave terms a forward–backward scheme is applied (Sun 1980), while the advective terms are solved by forward-in-time, cubic-spline upstream interpolation schemes. A forward–backward scheme is used for the Coriolis term. The vertical turbulent exchange terms are solved by an implicit scheme (Paegle et al. 1976). Finally, the terms supporting the surface temperature tendency, radiative heating and cooling of the air, cloudiness and precipitation processes and horizontal diffusion, are solved by forward-in-time Eulerian schemes.

(b) Simulation conditions

A typical summer thermal-low day (31 July 1985) has been simulated, beginning at 0000 UTC and lasting for a period of 30 hours. The simulation characteristics are as follows.

(i) Initial and boundary conditions. The model grid has been nested in the grid of the ECMWF model. The initial values in the model gridpoints were calculated by a biperabolic horizontal interpolation (Koehler 1977) from the ECMWF analyses \((0.5^\circ \times 0.5^\circ \text{ resolution})\) at the standard pressure levels, and linearly in the vertical (Gaertner 1994). Subsequently, a digital filter initialization is applied (Lynch and Huang 1992) to avoid the initial noise due to unbalance between the mass and wind fields. Although the use of an
initialization scheme may remove the initial high-frequency noise, it does not prevent the fields from having to adapt to the presence of a more detailed orography at the beginning of the simulation, as Majewski (1985) has pointed out.

A Davies-type boundary condition was used (Davies 1983) to determine the boundary values of the model from the 6-hourly ECMWF analyses linearly interpolated in time. To avoid noise generation at the upper boundary an absorbent layer is introduced (Klemp and Lilly 1978).

(ii) **Model domain and resolution.** The model domain is a 1300 × 1200 km² rectangle centred approximately on the IP, with its vertexes located at the following coordinates: (34.85 °N, 10.69 °W), (45.01 °N, 11.54 °W), (44.58 °N, 5.23 °E), (34.49 °N, 3.57 °E). The horizontal grid length is 20 × 10 km² and 31 variable resolution sigma levels are considered in the vertical, with the top level of the model situated at 300 hPa. The lower and upper layers are narrower than the intermediate layer.

(iii) **Topography and soil characteristics.** The model orography has been deduced from the ETOPO 5 data-base of the National Geophysical Data Center (Boulder, Col., USA) with a five-minute resolution in latitude and longitude. Subsequently an average of the height of all the points included in each grid was calculated, a bidimensional filter was applied in two steps: one to smooth and the other to restore the amplitude of waves larger than 2Δz. This technique reduces the generation of shortwave noise due to orographic forcing (Gaertner 1994).

In the surface processes parametrization schemes are necessary to assign the values of five parameters, related to the soil characteristics, in each grid cell of the model. These are albedo, longwave radiation emissivity, fraction of soil moisture available for evaporation, roughness length and thermal inertia of the soil. The values of all these parameters are assigned by considering different land-use categories as proposed by Benjamin and Carlson (1986) for summertime. The subsoil temperatures correspond to the climatological temperatures for July given by the ECMWF analysis.

(iv) **Synoptic setting of the simulated day.** At 0000 UTC on 31 July 1985 the surface synoptic situation was dominated by the presence of an anticyclone centred over the Azores, with a radius of influence extending to the western Mediterranean, while the Saharan depression was affecting the southern regions of the IP. The 1200 UTC synoptic chart still does not show the Iberian thermal low, probably because of its small size at that time and the intervals at which the isobars were drawn. However, the depression is evident at 1800 UTC, with its centre located over the south of the IP. An upper-level trough can be seen with its axis extending from SW England to the north-east of the IP. The skies were mostly clear across the IP, with no significant precipitation, except for some isolated afternoon storms in the Iberian mountain range.

4. **Results**

(a) **Verification of the model results**

Before commenting on the more significant results of the simulation, an assessment of its degree of agreement with the ECMWF analysis and surface observations has been made.

The simulation results were compared at 12, 18 and 24 hours of simulation time with the ECMWF analysis (geopotential height, temperature and horizontal wind components) at the standard pressure levels from 1000 to 500 hPa, after having made an interpolation from the model to the analysis grid. At the same time, mean-sea-level pressure, temperature
and wind observations from the surface stations in Spain and Portugal also were compared with the model's values interpolated at the surface points.

