

The effect of the island of Crete on the Etesian winds over the Aegean Sea

By VASSILIKI KOTRONI*, KOSTAS LAGOUVARDOS and DIMITRIS LALAS

National Observatory of Athens, Greece

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SUMMARY

The Etesians (northern sector winds), which blow over the Aegean Sea during summer, affect human activities in the area. The numerous islands of the Aegean and especially Crete (a mountainous island in the southern Aegean oriented perpendicular to the surface flow) seem to play an important role in the modification of the wind field during the Etesians. The Crete mountain ranges, surrounded as they are by water, are an excellent example of a major isolated topographic feature which significantly modifies the regional airflow and pressure; however, this modification can hardly be defined due to the lack of observing stations over the sea. For this reason, the available land surface and ship synoptic observations are used, together with ERS scatterometer wind data in order to identify the regions over the Aegean where the wind reaches its maximum intensity, and to assess the influence of Crete on the wind field. Moreover, numerical modelling is used to provide some further insight on the orographically disturbed wind flow. Sensitivity tests performed with the hydrostatic model BOLAM show that the interaction of the Etesian wind flow with the mountains of Crete produces deceleration of the Etesians up to almost 120 km upstream, the leftward deflection of the air as it approaches the mountains, and the associated intensification of the flow east of the island.

KEYWORDS: Orography Satellite winds Wind flow modification

1. INTRODUCTION

One of the most important meteorological events that occurs over the Aegean Sea during summer is the Etesian winds. The Etesian winds are northern sector winds blowing over the Aegean Sea during summer and early autumn. They are mainly north-easterly in the northern Aegean, northerly in the central and southern Aegean, and tend to become north-westerly near the south-western Turkish coasts. The air masses regularly originate from the region of southern Russia and the Caspian Sea and they are dry and relatively cool, contributing to the decrease of surface temperature and the moderation of summer heat and discomfort (Meteorological Office 1962). The sustained wind speed associated with the Etesians often attains surface values exceeding 15 m s^{-1} , (with gusts over $20\text{--}25 \text{ m s}^{-1}$), thus creating problems to sea transport within the Aegean Sea during the high season for tourists. Additionally, they can contribute to the rapid spread of forest fires. The importance of this wind regime is highlighted by the fact that it is a weather type incorporated by the US Navy in their fuzzy expert system to assist in the prediction of hazardous wind conditions in the Mediterranean basin (Kuciauskas *et al.* 1998).

The Etesians result from a combination of (Meteorological Office 1962; Brody and Nestor 1985):

- the monsoon effect that leads to a thermal low-pressure trough over Turkey, with higher pressures over the southern Balkans.
- the passage of cold fronts over the Balkans and the associated cold-air circulation behind them.

Once the synoptic situation for the onset of the Etesians is established, the wind flow over the area is also affected by the important topographic features. Indeed, the Aegean Sea is surrounded on three sides by complex topographical features, which include high mountains on continental Greece, southern Balkans and Turkey (Fig. 1). On the southern

* Corresponding author: National Observatory of Athens, Institute of Environmental Research, Lofos Nymfon, 11810 Athens, Greece. e-mail: kotroni@env.meteo.noa.gr

edge of the Aegean is the mountainous island of Crete, while the maritime area is dotted with a large number of islands of various sizes, with important channels between them. As pointed out by Brody and Nestor (1985), these topographical features frequently combine to create channelling of the flow by coastal valleys, channels between the islands and the mainland, as well as corner and obstacle effects over the various mountain barriers. Moreover, mountains oriented perpendicular to the Etesian direction (e.g. the mountains of Crete, see Fig. 1) block the flow and give calm seas on the lee side, while downwind from coastal valleys strong localized winds are frequently observed.

Although some studies have been devoted to the analysis of the synoptic conditions leading to the onset of Etesian winds over the Aegean Sea (Meteorological Office 1962; Metaxas 1977; Brody and Nestor 1985; Metaxas and Bartzokas 1994) as well as to the derivation of a linear correlation of wind speed with pressure gradients over the area (Prezerakos 1975), the detailed description of the wind field and its interaction with the complex topographic features of the Aegean have drawn little attention. This lack in the bibliography on these subjects can be partly attributed to the lack of detailed measurements (the only available observations being those of few synoptic stations over the Aegean islands and occasionally from ships) as well as on the high demand of computer power for fine-grid simulations over the area.

On the other hand, the interaction of the wind flow with complex topographic features has drawn the attention of many scientists in the past from the theoretical and analytical point of view. Smith (1982) presented a survey of synoptic data from the vicinity of major mountain ranges (New Zealand, Iceland, Taiwan, and the Alps) and identified two common aspects of orographic influence: a generated pressure-gradient difference across the mountains, and a deflection of the air approaching the mountains with considerable wind increase on the left-hand side (facing downstream). Hoiinka (1980) also described the leftward turn of low-level winds upstream of a mountain range.

The present study aims at investigating the effect of the complex topography of the southern Aegean, on the observed wind field, based on the combined use of model simulations and remote-sensing data. The island of Crete has a relatively simple rectangular shape, with its main axis oriented perpendicular to the prevailing northern airflow during the Etesians. The Cretan topography is dominated by high mountains on the western and the central part of the island (with peaks ~ 2400 m), while lower mountains are found on the eastern part of the island. Off the eastern coast of Crete, there is a relatively narrow gap between Crete itself and Karpathos and Rhodes islands (Fig. 1) where the strongest winds are observed during the Etesians, a fact that so far has been attributed to the channelling of flow within this gap (e.g. Brody and Nestor 1985). The mountain ranges along Crete seem to disturb the regional airflow and pressure significantly, but this type of influence has received no attention. The scale of the mountains and the existence of water bodies both to the north and to the south of the island is such that the existing synoptic network is not, and cannot be, dense enough to resolve fully the orographic influence. In this work, data from coastal stations and ships are used, but the bulk of observational data is provided by wind measurements collected from the European Remote-Sensing Satellite (ERS-2) scatterometer, which include mesoscale details not resolved by the present operational observational network, as well as data coverage over large maritime areas devoid of surface observing stations (e.g. the maritime area around Crete). In addition, for one selected case of a typical Etesian wind episode, fine-grid numerical simulations have been performed, with the aid of the BOlogna Limited Area hydrostatic Model (BOLAM). The application of the model provides a better understanding of the interaction of the wind field with