The verification indexes used were: mean error ($E$), root-mean-square error (RMSE), standard deviation of error ($S$), absolute correlation coefficient ($R$), standard deviation of model results (SDM), standard deviation of verifying values (SDV) and root-mean-square error after the removal of a constant bias (RMEC). Once these verification indexes had been determined, the criteria proposed by Keyser and Anthes (1977) were applied in such a manner that the model's results could be considered verified if they satisfied the following conditions:

$$(a) \ SDM \approx SDV, \quad (b) \ RMSE < SDV, \quad (c) \ RMEC < SDV.$$  

One fact that should be borne in mind in this comparison is that there is a difference in spatial resolution between the model and the ECMWF analysis grids or surface network. Thus, the smaller-scale features generated by the model might not be present in the analysis or observations fields.

As may be seen from Table I, the results obtained throughout the first 24-hour period show, in general, a high degree of agreement between the model and the ECMWF analysis, most of them fulfilling the requirements of Keyser and Anthes. The results do not diminish in quality when the time of simulation is increased. Also, agreement between the model and the ECMWF analysis improves as the height is increased. This may be related to the presence of simulated mesoscale-size structures in the lower layers not reflected in the analysis. These surface-related structures are not noticed at higher levels since the thermal thickness does not surpass the 700 hPa level. There is poorer agreement between model and analysis in the wind fields in the lower layers (1000 and 850 hPa) after 24 hours of simulation, which may be due to the abundant drainage flows simulated along the valleys of the main rivers (which will be commented on in the following section) that do not appear in the ECMWF analysis. It should also be noticed that the root-mean-square errors remain relatively constant throughout the simulation period. This was noted also by other authors in the case of limited-area models. Errico and Baumhefner (1987) ascribed this to the fact that, largely, errors in the interior of the domain are 'swept out' by the lateral boundary conditions used.

On the other hand, the results of the assessment with the observations show a good agreement as regards the mean-sea-level pressure and surface temperature fields, though the latter have the unresolvable problem of the topographic height difference between some of the stations and the corresponding model grids. However, verification fails in the case of the surface winds, which, we consider, is mainly because the observations that we used were instantaneous and strongly influenced by local effects.

(b) Main results of the simulation

Figure 5 shows the evolution of mean-sea-level pressure together with the simulated surface-wind fields from 0900 UTC 31 July to 0300 UTC 1 August, at 6-hourly intervals. It may be observed that at 0900 UTC, approximately four hours after sunrise, some surface low-pressure nuclei began to develop in the interior of the IP, one on the southern slope of the Pyrenees, another to the west of the Iberian mountain range and a third to the north of the Penibetic mountain range. As the day progressed, the first two disappeared while the third one predominated, reaching a minimum value lower than 1006 hPa at about 1500 UTC. At the same time, strong pressure gradients are observed, more intense in the north and west of the peninsula, with the isobars running almost parallel to the coast. The eastern area of maximum gradient is located at some distance from the Mediterranean
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**TABLE 1.** Verification indexes at different hours of geopotential (G), temperature (T), surface pressure (P) and wind components (u, v): E = mean error; M = model averages; V = verification averages; S = error standard deviation; RMSE = root-mean-square error; R = correlation coefficient; SDM = standard deviations of model results; SDV = standard deviations of verification values; RMEC = root-mean-square error (bias removed).
coast, joining another band of distinctive pressure gradient to the north-east of the Iberian range. Curiously, both areas of maximum gradient are orientated approximately along the line that divides the Mediterranean and the Atlantic basins of the IP.

A comparison between the characteristics of the simulated thermal-low and the ECMWF analysis for the same day shows the following differences. The simulated low centre (37.5°N, 3.0°W) is shifted approximately 250 km to the south-east relative to that of the analysis (38.5°N 5.5°W). Also, the simulated low is more intense than that of the analysis (6.8 hPa vs. 4.3 hPa) and of greater horizontal extent (4.4 hPa/3.9 $10^4$ km$^2$ vs. 0.5 hPa/3.9 $10^4$ km$^2$). However the vertical extent is similar (780 hPa vs. 750 hPa). These differences seem to be related to both the higher resolution and better surface parametrizations of the model compared to those of the ECMWF model. In Fig. 6 the temperature fields at a height of two metres from the ECMWF analysis (Fig. 6(a)) and from the PROMES model (Fig. 6(b)) are shown. The small-scale structures, only present in Fig. 6(b), seem to confirm this hypothesis. If this were true then the results of the statistical study of the Iberian thermal-low’s characteristics, previously presented in sections 2(b), (c) and (d), would require to be revised if higher-resolution analysis became available.
Figure 6. Temperature fields at two-metre heights at contour intervals of 2 K on 31 July 1985 at 1800 UTC from (a) ECMWF analysis and (b) PROMES model results.