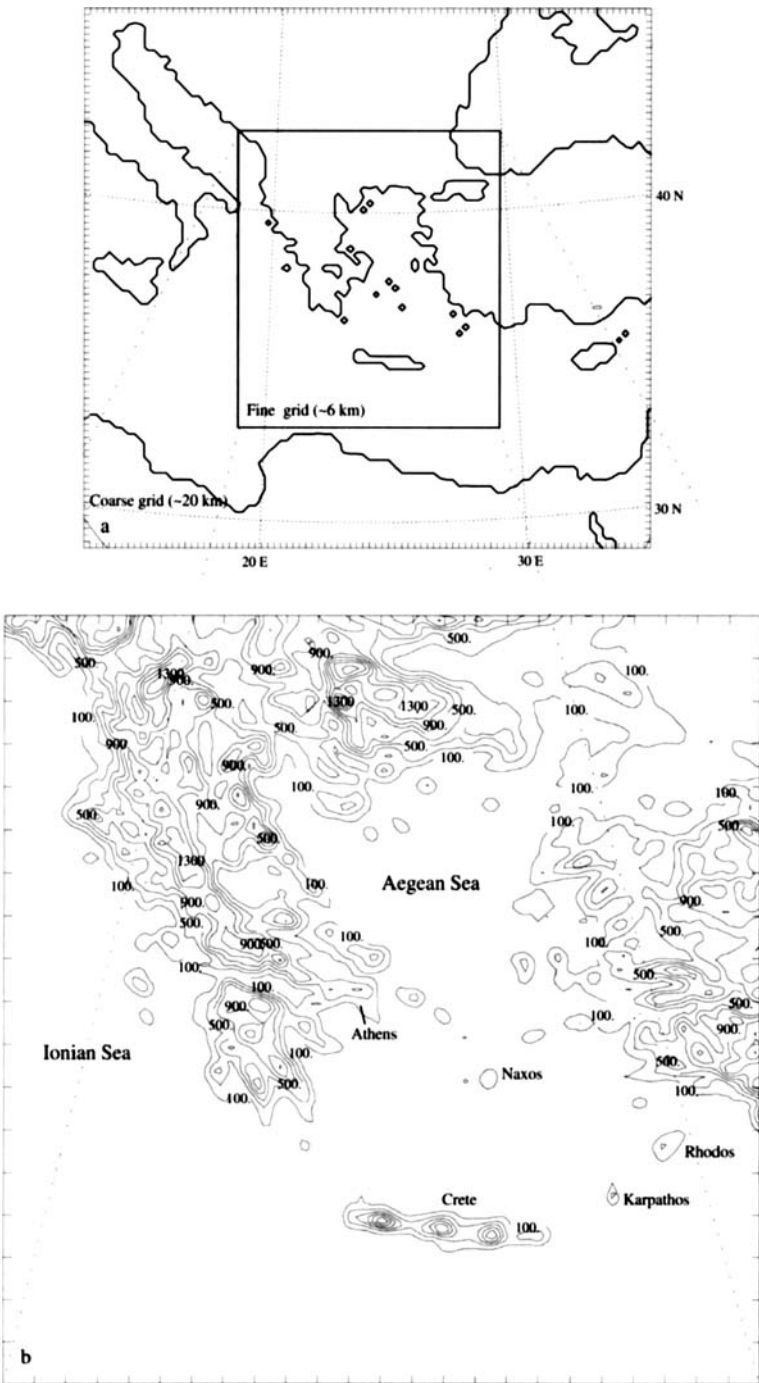


Figure 1. (a) Horizontal extension of the BOLAM coarse and fine grids. For each grid the approximate grid increment is shown. (b) Topography inside the BOLAM fine grid (at 200 m intervals, first contour line is at 100 m). The islands of Naxos, Crete, Karpathos and Rhodos, referred to in the text, are also shown.

the insular topography, leading to an improved description of the Etesian wind regime during summer.

The rest of this paper is organized as follows: section 2 presents surface wind observations from synoptic stations and ERS scatterometer data for four cases of Etesian winds for which ERS data were available and covered a significant part of the Aegean Sea. Section 3 outlines the main features of the numerical model used in this study, together with information about the settings of the various simulations that have been performed. The results of the simulations together with aspects of relevant theoretical results and estimations are presented in section 4, while appropriate conclusions are drawn and commented on in section 5.

2. SURFACE AND SATELLITE OBSERVATIONS

ERS-2 carries on board a scatterometer which is capable of providing surface wind data over water bodies. This instrument operates by recording the change in radar reflectivity of the sea due to the variation in backscatter from small ripples generated by the wind close to the surface. The energy in these ripples increases with wind speed and therefore the backscatter increases with wind speed. Inside the swath of the radar beam there is a regular grid of measurement points, the spacing of which is approximately 25 km across and along track. Measured winds are limited to the range $0.5\text{--}24\text{ m s}^{-1}$ and are referenced to a height of 10 m under neutral stability. Recent validation of wind speed and direction accuracy revealed root-mean-square differences of the order of 1.3 m s^{-1} and 12° for wind speed and direction, respectively (Quilfen 1992). A major problem that can arise when using wind scatterometer data is the directional ambiguity of 180° . This ambiguity can be removed easily by comparing the ERS wind directions with the European Centre for Medium-Range Weather Forecasts (ECMWF) 10 m wind direction interpolated to the grid points for the scatterometer winds (provided together with the scatterometer data).

A survey of all available ERS data since 1994 during Etesian episodes over the Aegean Sea was carried out. Four cases with good coverage of scatterometer data over the area were selected for further analysis. Figure 2 presents the surface wind fields at 0900 UTC 5 August 1997, 0900 UTC 6 September 1997, 0900 UTC 21 July 1998 and 2100 UTC 11 August 1998. Surface observations from island stations over the central and southern Aegean as well as from ships are also superimposed. It should be noted that the four panels of Fig. 2 constitute the most detailed observational evidence of the wind regime over the Aegean during Etesians available in the literature.

The major characteristics of the wind field as deduced by inspection of ERS data and synoptic observations are summarized below:

- The wind field presents a north-north-eastern direction over the north Aegean, turning northerly over the central Aegean and to a north-western direction over the south-eastern Aegean, which is the typical behaviour of the surface winds associated with the Etesians (Brody and Nestor 1985).
- The strongest wind intensity is observed over the central Aegean and over the maritime area east of the island of Crete (within the straits between Crete and Karpathos/Rhodos islands). This flow acceleration has been attributed by Brody and Nestor (1985) to the channelling of the flow between Crete and Karpathos/Rhodos. The validity of this assumption is discussed in section 4, based on the analysis of numerical simulations and sensitivity tests.
- Crete plays an important role in the flow modification both upstream and downstream of the island. The wind flow is clearly decelerated upstream of Crete, where the

wind speed is reduced by at least 2 m s^{-1} . The decreased surface wind is also reported by the synoptic surface stations located in the northern Cretan coasts (large wind symbols in Fig. 2). The deceleration zone reaches an upstream distance of about 120 km north of Crete. This feature is more evident upstream of the western part of the island (where the highest mountains are located, see Fig. 1) than upstream of the eastern part. On the leeward side of Crete, the northern flow is weaker overall, while there is evidence of a weak reverse flow in the wake of the mountains.

- There is a clear leftward (facing downstream) deflection of the upstream flow approaching Crete.

The aforementioned characteristics of the flow will be further discussed based on the analysis of model results, while the observations will be used further as a validation tool for the numerical simulations performed for one of the cases (21 July 1998).

3. MODEL AND SET-UP

For the present analysis, numerical simulations were performed using BOLAM, version 2000. BOLAM, version 2000, is a hydrostatic model based on previous versions of BOLAM described by Buzzi *et al.* (1997, 1998). The main features of the model are summarized below:

- hydrostatic primitive equations;
- dependent variables: surface pressure p_s , horizontal wind components u and v , potential temperature, specific humidity, and five microphysical variables;
- Arakawa C grid (rotated horizontal coordinates);
- new forward–backward (FB) 3-dimensional (3-D) advection scheme coupled with semi-Lagrangian advection of hydrometeors (Malguzzi and Tartaglione 1999);
- split–explicit time scheme (FB for gravity modes);
- 4th order horizontal diffusion and 2nd order divergence diffusion; and
- Davies–Kallberg–Lehmann (Lehmann 1992) relaxation scheme for lateral boundary conditions.

As it concerns the activated physical parametrizations, BOLAM includes:

- dry adiabatic adjustment;
- radiation: infrared and solar, interacting with clouds (Ritter and Geleyn 1992);
- vertical diffusion (surface layer and planetary boundary-layer parametrization) depending on Richardson number;
- surface thermal and water balance (three soil layers);
- explicit microphysical scheme (similar, in several aspects, to that proposed by Schultz (1995)) with five hydrometeors (cloud ice, cloud water, rain, snow, and hail/graupel); and
- convective parametrization scheme proposed by Kain and Fritsch (1990).

The model has the ability to perform one-way nested simulations. For that purpose, a first simulation is performed with a coarse grid interval and then the outputs of this coarse simulation are used as initial and boundary conditions on a subsequent run with finer grid resolution. For the purpose of this study, two one-way nested grids are used:

- the coarse nest with a grid of 90×84 points and 0.21° horizontal grid interval centred at 38°N , 24°E (approximately the position of Athens).
- the fine nest with a mesh of 140×160 points and 0.06° horizontal grid interval, with the same centre as the coarse simulation.

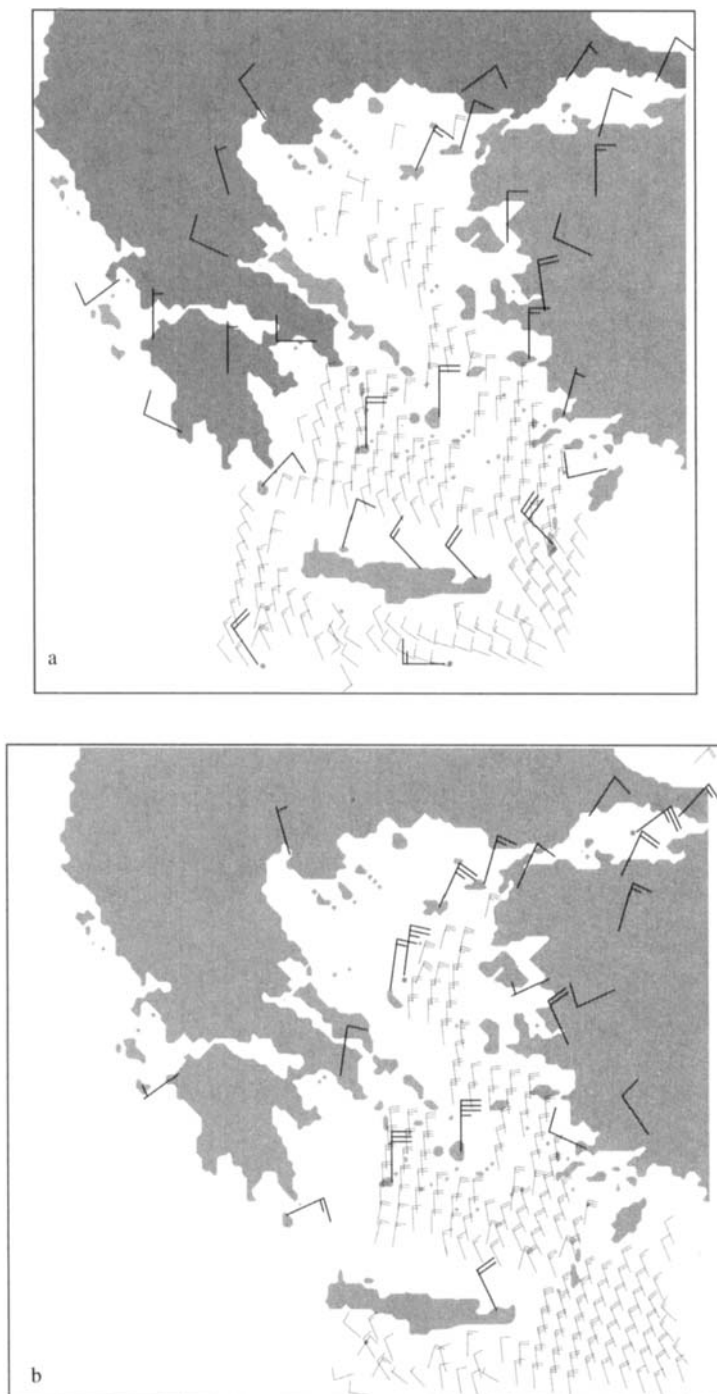


Figure 2. ERS-2 wind fields (one barb: 5 m s^{-1} , one half-barb: 2.5 m s^{-1}), valid at (a) 0900 UTC 5 August 1997, (b) 0900 UTC 6 September 1997, (c) 0900 UTC 21 July 1998, and (d) 2100 UTC 11 August 1998. Spacing between wind measurements is approximately 25 km. Large barb symbols denote observations from the synoptic network (ship reports are indicated by an asterisk).

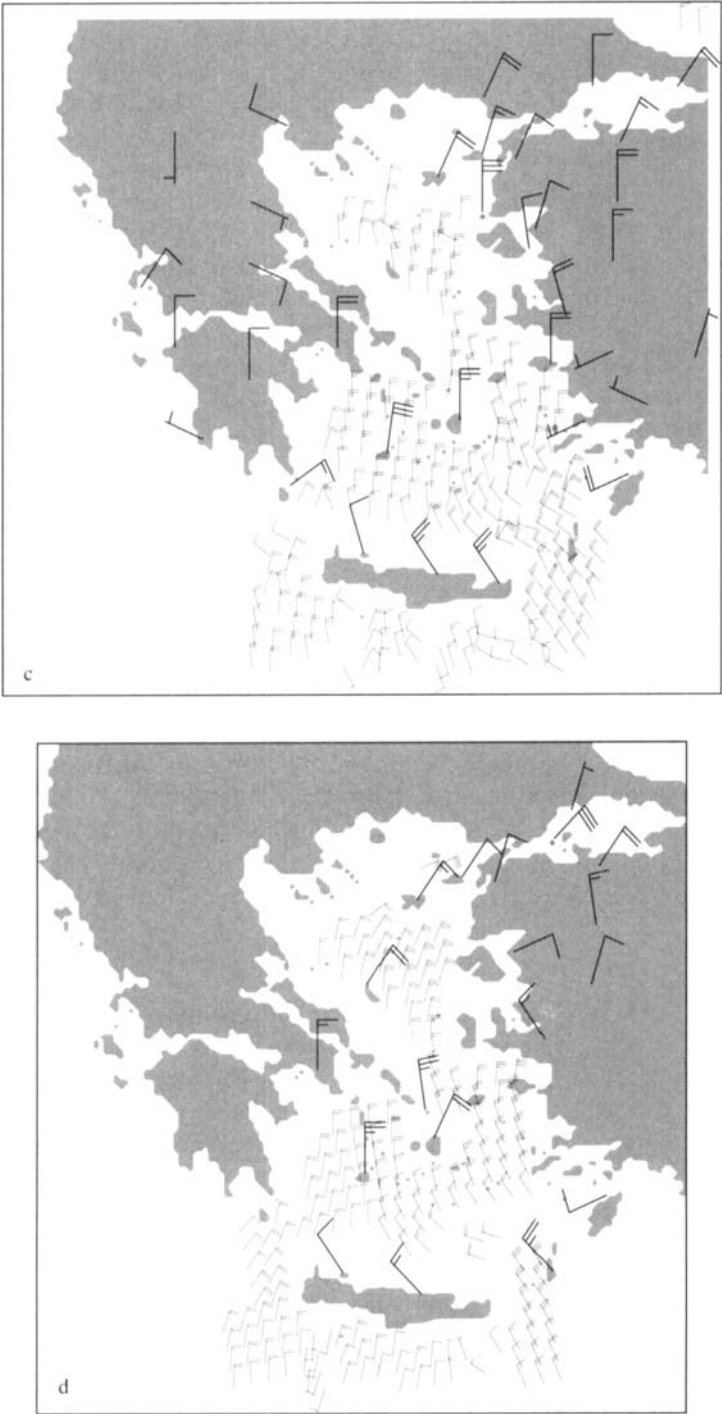


Figure 2. Continued.

The extent of both grids is shown in Fig. 1. In the vertical, 30 levels are used in the coarse nest and 40 levels in the fine nest, while model top on both nests is at about 10 hPa.

The ECMWF 0.5° latitude/longitude gridded analysis fields (at 6-hour intervals) were used to initialize the model and to nudge the boundaries of the coarse nest during the simulation period. These fields are objectively analysed by BOLAM on sigma levels (pressure/surface pressure) from which they are interpolated to the model grid points. Moreover, sea surface temperature of 0.5 degC resolution retrieved by the ECMWF, and topography derived from a 30 arcsec resolution terrain data have been used.

For the present study, BOLAM was initialized at 0000 UTC 19 July 1998 and the duration of the simulation was 72 hours. The choice of this Etesian wind event was based on the fact that this particular episode persisted for more than four days, and because of the existence of good quality ERS-2 data over the major part of the Aegean Sea. This will facilitate the comparison of model results with the observations.

The aforementioned set-up constitutes the control run (CONTROL hereafter). In order to test the role of barriers and channelling effects over the southern Aegean, two sensitivity tests were also performed: one with the removal of the topography of Crete, called NOCRETE hereafter, in order to investigate the role of the island on the upwind deceleration of the flow, and another with the removal of Rhodos and Karpathos topography (NORHODOS), in order to test the influence of channelling effects on the acceleration of surface winds over the sea east and south-east of Crete. In both sensitivity tests, the 'removed' islands existed as land distribution but their elevation was set to 0.5 m.