Figure 5 also shows that the simulated surface winds in the peripheral area of the IP are affected significantly by the thermal-low formation. Initially they were almost parallel to the coasts, and as the low intensified they suffered a deviation, acquiring a component towards the interior of the peninsula. These convergent surface flows usually follow the main river valleys and the mountain passes.

An interesting feature of the surface wind fields is the presence of some areas of strong convergence, which mainly coincide approximately with the aforementioned eastern band of maximum pressure gradient. Although it would be rather speculative, on the basis of the single simulation, to attribute a climatological character to this particular distribution of maximum convergence areas, nevertheless it is important to observe that some of these areas coincide with those having maximum frequency of summer storm formations (Font 1983).

After sunset, when the thermal low weakens, the surface winds over the peripheral areas tend to be almost parallel to the coasts again, the pressure gradients diminish noticeably, mainly in the southern and eastern areas, and significant drainage winds appear along the main river valleys, becoming stronger through the night.

Figure 7 shows the wind and geopotential height fields corresponding to the 850, 700 and 500 hPa levels at the time of maximum thermal-low intensity (1500 UTC). At 850 hPa the effect of the surface depression can still be observed, with convergence
approximately over the same areas, and with strong convergence at the surface. On the other hand, though in the north and west of the peninsula the surface winds are almost parallel to the coasts, this is not the case in the Mediterranean area. Here there is an inlet of eastern winds, apparently associated with a small and shallow cyclonic vortex located on the coast of Algeria. In the figure corresponding to the 700 hPa level, areas of greater divergence appear, coinciding approximately with the location of stronger convergence at lower levels. However, at the 500 hPa level the thermal low has no influence, a weak synoptic trough appearing to the west of the IP, with relatively intense quasi-geostrophic winds (notice the different scale of the vectors in the figure corresponding to this level).

The simulated potential temperature (dashed line) and geopotential height fields (continuous line) between 1000 and 500 hPa levels at 1500 UTC are shown in Fig. 8. Their appearance at the lower levels is coincident with the origin of the thermal low (its centre coincides with a maximum of the potential temperature field), while this warm nucleus does not appear at the 700 hPa level and aloft, so illustrating the small thickness of the system.

Figure 9 illustrates the zonal (east-west) and meridional (north-south) vertical cross-sections of potential temperature over the centre of the low at 1800 UTC. It shows the
Figure 8. Simulated (a) 1000 hPa; (b) 850 hPa; (c) 700 hPa; (d) 500 hPa potential temperature (dashed lines) at 2K contour intervals and geopotential height (continuous lines) at 10 geopotential-metre contour intervals on 31 July 1985 at 1500 UTC.

typical funnel-like structure of the isentropes, that also appear in the ECMWF analysis, though less intensely (not presented). In the zonal cross-section the great stability over the Mediterranean Sea (right-hand side of the figure), and the Atlantic Ocean (left-hand side) are clearly observed. The meridional cross-section shows the same feature over the Mediterranean (left-hand side of the figure) and, with less intensity, over the Cantabrian Sea (right-hand side). Stable layers such as these over maritime areas close to the IP may be related to divergent flows out of the thermal low and subsidence over the peripheral areas.

Likewise, in Fig. 10 the vertical cross-sections, over the centre of the low, of the vertical wind distribution at 1800 UTC are shown. In the east–west section the presence of a cell of upward velocities over the centre of the low may be observed, reaching maximum values of 30 cm s\(^{-1}\), approximately, at the 800 hPa level with downward-directed winds at both sides of the column. In the meridional cross-section, alternating ascending and descending wind cells appear: one ascending nucleus (positive values) just over the low centre and another located approximately over the northern foothills of the Iberian range.
that coincides with the strong horizontal convergence at the surface and at the 850 hPa level (see Figs. 5 and 7).

Finally, Fig. 11 shows the zonal and meridional vertical cross-sections of potential vorticity at 1800 UTC over the centre of the depression. In both cases a dome is seen over the low centre, with negative potential vorticity, this being a particular characteristic of the thermal-low system. The great relevance of the effects of friction and diabatic heating near the surface in this situation might be the cause of this, as Alonso et al. (1994) have pointed
Figure 10. (a) Zonal and (b) meridional vertical cross-sections of vertical wind at 8 cm s$^{-1}$ contour intervals over the low centre (37.5°N, 3°W) on 31 July 1985 at 1800 UTC. Negative and zero values are in continuous lines, and positive values in dashed lines. The centre of the low is indicated by a vertical arrow.

out. However, this feature becomes much clearer in the mesoscale model simulation than in the ECMWF analysis used by Alonso et al.

All these separate findings could be combined to give a dynamical picture of the simulated surface low-pressure system: The intense heating of the lower air layers over the Iberian peninsula during the morning induces an upwards expansion of the air, which, in turn, gives rise to a separation of the isobaric surfaces inside the heated layer and the
Figure 11. (a) Zonal and (b) meridional vertical cross-sections of potential vorticity at 0.2 PVU contour intervals (1 PVU is equivalent to $10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K} \text{ kg}^{-1}$) over the low centre ($37.5^\circ \text{N}, 3^\circ \text{W}$) on 31 July 1985 at 1800 UTC. Negative and zero values are in dashed lines, and positive values in continuous lines. The centre of the low is indicated by a vertical arrow.
formation of a zone of relatively high pressure near the top, where a divergence of air takes place. As a result, a thermal low that covers most of the peninsula develops, and a band with a pronounced pressure gradient surrounding the surface depression is observed to form. This mechanism induces a component directed inland in the peripheral surface air fluxes, which originally were almost parallel to the Atlantic and Mediterranean coasts. The height of the level where the divergence of mass takes place depends mainly on the intensity of the heating of the low-level airmass (albedo, soil moisture, etc.) and on the elevation of the ground, as Gaertner et al. (1993) have pointed out. According to the simulation results, the aforementioned level is usually located between the 850 hPa and the 700 hPa isobaric surfaces. This can be deduced from Fig. 7(b) from the weak distortion of the contours, and in Fig. 7(c) from the absence of disturbance at the 500 hPa heights.

Inside the thermal low, which covers most of the peninsula and which is situated close to mountainous areas, other depressions, more intense and of smaller horizontal dimensions, develop. However, the most intense simulated thermal-low centre (37.5°N, 3°W) is located in a zone where, the previous night, there had been a remnant of the surface depression formed the day before, as can be seen in the 0000 UTC ECMWF analysis used as initial condition in the mesoscale model. (These small-scale lows are not evident in the ECMWF analysis owing to its low resolution). These small-scale lows have a greater vertical depth, exceeding the 700 hPa level in the evening, as can be seen in Fig. 7(b), where it can also be noticed how those zones which have a pronounced flux divergence coincide with the most intense of the surface depression centres appearing in Fig. 5(b). These small lows induce marked convergence zones in the low-level layers where the fluxes, which initially are almost normal to the isobars, start circulating cyclonically at the end of the evening. However, at that time of day, the depression disappears and the flux weakens; therefore pronounced vortexes do not form (see Fig. 5(b) and (c)).

Finally, a displacement of the centres of the small-scale depressions is observed after sunset, as well as their weakening. This is evident in the case of the most intense low, as can be noticed in Fig. 5(b), (c) and (d). The movement of this low centre towards the northwest seems to coincide with the zone of negative potential vorticity around the depression (Fig. 11(a) and (b)). In the zonal cross-section (Fig. 11(a)), the negative potential vorticity area is situated to the west of the centre of the low and, in the meridional cross-section, to the north. This displacement provides a clue to the location of the centre of the most intense small-scale low that would develop the following day, as Alonso et al. (1994) suggested.

5. Conclusions

The results of a preliminary climatological study of the Iberian thermal lows based on nine years of surface observations and four years of ECMWF analysis data indicate the following features and characteristics of these typical mesoscale systems:

(i) Its maximum occurrence takes place in the middle of the summer (July and August), with a frequency of between 30 and 60%.