4. MODEL RESULTS

The Etesian wind episode studied in this work started on the morning of 18 July 1998, with northern wind reports exceeding 10 m s^{-1} over the Aegean. In the following, observations of sustained winds refer to the Naxos island station (see Fig. 1, also marked by an asterisk on Fig. 3), which is considered as the more representative station for wind reports in the Aegean. Maximum sustained winds of the order of 16 m s^{-1} were reported during the day hours on 21 July. The winds started to diminish in strength in the morning of 22 July.

(a) *Control run*

At 1200 UTC 19 July 1998, the Etesian wind regime was established over the Aegean. A low-pressure system was evident over southern Turkey, with an elongated low centre of 1006 hPa over south-west Turkey (Fig. 3(a)). This low-pressure system, often called the Turkish thermal low, is considered a western extension of the Asian thermal low (Meteorological Office 1962) and its persistency is related to the important ground heating over the area. Indeed, BOLAM 2 m surface temperature fields (not shown) give values as high as 38°C over southern Turkey, in very good agreement with the synoptic reports of 37°C at Mugla and 39.4°C at Antalya at the same time (denoted by letters M and A in Fig. 3(a), respectively). A strong east–west pressure gradient was evident over the Aegean, with a magnitude of the order of 10 hPa over a distance of 300 km. Synoptic reports from Naxos showed 14 m s^{-1} of sustained wind within the time period 0600–1200 UTC, with wind gusts reaching 18 m s^{-1} at 0900 UTC. At the 500 hPa level the geopotential pattern (Fig. 3(a)) reveals a weak trough over continental Greece. This configuration, according to Brody and Nestor (1985), is characteristic of the onset of an Etesian episode. At the 300 hPa level, geopotential heights were zonally oriented giving rise to a moderate wind from western directions over the area (not shown).

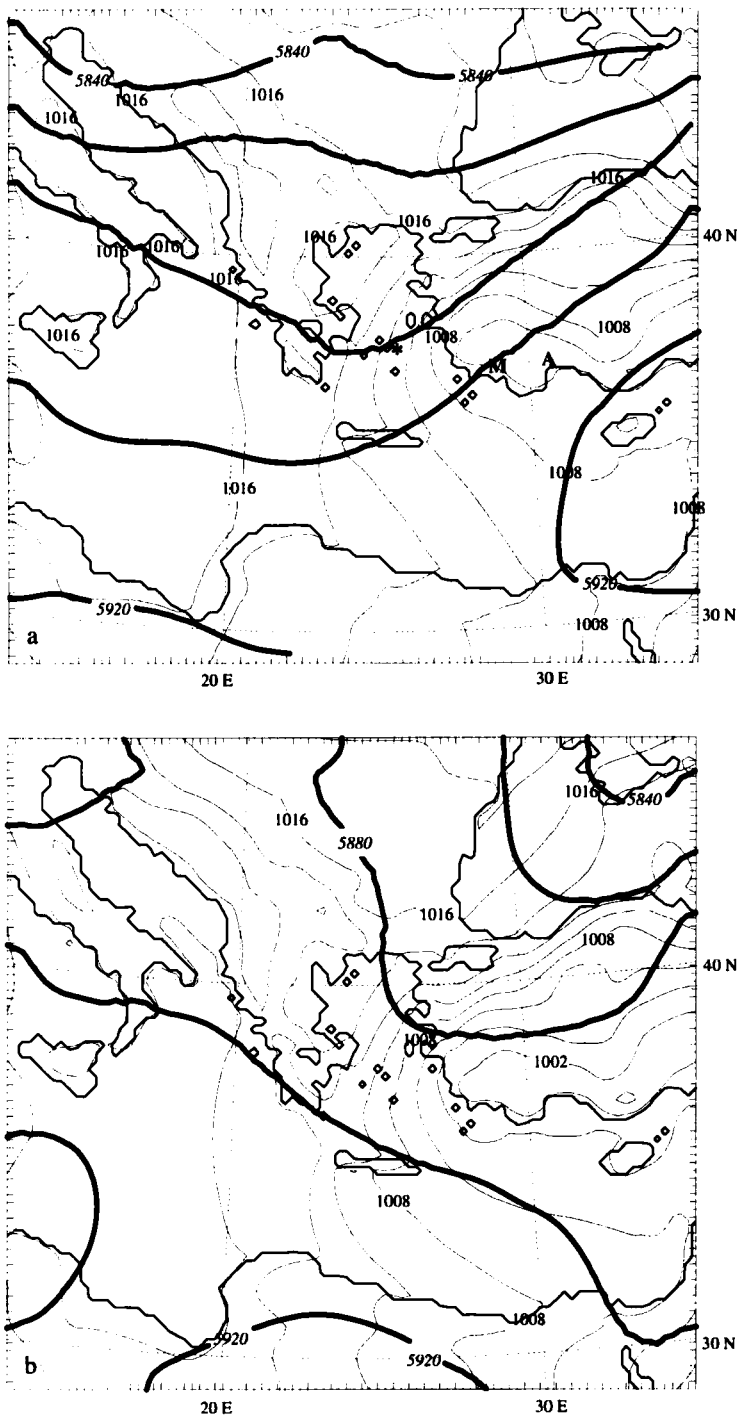


Figure 3. (a) BOLAM coarse-grid mean-sea-level pressure (at 2 hPa intervals, thin lines), and 500 hPa geopotential height (at 20 m intervals, bold lines) valid at 1200 UTC 19 July 1998, from the CONTROL run. (b) As in Fig. 3(a), except at 1200 UTC 21 July 1998.

Twenty-four hours later the situation had evolved, with the high-pressure system over the southern Balkans becoming more pronounced. The Turkish low had deepened to ~ 1004 hPa (1005 hPa was reported by Antalya station over south-west Turkey), while the pressure gradient over the Aegean had been slightly intensified.

At 1200 UTC 21 July (Fig. 3(b)), the low-pressure system over south-west Turkey had deepened to 1002 hPa, which is the lowest value attained during the simulation period. The pressure gradient exceeded 11–12 hPa over 300 km, while Naxos reported sustained winds of 14–16 m s^{-1} from 1200 to 1800 UTC, with wind gusts exceeding 21 m s^{-1} . The trough axis at the 500 hPa level was now evident over the eastern part of the domain, while a ridge had been established over the Ionian Sea. According to Brody and Nestor (1985), this synoptic configuration is characteristic of the beginning of the decay of the Etesian regime.

Further investigation of this Etesian episode is performed, based on data from the fine grid of simulation. The strongest surface winds were observed on 21 July, and our interest will be focused at 0900 UTC 21 July, due to the availability of an almost complete ERS-2 wind dataset over the Aegean (see Fig. 2(c)), which will facilitate the validation of model results. The mean-sea-level pressure field and 10 m wind from the fine grid at that time show the following characteristics (Figs. 4(a) and (b)):

- A pressure gradient of ~ 12 hPa, over a distance of ~ 300 km from the eastern coasts of continental Greece to the south-western Turkish coast, is evident over the Aegean Sea.

- A windward/leeward pressure difference is evident over Crete. The pressure is ~ 4 hPa higher on the upwind than on the leeward coasts of Crete. This behaviour of surface pressure is common in areas of airflow blocked by high mountains, where occasionally larger differences can be found (e.g. Smith (1982) reported a ~ 10 hPa difference between the windward and the leeward sides of the Alps).

- An upstream deceleration zone with a horizontal extent of ~ 120 km is clearly evident over the maritime area north of Crete. Within this zone, the wind decelerates from about 12.5 m s^{-1} to 7.5 m s^{-1} just offshore of Crete. This deceleration seems to be more pronounced upstream of the western part of the island where the highest mountains are located.

- The wind flow is deflected around Crete giving rise to strong winds (~ 12 –15 m s^{-1}) offshore of the western and eastern coasts. An asymmetry is evident in the upstream flow with a noticeable gradual leftward (facing downstream) turn of the wind offshore of the northern Cretan coast.

- A channelling of the wind flow within the Cretan valleys is evident offshore of the southern coast. At the exit of these valleys, the wind speed exceeds 18 m s^{-1} . There is also evidence of areas of a reversed weak southern flow to the lee of the high mountains of Crete, implying that the wake of the island consists of three pairs of small-scale counter-rotating vortices. This is noticed in the ERS-2 data as well. This flow reversal is rather shallow, as there seems to be no signs of it at the 850 hPa level. A similar, but not as clearly defined, situation appears in the area south-west of Peloponnisos.

The characteristics of the 10 m simulated wind flow shown in Fig. 4(b) and discussed in the previous paragraph are in very good agreement with the station and ERS observations (Fig. 2(c)). The strongest winds are reproduced by the model over the north Aegean (where both observations and model-predicted winds are of the order of 12.5 m s^{-1}), between the numerous islands in the central Aegean (with winds of the order of 12.5–15 m s^{-1} , note the excellent agreement of the model and the Naxos reported wind) and over the maritime areas west of Crete as well as over the area

between Crete and Karpathos. It should be noted that the model reproduces very well the observed (by ERS) deceleration of the flow north of Crete, the leftward (facing downstream) deflection of the upstream flow, and also the wake south of Crete.

The wind field at the 850 hPa level is shown in Fig. 4(c). It differs substantially from the surface flow (Fig. 4(b)) mainly over the south-eastern part of the domain. The strongest wind flow is found over the western part of the Aegean Sea, having a quasi-steady north-easterly direction and reaching an intensity of 17 m s^{-1} over the south-western Aegean (shaded areas in Fig. 4(c)).

It is instructive here to examine the modulation of the flow in the area of Crete, in the light of previous investigations which were based on analytical solutions and numerical simulations of wind flows over idealized mountains shapes. Smith (1982) presented analytical solutions of wind flow over isolated mountains in a rotating atmosphere, where he clearly demonstrated the asymmetric deflection of horizontal trajectories of air parcels due to the earth's rotation. His Fig. 10, shows the streamlines turning leftward (facing downstream) from the obstacle and then packing on the left of the elongated mountain, a feature which is also evident on the BOLAM 10 m wind field shown in Fig. 4(b). This asymmetry is attributed to the decreased Coriolis force experienced by the approaching air parcels within the upstream deceleration zone and thus to the leftward deflection of the air parcels by the large-scale pressure gradient. A discussion on the balance of forces inside the deceleration zone is provided in section 4(d). In addition, the sea-level isobaric pattern (shown in Smith's Fig. 11) exhibits a characteristic windward/leeward pressure difference, which is in striking resemblance with the BOLAM surface isobars depicted in Fig. 4(a).

Olafsson and Bougeault (1997) presented, through numerical simulations of a flow over a 3-D elliptical mountain, the significant effect of the Coriolis force on the intensification of wind on the left side of the mountain. Their Fig. 1 shows very well this asymmetry, giving greater wind speeds on the left (facing downstream) than on the right side of the mountain. Similar results regarding the asymmetry of the flow around an obstacle have also been reported by Sun and Chern (1994) and Peng *et al.* (1995).

The wind flow in the presence of complex topography is modulated by two free parameters (Pierrehumbert and Wyman 1985; Chen and Smith 1987; Smith 1989), the Froude number Fr , defined by $U/(Nh_m)$ and the Rossby number Ro , defined by $U/(fl_m)$, where U is the wind speed normal to the barrier, h_m is the height of the barrier, f is the Coriolis parameter, l_m is the half width of the barrier, N is the Brunt-Väisälä frequency (equal to $(g/\theta_0)\partial\theta/\partial z$, where θ is the potential temperature, θ_0 a reference value and g is gravity). In the rotating case, these two parameters are usually used to describe the atmospheric response to flow impinging upon mesoscale mountains. For certain values of Fr and Ro , blocking and deceleration of the flow appear, the threshold values being $Fr < 1$ for blocking and $Ro > 1$ for upstream deceleration. As shown by Pierrehumbert and Wyman (1985), if rotation is present, the deceleration zone upstream of the mountain is evident only over a limited distance upstream, of the order of the Rossby radius of deformation, equal to Nh_m/f .

Thus, in order to quantify the effect of the orographic influence on the wind flow and also to provide some basis of comparison with analytical solutions of upstream effects of mesoscale mountains, the Froude number, as well as the Rossby number, have been calculated based on BOLAM fine-grid data at 0900 UTC, 21 July. The calculated Brunt-Väisälä frequency has been evaluated from the BOLAM model potential-temperature field in the area north of Crete to a mean value of $\sim 10^{-2} \text{ s}^{-1}$, which with a mean flow (evaluated by the model) of the order of 10 m s^{-1} , a mountain height of

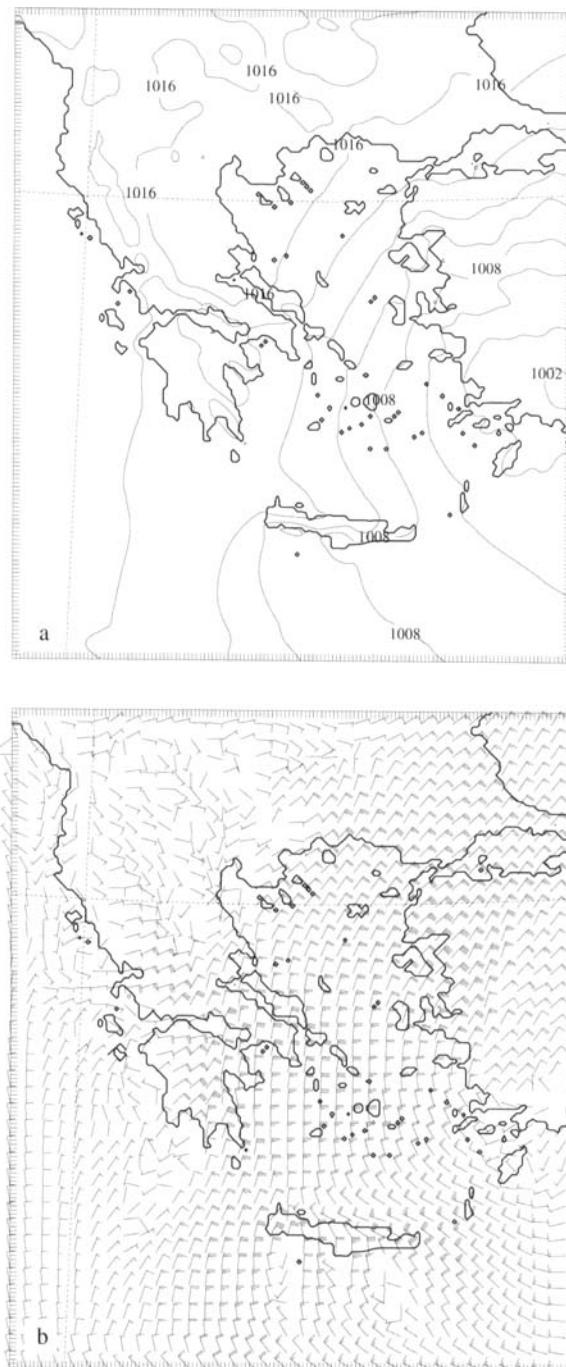


Figure 4. (a) BOLAM fine-grid mean-sea-level pressure (at 2 hPa intervals), valid at 0900 UTC 21 July 1998, from the CONTROL run. (b) As (a) but for the 10 m wind field (one barb: 5 m s^{-1} , one half-barb: 2.5 m s^{-1}). Wind symbols are plotted every fourth grid point. (c) As (b), but for the 850 hPa wind field. Lines CC' and NN' denote the positions of the cross-sections shown in Figs. 5(a), and 5(b), respectively. Shaded areas correspond to wind speeds exceeding 17 m s^{-1} .