(ii) The centre of these lows is located preferentially in a zone in the south-west of the IP.

(iii) The most intense thermal lows, and those of larger horizontal size, are observed usually in July.

(iv) The Iberian thermal lows have small vertical thickness compared to other similar systems that develop in subtropical regions—not exceeding the 550 hPa level in any of the days analysed.
Subsequently, the results of a simulation of a typical thermal-low day (31 July 1985) using a mesoscale model of $20 \times 20 \text{ km}^2$ horizontal resolution and 31 vertical sigma levels are presented. They underline certain aspects that are somewhat different from or do not appear in the ECMWF analysis corresponding to that day. A horizontally more extended and more intense thermal low with its centre located to the south-east of that indicated by the analysis results from the simulation. Although it has not yet been verified, this could be due to the fact that these systems were not well represented in the low-resolution ECMWF analysis for the years under consideration. Therefore, when a higher-resolution ECMWF analysis becomes available, the aforementioned climatological characteristics of the Iberian thermal lows should be rechecked.

The most relevant aspects of the simulated thermal low are these:

(i) Surface winds, that run approximately parallel to the coast during the early hours of the morning and turn towards the interior of the peninsula as the depression intensifies, showing a tendency to move up the valleys of the main rivers and mountain passes. This agrees with observations of pollutant trajectories from the coastal areas towards the interior of the IP on thermal-low days.

(ii) The stronger bands of surface pressure gradient appear somewhat distant from the thermal depression centre, parallel to and close to the northern and western coastal areas of the peninsula. In the Mediterranean zone this band is situated further towards the interior, almost coinciding with the orographic line that divides the Atlantic and Mediterranean basins of the IP.

(iii) Several areas with a strong surface convergence appear when the thermal low is well developed. Among them, the one simulated on the north-east of the peninsula is worth mentioning since it coincides with a zone having maximum climatological frequency of occurrence of summer storms.

(iv) The results obtained for the upper levels are very similar to the ECMWF analysis, with a clear thermal influence below the 850 hPa level, convergent flows up to the 700 hPa level and strong divergence just above that level, whereas the wind tends to be geostrophic at the 500 hPa level.

(v) Finally, the vertical distributions of the potential temperature and potential vorticity over the centre of the low at the surface are similar to those of the ECMWF analysis, though the characteristic structures of these variable fields become more noticeable in the model simulation.

This study of the Iberian thermal lows is expected to be completed in the future after further analysis which will be directed towards determining the sensitivity of this mesoscale system to other effects such as variations of soil characteristics, orography, etc., so completing an earlier study by Gaertner et al. (1993) using a 2-D version of the same numerical model applied in a zone in the north of the IP.

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REFERENCES


Barry, R. G. and Chorley, R. J. 1985 *Atmosfera, tiempo y clima*, 4th ed., Omega, Barcelona


Capel, J. J. 1981 *Los climas de España*. Oikos-tau, Barcelona


Fernández, C. 1992 ‘Desarrollo y aplicación de un modelo bidimensional para la simulación numérica de la atmósfera a mesoescala.’ Thesis Doctoral, Departamento de Física de la Tierra, Astronomía y Astrofísica I, Facultad de Ciencias Físicas, Universidad Complutense de Madrid


Gaertner, M. A. 1994 ‘Aplicación de un modelo numérico de predicción meteorológica a la simulación de flujos atmosféricos a mesoescala en la zona centro de la península Ibérica’. Thesis doctoral, Departamento de Geofísica y Meteorología, Facultad de Ciencias Físicas, Universidad Complutense, Madrid


Gilliland, R. P. 1980 ‘The structure and behavior of the California heat trough’. Master of Science Thesis in the Department of Meteorology, San José State University, U.S.A.


Hufty, A. 1984 *Introducción a la Climatología*. Ariel Geografía, Barcelona


Koehler, T. L. 1977 ‘A test of seven methods which perform grid observations interpolations’. Pp. 55–65 in Meteorological applications of satellite indirect soundings II, NOAA grant 04-4-158-2, University of Wisconsin


Soler, A. M. 1977 ‘Sitaciones meteorológicas locales típicas: su persistencia y parámetros o variables más característicos’. Tesis Doctoral, Facultad de Ciencias Físicas, Universidad Complutense, Madrid