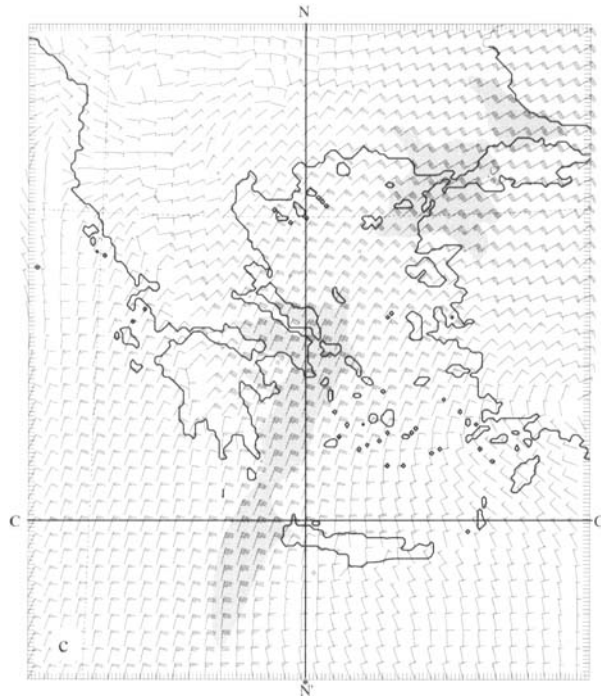


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1300 m (which is the resolved height of the mountains of Crete at the $6 \text{ km} \times 6 \text{ km}$ resolution) and a mountain half width $l_m \sim 20 \text{ km}$ gives $Fr = 0.77$ and $Ro = 5$. This implies that the parameters associated with the upstream flow, and thus the energy conditions, are theoretically favourable for a partial blocking of the upstream flow (as $Fr < 1$), and for a considerable deceleration of the flow (as $Ro > 1$) upstream of the mountain barrier, to occur. Analytical solutions by Pierrehumbert and Wyman (1985) give a maximum upstream extent of the decelerated flow of the order of Nh_m/f , which in our case (with $N \sim 10^{-2} \text{ s}^{-1}$ and $h_m \sim 1300 \text{ m}$) gives a calculated upsteam extent of the decelerated zone of $\sim 130 \text{ km}$. This value is consistent with the 120 km extent of the decelerated upstream flow as evidenced in both the simulated and ERS wind fields (Fig. 4(b), and Fig. 2(c), respectively).

Numerical experiments for a rotational flow past an elliptical mountain performed by Sun and Chern (1994) produced lee-side vortices extending $\sim 120 \text{ km}$ downstream from the obstacle, while an observational study of the wake of Hawaii island by Smith and Grubisic (1993) revealed that the reversed flow along the wake axis extends to a distance of almost 200 km. The model-simulated wind flow over Crete revealed the existence of counter-rotating vortices to the lee of the mountains, extending to about 100 km southwards, in general agreement with the aforementioned studies.

The vertical structure of the wind can be better investigated through the inspection of cross-sections which are provided in Fig. 5. Figure 5(a) shows an east–west cross-section just north from Crete (following the bold line CC' in Fig. 4(c)), at 0900 UTC 21 July. The strongest winds are found in the boundary layer in the areas both east and west of Crete (exceeding 12 m s^{-1} inside the surface–950 hPa layer), while strong winds exceeding 16 m s^{-1} are found in the 950–850 hPa layer, west of Crete, as this was also

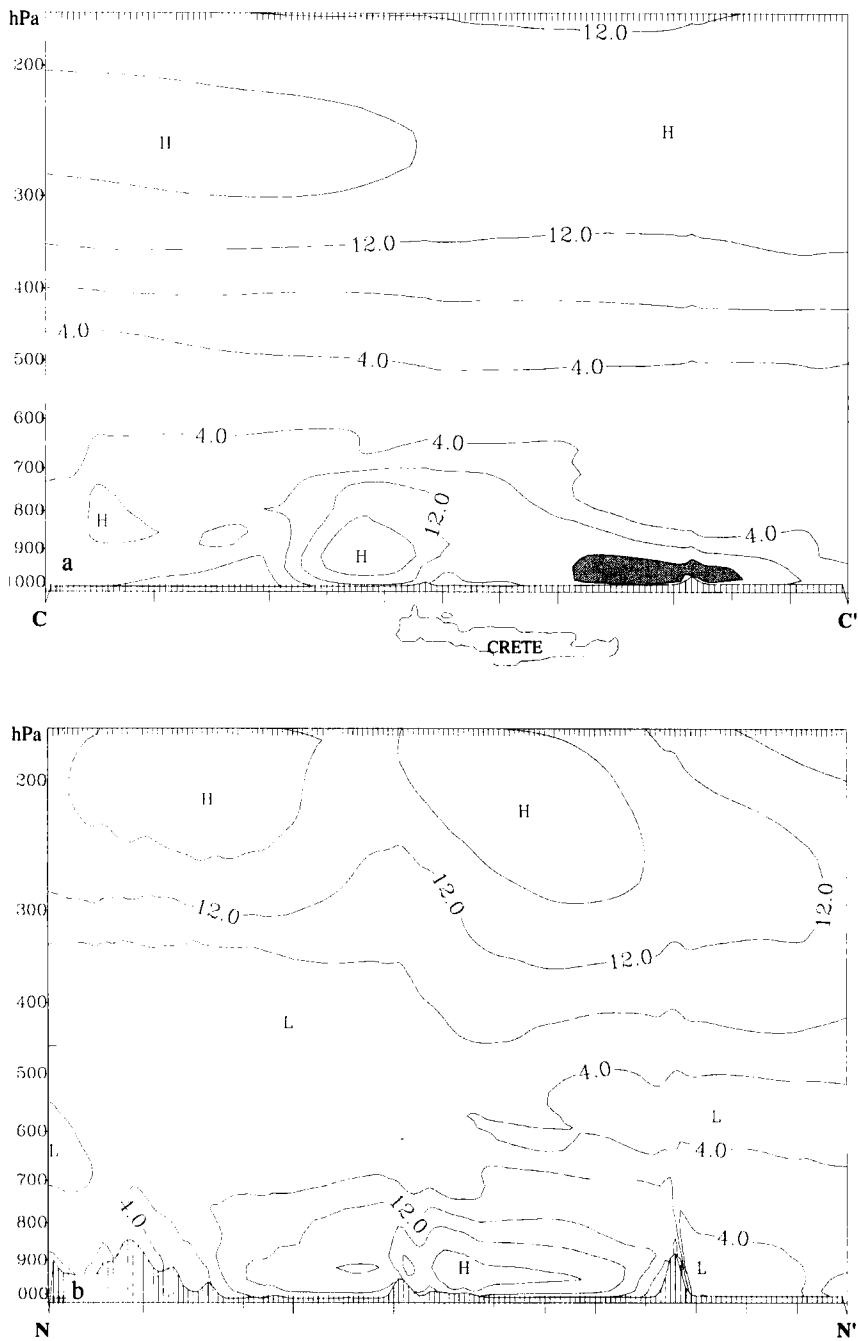


Figure 5. (a) Vertical cross-section (following line CC' shown in Fig. 4(c)) of horizontal wind speed (at 4 m s^{-1} intervals), from the fine-grid of BOLAM, at 0900 UTC 21 July 1998. The shaded area east of Crete represents wind speed greater than 12 m s^{-1} . (b) As (a), but following the line NN' shown in Fig. 4(c).

revealed by the 850 hPa wind flow depicted in Fig. 4(c). The strong wind shear in the surface–900 hPa layer west of Crete, strongly suggests the existence of a downward transport of momentum in this area. The important feature revealed in this figure is the closed contour of 12 m s^{-1} east of Crete which is rather shallow and is not supported further above, but seems to be caused by the presence of Crete. Indeed, inspection of a cross-section located 100 km northwards (not shown) reveals the presence of a well-formed jet crossing the Aegean in a NE–SW direction, as also evidenced in the 850 hPa field (shaded areas in Fig. 4(c)) and also a sharp reduction of the wind at that level north from the eastern tip of Crete.

The blocking of the upstream flow by the mountains is better assessed through inspection of a north–south cross-section, crossing the high mountains of western Crete (Fig. 5(b)). The wind speed decreases by $\sim 8 \text{ m s}^{-1}$ over a distance of 50 km upstream of the obstacle while to the lee of the mountain a wake is formed with a very weak wind speed ($< 4 \text{ m s}^{-1}$).

(b) *NORHODOS run*

The very strong surface winds observed over the maritime area east of Crete during an Etesian episode has been attributed by Brody and Nestor (1985) to the channelling between the eastern coasts of Crete and the neighbouring island of Karpathos. This concept (also widely accepted by Greek forecasters) needs to be re-evaluated in the light of the findings presented in section 4(a). In order to investigate to what extent the acceleration of the flow east of Crete is due to channelling or to flow deflection, or both, an identical experiment (NORHODOS) to the CONTROL experiment was conducted.

Comparison of the NORHODOS run against the CONTROL run revealed that the differences in most fields are insignificant. The only significant difference is found around Rhodes island, where the lee effect is now eliminated and the wind is about 2 m s^{-1} stronger on the south-eastern coasts of Rhodes in the NORHODOS run compared with the CONTROL run (not shown). Thus the presence of Rhodes and Karpathos is seen to have no effect on the wind-flow acceleration east of Crete, and the channelling within the Karpathos strait is shown to be of minor importance.

(c) *NOCRETE run*

Results of both the CONTROL and NORHODOS simulations support the crucial role of the mountainous island of Crete on both the upstream and downstream wind flow during an Etesian episode. To further ascertain this role, a sensitivity test (NOCRETE) was also performed.

Figure 6(a) presents the mean-sea-level pressure while Fig. 6(b) presents the 10 m wind at 0900 UTC 21 July 1998, from the fine grid of the NOCRETE run. Comparison with the CONTROL fields (Figs. 4(a) and 4(b)) shows that the pressure field over the central Aegean Sea remains unmodified in the NOCRETE run, while significant differences are found in the area around Crete. The windward/leeward pressure difference over Crete is now absent, while the wind flow is northerly over the whole Aegean Sea, with the exception of the cyclonic turn around the thermal low in southern Turkey.

The impact of the topography of Crete on the near-surface wind is clearly illustrated in Fig. 6(c), where the difference between the 10 m wind from the CONTROL and NOCRETE runs is plotted. A wind-flow deceleration of -2 m s^{-1} is evident at a distance of $\sim 120 \text{ km}$ upstream of Crete, reaching a value of -8 m s^{-1} just upstream of the high mountains of western Crete. On the leeward side of the island, a succession of negative and positive values of wind difference is evident, reflecting the blocking to the lee of the

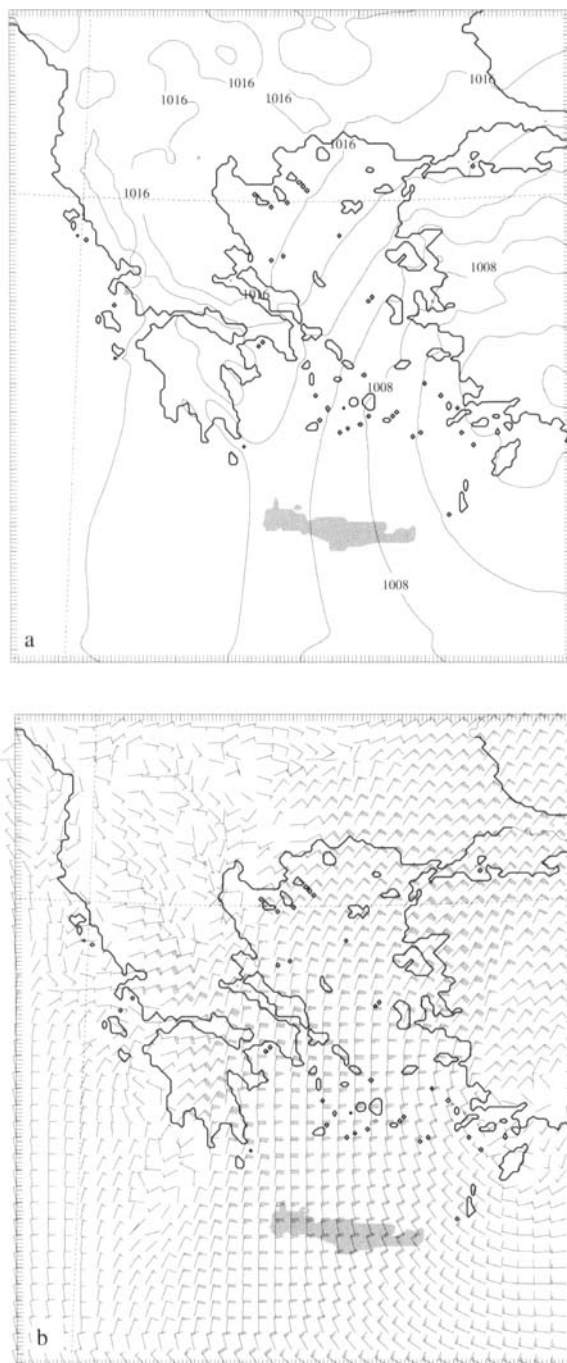


Figure 6. (a) BOLAM fine-grid mean-sea-level pressure (at 2 hPa intervals), valid at 0900 UTC 21 July 1998, from the NOCRETE run. The position of Crete is shown in grey for orientation purposes. (b) As in (a), but for the 10 m wind field (one barb: 5 m s^{-1} , one half-barb: 2.5 m s^{-1}). Wind symbols are plotted every fourth grid point. (c) The difference of 10 m wind speed between the CONTROL run and the NOCRETE run (at 2 m s^{-1} intervals). Dotted lines denote negative values (stronger winds on the NOCRETE run), while shaded contours denote positive values (stronger winds on the CONTROL run, scaled as follows: light shading: 2 m s^{-1} , medium shading: 4 m s^{-1} , dark shading: 6 m s^{-1} , black shading: 8 m s^{-1}).



Figure 6. Continued.

main mountain masses, the channelling of the wind flow inside the north–south oriented valleys, and the acceleration around the edges of the island.

The most prominent difference is observed over the eastern part of the island, where the wind speed in the CONTROL run is $\sim 8 \text{ m s}^{-1}$ stronger than in the NOCRETE run, indicating thus the strong impact of the topography of Crete on the modulation of the wind flow over the south-eastern Aegean.

(d) *Balance of forces at the surface*

It is of interest for a better understanding of the dynamics of the event to provide an analysis of the balance of forces at different points over the maritime area around Crete. This is accomplished through the computation of the horizontal component of the momentum equation:

$$\frac{d\mathbf{v}}{dt} = -f\mathbf{k} \times \mathbf{v} - \frac{1}{\rho}\nabla p + \text{friction} \quad (1)$$

where $\mathbf{v}$ is velocity, f the Coriolis force, ρ density, p pressure and $\mathbf{k}$ is the unit vector.

The three terms on the right-hand side of Eq. (1) represent accelerations resulting from the Coriolis force, the pressure-gradient force and friction, respectively. For the calculations based on model grid data, the total parcel acceleration is computed as a sum of the local and advective parts and friction is finally computed as a residual. This residual is a combination of the errors in computing all the other terms and of the physically significant stresses exerted on the flow.

Figures 7(a) and (b) show the horizontal force vector balance for the surface flow, computed by the fine-grid model data, at 0900 UTC 21 July 1998, at selected places

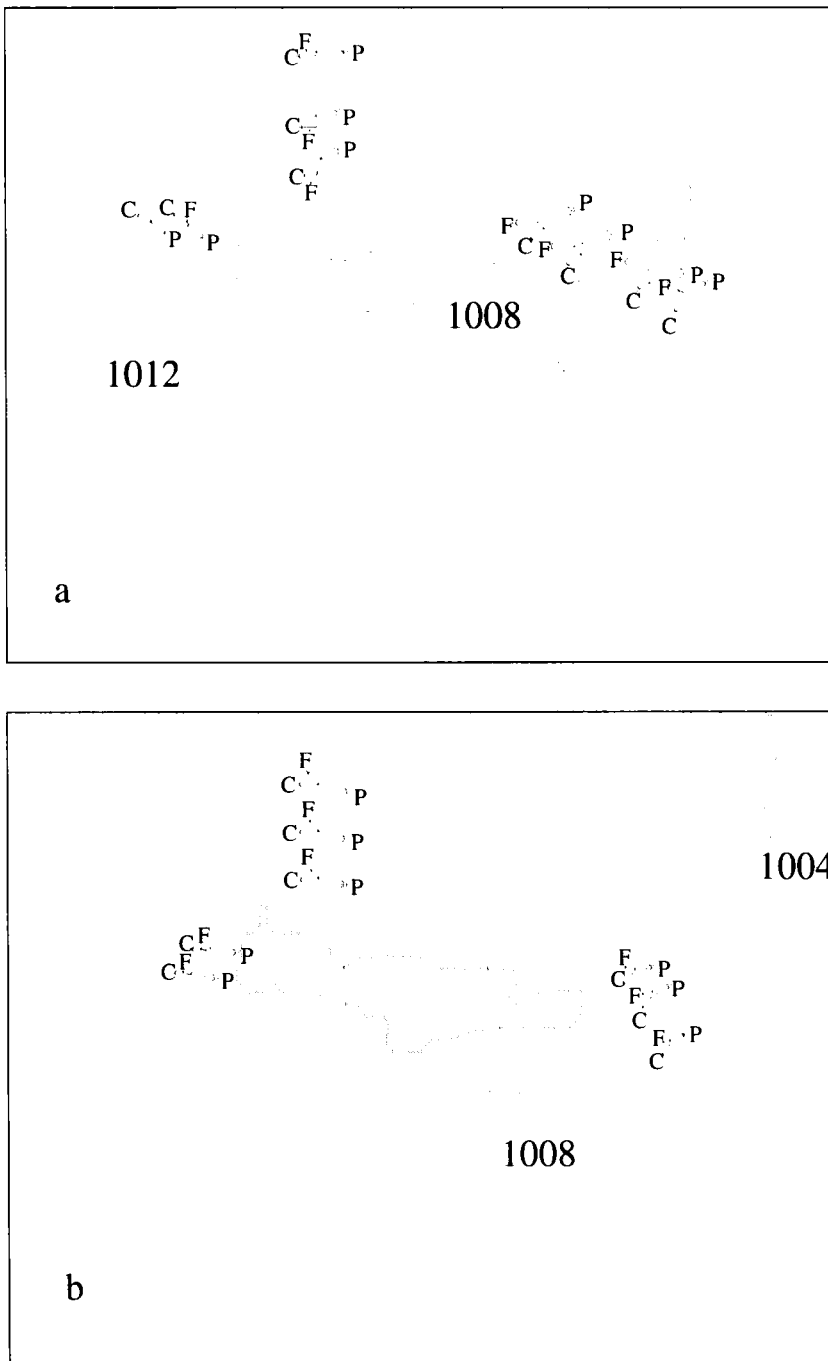


Figure 7. (a) Sea-level pressure (at 2 hPa intervals) and vector force balance of the surface flow at selected places, valid at 0900 UTC 21 July 1998, calculated using gridded data from the fine grid of the CONTROL run. Pressure-gradient force, Coriolis force and friction are labelled P, C and F respectively, while the unlabelled vector represents the parcel acceleration (vector magnitudes are scaled by the arrow plotted south of Crete, which is equal to 10^{-2} m s^{-2}). (b) As in (a), but for the NOCRETE run.

around Crete from the CONTROL and the NOCRETE run. North of Crete there is a cyclonic turning and increase of the pressure-gradient force due to the accumulation of air against the mountains, in the CONTROL run as compared with the NOCRETE run. The pressure-gradient force is sending the flow to the east as the decelerated flow 'feels' a decreased Coriolis force (note how the Coriolis force decreases from north to south approaching the northern coast of western Crete (Fig. 7(a)). The decelerated flow 'feels' also a decreased friction force compared with the NOCRETE run, while the acceleration term acts opposite to the flow and increases southwards.

In the area east of Crete, the acceleration term acts almost in the direction of the wind flow in the CONTROL run, while this term almost vanishes in the NOCRETE run. The accelerated flow results in an intensification of the friction term exerted on the air parcels. The direction of the friction force being almost opposite to the wind flow implies that the surface drag is the most important factor contributing to this term. Further south, there is a clear balance of the pressure-gradient force by the Coriolis force in the cross-flow direction, while the acceleration term is not significant. In the NOCRETE run, it is obvious that the weaker flow feels a weaker Coriolis force over this area.

5. CONCLUDING REMARKS

This study aims to provide some insight on the nature of the Etesian winds over the Aegean, and in particular to contribute to the understanding of the interaction of the wind flow under an Etesian wind regime with the main insular topographic feature of the Aegean Sea, the island of Crete. To achieve these objectives, remote-sensing data, surface observations, and fine-resolution numerical simulations have been used. Wind scatterometer data from the ERS-2 polar-orbiting satellite provided observational descriptions of a number of Etesian cases which occurred during recent years, giving information of unprecedented spatial resolution over the Aegean. Together with the available surface and ship synoptic data, ERS wind data helped to establish the repeatability of the main Etesian patterns, to identify the regions over the Aegean where the wind reaches its maximum intensity, as well as to pinpoint the influence of Crete on the modulation of the wind flow during the Etesians.

The observational dataset thus revealed three important characteristics: (i) the formation of a deceleration zone upstream of Crete; (ii) the deflection of the wind flow around the island, with the most prominent feature the asymmetrical deflection and intensification of the surface flow offshore its eastern coasts; and (iii) the formation of a wake in the lee of the island.

These characteristics of the surface wind flow have been further investigated through the analysis of model results provided by the BOLAM hydrostatic model. The numerical simulation of a selected Etesian episode reproduced well the characteristics of the surface flow as identified by the observational analysis. Indeed, the Crete mountain ranges surrounded by water provide an excellent example of a major isolated topographic feature forming a barrier to the northern wind field thus affecting both the upstream and downstream evolution of the flow. The energetics of the wind flow with an estimated Froude number of 0.77 favours strong flow retardation. Estimation of the Rossby number with a value of 5 implies that there is a significant deceleration of the flow upstream. The decelerated zone, following Pierrehumbert and Wyman (1985), extends at an upstream distance of the order of a Rossby radius of deformation, which in our case equals 130 km. This calculated distance is consistent with the 120 km extension of the decelerated surface wind flow evident in both the model and the observed wind

fields. Inside this zone of decelerated flow the air parcels feel a decreased Coriolis force, and consequently the pressure-gradient force deflects the flow to the east of Crete.

Numerical sensitivity tests revealed that the intensification of the flow east of Crete is related only to the interaction of the flow with the mountainous complex terrain of the island. Indeed, comparison of the CONTROL run with the NOCRETE run showed that the presence of Crete accounts for an important acceleration of the flow east of the island with a maximum wind speed increase of the order of 8 m s^{-1} near the islands eastern coasts, while the accelerated wind flow extends to about 200 km east of the island. The sensitivity test where the topography of Karpathos and Rhodos islands was excluded showed no difference when compared with the CONTROL run.

Although this paper helped to clarify our understanding of the interaction of the flow with the major obstacle of Crete under the Etesian wind regime, further analysis is required in order to understand better the organization of the flow within the Aegean Sea. More specifically, there is a need to investigate more thoroughly the role of channelling inside the narrow straits between the numerous islands of the Aegean, and the possible generation of an effective roughness because of them. This is considered of vital importance because modern society is demanding high quality and detailed weather information at the local scale for all human activities, including navigation. This task can be fulfilled mainly by the use of very-high-resolution non-hydrostatic modelling, and it is the authors' intention to continue this research to meet these objectives.

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